## 2.5.1 Basic Geologic and Seismic Information

The geologic and seismic information presented in this subsection provides a technical basis for evaluating potential geologic hazards at the Units 6 & 7 site. This subsection summarizes the current physiography, geomorphic processes, stratigraphy, tectonic features, stress regime, and the geologic history of the region within a 200-mile (320-kilometer)[1] radius of the site. This area is known as the site region (Figure 2.5.1-201). This subsection also provides similar information about the active plate boundary between North America and the Caribbean Plates located south of the site region. Both local and distant sources contribute to the seismic hazard at the site, including sources associated with the North America-Caribbean Plate boundary, whose closest approach is about 420 miles (675 kilometers) south of the Units 6 & 7 site (Subsections 2.5.1.1.2.2 and 2.5.1.1.2.3).

Subsection 2.5.1.1 describes the regional geology. Subsection 2.5.1.1.1 contains descriptions of the geologic and tectonic characteristics of the 200-mile radius site region. Information describing the geologic and seismic characteristics beyond the 200-mile radius site region is included in Subsection 2.5.1.1.2. The description of characteristics beyond the site region focuses on the North America-Caribbean Plate boundary, including potential seismic and tsunami sources in the Gulf of Mexico and Caribbean that may impact the Units 6 & 7 site. Subsection 2.5.1.2 describes the geologic and tectonic characteristics of the site vicinity, site area, and the site.

This subsection demonstrates compliance with the requirements of 10 CFR 100.23 (c). The geologic and seismic information was developed in accordance with NRC guidance documents RG 1.206 and RG 1.208.

The following paragraphs comprise a brief overview of the geologic evolution of the North America-Caribbean Plate boundary region and are intended to provide a context for more detailed discussions of available data presented in subsequent subsections.

### Overview of Tectonic Evolution

The Units 6 & 7 site is located on the southern portion of the North America Plate along the Atlantic passive margin, approximately 420 miles (675 kilometers) north of the North America-Caribbean Plate boundary (Figure 2.5.1-202). The North America Plate has been through multiple cycles of plate-tectonic spreading and convergence, resulting in the opening and closure of ocean basins. The cycles, with a period generally in the 100-million-year range, are known as Wilson cycles (References 201 and 202). The following provides a basic overview of the regional tectonic evolution of an expanded site region as a context for subsequent discussions of physiography, stratigraphy, structures, and seismicity of specific parts of the larger region.

#### Paleozoic Wilson Cycle

Most of Earth's landmass amalgamated into a supercontinent, Rodinia, between 1300 and 900 million years ago (Ma) (e.g., Reference 203) (Figure 2.5.1-203) (Subsection 2.5.1.1.3.1). Beginning about 750 Ma, rifting of Rodinia formed a vast proto-Atlantic ocean known as the Iapetus ocean that separated the paleocontinents of Laurentia (ancestral North America) and Gondwana (including ancestral Africa, South America, and Florida). The final closing of this ocean late in the Paleozoic (325 to 250 Ma) led to the formation of the supercontinent Pangea (Reference 204) (Figure 2.5.1-204). This ocean closure occurred in a series of three primary collisions known as the Taconic, Acadian, and Alleghany orogenies. These tectonic events led to the deformation of a belt of rocks that in present-day North America extends from Newfoundland to Alabama and as far west as Oklahoma and Texas. This belt of deformed rocks is known as the Appalachian-Ouachita orogen (Reference 205). The final deformation event related to the closure of the intervening oceans, the

1. The norm applied throughout FSAR Subsections 2.5.1, 2.5.2, and 2.5.3 regarding the presentation of English or metric units of measure is to present measurements in the units cited in the reference first, then to provide the conversion in parentheses. In general, the conversion of units is an approximation that reflects the significant figures of the original units.

late Mississippian-Pennsylvanian (Carboniferous)-Permian Alleghany orogeny, occurred more than 600 miles (~1000 kilometers) north of the Units 6 & 7 site in the southern Appalachians (Figure 2.5.1-204). Nevertheless, the Alleghany orogeny is significant because it transferred some of the crust from Gondwana to North America, in particular, the basement of the Florida and Bahama Platforms (e.g., Reference 206) (Subsections 2.5.1.1.1.2.1 and 2.5.1.1.1.2.2). This material was welded against North America along the Suwannee suture, located in southern Georgia and Alabama (References 207, 208, and 209) (Figure 2.5.1-205).

Mesozoic Break-up of Pangea

The supercontinent Pangea remained a single landmass for almost 150 m.y., from late Paleozoic (about 325 Ma) until the late Triassic-Early Jurassic (225 to 175 Ma). The earliest vestiges of rifting included widespread thinning of the underlying continental basement and intrusion of volcanics into rift valleys during the Triassic. By the Middle Jurassic (170 Ma), the supercontinent was rifted into discrete landmasses (e.g., Reference 209) (Figure 2.5.1-206) (Subsection 2.5.1.1.3.2), with present day North America moving away from present day South America and Africa. By the Late Jurassic (150 Ma), the Atlantic Ocean and Gulf of Mexico were fully opened. Prior to this rifting, the Yucatan block was located adjacent to the Texas coast, while basement underlying southern Florida may have been located farther northwest between the African and North America Plates (Reference 210) (Figure 2.5.1-206) (Subsections 2.5.1.1.2.1, 2.5.1.1.2.2, and 2.5.1.1.2.3). These blocks were transported out of the Gulf of Mexico Basin during the Mesozoic (Subsection 2.5.1.1.3.2) via rifting and/or strike-slip faulting. During the Late Triassic to Middle Jurassic, rifting and sea floor spreading formed oceanic crust in the central Gulf of Mexico while leaving the Suwannee terrane, a piece of Florida basement, stranded adjacent to the Laurentian margin (Reference 211). The Early Jurassic volcanic rocks beneath southern Florida are interpreted as transitional crust developed during this rifting (Reference 212). Marine seismic reflection profiles indicate that subsurface normal faults were active during the Middle Jurassic to Cretaceous and are interpreted to reflect this regional rifting event (Reference 211). By the Early Cretaceous, shallow water evaporites and carbonates were deposited over most of the Florida, Yucatan, and Bahama Platforms (Figures 2.5.1-207 and 2.5.1-208) (see discussion of carbonate platforms in Subsection 2.5.1.1.1.1.1.2) while the Gulf of Mexico and Atlantic coasts developed as a passive continental margin (Reference 213).

Cretaceous to Tertiary Caribbean-North America Plate Convergence

The plate tectonic regime changed in the Cretaceous when the Caribbean Plate impinged on the southern North America Plate from the west (Reference 214). An alternative hypothesis suggests in situ formation for the Caribbean Plate (Reference 608). In the Cretaceous, portions of Cuba were part of the Caribbean Plate and experienced calc-alkaline volcanism associated with the Greater Antilles volcanic arc as the North America Plate subducted southwestward beneath the Caribbean Plate (Subsection 2.5.1.1.1.3.2.4). This island arc generally was active in the Cretaceous, but volcanism occurred in different places and at different times as multiple discrete volcanic events (References 216 and 845). During subduction and subsequent collision, many of the Jurassic- to Cretaceous-age volcanic and sedimentary strata currently exposed in central and western Cuba underwent high-pressure, low-temperature metamorphism. This metamorphism and accompanying ophiolite emplacement occurred in the mid-Cretaceous to Paleogene (References 217 and 218).

Tertiary Transfer of Cuba to the North America Plate

During the early Eocene approximately 50 Ma, the Greater Antilles volcanic arc, with a westward-dipping subduction zone, approached the Bahama Platform and subduction ceased, resulting in the transfer of Cuba from the Caribbean Plate to the North America Plate (Reference 219) (Figure 2.5.1-206) (Subsection 2.5.1.1.3.3). In western Cuba, the collision between the Caribbean and North America Plates occurred in late Paleocene to early Eocene time (Reference 220). North- to northeast-directed thrusting and contraction related to this collision was widespread in the late Paleocene to middle Eocene along the Cuban fold-and-thrust belt. A southward dip in middle Eocene and older strata in the Straits of Florida also records the

overthrusting of the North America Plate by the Greater Antilles Arc (Reference 221) (Figure 2.5.1-209). After the northwestern portions of Cuba sutured to the Bahama Platform along northwest-trending contractional structures, the plate boundary shifted southward as new northeast-trending strike-slip faults initiated to the southeast (e.g., References 529 and 639). This process continued until Oligocene time, when collision ended in Cuba and the North America-Caribbean Plate boundary migrated to its approximate current position south of Cuba, where the opening of the Cayman Trough led to a more east-west oriented strike-slip system (References 217 and 222). Based on geologic mapping, tectonic reconstructions, and seismic imagery, the site region has been a largely tectonically quiescent portion of the North America Plate since the late Tertiary.

The stratigraphic and tectonic data that support an understanding of the geologic history and ongoing geologic processes affecting the site region and active plate boundary between the North America and the Caribbean Plates are the subject of this subsection. The data were developed from a review of previous reports for the existing Turkey Point units that were updated to include recent concepts, current published geologic literature, and interpretations of data obtained as part of the surface and subsurface field investigations at the Units 6 & 7 site.

### 2.5.1.1 Regional Geology

This subsection provides information on the physiography, geomorphic processes, stratigraphy, tectonic structures and setting, and geologic history within the site region and those features outside of the 200-mile (320-kilometer) radius site region pertinent to geologic hazards at the site. The stratigraphic and tectonic nomenclature for Florida used in this subsection is consistent with that used by the Florida Geological Survey (FGS) (Reference 223). The stratigraphic and tectonic nomenclature for the Gulf of Mexico, Bahama Platform, and the Caribbean used in this subsection generally is consistent with that used by the U.S. Geological Survey (USGS).

The regional geologic map (Figure 2.5.1-201) contains information on the geology within the site region and is discussed in Subsection 2.5.1.1.1. Regional geologic maps (Figures 2.5.1-202 and 2.5.1-210) of the larger Caribbean region contain information requested by the NRC and are discussed in further detail in Subsection 2.5.1.1.2.

While the tectonic and seismic history of the region is the basis for evaluating seismic hazards to the site, the climate history and history of geomorphic processes recorded in the stratigraphic record and current landforms are the basis for evaluating other potential geologic hazards to the site. The Deep Sea Drilling Program (DSDP) (1966 to 1983) (Reference 802), the Oceanic Drilling Program (ODP) (1983 to 2003) (Reference 803), and the present Integrated Ocean Drilling Program (IODP) have supplied large amounts of information used in reconstructing past climates and sedimentary events across the world's oceans. Much of the initial drilling work focused on the Gulf of Mexico, the Bahamas, and Caribbean regions. The result is a growing body of knowledge regarding the climatic, geomorphic, and tectonic history of the region around the Units 6 & 7 site. See Table 2.5.1-201 and Figure 2.5.1-211 for locations of DSDP and ODP sites referenced in this subsection.

A number of events and processes, many of which are interrelated, are common to the Gulf of Mexico, Florida, Bahamas, and Caribbean regions. These events and processes include the following:

- Karstification (Subsection 2.5.1.1.1.1.1.1)
- Carbonate platform development and demise (Subsection 2.5.1.1.1.1.1.2)
- Glacial cycles and associated sea-level fluctuations (Subsection 2.5.1.1.1.1.1.1)

- Megasedimentary events (Subsection 2.5.1.1.1.1.1.3)
- Oceanic and atmospheric reorganization events
- Large igneous province (LIP) magmatic events
- Interplate creation and intraplate modification of oceanic crust

Region-specific evidence for the first four of these events and processes are described in the subsections indicated above. The tectonic evolution of the North America-Caribbean plate boundary is a case study for the remaining three processes (oceanic and atmospheric reorganization and extinction events, LIP magmatic events, and the interplate creation and intraplate modification of oceanic crust). The significance of the remaining three processes (oceanic and atmospheric reorganization and extinction events, LIP magmatic events, and the interplate creation and intraplate modification of oceanic crust) on the tectonic evolution of the Gulf of Mexico, Florida, the Bahamas, and Caribbean regions is briefly discussed below.

### Oceanic and Atmospheric Reorganization and Extinction Events

The recent scientific focus on climate change has resulted in the identification of a number of large ecosystem perturbations (References 226 and 227), many of which were long known to paleontologists as mass extinction events. Some of the ecosystem perturbations are gradual but most have proven to be surprisingly catastrophic. The ecosystem perturbations resulting in major extinctions can generally be tied to a major tectonic change (e.g., emergence of the Isthmus of Panama), a magmatic episode (e.g., the Caribbean LIP), or a bolide impact (e.g., Chicxulub asteroid). While the immediate cause may be localized, the consequences may be global and the change is called a “boundary event.” Causes and effects of boundary events are still poorly understood (Reference 227) but several effects pertinent to the North America-Caribbean Plate boundary region have been identified. Figure 2.5.1-212 identifies seven parameters related to climate change over the past 600 m.y. Table 2.5.1-202 provides a summary of the major Mesozoic and Tertiary boundary events and how they affected the Gulf of Mexico, Florida, the Bahamas, and Caribbean regions.

A major effect of a boundary event is to alter oceanic current patterns. Changes in oceanic currents contribute to the geomorphic processes of marine erosion and deposition. Physical barriers (islands, submarine ridges, and shoals in the Caribbean and Gulf of Mexico) play an important role in controlling the characteristics and the quantity of deep ocean waters and currents that affect the physiographic features of the region. For example, the Bahama Platform, together with Florida and the Greater Antilles Platforms, separate the Atlantic Ocean from the Gulf of Mexico and the Caribbean. The connection between these oceans is limited to the seaways between these three landmasses. As such, the seaways act as a “valve” for the Florida Current that flows between Florida and the Bahamas into the North Atlantic Ocean, where it converges with the smaller Antilles Current to form the Gulf Stream (Reference 228) (Figure 2.5.1-213). The Florida Current is a strong surface current, producing erosion and hiatuses on the Miami Terrace and other terraces of the Florida Straits (Figure 2.5.1-214) and deposition of thick drift deposits in the deeper portions and in the lee of the current (References 229, 230, 231, and 232).

The long-term current strength is primarily controlled by paleotectonic and climatic factors (References 229, 233, and 234). Short-term sea level falls intensify the currents in the seaways because of restriction of the channel area (Reference 235). The tectonic or isostatic uplift of barriers or the filling of channels, such as has occurred for the Isthmus of Panama, the Aves Ridge, and the Suwannee Channel, can significantly affect currents (e.g., close ocean current passages or open passages for new currents, narrow or widen channels to affect the strength of currents, and modify ocean-floor depth to affect density-driven currents) (Reference 236).

## LIP Magmatic Events

LIPs consisting of tholeiitic basalt lava flows, sills and dikes have formed throughout the geologic history of the earth. In general, the LIPs are major bodies of extrusive igneous rock underlain by intrusive rock with crustal thickness ranging from 20 to 40 kilometers (12 to 25 miles). The crustal structure of LIPs is comprised of an extrusive upper crust and a lower crust characterized by high seismic velocities (7.0–7.6 kilometers/second or 4.3-4.7 miles/second) and are different from "normal" oceanic or continental crust. The possible compositions of lower crustal bodies on volcanic margins are gabbroic, strongly mafic, and ultramafic rocks. Some lower crustal bodies have been explained as magmatic underplating by accumulating mantle-derived material below the original crust (Reference 237). During the initial breakup of Pangea in the Late Triassic-early Jurassic, many LIPs formed as the result of rifting, basalt extrusion, and mafic intrusions. The most notable LIPs are the Central Atlantic Magnetic Province (CAMP) and the east coast margin igneous province (ECMIP). The total volume of lava for both the ECMIP, which is the source of the east coast magnetic anomaly (ECMA), and CAMP had exceeded 2.3 million cubic kilometers (0.55 million cubic miles) (Reference 239) to 2.7 million cubic kilometers (0.65 million cubic miles) (Reference 241). However, the two large igneous province (LIPs) that are significant to the tectonic evolution and boundary events that affected the southeastern North America and Caribbean regions are the CAMP and the Caribbean large igneous province (CLIP).

The CAMP is among the largest of the continental igneous provinces on earth, emplaced synchronously with, or just prior to, the Triassic-Jurassic boundary ca. 200 Ma. Magmatism associated with the CAMP occurred from 202 to 190 Ma. Intrusive CAMP magmatism began as early as 202 Ma. Extrusive activity initiated abruptly approximately 200 Ma, reaching peak volume and intensity around 199 Ma on the African margin. There were at least two phases over approximately 1.5 Ma, with magmatism commencing along the Africa-North American margins and slightly later along the South American margin (Reference 981). The extent of CAMP during the Mesozoic as described by McHone (Reference 239) was from "within Pangea from modern central Brazil northeastward approximately 5000 kilometers (3110 miles) across western Africa, Iberia, and northwestern France, and from Africa westward for 2500 kilometers (1550 miles) through eastern and southern North America as far as Texas and the Gulf of Mexico."

The precursor to the formation of the Central Atlantic Ocean (CAO) crust and the opening of the CAO was widespread groups of dike-fed fissure eruptions and flood basalts, which started during the Early to Middle Jurassic along sections of the central Atlantic rift (References 239, 241 and 977). The occurrence of CAMP magmatism and the volcanic rift margin adjacent to the newly forming oceanic crust along the eastern margin of North America is interpreted as subaerial volcanic flows or basalt wedges. This is also indicated by seaward-dipping seismic reflectors (SDRs) (Figure 2.5.1-284, Reference 239).

Within the South Georgia rift basin in the southeastern United States, the continental flood basalts overlap the SDRs. The SDRs are approximately 25-kilometer- (16-mile-) thick basalt and plutonic wedges that are approximately 55 kilometers (34 miles) wide along approximately 2000 kilometers (1240 miles) of the eastern North American margin. These SDRs comprise the ECMA, which has been referred to by Holbrook and Kelemen (Reference 976) as the ECMIP (References 239 and 979).

The ECMA marks the boundary between continental and ocean crust (Figure 2.5.1-266). It forms the seaward edge of the deep, sediment-filled basins and the landward edge of normal oceanic crust (Reference 973). The location of the ECMA segmented magnetic high (200–300 nT, positive magnetic anomaly) as seen in the total field magnetic anomaly, bathymetry, free-air gravity, isostatic anomaly and reduced-to-the pole (R-T-P) anomaly maps parallels the East Coast margin from the Blake Spur fracture zone (BSFZ) to Nova Scotia (Figures 2.5.1-382 and 2.5.1-383) (References 972, 975, 978, and 982). Holbrook and Kelemen (Reference 976) were able to create a velocity model that showed lateral changes in deep crustal structure across the ECMA margin. Lower crust velocities

average 6.8 kilometers/second (4.2 miles/second) at what were believed to be rifted crust areas, whereas the velocity (*Vp*) of units below the outer continental shelf were recorded as 7.5 kilometers/second (4.7 miles/second), decreasing to 7 kilometers/second (4.3 miles/second) with lateral shift to the oceanic crust. Holbrook and Kelemen (Reference 977) were able to determine that the high-velocity (*Vp*) lower crust and SDRs comprise a 100-kilometer- (62-mile-) wide, 25-kilometer- (16-mile-) thick oceanic-continental transition zone that they interpreted to be almost entirely mafic igneous material. This created an abrupt boundary between rifted continental crust and thick igneous crust, comprising only 20 kilometers (12 miles) of the margin. Holbrook and Kelemen (Reference 976) found that the Appalachian intracrustal reflectivity largely disappears across the boundary as velocity (*Vp*) increases from 5.9 kilometers/second (3.7 miles/second) to greater than 7 kilometers/second (4.3 miles/second), implying that the reflectivity is disrupted by massive intrusion and that very little (if any) continental crust can be found east of the SDRs.

The Blake Spur Magnetic Anomaly (BSMA) is a linear anomaly located near the Blake Escarpment, east of the southern portion of the ECMA (ECMA and the BSMA are conjugate anomalies, and they both coincide with the ocean-continent boundary) and north of the Little Bahama Bank (Figure 2.5.1-266) (Reference 466). The BSMA represents points of initiation of sea floor spreading between North America and northwest Africa (Reference 692) and is interpreted as a continental margin modified by a jump in the spreading center (Reference 424) during the early Callovian (middle Jurassic). Ridge jumps are possibly caused by the reheating of the lithosphere as magma penetrates it to feed near-axis volcanism (Reference 980) and can be related to plate interactions as North America separated from Gondwana (Reference 466). It has been suggested that the BSMA is the result of an eastward jump of the spreading center away from the ECMA prior to 170 Ma (Reference 466).

The CLIP formed between 95 and 69 Ma (with two magmatic pulses at 92 to 88 Ma and 76 to 72 Ma) most likely within the Pacific Ocean and most likely associated first with the East Pacific Rise spreading center and, later, with the Galapagos hotspot (References 242, 606, and 243) (Figure 2.5.1-207). Other models favor an intra-Caribbean origin; however, radiolarites of equatorial Pacific faunal affinities in Puerto Rico and the Dominican Republic are older (Reference 820) than the proposed opening of the proto-Caribbean and thus require large-scale lateral transport into their present location, consistent with formation of the CLIP in the eastern Pacific (Reference 244). The volume of the CLIP may be as large as 106 kilometers$^3$ (Reference 245).

The environmental impacts of LIPs are global (e.g., References 246, 247, 248, 249, and 876). Continental flood basalts and oceanic plateaus are formed by widespread and voluminous basaltic eruptions that released enormous volumes of volatiles such as $CO_2$, S, Cl, and F (Reference 246). These extensive eruptions likely caused massive melting of hydrates and explosive methane release where magma intrudes carbon-rich sedimentary strata along rifting continental margins (Reference 250). Courtillot and Renne (Reference 251) propose a strong correlation between LIP ages and the ages of mass extinctions or periods of global oceanic anoxia. In particular, Courtillot and Renne (Reference 251) correlate LIP events with the four largest mass extinctions in the last 260 Ma. Kerr (Reference 247) proposes a correlation between formation of the Caribbean Plateau and the 93 Ma (Cenomanian-Turonian boundary) global oceanic anoxia event (associated with worldwide occurrence of extensive, organic-rich, black-shale horizons) and the 89 Ma (end Turonian) extinction event. LIPs also appear to be associated with the occurrence of Cretaceous marine red beds, probably a secondary result of extrusion of iron-rich lava on the seafloor (Reference 252).

### Interplate Creation and Intraplate Modification of Oceanic Crust

The Caribbean Plate includes basins underlain by normal oceanic crust that transition to, or is faulted against, oceanic plateau (thickened) crust (Figure 2.5.1-215). The characteristics of the two types of crust influence their behavior as they impinge on boundary terranes, such as island arcs (Hispaniola and Puerto Rico) and the South and North America Plates (Reference 253).

The boundary between the North America and Caribbean Plates and the seismogenically active areas of the Caribbean is mostly submerged; thus, data from seismic reflection profiles have been a critical tool in elucidating the stratigraphic and tectonic structures of the region. Most of what is known about the stratigraphy and structure of the submerged areas, including the abyssal basins, spreading centers, trenches, platforms, and ridges, is based on seismic stratigraphy. The following paragraphs describe the basic seismic stratigraphy of: (a) normal oceanic crust created at interplate (divergent) spreading centers (e.g., oceanic crust of the abyssal Atlantic, Gulf of Mexico, Yucatan Basin, and proto-Caribbean Basins), and (b) thickened oceanic plateau crust created at divergent spreading centers and then modified by intraplate magmatic upwelling (e.g., oceanic plateaus of the Venezuelan and Colombia Basins, Beata Ridge, and possibly along the Atlantic margin of North America from the Bahamas to Canada).

Seismic reflection data gathered off the U.S. Atlantic coast in conjunction with early DSDP studies resulted in models of a relatively thin (about 6 kilometers or 3.7 miles thick), two-layer oceanic crust produced at mid-ocean ridge spreading centers (References 254 and 253). Following established naming conventions (A and B for the Atlantic, A' and B' for the Pacific, and A” and B” for the Caribbean), the seismic profiles of oceanic crust in the Gulf of Mexico, Yucatan, and Caribbean Basins exhibit both A” and B” horizons. A” is the higher marker horizon, signaling the change from overlying Tertiary oozes and turbidites to indurated Cretaceous sediments. B” is the lower marker horizon, signaling the change from indurated pelagic sediments and turbidites to igneous rocks (Reference 255).

The B” horizon was identified in the Gulf of Mexico abyssal basins as having a hummocky, rough lower layer (rough B”), signaling the regional basement unconformity between Cretaceous sediments and the normal-thickness oceanic basement of basalt flows intruded by diabase dikes (Reference 255). The rough character of the B” horizon is attributed to valleys separating gently tilted fault blocks, possibly similar to the ridged surface exposed today on either side of the Cayman Ridge spreading center. The rough basalt horizon is sometimes overlain by a thin sequence of mid-Mesozoic pre- and syn-rift sediments (References 255 and 256). In contrast, some of the B” horizon identified in seismic profiles of southern Caribbean Basins exhibits a smooth surface (called smooth B”). In seismic profiles analyzed by Diebold et al. (Reference 255), the rough B” horizon in the Venezuelan Basin-Beata Ridge transect pre-dates the emplacement of smooth B” volcanic material (Reference 255). According to Diebold et al. (Reference 255), the smooth B” represents highly mobile basalt flows that were able to spread in thin fingers over the valleys and faults of rough B” crust. In their model, Diebold et al. (References 255 and 778) postulate that the transition from smooth B” to rough B” crust coincides with shoaling of the Moho from as deep as 20 kilometers (12.4 miles) to as shallow as 10 kilometers (6.2 miles). This shoaling reflects the transition from oceanic plateau crust (Figures 2.5.1-215 and 2.5.1-216) to normal oceanic crust.

Sediments below the A” horizon are generally well laminated. Above the A” horizon, the sediments are more lenticular and hummocky for an interval defined by an overlying reflector, eM. The eM reflector correlates with the boundary between Early Miocene radiolarian ooze and Early to Middle Miocene calcareous ooze. Above the eM reflector, the sediments exhibit continuous stratification that drapes the seafloor topography underlain by thick and thin ocean crust (Reference 253).

#### 2.5.1.1.1 Regional Geology within the Site Region

The area within a 200-mile radius of Units 6 & 7 site is defined as the site region. The site region includes the lower half of the Florida Peninsula, a small area of northern Cuba that is part of the Greater Antilles deformed belt, and extensive marine platform areas, including parts of the Atlantic continental margin with the Blake Plateau, Florida Platform, and Bahama Platform and its western continuation through the Straits of Florida (Figure 2.5.1-201). For purposes of organizing the discussion of the physiography, geomorphic processes, stratigraphy, tectonic features, and geologic history of the site region, this subsection includes information on the Florida Peninsula, the Florida

Platform, and the Florida portion of the Atlantic Continental Shelf and Slope, including the Blake Plateau, the Bahama Platform, and the entire island of Cuba. Areas of the southernmost North America Plate and areas of the North America-Caribbean Plate boundary that are not included in Subsection 2.5.1.1.1 but may be relevant to the geologic hazards at the Units 6 & 7 site are described in Subsection 2.5.1.1.2.

#### 2.5.1.1.1.1 Regional Physiography and Geomorphic Processes

Physiography is synonymous with the term geomorphology or the study of the nature and origin of landforms. Landforms acquire their character by geomorphic processes, such as physical and chemical weathering, erosion, mass transport, biologic impacts, isostasy, and glacial impacts.

##### 2.5.1.1.1.1.1 Florida Peninsula and Continental Margin

The Florida Peninsula and continental margin consists of three physiographic areas (Figure 2.5.1-214):

- The Florida Peninsula
- The Florida Platform
- The Atlantic Continental Shelf and Slope

The Units 6 & 7 site is situated in the southern tip of the Florida Peninsula. The site region encompasses not only part of the Florida Peninsula, but also the southern half of the Florida Platform. A detailed discussion related to karst feature development in the Florida Peninsula physiographic province and in the southern Florida Platform is included in Subsection 2.5.1.1.1.1.1. Individual primary and secondary Florida Peninsula physiographic subprovinces that fall within that 200-mile radius site region are also described in detail in Subsection 2.5.1.1.1.1.1. The entire Florida Platform, including the Florida Escarpment (Figure 2.5.1-214), is discussed in Subsection 2.5.1.1.1.1.2. The unusually broad and shallow Atlantic Continental Shelf and Slope of Florida, including the Blake Plateau, while outside of the 200-mile site radius, are discussed in Subsection 2.5.1.1.1.1.3 because of their importance in protecting peninsular Florida from the effects of strong waves from the Atlantic Ocean.

###### 2.5.1.1.1.1.1.1 Florida Peninsula Physiographic Subprovinces

**General Geography and Geology of the Florida Peninsula**

The Units 6 & 7 site (Figure 2.5.1-214) is located within the Atlantic Coastal Plain physiographic province, comprising low-lying, gently rolling topography. The Atlantic Coastal Plain physiographic province is the portion of the Atlantic seaboard of the U.S. that extends eastward from the Fall Line to the coastline. The Fall Line is the physiographic boundary between the Coastal Plain province and the Piedmont province. In the eastern U.S. it parallels the north-northeast to south-southwest strike of the Appalachian orogen from Maine to Georgia, and then turns west into southern Georgia and Alabama, where it delineates the northeastern boundary of the Gulf of Mexico Coastal Plain.

The Florida Peninsula is a stable carbonate platform upon which surficial, dominantly Neogene and Quaternary terrigenous sediments have accumulated in an asymmetric fashion; relatively thick on the East Coast and generally thinning toward the west coast (Figure 2.5.1-224). The carbonate platform was prograding from the Mesozoic until the late Paleogene, when the early seaways connecting the Atlantic to the Gulf of Mexico (known through time as the Suwannee Channel, the Georgia Channel, and Gulf Trough) (Figure 2.5.1-218) were finally terminated (References 257, 258, and 234). The seaway system was a dynamic barrier that prevented siliciclastic sediments, eroded from the southern Appalachian Mountains, from inundating and smothering the carbonate production

to the south. Eventually the seaways were filled and the Florida Peninsula was covered by siliciclastics transported from the north via fluvial and longshore processes. The combination of antecedent topography on the carbonate bank and multiple sea level fluctuations heavily influenced distribution of the dominantly quartz sand cover (Reference 259).

Little of the ancient geologic history is revealed in the geomorphology of Florida due to its low, flat lying topography. Paleoshorelines, scarps, and terraces; present-day river drainage patterns; and karst features indicate that a combination of sea level and groundwater fluctuations and coastal geomorphic processes have strongly shaped Florida's current terrain. The fluctuations in sea level occurred through many cycles that extended through thousands to millions of years (References 562, 749, 877, and 298). Waves, winds, and currents have eroded sands from some locations and deposited the shifting sand along the beaches. The resulting principal geomorphic features in Florida are represented by the barrier island system of the Gulf Coastal Lowlands, the Reticulated Coastal Swamps of southwest Florida, and the swales and swamps of the Everglades (Figure 2.5.1-217).

At marine high stands, the central and southern areas of the Florida Peninsula were mostly submerged (Figure 2.5.1-219), and wave action eroded coast-parallel ridges and cliffs into emergent strata. During sea level lowstands, all of the Florida Peninsula and much of the Florida Platform were subaerially exposed (Figure 2.5.1-219), and the coast-parallel ridges and cliffs controlled surface water flow to further enhance coast-parallel valleys (Reference 265). Thus, central Florida is characterized by discontinuous highlands in the form of sub-parallel ridges separated by broad valleys. Southern Florida is characterized by a broad, flat, gently sloping, and poorly drained plain that is bounded on the east by the Atlantic Ridge. These low-lying lands of southern Florida do not reach elevations as high as the Atlantic Ridge because they were not exposed to high-energy processes of the Atlantic shore, which carried sand southward from northern sources to build beaches. The high-energy environment of the Atlantic shoreline decreases as it meets the Florida Current in the Florida Straits. There, nutrients carried by the warm Florida Current feed the reef-building corals of the Florida Keys and the current redistributes sand offshore as sand ridges (Reference 265).

Eight marine terraces (Hazlehurst [formerly Brandywine], Coharie, Sunderland, Wicomico, Penholoway, Talbot, Pamlico, and Silver Bluff) are mapped across the Florida landscape (Figure 2.5.1-220) (Table 2.5.1-203). Ward et al. (Reference 260) correlate nine marine terraces found in South Carolina (laterally equivalent to those in Florida) with stranded shorelines in Australia and Alaska. The Ward et al. (Reference 260) comparison suggests four high sea levels occurred in the last 0.25 Ma and two others about 0.4 Ma. Other stages of high sea level are dated at 0.76, 1.43, 1.66, and 1.98 m.y. The authors calculated altimetric ages from correlation of high sea level data with deep-sea core stages to provide estimated ages for the Florida/South Carolina high stands (Table 2.5.1-202). The ages of the South Carolina terraces generally correlate with global sea level high stands proposed by Ward et al. (Reference 260). There is some debate regarding the appropriate methodology for age estimation of the Florida marine terraces based on various radiometric and other age-dating techniques (Reference 261). These marine terraces are the former bottoms of shallow seas and are bounded along their inner margin by shoreline features such as relict beach ridges, swales or inner lagoons, seaward facing wave-cut scarps or sea cliffs, and offshore and bay bars (Reference 261).

The marine terraces in Table 2.5.1-203 were once thought to be the direct result of sea level fluctuations through the last glacial cycles but are now understood to be a result of complex interactions between sea level oscillation, subaerial exposure, a precipitation karstification function, and isostatic uplift (References 262 and 927). Since reefs form in a shallow marine environment, the organisms that comprise the Key Largo and Miami limestones preserve the record of Pleistocene sea level changes. These limestones in some places have been subaerially exposed. Investigators (References 928 through 933) studied the aforementioned limestones to understand the

Atlantic-Caribbean sea level changes. The record of Pleistocene sea level changes is preserved in the marine sequences Q1 through Q5, from oldest to youngest, which correlate to marine isotope stages MIS 11, 9, 7, and 5e (Table 2.5.1-209) (Reference 928). The marine sequences are defined as a stratigraphic sequence of marine strata that represents a population of benthic organisms. Marine isotope stages (MIS) are alternating warm and cool periods in the earth's paleoclimate history, inferred from oxygen isotope data reflecting changes in temperature.

Adams et al. (Reference 927) generated a model that calculates lithospheric uplift as a result of a precipitation-driven karstification function (decrease of bulk crustal density) and variations in subaerial exposure of a carbonate platform (i.e., Florida) due to oscillating sea level. The authors applied this model to north-central Florida to estimate the ages of beach ridges and depositional coastal terraces. The ages were based on the most recent estimates of sea level history since the Pliocene. The modeled ages of sea level highstands were then compared to the elevations of uplifted beach ridges and coastal terraces to evaluate plausible ages for deposition of the observed coastal geomorphic features (Reference 927). The geomorphic features were the Trail Ridge, the Penholoway Terrace, and the Talbot Terrace (Figure 2.5.1-370). The model produced the following ages for the three geomorphic features near the north Florida-southeastern Georgia Atlantic coast: (1) Trail Ridge approximately 1.44 m.y., (2) Penholoway Terrace approximately 408 k.y., and (3) Talbot Terrace approximately 120 k.y. (Table 2.5.1-203, Reference 927).

Hickey et al. (Reference 928) analyzed the $^{234}U/^{238}U$ ages of cores recovered at Grossman Ridge Rock Reef and Joe Ree Rock Reef in the Florida Everglades and revealed additional subaerial-exposure surfaces that are used to delineate subdivisions within the five marine sequences of the Pleistocene carbonates of south Florida (Figure 2.5.1-371). These five marine sequences with Hickey et al. (Reference 928) subdivisions in parentheses are as follows: Q1 (Q1a–Q1b), Q2 (Q2a–Q2d), and Q4 (Q4a–Q4b) and Q5 (Q5e) (Figure 2.5.1-372). These subdivisions delineated by Hickey et al. (Reference 928) within units Q1 through Q5 preserve evidence of at least ten separate sea level highstands, rather than five as indicated by previous studies (Reference 928). Q5e is the youngest Pleistocene subaerial exposure surface of the Florida Keys (Figure 2.5.1-372). The fossil content and the $^{234}U/^{238}U$ radiometric ages indicate that this morphostratigraphic unit was deposited during the peak sea level of the last interglacial marine isotope substage 5e (MIS 5e). Uranium-series ages on corals from this unit from Windley Key, Upper Matecumbe Key and Key Largo range from 130 to 121 ka after corrections for calculated high initial $^{234}U/^{238}U$ content (Reference 928). A Q4a sample from Point Pleasant near the island of Key Largo has a best estimate age range of 340–300 ka, which falls into the early part of marine-isotope stage 9 (MIS 9) (Reference 928). The age of a Q4b coral sample recovered from a spoil pile in a quarry within unit Q4 on Long Key, southwest of Key Largo is approximately 235 ka (corrected for calculated high initial $^{234}U/^{238}U$). This is consistent with the early part of MIS 7. Hickey et al. (Reference 928) concludes that the Q1 through Q3 units predate MIS 9 and that their preferred interpretation is that Q3 was deposited during MIS 11 and that Q2 and Q1 represent pre-MIS 11 interglacial intervals (Figure 2.5.1-373) (Reference 928). Lastly, Muhs et al. (Reference 929) obtained ages of corals from Windley Key, the island of Key Largo, and from Long Key to Spanish Harbor Keys (middle Florida Keys) using Uranium-series dating. $^{234}U/^{238}U$ age dates are as follows: approximately 114 to 122 ka (Windley Key), approximately 120 to 123 ka (island of Key Largo), and approximately 114 ka (Long Key to Spanish Harbor Keys) (Reference 929). Thus the ages obtained by Muhs et al. (References 929 and 933) correlate to MIS 5e and are consistent with the dates obtained by Hickey et al. (Figures 2.5.1-373 and 2.5.1-377 and Table 2.5.1-209) (Reference 928).

Although no post-Stage 5e dates have been reported from corals recovered from pits or cores from the exposed Florida Keys, several younger dates have been obtained from submerged corals recovered from the shelf to the east of the Florida Keys (References 930 and 931). These have been assigned to marine-isotope substages 5c, 5b, and 5a. These post-Q5e interglacial highstands were not high enough to flood the south Florida inner platform (Reference 928). Multer et al. (Reference 930) obtained dates for the Key Largo Limestone using thermal ionization

mass-spectrometric (TIMS) uranium-thorium (U-Th) dating. The dates from these rocks, 112.4 to 77.8 ka, correspond to the marine-isotope substages 5c and 5a (MIS 5c and MIS 5a). These rocks were found under the shelf edge at Conch Reef, Looe Key, under Carysfort Light area and at the shelf edge near Molasses Reef (Figures 2.5.1-374, 2.5.1-375, and 2.5.1-376) (Reference 930). Toscano and Lundberg (Reference 932) also used TIMS U-Th dating and obtained dates of 7.7 +/-0.7 ka and 8.6 +/-0.1 ka (basal Holocene) above the unconformity on the shelf edge (core SKSE) at Sand Key outlier reef (lower Keys) (Figures 2.5.1-374 and 2.5.1-375) (Reference 932). Below the unconformity, Toscano and Lundberg (Reference 931) obtained TIMS U-Th dates on corals from Sand Key outlier reef and Carysfort Light area of 86.2 +/-1.01 and 80.9 +/-1.7 ka (Figures 2.5.1-376 and 2.5.1-377).

The variations in orientation of shoreline features indicate variations in eustatic adjustment across the Florida Platform and Peninsula (Reference 262). Karstification effectively accomplishes the equivalent of isostatic compensation by decreasing the crustal mass within a vertical column of lithosphere. The rate of karstification (void space creation or equivalent surface lowering rate) within the north Florida Platform is about 3.5 times that of previous estimates (1 meter/11.2 thousand years [k.y.] vs. 1 meter/38 k.y.), and uplift rate is about two times higher than previously thought (0.047 millimeters/year vs. 0.024 millimeters/year) (Reference 262). This process has implications for landscape evolution in other carbonate settings and may play an underappreciated role within the global carbon cycle (Reference 262).

### Karst Processes and Features on the Florida Peninsula

Karstification is the process created by chemical dissolution when weakly acidic groundwater circulates through soluble rock (Figure 2.5.1-221). Carbon dioxide from the atmosphere is fixed or converted in the soil horizon to an aqueous state, where it combines with rainwater to form carbonic acid, which readily dissolves carbonate rock. Root and microbial respiration in the soil further elevates carbon dioxide partial pressure, increasing acidity (lowering pH). In tropical and subtropical regions such as Florida, abundant vegetation, high rainfall, and high atmospheric $CO_2$ values favor the rapid dissolution of the preexisting limestone.

### Carbonate Dissolution at Freshwater/Saltwater Interfaces

The freshwater/saltwater interface is defined as the location where seawater intrudes into a coastal aquifer and mixes with the discharging freshwater in a zone of mixed groundwater composition. The chemical reactivity of the mixing zone stems from the marked undersaturation with respect to carbonate minerals that develops from mixing a carbonate saturated freshwater with near surface seawater in a system closed with respect to carbon dioxide (Reference 945). Dissolution occurs when the two fluids of different salinities combine, even though both fluids are initially saturated with calcium carbonate (Reference 951). Because seawater saturated with calcium carbonate contains far less calcium carbonate than fresh groundwater saturated with calcium carbonate, the combined fluids become undersaturated with respect to calcium carbonate. This condition promotes dissolution of carbonate rocks.

Dissolution of limestone generally occurs where fresh, weakly acidic groundwater circulates through soluble carbonate rock or within zones of mixing fresh and seawater (References 263 and 965). The freshwater/saltwater interface within the Biscayne Aquifer is located approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Tables 2.4.12-210 and 2.4.12-211) and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, the site is not a location of fresh groundwater discharge or mixing of fresh and saltwater, and the mechanism necessary to form large solution cavities does not appear to be active on or near the site.

Rising sea level will increase the ocean hydrostatic head and tend to force intrusion of the freshwater/saltwater interface further inland and away from the site. Therefore, the mixing zone

mechanism necessary to increase the potential for carbonate dissolution and formation of large solution cavities on or near the site will not exist. Collapse of solution cavities is generally associated with lowering of groundwater levels and withdrawal of buoyant support. A rising sea level will counter this effect.

Conversely, any potential lowering of sea level would tend to move the freshwater/saltwater interface seaward and toward the site. However, the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), and sea level has been rising throughout the current interglacial stage of the Holocene. A significant lowering of sea level is not likely to occur until a future advance of continental glaciation, which is not likely to occur within the operating lifetime of Turkey Point Units 6 & 7. The magnitude of sea level lowering and the corresponding time necessary to move the interface to a location within the area of the site is not likely to occur within the operating lifetime of Turkey Point Units 6 & 7 (Subsection 2.4.5). Therefore, increased carbonate dissolution or formation of large solution cavities on or near the site due to a lowering of sea level is not likely to occur during construction or operation of the plant.

Several researchers (References 946, 947, 948, 949, and 950) indicate that carbonate dissolution associated with the mixing of freshwater and saltwater occurs predominantly at groundwater discharge sites or seafloor discharge zones. Mixing can also occur in surface water. The dissolution mechanisms are point source discharge and submarine groundwater discharge (SGD).

### Point Source Discharge

Point source discharge is a concentrated flow of spatially constricted fresh surface water into a saltwater body. The discharge can affect the local water chemistry equilibrium with the potential to alter the rate of dissolution or deposition of carbonates within the mixing zone in its vicinity. An example of a point source discharge is surface water released to Biscayne Bay through drainage canal discharge.

The freshwater/saltwater interface at the base of the Biscayne Aquifer is located approximately 9.6 kilometers (6 miles) inland of Turkey Point Units 6 & 7, as shown on Figure 2.4.12-207. The migration of saltwater inland along the base of the aquifer occurs along the entire coastal zone and is the result of the aquifer's high permeability, the lowering of inland groundwater levels from groundwater pumping and surface drainage, and rising sea level (Subsection 2.4.5). As shown on Figure 2.4.12-207, the position of the freshwater/saltwater interface was relatively consistent between 1984 and 1995 and, in fact, provisional data from the USGS (Reference 960) showing the 2008 freshwater/saltwater interface in southeast Florida indicates a similar pattern.

Under natural conditions and before anthropogenic activity (e.g., construction of canals and enlargement of the Miami River) (References 267, 722, 955, 961, 962, and 963), the freshwater/saltwater interface in southeastern Florida was close to the coastline and freshwater discharged from springs on the floor of Biscayne Bay. In the late nineteenth century construction of flood control levees, drainage canals, and urbanization changed the position of the freshwater/saltwater interface. Canals were first dug through the Everglades to drain water from the area south of Lake Okeechobee to enable agriculture to develop (Reference 267). These canals roughly follow the transverse glades (i.e., narrow valleys or channels in which the soils (marl and sand) and vegetation are similar to those in the Everglades). By the late 1920s, major canals were constructed and rivers in the transverse glades were modified to connect Lake Okeechobee with the Gulf of Mexico and Atlantic Ocean (Figure 2.4.12-207) (References 267 and 964). In the 1930s, the government initiated flood control measures including levee construction and drainage channel modification. By the 1970s, gated control structures were installed at the coastal end of the primary drainage canals to discharge excess water during the wet season and impede the landward movement of saltwater during the dry season. The final phase of canal development of the

Everglades-South Dade conveyance system in the 1980s was constructed to meet agricultural water supply needs, control flooding, and mitigate saltwater intrusion (Reference 267).

The increased fresh surface water discharge from the Everglades to Biscayne Bay and the Atlantic Ocean through the drainage canals and increased pumping from the freshwater aquifer has probably had an impact on coastal groundwater hydrology by contributing to inland migration of the freshwater/saltwater interface as shown in Figure 2.4.12-207. Point source discharge also may have increased the potential for dissolution of carbonate rocks in the immediate vicinity of the drainage canal outfalls. However, stratification of freshwater near the surface of the canal outfalls may limit carbonate dissolution to the near surface.

Outfalls of drainage canals closest to the site are the Model Land Canal (C107) outfall near the southeast corner of the Turkey Point cooling water canals, approximately 8 kilometers (5.0 miles) south of the site, and the Florida City Canal outfall, approximately 1.9 kilometers (1.2 miles) north of the site (Figure 2.4.1-203). Because of their distance from the site, the possible effect of freshwater stratification near the outfalls, and the effects of variable discharge from the outfalls related to operation of their control structures, variable rainfall, tidal fluctuations, and hurricanes, neither outfall is likely to induce formation of cavernous limestone with the potential for collapse at the site.

### Submarine Groundwater Discharge

SGD is defined as the “phenomenon that forces groundwater to flow from beneath the seafloor into the overlying ocean regardless of its composition, whether freshwater, recirculated seawater, or a combination of both” (References 946 and 952). SGD can be subdivided into “shoreline flow” (i.e., fresh groundwater flow through an aquifer to the nearshore ocean that is driven by an inland hydraulic head) and “deep pore water upwelling” (DPU) (i.e., fresh groundwater flow beyond the shoreline on the continental shelf through deeper confined permeable shelf sediments and rocks, driven by buoyancy and pressure gradients (Reference 946). Reference 953 states “SGD per unit length of coastline could be very significant as a discharge process, due to the length of coastline where SGD occurs; whether or not rivers are present.” The extent of SGD or saltwater intrusion at a given location is an issue of balance between hydraulic and density gradients in groundwater and seawater along a transect perpendicular to the shoreline (Reference 953). The two possible modes of submarine groundwater discharge, shoreline flow and deep pore water upwelling, are discussed below.

### Shoreline Flow

As stated above, shoreline flow to the sea occurs when fresh groundwater flow through an aquifer is driven by an inland hydraulic head. As the shoreline flow nears the sea, it encounters the saltwater that has infiltrated from the ocean. The density of freshwater is lower than that of saltwater and therefore it tends to flow above the saltwater. The freshwater flowing toward the sea encounters an irregular interface where mixing of the fluids is driven by diffusion and dispersion enhanced by ocean forces (i.e., tidal pumping, wave setup, storms, buoyancy, and thermal gradients). This freshwater/saltwater circulation pattern and mixing is similar to that in surface estuaries, leading to the term subterranean estuaries. Tidal forces operating in a mixed medium (i.e., bedrock) may enhance dispersion along the freshwater/saltwater interface and the permeability and preferential flow paths may be changed by chemical reactions within the aquifer. Precipitation of solids can restrict or block some paths, while dissolution will enlarge existing paths or open new ones (Reference 946). Examples of shoreline flow are:

- Freshwater springs along Biscayne Bay (approximately 25 kilometers [16 miles] northeast of the site)
- Cave development along the Atlantic Coastal Ridge (approximately 17 kilometers [11 miles] north-northeast of the site)

- Submarine paleokarst sinkhole in the Key Largo National Marine Sanctuary (approximately 13 kilometers [8 miles] south of the site)
- Blue holes of the Bahamas in eastern South Andros Island (approximately 190 kilometers [120 miles] southeast of the site)
- Karst development on emergent carbonate islands in the Bahamas (approximately 320 kilometers [200 miles] southeast of the site)
- Karst development on the Yucatan Peninsula, Quintana Roo, Mexico (approximately 560 kilometers [350 miles] southwest of the site)

### Freshwater Springs along Biscayne Bay

Fresh groundwater had discharged along the Atlantic Coastal Ridge shoreline and offshore as submarine springs before the drainage canals were built and before substantial lowering of surface water and groundwater levels in southeast Florida. The groundwater flow conduits still exist and are dissolution features within the Biscayne Aquifer. Springs reportedly discharged near shore as freshwater boils in the shallow waters of Biscayne Bay (References 721, 954, 955, and 1000). In the late 1800s and early 1900s, springs within the Biscayne Aquifer provided a source of freshwater for sailing ships in Biscayne Bay. Parks (Reference 956) describes a freshwater spring off Coconut Grove (south of Miami) (Figure 2.5.1-390) that was first documented in 1838 by Dr. Jacob Rhett Motte. Later a pump and platform was constructed to enable dories to tie up while filling wooden kegs with freshwater. This spring was marked as “freshwater” on Coast and Geodetic Survey Navigation Chart No. 166 (1896) (Reference 954).

Currently, karst spring catalogues maintained by FGS (Reference 1001) do not include entries for the Turkey Point Units 6 & 7 site vicinity or Broward or Miami-Dade counties. Evidence for spring flows (and inferred karst conduits) in the site vicinity (for example, at Coconut Grove and Devils Punch Bowl) is nonetheless provided by historical accounts (References 1002 and 1003). Anecdotal information suggests that submarine groundwater discharges into Biscayne Bay were particularly significant in the area between the Coral Gables Canal (near Coconut Grove) and the Mowry Canal, located approximately 5.1 kilometers (3.2 miles) north from the site, at least prior to canal construction (Reference 949).

Aerial imagery for the shoreline near Turkey Point Units 6 & 7 from 1938 clearly captures an offshore spring and groundwater seepage only 1500 meters (4921 feet) from the approximate site center-point (Figure 2.5.1-390). Gonzalez (Reference 1000) relocated the seepage/discharge point in 2004, but did not observe flow. Generally though, the approximately relocated spring site was characterized by sediment-filled, seagrass-covered karst holes.

At least 21 additional offshore springs (identified by green circles on Figure 2.5.1-391) were located in 2006 by Gonzalez (Reference 1000) in an area approximately mid-way between the aforementioned Mowry and Coral Gables canals. Generally, Gonzalez (Reference 1000) classified these seepage points as small, ephemeral openings in soft sediment, typically less than 15 centimeters (6 inches) across, or as more persistent, large diameter (1 to 4 meters [3 to 13 feet]) features. Discharge from the larger diameter features was described as strong with resulting exposure of the limestone surface and associated karst conduits, although dry season flow was apparently discernible only during low tide. Flow in the smaller, ephemeral springs was visible only in the wet season, or following precipitation events. Flow in all springs was diminished when nearby canal flood gates were opened.

Gonzalez (Reference 1000) reported that the spring waters were slightly acidic, and ranged in salinity from approximately 8 to 31 grams per liter (g/L) (equivalent to 8 parts per thousand [ppt] to 31 ppt). Foraminiferal assemblages associated with the springs were thus reported to include both brackish

and fresh water species. Significantly, Gonzalez (Reference 1000) indicated that foraminifera tests recovered from the springs exhibited extensive pitting, and thus suggests that some carbonate dissolution occurs at the discharge sites.

Because offshore spring flow (shallow submarine groundwater discharge) in the immediate site vicinity is relatively low, it is likely that associated dissolution is limited.

Langevin (Reference 948) suggested that the drainage canals are the present focal points for groundwater discharge into Biscayne Bay, intercepting fresh groundwater that would have discharged directly to the bay. Field observations by Langevin (Reference 948) suggest that Biscayne Bay has changed from a system controlled by widespread and continuous submarine discharge and overland sheet flow to one controlled by episodic releases of surface water at the mouths of drainage canals. The canals and pumping from the freshwater aquifer have lowered the water table and, thus, submarine groundwater discharge has decreased. The Turkey Point Units 6 & 7 groundwater model is consistent with Langevin's model (Reference 948).

### Cave Development along the Atlantic Coastal Ridge

Caves are not particularly common in the Turkey Point Units 6 & 7 site vicinity or in the wider southeastern Florida region (Figure 2.5.1-354). Cressler (Reference 955) described only 19 air-filled caves and one water-filled cave in southeastern Florida, although an additional seven caves have since been mapped by Florea (Reference 1004). Typically, these caves are located along the Atlantic Coastal Ridge or transverse glades (low relief, relict tidal channels) that cut across the Atlantic Coastal Ridge (Figures 2.5.1-390 and 2.5.1-391).

According to Cressler's (Reference 955) and Florea's (Reference 1004) field observations and descriptions, the caves within the Pleistocene limestones fall into four categories: (1) some are oriented along fractures, (2) some caves are concentrated along the margins of transverse glades, (3) some caves are composed of stratiform lateral passages, and (4) some caves have entrances along the margins of cave-roof collapse. Most of the caves discovered by Cressler (Reference 955) fall into the second category. The caves are concentrated along the margins of transverse glades. Entrances to the caves are either along the glade wall or occur as pits subjacent to the glade wall. Cressler (Reference 955) hypothesized that slightly acidic water from the Everglades could be a potent agent for dissolving limestone and forming the caves in the transverse glades in the Miami Limestone.

The most extensive karst development in Miami-Dade County lies within the boundaries of the Deering Estate County Park and Preserve (Reference 955) on the eastern flank of the Atlantic Coastal Ridge. The Deering Estate County Park and Preserve is located approximately 17.6 kilometers (11 miles) north-northeast of the site. Of the caves identified by Cressler (Reference 955) and Florea (Reference 1004), 11 are located in the Deering Estate Preserve.

Observations in the Deering Estate indicate that variations in Pleistocene stratigraphy (i.e., Miami Limestone) may have played an important role in the origin of many small caves, including the 95.4 meters (313 feet)-long Fat Sleeper Cave (Reference 1004). At Deering Estate, cave passages are commonly low, wide and sandwiched between crossbeds of oolitic limestone. These stratiform passages seem confined to a zone of rock with many centimeter-scale vugs related to complex burrow systems. It is hypothesized that the burrow-related porosity provided early preferential pathways for groundwater flow and concentrated dissolution. In some caves, solution pipes penetrate the upper cross-bedded limestone and connect to the land surface (References 954 and 955).

One of the most well known caves in Miami-Dade County, Palma Vista Cave, is located on Long Pine Key in the Everglades National Park (Figure 2.5.1-355). The entrance of the Palma Vista Cave probably formed by the collapse of a thin roof that spanned a stratiform cave (Reference 954). The

speleothems in the cave that are underwater are important because their presence implies that they developed in Palma Vista Cave during a previous, extended dry period (i.e., sea level low stand). Such a condition would have existed when sea levels were much lower, such as the period between approximately 80,000 and 6,000 years ago (Reference 957).

The Atlantic Coastal Ridge caves formed by solution enlargement of sedimentary structures in the Miami Limestone as groundwater entered the freshwater/saltwater mixing zone and discharged as shoreline flow on the margin of the coastal ridge. The freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the coast (Figure 2.4.12-207), shoreline flow at the Turkey Point Units 6 & 7 site is brackish to saline (Tables 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, the mixing-zone process that formed the caves along the flanks of the Atlantic Coastal Ridge is not likely to be active currently in formation of cavernous limestone with the potential for collapse in the area of the site.

### Submarine Paleokarst Sinkhole in the Key Largo National Marine Sanctuary

A large submarine, sediment-filled paleosinkhole in the Key Largo National Marine Sanctuary off Key Largo, Florida (Figure 2.5.1-360) is described as having a 600-meter (1970-foot) diameter with a depth likely to exceed 100 meters (328 feet) (Reference 959). The Key Largo submarine paleosinkhole lies beneath 5–7 meters (16-23 feet) of water, and is bordered by Holocene reefs to the east and marine grass and carbonate sand to the west. Shinn et al. (Reference 959) jet probed to 54.5 meters (179 feet) and did not reach the bottom of the sinkhole. Patches of marine grass grow on the carbonate sands in the circular feature, but corals are absent (Reference 959). The sediments as observed from the sediment cores consist of monotonous gray aragonite mud visually lacking sedimentary laminations and fossils. The composition of the sediment as analyzed by X-ray diffraction is approximately 95 percent aragonite and 5 percent calcite. The oldest $^{14}C$ age (from the bottom of the jet probe sampler) is 5650 +/-90 years before present. The youngest $^{14}C$ age (just below the overlying carbonate sand cap) is 3260 +/-60 years before present. The high percentage of aragonite and near absence of low-magnesium calcite indicate the sediment is of marine origin and the $^{14}C$ dates indicate rapid deposition (Reference 959).

Shinn et al. (Reference 959) postulate that the Key Largo sinkhole is a cenote that formed during the Pleistocene. Fluctuations in sea level related to advance and retreat of continental glaciers raised and lowered the fresh groundwater/seawater shoreline mixing zone in the area of the sinkhole and facilitated dissolution of carbonate rocks to a depth near the sea level low stand. As the Wisconsinan ice sheet began to retreat and sea level began to rise 15,000 years ago, the shelf off Key Largo was at least 100 meters (328 feet) above sea level. A shallow freshwater lake would have formed at the bottom of the sinkhole. The lake would have gradually deepened as the groundwater level adjusted to the rising sea level. By 6000 years ago, just before marine flooding of the shelf, the sinkhole would have been surrounded by wetlands. Infilling of the sinkhole most likely began with precipitated freshwater calcite muds (i.e., marl). As sea level continued to rise, fresh and brackish water were replaced by saline waters. Marine sediment began to settle into the sinkhole, at which time the sinkhole would have functioned like a giant sediment trap. The $^{14}C$ dates indicate that pulses of rapid sedimentation at 4.1 ka and 4.8 ka (thousand years before present) punctuated marine sedimentation. These pulses were likely the result of tropical hurricanes, which reworked and deposited the lime mud on the Florida reef tract. The lime mud sedimentation ceased and was replaced by sedimentation with skeletal carbonate sands approximately 3 ka. The eastern rim of the sinkhole is dominated by coral reefs which are assumed to be the major source of the carbonate sands that cap the muddy sediment (Reference 959).

In summary, it is postulated that the Key Largo submarine paleosinkhole began to form during the Pleistocene. Infilling of the sinkhole began approximately 15,000 years ago when sea level began to rise. The environment at the bottom of the sinkhole at that time was essentially that of a freshwater lake that became brackish and eventually evolved to the current marine environment, at which point

conditions conducive for continued limestone dissolution and sinkhole formation no longer existed. At approximately 6 ka the sinkhole was inundated by seawater and became a sediment trap. Rapid pulses of sedimentation occurred approximately 4.1 ka and 4.8 ka. At approximately 3 ka, coral reefs began to accumulate on the seaward side of the sinkhole.

Because the position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Tables 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), there is no fresh groundwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist and the process of shoreline flow that formed the Key Largo submarine paleokarst sinkhole is not a mechanism that is likely to produce cavernous limestone with the potential for collapse at the site.

### Blue Holes of the Bahamas, Eastern South Andros Island

The blue holes of the Bahamas beneath South Andros Island lead to an extensive system of underwater caves along nearshore fracture systems (Figure 2.5.1-365). Formation of the blue holes, which reach depths exceeding 100 meters (328 feet), began during a previous eustatic sea level low stand associated with advance of continental glaciation during the Pleistocene. Groundwater circulation to the blue holes is facilitated by the fracture permeability that exists within the fracture systems in the carbonate rock. Investigations into groundwater-seawater circulation in some of the holes offshore of South Andros Island indicate a brackish mixture in the caves that readily dissolves aragonite but not calcite, producing secondary porosity. The depletion of calcium in the saline groundwater indicates precipitation of calcite cement. Bacterial processes possibly due to submarine groundwater discharge also play a significant role in driving carbonate dissolution in the Bahamas (References 946 and 950).

A similar nearshore fracture system has not been identified in the limestones within the area of the Turkey Point Units 6 & 7 site. As noted previously, the position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Table 2.4.12-211), the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5), and there is no fresh groundwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist. For these reasons, conditions favorable for formation of dissolution features similar to the blue holes of the Bahamas do not appear to exist in the site area.

### Karst Development on Emergent Carbonate Islands in the Bahamas

In the Bahamas, flank margin caves (Figures 2.5.1-361 and 2.5.1-362) form on emergent carbonate islands due to the mixing of fresh and saltwater in the presence of organic matter. The presence of organic matter allows oxidation to produce carbon dioxide, which in turn produces carbonic acid that drives carbonate dissolution. This carbonate dissolution results in anoxic conditions in the mixing zone of the fresh groundwater lens. Complex oxidation/reduction reactions involving sulfur produce acids that lead to further dissolution (Reference 263). The morphology of the flank margin caves includes large, globular chambers, bedrock spans, thin bedrock partitions between chambers, tubular passages that end abruptly, and curvilinear phreatic dissolution surfaces. The flank margin caves are not conduits, but rather mixing chambers (Figure 2.5.1-362). They receive freshwater from the fresh groundwater lens in the island interior as diffuse flow, and discharge that water, after mixing, as diffuse flow to the sea. The caves develop without an external opening to the sea or the land. Current entry is possible due to surface erosion breaching into the cave (Reference 263). Examples of flank margin caves are Lighthouse Cave, San Salvador Island, Bahamas and Salt Pond Cave, Long Island, Bahamas (Reference 263).

In addition to flank margin caves, there are banana holes in the Bahamas (Figure 2.5.1-362). Banana holes form inland from the flank margin caves at the top of the fresh groundwater lens where the vadose and phreatic freshwaters mix. They are smaller phreatic dissolution voids that form due to collapse of a relatively thin bedrock roof resulting in a broad, vertical-walled depression up to 10 meters (33 feet) across (Reference 263). Both the flank margin caves and banana holes are found in the Bahamas at elevations of 1 to 6 meters (3.3 to 20 feet) above sea level. These caves formed during a glacioeustatic sea level high stand that reached elevations above modern sea level. According to Mylroie and Carew (Reference 263), these caves formed approximately 125,000 years ago. The duration of this high stand above modern sea level lasted approximately 15,000 years, during which time the Bahamas consisted of islands even smaller than today because all land below 6 meters (20 feet) in elevation was below sea level. Therefore, these phreatic caves formed in small freshwater lenses in as little as 15,000 years (Reference 263).

The process of shoreline flow that formed the flank margin caves may be active in the Bahamas today, but at an elevation closer to modern sea level. However, similar processes are not likely to be active at the Turkey Point Units 6 & 7 site because of the absence of fresh groundwater shoreline flow near the site. The position of the freshwater/saltwater interface is approximately 9.6 kilometers (6 miles) inland from the site (Figure 2.4.12-207), groundwater at the site is saline (Tables 2.4.12-210 and 2.4.12-211), and the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.2 meter (0.78 foot) per century (Subsection 2.4.5). Therefore, a freshwater/saltwater mixing zone that would promote carbonate dissolution at the site does not now exist.

### Karst Development on the Yucatan Peninsula, Quintana Roo, Mexico

The Yucatan Peninsula is outside of the 200-mile radius “site region” but karst development there provides evidence of shoreline flow and, therefore, is discussed here. In the Yucatan Peninsula, dissolution features intermediate in size between flank margin and epigenetic continental caves form along the margin of the discharging fresh groundwater lens as a result of freshwater/saltwater mixing. Fresh groundwater discharges are very substantial on the Yucatan carbonate platform, as they are fed by a large volume of allogenic recharge (i.e., recharge of the groundwater from an outside location) from the Yucatan interior (Reference 965). Smart et al. (Reference 965) believe that the Quintana Roo caves (Figure 2.5.1-363) represent a new cave type intermediate in size between flank-margin and epigenetic continental systems.

The Quintana Roo caves located several kilometers interior from the coast may display elements of a dendritic tributary pattern (typical of epigenetic continental caves). Downstream, this drainage passes into an extended zone characterized by a cross-linked anastomosing passage pattern that extends inland from the coast for maximum distances of 8 to 12 kilometers (5 to 7.5 miles) (Reference 965). Large isolated mixing chambers characteristic of the flank margin type caves are absent. Instead, large chambers occur as an element in the anastomosing zone and are generally associated with collapse. Rectilinear maze patterns are generally absent from the caves located in the interior; however, they do appear to be characteristic of some of the coastal caves where fractures have developed parallel to the flank margin (Reference 965).

The passage types in the Quintana Roo caves are horizontal elliptical tubes and canyon-shaped passages and are extensively modified by collapse, but many retain dissolutional wall morphology. The caves are actively enlarging because of undersaturation with respect to calcium carbonate, resulting from the mixing of fresh and saline water. However, according to Smart et al. (Reference 965), many caves in the interior are above the present mixing zone and are characterized by collapse and infill with surface-derived clays, speleothem deposits, and calcite raft sands. Cave sediment fill, speleothem, and ceiling-level data indicate multiple phases of cave development. These multiple phases are associated with glacioeustatic changes in sea level, and alternate in individual passages between active phreatic enlargement and vadose incision and sedimentation. Due to the continued accretion of carbonate rocks along the coast during the Pleistocene, caves that

are now located in the interior of the Yucatan Peninsula were formerly closer to the coast and have gone through multiple phases of cave development. Collapse of the cave roofs is extensive and ubiquitous, which results in the development of crown-collapse surface cenotes. Collapse is a result of the large roof spans caused by lateral expansion of passages at the level of the mixing zone, the low strength of the poorly cemented Pleistocene limestones, and the withdrawal of buoyant support during sea level low stands (Reference 965).

Two critical conditions that control the development of multiphase Quintana Roo caves following glacioeustatic variations in sea level are:

1. When the passage segments remain connected to the underlying deep cave systems and are occupied by the present mixing zone, substantial inflow of saline water maintains the rate of mixing-driven carbonate dissolution, and the predominantly carbonate rock is removed, allowing active passage enlargement to continue.

2. When the links between cave passages are absent, rates of dissolution are low, and passage enlargement ceases (Reference 965).

If the flow of freshwater through a passage is maintained by tributaries, the velocity may be sufficient to prevent accumulation of further sediments or to flush uncemented sediments from the passage and the cave void will remain open. If such freshwater flows are limited or absent due to blockage of the feeders, the passage segment will gradually become occluded by infill and roof collapse (Reference 965).

The greater topographic relief of the cenotes terrain of the Yucatan Peninsula provides a stark contrast with the flat topography seen at the Turkey Point Units 6 & 7 site and in the available bathymetric data for the near-site area of Biscayne Bay. The apparent origin of the greater topographic relief and a much more developed karst regime in the cenotes terrane relative to the Turkey Point Units 6 & 7 site and its vicinity is the relatively high rate of fresh groundwater discharge from a large inland watershed in the Yucatan that produces a more robust mixing zone and more carbonate dissolution (Reference 965). The absence of a more developed karst topography or an active mixing zone near the site (because of the location of the freshwater/saltwater interface as shown in Figure 2.4.12-207 and the presence of saline groundwater at the site as demonstrated by Tables 2.4.12-210 and 2.4.12-211) suggests that the process of shoreline flow that is instrumental in forming the caves on the Yucatan Peninsula is not a mechanism that is likely to produce cavernous limestone with the potential for collapse at the site.

### Hypogene Dissolution

Klimchouk (References 1005 and 1006) has generally described hypogene speleogenesis as dissolution-enlarged permeability (flow) structure development via ascending waters, driven by regional and/or more localized hydraulic potentials (i.e., hydrostatic pressures) or other convective circulation mechanisms. Given the vertical heterogeneity inherent in most sedimentary sequences, this upward groundwater flow implies some hydrological confinement (artesian conditions) rather than surface recharge. In southeastern Florida, confinement is largely provided by the Peace River and middle and upper (non-carbonate) Arcadia formations. Potential for ascending flow (and, by inference, hypogene speleogenesis) thus exists in the lowermost Arcadia Formation and the underlying Suwannee and Ocala limestones, and the Avon Park, Oldsmar, and upper Cedar Keys formations (i.e., the Floridan aquifer system).

In particular, Kohout (References 1007 and 1008) posited that thermally-induced convective circulation was occurring in the Floridan aquifer system within southern Florida. Specifically, Kohout (References 1007 and 1008) suggested upward flow from the lower Floridan aquifer through a middle, semi-confining unit in the aquifer (namely, the Avon Park Formation) and subsequent seaward flow within the upper Floridan aquifer. In the Turkey Point Units 6 & 7 vicinity, the

aforementioned upper Floridan aquifer includes the lower Arcadia, Suwanee, and uppermost Avon Park formations. Aquifer units ascribed to the Ocala limestones are missing in the site vicinity.

Specifically, the Kohout circulation mechanism assumes that horizontal and vertical temperature distributions in the Florida Straits (and Gulf of Mexico) allow cold, dense saline water to flow into the Florida Platform at depth. At depth, this water is warmed by geothermal flow. A corresponding reduction in density produces an upward convective circulation which brings saline water (seawater) into contact with fresh waters recharged via downward flow in central Florida karst regions. Mixing with fresh water results in further density reductions, and allows the diluted seawater (saltwater) to migrate (flow) seaward and discharge (by upward leakage through confining beds) into the shallow coastal zone or deeper submarine springs on the continental shelf and/or slope.

Meyer (Reference 1009) noted that groundwater ages and $^{14}C$ and uranium isotope concentration data within the Floridan aquifer substantiate Kohout convection, and suggested that lateral inland flows associated with the circulation pattern were as high as 52 meters (172 feet) per year in the early Holocene, at least in the so-named boulder zone in the Oldsmar Formation. Meyer (Reference 1009) estimated modern Kohout circulation inland flows (lateral) to be only about 1.5 meters (5 feet) per year. Morrissey et al. (Reference 1010) argued that this decreased flow was associated with increased coastal groundwater levels (i.e., hydraulic head) from long-term Holocene sea level rise, and subsequent reduced hydraulic gradients (and thereby flow velocities) across the Florida platform.

Morrissey et al. (Reference 1010) also suggested that the density difference between seawater and discharging freshwater alone could induce convection in the Floridan aquifer system, as similarly asserted by Sanford et al. (Reference 1011) and Hughes et al. (Reference 1012).

### Possible Hypogene Dissolution on the Florida Peninsula

It is important to note that hypogene karst features do not necessarily manifest at the surface (or should not be expected to manifest at the surface) owing to the aforementioned characteristic separation from meteoric recharge. Surface exposure is typically only provided via surface denudation (e.g., uplift and erosion). Accordingly, direct evidence for hypogene dissolution (from cave morphology) is not readily available for southeastern Florida, as only epigenetic caves are known and accessible.

Very few studies from southeastern Florida explicitly address (or invoke) hypogene dissolution processes as a cave or cavity/void forming mechanism. Most notably, Cunningham and Walker (Reference 958) proposed two hypogene mechanisms to possibly explain structural sags in Biscayne Bay and the Atlantic Ocean: (1) upward groundwater flow via Kohout convection and subsequent carbonate dissolution by mixed fresh and saline waters, and (2) dissolution associated with upward ascending hydrogen-sulfide-rich groundwater, sourced from calcium sulfates in deeper Eocene (or Paleocene) age rocks. These features are described in more detail below.

### Submarine Sag Structures Beneath Biscayne Bay

Cunningham and Walker (References 958 and 989) conducted a study east of the Miami Terrace using high-resolution, multichannel seismic-reflection data (Figure 2.5.1-356). The data exhibit disturbances in parallel seismic reflections that correspond to the carbonate rocks of the Floridan Aquifer system and the lower part of the overlying intermediate confining unit (Figure 2.5.1-357). The disturbances in the seismic reflections are indicative of deformation in carbonate rocks of Eocene to middle Miocene age. This deformation is interpreted to be related to collapsed paleocaves or collapsed paleocave systems and includes fractures, faults, and seismic-sag structural systems (Figure 2.5.1-358) (References 958 and 989).

In general, the seismic-sag structural systems exhibit one or more zones of vertically stacked, concave-upward arrangements of generally parallel seismic-reflection patterns (Figure 2.5.1-358) (References 958 and 989). Twelve seismic sag structural systems have been delineated on the seismic profiles of Cunningham and Walker (Reference 958). Two types of seismic-sag structural systems they have identified are “narrow” and “broad.” The type of system is defined based on the measured differences in the inner sag width of the deformed seismic reflectors. The inner sag width is defined as “the distance between inflection points (i.e., where the shape of the subsidence profile changes from concave to convex) on both sides of the structural trough” (Reference 958).

Collapse related to the “narrow,” seismic-sag structural systems is multistoried as shown in Figure 2.5.1-358 (Reference 958). The uppermost termination of zones of concave upward reflections displayed in many of the narrow sag structures may correspond to paleotopographic expression of the upper surface of paleosinkholes, since many are filled in with onlapping reflections. The onlapping reflections indicate passive sedimentary fill at the top of sagging reflections. This relationship is shown in zones 2 and 3 in the N1 profile in Figure 2.5.1-358. These two zones are indicative of cave collapse and suprastratal deformation during the Eocene. Cunningham and Walker (Reference 958) hypothesize that the association of narrow, seismic-sag structural systems with a possible single fault, in some cases, likely indicates a structural fabric and associated fracture/fault permeability. Although the more recent work by Cunningham and Walker confirms the existence of the seismic-sag structural systems in Biscayne Bay, the authors indicate that both faults and karst collapse systems that might cause disruption in confinement have only been imaged in the middle Eocene to Oligocene part of the Floridan Aquifer system (Reference 989). These faults may have a substantial control on the geographic distribution of some of the narrow seismic sag structural systems (References 958 and 989).

A major collapse event associated with the “broad” seismic-sag structural system is shown in Figure 2.5.1-359. This collapse event occurred in the Eocene based on the deformation of seismic-reflection stratigraphic layer 8 (SS8) reflections which are assigned to Eocene-age rocks. These SS8 reflectors appear to have downlapping relations onto the upper surface of the zone 2 sag structures and truncate reflectors at the top of the zone 2 structure (Reference 958).

Cunningham and Walker (Reference 958) suggested three possible mechanisms for the formation of the seismic sag structures: (1) “corrosion” or dissolution by an Eocene mixed freshwater/saltwater zone associated with regional groundwater flow, (2) upward groundwater flow during the Eocene driven by Kohout convection (the circulation of relatively warm saline groundwater deep in carbonate platforms and subsequent mixing with meteoric water as it rises), and (3) upward ascension of hydrogen sulfide-charged groundwater, with the hydrogen sulfide derived from the dissolution and reduction of calcium sulfates in the deeper Eocene or Paleocene rocks (Reference 958).

As noted above, the broad sag structures in Biscayne Bay are multi-storied (vertically stacked) features that can be interpreted as evidence for coalesced, collapsed, multi-story maze paleocave systems and associated deformation (fractures, faults, sagging, etc.). Narrower stacked sag structures, in turn, can be interpreted as evidence for more isolated (i.e., individual) subsurface void collapses. Generally, the hypogene dissolution process (speleogenesis) is associated with such multi-story maze caves and isolated subsurface cavities/voids. Although Cunningham and Walker (Reference 958) did not explicitly attribute the aforementioned sags to hypogene dissolution processes, the vertical stacking is consistent with collapse in a multi-story hypogene cave system, as described by Klimchouk (Reference 1005). Nevertheless, Cunningham (Reference 999) cites evidence (unspecified) for hypogenic karst collapse in just one southeast Florida location, a borehole (well) in the Miami-Dade Water and Sewer Department (MDWASD) northern wastewater injection field, at depths attributed to the much deeper and older Avon Park and Oldsmar formations. Critically, though, it should be noted that Cunningham and Walker (Reference 958) present no tangible evidence to support a hypogene origin for these features, either via Kohout circulation and fresh/salt water mixing or dissolution by hydrogen-sulfide-rich waters. Moreover, Cunningham (Reference 999)

has suggested that a different sag feature within the MDWASD's southern wastewater injection field could reflect subaerial exposure and sinkhole development (i.e., epigenetic dissolution) along a major sedimentation and subsidence stratigraphic/sequence boundary.

Regardless of the mechanism of formation of the submarine sags beneath Biscayne Bay, the geophysical data indicate the absence of deformation in rocks younger than Pliocene (Figures 2.5.1-357, 2.5.1-358, and 2.5.1-359). This finding suggests that if the same mechanism had been active at the Turkey Point Units 6 & 7 site during the Eocene, none of the strata younger than Pliocene would be deformed. These younger strata include the Miami Limestone, Key Largo Limestone, Fort Thompson Formation, and Upper Tamiami Formation.

### Onshore Sag Structures in Broward and Miami-Dade Counties

In addition to the 12 sag structures imaged in Biscayne Bay, Cunningham and others (References 999, 1013, 1014, and 1015) have identified 24 onshore sag structures in northeastern Miami-Dade and eastern Broward counties (Figure 2.5.1-391). These features are also interpreted as paleokarst sinkholes or faults and fractures and have the same formation history as the broad and narrow seismic sag structural systems in Biscayne Bay (Reference 958).

Cunningham (Reference 999) noted that some onshore sag structures show deformation-related features affecting younger units than those imaged beneath Biscayne Bay, where offset reflectors attributed to paleokarst-related faults and fractures were observed to have their upper extents in the Mid-Miocene aged units (Reference 958). Whereas most paleokarst faults and fractures in the onshore canals were observed to truncate at the upper surface of the Middle Miocene Arcadia Formation or lower, some in northern Broward County are observed to affect reflectors in the upper Miocene Peace River Formation as well as the overlying Ochopee Member of the Pliocene Tamiami Formation (References 1013 and 1015) (Figure 2.5.1-392). However, there is no deformation reported to affect units younger than the lower Pliocene, suggesting that the timing of karst formation and/or collapse into paleocave systems includes the Miocene and Pliocene Epochs. Furthermore, the onshore seismic sags closest to the site are not observed to affect any units above the top of the Arcadia Formation.

It is possible that the sags and recognized faults that cut through the upper surface of the Arcadia Formation form conduits for groundwater flow between the permeable zones of the Floridan aquifer system (Reference 999). This is indicated by detection of treated effluent in the uppermost major permeable zone of the lower Floridan aquifer that was injected into the deeper Boulder Zone (References 1013 and 999). The detection of the treated effluent implies density-related, upward migration of fluids being the result of the lack of confinement between the two permeable zones, presumably enhanced by paleokarst features associated with the karst-collapse structure imaged on the onshore profiles (References 1013 and 999).

The imaging of 24 seismic-sag features along 145 linear kilometers (90 miles) of seismic reflection acquisition would seem to imply that there are many more paleokarst collapse features that exist below Broward and Miami-Dade counties, and south Florida in general, that have yet to be discovered. Since none of these filled paleokarst collapse features have been observed to affect units younger than Early Pliocene and there is no known surface expression, it is unlikely that they pose any hazard to the stability of the south Florida ground surface. Likewise, if any such features would happen to exist below the site vicinity or site, there is no reason to believe that they would pose a threat to the surface collapse at the site due to the thickness of the overlying strata.

### Other Paleokarst Collapse Structures in Southern Florida

Several other paleokarst collapse structures have been identified on the southern Florida Peninsula and Florida Platform. These features are discussed below.

### Jewfish Creek Paleokarst Feature

A paleokarst feature of possible similar origin to the imaged sag structures in Biscayne Bay and Miami-Dade and Broward counties was identified during design work for a new bridge across Jewfish Creek and adjacent Lake Surprise on northern Key Largo (Figure 2.5.1-390) (Reference 1016).

Specifically, data from 34 geotechnical borings located on Jewfish Creek and within Lake Surprise provided evidence for localized loose sand layers that was interpreted as possible evidence for sediment transport (i.e., piping) into dissolution cavities (Reference 1016). At some locations, drilling water (circulation) losses were also observed, suggesting voids and/or highly permeable subsurface layers. For the most part, these water losses were concentrated at depths between 6 meters and 30 meters (20 feet and 100 feet).

Microgravity surveys over the same area provided evidence for a 100 microgal (µGal) anomaly centered between Jewfish Creek and Lake Surprise (Reference 1016). Generally, this gravity anomaly coincided with the aforementioned borehole locations showing evidence for cavities. Supplemental shallow and deep seismic reflection surveys in Lake Surprise also provided evidence for downward dipping reflectors located near the aforementioned gravity anomaly center and edges, and identified seven collapse (subsidence) structures filled with sediments derived from overlying materials. Generally, these collapse structures ranged in width from 30 to 60 meters (100 to 200 feet) and were distributed over a 580 meters (1900 feet) distance.

The aforementioned structures at Jewfish Creek/Lake Surprise were specifically interpreted as localized collapses, or collapse features associated with closely spaced and enlarged dissolution joints (Reference 1016). The largest subsidence structure in particular was interpreted as a cavity collapse in a soluble limestone layer, the Arcadia Formation, at depths below approximately 213 meters (700 feet). Corresponding subsidence in overlying Arcadia Formation layers, and in younger unconsolidated sands and capping limestone, inferred to be the Peace River, Tamiami, Caloosahatchee or possibly Fort Thompson, and Key Largo formations, was also interpreted from the seismic reflection data, at depths between approximately 21 meters and 213 meters (70 feet and 700 feet). Density logs from geotechnical borings located adjacent to the collapse structure indicated voids and porous zones in the shallower formations, primarily between 6.1 meters and 21.3 meters (20 feet and 70 feet).

A clear formation mechanism for the Jewfish Creek/Lake Surprise feature has not been indicated, but it has been intimated to be epigenetic (rather than hypogene) origin (Reference 1016). It should be noted that the collapse structures at Jewfish Creek/Lake Surprise are interpreted to be centered in the late Oligocene to early Miocene age Arcadia Formation (Reference 1016). It is possible, then, that the collapsed cavities (or collapsed joints) were formed (or enlarged) via subaerial exposure and downward meteoric dissolution (i.e., epigenic or hypergenic dissolution) during middle to late Miocene sea level lowstands, estimated to be 300 meters (985 feet) below modern sea level, with considerations/corrections for subsidence (Reference 951). Alternatively, void formation may be linked to eogenetic (or syngenetic) dissolution processes, namely submarine groundwater discharge during sea level highstands and consequent enhanced carbonate dissolution at a former freshwater/saltwater interface, as described previously.

Because the collapse structures (as inferred from dipping strata) at Jewfish Creek/Lake Surprise extend upward into the Key Largo Formation, void development and joint enlargement could also be attributed, in part, to epigenic dissolution by meteoric waters (at least in more near-surface layers) during Pleistocene sea level lowstands. It is important to note, however, that the deep collapse origination point (at least 152 meters [500-feet deep]) precludes subaerial exposure during the maximum estimated late Pleistocene sea level lowstand (125 meters [410 feet]). Void collapse (and subsidence) at Jewfish Creek/Lake Surprise may instead represent cave or joint enlargement via phreatic dissolution, and subsequent (later) cave or joint collapse in the Pleistocene due to sea level lowering induced buoyancy losses (but not exposure) and corresponding load changes within

overlying layers. Phreatic dissolution, in this case, would (again) likely have been associated with freshwater/saltwater mixing.

As noted above, the collapse structures at Jewfish Creek/Lake Surprise are also not entirely inconsistent with the narrow structural sag features described by Cunningham and Walker (Reference 958) and subsequently by Reese and Cunningham (Reference 1013) and Cunningham (References 999, 1014, and 1015) in Biscayne Bay and northeastern Miami-Dade and eastern Broward counties. However, these sag features are generally vertically stacked (multi-storied) and are not closely spaced or distributed horizontally as is the case at Jewfish Creek/Lake Surprise.

### Government Cut Collapse Structure

Another possible onshore paleo-collapse feature was identified during geophysical and geotechnical investigations for a tunneling project in Miami Harbor, under the Government Cut shipping channel (Figure 2.5.1-390) (Reference 1017). Located approximately 42 kilometers (26 miles) northeast from the Turkey Point Units 6 & 7 site and 25 kilometers (15 miles) north from the sag features in Biscayne Bay, this feature was described as exhibiting “soft zones” and cavities exceeding 3 meters (10 feet) diameter, at depths (below sea level) between 20 meters and 30 meters (65 feet and 100 feet). Because this structure occurs at relatively shallow depths, it is likely associated with subaerial exposure and epigenic dissolution during sea level lowstands.

### Submarine Sinkholes

Land et al. (Reference 1018) and, later, Land and Paull (Reference 951) mapped nine submarine sinkholes in the Florida Straits, on the Pourtalès and Miami terraces (Figure 2.5.1-390) at depths between 244 meters and 575 meters (800 feet and 1886 feet). Long and short axes and depths in the sinkholes average about 630 meters (2065 feet) and 440 meters (1444 feet) and 100 meters (328 feet) respectively. Land et al. (Reference 1018) (and Land and Paull [Reference 951]) interpreted the features as having formed underwater, and as rooted in Eocene and Oligocene limestones. Land and Paull (Reference 951) also suggested that some were possibly still active, given incomplete infilling.

Cunningham and Walker (Reference 958) suggested that the Pourtalès and Miami terrace sinkholes were evidence for freshwater/saltwater mixing resulting from upward flow driven by Kohout circulation, as described above. Kohout (References 1007 and 1008) predicted mixing and upward flow in these areas, but did not explicitly recognize the sinkholes as direct evidence for discharges. Land et al. (Reference 1018) and Land and Paull (Reference 951) only intimated that upward convective circulation could be responsible for the sinkholes, noting that the Pourtalès and Miami terrace sinkholes were laterally continuous with the Floridan aquifer and thus could simply represent past (or even present) freshwater discharge and dissolution associated with freshwater/saltwater mixing in the immediate discharge zone (i.e., non-hypogene mixing zone dissolution).

Generally, the Pourtalès and Miami terrace sinkhole locations are consistent with groundwater discharge loci (and inferred enhanced dissolution zones) generally predicted by numerical groundwater circulation models for the southern Florida Platform (Reference 1019). Under sea level highstand boundary conditions, these models predict increased saltwater encroachment (rather than discharge) and dissolution at depths exceeding 500 meters (1640 feet) (i.e., near the Florida Platform base). Contrastingly, lowstand model conditions predict increased groundwater discharge, and suggest that mixing will occur along the upper platform margin, at shallower depths.

Land et al. (Reference 1018) and Land and Paull (Reference 951) mapped only one sinkhole on the Pourtalès Terrace at depths greater than the aforementioned 500 meters (1640 feet) sea level highstand model dissolution limit. Land et al. (Reference 1018) noted that this karst feature, located near the base of the Pourtalès Terrace at a 575 meters (1886 feet) water depth, is positioned in a Quaternary sediment drape, suggesting recent formation, and thus is consistent with highstand model predictions for persistent submergence. Land and Paull (Reference 951) mapped the

remaining Pourtalès sinkholes at depths between approximately 350 meters and 460 meters (1148 feet and 1509 feet). Generally, then, these additional sinkholes are consistent with groundwater discharge during an earlier sea level lowstand (probably in the Miocene).

It should be noted that two other potential collapse features, described as sinkholes, have also been identified offshore from Broward County, approximately 90 kilometers (56 miles) north-northeast from the site. Specifically, two large depression features were identified by ENTRIX, Inc. during geophysical surveys conducted to support a now withdrawn application for the proposed Calypso liquefied natural gas (LNG) deep water port facility (Reference 1020). Centered roughly 10 kilometers and 16 kilometers (6 miles and 10 miles) from the shore, as depicted on Figure 2.5.1-390, these features were estimated as 670 meters (2200 feet) and 365 meters (1200 feet) in diameter, respectively.

ENTRIX, Inc. (Reference 1020) specified that the southernmost Calypso feature (Calypso Port 1 on Figure 2.5.1-390) exhibited surface expression and thus indicates possible continued subsidence. In contrast, the northern sinkhole feature (Calypso Port 2) was evident only in the subsurface. Given location and size, it is likely that the Calypso Port 1 and 2 features likely formed coincident with other submarine sinkholes in the Florida Straits described by Land and Paull (Reference 951) and Land et al. (Reference 1018).

### Crescent Beach Spring and Red Snapper Sink, Off the Coast of Northeast Florida

Crescent Beach Spring and Red Snapper Sink are located outside of the 200-mile radius site region, but the spring and sink are evidence of deep pore water upwelling and warrant discussion here. Crescent Beach Spring, a freshwater spring, is located approximately 4 kilometers (2.5 miles) east of Crescent Beach, Florida (Figure 2.5.1-364) and is considered a first-order magnitude spring with a flow rate of greater than 40 cubic meters/second (greater than 1400 cubic feet/second) (Reference 946). The spring is located at a depth of 18 meters (59 feet) in the Atlantic Ocean, and erosion of confining strata to a depth of 38 meters (125 feet) at the mouth of the vent has enabled direct hydrologic communication of confined groundwater in the Floridan Aquifer with coastal bottom waters (Reference 946).

The Red Snapper Sink (Figure 2.5.1-364) is located approximately 42 kilometers (26 miles) off Crescent Beach and is incised approximately 127 meters (417 feet) into the continental shelf at a water depth of 28 meters (99 feet). Divers investigating the site observed that seawater was flowing into small caves at the base of the hole, indicating possible recharge of the Floridan Aquifer, and that the water in the bottom of the hole was similar in salinity and sulfate content to ambient seawater. According to Moore (Reference 946), Red Snapper Sink was similar to Crescent Beach Spring before the piezometric head was lowered along the coast, and preservation of the feature suggests that a freshwater spring was active at this site in the recent past.

The existence of Crescent Beach Spring and, by inference, Red Snapper Sink indicates the presence of abundant fresh groundwater within confined aquifers on the continental shelf. Breaching of the confining layer overlying such aquifers by erosional or tectonic mechanisms has the potential to create similar submarine springs on the shelf off southern Florida. No capable faults that could induce a breach of the confining layer have been identified in the site vicinity (Subsection 2.5.3.6). Groundwater in the Biscayne Aquifer (the surficial aquifer) is saline (Tables 2.4.12-210 and 2.4.12-211). Therefore, dissolution of carbonate rocks in the vicinity of deep pore water upwelling from this aquifer into the overlying ocean is not probable. At the site, the underlying Tamiami Formation and Hawthorne Group combined comprise more than approximately 152 meters (500 feet) of low-permeability rocks and sediments that overlie and confine the Floridan Aquifer (Figures 2.4.12-202 and 2.4.12-204). Deep pore water upwelling generally occurs well offshore, where the slope of the shelf is steeper and erosion of this thickness of confining sediments is more likely. For this reason, carbonate dissolution associated with deep pore water upwelling from the Floridan Aquifer is not likely to pose a threat of surface collapse or sinkhole hazard at the site.

### Sea Level Changes and Migration of Freshwater/Saltwater Interface and Conclusion

The freshwater/saltwater interface and the zone of diffusion (potential dissolution) is located approximately 10 to 13 kilometers (6 to 8 miles) inland from the Turkey Point Units 6 & 7 site. The site, therefore, occupies a saline hydrogeochemical environment that is different from the locations of active carbonate dissolution and karst development discussed above. This difference provides an explanation for the apparent absence of large dissolution features.

Since a potential rise in sea level at the Turkey Point Units 6 & 7 site would cause the freshwater/saltwater interface to migrate inland and away from the site, a rise in sea level is not expected to result in dissolution of the subsurface limestones. Conversely, a fall in sea level would cause the freshwater/saltwater interface to migrate toward the site and could result in carbonate dissolution (Reference 956).

Karstification can occur at the ground surface resulting in sinkholes. These sinkholes create depressions that can be filled with water or sediment. There are three main types of sinkholes common to Florida (Reference 264):

- Solution sinkholes occur where limestone is exposed at the ground surface or is covered with a thin mantle of material. Dissolution is concentrated at the surface and along joints, fractures, or other openings in the rock. The development of such features is accomplished by a slow drop of the ground surface that results in the formation of a depression that is commonly filled with organic-rich sediments. These sinkholes typically manifest themselves as bowl-shaped depressions at the ground surface.

- Cover-collapse sinkholes occur where a solution cavity develops in the limestone to a size such that the overlying material cannot support its own weight. The result is generally a sudden collapse of the overburden into the cavity. These sinkholes are common in areas where limestone is close to the ground surface and under water-table conditions, with accelerated dissolution occurring in limestone zones at and just below the water table.

- Cover-subsidence sinkholes occur where the overburden is comprised of unconsolidated and permeable sands. They form when the sand slowly moves downward into space formerly occupied by other sediments, which have already moved downward into space formerly occupied by limestone that has been removed by dissolution. These sinkholes generally develop gradually.

The Florida Geological Survey (FGS) classifies sinkhole occurrences into four type areas (Reference 264) (Figure 2.5.1-222):

- Area I — Bare or thinly covered limestone. Sinkholes are few, generally shallow and broad, and develop gradually. Solution sinkholes dominate. The Units 6 & 7 site is located in Area I (Subsection 2.5.1.2.4).

- Area II — Cover is 9 to 60 meters (30 to 200 feet) thick and consists mainly of incohesive and permeable sand. Sinkholes are few, shallow, of small diameter, and develop gradually. Cover-subsidence sinkholes dominate.

- Area III — Cover is 9 to 60 meters (30 to 200 feet) thick and consists mainly of cohesive clayey sediments of low permeability. Sinkholes are numerous, of varying size, and develop abruptly. Cover-collapse sinkholes dominate.

- Area IV — Cover is more than 60 meters (200 feet) thick and consists of cohesive sediments interlayered with discontinuous carbonate beds. Sinkholes are few in number. However, the ones that do occur are generally large in diameter and deep. Cover-collapse sinkholes dominate.

It should be noted that the FGS have catalogued only five sinkhole openings (surface collapses) in Miami-Dade and Broward counties (Figure 2.5.1-390) (Reference 1021). FGS subsidence incident reports indicate that these features are not directly related to active rock dissolution and subsequent collapse, but instead to infrastructure issues, namely sediment piping associated with broken hydrants and water mains or leaky storm drain boxes (i.e., urban development).

Finally, karstification has a significant effect on the strength and density of thick carbonate sequences. Over time, limestone strata can become highly porous and weakened by karstification. The structural stability of extensive carbonate platforms in the Caribbean can be affected by karstification (Reference 263). The weakened platforms have undergone extensive escarpment collapse, similar to that described in Subsection 2.5.1.1.1.1.1.2. In addition, karstification during glacial periods of low sea level decreases the density of carbonate sediments by significant amounts and enhances the vertical tectonics of isostatic rebound (Reference 262), as discussed in Subsection 2.5.1.1.1.1.1.1 with respect to glacial cycles and sea-level fluctuations.

### Primary and Secondary Physiographic Provinces of the Florida Peninsula

Recent work by the Florida Department of Environmental Protection and the FGS now favors an organization by primary and secondary physiographic provinces. This subsection reflects the new organization and does not refer to the three former physiographic zones of Florida. The hierarchy of primary and secondary physiographic zones pertinent to the site region is outlined in the table below.

**Florida's Secondary Physiography Provinces within the 200-Mile Radius Site Region[1]**

| Atlantic Coastal Lowlands | Intermediate Coastal Lowlands | Gulf Coastal Lowlands | Central Highlands |
|---|---|---|---|
| • Atlantic Coastal Ridge<br>• Ten-Mile Ridge<br>• Eastern Valley<br>• Osceola Plain<br>• Crescent City Ridge<br>• Bombing Range Ridge | • Florida Keys, including Coral and Oolite Keys<br>• Florida Bay Mangrove Islands<br>• Everglades<br>• Big Cyprus Spur<br>• Immokalee Rise<br>• Southwestern Slope<br>• Southern Slope<br>• Okeechobee Plain<br>• Caloosahatchee Valley | • Reticulate Coastal Swamps and Ten Thousand Islands<br>• Gulf Coastal Swamps and Drowned Coastal Karst<br>• Gulf Coastal Lagoons and Barrier Chains<br>• Gulf Coastal Terraces<br>• Gulf Coastal Estuaries<br>• DeSoto Plain<br>• Polk Upland | • Lake Wales Ridge<br>• Winter Haven Ridge<br>• Lake Henry Ridge<br>• Lakeland Ridge |

Note: modified from Reference 266

The physiography of the Florida Peninsula and continental margin is divided by Schmidt (Reference 266) into seven primary physiographic provinces; Atlantic Coastal Lowlands, Intermediate Coastal Lowlands, Gulf Coastal Lowlands, Central Highlands, Northern Highlands, and the Marianna Lowlands. Two of these, the Northern Highlands and the Marianna Lowlands, are in northern Florida and are outside of the 200-mile radius site region.

Florida's terrestrial geomorphology (i.e., not including the Atlantic Continental Shelf and Slope) is further subdivided by Schmidt (Reference 266) into a number of secondary and tertiary

1. Small or scattered physiographic provinces may not be identified on Figure 2.5.1-217.

physiographic provinces. The secondary physiographic provinces associated with each primary province are listed in the above table and are shown in Figure 2.5.1-217. The tertiary physiographic provinces are not described in Section 2.5. The following subsections provide a description of geomorphologic features of the southern and central Florida Peninsula within the 200-mile radius site region. The paragraphs address the physiographic subprovinces of the Atlantic Coastal Lowlands, followed by those of the Intermediate Coastal Lowlands, the Gulf Coastal Lowlands, and, finally, the Central Highlands.

### Atlantic Coastal Lowlands Primary Physiographic Province

Atlantic Coastal Ridge Physiographic Subprovince

The Atlantic Coastal Ridge physiographic subprovince is a ridge of sand overlying limestone that ranges in elevation from approximately 10 to 50 feet (3 to 15 meters) above sea level (Figure 2.5.1-217). It averages approximately 5 miles (8 kilometers) wide and is breached in places by shallow sloughs (Reference 267). It is comprised of single and multiple relict beach ridges and bars (Reference 265).

The Atlantic Coastal Ridge (Figure 2.5.1-217) most likely formed during Pamlico time when sea level was approximately 30 feet (9 meters) higher than today. The eastern slope of the ridge mimics the present continental slope because of a sea level regression caused by a rapid onset of glaciation (see discussion of coastal features related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). Most of the eastern coast of Florida is an erosional shoreline rather than a prograding one (Reference 265).

In southern Florida, the narrow part of the Atlantic Coastal Ridge is a relict beach ridge that surmounts the crest of the remnant Pamlico offshore scarp (steep slope). The ridge and scarp are essentially preserved in their original form (Reference 265). The wider and higher relief of the Atlantic Coastal Ridge is due in part to deposition from terrigenous local sources. The terrigenous local sources consist of quartz sand deposits with broken fragments of contemporaneous shells in the beach sands. The Pamlico Scarp contributes quartzose sand along the Atlantic Coastal Ridge (Reference 265). The Atlantic Coastal Ridge disappears south of Florida City but reappears in the lower Florida Keys (Figure 2.5.1-217). The ridge was probably formed at approximately the same time as the Talbot terrace as an irregular limey bar (see discussion of dating of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 268).

The Atlantic Coastal Ridge, lagoons, and barrier chain is wider in central Florida than it is in southern Florida. The province is made of relict beach ridges and bars, sometimes single and sometimes multiple. The Atlantic Coastal Ridge appears to be almost wholly a product a time when sea level was about 30 feet (9 meters) higher than it now is. The eastern slope of the Atlantic Coastal Ridge closely resembles the present submarine slope that is very uniform offshore from the ocean beaches throughout the length of the Atlantic Coastal Plain. These offshore submarine slopes drop off seaward, steeply at first and then more gently until they are mostly flat at a depth of about 30 feet (9 meters). Because this slope is so persistent, the Pamlico offshore profile appears to have emerged rapidly without having been altered by later sea level changes. White (Reference 265) speculates that the clean, undamaged emergence may be due to the fact that marine regression from the Pamlico level was caused by rapid onset of glaciation.

Ten-Mile Ridge Physiographic Subprovince

Ten-Mile Ridge comprises the narrow, discontinuous remnants of a former coast ridge a few miles inland from the Atlantic Coastal Ridge of central Florida (Figure 2.5.1-217). Ten Mile Ridge is one of five distinct landscapes in Florida that preserves scrub vegetation, a rare and vanishing ecosystem. The scrub vegetation landscape is associated with ridges of well drained to moderately well drained soils. The five Florida landscapes preserving scrub vegetation are: (a) the recent barrier islands including Cape Canaveral, (b) Merritt Island, (c) the Atlantic Coastal Ridge, (d) the Ten Mile Ridge,

and (e) a small ridge in the southwest corner of Brevard County. These areas differ in age and topography, but all have a similar origin as coastal dunes (Reference 269).

Ten-Mile Ridge was once a line of dunes but now serves as the elevation upon which Interstate 95 runs. Although topographically low, it is sufficient to prevent water from draining from the interior into the Indian River lagoon. The result is the formation of a marsh west of the ridge that is some 20 miles (32 kilometers) wide and 30 miles (48 kilometers) long. The Ten-Mile Ridge is part of the Pamlico Terrace (Reference 270) (Figure 2.5.1-220), from the Late Pleistocene (Sangamonian) interglacial period (Reference 271).

Eastern Valley Physiographic Subprovince

The southern limit of the Eastern Valley province lies just north of the Atlantic Coastal Ridge and east of the Everglades subprovince of southern Florida (Figure 2.5.1-217). The Eastern Valley is a broad, flat valley with elevations of about 25 feet (7.6 meters), slightly lower than the level of the Pamlico stand. There are indications that the topographic surface was higher and has been reduced by solution of the constituent shell fraction (Reference 265). Thus, the area may precede the Pamlico stand. There are relict beach ridges throughout much of the length and width of the valley, indicating that the landform is a regressional or progradational beach ridge plain (Reference 265).

The Eastern Valley physiographic subprovince of central Florida (Figure 2.5.1-217) is a broad flat valley in Seminole and Indian River counties. Its elevation varies from 20 to 30 feet (6 to 9 meters) above sea level. There are relicts of beach ridges that at one time constituted a regressional or progradational beach ridge plain (Reference 265).

The head of the St. Johns River consists of a broad swampy valley with lakes. The river flows through each lake along its longest axis. This suggests that at one time there was a standing body of water that has been filled with sediments and vegetation between the upper levels of the lakes that eventually formed the flat, swampy flood plain; the unfilled places became the current chain of lakes in the St. Johns River's headwaters (Reference 265).

Southward of the St. Johns River, the topography has approximately 5 feet (1.5 meters) of local relief throughout the area. This topography is bounded by the headwaters of the St. Johns River at the north, the bounding scarp of the Eastern Valley on the west, the St. Lucie Canal on the south, and Ten Mile and Atlantic Coastal Ridges on the east (Figure 2.5.1-217). The surface of the entire area has elevations close to 25 to 30 feet (7.6 to 9 meters) (Reference 265).

Osceola Plain Physiographic Subprovince

The Osceola Plain physiographic subprovince (Figure 2.5.1-217) extends southeasterly through eastern Okeechobee County, extreme southwestern St. Lucie County, and into western Martin County. It is bounded on the west and northwest by the Lake Wales Ridge and the southern ends of the Mount Dora and Orlando Ridges. On the northeast, east, and south it is bounded by an outward-facing erosional ridge (Reference 265). The Osceola Plain reaches approximately 90 to 95 feet (27 to 29 meters) in elevation near its northern edge. It reaches an elevation of 80 feet (24 meters) east and northeast of Lake Kissimmee. Its local relief is very small, with variations of 10 feet (3 meters) across the entire subprovince (Reference 265).

The Kissimmee River passes roughly west of the Osceola Plain. The river is confined to a valley for 25 miles (40 kilometers) south of Lake Kissimmee. North of Lake Kissimmee, several lakes occupy most of the Osceola Plain (Reference 265). The Arbuckle Creek on the western side of the Osceola Plain (west of the Bombing Range Ridge) drains Lake Arbuckle into Lake Istokpoga below the southern bounding scarp of the Osceola Plain (Reference 265) (Figure 2.5.1-217).

Crescent City Ridge Physiographic Subprovince

Crescent City Ridge (Figure 2.5.1-217) is located in southeast Putnam County and northwest Volusia County in east-central Florida. The altitude of the land surface ranges from near mean sea level at Crescent Lake to approximately 120 feet (37 meters) above mean sea level in the sandy ridges. The topography is characterized by a series of terraces, or step-like surfaces of increasing elevation, which are the result of wave erosion and deposition during the advance and retreat of sea level during the Pleistocene glacial maxima. These marine terraces have been dissected by varying degrees of erosion and are capped by thin surficial sands (Reference 272).

The Crescent City Ridge also exhibits karst topographic features resulting from the dissolution of the underlying limestone formations. Such areas are characterized by high local relief, a lack of surface drainage features, subsurface drainage, sinkholes, and sinkhole-related lakes and springs. Due to the lack of surface drainage in the upland sandy ridges, almost all of the precipitation falling in these areas is either lost to evapotranspiration or drains downward through the permeable soils or sinkholes (Reference 272).

Bombing Range Ridge Physiographic Subprovince

Bombing Range Ridge is a small subprovince located east of elongate Lake Wales Ridge (Figure 2.5.1-217). The lakes of the Bombing Range Ridge and the northern Lake Wales Ridge are darker colored with higher nutrients than the lakes found on the southern Lake Wales Ridge. Elevations are 70 to 130 feet (~21 to 40 meters), and there are more extensive areas of poorly drained soils, such as the Satellite and Basinger series. Peaty muck Samsula soils border many of the lakes (Reference 291).

**Intermediate Coastal Lowlands Primary Physiographic Province**

Florida Keys Physiographic Subprovince

The Florida Keys physiographic subprovince (Figure 2.5.1-217) is a narrow chain of small islands at the southern tip of the Florida Peninsula. The Florida Keys consist of the High Coral Keys, Low Coral Keys, and the Oolite Keys. The keys are composed of Pleistocene reef sediments (Key Largo and Miami Limestones). There is not a clear dividing line between the High and Low Coral Keys; however, the Key Largo Limestone predominates in the High and Low Coral Keys whereas the Miami Limestone dominates in the Oolite Keys (Reference 274).

The Florida Keys extends for approximately 150 miles (240 kilometers) from Miami southwest to Key West. It is bounded by the Atlantic Ocean and the Bay of Florida, inland waters, and the Gulf of Mexico (Reference 274). The edge of the continental shelf parallels the Florida Keys approximately 7 miles (11.3 kilometers) offshore. Maximum elevations reach 10 to 12 feet (~3 to 4 meters) above mean sea level. On the offshore islands, elevations are less than 10 feet (3 meters) above mean sea level. Average depths in Hawk Channel, which parallels the shore in Florida State waters, are 13 to 15 feet (~4 to 5 meters). In the Florida Keys coral replaces the sand in barrier bars and islands along the southern coast of Florida (Reference 273).

A topographic high beneath Key Largo was the focus of reef growth (Reference 275). The late Tertiary siliciclastic sediments underlying the Quaternary carbonate rocks appears to control the position and arc shape of the recent shelf and slope of southern Florida (Reference 275). Additionally, the arc pattern of the Florida Keys is related to the bathymetry of the shelf edge and the Florida current. The growth of patch reefs depends on nutrient availability, sea level, and topography.

Florida Bay Mangrove Islands Physiographic Subprovince

The island and shoals of Florida Bay (Figure 2.5.1-217) are made of lime mud and their surfaces usually lie within a foot above or below normal water level. The westernmost of these banks are sandy and extend southeastward from East Cape near Cape Sable. These banks face the deeper,

open waters of the Gulf of Mexico on their west side. The remaining banks further east are made of mud and form a network of long, narrow vegetation-covered shoals and mangrove-covered islands (Reference 271).

Everglades Physiographic Subprovince

The Everglades physiographic subprovince (Figure 2.5.1-217) extends from Lake Okeechobee southward toward Florida Bay. It consists of the Everglades, Big Cypress Spur, Florida Bay, and coastal mangroves (References 267 and 276). The Everglades is a wetland prairie created by the overflow of Lake Okeechobee, whose water spreads in a slow-moving sheet flow across a slope of less than 2 inches/mile from Lake Okeechobee to the mangrove lined margins of southwest Florida near Florida Bay and the Gulf of Mexico (References 267 and 277).

Elevations in the Everglades range from 14 feet (4 meters) near Lake Okeechobee to sea level at Florida Bay. Before development of the canals in the northern portion of the Everglades, water discharged from the Everglades into the Florida Bay, the Gulf of Mexico, and the Atlantic Ocean through small rivers in the Atlantic Coastal Ridge or as seepage and spring flow into Biscayne Bay (see discussion of changes in drainage related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 267).

The vegetative wetland community and landscapes consist of a central core of peat that extended from Lake Okeechobee to Florida Bay. Organic soils (peat) overlie the limestone throughout most of the Everglades. The fibrous peat accumulates on limestones because the limestone can be dissolved down to the water table. This results in swampy conditions for the growth of fibrous swamp plants and their preservation as fibrous peat (Reference 265).

The water flowing through the Everglades is only a few inches deep, but is 50 miles (80 kilometers) wide. The predominant vegetation, sawgrass, is flooded during the wet season (summer) and burned (parched) during the dry season (winter/spring). During flood periods, the movement of water causes tree islands to develop an alignment pattern parallel to the lines of surface water flow (Reference 267). Interspersed through the sawgrass are deeper water sloughs where tree islands or hammocks appear. The coastal mangrove swamps form very thick, dense thickets at the coastline where the sawgrass and cypress swamps meet (References 267 and 276).

Big Cypress Spur Physiographic Subprovince

The Big Cypress Spur physiographic subprovince (Figure 2.5.1-217) is located west of the Everglades. The land surface is flat except for low-mounded limestone outcrops and small, oval, elongated depressions in the limestone. Water drains to the south and southwest through cypress strands into the coastal mangroves (Reference 267). The elevations are below 16 feet (5 meters), and the physiography consists of prairies, marshes, and stunted cypress (Reference 278). The subprovince is dominated by cypress, pine, and wet prairie. The Big Cypress Spur receives approximately 50 inches of rainfall per year but does not have overland flow similar to that of the Everglades (References 267 and 276).

Immokalee Rise Physiographic Subprovince

The Immokalee Rise physiographic subprovince (Figure 2.5.1-217) is north of the Big Cypress Spur, west of the Everglades, and south of the Caloosahatchee Valley. It appears to be a southern extension of Pamlico marine sand because it exhibits several relicts of Pamlico shoreline features (see discussion of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). According to White (Reference 265), the Immokalee Rise was built as a submarine shoal that extended southward from a mainland cape at the south end of the DeSoto Plain. The Immokalee Rise is ringed with small peripheral lakes that formed as a result of limestone dissolution (Reference 265). Elevations range between 30 and 40 feet (9 to 12 meters) (Reference 278).

The Immokalee Rise includes most of Hendry County and eastern Lee County. It is about 8 meters (25 feet) in elevation, but can peak at 11 meters (36 feet) and 13 meters (43 feet) in some areas. All soils are deep, nearly level, and poorly drained, with a water table less than 25 centimeters (10 inches) from the surface during at least part of the year (Reference 279).

Southwestern Slope Physiographic Subprovince

The Southwestern Slope physiographic subprovince (Figure 2.5.1-217) is located along the eastern shore of the Gulf of Mexico, west of the Immokalee Rise and Big Cypress Spur. The landscape is relatively flat and underlain with an uneven bedrock surface that is usually covered by a veneer of soils. The soils are relatively modern and in the process of formation from surficial sediments such as sand and calcareous marl mixing with organic peat and muck components. Tidal marshes, sea grass beds, and mangroves develop in this region of Florida because of the low wave energy (Reference 267).

Southern Slope Physiographic Subprovince

The Southern Slope physiographic subprovince (Figure 2.5.1-217) is located along the southern shore Florida Bay, south of the Everglades and the Big Cypress Spur. The topography is at or near sea level, and the area consists of broad bands of swamps and marshes that are flooded by tides or by freshwater run-off. Tidal marshes, seagrass beds, and mangroves develop in this region of Florida because of the low wave energy (Reference 267).

Okeechobee Plain Physiographic Subprovince

The Okeechobee Plain physiographic subprovince (Figure 2.5.1-217) is within Okeechobee County and includes part of Lake Okeechobee. The southern part of this plain abuts the Everglades with Lake Okeechobee bisecting the plain. The Okeechobee Plain is divisible from the Everglades by its slightly better drainage and slightly steeper slope and a higher mineral content in its soils. The Okeechobee Plain slopes gradually south to approximately elevation 20 feet (6 meters) at the northern shore of Lake Okeechobee (Reference 265).

Caloosahatchee Valley Physiographic Subprovince

The Caloosahatchee Valley physiographic subprovince is part of the “Caloosahatchee Incline.” This broad gentle incline forms the valley wall of the Caloosahatchee River, which runs through the valley. The Caloosahatchee incline slopes eastward from the eastern toe of the Lake Wales Ridge at 50 to 60 feet (15 to 18 meters) elevation down to 30 to 35 feet (9 to 10.7 meters) at the edge of the Okeechobee Plain. The Caloosahatchee incline is a remnant of a submarine shoal (Reference 265).

The Caloosahatchee River subprovince encompasses the Caloosahatchee River watershed, the lower Charlotte Harbor estuarine system (which includes the San Carlos Bay, Matlacha Pass, and Pine Island Sound), the Estero Bay estuary and watershed, and the Immokalee Rise. This area is approximately 516,000 hectares (1.28 million acres), and includes most of Lee County, the southeastern portion of Charlotte County, western Hendry County, and southern Glades County. The major physiographic provinces of the subregion are the Caloosahatchee Valley, Gulf Coast Lowlands, DeSoto Plain, and the Immokalee Rise (Reference 279).

### Gulf Coastal Lowlands Primary Physiographic Province

Reticulate Coastal Swamps Physiographic Subprovince

Reticulate Coastal Swamps (Figure 2.5.1-217) consist of a broad band of swamps and marshes south of the Everglades and the Big Cypress Swamp. The land is at or near sea level and is often flooded by tides or by freshwater runoff. Salinities range from freshwater to hypersaline, depending on the amount rainfall and runoff and tide levels. The gradual slope of the land continues offshore across the broad west Florida Platform into the Gulf of Mexico. Much of the southern Florida Gulf

Coast receives low wave energy that is favorable to the development of tidal marshes, seagrass beds, and mangrove forests (Reference 267).

Gulf Coastal Swamps and Drowned Karst Physiographic Subprovince

The Gulf Coastal swamps and drowned coastal karst include coastal areas with too little sand to build beaches. The areas start in the south at Florida Bay and continue northward as far as Naples and then from Tarpon Springs to the west side of Apalachee Bay. Coastal lagoon areas include large areas of swamp where they have accumulated sediment and covered with vegetation. Continuous areas of drowned karst along the Gulf Coast of Florida are most prevalent north of Tampa Bay, but small, isolated areas occur south of Tampa Bay to Florida Bay (not shown on Figure 2.5.1-217).

Gulf Coastal Lagoons and Barrier Chains Physiographic Subprovince

The Gulf coastal lagoons and barrier chains are found the entire length of the central Gulf Coast area north of the Caloosahatchee River estuary and continuing northward along the Florida Panhandle. The barrier lagoon system of the Gulf Coast of Florida differs from those on the Atlantic coast in that the sands that build the barriers appear to have been locally derived by erosion of the headlands (Reference 265). The barriers appear to have been developed by erosion of the coastal prominences that separate the large estuaries. White (Reference 265) notes that the process of building spit-like barriers from headlands at the drainage divides appears to have been the habit of this part of the Gulf Coast for some time.

Gulf Coast Terraces Physiographic Subprovince

The Gulf Coastal Terraces physiographic subprovince (Figure 2.5.1-220) can be subdivided by the terraces created by marine regression of the Gulf of Mexico (see discussion of marine terraces related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1). The topography is dominated by broad marine plains and gentle depositional slopes; the regional slope steepens and narrows northward and becomes terraced. The main terraces in the Gulf Coast Terraces province, from oldest to youngest, are the Wicomico, Penholoway, Talbot, and Pamlico (Reference 281). According to Walker and Coleman (Reference 282), these four terraces and three intervening cemented ridges represent the late Pleistocene near shore environments. However, it has been suggested by Alt and Brooks (Reference 283), Lichtler (Reference 284), and Healy (Reference 261) that the elevation of the terraces above 100 to 170 feet (30 to 52 meters) within Florida are not representative of the Pleistocene marine terraces but of older Pliocene and upper Miocene deposits.

Gulf Coast Estuaries Physiographic Subprovince

Estuaries are found along the length of coastal areas of the western Florida Peninsula. The most noteworthy estuaries of southwest Florida are the Charlotte Harbor, Lemon Bay, Coastal Venice, Pine Island Sound, the Caloosahatchee River, Estero Bay, Wiggins Pass/Cocohatchee River, Naples Bay, Rookery Bay, and the Ten Thousand Islands estuaries. Estuaries are semi-enclosed areas, such as bays and lagoons, where fresh water meets and mixes with salty ocean waters. Estuaries are dynamic systems with constantly changing tides and temperatures where salinity varies temporally and spatially. The rivers and streams that drain into the estuaries bring nutrients from the uplands. The nutrients are a food source for a wide range of species, from floating phytoplankton and large algae attached to the estuary floor, to rooted plants including mangrove, marsh grasses, and seagrass. As these endemic plants and animals die, they decompose and provide food for fish, crustaceans and shellfish for which estuaries serve as nurseries (Reference 285).

DeSoto Plain Physiographic Subprovince

The DeSoto Plain physiographic subprovince (Figure 2.5.1-217) in DeSoto County is approximately 45 to 50 miles (72 to 80 kilometers) in length and varies in width from 25 miles (40 kilometers) at the southern edge to 50 miles (80 kilometers) at the northern edge. The northern edge is adjacent to the Polk Uplands. This subprovince has low relief; the northern edge has elevations of 75 to 90 feet (23

to 27 meters); the southern edge has an elevation of 60 feet (18 meters) (Reference 265). The DeSoto Plain and the Osceola Plain represented the southernmost edges of the Florida Peninsula during the Pliocene. The two shorelines were separated by a shallow embayment, the Kissimmee Embayment, which is still recognizable as the Okeechobee Plains lowlands (Reference 286).

The Peace River transverses the DeSoto Plain and is entrenched into the plain 30 to 40 feet (9 to 12 meters). The DeSoto Plain consists of a line of elongated cypress swamps underlain by clay deposits (Reference 278). The relict depositional environment is that of a lagoon that existed during the emergence of the DeSoto Plain. The plain consists of Long Island Marsh, Rainsy Slough, and Valley of Fisheating Creek (Reference 265).

Polk Upland Physiographic Subprovince

The Polk Upland physiographic subprovince (Figure 2.5.1-217) is within Polk County. It is surrounded by the DeSoto Plain on the south, the Gulf Coastal Lowland on the west, the valley of the Hillsborough River and upper Withlacoochee River on the north, and by the Lake Wales Ridge on the east. The Winter Haven and Lakeland Ridges rise from its surface from the northeast. The elevation of the Polk Upland ranges from 100 to 130 feet (30 to 40 meters) (Reference 265).

From the south, a ridge separates the Polk Upland from the DeSoto Plain. The ridge turns 90° N at the southwestern corner of the Polk Upland and terminates approximately halfway on its western side. The edge of the ridge is at an elevation of about 75 to 85 feet (23 to 26 meters), and the crest is at an elevation of about 100 feet (30 meters) (Reference 265). A second ridge is located at the northern end of the western boundary of the Polk Upland and a third irregular ridge is located at the southern boundary. These ridges are most likely erosional marine scarps formed as Gulf of Mexico shorelines during Wicomico sea level (see discussion of marine scarp features related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (Reference 265).

Central Highlands Primary Physiographic Province

The Central Highlands are comprised of coast-parallel scarps and constructed sand ridges (see discussion of shoreline variations related to glacial cycles and sea-level fluctuations in Subsection 2.5.1.1.1.1.1.1) (References 265 and 287). The ridges are all long, narrow, and elongated in the same orientation of the relict beach ridges along the eastern shore of the peninsula. The region consists of xeric (arid) residual sand hills, beach ridges, and dune fields interspersed with numerous sinkhole lakes and basins caused by the dissolution of the underlying limestone bedrock (Reference 265). Millions of years ago, these ridges were formed by rising and falling sea levels. During the periods when the sea level was high and flooded most of peninsular Florida, these ancient islands became refuges for plants and animals. Populations were isolated from the mainland for thousands of years and evolved within these small, sandy habitats. The central inland ridges are older, having remained islands while coastal ridges were flooded, and have a greater concentration of endemic species (Reference 288).

Lake Wales Ridge Physiographic Subprovince

The Lakes Wales Ridge physiographic subprovince (Figure 2.5.1-217) is a unique mosaic of elevated sandy ridges encompassing an area from about the southern Highlands County boundary 160 kilometers (99 miles) north to near Orlando. The Lake Wales Ridge averages about 7.5 kilometers (4.6 miles) wide (Reference 289). Though the name implies a single physiographic area, the Lake Wales Ridge actually consists of three elevated sandy ridges that were once the beach and dune systems of Miocene, Pliocene, and early Pleistocene seas (Reference 290). These relict dunes and the deep, sandy, well-drained soils support a number of plant communities that have adapted to xeric conditions over millions of years. Due to the elevation and geologic age of the soils of Lake Wales Ridge scrubs, it has been estimated that the highest hilltops in this area have supported upland vegetation for about 25 million years. On the Lake Wales Ridge, an estimated 200 ancient scrub islands have been identified (Reference 290). Between ridges and at the base of hills, the soils

become fine and compacted and often retain surface water, forming wetlands and lakes. Rainfall, seepage, and elevated water tables provide the sources of water for these aquatic systems. Combined with the aquatic and wetland communities that now exist between and within the ridges, this subregion consists of a complex mosaic of habitats, some unique to Florida (Reference 291).

Winter Haven Ridge and Lake Henry Ridge Physiographic Subprovinces

The Winter Haven Ridge and Lake Henry Ridge physiographic subprovinces (Figure 2.5.1-217) are an upland karst area that is 130 to 170 feet (40 to 52 meters) in elevation with an abundance of small- to medium-sized lakes. Candler-Tavares-Apopka is the soil association of the well-drained upland area, with longleaf pine and xerophytic oak natural vegetation. Pliocene pebbly quartz sand and the phosphatic Bone Valley Member of the Peace River Formation comprise the underlying geology. The lakes of the area can be characterized as alkaline, moderately hardwater lakes of relatively high mineral content, and are eutrophic (Reference 291).

Lakeland Ridge Physiographic Subprovince

The Lakeland Ridge physiographic subprovince (Figure 2.5.1-217) includes the sand hills of the Lakeland Ridge. These sand hills are covered by phosphatic sand or clayey sand from the Miocene-Pliocene Bone Valley Member of the Peace River Formation. The region generally encompasses the area of most intensive phosphate mining, but phosphate deposits and mining activities are also found south of this region. The dominant characteristic of all lakes in this region is high phosphorus, nitrogen, and chlorophyll-a values. The lakes are alkaline, with some receiving limestone-influenced groundwater (Reference 291).

#### 2.5.1.1.1.1.1.2 Florida Platform

This subsection describes the submerged portion of the west coast of Florida projecting into the Gulf of Mexico (Figure 2.5.1-201). This submerged platform is underlain in its entirety by a large carbonate province of middle Jurassic to Late Cretaceous age, which in turn rests upon what is believed to be highly extended transitional continental-oceanic crust. The west Florida coastal zone is characterized by a series of barrier beaches and back barrier lagoons. The subsiding platform shelf with spotty carbonate production and the quantity of siliciclastic sediments moving along the west Florida coast are evidence that the carbonate platform has been “drowned” by climatic and tectonic changes that occurred since the Miocene.

The siliciclastic-dominated, west-central barrier island coast of peninsular Florida and its adjacent inner continental shelf lie at the center of a huge, ancient carbonate platform that forms the proximal portion of the western Florida shelf/slope system. This platform's east-to-west lateral dimensions extend 310 miles (500 kilometers) from the 650 feet (200 meters) water depth shelf break off the east coast of Florida to the 5000 feet (1500 meters) top of the Florida Escarpment, which drops precipitously nearly another 6000 feet (1800 meters) into the deep Gulf of Mexico (Reference 292) (Figures 2.5.1-201 and 2.5.1-214). For the purpose of this discussion, the Florida Platform is defined as that area between the western coastline of the Florida Peninsula to the edge of the Florida Escarpment (Figure 2.5.1-214). The distinction of where to draw the eastern boundary of the Florida Platform is somewhat arbitrary in that the thick sequence of Upper Jurassic carbonate deposits extends inland to the edge of the Paleozoic rocks forming the core of the Florida Peninsular Arch.

“Carbonate platforms” is a morphological term for a three-dimensional structure and a stratigraphic term for thick sequences of shallow-water carbonates (Reference 293); in this subsection carbonate platforms are referred to in the morphological sense. All carbonate platforms are uniquely sensitive environments that have formed and disappeared many times throughout the Proterozoic and Phanerozoic. The most important controls on carbonate platform development are a shallow substrate, temperature, salinity, and light intensity (Reference 293).

Beginning in the Jurassic, a carbonate platform began to form along the length of the Appalachian-Ouachitan suture/Pangean rift zone. The thinned continental crust foundered to create a long, shallow basin. Seawater incursions were sporadic at first, leaving accumulations of evaporites. Eventually, seawater flooded the shallow basins, building a carbonate megaplatform that deposited 7- to 11-kilometer (4.3- to 6.8-mile) thicknesses of limestones, dolomites, and evaporites (Reference 294). The megaplatform stretched continuously from the Bahamas along the entire Atlantic margin in Late Jurassic to Early Cretaceous time (Reference 424), eventually extending with little or no interruption over 3700 miles (6000 kilometers) from the Gulf of Mexico to the Grand Banks. This carbonate “gigaplatform” probably was one of the largest carbonate platform systems in earth's history (Reference 296).

Because of their sensitivity to depth, temperature, salinity, and light intensity, carbonate platforms are excellent markers for climate change induced by tectonics, igneous activity, reorganization of ocean currents, and other causes. The two types of events that can immediately shut down carbonate production globally or over very wide regions, namely rapid transgression and carbonate crashes, are discussed in the following paragraphs.

Rapid transgression (drowning events) may submerge the carbonate platform below the photic zone and subsequently shut down carbonate production. The rate of transgression must exceed the rate of carbonate accumulation for this to occur. These events may create unconformities that transition from carbonate to siliciclastic deposition. Platform drowning may also occur from episodic rapid subsidence induced by tectonism. If the depth of drowning is moderate, sediment accumulation may allow the seafloor to rise into the photic zone and again begin carbonate production. While reefs may keep up with sea level rise due to high accumulation rates, shallow tidal carbonates, with lower growth rates, may be drowned more easily thus forming a rimmed platform (Reference 297).

A “carbonate crash,” a time of increased regional carbonate dissolution, occurred in the Gulf of Mexico and Caribbean region at the middle to late Miocene transition (Figure 2.5.1-223). Miller et al. (Reference 298) proposes that the early Eocene peak in global warmth and sea level was due not only to slightly higher ocean-crust production but also to a late Paleocene-early Eocene tectonic reorganization caused by emplacement of geographic barriers. The Miocene crash is marked by five dissolution episodes, occurring from 12 to 10 Ma and characterized by significant reductions in the mass accumulation rates of carbonate, noncarbonated, and bulk sediments. Antarctic Intermediate Water filled the Caribbean to abyssal depths during the Quaternary interglacial stages and initiated the North Atlantic Deep Water production. The combination of the initiation of North Atlantic Deep Water, the partial closing of the Isthmus of Panama, and the opening of the Pedro Channel in the northern Nicaraguan Rise led to a change in the global thermohaline circulation. The increase of return flow that passes through the Caribbean may have brought corrosive Antarctic Intermediate Water into the Caribbean, which then caused dissolution of carbonate sediment at the sea floor (References 299, 236, 300, and 302) (Figure 2.5.1-213).

The carbonate crash occurred on both sides of the Isthmus of Panama and in the northwest Caribbean and thus is related to a major reorganization of the ocean circulation during the late Neogene and possibly to the first establishment of a pattern of global thermohaline ocean circulation approaching that of today (Reference 301) (Figure 2.5.1-213). The opening of major seaways would have caused a major reorganization of the deep and intermediate oceanic circulation and maybe the initial production of the North Atlantic Deep Water. The influx into the Caribbean Basins of Antarctic Intermediate Water would have been initiated at this time to replenish the waters sinking in the northern latitudes of the North Atlantic Ocean. This influx of southern source intermediate waters that were corrosive toward carbonate sediments would explain the occurrence of the systematic carbonate dissolution observed in the Caribbean Basin at the middle to late Miocene transition (References 299, 236, and 302).

The geomorphic processes acting on the Florida Platform today include the scouring action of the Loop Current and the influx of terrigenous sediments at the mouths of rivers, generally represented by large embayments, most with backwater lagoons and estuaries (e.g., the Waccasassa, Tampa, Sarasota, and Charlotte Bays). In general, siliciclastics are introduced into the embayments where they may be trapped by persistent carbonate collapse-related sedimentary basins (e.g., Tampa Bay) (Reference 303) or, more likely, the siliciclastics are redistributed by the Loop Current to deep-water basins (see discussion of currents related to changes in oceanic circulation patterns in Subsection 2.5.1.1.1.1.1.2). Very little clastic deposition (<66 feet or 20 meters of accumulated Neogene and Quaternary sediments) occurred in the northern half of the Florida Platform at this time. The southern portion of the Florida Platform is characterized by thick (>330 feet or 100 meters) to very thick (>660 feet or 200 meters) Neogene and Quaternary siliciclastic and carbonate deposition. At intermediate water depths, coarse sand clastics move along the eroded carbonate base of the Florida Platform as coast-parallel sand ridges (Reference 304).

The Florida Peninsula is the emergent portion of the wide, relatively flat geologic feature called the Florida Platform, which forms a rampart between the deep waters of the Gulf of Mexico and the Atlantic Ocean (Figures 2.5.1-201 and 2.5.1-214). The Florida Peninsula is located on the eastern side of the platform. The western edge of the platform lies over 100 miles (161 kilometers) west of Tampa, while on the east side of Florida it lies only 3 or 4 miles (5 to 6.4 kilometers) off the coast from Miami to Palm Beach. Within relatively short distances from the basinal edge of the Florida Platform water depths increase sharply, eventually reaching “abyssal” depths of over 10,000 feet (3050 meters). This relatively sharp break in seabed topography is the Florida Escarpment. The Florida Escarpment is analogous to the Campeche Escarpment and the Bahamas Escarpment, representing the margin of a once actively prograding carbonate platform at a passive margin with an ocean basin (e.g., the two spreading centers in the Gulf of Mexico on the northwestern and northeastern edges of the Yucatan Platform, discussed in Subsection 2.5.1.1.2.1.1, and the mid-Atlantic Ridge east of the Bahama Platform). As discussed in Subsection 2.5.1.1.1.1.2, none of the escarpments formed as the result of Quaternary faulting, and all are retreating on their seaward edges by ocean current erosion and mass wasting into the nearby abyssal basins (Reference 305). Thus, the former spreading center (inactive) faults are likely many kilometers basinward from the current edge of the escarpments.

Diving expeditions with the deep submersible *Alvin* along the escarpment west of Tampa found the escarpment there consisted of a limestone cliff that rose over 6000 feet (1830 meters) above the 10,700-foot (3260-meter) deep Gulf of Mexico floor. Based on seismic evidence from petroleum exploration, carbonate and evaporitic rocks may underlie the southern tip of Florida at depths greater than 20,000 feet (6100 meters) below ground surface (Reference 287).

The Florida Platform is part of the larger Florida-Bahama Platform, which is approximately 600 miles (900 kilometers) long, 620 miles (1000 kilometers) wide, and over 7.5 miles (12 kilometers) thick (Reference 307). The Florida-Bahama Platform represents an enormous shallow-water, carbonate sedimentary province (Figures 2.5.1-201, 2.5.1-202, and 2.5.1-224). The marginal-reef facies along the Florida Platform have been eroded back by as much as 4 to 16 miles (6 to 26 kilometers) since the close of the Oligocene (References 305 and 308).

The Florida Platform is a broad smooth shelf with widths of nearly 124 miles (200 kilometers) and little morphologic variability. Vertical relief is shown in the coral algae ridges (Reference 282). Facies distribution across the Florida Platform is shown in Figure 2.5.1-225. Before the Florida Platform was drowned in the Late Jurassic, the facies distribution of active carbonate accretion was very similar to that of the current Bahama and Yucatan Platforms. Today, carbonate production on the Florida Platform is much reduced (see discussion of carbonate platforms in Subsection 2.5.1.1.1.1.1.2). The “drowning” of the Florida Platform did not occur as a single event. Carbonate production kept pace with subsidence across the platform through the Early Cretaceous (until about 100 Ma). Siliciclastic sedimentation on the area near the Florida Panhandle began to replace carbonate production

sometime in the mid-Cretaceous. Carbonate production in the southern portion of the Florida Platform kept pace with subsidence until early Late Cretaceous (Cenomanian) time, when the Florida Platform and the other carbonate banks of the Gulf of Mexico were drowned and covered by neritic marl containing some siliciclastic debris. Only the south-southwestern most parts of the west Florida and Bahama Platforms continued to be areas of active carbonate sedimentation (Reference 309). Elsewhere, the juxtaposition of underlying mollusk-rich, shallow-water limestones with the overlying pelagic facies indicates that the drowning was rapid and terminal (Reference 309). The drowning event is recognized in the stratigraphy across the Gulf of Mexico and West Florida area and is known as the mid-Cretaceous sequence boundary or unconformity (see discussion of the oceanic and atmospheric reorganization events in Subsection 2.5.1.1). The general distribution of facies prior to drowning is shown in Figure 2.5.1-225. Like other drowned carbonate platforms, today's Florida Platform has subsided to bathy depths, forming a deep-water plateau as the increasingly deeper-water sedimentary facies were unable to keep pace with subsidence (Reference 305). By the end of the Late Cretaceous, water depths exceeded 900 meters (2950 feet) over the marginal parts of the former shallow-water carbonate bank. Currently, the Gulf Coast continental shelf and slope of the Florida Platform constitutes a ramp system that connects the Gulf of Mexico abyssal plain with the Florida Peninsula. Since the end of Late Cretaceous (Maastrichtian) time (66 Ma), the Florida Platform has undergone phases of aggradation and progradation with sea-level fluctuations. The sea-level fluctuations, together with the action of the Loop Current, are reflected in the character and continuity of Cenozoic sediments that overlie the platform.

#### 2.5.1.1.1.1.1.3 Atlantic Offshore Continental Shelf and Slope

The Atlantic Continental Shelf is generally a continuation of the coastal plain as it dips beneath sea level at a low angle (0.1 to 0.5°) to a depth of about 150 meters (500 feet). The shelf is characterized by a shallow gradient to the southeast (Reference 313) and many shallow water features that are relicts of lower sea levels. Below 150 meters of water depth, the shelf abruptly transitions to a steeper continental slope that dips seaward at an angle of between 3 and 6° to a depth of about 1500 meters (5000 feet). At the base of the continental slope is a fan of turbidite sediments that make up the continental rise, with a slope of 0.5 to 1.0° (Reference 310), and extends to depths of 1500 to 5000 meters (5000 to 16,500 feet), which in turn transitions to the abyssal plain between water depths of 3000 and 10,000 meters (10,000 to 33,000 feet) (Reference 311).

The Atlantic coast of Florida, however, does not fit easily into the morphological characterization described above. Here, the continental shelf (defined based on water depth) is represented by the narrow Florida-Hatteras Shelf extending to 80 meters (260 feet) water depth, after which the seafloor drops precipitously to several hundred meters depth. At a depth of 400 to 1250 meters (1300 to 4100 feet), a broad, flat carbonate plateau, the Blake Plateau, forms part of the Atlantic Continental Slope. The plateau is an 8- to 12-kilometer thick sequence of flat-lying Middle Jurassic and lower Cretaceous carbonate rocks that are continuous with the limestone platform rocks underlying the Florida, Bahama, and Yucatan Platforms. The Blake Plateau and Escarpment separate the wedge of Upper Cretaceous and Cenozoic siliciclastic sediments from the Atlantic Abyssal Plain (Reference 487). The Blake Plateau represents a carbonate platform that “drowned,” or ceased carbonate production in the early Cretaceous (see discussion in Subsection 2.5.1.1.1.1.1.2) (Reference 312). In contrast to the Blake Plateau, carbonate production kept pace with subsidence across most of the Bahama Platform, and the platform's northern shallow water areas now form part of the continental shelf of the North America Plate. Therefore, the Blake Plateau and the canyons and channels that cut the Bahama Platform are considered part of the Atlantic Continental Slope, whereas the shallower portions of the Bahama Platform are considered to be part of the Atlantic Continental Shelf (Figure 2.5.1-214).

The continental shelf/slope along Florida's Atlantic coast has four distinct physiographic areas illustrated in Figure 2.5.1-214:

- Florida-Hatteras Shelf, from the shoreline to 80 meters (260 feet) water depth.
- Florida-Hatteras Slope, from 80 to 400 meters (260 to 1300 feet) water depth.
- Blake Plateau, from 400 to 1200 meters (1300 to 4000 feet) water depth.
- Blake Escarpment, the nearly vertical escarpment beginning at 1200 meters (4000 feet) water depth, formed by erosional retreat of the Blake carbonate platform (Reference 305). The Blake Escarpment connects the Blake Plateau to the Atlantic's abyssal Blake-Bahamas Basin, found at 4000 to 5400 meters (13,000 to 17,700 feet) water depth (Figure 2.5.1-214).

Sedimentary processes and sediment instability on the Florida-Hatteras Shelf and Slope appear to be related to evidence of potential megasedimentary events that have occurred from shallow water sources into deeper basins and troughs in the Gulf of Mexico, Straits of Florida, Florida Atlantic continental margin, Bahama Platform, and Caribbean region. There are indications that the “normal” processes seen today have, in the past, occurred on a much larger scale than has been often witnessed in human history. Based on evidence gathered by the DSDP and ODP, turbidite megasedimentation events have occurred with relative frequency in the Caribbean and Atlantic Ocean Basins (Reference 315).

Pilkey (Reference 315) roughly estimates the probability of recurrence of such megaturbidite events at once every 50,000 years in the largest ocean basins and more frequently in smaller basins. In the basin plains studied by Pilkey (Reference 315), hemipelagic sediments usually make up less than 20 percent of the sediment volume. Convulsive megasedimentation events flatten out the topographic irregularities formed by the deposition of small gravity flows between the large events. Large-scale events and the sediments they deposit maintain the flat plain floor. In some instances, a single giant flow may arrive on the basin simultaneously from geographically widespread basin entry points, indicating that the initiating mechanism, probably an earthquake, was regional in scope (Reference 315). The water displacement associated with these megaturbidites likely creates tsunamis that may have affected any nearby coastal areas. The largest event measured by Pilkey in modern basin plains is the Black Shell Turbidite of the Hatteras Abyssal Plain. The Black Shell Turbidite is at least 100 kilometers$^3$ (24 miles$^3$) in volume and occurred about 16,900 years ago, based on radiocarbon dating of the youngest shell material incorporated in the deposit (Reference 316). Major events such as the Black Shell Turbidite should not be assumed to be restricted to times of lowered sea level. The 1929 Grand Banks slump, which occurred in the current interglacial period, may have involved as much as 400 kilometers$^3$ (96 miles$^3$) of material (Reference 317). This slump resulted in a turbidity current that traveled 500 kilometers (300 miles) to the Sohm Abyssal Plain, but the full areal extent of the resulting turbidite is still unknown (Reference 315). In this case, the slump and resulting tsunami were induced by the $M_w$ (moment magnitude) 7.2 earthquake that occurred on November 18, 1929 (Reference 318). See further discussion of the Black Shell Turbidite in Subsection 2.5.1.1.5.

Subsections 2.5.1.1.2.1.1 and 2.5.1.1.2.2.3 provide discussions of megasedimentation events in the Gulf of Mexico and Caribbean regions. Subsections 2.5.1.1.5 and 2.4.6 provide discussions of tsunami hazards from seismic or non-seismogenic submarine landslides.

There is no evidence to indicate that a feature similar to the Florida-Hatteras Slope existed prior to the Eocene. Paleocene strata are deeply eroded in the subsurface, and this erosion may mark the initial appearance of the Gulf Stream in this region (see discussion of changes in oceanic circulation patterns in Subsection 2.5.1.1.1.1.1.2). During the Eocene and Oligocene, another wedge of shelf sediments prograded across the Blake Plateau but were interrupted by erosion at the end of the Oligocene. A prograding wedge of Miocene to Holocene sediments covers the erosional unconformity at the base of the Miocene. The accumulation of post-Paleocene sediment at the foot of

the Florida-Hatteras Slope and seaward onto the Blake Plateau has been very slow, due in part to the reduction in sediment supply as well as to erosion by the modern Gulf Stream (Reference 487).

The Blake Plateau underlies the continental shelf off the east coast of Florida, northeast of the northwest-striking Jacksonville Fracture Zone interpreted from magnetic anomaly maps (Reference 212). The plateau is dominantly underlain by Jurassic to Cretaceous carbonates and thought to be built upon transitional crust of African affinity, like the Florida Platform (Reference 307). The east edge of the plateau is near the Blake Spur magnetic anomaly, which may mark the boundary between transition and oceanic crust (Reference 307). The Jurassic and younger strata of the Blake Plateau are unfaulted, though an apparent left-lateral offset in the M-25 magnetic anomaly may reflect dextral shearing along the Jacksonville Fracture Zone (called the Bahama Platform-Blake Plateau boundary or Great Abaco Fracture Zone) (Reference 307) (Figures 2.5.1-206 and 2.5.1-229). Faulting here is pre-Miocene age.

At the northern terminus of the Blake Plateau lies the Blake Ridge contourite drift. The Blake Ridge contourite deposits (Figure 2.5.1-226), located off the U.S. coast, east of Charleston, South Carolina, and Savannah, Georgia, represent one of the largest methane hydrate provinces on Earth. Hornbach et al. (Reference 319) analyze high-resolution three-dimensional seismic data to map seismic indicators of concentrated hydrate and fluid flow. Their analysis reveals that the Blake Ridge gas hydrate system is significantly more dynamic than previous studies suggest, and hypothesize that fluctuating sedimentation and erosion patterns cause hydrate phase-boundary instability that triggers fluid flow. The surface morphology of Blake Ridge is controlled by the western boundary undercurrent, which erodes sediment from the eastern flank of the ridge and redeposits it on the western half. The Blake Ridge contourite deposits apparently are affected by complex ocean currents created by the overriding Gulf Stream mixing with the Western Boundary Undercurrent at intermediate bottom waters. Therefore, what appears to be an anomalously linear western margin of the Blake Ridge is caused by circulating currents, erosion, and deposition and not by faulting (Subsections 2.5.1.1.1.3.2.2 and 2.5.1.1.1.3.2.3).

Twichell et al. (Reference 320) report numerous underwater landslide scars on the sea floor off the U.S. Atlantic coast (Figure 2.4.6-202). They find that landslide scars cover 13 percent of the sea floor on the Atlantic Continental Slope and Rise between Cape Hatteras and the Blake Plateau. Landslides can be divided into two categories based on their source areas: those with sources located in submarine canyons and those with sources located on the open Atlantic Continental Slope and Rise. The deposits from both landslide categories are generally thin (mostly 20 to 40 meters [65 to 131 feet] thick) and primarily comprise Quaternary material, but the volumes of the open-slope sourced landslide deposits can be larger (1 to 392 kilometers$^3$) than the canyon sourced ones (1 to 10 kilometers$^3$) (Reference 320).

Canyons are absent south of Cape Hatteras. Two large submarine landslides are identified south of Cape Hatteras, the Cape Fear slide and an older Cape Lookout slide that is crossed by the Cape Fear slide (References 320 and 321) (Figure 2.4.6-202). Both slides have been correlated with salt diapirism and methane hydrate layers. The Cape Fear landslide is about 100 meters (330 feet) thick, 25-kilometer (15.5-mile) wide, involving an area of about 5000 kilometers$^2$ (Reference 322), with a volume that is likely in excess of 200 kilometers$^3$ (Reference 321). Using $C^{14}$ of sediment retrieved from the uppermost Cape Fear Slide plain, the slide is dated at early- to mid-Holocene (between 3800 to 10,000 years before present) (Reference 322). Based on multibeam bathymetry data and seismic Chirp data, Hornbach et al. (Reference 323) interpret structural controls between the salt diapirs near the Cape Fear landslide and the diapirs associated with the Blake Ridge gas hydrate field (Figure 2.5.1-226). A 40-kilometer (25-mile) long fault, imaged using Chirp data, stretches between the diapirs and may serve as a point of failure on the headwall of submarine slides.

Submarine slope failures are caused by lower sediment strength and an increase in shear stresses (Reference 324). The dissociation of natural gas hydrates found in the seafloor can cause a

reduction in sediment shear strength and an increase in pore water pressure along the potential failure plane (Reference 324). Both of these mechanisms can contribute to seafloor slope failures. Hornbach et al. (Reference 323) report that at least 10 major (>100 kilometers$^3$) (>24 miles$^3$) mass wasting events have been documented in the North Atlantic Ocean during the past 40 k.y. They indicate that potential triggers of submarine slope failures may act alone or in concert include earthquakes, mechanical failure of overpressured sediments, storm waves, groundwater seepage, failure of oversteepened slopes, gas hydrate dissociation, and sea level change.

Gas hydrates are most susceptible to decomposition in response to lowered sea level if they occur in 200- to 600-meter (650- to 1970-feet) water depths. The Blake Ridge contourite is in water depths of over 2000 meters (6500 feet) (Figure 2.5.1-226). However, Paull et al. (Reference 325) and Tappin (Reference 326) indicate that an increase in submarine slumping occurred during the last glacial maximum that is consistent with the prediction that a causative relation exists between gas hydrate decomposition induced by lowered sea level and slumping frequency. A further consideration is that the Blake Ridge contourite deposits are located offshore of the Charleston, South Carolina seismic zone and large mass wasting events associated with gas hydrate decomposition could represent an additional consequence of earthquake activity there. Twichell et al. (Reference 320) conclude that large landslides may be related to the upward migration of salt along normal faults in the Carolina Trough (Figure 2.4.6-202). It appears that oversteepening of the sea floor due to salt movement, aggravated by any earthquakes in the region, however small, could lead to repeated slope failures and landslides. See Subsections 2.5.1.1.5 and 2.4.6 for further discussion of potential landslide and tsunami hazards to the Units 6 & 7 site.

#### 2.5.1.1.1.1.2 Bahama Platform

From the Late Jurassic to Early Cretaceous, the Bahama Platform was contiguous with the Florida Platform, the Yucatan Platform, and the carbonate platform of the Gulf of Mexico (Reference 307). Together, the complex was part of one of the most extensive carbonate systems in Earth's geologic history. The carbonate platform stretched nearly 7000 kilometers (4400 miles) from the north-central Gulf of Mexico along the eastern North American continental margin to Canada (Reference 327). Carbonate production on the Bahamas-Grand Banks megaplatform began to shut down in the mid- to Late Cretaceous (Reference 327) (Subsection 2.5.1.1.1.1.1.2).

Located immediately east and south of the Florida Platform, the Bahama Platform is a broad, shallow marine platform with shallow banks (small portions of which are emergent as the Bahamas Islands) and intervening deep-water channels with depths of up to 13,100 feet (4000 meters) (Reference 307). The Bahama Bank, consisting of the Great and Little Bahama Banks, is separated from the Florida Peninsula by the Straits of Florida. The Bahama Bank mostly comprises shallow continental shelf and lagoon sediments. The windward side comprises a small percentage of active reefs. The reef pinnacles are found on the outer margin of the shelf. The sediment surface consists of grass flats of carbonate sands with relict sand bodies and oolite shoals (Reference 282).

The geomorphology of the Bahama Platform is controlled by the progradation of reefs, the brecciated fore-reef escarpments, and the accumulation of back-reef carbonate sediments reworked by currents, waves, and winds. A minor component of platform sediments is terrigenous sands transported from Africa by the prevailing trade winds (Reference 328).

Anselmetti et al. (Reference 228) analyze high-resolution seismic data collected across the Great Bahama Bank margin and the adjacent Straits of Florida to determine that the deposition of strata in this transect over the past 23 m.y. was and is controlled by two sedimentation mechanisms: (a) west-dipping layers of the platform margin, which are a product of sea-level-controlled, platform-derived downslope sedimentation; and (b) east- or north-dipping drift deposits in the basinal areas, which were deposited by ocean currents. These two sediment systems are active simultaneously and interfinger at the toe-of-slope. The Neogene slope sediments consist of

peri-platform oozes intercalated with turbidites, whereas the basinal drift deposits consist of more homogeneous, fine-grained carbonates that were deposited without major hiatuses by the Florida Current starting at approximately 12.4 Ma. Glacial sea-level fluctuations, which controlled the carbonate production on Great Bahama Bank by repeated exposure of the platform top, controlled lithologic alternations and hiatuses in sedimentation across the transect.

Droxler and Schlager (Reference 329) analyze the sedimentation rates during glacial and interglacial periods. Based on radiometric and faunal assemblage age dating for stratigraphic sequences, Droxler and Schlager (Reference 329) find the mean of bulk sedimentation rates at the Great Bahama Bank is four to six times higher in interglacial periods than during glacial periods; average accumulation rates of recognizable turbidites are higher during interglacial than glacial periods by a factor of 21 to 45, and interglacial turbidite frequency is higher by a factor of 6 to 14 during glacial periods. Sediment composition indicates that increased interglacial sedimentation rates are due to higher accumulation of platform-derived material. Additional data from other Bahamian basins as well as published material from the Caribbean strongly suggest that highstand shedding is a general trend in pure carbonate depositional systems. Carbonate platforms without a siliciclastic component export more material during highstands of sea level when the platform tops are flooded and produce sediment. The response of carbonate platforms to Quaternary sea-level cycles is directly opposite of that of siliciclastic ocean margins, where sediment is stored on the inner shelf during highstands and passed on to continental rises and abyssal plains during lowstands of sea level (Reference 329).

Carbonate production on the Bahama Platform is controlled largely by sea-level conditions. Sediment production and off-bank transport is highest during sea-level highstands when the platform is flooded (References 330, 331, and 332). During these times, more sediment is produced than can be accumulated on the platform top, and a large amount of sediment is transported off-bank onto the slopes (References 330, 333, 863, and 334). Light dependency forces the carbonate-secreting organisms to maintain the depositional systems close to sea level, resulting in a nearly flat sediment surface across the entire platform. As a result, falling sea level exposes the platform and restricts sediment production to the fringes of the platform. With renewed flooding of the platform, sediment production and off bank transport resumes, depositing a highstand wedge on the leeward slope of the western Great Bahama Bank (Reference 334).

#### 2.5.1.1.1.1.3 Cuba

Unlike the Bahama Platform, the physiography of Cuba includes mountainous terrain. The archipelago of Cuba is formed of about 1600 islands, the largest of which, Cuba, lies approximately 150 miles (240 kilometers) south of the Units 6 & 7 site. The Windward Passage separates Cuba from Jamaica, the Bahamas, and Haiti. Cuba is about 1200 kilometers (745 miles) east to west and between 40 and 290 kilometers (25 and 180 miles) north to south. The island covers 107,500 kilometers$^2$ (44,000 miles$^2$) and is mountainous for 20 percent of its land surface. Three different mountain ranges have been identified: the highest (up to 2000 meters or 6600 feet) in the Sierra Maestra of southeastern Cuba; the central low ranges of Escambray Mountains (Sierras Cienfuegos and Sancti Spiritus), and in the west, the Pinar del Rio range (Sierra de los Organos) (Figure 2.5.1-227). The other 80 percent of Cuba's land surface consists of more gently rolling hills and extensive lowlands, with deep red sandy clays and fertile alluvial soils in the flood plains (Reference 335). About two-thirds of the island consists of limestone and karstic features are well developed, particularly in the eastern section (Reference 336).

Although rivers are plentiful, the island's narrow, elongated form means that much of the fresh water runs off quickly seawards, with little retention other than where captured by human activities. The longest is the Cauto River at 370 kilometers (230 miles), followed by the Sagua la Grande River at 163 kilometers (101 miles) and Zaza River at 155 kilometers (96 miles) (Figure 2.5.1-227). The east coast is subject to hurricanes from August to October, and droughts are also common (Reference 336).

#### 2.5.1.1.1.2 Regional Stratigraphy within the Site Region

The stratigraphy described in this subsection has been developed from the analysis of surface and subsurface geologic and geophysical investigations performed at the site and reported in peer-reviewed publications. The stratigraphy of southern Florida is characterized by a thick sequence of Jurassic to Holocene sediments that lie unconformably on Jurassic basement volcanic rocks. Although most of the units in the sedimentary sequence are carbonates, deposition of Appalachian derived siliciclastic sediments occurred during the Miocene and Pliocene. The oldest stratum exposed at the surface in the southern Florida region is the Miocene-Pliocene Peace River Formation that crops out in Hardee and DeSoto counties (Reference 283).

An important aspect of describing stratigraphy is identifying the depth to the basement. Basement may be defined as structural, stratigraphic, seismic, or petrologic. Recent literature has generally accepted the pre-Cretaceous surface as an appropriate upper limit of the basement stratigraphy of the Gulf of Mexico and North America-Caribbean Plate boundary (References 212 and 337). In this context, “basement” refers to stratigraphic basement below a regionally recognizable and tectonically significant unconformity. The basement unconformity separates pre- to syn-rift rocks formed during the breakup of Pangea from overlying rocks. The information presented in this subsection refers to the basement surface as pre-Middle Jurassic. This age designation more accurately constrains the Mesozoic post-rift unconformity that can be correlated across most of in the subsurface of Florida and across the Gulf of Mexico, Yucatan and Bahama Platforms, and Caribbean region.

##### 2.5.1.1.1.2.1 Stratigraphy of the Florida Peninsula and Platform

This subsection describes the stratigraphy of the Florida Peninsula and the remainder of the submerged Florida Platform.

###### 2.5.1.1.1.2.1.1 Stratigraphy of the Florida Peninsula

Two basement lithologic regions are recognized in Florida: a central and northern suite of Proterozoic rocks collectively known as the Suwannee terrane and a southern early Middle Jurassic volcanic province (References 338 and 339) (Figure 2.5.1-205).

The depth to metamorphic or crystalline basement beneath the Florida Peninsula and Platform is variable, from depths of 5 to 7 miles (8 to 11 kilometers) beneath the North Atlantic Ocean shorelines (Reference 282). Basement rocks are overlain by up to 15,000 feet (4570 meters) of relatively flat-lying Mesozoic evaporate and carbonate units (Figure 2.5.1-228), which are in turn overlain by up to 6000 feet (1830 meters) of Cenozoic carbonate and siliciclastic sediments (References 339, 338, and 340). The basement is found at depths of about 15.5 miles (25 kilometers) beneath the southern Florida shoreline, within the South Florida Basin. There, basement rocks are overlain by up to 25,000 feet (7600 meters) of relatively flat-lying Mesozoic evaporate and carbonate units, which are in turn overlain by up to 5250 feet (1600 meters) of Cenozoic carbonate and siliciclastic sediments (Reference 341).

**Proterozoic Stratigraphy of the Florida Peninsula**

The Suwannee terrane (Figures 2.5.1-205 and 2.5.1-228) comprises the following:

- Low-grade, felsic metavolcanics of the Osceola volcanic complex
- The undeformed Osceola Granite
- A suite of high-grade metamorphic rocks, such as gneiss and amphibolite, belonging to the St. Lucie complex

- A succession of generally undeformed Paleozoic sedimentary rocks

The Osceola Volcanic Complex

The Osceola volcanic complex (Figures 2.5.1-205 and 2.5.1-228) is a group of calc-alkaline, felsic, low-grade metaigneous rocks (Reference 338). Lithologic variations within the complex include felsic vitric tuff, felsic ash-flow tuff, and tuffaceous arkose with subordinate andesite and basalt. The rocks are generally undeformed but almost always display low-grade metamorphic assemblages (Reference 342). There is no consensus in the published literature on the age of the rocks belonging to the Osceola volcanic complex.

Dallmeyer (Reference 338) correlates rocks of this volcanic complex with a West African calc-alkaline metaigneous sequence dated at 650 Ma. Lithologic comparisons were used to propose that the north Florida volcanic suite is directly correlative with Late Proterozoic calc-alkaline volcanic rocks of the Niokolo-Koba Group in Senegal, West Africa, which, in turn, were proposed to be coeval with granites dated by $^{40}Ar/^{39}Ar$ and $^{87}Rb/^{86}Sr$ at 650 to 700 Ma (late Proterozoic) (Reference 343).

Chowns and Williams (Reference 344) suggest a Late Proterozoic to early Paleozoic age for the rock based on core recovered from a well drilled in central Florida where the felsic igneous complex appears to be unconformably overlain by Lower Ordovician sandstone. Whole rock K-Ar ages for this igneous complex range from about 165 to 480 Ma. Unpublished whole rock $^{40}Ar/^{39}Ar$ data reported by Horton et al. (Reference 342) on a suite of seven volcanic samples indicate that all of the samples have a noticeably discordant age spectra. A slate sample from 11,600 feet (3500 meters) displays an internally discordant $^{40}Ar/^{39}Ar$ age spectrum, defining a total-gas age of about 341 Ma. Similarly, a felsic metavolcanic rock recovered from a depth of 12,350 feet (3800 meters) displays an internally discordant age spectrum; however, intermediate and high temperature increments correspond to a plateau age of about 375 Ma (Reference 342).

Heatherington and Mueller (Reference 343) report on the age of a suite of volcanic rocks; basaltic andesites to rhyolites, from northeastern and north central Florida. These rocks from Putnam and Flagler counties yielded $^{40}Ar/^{39}Ar$ measurements corresponding to approximately 410 to 420 Ma, while whole rock $^{87}Rb/^{86}Sr$ data suggest a composite isochron corresponding to an age of about 480 ± 60 Ma. Heatherington and Mueller (Reference 343) report that, while these are the only dates available for these rocks, they should be viewed as “lower limits” only. This limitation appears to be due to: the complexity of the $^{40}Ar/^{39}Ar$ data, the possibility that the whole-rock $^{87}Rb/^{86}Sr$ data may represent a mixing array, and that both the $^{40}Ar/^{39}Ar$ and $^{87}Rb/^{86}Sr$ systems are easily reset in these types of rocks by low-grade thermal and hydrothermal events.

The Osceola Granite

The Osceola Granite (Figures 2.5.1-205 and 2.5.1-228) comprises undeformed diorite to batholithic granodiorite (Reference 345). This rock has a granitic texture with coarse pink sodic plagioclase feldspar, abundant quartz, albite-oligoclase, and some potash feldspar, ilmenite, and apatite (Reference 340). Dallmeyer et al. (Reference 337) describe the pluton as heterogeneous and predominantly comprising biotite granodiorite, leucocratic biotite quartz monzonite, and biotite granite. According to Horton et al. (Reference 342), most of the samples examined by Dallmeyer et al. (Reference 337) were predominantly composed of oligoclase, quartz, perthitic alkali feldspar, and biotite. Depth to the top of this granite is approximately 8000 feet (2400 meters) in Osceola County (Reference 339).

Several dates are reported for the Osceola Granite. Biotite samples collected from two wells in Osceola and Orange counties yielded $^{40}Ar/^{39}Ar$ dates of 527 and 535 Ma (Reference 337). Dallmeyer et al. (Reference 337) suggest that these ages closely date emplacement of the pluton in view of its high-level petrographic character and apparently rapid postmagmatic cooling. $^{87}Rb/^{86}Sr$ analytical results from several density fractions of feldspar collected from a well in Osceola County

reflect a crystallization age for the granite of about 530 Ma (Reference 346). Mueller et al. (Reference 347) report different ages obtained from whole-rock samples taken from two wells in Osceola County, suggesting that the Late Proterozoic to Early Cambrian Osceola Granite was derived from two or more older sources with different ages, at least one of which was Archean. Reconnaissance single-grain ion-probe analyses of zircons from the Osceola Granite corroborate both the $^{40}Ar/^{39}Ar$ cooling age of approximately 530 Ma determined by Dallmeyer et al. (Reference 337) and the Archean component suggested by Mueller et al. (Reference 347). According to Heatherington and Mueller (Reference 343), several grains produced $^{206}Pb/^{238}U$ dates of about 550 to 600 Ma consistent with the $^{40}Ar/^{39}Ar$ date of 530 Ma as a cooling age.

St. Lucie Metamorphic Complex

The St. Lucie metamorphic complex (Figures 2.5.1-205 and 2.5.1-228) is immediately south of, and associated with, the Osceola Granite. It is a suite of high-grade metamorphic rocks and variably deformed igneous rocks. Predominant rock types include amphibolite, biotite-muscovite schist, chlorite schist and gneiss, and quartz diorite. The complex has a distinctive aeromagnetic signature with marked northwest-trending magnetic lineations that may reflect structural strike (Reference 342). Depth to the amphibolite in St. Lucie County is approximately 12,500 feet (3800 meters) (Reference 338).

Core recovered from wells drilled in the St. Lucie metamorphic complex in St. Lucie and Marion counties are predominantly amphibolites with schist and layers of quartz diorite (Reference 346). Radiometric dates include K/Ar dates of 503 and 470 Ma for hornblende from amphibolite recovered from a well drilled in St. Lucie County (Reference 346) and a reportedly more reliable $^{40}Ar/^{39}Ar$ date of 513 ± 9 Ma for a hornblende concentrate from amphibolite recovered from another well in St. Lucie County (Reference 338). On the basis of this later date, Dallmeyer (Reference 338) suggests that the St. Lucie amphibolite is correlative with amphibolites from the northern Rokelide orogen in Sierra Leone, West Africa, which have similar cooling ages.

### Paleozoic Stratigraphy of the Florida Peninsula

The Paleozoic sedimentary suite is composed of a succession of undeformed, Lower Ordovician quartzitic sandstones and Middle Devonian black shales and siltstones overlying the Peninsular Arch (Reference 339) (Figures 2.5.1-205 and 2.5.1-228). Muscovite within the sandstone records an $^{40}Ar/^{39}Ar$ age of 504 Ma (Reference 338).

The base of the subsection, the Lower Ordovician littoral quartz sandstone (Reference 342), consists of white to reddish quartz sandstone with *Skolithos* burrows and interbedded micaceous shales. This portion of the subsection is the most widely distributed and possibly the thickest of all the Paleozoic sedimentary units. The sandstone is Early Ordovician in age and is defined on the basis of Arenig age graptolites and inarticulate brachiopods (References 207 and 348).

Overlying the sandstone is Ordovician to Middle Devonian shale with locally significant horizons of siltstone and sandstone (Reference 342). The dark-gray to black shales are interbedded with gray fine-grained micaceous sandstone and locally medium- to coarse-grained quartz sandstone. Based on paleontologic data, the shales are divided into three sections. The lowest section consists of Middle to Upper Ordovician fauna including trilobites, inarticulate brachiopods, conulariids, conodonts, and chitinozoans. The middle section consists of Late Silurian to Early Devonian shale with bivalves, gastropods, orthocone cephalopods, tentaculitids, brachiopods, crinoids, eurypterids, ostracods, and chitinozoans. The upper section consists of shales and sandstones containing Middle Devonian land plants, bivalves, ostracods, and marine microfossils (Reference 207). According to Thomas et al. (Reference 207), the contacts between several of the units are undefined and the section might be continuous. However, there is a possible discontinuity based on the absence of Early Silurian faunas. The thickness of the sedimentary units is uncertain, but based on gravity modeling and seismic profiles the thickness of the entire section ranges from 8202 feet (2500 meters)

in parts of north-central Florida, to 32,808 feet (10,000 meters) in the Panhandle (References 207 and 342).

A genetic relationship between the southern Florida basement (Figure 2.5.1-204) and West African rock sequences has been suggested by many investigators (References 339, 338, 349, and 350) based on the following:

- A correlation between lithology and the radiometric age of calc-alkaline felsic igneous complex rocks in central Florida and West Africa (Reference 338). There is a correlation between the Osceola Volcanic Complex and West Africa calc-alkaline, metamorphosed igneous sequence (i.e., Niokola-Koba Group) along western portions of the Mauritanide, Bassaride, and northernmost Rokelide orogens (Reference 337). The isotopic ages of the sequence are about 650 and 700 Ma (Reference 342).

- A correlation between radiometric ages and petrography of the Osceola Granite with post-tectonic granite plutons in Guinea West Africa (Reference 338). The inferred correlation proposed by Dallmeyer et al. (Reference 337) for the Osceola Granite is with the Coya Granite in West Africa (northern Rokelide orogen in Guinea). The Osceola and Coya granites both have crystallization ages of about 530 Ma and display similar petrographic characteristics (Reference 342).

- A correlation between the stratigraphic, geochronologic, and geochemical data of the southern Florida tholeiitic volcanic sequence and Liberian tholeiites indicates that both may have had the same parental magma (Reference 339). Hornblende $^{40}Ar/^{39}Ar$ cooling ages of 510 to 515 Ma from the St. Lucie Metamorphic complex are similar to the Rokelide orogen where hornblende K-Ar and $^{40}Ar/^{39}Ar$ dates are approximately between 485 and 530 Ma (Reference 342).

- A correlation of the Paleozoic sedimentary sequence overlying the Peninsular Arch with the lithology, radiometric ages, paleontology, and paleomagnetic data from rocks in Senegal and Guinea (Reference 338). The correlation between the subsurface Paleozoic sedimentary sequences in the North Florida Basin with sequences of similar age in the Bové Basin of Senegal and Guinea is suggested by similarities in fauna and stratigraphic succession (Reference 344). In addition, an $^{40}Ar/^{39}Ar$ 505 Ma date of detrital muscovite from subsurface Ordovician sandstone in Marion County, Florida, suggests a metamorphic source similar in age to the rocks of the Bassaride and Rokelide orogens that yield $^{40}Ar/^{39}Ar$ muscovite ages of about 500 to 510 Ma (Reference 206). Opdyke et al. (Reference 351) report 1650 to 1800 Ma U-Pb ages for detrital zircons from Ordovician-Silurian age sandstone in Alachua County, Florida, which suggests a source similar in age to the basement of the West African craton.

The lithologic and geochronologic characteristics described above suggest that these Florida basement provinces originated in West Africa and represent a fragment of Gondwana that accreted to Laurentia during the late Paleozoic formation of Pangea (References 339, 338, and 349). The Floridian piece of Africa remained attached to North America when Pangea broke apart during the opening of the Atlantic Ocean in the Jurassic (Reference 338). This fragment of the African Plate provided the base for the development of a carbonate platform that included the Florida Platform (Reference 349) (Subsection 2.5.1.1.2.1).

### Jurassic Stratigraphy of the Florida Peninsula

The Florida carbonate platform began to develop following the establishment of a continental base in North America. Knowledge of the Mesozoic sequence is principally derived from limited oil exploration borings and geophysical data reported in the literature (References 352, 353, 354, 355, and 356).

The Peninsular Arch and other structural/topographic high points (Figure 2.5.1-229) controlled the type and distribution of carbonate depositional facies of Jurassic and Cretaceous sediments including reef complexes that onlapped and wedged or pinched out against these bathymetric highs. A major barrier reef complex of continual reef growth existed in southern Florida from the Cretaceous to Holocene. The presence of nearly continuous carbonate-evaporite cyclical deposition from the Jurassic to the present indicates that sedimentation in the southern Florida Basin kept pace with subsidence (Reference 355). Several wells have been drilled in the southern Florida Basin through carbonate and evaporite sequences to depths as much as 5300 feet (1615 meters) below the Punta Gorda Anhydrite (Figure 2.5.1-228). The deepest well penetrated igneous basement rocks at a total depth of 18,670 feet (Reference 353).

The Upper Jurassic and Lower Cretaceous Wood River Formation (Figure 2.5.1-228) is the stratigraphically lowest sedimentary unit in southern Florida and rests unconformably on rhyolite porphyry dated at 189 Ma (Reference 354). A 100- to 150-foot (46-meter) thick clastic unit forms the basal part of the Wood River Formation and consists of dark-red shale, sandy dolomite, and fine- to coarse-grained arkosic sandstone and calcareous sandstone. These basal clastic units may represent fan, fan-delta, and fluvial-lacustrine or marine deposits. Overlying these clastic rocks is a thick sequence of anhydrite, dolomite, and limestone with occasional interbedded salt stringers, indicating marine transgression. Marine beds are predominant in the formation, and the depositional environment, especially in the southern part of the depositional area, probably favored reef growth (Reference 354). The dolomite is microcrystalline and brown with relict oolitic texture (Reference 356). Interbedded anhydrite, salt stringers, and micritic limestones act as impermeable layers within the more porous dolomite (Reference 355). The thickness of the Wood River Formation ranges from 1700 to 2100 feet (520 to 640 meters) and is generally encountered at depths exceeding 15,000 feet (4572 meters) (Reference 353).

### Cretaceous Stratigraphy of the Florida Peninsula

A major barrier reef complex of continual reef growth existed in southern Florida from the Cretaceous to the Holocene (Reference 355). From the Late Jurassic (161 Ma) through the Early Cretaceous (99 Ma), the continental margin was occupied by a carbonate complex that restricted marine circulation in some areas. In the southeast, this barrier caused the deposition of lagoonal carbonates and anhydrites that formed the Bone Island and Pumpkin Bay formations and the Glades, Ocean Reef, Big Cypress, and Naples Bay Groups (Figures 2.5.1-228 and 2.5.1-230). Carbonate-evaporite deposition in the south ended at the close of the Early Cretaceous and was followed by the deposition of chalk and chalky limestone of the Late Cretaceous Pine Key Formation (Figure 2.5.1-228). By the middle of the Late Cretaceous, the Rebecca Shoal barrier reef had appeared on the upthrown northern side of the straits. This barrier reef expanded to encircle the Florida Peninsula completely. Winston (Reference 357) indicates that following this encirclement, the Paleocene (65 to 56 Ma) Cedar Keys Formation (Figure 2.5.1-231) lagoonal dolomite-anhydrite appears to have been deposited within this enclosed environment.

The Lower Cretaceous Bone Island Formation conformably overlies the Upper Jurassic Wood River Formation (Figure 2.5.1-228). The Lower Cretaceous Bone Island Formation is a sparsely oolitic brown limestone with occasional similarly textured dolomites and anhydrites (Reference 353). Winston (Reference 356) and Applegate et al. (Reference 353) indicate the Bone Island Formation is capped by a regionally persistent 200-foot (61-meter) thick anhydrite layer and contains a 100-foot (30.5-meter) thick lens of dolomite in the type section (well). The Bone Island Formation is approximately 1300 to 2000 feet (400 to 600 meters) thick in southern Florida (References 353 and 356).

The Lower Cretaceous Pumpkin Bay Formation conformably overlies the Cretaceous Bone Island Formation (Figure 2.5.1-228). The Pumpkin Bay Formation is composed of limestone, except at its northern limit, where dolomite is the dominant lithology. A 350-foot thick dolomite zone occurs in the middle and upper parts of the formation and exhibits pinpoint intercrystalline to vuggy porosity

(Reference 355). The limestone is brown and sparsely oolitic with occasional oolitic textured dolomite and two thick (200 feet or 61 meters) anhydrite lenses (Reference 356). Pollastro and Viger (Reference 354) describe organic-rich beds in the upper Pumpkin Bay Formation. Anhydrite and dolomite are predominant in the lower part of the formation (Reference 353). Within the Florida Peninsula, the Pumpkin Bay Formation is as much as 1200 feet (600 meters) thick and thickens westward into the southern Florida Basin depocenter. Onshore, the Pumpkin Bay Formation is found at depths from approximately 12,500 to more than 15,000 feet (3810 to more than 4570 meters) (Reference 354).

The Lower Cretaceous Glades Group conformably overlies the Pumpkin Bay Formation (Figure 2.5.1-228). The Glades Group consists of the Lehigh Acres Formation and Punta Gorda Anhydrite and exhibits a continuous lagoonal carbonate depositional environment. The Lehigh Acres Formation conformably overlies the Cretaceous Pumpkin Bay Formation. The Lehigh Acres Formation is divided into the basal West Felda Shale Member, the Twelve Mile Member, and the uppermost Able Member. The members represent a backreef depositional cycle of limestone, dolomite, and anhydrite with a total thickness that varies from 530 to over 700 feet (References 353 and 356). The West Felda Shale Member consists of dark gray, micaceous, calcareous shale with thin interbeds of brown, micritic limestone up to 200 feet thick. The overlying Twelve Mile Member is composed of relatively thin limestone beds within the main thick, vugular, porous dolomite unit. The Able Member consists of a regionally persistent white to gray anhydrite interbedded with limestone and occasional dolomite beds (References 352 and 353). The Lower Cretaceous Punta Gorda Anhydrite of the Glades Group conformably overlies the Cretaceous Lehigh Acres Formation. Punta Gorda Anhydrite layers have been used as marker beds throughout the Gulf of Mexico and are divided into nine individual anhydrite beds traceable from southern Mississippi to southern Florida (Reference 352). Applegate et al. (Reference 353) describe the Punta Gorda Anhydrite as a series of anhydrite layers approximately 800 feet (244 meters) thick and indicate that it has been found at an elevation of -12,000 feet (-3660 meters) in Collier County. It serves as a regional impermeable seal for hydrocarbon deposits both above and below (Reference 353). Winston (Reference 356) also indicates the presence of salt stringers in the upper Punta Gorda Anhydrite.

The Punta Gorda Anhydrite beds appear to thicken and interfinger with carbonates of the Lower Cretaceous shelf edge reef (Figure 2.5.1-228). Paleontologic data indicate that these interbedded carbonates were deposited in water depths that ranged up to 300 feet (91 meters). Carbonates and anhydrites were deposited simultaneously with carbonate patch reefs developing on crests of paleo highs while evaporites precipitated out of a hyper-saline solution on the flanks. Areas where poor anhydrite bed development occurs indicate areas of patch reefs. Some anhydrite beds are regional whereas others are more restricted. Isopach maps show that these beds were deposited with the long axis of the southern Florida Basin parallel to the reef trend. Anhydrite deposition occurred where evaporation of restricted highstand waters was behind reefs that rimmed the shelf edge (Reference 352). The presence of micritic, calcareous mudstone immediately above the anhydrite provides evidence of the termination of marine regression (Reference 296).

The Lower Cretaceous Ocean Reef Group conformably overlies the Glades Group (Figure 2.5.1-228). The Ocean Reef Group consists of the Sunniland, Lake Trafford, and Rattlesnake Hammock formations. The units of the Ocean Reef Group are typically composed of evaporites and carbonates formed during transgressive-regressive cycles (Reference 360). The units consist of limestones, anhydrites, and dolomites that have been subdivided into multiple formations and groups based on regionally persistent anhydrites that form the uppermost lithologic unit of each formation (Reference 356). Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present, especially in the anhydrite and limestone intervals. The limestones range from white to gray to tan to dark brown and are usually micritic, chalky, and calcarenitic, with skeletal fragments of gastropods, algae, and other fossils. The anhydrite is nodular, microcrystalline, or crystalline and is commonly bedded. Dolomite crystals are euhedral to sucrosic and occur in approximately 30 percent of each formation (Reference 360).

The Sunniland Formation conformably overlies the Punta Gorda Anhydrite (Figure 2.5.1-228). The Sunniland Formation is the basal unit of the Ocean Reef Group. The formation is relatively uniform in thickness within the region and consists of limestone, dolomite, and anhydrite composed of reefs, shoals, carbonate mounds, bioherms, and related features (Reference 354). These shelf carbonates were deposited in beach and shoal-type environments along a high energy, reef-forming band between the shallow-water, low-energy chalky beds and the quiet, deep-water dark micrites to the southwest. Almost all of the effective porosity is in this northwest-southeast band where reef buildup occurred. Secondary dolomitization, which appears to be important in the higher porosity, decreases abruptly perpendicular from this band both to the northeast and the southwest (References 361 and 362). The upper Sunniland Formation represents a shoaling-upward depositional cycle that extends throughout onshore and offshore southern Florida. This slowly oscillating transgression-regression cycle continues to the top of the Sunniland Formation, where it gives way to a major marine regression with the deposition of the Lake Trafford Formation (Reference 359).

The lower Sunniland Formation is composed of brown and medium-dark-gray micritic and argillaceous limestones that are commonly algal laminated. The dark carbonate facies varies in thickness up to 150 feet and thins toward the eastern and southern margins of the southern Florida Basin (Reference 355). The dark carbonate unit called the “rubble zone” in the lower Sunniland Formation is burrowed, fractured, and stylolitized. The lower zone is enclosed by impermeable, micritic, tidal-flat, calcareous mudstones above and sealed below by the Punta Gorda Anhydrite (Reference 363). At the base of the Sunniland Formation, at the top of the Punta Gorda Anhydrite, is evidence of a slow marine transgression and the termination of a major regression (Reference 359).

The upper Sunniland Formation consists of isolated fossil-shell, skeletal-petal, porous, and permeable grainstone mounds enclosed by impermeable lagoonal mudstones, wackestones, nodular anhydrite beds, and micritic carbonates-some of which have been dolomitized (Reference 354). This facies may represent storm deposition as shoals in a regionally restricted, back-reef lagoonal area in a warm, shallow marine-shelf setting (References 361 and 364). Biotic abundance and content of the fragmented biotic material suggests that the debris mound facies were deposited on micritic tidal mud flats during a sea level rise (Reference 365). During a later regression, the upper portions of these porous shoal mounds were sub-aerially exposed, leached, and dolomitized during a low sea-level stand, further increasing the porosity of the upper Sunniland Formation carbonates. Individual mounds are between 40 and 100 feet thick (Reference 363). These highly porous bioclastic mounds accumulated along the southeastern coast of the Florida Peninsula on subtle topographic/bathymetric highs that were probably related to underlying basement structure (Reference 363). Pollastro (Reference 363) also describes an anhydrite-cemented nonporous sabkha-like facies near the southern boundary of the upper Sunniland Formation that formed in supratidal arid conditions on restricted coastal plains above high tide level (Reference 354).

Pollastro et al. (Reference 366) indicate that the depth to the top of the Sunniland Formation in southern Florida is approximately 10,000 feet and increases to greater than 12,000 feet in the Florida Basin to the southwest. The thickness of the Sunniland Formation in southern Florida is approximately 200 to 300 feet and increases toward the south (Reference 355).

The Lake Trafford Formation conformably overlies the Sunniland Formation (Figure 2.5.1-228). A major regression in the continuing oscillating transgression occurred with the deposition of the Lake Trafford Formation (Reference 359). The Lake Trafford Formation consists of a limestone-dolomite unit with a thin (<100 feet) anhydrite lens. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Lake Trafford Formation is approximately 150 feet thick in southern Florida (Reference 356).

The Rattlesnake Hammock Formation conformably overlies the Lake Trafford Formation (Figure 2.5.1-228). The Rattlesnake Hammock Formation consists of a 200-foot thick anhydrite cap underlain by cyclic deposits of limestone, anhydrite, dolomite, anhydrite, and limestone units

successively. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Rattlesnake Hammock Formation is approximately 600 feet thick in southern Florida (Reference 356).

The Lower Cretaceous Big Cypress Group conformably overlies the Rattlesnake Hammock Formation of the Ocean Reef Group (Figure 2.5.1-228). The Big Cypress Group consists of the Marco Junction, Gordon Pass, and Dollar Bay formations. The units of the Big Cypress Group are typically composed of evaporites and carbonates formed during transgressive-regressive cycles (Reference 360). The units consist of limestones, anhydrites, and dolomites that have been subdivided into multiple formations and groups based on regionally persistent anhydrites that form the uppermost lithologic unit of each formation (Reference 356). Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present, especially in the anhydrite and limestone intervals. The limestones range from white to gray to tan to dark brown and are usually micritic, chalky, and calcarenitic, with skeletal particles of gastropods, algae, and other fossils. The anhydrite is nodular, microcrystalline, or crystalline and is usually bedded. The dolomite crystals are euhedral to sucrosic and occur in approximately 30 percent of each formation (Reference 360).

The Lower Cretaceous Marco Junction Formation conformably overlies the Rattlesnake Hammock Formation (Figure 2.5.1-228). The Marco Junction Formation consists of a relatively thin (<100 feet) anhydrite cap underlain by a sequence of limestones and dolomites and a second thin anhydrite lens. The Marco Junction Formation is approximately 350 feet thick in southern Florida (Reference 356).

The Gordon Pass Formation conformably overlies the Marco Junction Formation (Figure 2.5.1-228). The Gordon Pass Formation consists of a thick (>100 feet) anhydrite cap underlain by a sequence of limestones and dolomites and a second thin anhydrite lens. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Gordon Pass Formation is approximately 475 feet thick in southern Florida (Reference 356).

The Dollar Bay Formation is the uppermost unit of the Big Cypress Group; it conformably overlies the Gordon Pass Formation (Figure 2.5.1-228). The Dollar Bay Formation commonly consists of evaporite-carbonate beds of limestone, dolomite, and anhydrite formed during a transgressive-regressive cycle (Reference 354). The Dollar Bay Formation consists of reefs, shoals, carbonate mounds, bioherms, and related features, forming organic-rich calcareous units inter-bedded with the carbonates (Reference 355). The limestone, dolomite, and anhydrite units occur in a series of cycles that typically begin and end with anhydrite. Porous carbonate units were deposited as tidal shoal deposits and patch reefs in a tidal flat, lagoonal, restricted-marine setting and in a sub-tidal platform, open-marine setting (Reference 354). Environments from shallow shelf to euxinic are present (Reference 367).

The Dollar Bay Formation consists of a 55-foot thick dark brown, fine crystalline dolomite with intercrystalline porosity that is typically found at the base of the formation. Above this lies a sedimentary cycle averaging 325 feet thick consisting characteristically of chalky dolomite and limestone interspersed with beds of fine-grained calcarenite (Reference 367). Leached limestone units (Reference 354) formed from isolated patch reefs (Reference 360) are present in the middle part of the Dollar Bay Formation. A porous dolomite unit forms the upper part of the formation (References 354 and 360). These units are capped with an impermeable tidal flat deposit of micritic, argillaceous lime mudstone and an uppermost anhydrite unit (Reference 354). The Dollar Bay Formation occurs at depths of more than 10,000 feet (3050 meters) and averages 450 feet (140 meters) thick but ranges up to as much as 620 feet (190 meters) thick in some parts of the southern Florida Basin (Reference 354). All contacts above, below, and within the Dollar Bay Formation are conformable (Reference 367).

The Lower Cretaceous Naples Bay Group conformably overlies the Big Cypress Group (Figure 2.5.1-228). The Naples Bay Group consists of the Panther Camp, Rookery Bay, and

Corkscrew Swamp formations. The units of the Naples Bay Group are typically composed of evaporites and carbonates formed during transgressive-regressive cycles (Reference 360). The units consist of limestones, anhydrites, and dolomites that have been subdivided into multiple formations and groups based on regionally persistent anhydrites that form the uppermost lithologic unit of each formation (Reference 356). Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present, especially in the anhydrite and limestone intervals. The limestones range from white to gray to tan to dark brown and are usually micritic, chalky, and calcarenitic, with skeletal particles of gastropods, algae, and other fossils. The anhydrite is nodular, microcrystalline, or crystalline and is usually bedded. The dolomite crystals are euhedral to sucrosic and occur in approximately 30 percent of each formation (Reference 360).

The Panther Camp Formation conformably overlies the Dollar Bay Formation (Figure 2.5.1-228). The Panther Camp Formation consists of a thin (<100 feet) cap of anhydrite underlain by two limestone-dolomite units separated by an anhydrite layer. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Panther Camp Formation is approximately 350 feet thick in southern Florida (Reference 356).

The Rookery Bay Formation of the Naples Bay Group conformably overlies the Panther Camp Formation (Figure 2.5.1-228). The Rookery Bay Formation consists of a thin (<100 feet) cap of anhydrite underlain by two limestone-dolomite units separated by a thin anhydrite layer. This, in turn, is underlain by a dolomite unit and a limestone unit. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Rookery Bay Formation is approximately 500 feet thick in southern Florida (Reference 356).

The Corkscrew Swamp Formation of the Naples Bay Group conformably overlies the Rookery Bay Formation (Figure 2.5.1-228). The Corkscrew Swamp Formation consists of a thin (<100 feet) cap of anhydrite underlain by a dolomite unit and two limestone units separated by a second 100-foot thick anhydrite layer. Calcareous shales, mudstones, salt, and lignitic carbonaceous materials are also present. The Corkscrew Swamp Formation is approximately 450 feet thick in southern Florida (Reference 356).

The Upper Cretaceous Pine Key Formation conformably overlies the Corkscrew Swamp Formation of the Naples Bay Group and is the uppermost Mesozoic formation in southern Florida (Figure 2.5.1-228). Its lower contact is conformable in southern and eastern Florida, but unconformable to the west (Reference 368). The Pine Key Formation is essentially made up of two facies: (a) a white chalk and chalky limestone formed in a lagoonal environment that interfingers with and is replaced by (b) the lower tongues of a regional barrier reef complex composed of tan, cream, light gray, and brown very fine microcrystalline to coarse crystalline euhedral and anhedral dolomite (References 369, 357, and 370). The reef facies is characterized by vugs and reports of cavities and wall collapse zones (References 371, 357, and 370). Neither evaporites nor dolomites are present within the lagoonal facies of the Pine Key Formation indicating that the Rebecca Shoals barrier-reef complex did not completely encircle Florida or otherwise restrict circulation during the Late Cretaceous (Reference 370). In the South Florida back-reef basin, deposition of the lagoonal chalk facies of the Pine Key Formation persisted until the barrier reef had completely encircled the Florida Peninsula during the Paleocene (Reference 357). The Pine Key Formation is as much as 3000 feet thick in southern Florida at a depth of approximately 5500 to 6000 feet (References 356 and 369).

### Cenozoic Stratigraphy of the Florida Peninsula

The early part of the Cenozoic consists of a depositional shallow marine environment of carbonate rocks (limestone and dolostone with some evaporites). These carbonate rocks include the Cedar Keys Formation, Oldsmar Formation, Avon Park Formation, Ocala Limestone, Suwannee Limestone, and part of the basal Arcadia Formation (Figure 2.5.1-231). The occurrence of gypsum and anhydrite during the Cenozoic indicates that seawater circulation in the shallow marine environment was

periodically restricted (Reference 287). During the Cenozoic (last 65 m.y.), sea level fluctuated ±100 feet above and below the present-day sea level (Reference 287).

The oldest Cenozoic sediment that crops out in the site region is the Miocene-Pliocene Peace River Formation, exposed in Hardee and DeSoto counties (Figure 2.5.1-201). All deeper Cenozoic units occur only in the subsurface. The Cenozoic sedimentary section in southern Florida averages approximately 5000 to 6000 feet thick and consists of a sequence of carbonate deposition interrupted by Appalachian derived siliciclastic sediments during the Miocene and Pliocene (References 373, 374, 375, 376, 377, 378, 356, and 379). The regional Cenozoic stratigraphic section is shown in Figure 2.5.1-231. Figures 2.5.1-232, 2.5.1-233, 2.5.1-234, 2.5.1-235, and 2.5.1-236 provide geologic cross sections illustrating the regional Cenozoic stratigraphy.

### Paleocene Stratigraphy of the Florida Peninsula

By the Paleogene, the Appalachian Mountains had been gradually lowered by erosion. This not only reduced the supply of siliciclastic material, but also resulted in a lower stream gradient that limited the transport of siliciclastic sediments to the Florida Platform. In addition, the currents in the Suwannee Channel (also known as the “Gulf Trough” or “Suwannee Straits”) (Figures 2.5.1-218 and 2.5.1-229) acted as a barrier to siliciclastic transport. These currents protected the carbonate depositional environment of the Florida Platform from the influx of siliciclastic sediments resulting in predominantly carbonate deposition during the Paleogene (References 338 and 349).

The oldest Cenozoic formation on the Florida Platform is the Paleocene Cedar Keys Formation that conformably overlies the Late Cretaceous Pine Key Formation (Figure 2.5.1-231). The Cedar Keys Formation is a marine lagoonal facies that occurs within the confines of the Rebecca Shoal barrier reef (Reference 369). In southern Florida, the Cedar Keys Formation consists primarily of gray dolomite, gypsum, and anhydrite with a minor percentage of limestone. The upper part of the Cedar Keys Formation consists of coarsely crystalline, porous dolomite. The lower part of the Cedar Keys Formation contains more finely crystalline dolomite interbedded with anhydrite (Reference 369). The configuration of the Paleocene sediments in Peninsular Florida reflects depositional controls inherited from pre-existing Mesozoic structures such as the Peninsular Arch and the southern Florida Basin. The upper unit of porous dolomite in the Cedar Keys Formation forms the base of the Floridan aquifer system (Subsection 2.4.12) throughout southern Florida (Reference 349), where it is found at elevations ranging from -3000 to -4000 feet (-900 to -1200 meters) (Reference 389). The Cedar Keys Formation varies from approximately 500 feet up to 2000 feet (150 meters up to 600 meters) thick in southern Florida (References 356 and 375).

### Eocene Stratigraphy of the Florida Peninsula

The Eocene Oldsmar Formation within southern Florida conformably overlies the Paleocene Cedar Keys Formation (Figure 2.5.1-231). The Oldsmar Formation primarily consists of a sequence of white, cream to gray, micritic to chalky limestones interbedded with tan to light-brown microcrystalline, vuggy dolomite. Dolomitization is usually more extensive in the lower part of the formation that is also noticeably unfossiliferous (References 390, 376, and 349). Gypsum and thin beds of anhydrite occur in some places. According to Winston (Reference 379), the top of the Oldsmar Formation in southern Florida is not identifiable or distinguishable on the basis of lithologic and faunal criteria. However, in southern Florida, the top of the uppermost thick dolomite unit is marked by glauconitic limestone (Reference 376). The “Boulder Zone,” a regional hydrostratigraphic unit recognized in the subsurface of South Florida (Subsection 2.4.12), forms part of the lower Oldsmar Formation and characteristically contains fractured dolomite (Reference 376). The Oldsmar Formation occurs in the subsurface at elevations ranging from -1950 to -2250 feet (-590 to -690 meters) (Reference 375). It ranges from 500 to as much as 1500 feet (150 to as much as 460 meters) thick in southern Florida (Reference 376). Observations recorded during the construction of the Class V exploratory well EW-1 at the Turkey Point Units 6 & 7 site provide a site-specific measurement for depth to the top of the Oldsmar formation of approximately 2580 feet below ground

surface (bgs). All depths for well EW-1 are reported as below pad level, which represents the depth below the top of the 64-inch-diameter pit pipe. The pit pipe was surveyed and found to be at elevation 7.18 feet North American Vertical Datum of 1988 (NAVD 88), which is approximately 0.4 feet above the final ground surface (6.8 feet NAVD 88) at the exploratory well (Reference 970).

The Eocene Avon Park Formation overlies the Oldsmar Formation (Figure 2.5.1-231). A regional unconformity in southern Florida has been proposed at the top of the Oldsmar Formation/base of the Avon Park Formation (Reference 375). The Avon Park Formation consists of cream to light brown or tan, poorly indurated to well-indurated, variably fossiliferous, marine limestone (grainstone, packstone, and wackestone, with rare mudstone). These limestones are interbedded with light brown to orange-brown to dark brown or black, very poorly indurated to well indurated to dense, sucrosic to very fine to medium crystalline, fossiliferous (molds and casts), vuggy dolomites. Fine- to medium-grained calcarenite that is moderately to well sorted is intermittently present. Portions of the middle Avon Park Formation are very fine-grained with low permeability and act as confining beds separating the Avon Park Formation into upper and lower (formerly Lake City) parts (Reference 349). The fossils present include mollusks, foraminifera, echinoids, algae, and carbonized plant remains (References 377 and 376). The top of the Avon Park Formation is marked in some places by light brown, finely crystalline to fossiliferous dolomitic limestone or dolomite, thinly interbedded with limestone. Thick intervals containing mostly dolomite, but in some places interbedded with limestone, are commonly present in the middle to lower part of the Avon Park Formation in southern Florida. High permeability due to fracturing is common, particularly in dolomite units. Gypsum and anhydrite also occur in the lower part of this formation in southwestern Florida, either as bedded deposits or more commonly as intergranular or pore-filling material in the carbonate rocks. An upper marker horizon separates the more thinly bedded strata of the upper Avon Park Formation from more thickly bedded and massive units of the lower Avon Park Formation (Reference 376).

The shallow marine limestones and dolomites of the Avon Park Formation were deposited primarily on the inner part of a broad, flat-lying carbonate ramp that sloped gently toward the Gulf of Mexico during the Eocene (Reference 391). Carbonates of the Avon Park Formation are the oldest sediments exposed in the state and crop out in a limited area on the crest of the Ocala Platform in the central peninsula (Reference 377). The Avon Park Formation varies from 400 feet up to 1200 feet thick in southern Florida (References 375 and 356) and occurs at elevations ranging from -1000 to -1300 feet (Reference 375). Observations recorded during the construction of the Class V exploratory well EW-1 at the Turkey Point Units 6 & 7 site provide a site-specific measurement for the Avon Park Formation from a depth of 1255 to 2580 feet bgs (1255 to 2580 feet below pad level).

The Eocene Ocala Limestone overlies the Eocene Avon Park Formation (Figure 2.5.1-231). A regional unconformity in southern Florida has been proposed at the top of the Avon Park Formation/base of the Ocala Limestone (Reference 375). The Ocala Limestone consists of white to cream, micritic or chalky marine limestones, calcarenitic limestone, coquinoid limestone, and occasional dolomites (References 376 and 392). Generally the Ocala Limestone is soft and porous, but in places it is hard and dense because of cementation of the particles by crystalline calcite. The deposit is unique in that it is composed of almost pure calcium carbonate from shells and micritic chalky particles (Reference 392). It can be subdivided into lower and upper facies on the basis of lithology. The lower unit is composed of a white- to cream-colored, fine- to medium-grained, poorly to moderately indurated, very fossiliferous limestone (grainstone and packstone). The lower facies may not be present throughout the areal extent of the Ocala Limestone and may be partially to completely dolomitized in some regions. The upper facies is a white, poorly to well indurated, poorly sorted, very fossiliferous limestone (grainstone, packstone, and wackestone). Silicified limestone is common in the upper facies. Fossils present in the Ocala Limestone include abundant large and smaller foraminifera, echinoids, bryozoans, and mollusks. Where the Ocala Limestone is at or near the surface, it exhibits extensive karstification (see discussion of hydrologic features related to karstification processes in Subsection 2.5.1.1.1.1.1.1) (Reference 377). The limestone is characterized by abundant large benthic foraminifera, which have been used by various workers to

distinguish the Ocala Limestone from the overlying Suwannee Limestone and the underlying Avon Park Formation (Reference 376).

The fine-grained carbonates of the Ocala Limestone (Figure 2.5.1-231) were deposited on the middle to outer-ramp setting at water depths generally below storm wavebase (Reference 391). The Ocala Limestone occurs at the surface in a few locations, but appears to be absent even in the subsurface of the southernmost part of southeastern Florida (most of Miami-Dade County and southeastern Broward County). In the remainder of southern Florida, the thickness of the Ocala Limestone varies from 200 to 400 feet (61 to 122 meters) (References 376 and 356) and occurs at elevations ranging from -980 to -1100 feet (-300 to -335 meters) (Reference 375).

### Oligocene Stratigraphy of the Florida Peninsula

A significant increase of siliciclastic sediments occurred during the Oligocene, possibly due to renewed uplift of the Appalachian Mountains. The Suwannee Channel (also known as the Suwannee Straits) was filled with a flood of siliciclastic sediments as a possible result of longshore transport and currents. As a result of filling the Suwannee Channel, the carbonate depositional environment was replaced with sands, silts, and clays (Reference 349). The siliciclastic sediments appear in the early Miocene in northern Florida; however, in southern Florida, carbonates continued to be deposited until at least mid-Miocene. The siliciclastics spread southward along the east coast of Florida due to active transport conditions along the Atlantic coastline (Reference 287). The siliciclastic depositional environment moved further south due to longshore transport and currents until almost the entire Florida Platform was covered with sands and clays. The incursion of siliciclastics diminished during the later Pleistocene (Reference 349).

Karst features began to form at least as early as the latest Oligocene as determined from the occurrence of terrestrial vertebrate faunas (Reference 349). Karst features such as sinkholes, dissolution valleys, and collapse depressions formed when groundwater flowed through Florida's Eocene, Oligocene, and Miocene limestones and dissolved these carbonate sediments (Reference 287).

The Early Oligocene Suwannee Limestone overlies the Eocene Ocala Limestone (Figure 2.5.1-231). A regional unconformity in southern Florida has been described at the top of the Ocala Limestone/base of the Suwannee Limestone (References 375 and 393). The Suwannee Limestone consists of a white to cream, poorly to well indurated, fossiliferous, vuggy to moldic marine limestone (grainstone and packstone) with minor amounts of quartz sand and rare-to-absent phosphate mineral grains. The dolomitized parts of the Suwannee Limestone are gray, tan, light brown to moderate brown, moderately to well indurated, finely to coarsely crystalline, dolomite with limited occurrences of fossiliferous (molds and casts) beds. Silicified limestone and chert are common (References 392 and 377). Up to seven lithofacies have been identified in the Suwannee Limestone based on biotic content and texture (Reference 373). Characteristic porosity and permeability in the Suwannee Limestone is interparticle to moldic or vuggy (Reference 376). Mollusks, foraminifers, corals, and echinoids are present in the Suwannee Limestone (Reference 392).

During deposition of the early Oligocene Suwannee Limestone, a series of clean siliciclastic shoreline deposits began to prograde onto the southern Florida Platform that extended along the present west coast of Florida; however, these siliciclastic sediments did not affect the continued deposition of carbonate sediments in an open circulation shelf setting (Reference 394). The Suwannee Limestone represents the continued deposition in shallow marine conditions during the early Oligocene (Reference 391). The Suwannee Limestone exhibits numerous cycles of limestone capped by brecciated karst suggesting subaerial exposure; each cycle is overlain by a landward shift in sedimentary facies (marine flooding) (Reference 373). Various publications contain opposing interpretations concerning the presence or absence of the Suwannee Limestone in southeastern Florida. The Suwannee Limestone may be absent from the eastern side of the Peninsular Arch (Figures 2.5.1-229, 2.5.1-232, and 2.5.1-234) due to erosion, nondeposition, or both

(References 232, 267, and 377). In southern Florida, the thickness of the Suwannee Limestone varies from 200 feet to as much as 600 feet in Lee and western Collier counties (References 356 and 376) and occurs at elevations ranging from -900 to -1300 feet (Reference 375).

The Oligocene-Miocene-Pliocene Hawthorn Group unconformably overlies the Oligocene Suwannee Limestone (Figure 2.5.1-231). The Hawthorn Group consists of an interbedded sequence of widely varying lithologies and components that include limestone, mudstone, dolomite, dolomitic silt, shells, quartz sand, clay, abundant phosphate grains, and mixtures of these materials. The characteristics that distinguish the Hawthorn Group from underlying units are (a) high and variable siliciclastic and phosphatic content; (b) color, which can be green, olive-gray, or light gray; and (c) a distinguishing gamma-ray log response. Intervals high in phosphate sand or gravel content are present and have high gamma-ray log activity, with peaks of 100 to 200 API gamma ray units or more (Reference 376). In southern peninsular Florida, the Hawthorn Group consists of the basal Oligocene-Miocene Arcadia Formation, including the Tampa Member and the uppermost Miocene-Pliocene Peace River Formation with its Bone Valley Member (Reference 377). Zones of dissolution of Oligocene rocks indicate that post-Oligocene erosion was extensive (Reference 375). The complete Hawthorn Group varies from 500 to 800 feet thick in southern Florida (References 373 and 394).

A regional unconformity in southern Florida has been proposed at the top of the Suwannee Limestone/base of the Arcadia Formation of the Hawthorn Group (References 373 and 394). Zones of dissolution of Oligocene rocks indicate that post-Oligocene erosion was extensive (Reference 375). The Arcadia Formation is predominantly a carbonate unit with a variable siliciclastic component, including thin beds of quartz sands. The Arcadia Formation (with the exception of the Tampa Member) is composed of yellowish gray to light olive gray to light brown, micro to finely crystalline, variably sandy, clayey, and phosphatic, fossiliferous limestones and dolomites. Thin beds of sand and clay are common. The sands are yellowish gray, very fine- to medium-grained, poorly to moderately indurated, clayey, dolomitic, and phosphatic. The clays are yellowish gray to light olive gray, poorly to moderately indurated, sandy, silty, phosphatic, and dolomitic. Molds and casts of mollusks are common in the dolomites (Reference 377). Sediments within the Arcadia Formation show an upward and geographically northward (Reference 395) change from predominantly carbonate with some quartz sand to an equal mix of siliciclastics and carbonates (Reference 394). The Tampa Member occurs near the base of the Arcadia Formation and is predominantly a white to yellowish gray fossiliferous marine limestone (mudstone, wackestone, and packstone) with subordinate dolomite, sand, clay, and phosphate. The Tampa Member is usually a hard, massive crystalline rock, and in some areas it contains small moldic cavities. Mollusks and corals, foraminifera, and algae are common in the Tampa Member.

Subsurface data show the Arcadia Formation as a gently sloping carbonate ramp upon which was deposited multiple high frequency, fining/coarsening upward, eustatically-driven siliciclastic sequences. Fossil evidence suggests a shift from tropical to subtropical oceanic conditions during Arcadian deposition (Reference 373). The Tampa Member and the lower part of the Arcadia Formation form the upper part of the Floridan aquifer system (Subsection 2.4.12) in parts of southern Florida (References 377 and 392). The thickness of the Arcadia Formation in southern Florida varies from 100 to 700 feet and occurs at elevations ranging from -300 to -650 feet (-91 to -200 meters) (References 373 and 394).

### Miocene Stratigraphy of the Florida Peninsula

During the Miocene, siliciclastics covered the Florida Platform providing a semipermeable barrier that reduced dissolution of the underlying carbonates. However, erosion of these siliciclastics during the early Pleistocene renewed the dissolution of the underlying limestones formations. This dissolution led to increased karst and an enhanced secondary porosity of the sediments of the Floridan aquifer system.

During the early Miocene a strong southward flood of terrigenous coarse clastics, presumably from the southern Appalachian Mountains, prograded over most of the Florida Platform (References 368 and 393) (Figure 2.5.1-237). The Hawthorn Group of shallow marine to non-marine coastal and deltaic sandstones and mudstones prograded out over the older carbonate platform during the late Oligocene to Pliocene (Reference 391). By the end of the Oligocene, the influx of siliciclastic sediments, principally from the Appalachians, increased in volume and fines content. Carbonate production was significantly reduced in the east; however, the slow rate of sediment influx and lack of significant clay content allowed continued carbonate growth to continue into the mid-Miocene in the central portion of the Florida Platform (Reference 394). This drowning/burial by siliciclastics is not the only interpretation for the reduction of the carbonate-producing organisms. McNeill et al. (Reference 395) suggest sea level rise, environmental deterioration, and the influence of local ocean currents as viable alternatives (see discussion of carbonate platforms: growth, shut downs and crashes in Subsection 2.5.1.1.1.1.1.2).

The middle Miocene-early Pliocene Peace River Formation of the Hawthorn Group unconformably overlies the Oligocene-Miocene Arcadia Formation (Figure 2.5.1-231). The base of the Peace River Formation is a regional unconformity that is identified by a thin, black phosphorite layer that appears to be the source of a strong gamma log response (Reference 395) (Figure 2.5.1-234). The Peace River Formation is composed of interbedded sands, clays, and carbonates. The sands are generally light gray to olive gray, poorly consolidated, clayey, variably dolomitic, very fine- to medium-grained, and phosphatic. The clays are yellowish gray to olive gray, poorly to moderately consolidated, sandy, silty, phosphatic, and dolomitic. The carbonates are usually light gray to yellowish gray, poorly to well indurated, variably sandy, clayey, and phosphatic dolomites. Two distinct lithologies are present in the subsurface in southern Florida: a lower diatomaceous mudstone unit and an upper unit of mud-rich, very fine quartz sandstone. Fossil mollusks occur as reworked casts, molds, and limited original shell material. The Bone Valley Member of the Peace River Formation crops out in a limited area on the southern part of the Ocala Platform in Hillsborough, Polk, and Hardee counties. Where it is present, the Bone Valley Member is a poorly consolidated clastic unit consisting of sand-sized and larger phosphate grains in a matrix of quartz sand, silt, and clay. The lithology is highly variable, ranging from sandy, silty, phosphatic clays, and relatively pure clays to clayey, phosphatic sands to sandy, clayey phosphorites. Colors range from white, light brown, and yellowish gray to olive gray and blue green. Vertebrate fossils occur in many of the beds within the Bone Valley Member. Shark's teeth are often abundant (Reference 377). The Peace River Formation may be, in part, correlative to the proposed Long Key Formation (Reference 373).

Cunningham et al. (Reference 396), McNeill et al. (Reference 395), and Ward et al. (Reference 391) suggest that the Peace River Formation represents the southward transport and deposition of continental siliciclastics in a fluvial-deltaic system, which eroded and prograded out over the older carbonate platform environment. Well data show intervals of quartz sand localized as a wide north-south pathway from the central part of the peninsula to the middle Florida Keys (Reference 393) (Figure 2.5.1-237). This pathway is interpreted as a record of a strong, southward-moving shoreline and channeled deposition or a regional prograding spit (Reference 393). The ultimate source of these siliciclastics is considered to be the distant Appalachian highlands (References 395, 393, and 368). The Peace River Formation is widespread in southern Florida. It is part of the intermediate confining unit between the surficial and Floridan aquifer systems (References 376 and 377) (Subsection 2.4.12). The thickness of the Peace River Formation in southern Florida varies from 100 to 650 feet and occurs at elevations ranging from -100 to -250 feet (-30.5 to -76 meters) (References 373 and 394).

### Pliocene Stratigraphy of the Florida Peninsula

The Pliocene Tamiami Formation unconformably overlies the Miocene-Pliocene Peace River Formation of the Hawthorn Group and interfingers with the contemporaneous Cypresshead Formation (Figure 2.5.1-231). The Tamiami Formation in southern Florida is a poorly defined lithostratigraphic unit containing a wide range of mixed carbonate-siliciclastic lithologies that include:

(a) light gray to tan, unconsolidated, fine- to coarse-grained, fossiliferous sand; (b) light gray to green, poorly consolidated, fossiliferous sandy clay to clayey sand; (c) light gray, poorly consolidated, very fine- to medium-grained, calcareous, fossiliferous sand; (d) white to light gray, poorly consolidated, sandy, fossiliferous limestone; and (e) white to light gray, moderately to well indurated, sandy, fossiliferous limestone (Reference 377). Phosphatic sand- to gravel-sized grains are present in small quantities within virtually all the lithologies. Fossils present in the Tamiami Formation occur as molds, casts, and original material. The fossils present include barnacles, mollusks, corals, echinoids, foraminifera, and calcareous nannoplankton (Reference 377). The occurrence of limestone lenses in the Tamiami Formation appears to be related to fluctuations of the water table accompanied by cementation with calcium carbonate. The faunal assemblage of the Tamiami Formation commonly contains a variety of mollusks (Reference 397). The lower unit of the Tamiami Formation includes greenish sandy, clayey silt beds of low permeability that vary in thickness and extent and conform with the surface of the underlying Hawthorne Formation. The argillaceous content of the lower Tamiami and underlying Peace River strata is expressed in well logs regionally and at the Turkey Point site by an increase in activity on the gamma ray log (References 391 and 708). The complex mix of permeable and impermeable lithologies makes the Tamiami Formation part of both the surficial aquifer system and the intermediate confining unit between the surficial and Floridan aquifer systems (References 376, 377, and 862) (Subsection 2.4.12). The Tamiami Formation may be, in part, correlative to the proposed Long Key Formation (Reference 373).

Cunningham et al. (Reference 396) suggest that the presence of minor carbonate in the Tamiami Formation reflects a shift from the progradation of siliciclastics to aggradation of a vertical mix of carbonates and siliciclastics. The top of the Tamiami Formation is an undulating surface that varies as much as 25 feet (7.6 meters) in elevation within a distance of 8 miles (13 kilometers) (Reference 397). This unevenness indicates that the upper part has been subjected to erosion. The deposition of the Caloosahatchee Formation on top of and along the flanks of erosional remnants indicates that the Tamiami Formation was dissected prior to Pliocene deposition and again during the Pleistocene. Apparently the deeper valleys were developed during the Pleistocene (Reference 397) in response to lower sea levels caused by glaciation. The Tamiami Formation occurs at or near the land surface in Charlotte, Lee, Hendry, Collier, and Monroe counties (Reference 377). In Collier and Lee counties, Schroeder and Klein (Reference 397) found the Tamiami Formation to be approximately 50 feet (15 meters) thick, while in Miami-Dade County various reports (References 397 and 398) indicate it ranges in thickness from 25 to 220 feet (7.6 to 67 meters).

The Pliocene Cypresshead Formation unconformably overlies the Miocene-Pliocene Peace River Formation of the Hawthorn Group and interfingers with the contemporaneous Tamiami Formation (Figure 2.5.1-231). The Cypresshead Formation consists of reddish brown to reddish orange, unconsolidated to poorly consolidated, fine- to very coarse-grained, clean to clayey sands. Cross-bedded sands are common within the Cypresshead Formation. Discoid quartzite pebbles and mica are often present. Clay beds are scattered and not really extensive. Original fossil material is not present in the sediments although poorly preserved molds and casts of mollusks and burrow structures are occasionally present. The Cypresshead Formation is at or near the surface from northern Nassau County southward to Highlands County forming the peninsular highlands (Lakeland, Lake Henry, Winter Haven, and Lake Wales Ridges) and appears to be present in the subsurface southward and to underlie the Florida Keys (Figure 2.5.1-217). The Cypresshead Formation formed in a shallow marine, near-shore environment and consists of deltaic and prodeltaic sediments (Reference 377). The Cypresshead Formation may be in part correlative to the proposed Long Key Formation (Reference 373). The Cypresshead Formation is approximately 50 to 60 feet thick in Polk County (Reference 399).

The Pliocene-Pleistocene shell beds have attracted much attention due to the abundance and preservation of the fossils but the biostratigraphy and lithostratigraphy of the units has not been well defined. The “formations” previously recognized within the latest Tertiary-Quaternary section of southern Florida include the Late Pliocene-Early Pleistocene Caloosahatchee Formation and the

Late Pleistocene Fort Thompson Formation (Figure 2.5.1-231). Lithologically these sediments are complex, varying from unconsolidated, variably calcareous and fossiliferous quartz sands to well indurated, sandy, fossiliferous limestones (both marine and freshwater). Clayey sands and sandy clays are present. These sediments form part of the surficial aquifer system (Reference 377) (Subsection 2.4.12). The identification of these units is problematic unless the significant molluscan species are recognized (Reference 377); over 680 species are presently recognized (Reference 397). Often the collection of representative faunal samples is not extensive enough to properly discern the biostratigraphic identification of the formation. In an attempt to alleviate the inherent problems in the recognition of lithostratigraphic units, Scott (Reference 349) suggests grouping the latest Pliocene through late Pleistocene Caloosahatchee Formation and Fort Thompson Formation into a single lithostratigraphic unit. This unit may be in part correlative to a proposed Long Key Formation (Reference 373) (Figure 2.5.1-231). In mapping these shelly sands and carbonates, a generalized grouping termed the Tertiary-Quaternary shell-bearing units was used by Scott (Reference 377) in the preparation of the Geologic Map of Florida. A more detailed description of the units identified as the Caloosahatchee and Fort Thompson formations follows.

The Pliocene-Pleistocene shell-bearing sediments, also known as the Caloosahatchee Formation, unconformably overlie the Pliocene Tamiami Formation (Reference 397) (Figure 2.5.1-231). The Caloosahatchee Formation consists of fossiliferous quartz sand with variable amounts of carbonate matrix interbedded with variably sandy, shelly limestones. Freshwater limestones are commonly present within the Caloosahatchee Formation (Figure 2.5.1-231). Fresh unweathered exposures are generally pale cream-colored to light gray, although green clay marls have been included in the formation. Green silty sands or sandy silts in the Caloosahatchee Formation appear to be restricted to the flanks of the hills of the Tamiami Formation. The greenish clastics are considered redeposited green clay marls of the Tamiami Formation (Reference 397). Mollusks are typically the predominant fossils, along with corals, bryozoans, echinoids, and vertebrates (Reference 392). The sand and shell variations of the Caloosahatchee Formation can be separated from the Pleistocene marine formations by identification of the mollusk faunas (Reference 397).

Sediments identified as part of the Caloosahatchee Formation occur from Tampa south to Lee County and to the east coast (Reference 349). The Caloosahatchee Formation is present in southern Florida as discontinuous erosion remnants. The most continuous exposures occur as thin beds along the Caloosahatchee River and other rivers along the southwest Florida coast (Reference 397). The Caloosahatchee Formation has not been identified on the southeast Florida mainland (Reference 393). The Caloosahatchee Formation is at least 10 feet thick along the Caloosahatchee River and may be as much as 20 feet thick near Lake Hicpochee (Reference 397).

The Pliocene-Pleistocene shell-bearing sediments, also known as the Fort Thompson Formation, appear to conformably overlie the Pliocene Tamiami Formation but lie unconformably on the Caloosahatchee Formation (Reference 397) (Figure 2.5.1-231). The discontinuity surfaces within the Fort Thompson Formation can include dense, well-indurated laminated crusts (Reference 400). Both Sonenshein (Reference 401) and Wilcox et al. (Reference 402) split the Fort Thompson Formation into an upper and lower unit based on lithologic and core data. The Fort Thompson Formation is typically composed of interbedded marine limestone, minor gastropod-rich freshwater limestone, shell marl, sandy limestone, and sand (References 403, 397, and 349). The shell beds are variably sandy and slightly indurated to unindurated. The sandy limestones were deposited under both freshwater and marine conditions. The sand present in the Fort Thompson Formation is fine- to medium-grained quartz sand with abundant mollusk shells and minor but variable clay content (Reference 349). Descriptions of core indicate that the Fort Thompson Formation is a vuggy, solution-riddled, well to poorly indurated, dense to friable limestone. Numerous vertical features in the formation are characteristic of shallow solution pipes or vugs. The features commonly penetrate through more than one horizon and may be conduits for vertical water flow through the formation (Reference 403).

The depositional environment of the Fort Thompson Formation can be related to late Quaternary sea level fluctuations (References 397 and 400). This formation is composed of a group of high-frequency depositional cycles within a progradational environment building on the Tamiami clastic ramp (Reference 404). According to Cunningham et al. (Reference 405), the depositional environments for the Fort Thompson Formation include (a) platform margin to outer platform, (b) open marine, restricted, and brackish platform interiors, and (c) freshwater terrestrial. The Fort Thompson Formation covers the greatest geographical area of all Quaternary formations in southern Florida (Reference 397). The thickness of the Fort Thompson Formation varies from approximately 40 to 80 feet in Miami-Dade, Broward, and Palm Beach counties, where it constitutes the highly productive zone of the Biscayne aquifer (References 400 and 397) (Subsection 2.4.12). In southern Florida the thickness of the Fort Thompson Formation ranges from approximately 50 to 100 feet (References 403 and 398).

### Pleistocene Stratigraphy of the Florida Peninsula

During the Pleistocene, glaciation and fluctuating sea levels occurred worldwide (Figure 2.5.1-212) (Subsection 2.5.1.1.1.1.1.1). Growth of continental glaciers resulted in a drop in sea level as water was retained in the ice sheets. As a result, Florida's land area increased significantly (Figure 2.5.1-219). Based on sea levels during peak glacial periods, Florida's Gulf of Mexico coastline was probably situated some 100 miles (161 kilometers) west of its current position. Warmer interglacial intervals resulted in the glacial melting and a rise in sea level that flooded Florida's land area. At the peak interglacial intervals, sea level stood approximately 100 feet (30 meters) above the current sea level (Reference 287). During this time wave action and currents eroded the existing landforms that became filled with quartz sands originating from the erosion of the Appalachian Mountains and other upland areas. Due to a rise in sea level during the Pleistocene, nutrient rich waters flooded the southern portion of the Florida Peninsula and broken shell fragments along with chemically precipitated particles became the main source of carbonate sediments (Reference 287).

The Pleistocene Anastasia Formation overlies the Pliocene-Pleistocene shell-bearing formations and transitions into the contemporaneous Key Largo Limestone and Miami Limestone (Figure 2.5.1-231). The Anastasia Formation is composed of interbedded sands and coquinoid limestones. The most recognized facies of the Anastasia Formation sediments is an orange-brown, unindurated to moderately indurated coquina of whole and fragmented mollusk shells in a matrix of sand commonly cemented by sparry calcite. Sands occur as light gray to tan and orange-brown, unconsolidated to moderately indurated, unfossiliferous to very fossiliferous beds. The Anastasia Formation forms part of the surficial aquifer system (Reference 377) (Subsection 2.4.12).

The Anastasia Formation includes the coquina, sand, sandy limestone, and shelly marl of Pleistocene age that lies along both the east and west coasts of Florida (Figure 2.5.1-231). The typical coquina of the Anastasia Formation in the type locality does not occur in the western part of southern Florida. Sand, shell beds, marl, and calcareous sandstone are the most common materials. In southern Florida, molluscan faunas establish a Pleistocene age for the Anastasia Formation (Reference 397).

The Atlantic Coastal Ridge (Figure 2.5.1-217) is underlain by the Anastasia Formation from St. Johns County southward to Palm Beach County. The Anastasia Formation generally is recognized near the coast but extends inland as much as 20 miles (32 kilometers) in St. Lucie and Martin counties. To the south of Palm Beach County, the Anastasia Formation grades laterally into the Miami Limestone and is not present in southern Miami-Dade County (Reference 377). Thin marine sandstones of the Anastasia Formation are also present along the southwest coast and extend as a tongue into Collier and Hendry counties (Reference 397). The thickness of the Anastasia Formation varies up to a maximum of 140 feet in southern Florida (Reference 398).

The Pleistocene Key Largo Limestone overlies the Pliocene-Pleistocene shell-bearing sediments and transitions into the contemporaneous Anastasia Formation and Miami Limestone

(Figure 2.5.1-231). The Key Largo Limestone is a white to light gray, moderately to well indurated, fossiliferous, coralline marine limestone composed of coral heads encased in a calcarenitic matrix (Reference 377). Some of these corals have been partially dissolved by groundwater, and the spaces remaining have been filled with crystalline calcite (Reference 392). Little to no siliciclastic sediment is found in these sediments. Fossils present include corals, mollusks, and bryozoans. The Key Largo Limestone is highly porous and permeable and is part of the Biscayne aquifer of the surficial aquifer system (Reference 377).

The Key Largo Limestone is a fossil coral reef that is believed to have formed in a complex of shallow-water shelf-margin reefs and associated deposits along a topographic break during the last interglacial period (Reference 406). The Key Largo Limestone is exposed at the surface in the Florida Keys from Soldier Key on the northeast to Newfound Harbor Key near Big Pine Key on the southwest and from Big Pine Key to the mainland. On the mainland and in the southern Florida Keys from Big Pine Key to the Marquesas Keys, the Key Largo Limestone is replaced by the Miami Limestone (Reference 377). The thickness of the Key Largo Limestone varies widely and is more than 180 feet in southern Florida (Reference 406).

The Pleistocene Miami Limestone overlies the Pliocene-Pleistocene shell-bearing sediments and transitions into the contemporaneous Key Largo Limestone and Anastasia Formation (Figure 2.5.1-231). The Miami Limestone (formerly the Miami Oolite) is a Pleistocene marine limestone. Johnson (Reference 407) has identified six lithofacies in the Miami Limestone: ooid calcarenite, oomoldic-recrystallized, calcirudite, breccia, sandy, and microsparry-coralline. The oolitic facies is the most common and consists of white to orange gray, oolitic limestone with scattered concentrations of fossils. Fossils present include mollusks, bryozoans, and corals; molds and casts of fossils are common (Reference 392).

The Miami Limestone occurs at or near the surface in southeastern peninsular Florida from Palm Beach County to Miami-Dade and Monroe counties. It forms the Atlantic Coastal Ridge and extends beneath the Everglades (Figure 2.5.1-217) where it is commonly covered by thin sediment. The Miami Limestone occurs on the mainland and in the southern Florida Keys from Big Pine Key to the Marquesas Keys. From Big Pine Key to the mainland, the Miami Limestone is replaced by the Key Largo Limestone. To the north, in Palm Beach County, the Miami Limestone grades laterally northward into the Anastasia Formation (Reference 377). The depositional environment of the Miami Limestone can be related to late Quaternary sea level fluctuations (Reference 400). This formation is composed of a group of high-frequency depositional cycles within an aggradational environment (Reference 404). According to Cunningham et al. (Reference 405), the depositional environments for the Miami Limestone include both open marine platform interior and freshwater terrestrial. The highly porous and permeable Miami Limestone forms much of the Biscayne aquifer of the surficial aquifer system (Reference 377) (Subsection 2.4.12). The thickness of the Miami Limestone varies from 10 to 40 feet in southeastern Florida (References 406, 398, and 766). Undifferentiated Quaternary sediments overlie the Pliocene-Pleistocene shell-bearing sediments and the Pleistocene Anastasia Formation, Key Largo Limestone, and Miami Limestone. These undifferentiated sediments consist of siliciclastics, organics, and freshwater carbonates that vary in thickness. The siliciclastics are light gray, tan, brown to black, unconsolidated to poorly consolidated, clean to clayey, silty, unfossiliferous, variably organic-bearing sands to blue green to olive green, poorly to moderately consolidated, sandy, silty clays. Organics occur as plant debris, roots, disseminated organic matrix, and beds of peat. Freshwater carbonates, often referred to as “marls” are scattered over much of the region. In southern Florida, freshwater carbonates are nearly ubiquitous in the Everglades. These sediments are buff colored to tan, unconsolidated to poorly consolidated, fossiliferous carbonate muds. Sand, silt, and clay may be present in limited quantities. These carbonates often contain organics. The dominant fossils in the freshwater carbonates are mollusks (Reference 377).

Where these sediments exceed 20 feet in thickness, Scott (Reference 377) maps them as discrete units. Those sediments occurring in flood plains are termed alluvial and flood plain deposits.

Sediments exhibiting the surficial expression of beach ridges and dunes are shown separately. Terrace sands are not identified individually. The subdivisions of the undifferentiated Quaternary sediments are not lithostratigraphic units but are used to facilitate a better understanding of the geology (Reference 377).

### Holocene Stratigraphy of the Florida Peninsula

Much of Florida is covered by a blanket of Pliocene to Quaternary undifferentiated siliciclastic sediments that range in thickness from less than 1 foot (<0.3 meter) to greater than 100 feet (30 meters). The Holocene sediments in Florida occur near the present coastline at elevations generally less than 5 feet (1.5 meters). These sediments include quartz sands, carbonate sands, and muds with organic materials (Reference 377).

Because of the scouring effect of hurricanes in southern Florida (References 756, 865, and 866), Holocene sediment sequences are preserved only in protected depositional environments. Much of the recent work on these deposits has focused on low energy, low relief areas sheltered by barrier islands, such as the mangrove-capped oyster bars that separate Florida Bay from open marine influences (Reference 755). The following description of Holocene stratigraphy of southern Florida, indicating a general history of sea-level transgression, regression, transgression during the Holocene (References 749 and 757), is based on: deposits preserved in Blackwater Bay on the southwest Gulf coast of Florida (Reference 750); deposits preserved in Sarasota Bay and Little Sarasota Bay on the west-central Gulf coast of Florida (Reference 753); deposits preserved in Whitewater Bay near Cape Sable, on the southern tip of Florida (Reference 800); and the hurricane-disrupted deposits of Biscayne Bay, on the southeastern coast of Florida (Reference 754).

Based on six core samples retrieved from Blackwater Bay, Lowrey (Reference 750) notes that this portion of the southwest Florida shoreline has experienced three major phases of relative sea-level change during the Holocene eustatic rise. Using vibracore samples, Lowrey (Reference 750) developed a stratigraphic sequence that is consistent across the entire bay. Pliocene limestone bedrock (described in References 751 and 752), at the base of cores 6 and 1, is overlain by units A (oldest) to D (youngest). These units were classified as sediment type A (quartz packstone or a clayey quartz sand), sediment type B (quartz grainstone), sediment type C (*Rhizophora*, red mangrove, peat), and sediment type D (shelly quartz packstone to wackestone). The base of the peat in core 1 was dated at 4170 + 40 years before present using radiocarbon techniques, and the upper surface of the peat in core 6 was dated at 1090 + 40 years before present.

Vertical and lateral relationships of units A to D in the cores suggest that Blackwater Bay has undergone three phases of local sea-level change during the eustatic Holocene transgression. Each sedimentary sequence represents a time transgressive unit, as changes in sea level caused migration of depositional environments. Sediment types A and B formed during the early transgressive phase, as interpreted by Parkinson (References 751 and 752), as shoreline approached the study site. The occurrence of sediment type C represents the shoreline intersection with the site, followed by a stabilization and possible regression of the shoreline at approximately 4100 years before present with the accumulation of thick peat sequences. The facies change to sediment type D at a uniform elevation indicates a significant event at approximately 1000 to 1090 years before present, possibly a storm or series of storms, inundated the mangroves in all cores, reinitiating a relative sea-level rise and a return to deeper water conditions.

Davis et al. (Reference 753) conducted studies of the Holocene stratigraphy of Sarasota Bay and Little Sarasota Bay, coastal bays located landward of a Holocene barrier/inlet complex on the west-central, microtidal Gulf Coast of Florida. In addition to evidence for cyclic sea-level change, the sand and shell gravel deposits sampled in cores from both bays were deposited by at least four storms. Three storm units from Sarasota Bay have been radiocarbon dated at 2270, 1320, and 240 years before present. Historically documented severe hurricanes influenced this coast in 1848 and

1921. Hurricanes interrupted the normal, low energy, slow deposition in the bays and caused inlets to open and close (Reference 753).

Vlaswinkel and Wanless (Reference 800) find that, under conditions of sea-level rise, natural and cut tidal channels contribute to a larger tidal flow and thus bring increased volumes of sediment-laden tidal water into estuaries and coastal lakes (e.g., Lake Ingraham and adjacent southern lakes). Rapidly forming flood tidal mud deltas are filling these lakes and bays (e.g., Whitewater Bay near Cape Sable) at rates of 1 to 20 centimeters per year. Organic content of these carbonate sediments is up to 40 percent (in contrast to 2 to 10 percent in most of the Florida Bay mud banks). This rapid pulse of coastal sedimentation in response to small sea-level changes and coastal instability may be more common in building a stratigraphic record than presently appreciated (References 756 and 800).

According to Wanless et al. (Reference 756), the Pleistocene and Holocene coastal dune ridges found around the Gulf of Mexico and Atlantic Ocean coasts of Florida are stabilized mostly by vegetation and do not appear to be producing layered sequences because the sand, as it gradually accumulates, is bioturbated by root processes. Some of these ridges are formed during and just following major storm events as large volumes of sediment are scoured and recycled. Wanless et al. (Reference 756) identify rapid pulses of growth during times of rising sea level as large volumes of coastal and shelf sediment became exposed and unstable. Waves and currents rapidly erode and deposit these sediments inland from the coast producing pulses of dune-ridge growth. These thickly layered sequences are then followed by a time of vegetative stabilization, bioturbation of the upper portion, and minor trapping of sand that is blown or washed in. This process is occurring along sections of the southwest Florida coast today. Probably with the help of the 23-centimeter relative rise of sea level during the past 70 years, the marl (firm carbonate mud) and organic peat of the southwest coast of Florida is rapidly eroding (200 to 400 meters since the earliest 1928 aerial photographs) and large volumes of sediment are being redistributed (Reference 756).

Hurricanes complicate the preservation of Pleistocene and Holocene deposits on the east and west coasts of the Florida Peninsula by eroding these deposits and redepositing them elsewhere. As an example, Hurricane Andrew impacted the shallow marine environments of south Florida in August, 1992 (References 754, 865, and 866). Tedesco and Wanless (Reference 754) maintained an extensive set of pre-storm data and monitor a broad spectrum of environments on both Florida coasts since immediately after the storm. They report that the most pronounced long-term change occurred on the high energy shallow marine carbonate banks forming the seaward margin of Biscayne Bay. These banks experienced accelerated surge currents at areas of shoaling or confinement. Seagrass blowouts covered a broad expanse of the seaward bank margin and up to 1 meter of initial erosion of muddy substrates resulted. A bankward thinning wedge of skeletal sand and gravel, which originated from erosional areas, was deposited on the banks. Destabilized areas exposed to lower energy storm events have continued to be reworked. The overwash lobes of skeletal sand and gravel have prograded bankward more than 60 meters (200 feet). Sediment for lobe progradation came initially from seaward erosional areas, but now originates from portions of the overwash lobe itself. Skeletal sand and gravel reflecting the initial storm deposit has been eroded and reincorporated into an evolving, migrating sediment wedge up to 15 centimeters (6 inches) thick. New fauna has been incorporated into the deposit. This beach-building process appears to be continuing (Reference 756).

#### 2.5.1.1.1.2.1.2 Stratigraphy of the Florida Platform

The Florida Platform is a broad low-relief marine platform ranging in elevation from -656 feet to the shoreline (-200 to 0 meters). It includes the shallow portion of the continental shelf currently underwater, stretching from the Florida Escarpment to Florida's Gulf coast, roughly 300 miles (480 kilometers) across from west to east (Figure 2.5.1-214). Geophysical data indicate that the platform is underlain by continental crust at the axis of the peninsula to thinned continental crust at its

periphery. The crust beneath southern Florida may be more mafic or transitional (Reference 409). The structure and tectonic evolution of the Florida Platform are described in detail as part of the larger Florida Platform, including the modern peninsular Florida and its surrounding areas of continental shelf and slope (Subsection 2.5.1.1.3), which are located immediately west of the Florida Platform. The Florida Escarpment represents the transition to the Gulf of Mexico, the deep-water basin that opened in the middle Jurassic (Reference 410).

The structural relationships between the large, separated areas of carbonate platform in the Gulf of Mexico, the Bahamas, the Blake Plateau, and elsewhere in the Caribbean are not clear. The following discussion is pertinent to the continuity of basement and overlying stratigraphy between these carbonate platforms, as described in Subsections 2.5.1.1.1.2.1.2, 2.5.1.1.1.2.2, 2.5.1.1.2.1.2, and 2.5.1.1.2.1.3. Two different hypotheses have been proposed for the character and continuity of the basement rocks between the Florida and Bahama Platform. These hypotheses probably also apply to the relationship between those two platform areas and the Yucatan Platform. Mullins and Lynts (Reference 411) postulate that the Bahama Bank formed during the Jurassic on top of a rift-generated horst-and-graben topography (known as the “graben hypothesis”). According to this hypothesis, the seaways now separating the banks originally formed as structural lows and that the Florida-Bahamas megabank was situated on structural highs. During long-term subsidence following initial rifting, carbonate-derived sedimentation kept pace across the megabank topographic highs, forming up to 14 kilometers (9 miles) of shallow-water limestones. The basins also accumulated great thicknesses of both shallow and deep-water carbonate sediments but lagged behind the banks, thereby amplifying the original depositional relief.

An alternate hypothesis is proposed by Sheridan et al. (Reference 307), who postulate that a large, continuous megabank, extending from the Florida Escarpment to the Blake-Bahamas Escarpment, had formed by the Late Jurassic on a basement not segmented by horsts and grabens (Sheridan et al.'s hypothesis is known as the “megabank hypothesis”). The continuous platform may have had deep-water reentrants (Reference 501), but most of the area from the Florida Platform to the Blake-Bahamas Escarpment was covered by shallow-water carbonate depositional environments that persisted until the mid-Cretaceous. Deep sea drilling has confirmed that, prior to the mid-Cretaceous, shallow-water limestones were deposited in the Straits of Florida, in contradiction to the graben hypothesis.

Beginning in the mid-Late Cretaceous, the Cuban and Antillean orogenies produced left-lateral shearing between the North America and Caribbean plates. Faults and folds developed, preferentially aligned with the margins of the carbonate banks, including the eastern margin of the Florida Platform. As a result, the megabank broke up into a number of banks and basins in the Florida-Bahamas region. On the basis of undeformed sediments visible on seismic reflection surveys, Hine (Reference 309) concludes that the eastern margin of the Florida Platform detached from the Bahama Platform and the Straits of Florida formed during the mid-Late Cretaceous. Hines (Reference 309) notes the continuity of flat-lying, shallow water limestones across the Florida to Bahama Platforms in the mid-Late Cretaceous, indicating the initiation of Florida Current activity at a mid-Late Cretaceous (Coniacian) unconformity in what would become the Straits of Florida. Later (post-Coniacian) strata are thinner in the Straits of Florida than on the west Florida and Bahama Platforms. It is likely that the Straits of Florida (and perhaps other channels across the Bahama Platform) are the result of new current circulation patterns that may have been caused by tectonic events such as the emergence of the Isthmus of Panama (as discussed in Subsection 2.5.1.1).

Persistent structural controls on carbonate sedimentation across the larger Florida Platform are discussed in Subsection 2.5.1.1.1.3.2.1. Stratigraphic relationships indicate, for example, that the Peninsular Arch has been a structural high since Late Jurassic time and possibly as early as mid-Paleozoic time, while the Sarasota Arch has been a structural high since late Paleozoic time and possibly as long as early Paleozoic (Reference 413). According to Winston (Reference 413), the Florida arches were formed not by uplift, but by subsiding more slowly than contiguous basin areas.

The depositional basins identified across the Florida Peninsula and Florida Platform have been structural lows since the Late Cretaceous (Coniacian) time and may be downwarping even today (Reference 413). This differential subsidence has limited the continuity of carbonate and evaporite deposition since the Late Jurassic. These structural controls have been the main determinant as to the lateral continuity and thickness of the shallow carbonate and evaporate units across peninsular Florida and the Florida Platform.

The Middle Jurassic through Paleogene carbonate strata of the Florida Platform is essentially the same, albeit thicker, as the units described for the Florida Peninsula. As the Pangea supercontinent began to break apart in the Early Jurassic (about 175 Ma), the rocks along the rifted margins were faulted into large blocks that began to subside. The resulting basins began to fill with sediments eroded from the blocks and from the adjacent continents. As these basins subsided below sea level, they were invaded by seawater whose restricted circulation and high evaporation rates caused deposition of thick evaporitic deposits in some areas (References 414, 415, 416, 417, 418, and 886).

The oldest Cenozoic formation on the Florida Platform is the Paleocene Cedar Keys Formation that conformably overlies the Late Cretaceous Pine Key Formation (Figure 2.5.1-231) (References 357 and 369). The Cedar Keys Formation is a lagoonal facies that occurs within the confines of the Rebecca Shoal barrier reef (Reference 369). In southern Florida, the Cedar Keys Formation consists primarily of gray dolomite, gypsum, and anhydrite with a minor percentage of limestone. The upper part of the Cedar Keys Formation consists of coarsely crystalline, porous dolomite. The lower part of the Cedar Keys Formation contains more finely crystalline dolomite interbedded with anhydrite (Reference 369). Based on structural/stratigraphic analyses of borehole data, the configuration of the Paleocene sediments in peninsular Florida reflects depositional controls inherited from preexisting Mesozoic structures such as the Peninsular Arch and the southern Florida Basin (References 389 and 419). The upper unit of porous dolomite in the Cedar Keys Formation forms the base of the Floridan aquifer system (Subsection 2.4.12) throughout southern Florida (Reference 349), where it is found at elevations ranging from -3,000 to -4,000 feet (~-914 to -1220 meters) (Reference 389). The Cedar Keys Formation varies from approximately 500 feet up to 2000 feet (~152 to 610 meters) thick in southern Florida (References 356 and 375).

#### 2.5.1.1.1.2.1.3 Stratigraphy of the Atlantic Offshore Continental Shelf and Slope

The southern portion of the Atlantic margin from Florida and the Bahamas northward to the Newfoundland Fracture Zone represents a fully developed, passive margin with a sedimentary record spanning the mid-Jurassic to Recent (Reference 327). The strata rest on Triassic-Early Jurassic rift basins. The Atlantic margin includes a generally broad continental shelf underlain by extended continental crust transitioning to thick oceanic crust, including the Blake Plateau and the Bahama Platform. The basins of the Atlantic margin display the two-phase architecture characteristic of extension (passive) continental margins. Rift basins, formed by the brittle failure during the initial phase of crustal stretching, are followed by a broad seaward-thickening sediment wedge deposited during the phase of flexural subsidence that accompanies regional cooling and subsidence as the rifted margins move away from the oceanic spreading center. The sediment piles characteristically onlap the continental margin over periods of tens of millions of years as the thinned continental crust gradually subsides (Reference 327).

Rifting was accompanied by the extensive volcanic activity of the central Atlantic magmatic province or CAMP (Subsection 2.5.1.1). Basaltic dikes, sills, and flows formed during a 25 m.y. period spanning the Triassic-Jurassic boundary over vast areas of the Pangean suture between the proto-North America, South America, and African cratons. The rift basins include the South Georgia Rift, the Suwannee Basin, and others buried beneath Coastal Plain and continental shelf sedimentary cover (Reference 421) (Figure 2.5.1-229). Further subsidence of the rifted margin included development of shallow seas and the deposition of extensive evaporites, present under the South

Georgia Basin, the Carolina Trough, the Blake Plateau Basin, and the Bahama Platform (Reference 327).

The transition from rifting to drifting in the middle Atlantic margin from northern Florida to Newfoundland began in the Middle Jurassic (Reference 421). Near the Blake Plateau, the transition from rifting to drifting appears to have occurred slightly later in the Middle Jurassic (e.g., Reference 341). The sedimentary succession in the Atlantic marginal basins is about 9 kilometers (5 miles) thick and consists of a variegated clastic succession of conglomerate, felsic and lithic arenite, siltstone, shale and mudstone, with interbedded basaltic lava flows. Evaporites, eolian sands, coal, and kerogen-rich beds are locally important. Siliceous tufas formed locally from hydrothermal systems associated with the lava fields. Fossil remains include fish, algae, zooplankton, spores, and pollen; the organic remains occurring in sufficient abundance in some cases to quality the fine-grained deposits as oil shales. Varved deposits attest to cyclic climatic conditions (Reference 327).

Sedimentary cover across Florida's Atlantic Continental Shelf and Slope began in post-Albian (post-Early Cretaceous) time when the area became a marine province. Siliciclastic sediments shed from the southern Appalachians, moved by rivers running north-south across the peninsula, accumulated on the level carbonate platform of Florida's Atlantic margin. Regional unconformities at the tops of the Albian, Santonian, Maastrichtian, Paleocene, and Oligocene units as well as one between Turonian and the Santonian have been mapped. Two styles of sedimentary accumulation have been active: (a) platform upbuilding and (b) platform outbuilding or progradation of the shelf. In post-Albian time, the area became a marine province, and sediment accumulated on a level platform. During the Santonian-Coniacian a shelf prograded seaward across this platform, but during the Campanian, Maastrichtian, and Paleocene deposition on a level plateau resumed (Reference 234).

The persistence of Gulf Stream erosion during the Cenozoic, in conjunction with crustal subsidence, transformed the distal edge of this continental margin sector into a deep-water, sediment-starved environment, the Blake Plateau (Reference 234). The Blake Plateau comprises an 8- to 12-kilometer (5- to 7.5-mile) thick sequence of Jurassic and lower Cretaceous limestones that are capped by less than 1 kilometer (0.6 mile) of Upper Cretaceous and Cenozoic deposits (Reference 422). The limestone platform extends beneath the emergent Florida Peninsula and continues west beneath the Florida Platform. The carbonates apparently also extend, uninterrupted, beneath the Bahama and Yucatan Platforms.

The Continental Offshore Stratigraphic Test Well (COST GE-1) was drilled in the center of the Southeast Georgia Embayment and penetrated more than 4 kilometers (2.5 miles) of marine and continental sedimentary strata, terminating in Paleozoic metamorphic rocks (Reference 423). Based on well data and seismic reflection profile data, Paull and Dillon (Reference 487) summarize the stratigraphy of the Atlantic Continental Shelf and Slope across the Florida-Hatteras Shelf and the Blake Plateau. Lower Cretaceous sediments were continental at the end of the Cretaceous, the entire area was a broad, level, submerged carbonate platform. The Mesozoic-Cenozoic boundary is marked by a small but not particularly distinct unconformity. A sequence of Paleocene strata about 100 meters (330 feet) thick overlies the Cretaceous units. The top of the Paleocene section is irregularly eroded and has relief of as much as 100 meters. The erosion is related to the initiation of the Gulf Stream. The late Paleocene unconformity is buried by a large seaward progradational wedge of Eocene to Oligocene age. The Eocene sections consist primarily of limestone and marl that grade northward into sandy limestone. Progradation was terminated by erosion at the end of the Oligocene. The Late Oligocene erosional surface was buried by another progradation of shelf and slope during Miocene to Holocene time. Tertiary accumulations under the shelf are much thicker than on the Blake Plateau.

#### 2.5.1.1.1.2.2 Stratigraphy of the Bahama Platform

The stratigraphy of the Bahama Platform is based on large quantities of seismic reflection data as well as core recovered from numerous DSDP and ODP drilling sites (e.g., Reference 787) (Figure 2.5.1-211) (Table 2.5.1-201). A limited number of deep drill holes are located on the Bahama Platform. Petroleum exploration holes reached depths of 5700 meters (16,400 feet), bottoming in the Upper Jurassic carbonates and evaporites. DSDP and ODP holes reached depths greater than 3000 meters (9800 feet).

The northwestern portion of the Bahama Platform is a passive continental margin that was not significantly affected by the tectonism that sutured Cuba and the Yucatan Basin during the Late Cretaceous though Eocene (Reference 220). Based on numerous reprocessed seismic surveys (Reference 424), the basement of the carbonate platforms of the Atlantic margin and Gulf of Mexico comprises highly rifted transitional continental crust consisting of probable Paleozoic crystalline metamorphics overlain by late Paleozoic sediments, intruded by Early to Middle Jurassic volcanics. The surface topography of the acoustic basement, as deduced from seismic reflection data, indicates a landscape of ancient horsts and grabens. The acoustic basement beneath the Bahama Platform is estimated to be between 7.5 to 8.7 miles (12 to 14 kilometers) (Reference 425) thick and is overlain by laterally continuous sedimentary units with large impedance variations, such as alternations of volcaniclastics, evaporites, and limestones (Reference 426) (Figure 2.5.1-244). The pre-rift sediments, where present, have been rotated with the underlying fault blocks. The sediments are often missing on high-standing fault blocks.

According to Case et al. (Reference 427), the Florida and Bahama Platforms are one continuous tectonic entity, that is, there is no single tectonic discontinuity or boundary identified that separates the two. The basement beneath the Florida Platform and the relationship between the Florida and Bahama Platforms (and possibly also the Yucatan Platform) are described in Subsection 2.5.1.1.1.2.1.2. The Bahama Platform is built upon the same rifted fragments of continental or transitional crust as the Florida Platform. However, gravity and magnetic data indicate that the crust of the Bahamas is more variable, with the southeastern portion of the platform beyond the site region potentially consisting of thick oceanic crust that was formed during Jurassic volcanism (Reference 428). The Middle Jurassic was a period of widespread production of the thickened oceanic crust, some of which floors the Caribbean basins and surrounding regions (Reference 250). The thickened oceanic crust is believed to be related to anomalously high heat flow due to Pangean rifting, related to emplacement of the ECMIP in Subsection 2.5.1.1.

Mesozoic rifting of the Atlantic Ocean and the Gulf of Mexico resulted in modification of continental crust along both margins and the creation of new oceanic crust farther offshore. The crust beneath the Bahama Platform (and within the site region) is characterized by several studies as having similar characteristics to the crustal types within the Gulf of Mexico. The Gulf of Mexico and the site region are also similarly characterized as seismically quiescent. Studies that characterize similar crust between these two regions include:

- Sawyer et al. (Reference 410) divide basement rock in the Gulf of Mexico region into four main types on the basis of the manner in which crust was created or modified by Mesozoic rifting: oceanic, thin transitional, thick transitional, and continental crust (Figure 2.5.1-238).
- Ewing (Reference 430) includes Florida as part of the Gulf of Mexico.
- Crustal-scale cross sections by Salvador (References 368 and 839) depict thick transitional crust beneath the Florida Peninsula and Shelf (Figures 2.5.1-239, 2.5.1-240, 2.5.1-241, and 2.5.1-242).

- The Phase 1 and 2 earthquake catalogs (Subsections 2.5.2.1.2 and 2.5.1.2.3, respectively) indicate sparse seismicity throughout the Florida and the Bahama Platform, in contrast with the island arc terranes of Cuba that show abundant seismicity (Figure 2.5.2-201).

The deepest wells in the Bahamas have encountered a basement of arkosic rhyolitic volcaniclastic deposits overlain by Upper Jurassic limestones, dolomites, and evaporites (Reference 307) (Figure 2.5.1-243). The overlying stratigraphic section is more than 3 miles (5 kilometers) thick and indicates that shallow shelf and platform carbonate deposition continued essentially uninterrupted to the present time and was primarily controlled by eustatic sea level changes (Reference 211) (Figure 2.5.1-208). The Great Isaac I Well (Figure 2.5.1-243) reached volcaniclastic sediments beneath the Jurassic carbonates. Sheridan et al. (Reference 307) suggest that the larger, western platform near Andros Island and Grand Bahama is underlain by transitional continental-oceanic crust formed during an aborted rifting phase in the Mid-Jurassic. Later seismic reflection survey results show that the area is underlain by anomalously thick oceanic crust that extends from the Georges Bank to the tip of Florida (see discussion of the ECMIP in Subsection 2.5.1.1).

The Great Isaac I well, located southwest of the Little Bahama Bank, penetrated 7000 feet (2100 meters) into the carbonates without reaching crystalline basement (Reference 432) (Figure 2.5.1-243). However, the core revealed shallow-water carbonate-evaporite deposits (mid-Cretaceous and older) overlain by deep-water deposits with upward-increasing neritic debris (Late Cretaceous-Tertiary) and capped by bank-margin deposits (Plio-Pleistocene). This succession is interpreted by Schlager et al. (Reference 432) as a restricted carbonate platform that was drowned in the late Albian or Cenomanian (~100 m.y.) and subsided to over 2000 feet (600 meters) water depth (see discussion of carbonate platforms in Subsection 2.5.1.1.1.2.1.2). The drowned carbonate platform was subsequently reintegrated into the Great Bahama Bank by westward progradation of the platform (Reference 432) and tectonic uplift (Reference 327). The tectonism was associated with the collision of the Greater Antilles Arc with the North America Plate, starting in the mid-Late Cretaceous and continuing through Eocene.

Multichannel seismic line MC92 was run across the northern Straits of Florida to tie to the Key Largo well, KL, and the Great Isaac Island well, GI-1. The seismic-stratigraphic evidence seen in line MC92 indicates that the Straits of Florida first began to develop as a deepwater area during the Cenomanian (lower Upper Cretaceous). Before this, Albian (upper Lower Cretaceous) and older sedimentary units were deposited on a shallow-water bank, which was continuous from southern Florida to the Great Bahama Bank (References 307 and 424).

Sheridan et al. (References 307 and 424) interpreted the seismic reflectors at the top of the upper Oligocene-Holocene (HOLO.-UP. OLIG.) as an asymmetric ridge. However, according to Eberli et al. (Reference 983), this feature may also be interpreted as a possible clinoform. As defined by Miall (Reference 985), a clinoform is a sloping dipping surface that is commonly associated with strata prograding into deepwater. SEPM (Reference 986) describe sigmoid clinoforms (s-shaped reflection patterns) to be interpreted as strata with thin, gently dipping upper and lower segments, and thicker, more steeply dipping middle segments. Twenty degrees is the angle of repose for carbonate sediments (Reference 983). In general, sigmoid clinoforms tend to have low depositional dips or angles for the upper segments, typically less than 1 degree, and are parallel with the upper surface of the facies unit with no strata termination with the bounding surfaces (Reference 986). According to Eberli et al. (Reference 983), the sigmoid clinoforms that formed in the northern Straits of Florida nearly match the third order sea level fluctuations on the global cycle chart. SEPM (Reference 987) defines a third order sea level fluctuation or sequence as a depositional sequence that has a duration in the order of 1 million to 10 million years with a relative sea level amplitude of 50 to 100 meters (164 to 328 feet) and a relative sea level rise/fall rate of 1 to 10 centimeters (0.4 to 3.9 inches) per 1000 years (Reference 987).

Bergman (Reference 906) and Anselmetti et al. (Reference 228) interpret the progradation of sediments that form the clinoforms in the Straits of Florida to be caused by a sea level drop. The drop in sea level occurred approximately during the middle Miocene. Eberli et al. (Reference 983) interpret the data and the results of modeling studies to indicate that progradation occurred in pulses during rises in sea level subsequent to drops in sea level. More sediment is produced on the bank surface than can be accommodated and, thus, excess sediment is transported down the leeward slope and deposited as apron sediments and turbidites (Reference 983). In addition to eustatic changes in sea level, a western boundary paleo-Florida Current had developed and the Straits of Florida became the major pathway for the Florida-Gulf Stream surface current system by the middle Miocene (References 228 and 906).

Since the apparent slump is interpreted as a progradational depositional feature, it does not bear upon the tsunami hazard in the site region. Based on more recent data presented in Mulder et al. (Reference 984), the turbidite deposits might have resulted from a slope failure on the western margin of the Great Bahamas Bank. For Probable Maximum Tsunami purposes, a potential landslide-induced tsunami is discussed in Subsection 2.4.6.

Lower Cretaceous stratigraphy includes dolomite and layers of anhydrite. The upper Cretaceous sedimentary sequence is predominantly shallow water limestones, but deepwater oozes, chalks, and cherts of Late Cretaceous to Tertiary age occur in the Providence Channel (Figure 2.5.1-208). The finding of deep-water sediments in the Providence Channel indicates that deep-water channels and troughs of the Bahama Platform were in existence at that time (Reference 307).

A review of paleogeography developed by Salvador (Reference 368) from a combination of seismic profiles and drilling data indicates that shallow shelf and platform carbonate deposition on the Bahama Platform continued essentially uninterrupted from the Early Cretaceous until the present. The carbonate shelves and platforms were often fringed along their basinal margins by high-energy shoals, rudist-dominated reefal buildups, and barrier islands interrupted by tidal channels and passes. Banks, patch reefs, and occasional evaporites were often formed in intra-shelf basins and back-reef lagoons. Unlike Florida, there was no deposition of siliciclastics to interrupt carbonate deposition (References 211 and 368).

Based on their similar depositional environments and marine geomorphology, Hoffmeister et al. (Reference 384) state that the geologic features of Florida and the Bahama Platforms are mirror images of each other. Mesozoic sediments are largely shallow water carbonates with some evaporites.

Drilling at DSDP Site 627 (Figure 2.5.1-211) recovered mid-Cretaceous (Albian) through Quaternary sediments. A short section of Upper Pleistocene (160 centimeters [5.3 feet] in Core 627A-1H) is separated from the lower Pleistocene by an unconformity. The underlying section suggests relatively continuous sedimentation from late Miocene through early Pleistocene time. A possible hiatus separates this section from the underlying uppermost lower Miocene to middle Miocene. A substantial unconformity separates Neogene sediments from siliceous sediments deposited during the early to middle Eocene. Below this lies a thin section of Paleocene sediments. The range of ages and the thinness of the section suggest that sedimentation during this time may have been either punctuated or condensed (Reference 436).

The Cenozoic section is separated from the Mesozoic section by a substantial hiatus that includes the earliest Paleocene and all of the Late Cretaceous (Maastrichtian). The underlying upper Campanian section consists of purely pelagic sediment with open-ocean micro faunas and micro floras. This relatively thick Campanian section is underlain by a sequence of lower to middle Cenomanian hemipelagic marls. The microfaunas indicate that the top of this Cenomanian section represents an outer-shelf environment, whereas the base is an inner shelf. The underlying sequence of shallow-water-platform dolostones and evaporites is Albian and most probably late Albian in age.

These stratigraphic relationships in the mid-Cretaceous suggest that little time elapsed between deposition of the shallow-water-platform sediments and subsequent deposition of the hemipelagic sequence (Reference 436).

Cenozoic sediments are dominated by low magnesium carbonates with varying amounts of aragonite and dolomite. Based on deep drilling and seismic reflection data, the base of Cenozoic sediments ranges in depth from approximately 3200 feet (975 meters) (References 385 and 437) to 8000 feet (2438 meters) (Reference 387). The Great Bahama Bank is considered to be an excellent indicator of sea-level changes in the Atlantic Ocean. Eberli et al. (Reference 385) report that deep core borings on the Great Bahama Bank indicate that the frequency and amplitude of sea-level changes have had a significant effect on the progradation, thickness, and diagenesis of the platform strata (Figures 2.5.1-244 and 2.5.1-245).

A combination of seismic, magnetic, and gravity profiles and drilling data provide information about the Quaternary subsurface stratigraphy of the Bahamas. Carew and Mylroie (Reference 438) draw the following conclusions about the Quaternary stratigraphy (Figure 2.5.1-246):

- A transition from Pliocene skeletal and reefal facies to Quaternary oolites and eolianites is located at the margins of the Great Bahama Bank.
- Shallow coring has indicated that Pleistocene-Holocene sediments are about 79 feet (24 meters) thick on the Little Bahama Bank and as much as 131 feet (40 meters) thick on Great Bahama Bank.
- The thickness of the Quaternary sediments does not vary systematically across the Bahamas.

Pleistocene stratigraphic units exposed on the Bahama Bank include the Owl's Hole, Grotto Beach, and Rice Bay formations (Figure 2.5.1-246). The Pleistocene Owl's Hole Formation consists of eolianite deposits overlain by terra-rosa paleosol that are, in turn, overlain by either a highly oolitic eolianite deposit capped by a second terra-rosa paleosol or by subtidal deposits. The Owl's Hole eolianites consist of fossiliferous and peloidal grainstones and oolitic rocks. The oolitic rocks are micritized at the exposed surface but portions remain weakly cemented. The top of the unit is a hard, red micritic terra-rosa paleosol overlain by younger oolitic eolianites (Reference 438).

The Pleistocene Grotto Beach Formation consists of eolianite and beach-face to subtidal marine limestones subdivided into the French Bay and Cockburn Town Members. The Grotto Beach Formation is capped by a terra-rosa paleosol except where it has been eroded. The formation is characterized by well-developed ooids. The regressive stage and subtidal facies eolianites are predominantly peloidal or bioclastic and also contain ooids. The transgressive stage eolianites, a beach facies, are represented by the French Bay Member, consisting of fine to medium oosparites (oolitic grainstones), while the subtidal and stillstand through regressive stage beach and eolian deposits are represented by the Cockburn Town Member of the Grotto Beach Formation (Reference 438).

The Holocene Rice Bay Formation consists of all the rocks overlying the paleosol that caps the Grotto Beach Formation. In some places it is subdivided into the North Point and Hanna Bay Members (Figure 2.5.1-246). The Rice Bay Formation consists of eolianites and beach facies rocks that have been deposited during the transgressive and stillstand stages of the current sea level highstand. The rocks of the Rice Bay Formation are characterized by a low abundance of ooids, the small size of the ooids, the dominance of peloids and bioclasts, limited diagenetic micritization, and low magnesium calcite cement. The transgressive stage of eolianites is represented by the North Point Member. The rocks are mostly peloidal and cemented (by either water from the vadose zone or by marine water). The stillstand stage beach and eolian facies of the Rice Bay Formation consist of

the Hanna Bay Member. This member consists of peloidal/bioclastic grainstones with low-magnesium calcite cement. Lithification of rocks from this member occurred while at the current sea level (Reference 438).

#### 2.5.1.1.1.2.3 Stratigraphy of Cuba

Cuba comprises several lithostratigraphic components including the following (Figure 2.5.1-247):

- The Socorro Complex, Grenvillian basement rocks in central Cuba
- North American-derived passive margin strata of Jurassic to Eocene age
- Jurassic to Cretaceous metamorphic terranes (the Escambray, Pinos, and Guaniguanico terranes)
- An allochthonous Cretaceous volcanic arc and associated ophiolites and mafic metamorphic rocks
- An in situ Tertiary volcanic arc at the southeastern edge of Cuba
- Undeformed upper Eocene to Recent cover

**The Socorro Complex**

Renne et al. (Reference 689) document the presence of Grenvillian basement rocks in central Cuba (Socorro Complex, 0.904 Ga, Fig. 1) that they propose formed part of a continuous band of Grenville rocks extending between southwestern North America, Central America (Chortis block) and northwestern South America (Reference 442).

**North American-Derived Passive Margin Strata**

The northeastern edge of Cuba, onshore and offshore, includes an approximately 16,000-foot (4900-meter) thick carbonate platform, carbonate slope, and deep-water basin in a northwest-trending exposure (Reference 439) (Figures 2.5.1-247 and 2.5.1-248). These sections range in age from upper Jurassic to lower-Middle Eocene. The platform section is located farther northeast, while the deep-water basin is exposed farthest to the southwest (Figure 2.5.1-248). The platform facies commonly includes an Late Triassic to Early Cretaceous, approximately 6600 to 13,100 feet (approximately 2000 to 4000 meters) thick, siliciclastic-evaporite-carbonate section. The Late Cretaceous platform section is more variable, but mostly includes shallow-water limestone, dolomite, and minor chert (Reference 440). The slope facies is represented by the Camajuani belt, and includes uppermost Jurassic to Late Cretaceous deep-water limestone, chert, and calcareous clastic beds. The Placetas belt, to the southwest, is the most variable and represents a deep-water basin depositional environment. The stratigraphic section includes a Late Jurassic to Late Cretaceous deep-water limestone, chert, and clastics overlain by a Late Cretaceous calcareous megaturbidite, overlain by deep-water limestones (Reference 440). A comparison with wells drilled in Florida and in the Bahamas indicates that the Cuban passive margin strata are similar to the North American passive margin stratigraphic sequence (Reference 441). In addition, the transition from carbonate platform to slope deposits likely reflects the southern boundary of the Cretaceous Florida-Bahama carbonate platform.

The Late Cretaceous (Maastrichtian) to upper Eocene sections of the passive margin terrane display a characteristic transition from exclusively carbonate to terrigenous clastic deposition. This transition is variable across the island of Cuba and reflects the diachronous approach of the Greater Antilles volcanic arc (Figure 2.5.1-250). The southwestern portions of the passive-margin sequence are relatively more deformed and have the thickest syn-orogenic Paleocene to Eocene foreland basin

deposits. To the northeast, the carbonate platform sections display less deformation and a thin clastic cap (Reference 440).

### Metamorphic Southwestern Terranes

Three main exposures of Jurassic to Cretaceous metasedimentary rocks are collectively known as the southwestern terranes. The southwestern sedimentary terranes are characterized by a thick section of continentally derived clastics of Middle Jurassic age. From west to east these terranes are the Guaniguanico, the Pinos, and the Escambray (Figures 2.5.1-251 and 2.5.1-247). The terranes were originally Laurentian rocks that may have originated from the Yucatan Peninsula as the Greater Antilles Arc pushed eastward past the Yucatan block (References 442, 443, and 444).

In western Cuba the base of the Guaniguanico terrane is the San Cayetano clastic sequence that is as old as Early Jurassic. This unit, consisting of sandstones and conglomerates, was deposited over a rifting basement that is probably composed of continental crust (Reference 439). The provenance of the San Cayetano has been used to determine the paleo-position of Cuba before Early to Middle Jurassic rifting of Pangea. Detrital mica ages indicate that the southern Yucatan is the source for this unit (Reference 442), whereas detrital zircons indicate a northern South American or Yucatan provenance (Reference 445). Hence, this suggests that the Guaniguanico was also allochthonous. The terrane also includes an Late Jurassic thick basalt sequence with major and trace element geochemistry consistent with continental rifting (Reference 443). The overlying Jurassic to -Cretaceous sequence includes shallow and deep-water carbonates. An hiatus in deposition is present in the uppermost Cretaceous to earliest Paleocene before deposition transitions to Paleocene to lower Eocene foreland strata deposited in a piggyback basin. There, late Campanian (Late Cretaceous) strata (77 to 79 Ma) record the first fine-grained input from the approaching Greater Antilles Arc to reach the southern North American margin, indicating that the arc was approaching the Florida and Bahama Platforms at that time. The collision occurred later, however, because syn-orogenic strata of western Cuba contain nanofossils of late Paleocene to early Eocene age (Reference 220).

South of the main island of Cuba, the Pinos terrane is exposed on a small, circular island (Isla de la Juventud) (Figures 2.5.1-247 and 2.5.1-251). This terrane also includes Jurassic-Cretaceous metasiliciclastics, with marbles and amphibolites near the top of the section. The terrane was subjected to Late Cretaceous metamorphism (References 440 and 446).

The Escambray terrane, or massif, is exposed in south-central Cuba, outside the site region (Figures 2.5.1-247 and 2.5.1-251). It consists of two metamorphic domes separated by a Paleogene sedimentary cover. It is composed of Jurassic to Cretaceous age siliciclastic metasedimentary rocks and minor marbles, metabasic rocks, and serpentinites. The Escambray terrane is divided into a lower unit of greenschist grade pelites and carbonates, overlain by a middle unit of blueschist-facies metasediments with ultramafic boudins, which are in turn overlain by an upper unit of metasedimentary sequences with eclogite lenses in a serpentinite matrix. These rocks have $^{40}Ar/^{39}Ar$ and $^{87}Rb/^{86}Sr$ ages that reflect cooling from high temperatures at 70 Ma (Reference 218).

### Cretaceous Volcanic Arc

The Cretaceous volcanic arc unit includes ophiolites, subduction-related metamorphic rocks, early tholeiitic island arc rocks, calc-alkaline volcanic and intrusive igneous rocks, and Paleocene to Eocene arc-derived strata (References 443 and 220) (Figure 2.5.1-247). These rocks are the result of the Greater Antilles Arc, produced during the subduction of the North America Plate. The volcanic arc rocks are in fault contact with the underlying passive margin strata and are deformed into a synclinorium along the length of Cuba (Reference 439) (Figures 2.5.1-247 and 2.5.1-252). The fault at the base of the Cretaceous arc section is the north-vergent Domingo thrust fault (Figure 2.5.1-247). Paleomagnetic data indicate that these units have poles that are discordant from North America in the mid-Cretaceous, indicating the arc was transported from southwest to northeast

during the Cretaceous to Eocene (References 447 and 448). The Cretaceous arc terrane is sometimes known as the Zaza tectonic terrane and is allochthonous to the passive margin sequence (Figure 2.5.1-248).

Generally exposed directly above the Domingo thrust (Figure 2.5.1-247), the northern ophiolites include sheared serpentinites, interlayered gabbros, and metamorphosed volcanic and sedimentary rocks (such as basalt, chert, limestone, and shale) (Reference 440). Some of these deposits have been identified as back-arc basin related (References 440 and 443). Eclogites from this unit underwent high-pressure (>15 kbar from thermobarometry) metamorphism and $^{40}Ar/^{39}Ar$ amphibole and $^{87}Rb/^{86}Sr$ ages indicate they were emplaced in the early Late Cretaceous (Reference 449).

The arc includes the Mabujina complex, an amphibolite interpreted as a pre-Cretaceous arc basement, Aptian to Campanian tholeiitic to calc-alkaline extrusives and volcaniclastics, and intruding bodies of arc-related granodiorite (Reference 440). The earlier extrusives and volcaniclastics (pre-Albian) are tholeiitic basalts and rhyolites, while the post-Albian volcanic rocks are calc-alkaline andesites (Reference 216). The latest Cretaceous section may be dominated by high-alkaline compositions (Reference 443). $^{40}Ar/^{39}Ar$ mica ages from rhyolites, granodiorites, and other arc products indicate that the Cretaceous arc was uplifted and cooled between 75 and 70 Ma (Reference 770).

### Tertiary Volcanic Arc

Paleocene to Eocene arc rocks are found primarily in southeasternmost Cuba, in the Oriente province (Figures 2.5.1-251 and 2.5.1-247). The southern portion of this arc exposure is dominated by calc-alkaline extrusives, while the northern portion consists of pyroclastics and sedimentary sequences that are more consistent with a back-arc depositional setting. Thin tuffaceous layers found in central and northern Cuban sedimentary successions indicate the distal influence of this younger arc (Reference 440).

### Upper Eocene and Younger Strata

The complex folding and thrust-related deformation present in all of the older units in Cuba is not present in the unconformably overlying upper Eocene to recent sedimentary strata that drape the island. These post-orogenic sediments consist of Late Eocene argillaceous limestones overlain by a thick sequence of Oligocene through Pliocene limestones (Reference 439) (Figures 2.5.1-248 and 2.5.1-252). Iturralde-Vinent (Reference 451) identifies an unconformity that is followed by deposition of Paleocene submarine debris flows and Eocene calcareous shaly flysch. The Eocene-Oligocene contact is at a depth of approximately 4500 feet (1370 meters). The Oligocene unit consists of up to 600 feet (183 meters) of deep-water chalk and limestone that grades laterally into an arenaceous and shaly limestone deposited in marine water of intermediate depth. This is overlain by 400 to 1000 feet (120 to 300 meters) of Miocene sediments consisting of deep-water marl, siltstone, and shaly limestone that grade into arenaceous and calcareous sediments with intercalated, fossiliferous sandy limestone deposited in a neritic environment (Reference 382). Late Tertiary deposits occur in the northern coastal area and dip gently toward the north.

Along Cuba's north coast in the site region, the marine terraces that dip gently seaward (to the north) consist primarily of Miocene through Pleistocene age limestones (References 923 and 924) and extend laterally along the north coast (Reference 848) except where rivers have eroded gaps in the terraces (Reference 926). The terraces are wide, with gentle slopes, the karst processes are more pronounced (i.e., the formation of caves and caverns and sinkholes), and notches (a cut along the base of a sea cliff near the high water mark that forms by undercutting the sea cliff due to wave erosion and/or chemical solution) are pronounced (Reference 921). The Miocene rocks that the marine terrace deposits formed are divided into the Cojimar Formation marls and the Güines Formation carbonates (chalks, argillaceous bioclastic limestones, and reef limestones) that outcrop from Havana to Matanzas. The Cojimar Formation marls represent a middle Miocene deep open

shelf that is overlain unconformably by the Güines Formation. The Güines Formation represents a carbonate platform that covered almost the entire Greater Antilles from the second half of the middle Miocene up to the late Miocene. Late Miocene-Pliocene deposits are only locally developed at the Morro Castle of Havana (the Morro limestones) and near Matanzas City at El Abra de Yumurí (El Abra Formation). The El Abra Formation is a fluvio-marine unit. Pleistocene carbonates of the Jaimanitas Formation (coral reef limestones and calcarenites) are exposed along the coastal plain of Havana and Matanzas (References 383 and 919) and along much of the north coast of Cuba (Reference 925).

Terraces in Cuba near Matanzas are classified as erosional, depositional/cumulative and constructional (References 920 and 923). Erosional terraces on Cuba's northern coastline are located east of Boca de Juruco, province of Havana and in the vicinity of the Bay of Matanzas (Reference 923). Cumulative terraces are described as: (a) having a sandy beach with an inner edge of 1 to 1.5 meters (3.3 to 4.9 feet) above sea level, and (b) storm bank with heights of 2 to 3 meters (6.6 to 9.8 feet) above sea level. Cumulative terraces occur on the northern coastline of Cuba, east of Havana. Constructional coral reef terraces are located on the north coast west of Havana to Mariel and the suburbs of Havana and Santa Fe Jaimanitas (References 920 and 923).

Four marine terraces near Havana occur at elevations 200, 100, 10–15 and 4–5 feet (61, 31, 3.1–4.6 and 1.2–1.5 meters) above mean sea level (References 383, 917, 918, and 926). Near Matanzas, six terraces have been observed at elevations 400, 300, 200, 140, 30, and 5–6 feet (122, 91, 61, 43, 9, and 1.5–1.8 meters) above sea level (References 917, 918, and 926). At Matanzas Bay, Ducloz (Reference 915), Shanzer et al. (Reference 923), and Penalver Hernandez et al. (Reference 921) observed four terraces at the following approximate elevations 25–51 meters (82–167 feet (Rayonera), 15–33 meters (49–108 feet) (Yucayo), approximately16 meters (approximately 52 feet) (Puerto), and 4–10 meters (13–33 feet) (Terraza de Seboruco) (Table 2.5.1-208). The Rayonera terrace is strongly karstic. The presence of sinkholes and caves indicate that the outer edge of the terrace has a height of 39 meters (128 feet), whereas the inner edge is approximately 51 meters (167 feet) giving this surface a topographic slope of approximately 3 to 4 degrees towards the coast. The rocks of this terrace are Pliocene-Pleistocene in age. As noted by its name, the Yucayo terrace is "narrow." It has an average height of 30 meters (98 feet) near the Bay of Matanzas. The terrace is cut off from the sea by a vertical cliff that is approximately 6 to 14 meters (20 to 46 feet) high. Sea caves are present and are indicative of coastal erosion. The Pliocene-Pleistocene rocks of this terrace are algal conchiferas, with hard, massive, and recrystallized limestone reefs. The Pliocene-Pleistocene Puerto terrace is similar to the Yucayo and Rayonera terraces. All three are characterized by the development of karst, sinkholes and a very sharp weathering surface known as "diente de perros" (dog's teeth) (References 915 and 921). The Terraza de Seboruco, the youngest of these terraces is located west of Matanzas Bay. It rises just a few meters (2 to 3 meters) (6.6 to 9.8 feet) above mean sea level with paleolagoonal facies extending inland 1 or more kilometers. Near Havana and Matanzas, the elevation of the Terraza de Seboruco ranges from 2 to 3 meters (6.6 to 9.8 feet) above mean sea level to 4 to 5 meters (13 to 16 feet) above mean sea level, respectively. The terrace is described as porous or cavernous fossilized limestone from the Pleistocene Jaimanitas Formation with a weathering surface of "diente de perros" (References 915 and 925).

The terraces and sea cliffs form a stair-step sequence, which suggests that reef deposition was followed by high sea level stands that cut the bench-like features in the sea cliffs (Reference 912). Several alternate processes can explain or partially explain the stair-step morphology and bench-like features that were described by Agassiz (Reference 912), Spencer (Reference 924), and Ducloz (Reference 915). The alternate hypotheses for what might have contributed to terrace formation as discussed in FSAR subsections are eustatic changes in sea level (Subsection 2.5.1.1.1.1.1.1), changes in ocean circulation pattern (Subsection 2.5.1.1), rise and fall in sea level as a direct result of melting and formation of the continental glaciers (Subsection 2.5.1.1.1.1.1.1), and tectonic activity (Subsections 2.5.1 and 2.5.1.1.3.3).

U-Th dates were obtained on corals (two very large *Montastrea* sp. and one *Acropora* palmate) from the Terraza de Seboruco at the Cantera Playa Baracoa quarry and in the Santa Cruz del Norte canal. When corrected from the initial Uranium age dates, the ages of the samples correspond to the Marine Isotope Stage 5e sea level high stand at approximately 120–130 ka (Reference 925). Toscano et al. (Reference 925) observe that similar age terraces throughout "stable" portions of the Caribbean area are at similar elevations, which is evidence for the absence of active uplift near Matanzas in the past 120–130 ka. Therefore, based on the U-Th dates, the Terraza de Seboruco is correlative to the Cockburntown reef (Bahamas) (Reference 914), Barbados III (Barbados) (Reference 916), and Key Largo Limestone (Florida) (References 913 and 922).

#### 2.5.1.1.1.3 Regional Tectonics within the Site Region

This subsection describes the principal tectonic structures and features in the southeastern U.S. that are located at least partially within the site region, including descriptions of the regional gravity and magnetic fields. Additionally, because Cuba is located partially within the site region, this subsection describes the principal tectonic structures and features of Cuba.

##### 2.5.1.1.1.3.1 Gravity and Magnetic Fields

Gravity and magnetic data for the site region are described in this subsection, with a focus on anomalous features in the gravity and magnetic fields. Data for the gravity and magnetic fields for the site region were obtained from the National Geophysical Data Center. The gravity (Reference 452) and magnetic data (Reference 453) were originally produced for the Decade of North America Project (DNAG).

The original DNAG gravity data were presented on a 6-kilometer grid (and subsequently regridded by the National Geophysical Data Center to a 2.5-minute grid spacing [Reference 454]), which represented free-air gravity anomalies over the ocean and Bouguer anomalies on land. Terrain corrections were only computed and applied in high relief areas of the continent. Large portions of the site region are located offshore and, as a result, this subsection describes both free-air and Bouguer anomalies for the offshore and onshore portions of the site region.

The original DNAG magnetic data were presented on 1.2-mile grid spacing based on a spherical North American Transverse Mercator projection. These data were subsequently also regridded on a 2.5-minute grid spacing by the National Geophysical Data Center. These data sets were selected because they extend farther and give better coverage offshore in southern portions of the site region.

Magnetic highs located near the eastern portion of the Bahamas Platform (including the Little Bahamas Bank) represent both positive and negative magnetic anomalies related to structural controls from post Lower Cretaceous folding or faulting in the basement rocks of the Bahama Platform (Subsection 2.5.1.1.1.3.2.2). The density contrasts below the sea floor as seen from the Bouguer gravity anomaly map and profile (Figures 2.5.1-254 and 2.5.1-255) in this area show that the intraplatform straits (and channels) and basins correlate with the present day platform topography (Figure 2.5.1-254). In general, the straits and basins are areas of negative anomaly while the platforms are generally areas of positive anomaly. The negative anomalies coincide with areas of structural lows, probably downfaulted, and their negative signature stems from a combination of downfaulting of relatively light material and infilling of the resulting lows with low density sediments (Reference 971).

The Blake-Bahamas Platform is related to, and limited by, a regional structural feature, the crustal transition zone, in which Jurassic volcaniclastics are tilted in fault blocks (References 307, 424, and 971). The transitional crust has a smooth, circular magnetic anomaly pattern (Figures 2.5.1-256 and 2.5.1-257). This zone is known as the ECMA and includes the BSMA (Figure 2.5.1-266) (References 466 and 974). The presence of thick volcanics on the oceanic basement of the Blake

Plateau infers that the Bahamas Platform and Little Bahamas Bank overlie seamounts that were produced by CAMP magmatism. This also infers that volcanic activities continued as seafloor was accreted in the CAO (References 466 and 974). A basement map of the Florida-northern Bahamas region using seismic data compiled by Sheridan et al. (Reference 307) (Figure 2.5.1-384) shows the continental, transitional, and oceanic basement rocks along with their approximate ages. Sheridan et al. (Reference 307) concluded that the "transitional crust underlies the northwestern Bahamas to the projected BSMA, and that oceanic crust underlies the Bahamas farther southeast."

### Regional Gravity Field

The gravity field for the site region is shown in Figure 2.5.1-254, which indicates locations of representative field profiles that run through the site region: one oriented north-south parallel to and along the Florida Peninsula and one oriented east-west across the southern portions of the Florida Platform to the northern portions of the Bahama Platform (cross sections A-A' and B-B' in Figure 2.5.1-255). The first-order pattern of the anomalous gravitational field correlates well with bathymetry. Higher gravitational field values are in the range from approximately -20 up to +20 to +40 mGals. These higher values generally are surrounded by steep gravity gradients that generally outline and coincide with the exposed land and surrounding shallow waters of the Florida Platform and Bahama Platform. Intervening deep-water areas such as the Straits of Florida and deep-water channels and basins in the Bahama Bank are characterized by negative values of the gravity field. In addition to the low-density water column, the deep-water channels of the Straits of Florida, Providence Channel, and Tongue of the Ocean have been shown to be floored by thick sequences of low-density sediment, compared to the surrounding platform carbonates (Reference 307). This density contrast and the steep nature of the escarpments that form the boundaries of the Bahama Platform (Reference 307) provide the well-defined nature of the platforms in the gravity field.

The overall positive anomaly associated with the Bahama Platform exhibits a positive gradient to the east from approximately +20 to +40 mGals to over +100 mGals, roughly 225 miles (360 kilometers) east of the Units 6 & 7 site, except where modified by bathymetric effects as described above (Figure 2.5.1-254). The easterly increasing gradient is interpreted as the result of the decreasing importance of transitional crust with more mafic and denser oceanic crust progressively occurring at shallower levels (Reference 307) towards the Atlantic Ocean Basin. This long wavelength, easterly increasing gradient is locally modified by relatively subdued, low amplitude circular to elongate anomalies with amplitudes up to 10 mGals.

The Florida Platform occupies the northwest quadrant of the site region. This portion of the Florida Platform is transected by the Central Florida Gravity Lineament, a well defined northwest-southeast-oriented linear gravity high with an amplitude of a little over +30 mGals in central Florida, at a location approximately 100 miles (162 kilometers) north of the Units 6 & 7 site on gravity profile A-A' (Figures 2.5.1-254 and 2.5.1-255). However, the amplitude of this anomaly is not constant along strike. It appears to decrease slightly to the northwest of profile A-A' and shows higher anomalous values near the coastline to the southeast of profile A-A'. This subsection contains an interpretation of this linear anomaly.

Northeast of the Central Florida Gravity Lineament the Florida Platform is characterized by relatively low values of the gravity field down to a little less than -20 mGals, with the exception of a short wavelength (approximately 12 miles or 19 kilometers) circular gravity high located at approximately mile 40 on profile A-A', which reaches a maximum of a little less than +40 mGals. In contrast, the gravity field associated with the Florida Platform southwest of the Central Florida Gravity Lineament is characterized by relatively long wavelength, on the order of 62 to 93 miles (100 to 150 kilometers), gravitational anomalies that range from 0 to approximately 30 mGals on gravity profile B-B' (Figure 2.5.1-255). These observations indicate that the Central Florida Gravity Lineament effectively forms a boundary that separates the Florida Platform into a northern portion characterized by relatively low gravity field values and short wavelength anomalies and a southern portion characterized by a relatively high anomalous field with broader longer wavelength anomalies.

To the south-southeast of the Units 6 & 7 site a positive anomalous field of approximately 40 mGals is associated with the Cay Sal Bank, which exhibits similar field characteristics to the Bahama Platform just to the east. In the extreme, southern portions of the site region a positive anomalous field occurs in association with the Cuban mainland resulting from relatively dense igneous, metamorphic, and sedimentary basement rocks and their carbonate cover at shallow levels.

A gravitational low anomaly, whose minimum gravity field value of approximately –15 mGals, occurs just north of the site on gravitational profile A-A'.

### Regional Magnetic Field

The regional aeromagnetic field for the site region is shown in Figures 2.5.1-256 and 2.5.1-257, which also indicate the location of magnetic profile A-A' (Figure 2.5.1-258). In distinct contrast with the gravity field in the site region, the magnetic field shows no strong correlation with bathymetry. This is the result of thick carbonate successions that cap the Florida and Bahama Platforms that are essentially nonmagnetic. The magnetic field sources all lie in the sub-Cretaceous basement. However, the fundamental nature of the Central Florida Gravity Lineament (Figure 2.5.1-255) is reflected in the nature and anomalous patterns of the magnetic field in the site region (Central Florida Magnetic Lineament in Figure 2.5.1-256).

The Central Florida Gravity Lineament is associated with similar trending and coincident low values in the magnetic field that exhibit values of -500 nanoteslas (nT) on magnetic profile A-A' (Figure 2.5.1-258) approximately 100 miles (161 kilometers) north of the Units 6 & 7 site. An anomalous magnetic high area on its northwest extension complicates the pattern of this feature, which generally is coincident with the area where the gravitational field associated with this linear feature is also diminishing (Figure 2.5.1-254).

The Florida Platform to the southwest of the Central Florida Gravity and Magnetic Lineament (Figures 2.5.1-256 and 2.5.1-254) and the Bahama Bank exhibit magnetic anomalous features that are circular to ellipsoidal in shape, have relatively long wavelengths, are relatively subdued, and generally trend northwest. Conversely, in central Florida, in the site region and in areas north of the site region, the magnetic field is characterized by relatively high gradient, short wavelength anomalies that are more elongated and linear in shape. In addition, these elongated and linear anomalies are oriented more or less uniformly northeasterly in concert with the structural grain of the Appalachian tectonic province. These trends are truncated by the Central Florida Gravity Lineament. These relationships in the magnetic field further support the fundamental nature of the Central Florida Gravity Lineament, and indicate that it separates two crustal provinces, each of which exhibits different gravity and magnetic field characteristics.

The circular anomalous gravity high noted at mile 40 on gravity profile A-A' (Figures 2.5.1-254 and 2.5.1-255) described in the previous subsection is associated with an apparent magnetic dipole source with a magnetic field high of +300 to +400 nT to the south of a magnetic low of -400 nT. The close association of a dense and magnetic source at this location would suggest that the potential field source for this feature is probably a mafic intrusion in the basement (Reference 212).

Within the site vicinity, a steep northerly increasing magnetic gradient marks the transition from a magnetic field minimum of approximately -200 nT to the south to a magnetic field high of approximately +200 nT to the north. This relationship is opposite of that expected for a magnetic dipole in the northern hemisphere so the significance of the magnetic field is uncertain in terms of a primarily induced field interpretation. However, the occurrence of several magnetic high and low anomalous field regions in the vicinity most probably indicates a magnetized basement with a significant mafic component. However, as noted, the site vicinity is located in association with a gravity low anomaly indicating that the basement in the vicinity is less dense than would be expected from a purely mafic composition and possibly reflects basement lithologic variation.

### Summary of Gravity and Magnetic Findings

Klitgord et al. (Reference 212) and Sheridan et al. (Reference 307) present a synthesis and discussion of the anomalous gravitational and magnetic fields for the Florida Bahama Platform including the site region. The following description summarizes the pertinent points that relate to the site region presented in these sources.

In addition to the contrasting character exhibited by the gravity and magnetic fields on either side of the Central Florida Gravity Lineament, the location of the lineament is also correlated with other basic changes in the nature of the crust. The lineament marks a transition in the composition of the crust in the subsurface. In addition, the top of the basement surface becomes much deeper to the southwest. North of the lineament, in the site region, the basement of the Florida Platform is relatively shallow and characterized by Paleozoic igneous and metamorphic rocks that compose the Central Florida basement complex, which roughly corresponds with the Suwannee terrane (Figure 2.5.1-257). On the southwestern side of the central Florida lineament the carbonate platform cap becomes much thicker, and the composition of the crust becomes more “transitional” in nature and is composed of Jurassic volcaniclastic sequences.

Klitgord et al. (Reference 212) notes the fact that the Central Florida Gravity and Magnetic Lineament aligns with the Bahamas Fracture Zone (Figures 2.5.1-257 and 2.5.1-205) and suggests that it likely represents a Jurassic transform that separates a Late Jurassic spreading center in the Gulf of Mexico from the Central Atlantic spreading center. North of this transform plate boundary the crust is continental in nature and exhibits structural features associated with Appalachian tectonics as shown in the distinct northeast trend in the potential field anomalies. The relatively shallow basement and consequent relatively thin carbonate cap results in little attenuation of the high frequency components of the potential field anomalies. For that reason, the high gradient characteristics of the gravity and magnetic sources are preserved.

In contrast, the basement of the Florida Platform and other areas southwest of the transform boundary is much deeper, with a correspondingly thicker carbonate cap on the Florida and Bahama Platforms. The basement in these areas contains rotated and tilted blocks of volcaniclastic rocks. These deeply buried volcaniclastic sources result in subdued potential field anomalies whose high-frequency components have been attenuated and the field gradients subdued; consequently, they exhibit broad long wavelength characteristics. Also, detailed interpretation of the magnetic field anomalous sources associated with the volcaniclastic basement to the southwest of the transform boundary is complicated by the fact that some of these volcanic rocks may contain a component of remnant magnetization that may significantly modify the anomaly produced by the induced field.

#### 2.5.1.1.1.3.2 Principal Tectonic and Structural Features

The site region is covered by a thick sequence of sedimentary rocks and deposits that obscure any Precambrian to Paleozoic tectonic features associated with the formation of Pangea (Figures 2.5.1-240, 2.5.1-242, and 2.5.1-201). In fact, this region has generally recorded only sedimentary processes since Mesozoic rifting, with the exception of the possible tectonic activity associated with the Cuban fold and thrust belt, possibly active faults in northern Cuba, adjacent Straits of Florida normal faults, the Santaren anticline, and the Walker’s Cay fault. The Florida Platform has been a site of stable carbonate platform deposition continually since the Cretaceous. Variations in sediment thickness are interpreted as a series of arches, uplifts, basins, or embayments from geophysical or borehole data (Reference 413). Generally, these arches and basins are sedimentary responses to minor warping, regional tilting, sedimentary compaction, or sea level changes and are not considered associated with faulting or tectonic events (Reference 413). In some cases, the highs or lows seen in the stratigraphy may be mimicking Mesozoic paleotopography. The Bahama Platform is also largely undeformed, but does include sparse post-rift faulting or deformation, generally adjacent to the Cuban orogen. The EPRI (Reference 456) earthquake catalog and the updated earthquake catalog completed for the Units 6 & 7 site investigation

(Subsection 2.5.2.1) indicate that north of Cuba and the northern Caribbean seismic source model (Subsection 2.5.2.4.4.3) earthquakes are sparsely and randomly distributed within the site region and that none of the earthquakes can be associated with a known geologic structure (Subsection 2.5.2.3).

#### 2.5.1.1.1.3.2.1 Florida Peninsula and Platform Tectonic and Structural Features

##### Structures of the Florida Peninsula and Platform

The Florida Peninsula exposes only flat, unfaulted strata at the surface (Figure 2.5.1-201), so all tectonic features are identified with subsurface methods, usually from drilling or seismic reflection data. Tectonic and structural features identified within the Florida Platform province (which includes the emergent peninsula and the submerged carbonate platform) are mostly a series of gentle highs and lows in the regional stratigraphy. In some cases, these may be reflecting original basement topography (such as the Peninsular Arch) or may reflect changes in sedimentation later in time (Figures 2.5.1-259, 2.5.1-260, and 2.5.1-261). Both surface exposures and well data indicate that Cretaceous and younger strata on the Florida Platform are generally gently dipping to horizontal (Figures 2.5.1-261, 2.5.1-240, 2.5.1-242, and 2.5.1-232). Local and regional seismic data and high-resolution bathymetric data indicate that the shallow stratigraphy in southern Florida is undeformed by tectonic faulting (e.g., References 798, 799, and 398). In a few instances, variations in pre-Miocene stratigraphy recorded in boreholes and seismic-reflection data have been interpreted as possible faulting (for example, the queried fault on Figure 2.5.1-234 [Reference 373] or the fault from Cunningham [Reference 999]), and erosional paletopography or karst may complicate these interpretations. More typically, local and regional seismic data and high-resolution bathymetric data indicate that the shallow stratigraphy in southern Florida is undeformed by tectonic faulting (e.g., References 798, 799, and 398). Similarly, continuous, unfaulted prograding strata drape the edges of the Florida and Bahama Platforms along the Straits of Florida (Figure 2.5.1-262) (Subsection 2.5.1.1.1.2.2 provides a discussion of the prograding strata).

##### Basement Faults of the Florida Peninsula and Platform

Faults cutting pre-Mesozoic basement of the Florida and Bahama Platforms are of two types: (a) normal faults responsible for the extension and thinning of the continental crust during rifting, and (b) hypothetical strike-slip faults inferred from plate reconstructions. Seismic reflection data indicate that the basement beneath the Bahama and Florida Platforms is faulted (e.g., Figure 2.5.1-263), although many regional-scale cross sections beneath the Florida Platform indicate it has a smoothly varying basement surface (Reference 457) (Figure 2.5.1-264). Normal faults have not been mapped confidently on the platform, but some have been preliminarily identified using well data to determine depths to basement in northern and central Florida, outside of the site region (Reference 457) (Figures 2.5.1-261 and 2.5.1-263). Because drilling data and mapping of exposed Eocene and younger strata as well as available seismic reflection profiles (largely offshore) indicate that the Cretaceous and younger strata are unfaulted (Figures 2.5.1-230, 2.5.1-263, and 2.5.1-265), the age of this faulting is pre-Cretaceous (References 457 and 458).

A series of hypothetical faults have been drawn on various tectonic maps depicting the subsurface, pre-Mesozoic lithology of the Florida Platform, including strike-slip structures across central and southern Florida (e.g., References 458 and 212) (Figure 2.5.1-253). These primarily serve to separate the different lithologies encountered through subsurface drilling or potential correlations in magnetic data, or have been proposed to accommodate potential misfits in plate reconstructions (e.g., References 459 and 460). The structures, if they do exist, are expected to have been inactive since the end of the Jurassic because no offset is seen in any younger strata (Reference 458). This includes structures variably referred to as the West Florida-Ouachita megashear (Reference 461), the Jay fault (Reference 351), and the Bahamas Fracture Zone (Reference 212). Most notably, the Bahamas Fracture Zone is drawn through the site region, often along the boundary between the Suwannee terrane and Jurassic basement of southern Florida (Reference 212) (Figures 2.5.1-257,

2.5.1-205, and 2.5.1-229). The Bahamas Fracture Zone is a northwest-trending boundary seen in gravity and magnetic data (Reference 212) and the distribution of basement lithologies (e.g., Reference 463). It is interpreted as a Jurassic transform plate boundary (Reference 212). Offset has been interpreted in the Late Jurassic to Early Cretaceous strata in northwestern Florida, and attributed to a Bahamas Fracture Zone-style structure (Reference 464). No Cretaceous or younger faulting or offset is depicted in regional cross sections that transect this structure (References 344, 212, and 465), nor is any faulting seen at the surface (Figures 2.5.1-260, 2.5.1-261, and 2.5.1-230). In particular, interpreted seismic reflection lines that cross the Bahamas fracture zone show unfaulted Cretaceous and younger strata (e.g., Figures 2.5.1-263 and 2.5.1-262). The most recent mapping of magnetic anomalies on the Atlantic Ocean seafloor indicates that the Bahamas Fracture Zone ends east of the Bahama Bank and outside of the site region (Figure 2.5.1-266). Similarly, any offset of magnetic anomalies across the Blake Spur Fracture Zone (which offsets Chrons 25 by 30 to 50 kilometers [19 to 31 miles]) ends east of the Blake Spur magnetic anomaly, approximately 300 miles (480 kilometers) from the site (Reference 466).

### Peninsular Arch

The Peninsular Arch is a northwest-trending feature that formed a relative topographic high until the Cretaceous Period (Reference 467) (Figures 2.5.1-229 and 2.5.1-259). East-west geologic cross sections across northernmost Florida show that Triassic/Jurassic and lower Cretaceous strata are truncated against this basement high (Reference 465) (Figure 2.5.1-259). Upper Cretaceous beds were deposited over the crest of this arch, indicating it had ceased to be a high by that time as the sea-level rose to an elevation that permitted continuous deposition across the arch (References 465 and 467) (Figure 2.5.1-260). However, Oligocene deposition may have been affected by the Peninsular Arch, but Neogene to Holocene sediments are unaffected (Reference 304).

### Sarasota Arch

The Sarasota Arch is a northeast-trending basement high extending southwest from the Peninsular Arch in the west-central portion of the Florida Platform (Figure 2.5.1-229). It borders the South Florida Basin to the south and the Tampa Basin to the north. This basement high controlled deposition during the Jurassic and early Cretaceous (Reference 413). Seismic data interpreted by Ball (Reference 468) indicate a rough and faulted basement surface overlain by horizontal strata.

### Tampa Basin

The Tampa Basin lies just north of the Sarasota Arch and primarily affects basement rocks of the Florida Platform (Figure 2.5.1-229). Ball (Reference 468) interprets a gentle syncline in Paleozoic reflections, but horizontal strata of Mesozoic and younger age persist upsection across the feature.

### Broward Syncline

The Broward Syncline (Figure 2.5.1-229) is a northwest-southeast-trending syncline mapped in the subsurface Cretaceous strata (References 457 and 467) but its presence cannot be “unequivocally demonstrated because of lack of drilling” (Reference 469). This feature is located in Broward and Palm Beach counties, outside of the site vicinity (Figure 2.5.1-229).

### South Florida Basin

The South Florida Basin (Figure 2.5.1-229) is a sedimentary basin filled with 3 to 8 miles (5 to 13 kilometers) of Jurassic to Holocene strata (Reference 409) that slowly subsided at the southern end of the Florida Platform (Figure 2.5.1-261). These deposits are the thickest in the Atlantic coastal plain. The Upper Jurassic and younger strata are shallow-water limestones and dolomites, evaporites, and deep-water limestones. The thickening to the southwest displayed in Cretaceous strata is sometimes referred to as the South Florida Shelf (Reference 457). The South Florida Basin is separated from the Tampa Basin by the Sarasota Arch (Reference 368) (Figure 2.5.1-229). To the south, the edge of the South Florida Basin is sometimes delimited by the Pine Key Arch or the Largo

High. However, these features, defined by variations in sedimentary facies, are not consistently identified or described (References 355, 413, and 469).

### Ocala Uplift

The Ocala Uplift, or Ocala Platform, is a feature characterized by thickness variations in the Eocene strata in northwest Florida approximately 300 miles (480 kilometers) from the Units 6 & 7 site (Figure 2.5.1-229). The stratigraphic variations are interpreted to result from either compaction shortly after deposition or sedimentary build-ups and do not have a fault-controlled origin (References 389 and 470). The Ocala Uplift was subaerially exposed during the Eocene to Miocene, but the cause of this is uncertain. Mid-Miocene and younger deposition is undisturbed across the Ocala Uplift (Reference 304).

### South Georgia Rift Basin

The South Georgia Basin (or Rift) is the southernmost continental rift basin associated with the opening of the Atlantic Ocean (Reference 471), and is located more than 350 miles (560 kilometers) north of the Units 6 & 7 site (Figures 2.5.1-229 and 2.5.1-261). Approximately 50 borings in northern Florida, southern Georgia, and southeastern Alabama have penetrated nonmarine clastics and diabase dikes and sills of Late Triassic and Early Jurassic age, known as the Eagle Mills Formation (References 458 and 344). These strata have been correlated with the Newark Supergroup, which fill rift basins from Newfoundland to Alabama along the East Coast and result from the widespread Mesozoic rifting that accompanied the opening of the Atlantic Ocean. In northern Florida, these strata may be 1830 meters (6000 feet) thick (Reference 211). The clastics have been interpreted to fill a graben up to 60 miles (100 miles) wide (Reference 458).

### Suwannee Channel

The Suwannee Channel is a general term for the Suwannee Straits and the Gulf Trough (Figures 2.5.1-229 and 2.5.1-218), two marine channels that are located in roughly the same position, which separated southeastern North America from the Florida carbonate platform during two different time periods. The Suwannee Straits were located in southern Georgia, panhandle Florida, and southernmost South Carolina in the Middle Cretaceous to Late Paleocene (References 258 and 257). The Gulf Trough is middle Eocene to early Oligocene in age (Reference 473). After the Oligocene, this trough had been filled, which allowed for clastic deposition on the central peninsula in the late Early Miocene (Reference 389). The currents from the Gulf of Mexico that flowed through these features to the Atlantic Ocean might have prevented clastics from the Appalachians and southeastern North America from reaching the platform (Reference 389). There is no faulting or tectonic activity associated with this feature, which is located approximately 400 miles (640 kilometers) to the north of the Units 6 & 7 site (Figure 2.5.1-229).

### Queried Fault from Cunningham et al.

Cunningham et al. (Reference 373) postulate that a fault or paleotopography could be responsible for elevation variations in the Arcadia Formation in southwestern Florida (Figure 2.5.1-229). The queried structure is between 50 and 60 kilometers long (30–37 miles) (Figure 2.5.1-229), and at its nearest approach, the eastern end of the queried fault is approximately 41 kilometers (25 miles) west of the Turkey Point Units 6 & 7 site.

Figure 2.5.1-234 shows a cross section across southern Florida that was developed with data from eight wells in southern Florida, with variable horizontal scale between pairs of wells and, thus, with variable vertical exaggeration. Between the southernmost two wells, Cunningham et al. (Reference 373) postulate the existence of a fault that cuts up through Avon Park Formation, Suwannee Limestone, and Oligocene-Miocene-age Arcadia Formation, and potentially places the Arcadia Formation in fault contact with the lower portion of the overlying Miocene-Pliocene Long Key Formation. Alternatively, the Long Key Formation may be interpreted as deposited across a

paleoscarp. Reference 373 labels the postulated fault on this cross section with two question marks, indicating the speculative nature of this fault.

In cross section, the postulated fault cuts units as young as the Miocene Arcadia Formation, and although the Miocene to Pliocene Long Key Formation and the Pleistocene Key Largo are depicted as unfaulted, they have thickness and elevation differences across the structure (Figure 2.5.1-234). Higher up-section above the queried fault tip, Cunningham et al. (Reference 373) cross section shows marine carbonate stringers that could be interpreted as deformed by slip on the underlying fault. Alternatively, these marine carbonate stringers could represent deposition draped across a paleoscarp and thus could post-date slip on the underlying postulated fault.

Although the postulated fault in Figure 2.5.1-234 would not represent a Quaternary faulting hazard for the site if it existed, in detail the thickness and stratigraphic variations may instead be related to paleotopography. Indeed, the top of the Arcadia Formation is known to be an erosional unconformity with significant paleotopographic variation. For example, “A distinct regional unconformity and subaerial exposure surface at the top of the Arcadia Formation separates the Long Key and Arcadia formations” (Reference 393). A cross section presented by Reference 393 depicts 90 meters (295 feet) of relief on the top of the Arcadia Formation surface in southern Florida, while the thickness of the Arcadia Formation varies from 200 meters (656 feet) in the central portion of the Florida peninsula to between 0 and 20 meters (0 and 66 feet) farther east (Reference 394). A study in southern Florida determined that intensification of marine currents increased the erosion of marine carbonates and led to a significant time hiatus (more than 4 m.y.) following deposition of the Arcadia Formation (Reference 934) and the influence of Arcadia Formation paleotopography on highs in subsequent carbonate and clastic deposition in southernmost Florida has been recognized (Reference 395).

On Key Largo, relief on the top of the Arcadia Formation as large as 40 meters (131 feet) was found between borings only a few kilometers apart (Reference 393). Furthermore, in other cross sections presented by Reference 273, the elevation of the top of the Arcadia Formation varies by approximately 100 meters (328 feet) between wells W-3174 and W-17086 (88 kilometers or 55 miles apart), by 50 meters (164 feet) between wells W-17156 and W-12554 (56 kilometers or 35 miles apart), and by 25 meters (82 feet) or 1.2 miles between wells W-3011 and W-17157 (2 kilometers apart), all interpreted without faulting. The slope required to achieve this latter elevation variation, 0.7 degree, is actually greater than the slope required to achieve the elevation variation observed in the Arcadia Formation between the Everglades Park and Gulf Oil wells, where the queried fault is depicted in Figure 2.5.1-234 (approximately 100 meters [328 feet] over 18 kilometers [11 miles] of distance, or a 0.3-degree slope). Numerous other examples exist throughout southern Florida of steeper paleotopographic slopes on the top of the Arcadia Formation that are not associated with faulting. In addition, the down-to-the-south separation depicted on the postulated fault in Figure 2.5.1-234 is consistent with, and may, in part, be attributed to the regional southward dip of the strata towards the South Florida Basin in the area (References 377 and 389).

The karst-influenced paleotopography of the Arcadia Formation is detailed in Reference 936. While using borings at a much finer spacing than Cunningham study, the Hine study documents karst sub-basins with as much as 100 meters (328 feet) of relief over distances of kilometers to tens of kilometers on the top of the Arcadia Formation in west-central Florida. They attribute this relief to a mid- to late-Miocene sea level lowstand that caused dissolution in the deeper carbonates, such as the Arcadia Formation, and formed paleotopographic depressions and non-tectonic deformation in the Arcadia Formation (Reference 936).

Alternative interpretations of well data in southern Florida, often including the three wells closest to the postulated fault, provide evidence for unfaulted Eocene to Pliocene stratigraphy in the same location (References 389, 393, 934, 935). For example, Reference 934 provides a stratigraphic correlation diagram across the projection of the queried fault from Cunningham et al.

(Reference 373) and interprets no faulting. This diagram also displays similar relief between boreholes on the top of the Arcadia to the north. Likewise, the regional north-south-oriented cross section shown in Figure 2.5.1-233 intersects the projection of the queried fault and does not indicate faulting in the area.

As shown in Figure 2.5.1-381, there are three wells adjacent to the queried structure: Gulf Oil W-3510 south of the postulated fault and W-1115 and W-2404 north of it. The Gulf Oil well W-3510 appears to control the set of structure contours used to delineate the area of faulting (Figures 2.5.1-234 and 2.5.1-381). Yet, other published contour maps of the same well data use dashed contours and question marks to indicate uncertainty in contouring such sparse data in the Florida Bay area (Reference 393). A later publication (Reference 935) also provides interpretations of unfaulted Miocene to Pliocene stratigraphy in the same location as the postulated fault from Reference 273.

In summary, numerous other sources using similar well data indicate unfaulted strata that gently dips to the south in this location, reflecting the influence of the South Florida Basin (References 389, 393, 396, and 827). The fault postulated by Reference 273 has not been documented in any subsequent investigations and numerous examples of paleotopographic variation in the top of the Arcadia support a non-fault-related origin for the stratigraphic variations seen in Figure 2.5.1-234.

### Tectonic Fault from Cunningham

Cunningham et al. (Reference 989) and Cunningham (Reference 999) identified a buried, north-northeast-striking fault beneath Biscayne Bay approximately 11 kilometers (6.8 miles) north of Turkey Point Units 6 & 7 (Figures 2.5.1-229 and 2.5.1-331). This feature is identified on five seismic-reflection profiles and is interpreted to have vertical separations of approximately 12 meters (40 feet) of the tops of the Avon Park and Arcadia formations (Reference 999; Figure 2.5.1-393). Greater separation at the top of the Arcadia Formation surface is due to offset paleotopography (Reference 999; Figure 2.5.1-393). A range of dips from 89°W to 85°E is observed on the five seismic profiles, and the change in dip direction and steepness of the fault dips is cited by Cunningham (Reference 999) as evidence for strike-slip faulting. Cunningham (Reference 999) indicates that the latest movement occurred in the middle Miocene to early Pliocene epochs based on the absence of offset reflectors in the overlying Miocene Peace River, Pliocene Tamiami and younger strata (Figure 2.5.1-393). Therefore, this fault does not represent a capable tectonic fault or pose a fault rupture hazard for Turkey Point Units 6 & 7. Cunningham (Reference 999) suggests this strike-slip fault could be related to the buried regional transform fault systems related to the opening of the Atlantic Ocean.

### Reverse Faults and Anticline on Miami Terrace from Cunningham

In addition to the strike-slip faulting in Biscayne Bay, reverse faults and an anticline on the offshore Miami Terrace were imaged in seismic-reflection profiles approximately 30 miles (45 kilometers) northeast of Turkey Points Units 6 & 7 (References 989 and 999) (Figures 2.5.1-394 and 2.5.1-395). Both of these features, the anticline and the reverse faults, were each only imaged in one location, so no information about the extent or strike of these structures is currently available. However, based on the seismic-reflection data, the reverse faults offset strata of the upper Floridan aquifer in a down-to-the-west sense (Figure 2.5.1-394). Yellow arrows point to seismic-reflections indicating structural offset between fault blocks (Reference 999) (Figure 2.5.1-394). The anticline, located 3.7 miles (6 kilometers) west of the group of reverse faults, comprises folded strata of the Eocene Avon Park to lower Miocene Arcadia formations (Figure 2.5.1-395). Subsequent to this folding, erosion resulted in a set of horizontal wave-cut terraces incising the anticline, and finally, undeformed late Pliocene- to early Pleistocene-age strata unconformably onlap and downlap these terraces (Reference 999). Cunningham interprets the compression causing these faults and the fold to have occurred after the erosion of the Oligocene or early Miocene lower Arcadia Formation and before deposition of the overlying undeformed late Pliocene or early Pleistocene strata, and thus ‘sometime

during the Oligocene to early Pleistocene' (Reference 999, p. 18). The timing of this contraction is similar to the uplift of the Santaren anticline and the tectonic strike-slip fault from Cunningham (Reference 999) in Biscayne Bay. Given their age and absence of any nearby Quaternary tectonic structures or capable faults, the reverse faults and fold described by Cunningham et al. (2012) and Cunningham (2015) (References 989 and 999) do not represent capable tectonic sources.

#### Seismicity of the Florida Peninsula and Platform

The Phase 1 earthquake catalog (Subsection 2.5.2.1) indicates sparse seismicity in the Florida Peninsula tectonic province (Figures 2.5.3-203 and 2.5.1-267). There are only a few widely distributed events within this province, the largest being an Emb 4.3 earthquake that occurred in 1879, located 479 kilometers (298 miles) from the site. The Emb (best estimate body-wave magnitude) is described in Subsection 2.5.2.1. The nearest event, located 53 kilometers (33 miles) from the site, occurred on December 22, 1945, and had an Emb magnitude of 2.7. The overall seismicity pattern and the above-mentioned events show no correlation with geologic or tectonic features (Subsection 2.5.2.3).

### 2.5.1.1.1.3.2.2 Bahama Platform Tectonic and Structural Features

#### Structures of the Bahama Platform

The Bahama Platform, like the Florida Platform, is best characterized by continuous, horizontal carbonate deposition, rarely interrupted by faulting or other deformation (Figure 2.5.1-245). Because the platform is largely submerged, all information about potential structures is gained from interpretations of seismic lines, and therefore is subject to limitations. The vast majority of seismic lines inspected and available to this study confirm the unfaulted nature of Cretaceous and younger strata across the Bahama Platform and southern Florida Platform (Figures 2.5.1-263, 2.5.1-268, 2.5.1-269, 2.5.1-270, 2.5.1-271, and 2.5.1-272). However, a few exceptions to this exist, such as the deformation associated with the Santaren anticline (Figure 2.5.1-278) normal faults in the Straits of Florida (Figure 2.5.1-273), the Walkers Cay fault (Figure 2.5.1-275), and the eastern Bahama Platform (right panel of Figure 2.5.1-264).

#### Mesozoic Normal Faults of the Bahama Platform

As described above, the openings of the Gulf of Mexico and Atlantic Ocean led to the development of Mesozoic normal faults that extended the basement beneath the Florida and Bahama Platforms. No detailed maps of the entire subsurface Bahama Platform exist, but limited mapping of such faults has been done in conjunction with large-scale seismic surveys. For example, a seismic line in the Straits of Florida identified several minor normal faults cutting strata above a mid-Cretaceous shallow-water carbonate platform at a depth of 940 meters (3084 feet) below the seafloor (Figure 2.5.1-274). More commonly, the basement of the Bahama Platform is depicted as a series of fault blocks with syntectonic Triassic to Jurassic strata, draped by undeformed Lower and/or Upper Cretaceous strata. In the eastern Bahama Platform, Sheridan et al. (Reference 307) interpret normal faults cutting Lower Cretaceous strata that are draped by unfaulted Upper Cretaceous (Santonian or Cenomanian) strata (right panel of Figure 2.5.1-264). On Figure 2.5.1-263, a north-south seismic line located east of the site indicates normal faulted basement of Paleozoic to Jurassic strata draped by unfaulted Upper Jurassic to Lower Cretaceous strata. Similarly, the seismic line interpretation on Figure 2.5.1-243 indicates faulted basement covered by undeformed Upper Jurassic and younger strata. On Figures 2.5.1-268 and 2.5.1-269, flat unfaulted Lower Cretaceous and younger strata cover the Bahama Platform.

#### Walkers Cay Fault

The Walkers Cay fault was initially identified by Mullins and Van Buren (Reference 474) north of Little Bahama Bank based on seismic reflection data. As later mapped by Van Buren and Mullins (Reference 791), the fault is a 33-kilometer-long (21-mile-long) structure that strikes north-northeast

(Figures 2.5.1-275 and 2.5.1-366). In contrast, Austin et al. (Reference 785) depict a broad zone of faulting by mapping the northwest and southeast boundaries of the zone, but do not map the extent of any individual strands in the zone. These boundaries of the Walkers Cay fault zone are defined as having a more easterly strike than the fault of Van Buren and Mullins (Reference 791) and a similar length (Figure 2.5.1-366). The spatial coincidence of the faulting expressed in Oligocene- to Cretaceous-age strata with a magnetic anomaly has been used to interpret the Walkers Cay fault as a basement-involved structure (References 307 and 474).

In the vicinity of the Walkers Cay fault, five seismic reflection lines with variable levels of interpretation have been reproduced in the published literature. Mullins and Van Buren (Reference 474) and Van Buren and Mullins (Reference 791) present two air gun seismic reflection profiles (Profile 4 and Profile E) in their reports about this structure (Figure 2.5.1-366). Shortly thereafter, the Ocean Drilling Program (ODP) Leg 101 conducted another seismic survey of the north slope of Little Bahama Bank and these data are discussed in several publications (References 476, 785, 937, 938, and 940). Three published seismic lines from that work (Lines LBB-13, LBB-17 and LBB-18) depict the Walkers Cay fault or a splay of the fault (Figure 2.5.1-366). In at least one of those lines (Line LBB-18), the authors interpret a normal fault “believed to be the Walkers Cay normal fault” as extending up to the seafloor, suggesting possible Quaternary activity (Reference 785) (Figure 2.5.1-367). Normal faulting is reportedly visible on lines LBB-5, LBB-6, and LBB-15 (Reference 476), but these seismic lines have not been reproduced in any of the publications from ODP Leg 101.

Borehole data in the vicinity of these seismic profiles indicate that the Quaternary section is limited to a thin veneer. At ODP sites 627, 628, and 630, analysis of cores found that planktonic foraminifers associated with Pleistocene sediments are limited to approximately the uppermost 15.5 meters (50.9 feet), 3.6 meters (11.8 feet), and 18.2 meters (59.7 feet), respectively (Figure 2.5.1-366) (References 937, 938, 939). There is no indication of abrupt thinning or thickening of layers in the seismic profiles that would suggest these observations are spatial anomalies. Furthermore, these thicknesses are in agreement with regional mapping (Reference 941) that indicates Neogene (Pliocene or Miocene) strata are within 20 meters (66 feet) of the seafloor in this area. Thus, to determine if the Walkers Cay fault is a Quaternary structure, seismic data would need to resolve displacement within approximately the uppermost 20 meters (66 feet) of seafloor sediments.

In summary, on only one of the five interpreted seismic lines that cross the Walkers Cay fault (LBB-18) do the authors interpret a fault reaching the seafloor (Figure 2.5.1-367) (Reference 785). This is consistent with the summary of seismic lines collected in ODP Leg 101 that states that “throughout the area it [the Walkers Cay fault] has only a minimal effect on sediments younger than middle Miocene” (Reference 476). The possibility of Quaternary slip on the Walkers Cay fault cannot be precluded by the available data. For this reason, a hazard sensitivity calculation for a Walkers Cay fault source is presented in Subsection 2.5.2.4.4.3.4.

### Jacksonville Fracture Zone

Several steep, buried normal faults located outside of the site region northeast of the Bahama Platform are known as the Jacksonville Fracture Zone (also known as the Great Abaco Fracture Zone) (References 424 and 307) (Figures 2.5.1-208 and 2.5.1-253). The faults cut Cretaceous strata and are covered by flat and unfaulted lower Miocene units (Reference 307). Because strata from Upper Cretaceous to Oligocene are missing, it is impossible to know where within this interval the faulting occurred (Figure 2.5.1-208). Sheridan et al. (Reference 307) speculates that the shape of the northern margin of the Little Bahama Bank is controlled by post-Cretaceous faulting on the Jacksonville Fracture Zone. Because lower Miocene and younger strata are unfaulted by these faults (Figure 2.5.1-208), these faults do not represent a capable tectonic source.

### Possible Faults and Folds Near Cay Sal Bank

Based on new limited-penetration seismic reflection and bathymetry data, Kula (2014) (Reference 1022) and Eberli et al. (2015) (Reference 1023) proposed that four faults and related folding east and northeast of Cay Sal Bank may be tectonically related to a northeast extension of the Cuban fold-and-thrust belt, similar to previous suggestions by Bergman (Reference 906). As shown by Bergman (Reference 906), movement on the faults was primarily older than Paleocene and Eocene in age, but some waning movement may have continued into the late Miocene and possibly Pliocene. Kula's (Reference 1022) interpretation of the two eastern-most and longer faults suggests ages similar to Bergman (Reference 906), but also suggests that two shorter faults (Faults A and B) east and north of Cay Sal Bank have evidence of neotectonic activity including seafloor breaks up to 50 meters high (Figure 2.5.1-229).

Evaluation of additional seismic data, discussions with Kula's principal thesis advisor, Prof. G. Eberli, and comparisons to features identified by other researchers working with similar high-resolution data in the region (e.g., Tournadour et al. [Reference 1024]; Jo et al. [Reference 1025]) indicate that most of the features interpreted by Kula (Reference 1022) and Eberli et al. (Reference 1023) as evidence of neotectonic activity are more credibly interpreted as evidence of mass movements and gravity-driven slope failures, and that other non-seismogenic processes including polygonal faults, fluid expulsion, and sediment compaction are likely the primary origin of features mapped by Kula (Reference 1022). The youngest fault interpreted by Kula (Reference 1022) which extends to the seafloor is not continuous either in depth or length, indicating a non-tectonic origin. Therefore, the possible faults and folds near Cay Sal Bank do not represent capable tectonic sources.

### Santaren Anticline

The northwest-trending detachment fold primarily affects Cretaceous to Miocene strata and represents the northern limit of the Cuban fold-thrust belt (Reference 501, Figures 2.5.1-229 and 2.5.1-350). Initial work indicated that folding initiated in the Late Cretaceous, reached maximum expression in the early Cenozoic, and experienced differential compaction in the late Cenozoic (Reference 501), a timeline consistent with the end of Cuban orogeny in the latest Eocene.

Detailed analysis of the stratigraphy indicates that the syntectonic growth strata are Eocene and younger and was used to infer Pliocene or potential early Quaternary activity on the structure (References 426 and 479, Figure 2.5.1-278). References 426 and 479 use the geometries and inferred ages of growth strata associated with the Santaren anticline to respectively model the shortening rate and the temporal variability in sedimentation and fold-growth rates since Late Oligocene time. The authors conclude that the geometry of Santaren anticline growth strata results from the interplay between sedimentation and tectonic fold uplift and that sedimentation and fold-growth rates have been highly variable over time, though sedimentary processes, such as localized bottom-current erosion and sediment compaction (Reference 501), could be responsible for the stratigraphic variations.

Reference 479 interprets the variation as indicating that the "evolution of the Santaren anticline consists of cycles that involved tectonically active periods separated by interruptions in which the tectonic activity fell to zero" (Reference 479). Furthermore, their analysis suggests that the preponderance of tectonic growth of the Santaren anticline occurred before 20 Ma (i.e., before bed E in Figure 2.5.1-278). Since that time, the average fold uplift rate is approximately 0.03 millimeter per year, corresponding to a shortening rate of 0.001 millimeter per year (Reference 426). Reference 479 concludes that, for the time period 6.2 Ma to present, "there were many lapses in this [time period] during which no tectonic uplift occurred" (Reference 479) and that the greatest fold uplift rate since approximately 6.2 Ma occurred during or just before deposition of beds K2 and K3, which are assigned Late Miocene age. Since deposition of beds K2 and K3, Santaren anticline fold uplift rates have been at or near zero (Figure 2.5.1-278). The youngest interval for which a non-zero uplift rate was calculated was the early Quaternary M2–M3 interval, which has a 0.05 millimeter-per-year

(0.002-inch-per-year) fold uplift rate (Reference 479, Figure 2.5.1-278). However, it should be noted that the calculation of fold uplift, based on differences in crestal relief, incorporates measurements, calculations, and assumptions that are assigned an error of 10 percent (Reference 426). The calculated crestal relief between beds M2 and M3 is less than 24 meters (79 feet), and this reflects the difference between two values that each may vary by ±130 meters (427 feet) (References 426 and 479). The preponderance of data indicate that this structure was predominantly active in the Eocene, with waning activity throughout the Miocene, and possible, yet questionable, deformation into the early Quaternary. However, beds with ages of approximately 1 Ma and younger show no evidence of deformation (Reference 479, Figure 2.5.1-278). Horizontal shortening rates over the last 3.6 Ma are estimated to be 0.0003 millimeter/year and seismicity near this structure is sparse (Reference 426, Figure 2.5.1-350). The fold may be rooted in Jurassic evaporites, such as the Punta Allegre Formation (References 307 and 477), which could account for this structure's apparent longevity without clear tectonic mechanisms.

Figure 2.5.1-350 and Figure 2.5.1-368 (Sheet 2 of 3) show mainshock and dependent seismicity in the vicinity of the Santaren anticline. There is one $M_w$ 3.26 event on June 2, 1990 that is located 45 kilometers northwest of the northwest extremity of the Santaren anticline, one $M_w$ 2.7 event on November 16, 1984 located 42 kilometers south-southwest of the northern end of the anticline, and one $M_w$ 2.3 earthquake on August 18, 1990 on its southeast end. However, given that nothing is known about the location accuracy of these earthquakes and published statements (Reference 489) regarding the inadequacy of Cuba earthquakes for the purpose of identifying seismogenic structures (Subsection 2.5.2.4.4.3.2.1), there does not appear to be a spatial association between historical seismicity and the Santaren anticline.

### Straits of Florida Normal Faults

A series of short, steep normal faults exist in the western Straits of Florida southwest of Turkey Point (Reference 480) (Figure 2.5.1-229). These faults are mapped using seismic data in Paleocene and Eocene strata and are buried by undeformed Miocene and younger strata (Figures 2.5.1-209 and 2.5.1-273). This faulting represents syntectonic deformation of the Cuban foreland basin during its collision with the Florida-Bahama Platform (References 794 and 482). Seismic studies in central Straits of Florida indicate that Paleocene to Eocene strata dip to the south indicating the flexure of the southern margin of the Bahama Platform in response to loading from the Cuban orogeny (Reference 221). These syntectonic Paleocene and Eocene strata are terrigenous and were shed directly from Cuba into northward tapering wedges observed in seismic data (Figure 2.5.1-209). In contrast, the late middle Eocene to early middle Miocene strata were deposited uniformly over most of the southern straits of Florida, with pelagic to hemipelagic sedimentation, indicating that the Straits of Florida had subsided to 'near-modern' depths with a change in tectonic regime. The development of sediment drifts in Middle Miocene and younger strata reveal increased current strength in the Straits of Florida at this time (Reference 221). Just outside of the site region, but in a comparable tectonic environment in the southeastern Gulf of Mexico, interpretation of seismic lines indicate that generally no major displacements affect strata above an upper Eocene unconformity (Reference 482), and lines in the site region indicate unfaulted strata above the late middle Eocene unconformity (Reference 221). However, cases of later Tertiary reactivation of faults in the area have been documented (Reference 484).

Also in the Straits of Florida, initial workers hypothesized faulting along the edges of the Pourtales and Miami terraces and along other seafloor escarpments, but also suggested that the escarpments could be original sedimentary features associated with sediments deposited against the steeper face of old reef fronts (Reference 967) (Figure 2.5.1-379). Higher resolution, more detailed seismic imaging has allowed the Pourtales escarpment and similar steep-sided escarpments throughout the Gulf of Mexico, Straits of Florida, and Bahamas to be recognized as relict carbonate platform margins, sometimes steepened and modified by erosion, with drifts of younger sediment resting adjacent (Figure 2.5.1-380) (References 687, 951, and 968). For example, Mullins and Neuman (Reference 968) conclude that there is no evidence for faulting at the eastern edge of the Miami

terrace and that truncated reflectors near the surface indicate erosion was responsible for the observed stratigraphic variations.

South of the Straits of Florida normal faults, thrust faults are expected within a narrow apron offshore of the Cuban coastline. These thrusts, such as the Nortecubana fault, are discussed as part of the Cuban fold-and-thrust belt.

### Cuban Fold and Thrust Belt

North American passive margin strata are deformed in a series of north-vergent imbricate thrusts and anticlines along the northern edge of Cuba (Figures 2.5.1-248, 2.5.1-251, 2.5.1-252, 2.5.1-279, 2.5.1-280, and 2.5.1-281). These faults and folds are exposed onshore, particularly in western Cuba, but imaged with seismic data offshore, within about 20 miles (32 kilometers) of the Cuban coastline (References 221, 484, and 485) (Figure 2.5.1-248). Syntectonic strata of foreland and piggyback basins are well dated onshore and indicate that the thrust faulting is Eocene in age (References 220, 485, and 439). In two offshore seismic lines, Reference 497 indicates that north-vergent thrusts terminate either above an Upper Cretaceous horizon (Figure 2.5.1-281) or just below a Tertiary horizon (Figure 2.5.1-280). Based on a series of north-northeast-trending seismic lines extending north from the Cuban shoreline in the Straits of Florida, Moretti et al. (Reference 484) conclude that the foreland fold and thrust belt developed in the Eocene and indicate that post-tectonic Tertiary and Quaternary sediments are undeformed by the thrusts. For example, in Figure 2.5.1-287, seismic horizons are not traced near the imbricate thrusts, but the faults terminate upward between 0.3 and 0.7 seconds below the seafloor (two-way travel time). Moretti et al. (Reference 484) do note occasional Miocene reactivations of either the early Tertiary thrusts or Jurassic normal faults. On the basis of well-dated Eocene syntectonic strata (References 220, 439, and 485) and published structural interpretations indicating unfaulted Quaternary strata above these structures offshore (References 484 and 485), these faults are concluded to be Tertiary in age and not capable tectonic structures. This age determination is also in agreement with published summaries of the tectonic evolution of Cuba (References 217and 440). Moreover, recent studies of the marine Substage 5e terrace that formed approximately 122 ka preserved on Cuba's north coast between Matanzas and Havana are consistent with the lack of ongoing or recent tectonic uplift (References 920 and 925).

While most investigators confine the extent of the Cuban fold and thrust belt near the north shore of Cuba as described above, others have suggested that the relatively more subdued folding and faulting found to the northeast onto the Bahama Platform represents an extension of the Cuban fold and thrust belt (Masaferro et al. [References 426 and 479] Bergman [Reference 906]). Structural features such as the Santaren anticline (Masaferro et al. [References 426 and 479]) and reverse faults interpreted by Bergman (Reference 906) in the southern and central Santaren Channel are taken as an indication of a greater northeastern extent for the Cuban fold and thrust belt-style of deformation.

### Submarine Surficial Slumps

Marine seismic reflection data have recognized evidence for gravity-driven slumping of surficial strata in the site region. Shallow slumps have been identified along the margin of the Little Bahama Bank, in Exuma Sound, (Reference 476) and in the southeastern Gulf of Mexico (Reference 482). These gravitational features are generally confined to submarine valleys or escarpments (Reference 476). Evidence for submarine landslides in and around the Bahama Platform is discussed in more detail in Subsection 2.5.1.1.1.1.2.

### Seismicity of the Bahama Platform

The Phase 1 earthquake catalog (Subsection 2.5.2.1.2) indicates sparse seismicity within the Bahama Platform (Figure 2.5.1-267). Earthquakes within the Bahama Platform are widely distributed, the largest being an Emb 4.3 earthquake that occurred near Ackins Island, approximately 700 kilometers (430 miles) southeast of the Units 6 & 7 site. Two earthquakes are located northeast of the

site at distances of 53 and 175 kilometers (33 and 109 miles) with Emb 2.7 and 3.2, respectively. About a dozen earthquakes are located about 600 kilometers to the southeast in the vicinity of the central portion of the Bahamas Islands. The dates of these earthquakes range from 1894 to 2007, suggesting that this is a zone of low-level but persistent activity. Ten earthquakes are located within a few tens of kilometers of the northern coastline of Cuba. The overall seismicity pattern within the Bahama Platform shows no correlation with geologic or tectonic features (Subsection 2.5.2.3).

#### 2.5.1.1.1.3.2.3 Continental Slope and Rise

##### Structures of the Continental Slope and Rise

The site region includes a small corner of the Blake Plateau, the intermediate depth plateau just north of Little Bahama Bank (Figure 2.5.1-229). North of the Little Bahama Bank, the Atlantic Continental Shelf extends seaward from the shoreline to a steeper continental slope, located approximately 50 miles (80 kilometers) offshore. This slope has been in existence since the Eocene, but prior to that, the Florida Platform and Blake Plateau were continuous. Seaward, the Blake Plateau extends up to 300 kilometers (185 miles) to the Blake Escarpment, the steep transition to deep ocean basin (Reference 487) (Figure 2.5.1-283). It is east of the Blake Escarpment, where the Blake Spur magnetic anomaly likely represents a transition to oceanic crust, rather than rifted continental material that underlies the Florida Platform, Bahama Platform, and Blake Plateau (Reference 409) (Figure 2.5.1-229).

The plateau is dominantly underlain by Jurassic to Cretaceous carbonates (Reference 307) (Figure 2.5.1-284). The Jurassic and younger strata of the Blake Plateau are generally flat and unfaulted, but Paull and Dillon (Reference 487) identify minor faulting on the Blake Plateau, beyond the site region. This minor faulting is characterized as vertical normal faults exhibiting throws of less than 10 meters that do not affect beds younger than Cretaceous, and are interpreted to be the result of sediment compaction (Reference 487) (Figure 2.5.1-285). In addition, shallow slumps or other gravity-driven faulting has occasionally been noted in the Blake Plateau (Reference 487).

##### Seismicity of the Continental Slope and Rise

The Phase 1 earthquake catalog (Subsection 2.5.2.1.2) indicates sparse, low-magnitude seismicity in the Atlantic Continental Shelf and Slope region, north of the Bahama Platform (Figure 2.5.1-267). According to the updated Phase 1 earthquake catalog, the two earthquakes in the Atlantic Continental Slope and Rise that are nearest to the Units 6 & 7 site are the June 3, 2001, Emb 3.30 and the June 11, 2001, Emb 3.30 earthquakes, at distances of approximately 310 and 330 miles (500 and 530 kilometers) from the site, respectively.

#### 2.5.1.1.1.3.2.4 Cuba

This subsection discusses available geological and geophysical information pertaining to seismic hazard characterization for Cuba. While only a small portion of northern Cuba is within the site region, a discussion of the regional structures on the entire island is presented. Within the past ten years, international groups have published research conducted in Cuba, though many of these concentrate on geochemistry of the arc-related rocks (e.g., Reference 488), rather than any potential recent faulting or seismicity. From a seismic hazard perspective, potential seismic sources in Cuba are summarized by Garcia et al. (References 489 and 490) and Cotilla-Rodríguez (Reference 494) to support seismic hazard mapping.

The major geologic units and their stratigraphic relations are described in Subsection 2.5.1.1.1.2.3. The plate tectonic history of Cuba and the northern Caribbean, including the origin and emplacement timing of the geologic units, are discussed in Subsection 2.5.1.1.3.

### Structures of Cuba

Most regional faults in Cuba, particularly in northern Cuba, are north-directed thrusts or east- to northeast-striking strike-slip faults responsible for transferring the Cretaceous Greater Antilles Arc onto the Bahama Platform (Figures 2.5.1-247, 2.5.1-250, and 2.5.1-251). The Oriente fault zone, located directly off the southern coast of the island, forms the boundary between the modern North America Plate and the Gonâve microplate and is a capable tectonic source. The Oriente fault zone is discussed further in Subsection 2.5.1.1.2.3.1.2, and its characterization in the Cuba and northern Caribbean seismic source model is described in Subsection 2.5.2.4.4.3.

In an effort to explain seismicity that continues on intraplate Cuba, 12 faults on the island of Cuba are designated by Cotilla-Rodriguez et al. (Reference 494) as “active” based on their ambiguous definition of the term. For many faults in intraplate Cuba, the Cotilla-Rodriguez et al. (Reference 494) analysis does not provide sufficient information to conclude that a structure is a capable tectonic source according to RG 1.208. Table 2.5.1-204 provides a summary of these and other regional fault zones of Cuba. Available geologic and tectonic maps are 1:250,000 (Reference 846) and 1:500,000 scale (References 848 and 847) and therefore do not have sufficient detail to properly characterize fault activity based on map relations alone. Available information for the regional Cuban faults that extend to within the site region, and several that lie beyond it, is summarized below.

In the Cuba region, it is acknowledged that seismicity that has not undergone the declustering process, as well as smaller magnitude earthquakes with $M_w$ less than 3, could be used in analyzing or identifying potentially active structures. Therefore, Figure 2.5.1-368 as well as the discussions of individual faults include both these dependent earthquakes (i.e., foreshocks, aftershocks, and cluster events) and the earthquakes of smaller magnitude.

Baconao Fault

The Baconao fault is a northwest-striking fault located in southeastern Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 3). At its nearest point, the Baconao fault is approximately 530 kilometers (330 miles) from the Turkey Point Units 6 & 7 site. Garcia et al. (Reference 489) provide only minimal discussion of this fault but describe it as “better defined in its eastern part, where it has a clear expression mainly in relief and significant seismic activity at the intersection with the (Oriente fault zone).”

Cotilla-Rodríguez et al. (Reference 494) characterize the Baconao fault as active, based on geologic map relations, geomorphology, and its possible association with seismicity. Cotilla-Rodriguez et al. (Reference 494) describe the Baconao fault as “normal and reverse type with left strike-slip.” Cotilla-Rodriguez et al. (Reference 494) note that, along the easternmost portion of the fault near the modern plate boundary, there are “vast, continuous and abrupt escarpments and many distorted and broken fluvial terraces of the Quaternary and Pleistocene.” These observations, coupled with the proximity to the modern plate boundary (i.e., Oriente fault, Figure 2.5.1-247), suggest that the eastern portion of the Baconao fault may be Quaternary active.

Cotilla-Rodriguez et al. (Reference 494) list five earthquakes that they suggest may have occurred on the Baconao fault, all of which occurred between 1984 and 1987. Each of these five earthquakes is assigned Medvedev-Sonheuer-Karnik (MSK) intensity IV (approximate Modified Mercalli Intensity [MMI] IV) (Reference 494). As shown on Figure 2.5.1-368 Sheet 3, however, there is little to no seismicity from the Phase 2 earthquake catalog at the $M_w$ 3 level and greater along much of the length of the Baconao fault, especially along the northwestern two-thirds of its length northwest of the intersection of the Nipe fault. However, several $M_w < 3$ events are seen along the central and northwestern part of the fault. Subject to the limitations in location accuracy stated in Subsection 2.5.2.4.4.3.2.1, these may or may not indicate seismic activity on the fault. Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

The Baconao fault is not shown on Case and Holcombe's (Reference 480) 1:2,500,000 scale map of the Caribbean. Perez-Othon and Yarmoliuk (Reference 848), however, show an unnamed, dashed fault on their 1:500,000 scale geologic map of Cuba. This unnamed fault is located in the vicinity of the Baconao fault and is depicted cutting Oligocene-Miocene strata, but covered by apparently unfaulted mid-Quaternary-age strata (Reference 848). According to mapping by Perez-Othon and Yarmoliuk (Reference 848), the Baconao fault appears to be offset in a right-lateral sense by two strands of the northeast-striking Nipe fault. As an inset to their geologic map, Perez-Othon and Yarmoliuk (Reference 848) provide an additional map that shows their estimates of fault ages in Cuba. A modified version of their inset map is provided as Figure 2.5.1-369. The inset map presented in Figure 2.5.1-369 was modified by enhancing the color-coding of the Perez-Othon and Yarmoliuk (Reference 848) age estimates and by adding fault name labels based on their relative locations. Most of the fault name labels added to the inset map are queried, however, indicating the uncertainty regarding which faults are, and which are not, shown on the inset map. If the unnamed fault depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba represents the Baconao fault, as is assumed on Figure 2.5.1-369, then they indicate a Neogene-Quaternary age for the southeastern one-third of the Baconao fault. The northwestern two-thirds of the Baconao fault as shown on Figure 2.5.1-368 Sheet 3 does not clearly appear on Perez-Othon and Yarmoliuk's (Reference 848) inset map (Figure 2.5.1-369).

The Nuevo Atlas Nacional de Cuba includes a 1:1,000,000 scale geologic map of Cuba (Reference 944, plate III.1.2-3) and a 1:2,000,000 scale neotectonic map of Cuba (Reference 944, plate III.2.4-8). No fault names appear on these two maps so it is not clear whether the Baconao fault is shown. The geologic map of Cuba from this atlas shows an approximately 50-kilometer-long (30-mile-long), northwest-striking fault near Santiago de Cuba that may be the Baconao fault, but this fault is restricted to southernmost Cuba, southeast of the Nipe fault. This fault appears to cut middle Eocene strata. Likewise, the neotectonic map of Cuba from this atlas shows an approximately 75-kilometer-long (45-mile-long), northwest-striking fault in the same area of southernmost Cuba that could be the Baconao fault. The Baconao fault is depicted and labeled on the 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11). The Baconao fault is shown and labeled on Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba.

Camaguey Fault

The Camaguey fault is a northeast-striking fault located in southeastern Cuba (Figures 2.5.1-247, 2.5.1-251, 2.5.1-368 Sheet 2, and 2.5.1-368 Sheet 3). At its nearest point, the Camaguey fault is approximately 530 kilometers (330 miles) from the Turkey Point Units 6 & 7 site. Garcia et al. (Reference 489) describe the Camaguey fault as a “regional transverse fault with lateral displacement that affects the whole crust and constitutes the boundary between two megablocks” and that “cuts young as well as old sequences.” In their Figure 5, Garcia et al. (Reference 489) show the Camaguey fault as a normal fault with unspecified dip direction and sense of throw. Garcia et al. (Reference 489) also note that “the gravimetric and magnetic fields show apparent inflections.”

Cotilla-Rodríguez et al. (Reference 494) classify the Camaguey fault as active based on the possible association of seismicity with the fault. Cotilla-Rodriguez et al. (Reference 494) describe the Camaguey fault as a sinistral strike-slip fault with an almost vertical plane associated with a low “level of seismic activity.” They list ten earthquakes that they suggest may have occurred on the Camaguey fault. Three of these earthquakes are assigned MSK intensity III–IV (approximately MMI III–IV), with the remaining seven unspecified (Reference 494). As shown on Figures 2.5.1-368 Sheets 2 and 3, however, there is little to no seismicity from the Phase 2 earthquake catalog located along the length of the Camaguey fault, with the possible exception of a single, minor-magnitude earthquake near the northeastern end of the fault. Alternatively, this minor earthquake may be associated with the northwestern end of the Baconao fault or some other unmapped structure (Figure 2.5.1-368 Sheet 3). Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

The Camaguey fault is not consistently shown on geologic and tectonic maps of Cuba. For example, it is not labeled on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba, Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba, the Nuevo Atlas Nacional de Cuba 1:1,000,000 scale geologic map (Reference 944, plate III.1.2-3), and van Hinsbergen et al.'s (Reference 500) mapping of the Camaguey area. The Camaguey fault is depicted and labeled on the 1:2,000,000 scale lineament map from the national atlas (Reference 944, plate III.3.1-11) and shown but not labeled on the 1:2,000,000 scale neotectonic map from the same atlas (Reference 944, plate III.2.4-8). Because they do not label faults by name, it is not clear whether the Camaguey fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Paleogene age for an unnamed fault in the vicinity of the Camaguey fault (Figure 2.5.1-369).

Cochinos Fault

The Cochinos fault is a north- (References 494 and 770) to north-northwest-striking (Reference 493) fault in south-central Cuba. Figures 2.5.1-247, 2.5.1-368 Sheet 1, and 2.5.1-368 Sheet 2 show the location of the Cochinos fault after Hall et al. (Reference 770). As mapped by Hall et al. (Reference 770), the fault at its nearest point is approximately 330 kilometers (205 miles) from the Turkey Point Units 6 & 7 site. Alternatively, mapping by Cotilla-Rodriguez et al. (Reference 494) suggests this fault may extend northward to within 280 kilometers (175 miles) of the site, whereas mapping by Mann et al. (Reference 493) indicates a closest distance of approximately 340 kilometers (210 miles). The Cochinos fault is the only onshore feature in intraplate Cuba identified as “neotectonic” by Mann et al. (Reference 493) (Figure 2.5.1-286). They map the Cochinos fault as two parallel, north-northwest-striking normal faults that form a graben (Figures 2.5.1-286, 2.5.1-368 Sheet 1, and 2.5.1-368 Sheet 2). The morphology of Bahia de Cochinos is consistent with this interpretation and suggests the possibility of fault control on the landscape.

Cotilla-Rodriguez et al. (Reference 494) describe the Cochinos fault as a “normal fault with a few inverse type sectors which demonstrates transcurrence to the left” and “normal and reverse type with left strike-slip.” Recorded seismicity near the Cochinos fault is sparse. They list six earthquakes that they suggest may have occurred on the Cochinos fault. The largest of these is the December 16, 1982 $M_S$ 5.0 earthquake. The Phase 2 earthquake catalog developed for the Turkey Point Units 6 & 7 site does not include an earthquake on that date with similar magnitude and location. The Phase 2 earthquake catalog does, however, include an $M_w$ 5.4 earthquake near the Cochinos fault that occurred on November 16, 1982 (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Based on the similarity in location, magnitude, and year for the December 16 and November 16 earthquakes, it is assumed that these are the same earthquake and that the discrepancy in month is the result of a typographical error in Cotilla-Rodríguez et al.'s (Reference 494) manuscript. The remaining five earthquakes that Cotilla-Rodriguez et al. (Reference 494) associate with the Cochinos fault “are all of low [and unspecified] intensity.” In the Phase 2 earthquake catalog, the 1982 earthquake is located approximately 5 kilometers (3 miles) northwest of the Cochinos fault trace (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Cotilla-Rodriguez et al. (Reference 494) suggest that the 1982 earthquake may instead have occurred on the Habana-Cienfuegos fault. In addition to the 1982 earthquake, the Phase 2 earthquake catalog shows only four other earthquakes within 32 kilometers (20 miles) of the Cochinos fault, the largest of which is assigned $M_W$4.1 (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Cotilla- Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

Cotilla-Rodríguez et al. (Reference 494) classify the Cochinos fault as active based on the possible association of seismicity with the fault. Cotilla-Rodríguez et al. (Reference 494) provide no geologic evidence for activity on the Cochinos fault and describe the fault as “covered by young sediments.” Indeed, the most detailed geologic maps inspected in the area (1:250,000 scale) show no fault cutting Miocene and younger strata (Reference 846). Because they do not label faults by name, it is not clear whether the Cochinos fault is depicted on Perez-Othon and Yarmoliuk’s (Reference 848) inset map of fault ages in Cuba, but they indicate a Paleogene age for a northern extension of this

fault (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba shows and labels the approximately 100-kilometer-long (60-mile-long) Cochinos fault. The southern approximately 80 kilometers (50 miles) of this fault are shown as a dashed line. Garcia et al. (Reference 489) provide no discussion of the Cochinos fault.

The Cochinos fault is depicted differently on various maps from the Nuevo Atlas Nacional de Cuba (Reference 944). The 1:1,000,000 scale geologic map of Cuba from this atlas (Reference 944, plate III.1.2-3) shows an approximately 140-kilometer-long (87-mile-long) unnamed fault in the vicinity of the Cochinos fault that extends from Cuba's northern coast where it is mapped in Pliocene-age deposits southward into the Bahia de Cochinos. The southernmost 30 kilometers (18 miles) of this fault are shown by a dashed line. The 1:2,000,000 scale neotectonic map of Cuba from this atlas (Reference 944, plate III.2.4-8) shows an approximately 140-kilometer long (87-mile-long) unnamed fault in the vicinity of the Cochinos fault, the southernmost 50 kilometers (30 miles) of which is offshore southern Cuba and shown by a dashed line. To the north, this fault on the neotectonic map is truncated by the Hicacos fault. The Cochinos fault is depicted and labeled on the 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11). The 1:1,000,000 scale geomorphic map from the Nuevo Atlas Nacional de Cuba (Reference 944, plate IV.3.2-3) shows an approximately 60-kilometer-long (37-mile-long) unnamed fault in the vicinity of the Cochinos fault. The map explanation indicates that this fault cuts a Quaternary-age marine abrasion platform that is at an elevation of either 2– 3 meters (6.6-9.8 feet) or 5–7 meters (16.4-23 feet) above sea level. They do not provide explanation for the lack of specificity in elevation of the platform nor do they provide a precise age for the Quaternary abrasion platform.

Cubitas Fault

The Cubitas fault is a northwest-striking, steeply south-dipping fault located in southeastern Cuba (Figures 2.5.1-247, 2.5.1-368 Sheet 2, and 2.5.1-368 Sheet 3). At its nearest point, the Cubitas fault is approximately 435 kilometers (270 miles) from the Turkey Point Units 6 & 7 site. Garcia et al. (Reference 489) describe the Cubitas fault as a "deep fault that constitutes a portion of the Cuban marginal suture and is considered to be the main structure in central Cuba. It is cut by the Camaguey and the La Trocha transverse faults, where seismicity is documented." They associate the 1974 $M_s$ 4.5 MSK VII Esmeralda earthquake (month and day unspecified) with the Cubitas fault.

Cotilla-Rodríguez et al. (Reference 494) characterize the Cubitas fault as active based on its possible association with seismicity. Cotilla-Rodriguez et al. (Reference 494) describe the Cubitas fault as "an almost vertical normal fault with some sectors of inverse type" and as "normal and reverse type." They describe large scarps associated with this fault but do not provide additional descriptions of the scarps. They assign a Pliocene to Quaternary age for this fault. Cotilla-Rodriguez et al. (Reference 494) list 15 earthquakes that they suggest may have occurred on the Cubitas fault. Eight of these earthquakes are assigned MSK intensity III–V (approximately MMI III–V), with the remaining seven unspecified (Reference 494). The Phase 2 earthquake catalog includes several low-magnitude earthquakes that may be spatially associated with the northwestern half of the Cubitas fault (Figures 2.5.1-368 Sheet 2 and 2.5.1-368 Sheet 3). The central and southeastern portions of the fault appear largely devoid of seismicity. The Phase 2 earthquake catalog indicates $M_w$ 4.0 and $M_w$ 5.1 earthquakes occurred approximately 24 kilometers (15 miles) south of the mapped trace near the northwestern end of the fault in 1974 and 1984, respectively, which may be associated with the Cubitas fault. Cotilla-Rodriguez et al (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

Van Hinsbergen et al. (Reference 500) describe the Cubitas fault as a post-Middle Eocene, south-dipping normal fault that forms a steep slope along the southern margin of the Cubitas Hills. They describe approximately 200 meters (650 feet) of uplift associated with the Cubitas Hills that post-dates deposition of Pliocene-Pleistocene (?) fluvial deposits north of the hills. If this interpretation is correct, then this uplift may have occurred in the hanging wall of the Cubitas fault, which may be Quaternary-active (Reference 500).

Pushcharovskiy et al. (Reference 846) do not label the Cubitas fault on their 1:250,000 scale geologic map. Pushcharovskiy (Reference 847) shows the Cubitas fault as an approximately 85 kilometers long (50-mile-long), south-dipping thrust fault on the 1:500,000 scale tectonic map. Because they do not label faults by name, it is not clear whether the Cubitas fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Mesozoic age for an unnamed fault in the vicinity of the Cubitas fault (Figure 2.5.1-369).

The Cubitas fault does not appear on the 1:1,000,000 scale geologic map of Cuba from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3), but seemingly does appear as an unnamed fault on the 1:2,000,000 scale neotectonic map from this same atlas (Reference 944, plate III.2.4-8). The 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11) labels an approximately 85-kilometer-long (50-mile-long) feature as the Cubitas fault.

Domingo Fault

At its nearest point, the low-angle Domingo fault is located 282 kilometers (175 miles) south of the Turkey Point Units 6 & 7 site. This northwest-striking, south-dipping thrust fault carried the Cretaceous arc and serpentinites over the carbonate platform rocks and can be considered the former suture between North America and Caribbean plates (References 439 and 440) (Figure 2.5.1-247). The Domingo fault does not cut the uppermost Eocene and younger sedimentary units, and is late Eocene in age (References 439 and 440). A myriad of other thrusts are mapped in detail (though not shown Figure 2.5.1-247), which imbricate both the autochthonous and allochthonous units on the island (Reference 439). On 1:250,000 scale maps and interpreted cross sections, these faults also do not cut the uppermost Eocene and younger deposits, and so are not Quaternary in age (References 439, 440, 497, and 846) (Figure 2.5.1-248).

Guane Fault

The subsurface Guane fault is a northeast-striking fault in western Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 1). At its nearest point, the Guane fault is approximately 370 kilometers (230 miles) from the Turkey Point Units 6 & 7 site. Garcia et al. (Reference 489) provide no discussion of the Guane fault.

Cotilla-Rodríguez et al. (Reference 494) characterize the Guane fault as active based on its possible association with seismicity. Cotilla-Rodriguez et al. (Reference 494) describe he Guane fault as a “large and complex structure totally covered by young sediments in the Palacios Basin” that is “predominantly vertical with left transcurrence.” They list 19 earthquakes that they suggest may have occurred on the Guane fault, many of which are listed by year only without month, day, intensity, and magnitude information. The largest of these is the January 23, 1880 $M_w$ 6.1 San Cristobal earthquake. In the Phase 2 earthquake catalog, seismicity in the vicinity of the Guane fault is sparse, but other light- to-moderate magnitude earthquakes within 32 kilometers (20 miles) of the fault include the May 20, 1937 $M_w$ 5.1, December 20, 1937 $M_w$ 5.1, October 12, 1944 $M_w$ 4.0, and September 11, 1957 $M_w$ 4.0 earthquakes (Figure 2.5.1-368 Sheet 1). Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

Based on their review of aerial photographs and satellite imagery, Cotilla-Rodriguez and Cordoba-Barba (Reference 942) note two rivers in the Palacios Basin (Bayate and San Cristobal rivers) that show, in plan view, what they call “fluvial inflections” that they interpret as the result of surface deformation associated with the Guane fault. Cotilla-Rodriguez and Cordoba-Barba (Reference 942) indicate this allows for “the identification of an SW-NE alignment on the south plain of Pinar del Rio, corresponding to the Guane fault, whith *[sic]* was responsible for the San Cristobal earthquake on the 28.01.1880.” However, other rivers along strike to the northeast and southwest do not appear to show such inflections. Moreover, Cotilla-Rodriguez et al. (Reference 494) indicate the Guane fault is “totally covered by young sediments in the Palacios Basin.” Likewise, Cotilla-Rodriguez and Cordoba-Barba (Reference 943) indicate the Guane fault “is located under

ample thicknesses of sediments of the plain in southern Pinar del Rio." The Cotilla-Rodríguez et al. (Reference 494) and Cotilla-Rodriguez and Cordoba-Barba (Reference 943) studies do not specify a burial depth for the Guane fault, but seemingly are at odds with Cotilla-Rodriguez and Cordoba-Barba's (Reference 942) interpretation of surface manifestation of deformation.

Cotilla-Rodriguez and Cordoba-Barba (Reference 943) describe historical accounts of the January 23, 1880 earthquake, including first-hand observations of earthquake damage in San Cristobal, Candelaria, and elsewhere in the region. They note that the most severe and concentrated damage was located not in the mountainous regions of the Sierra del Rosario and Sierra de los Organos near the Pinar fault (discussed below) but rather within the Palacios Basin near the Guane fault. Cotilla-Rodriguez and Cordoba-Barba (Reference 943) cite this as evidence that the 1880 earthquake occurred on the Guane fault. Cotilla-Rodriguez and Cordoba-Barba (Reference 943) conclude that the Pinar fault "is not the seismogenetic element of the January 23, 1880 earthquake" and that it is "subordinate to" the Guane fault. Alternatively, however, the pattern of 1880 damage could be explained by possible focusing of seismic waves within the basin, possible hanging-wall focusing effects, possible liquefaction, or possible differences in population density and building styles. In other words, the pattern of 1880 damage is not conclusive evidence that the earthquake occurred on the Guane fault, as opposed to on the Pinar fault or other structure.

The Guane fault is not depicted on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba. Perez-Othon and Yarmoliuk (Reference 848) show an unnamed, dashed fault on their 1:500,000 scale geologic map of Cuba in the vicinity of the Guane fault that cuts Miocene strata, but is covered by unfaulted Pliocene-Pleistocene units. Because they do not label faults by name, it is not clear whether the Guane fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Paleogene age for an unnamed fault in the vicinity of the Guane fault (Figure 2.5.1-369). The Guane fault does not seem to appear on any maps in the Nuevo Atlas Nacional de Cuba (Reference 944).

Habana-Cienfuegos Fault

The Habana-Cienfuegos fault is a northwest-striking, left-lateral strike-slip fault in western and central Cuba (Figures 2.5.1-247, 2.5.1-368 Sheet 1, and 2.5.1-368 Sheet 2). At its nearest point, the Habana-Cienfuegos fault is approximately 355 kilometers (220 miles) from the Turkey Point Units 6 & 7 site. Cotilla-Rodriguez et al. (Reference 494) map the Habana-Cienfuegos fault as extending offshore in northern Cuba, where it terminates at or south of the Nortecubana fault, with which it forms a "morphostructural knot" (Reference 494) (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Offshore of southern Cuba, the Habana-Cienfuegos fault is shown as intersected and terminated by the Surcubana fault in a similar "morphostructural knot" (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2, and Figure 5 of Reference 494). Cotilla-Rodriguez et al. (Reference 494) indicate that the Habana-Cienfuegos fault is expressed in the topography in the northwest at Havana Bay and in the southeast at Cienfuegos Bay.

Garcia et al. (Reference 489) provide minimal discussion of the Habana-Cienfuegos fault. Garcia et al. (Reference 489) indicate "although the earthquakes reported in Havana and some locations of its province cannot be attributed to the western portion of the Norte Cubana seismic region, the seismic activity of the Havana fault system is still under debate." Further to the southeast, Garcia et al. (Reference 489) indicate that the Cienfuegos fault "coincides with a deep fault located under younger tectonic sequences, it does not have a well-defined character."

In the Phase 2 earthquake catalog, seismicity is sparse in the vicinity of the Habana-Cienfuegos fault (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Cotilla-Rodriguez et al. (Reference 494) list nineteen earthquakes that they suggest may have occurred on the Habana-Cienfuegos fault, many of which are listed by year only without month, day, intensity, and magnitude information. The largest of these earthquakes is the December 16, 1982 $M_S$ 5.0 earthquake. The Phase 2 earthquake catalog developed for the Turkey Point Units 6 & 7 site does not include an earthquake on that date with

similar magnitude and location. The Phase 2 earthquake catalog does, however, include an $M_W$ 5.4 earthquake near the Cochinos fault that occurred on November 16, 1982 (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). Based on the similarity in location, magnitude, and year for the December 16 and November 16 earthquakes, it is assumed that these are the same earthquake and that the discrepancy in month is the result of a typographical error in Cotilla-Rodríguez et al.'s (Reference 494) manuscript. In the Phase 2 earthquake catalog, this earthquake is located approximately 11 kilometers (7 miles) north of the Habana-Cienfuegos fault trace (Figure 2.5.1-368 Sheet 1). Cotilla-Rodriguez et al. (Reference 494) alternatively suggest that this earthquake may have occurred on the Cochinos fault instead. They also associate an $M_S$ 2.5 earthquake and nine MSK intensity III–V earthquakes (approximately MMI III–V) with the Habana-Cienfuegos fault. Cotilla-Rodriquez et al. (Reference 494) suggest that the March 9, 1995, $M_S$ 2.5 earthquake could have occurred on the Habana-Cienfuegos fault or on the nearby Guane fault. Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

The Habana-Cienfuegos fault is not shown on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba and Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba. Because they do not label faults by name, it is not clear whether the Habana-Cienfuegos fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Paleogene age for an unnamed fault in the vicinity of the Habana-Cienfuegos fault (Figure 2.5.1-369).

The 1:1,000,000 scale geologic map of Cuba from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3) shows an approximately 40-kilometer-long (25-mile-long) unnamed fault near Havana in the vicinity of the northwestern-most portion of the Habana-Cienfuegos fault as shown on Figure 2.5.1-368 Sheet 1. Similarly, the 1:2,000,000 scale neotectonic map of Cuba from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.2.4-8) shows an approximately 60-kilometer-long (37-mile-long) unnamed fault in the same vicinity, the southeastern 20 kilometers (12 miles) of which is shown as a dashed line. Neither of these maps from the Nuevo Atlas Nacional de Cuba (Reference 944, plates III.2-3 and III.2.4-8) shows a fault extending from Havana southeastward to the southern coast of Cuba, as shown by Cotilla-Rodríguez et al. (Reference 494).

Hicacos Fault

The Hicacos fault is an east-northeast-striking fault in north-central Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 1). At its nearest point, the Hicacos fault is approximately 250 kilometers (155 miles) south of the Turkey Point Units 6 & 7 site. Based on mapping by Cotilla-Rodriguez et al. (Reference 494), the Hicacos fault is the nearest fault in Cuba to the site identified as active by these authors. Some publications (Reference 769) refer to this fault as the Matanzas fault.

Garcia et al. (Reference 489) provide minimal discussion of the Hicacos fault. They indicate it is “a deep fault above Paleocene-Quaternary formations, splitting the ophiolites sequence that makes the main Cuban watershed deviate abruptly, causing different types of fluvial networks.” Garcia et al. (Reference 489) state that the “earthquakes reported in Matanzas and more recently in the Varadero-Cardenas area are associated with this structure.” They provide no additional information regarding these earthquakes.

Cotilla-Rodríguez et al. (Reference 494) characterize the Hicacos fault as active based on its possible association with seismicity. Cotilla-Rodriguez et al. (Reference 494) describe the Hicacos fault as a “normal fault, transcurrent to the left” that is “expressed throughout the Peninsula de Hicacos and is internal in the island territory by the eastern edge of Matanzas Bay, delineating very well the Matanzas Block.” Further to the west-southwest, Cotilla-Rodriguez et al. (Reference 494) indicate that the Hicacos fault is “weakly represented” in the geomorphology.

Seismicity in the vicinity of the Hicacos fault is sparse (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). The nearest epicenters from the Phase 2 earthquake catalog to the Hicacos fault are four

co-located $M_w$ 3.1 to 3.7 earthquakes that occurred near the central portion of the fault in 1812, 1852, 1854, and 1970. A small event from the Phase 2 project earthquake catalog with $M_w < 3$ is seen very close to the southwest terminus of the fault on Figure 2.5.1-368 Sheet 1. Another earthquake occurred in 1777 with $M_w$ 3.7, located on strike with, but approximately 11 kilometers (7 miles) southwest of, the mapped fault trace. Likewise, Cotilla-Rodriguez et al. (Reference 494) indicate sparse seismicity near the Hicacos fault, and note that no focal mechanisms are associated with earthquakes in the vicinity of this fault. According to Cotilla-Rodriguez et al. (Reference 494), historical accounts suggest ten earthquakes of less than or equal to MSK intensity V (approximately MMI V) occurred in the vicinity of the Hicacos fault (Reference 494). However, the association of these earthquakes with the Hicacos fault or another mapped or unmapped fault is problematic due to the uncertainties associated with the locations of both faults and earthquakes in Cuba and the paucity of available focal plane solutions.

Case and Holcombe's (Reference 480) 1:2,500,000 scale map of the Caribbean region shows segments of the Hicacos fault cutting upper Tertiary rocks. Perez-Othon and Yarmoliuk's (Reference 848) 1:500,000 scale geologic map of Cuba shows an unnamed fault in the vicinity of the Hicacos fault that extends from Matanzas for approximately 80 kilometers (50 miles) to the southwest. Because they do not label faults by name, it is not clear whether the Hicacos fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba. They indicate, however, a Mesozoic age for an unnamed fault in the vicinity of the northeastern-most portion of the Hicacos fault (Figure 2.5.1-369). Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba shows an unnamed fault cutting lower Miocene rocks in the vicinity of the central Hicacos fault as shown on Figure 2.5.1-368 Sheet 1, but their mapping does not extend this fault as far northeast as the north coast of Cuba. The locally northeast-trending shoreline and a narrow peninsula near Matanzas are notably linear and on-trend with the fault, likely influencing where the fault is mapped in other representations. Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba shows the northeastern extent of the Hicacos fault similar to the depiction shown in Figure 2.5.1-368 Sheet 1, and terminating to the southwest at Cuba's southern coast.

The Hicacos fault is depicted differently on different maps from the Nuevo Atlas Nacional de Cuba (Reference 944). The 1:1,000,000 scale geologic map from this atlas (Reference 944, plate III.1.2-3) shows an unnamed, northeast-striking, approximately 40-kilometer-long (25-mile-long) fault in the in the vicinity of the Hicacos fault. This unnamed fault is mapped within lower to middle Miocene-age deposits and does not appear to cut Holocene-age deposits near Matanzas at the northeastern end of the fault. The 1:1,000,000 scale geomorphic map from this atlas (Reference 944, plate IV.3.2-3) shows an unnamed fault offshore along the narrow peninsula that may be the Hicacos fault, but this offshore fault does not extend onshore to the southwest. The Hicacos fault is labeled on the lineament map from this atlas (Reference 944, plate III.3.1-11) as an approximately 175-kilometer-long (110-mile-long), northeast-trending feature that extends from near Cuba's south coast, across Cuba, and along the narrow peninsula near Matanzas on Cuba's north coast. On the lineament map, the northeastern-most 35 kilometers (20 miles) of this feature are shown as a dashed line. The 1:2,000,000 scale neotectonic map from this atlas (Reference 944, plate III.2.4-8) shows an unnamed, northeast-striking fault in the vicinity of the Hicacos fault that extends from Cuba's south coast, across Cuba, and along the narrow peninsula near Matanzas, and offshore where it is terminated by an unnamed fault that likely is the Nortecubana fault.

Various researchers describe elevated marine terraces west of Matanzas Bay near the Hicacos fault along Cuba's north coast. Continuous and planar geomorphic surfaces like these can be used as Quaternary strain markers with which to assess the presence of tectonic deformation. Ducloz (Reference 915) and Shanzer et al. (Reference 923) provide observations of Pleistocene-age terraces in this region, including the Terraza de Seboruco terrace, which is currently a few meters above modern sea level. Both Ducloz (Reference 915) and Shanzer et al. (Reference 923) speculate that Pleistocene-age terraces in this region may have formed as the result of both tectonic uplift and global fluctuations in sea level.

More recent studies, however, conclude that tectonic uplift is not required to explain the present elevation of the Pleistocene-age Terraza de Seboruco terrace west of Matanzas Bay and near the Hicacos fault. Toscano et al.'s (Reference 925) radiometric age dating of coral samples collected from the Terraza de Seboruco terrace indicates this surface formed at approximately 120–140 ka. Based on these ages, they associate the Terraza de Seboruco terrace with the global Substage 5e sea level high-stand at approximately 122 ka. Toscano et al. (Reference 925) also observe that this terrace in the Matanzas area is just a few meters above mean sea level, similar to the elevation of other Substage 5e reef deposits throughout “stable” portions of the Caribbean and, therefore, can be explained solely by changes in sea level. Toscano et al. (Reference 925) conclude that “no obvious tectonic uplift is indicated for this time frame along the northern margin of Cuba.” Similarly, Pedoja et al. (Reference 920) investigated late Quaternary coastlines worldwide and observe minor uplift relative to sea level of approximately 0.2 millimeter/year, even along passive margins, outpacing eustatic sea level decreases by a factor of four. They suggest that, when accounting for eustatic changes in sea level, the Substage 5e terrace in the Matanzas area (i.e., the Terraza de Seboruco terrace) has been uplifted at an average rate that ranges from approximately 0.00 to 0.04 millimeters/year over the last approximately 122 ka, consistent with uplift rates observed from other stable margins worldwide. If the effects of eustasy are ignored, Pedoja et al.'s (Reference 920) data allow for an uplift rate at Matanzas of approximately 0.06 millimeter/year over the last approximately 122 ka, following this “conservative” (Reference 920) approach.

Whereas recent studies indicate that tectonic uplift is not required to explain the present elevation of the Terraza de Seboruco terrace west of Matanzas Bay (References 920 and 925), these data do not preclude activity on the Hicacos fault. As described above, the location and extent of the Hicacos fault differs between various geologic maps and published figures, so it is unclear whether the Hicacos fault is overlain by the Terraza de Seboruco terrace. Furthermore, if the sense of slip on the Hicacos fault were primarily strike-slip as opposed to dip-slip, it could be difficult to observe surface manifestation of fault-related deformation on the Terraza de Seboruco terrace.

La Trocha Fault

The La Trocha fault is a northeast-striking fault in central Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 2). At its nearest point, the La Trocha fault is approximately 420 kilometers (260 miles) south of the Turkey Point Units 6 & 7 site. Rosencrantz (Reference 529) maps a northeast-striking structure across the Yucatan basin south of Cuba (Figure 2.5.1-286) and interprets it as the southwestern extension of the La Trocha fault.

Garcia et al. (Reference 489) provide minimal discussion of the La Trocha fault. Garcia et al. (Reference 489) indicate it is a “deep fault more than 180 kilometers (112 miles) long, with neotectonic transcurrent activity” and “its seismicity is documented by the earthquakes in the Santi Spiritus region.” They also indicate that the La Trocha fault is expressed in geophysical data, but they do not elaborate.

Cotilla-Rodríguez et al. (Reference 494) assign the La Trocha fault an age of Pliocene-Quaternary and also suggest a possible association with seismicity. Cotilla-Rodriguez et al. (Reference 494) describe the La Trocha fault as “a fault zone transcurrent to the left with a large angle.” They suggest a possible association between three earthquakes of less than or equal to MSK intensity V (approximately MMI V) and the La Trocha fault. The Phase 2 earthquake catalog shows very sparse seismicity associated with the La Trocha fault (Figure 2.5.1-368 Sheet 2). The largest earthquakes from the Phase 2 earthquake catalog near the La Trocha fault are the March 10, 1952 $M_W$ 4.0 and January 1, 1953 $M_W$ 4.3 events. Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

Leroy et al. (Reference 499) interpret the La Trocha fault as the northern transform limb of a proto-Cayman spreading center that was active in the early Eocene (53 Ma) and was abandoned by

49 Ma. This interpretation is the result of the southward migration of the left lateral strike slip faults that make up the Caribbean-North America plate boundary (Reference 639).

The La Trocha fault is not shown on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba. Review of Pushcharovskiy et al.'s (Reference 846) maps in the vicinity where Cotilla-Rodriguez et al. (Reference 494) map the La Trocha fault indicates no northeast-striking faults cutting Miocene and younger strata. Potentially, this structure is buried by the overlying strata and could be pre-middle Miocene in age. Pushcharovskiy's (Reference 847) tectonic map of Cuba, however, clearly depicts and labels the La Trocha fault with extent and location similar to the La Trocha fault shown in Figure 2.5.1-368 Sheet 2. Because they do not label faults by name, it is not clear whether the La Trocha fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Neogene-Quaternary age for an unnamed fault in the vicinity of the La Trocha fault (Figure 2.5.1-369).

The La Trocha fault is depicted differently on various maps from the Nuevo Atlas Nacional de Cuba (Reference 944). The 1:1,000,000 scale geologic map of Cuba from this atlas (Reference 944, plate III.1.2-3) does not include the La Trocha fault. The 1:2,000,000 scale neotectonic map of Cuba from this atlas (Reference 944, plate III.2.4-8) shows an unnamed fault in the vicinity of the La Trocha fault. This unnamed fault is mapped as terminating northward at the northern coast of Cuba. The 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11) depicts and labels the La Trocha fault as an approximately 150-kilometer-long (90-mile-long), northeast-trending feature that extends from Cuba's southern to its northern coast.

Las Villas Fault

The Las Villas fault is a northwest-striking fault in central Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 2). At its nearest point, the Las Villas fault is approximately 250 kilometers (155 miles) south of the Turkey Point Units 6 & 7 site. Pardo (Reference 439) maps the Las Villas fault as a south-dipping thrust with up to approximately 30 kilometers (18 miles) of horizontal displacement. According to Pardo (Reference 439), the Las Villas fault displaces middle Eocene units, but exhibits greater displacement of older units, indicating that most of its movement was pre-middle Eocene.

Garcia et al. (Reference 489) describe the Las Villas fault as a “deep fault that divides the younger coastal formations of the north from the older ones of the south, it appears as a negative anomaly in the gravimetric map and with positive and negative anomalies in the magnetic field. Medium-magnitude seismicity is associated with this fault.”

Cotilla-Rodríguez et al. (Reference 494) characterize the Las Villas fault as active based on its possible association with seismicity and geomorphic expression. Cotilla-Rodríguez et al. (Reference 494), however, provide only the following minimal description of the Las Villas fault:

> This fault maintains the prevailing strike of the island on the southern part of the Alturas del Norte de Las Villas, from the surroundings of the Sierra Bibanasi to the Sierra de Jatibonico. It is a normal type fault with a large angle, with inverse type sectors. It is intercepted to the east by the La Trocha fault. Its outline has young eroded scarps. It is of Pliocene-Quaternary age. The associated seismic events are: 15.08.1939 ($M_s$ = 5.6), 01.01.1953 (I = 5 MSK), I = 4 MSK, (03.02.1952 and 25.05.1960), 22.01.1983 (I = 3 MSK), and noticeable without specification 04.01.1988.

Cotilla-Rodríguez et al. (Reference 494) do not describe their basis for concluding that the Las Villas fault is Pliocene -Quaternary in age and they do not provide reference to other publications that provide this information. Likewise, Cotilla-Rodríguez et al. (Reference 494) do not provide additional discussion of the “young eroded scarps,” nor do they provide reference to other publications that provide this information. It is not clear from this limited description if these are fault scarps formed directly by recent slip on the Las Villas fault or if they are fault-line scarps formed by recent differential

erosion along the fault trace. It is also possible that these “young eroded scarps” formed by preferential erosion of sheared rocks within the fault zone. Based on the scant information provided in Cotilla-Rodríguez et al. (Reference 494), it is not possible to distinguish between these alternatives. There are no known paleoseismic trench studies or detailed geomorphic assessments of the Las Villas fault with which to assess recent earthquake activity on this fault. Where faults exhibit scarps in young deposits or surfaces, such as the Baconao fault in southernmost Cuba, Cotilla-Rodriguez et al. (Reference 494) provide clear description and do not include “eroded” in the description.

Figure 2.5.1-368 Sheet 2 indicates moderately sparse seismicity from the Phase 2 earthquake catalog that may be roughly aligned with the Las Villas fault, as mapped by Pardo (Reference 439). A few dozen earthquakes are seen in the vicinity of the fault. The largest earthquake near the Las Villas fault is the August 12, 1873 $M_W$ 5.1 earthquake, located approximately 5 kilometers (3 miles) northeast of the fault (Figure 2.5.1-368 Sheet 2). Cotilla-Rodriguez et al. (Reference 494) indicate focal mechanisms for these earthquakes are unavailable, so it is not possible to assess whether these possibly roughly aligned epicenters occurred on the Las Villas fault or on another fault or faults. Cotilla-Rodriguez et al. (Reference 494) suggest that the largest recorded earthquake associated with the Las Villas fault is the $M_S$ 5.6 event on August 15, 1939 (listed in the Phase 2 earthquake catalog as $M_W$ 5.84). Based on the fault mapping of Pardo (Reference 439) and the location of this earthquake from the Phase 2 earthquake catalog, however, this earthquake is located approximately 32 kilometers (20 miles) northeast of this southwest-dipping fault (Figure 2.5.1-368 Sheet 2), suggesting a fault other than the Las Villas ruptured during this event.

Review of geologic mapping (References 480, 846, and 848) reveals that no units of Quaternary age are faulted, but the coarse scale of mapping (1:250,000 to 1:2,500,000) does not preclude recent activity. Because they do not label faults by name, it is not clear whether the Las Villas fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Mesozoic age for an unnamed fault in the vicinity of the Las Villas fault (Figure 2.5.1-369).

The Las Villas fault is not shown on the 1:1,000,000 scale geologic map of Cuba from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3). The 1:2,000,000 scale neotectonic map of Cuba from the same atlas (Reference 944, plate III.2.4-8) shows an unnamed fault in the vicinity of the Las Villas fault. Likewise, the 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11) depicts and labels the Las Villas fault as an approximately 190-kilometer-long (120-mile-long), northwest-trending feature.

Nipe Fault

The Nipe fault is a northeast-striking fault in southern Cuba (Figures 2.5.1-247 and 2.5.1-368 Sheet 3) that separates the mountainous Sierra Maestra province on the east from the Camaguey terrane on the west. At its nearest point, the Nipe fault is approximately 675 kilometers (420 miles) from the Turkey Point Units 6 & 7 site. Other names for this fault include the Cauto, Cauto-Nipe, Guacanayabo, and Nipe-Guacanayabo fault.

Leroy et al. (Reference 499) and Rojas-Agramonte et al. (Reference 445) interpret the Nipe fault as the southern transform limb of the early Cayman spreading center. In their models, the Nipe fault was abandoned by the early Oligocene (approximately 20 Ma) as the plate boundary shifted south to its present location at the Oriente fault.

Cotilla-Rodríguez et al. (Reference 494) characterize the Nipe fault as active based on possible association of seismicity with the fault and gross geomorphic expression. Cotilla-Rodriguez et al. (Reference 494) describe the Nipe fault as “a fault system with transcurrence to the left” whose “outline is labeled by several epicenters” including “some epicentral swarms” near its northeastern end. The Phase 2 earthquake catalog shows sparse seismicity associated with the Nipe fault (Figure 2.5.1-368 Sheet 3). The largest earthquakes in the vicinity of the fault include the August 3,

1926 $M_W$ 5.3 and July 19, 1962 $M_W$ 5.36 earthquakes (Figure 2.5.1-368 Sheet 3). Cotilla-Rodriguez et al. (Reference 494) indicate there are no earthquake focal mechanisms associated with this fault.

Unnamed faults in the vicinity of the Nipe fault are shown on Perez-Othon and Yarmoliuk's (Reference 848) 1:500,000 scale geologic map of Cuba. Because they do not label faults by name, it is not clear whether the Nipe fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Paleogene age for an unnamed fault in the vicinity of the mapped position of the Nipe fault (Figure 2.5.1-369). Unnamed faults in the vicinity of the Nipe fault also are shown on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic map of Cuba. Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba depicts and labels the Nipe fault as the “Cauto-Nipe” fault.

The Nipe fault is not shown on the 1:1,000,000 scale geologic map of Cuba from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3). The 1:2,000,000 scale neotectonic map of Cuba from the same atlas (Reference 944, plate III.2.4-8), however, shows two subparallel, unnamed faults in the vicinity of the Nipe fault. The 1:2,000,000 scale lineament map from this atlas (Reference 944, plate III.3.1-11) labels two faults as “Cauto I” and “Cauto II” in the vicinity of the Nipe fault. On this map, Cauto I strikes northeast and extends from Cuba's southern to its northern coast. Cauto II is more northerly striking and is truncated by Cauto I.

Nortecubana Fault

The Nortecubana fault system is the main structure within the Cuban fold-and-thrust belt offshore of, and nearshore to, northern Cuba (Figures 2.5.1-247, 2.5.1-368 Sheet 1, 2.5.1-368 Sheet 2, and 2.5.1-368 Sheet 3). The Nortecubana fault system dips south with a dip angle that varies along strike. At its nearest point, the Nortecubana fault system is approximately 240 kilometers (150 miles) from the Turkey Point Units 6 & 7 site.

The role of the Nortecubana thrust in the evolution of the Caribbean-North America plate boundary has been interpreted in different ways. The Nortecubana fault system may represent the ancestral subduction zone that was abandoned as the plate boundary shifted southward towards its current location south of Cuba. Alternatively, the Nortecubana thrust fault has been interpreted to represent the frontal decollement of an accretionary wedge associated with the collision of the Greater Antilles Arc and the North America plate south of Cuba (References 439 and 786). Regardless of its ancestral origins, the Nortecubana fault system underlies the preponderance of folding and deformation within and just north of Cuba, which is collectively referred to as the Cuban fold-and-thrust belt. Wells drilled directly offshore of northeastern Cuba have encountered faults and repeated stratigraphy indicating Eocene thrusting (Reference 439), and seismic reflection data have imaged northward thrusting of basin deposits (Reference 307). Seismic lines typically indicate that the offshore north-vergent thrusts are draped by unfaulted late Tertiary to Quaternary sediments (Figures 2.5.1-279, 2.5.1-280, 2.5.1-282, 2.5.1-287, and 2.5.1-288).

Cotilla-Rodríguez et al. (Reference 494) characterize the Nortecubana fault as active based on its possible association with seismicity. They note that the preponderance of this seismic activity is associated with eastern portions of the fault nearest the modern plate boundary. In the Phase 2 earthquake catalog developed for the Turkey Point Units 6 & 7 site, seismicity along the west and central portions of the Nortecubana fault is sparse (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2), relative to the easternmost portion of the fault (Figure 2.5.1-368 Sheet 3). The Phase 2 earthquake catalog includes a $M_W$ 6.29 earthquake that occurred on February 28, 1914 off the north coast of southeastern Cuba (Figure 2.5.1-368 Sheet 3). Cotilla-Rodríguez et al. (Reference 494) suggest this earthquake occurred on the Nortecubana fault. Due to the absence of a permanent seismic monitoring network in Cuba, however, this epicenter is poorly located. The given location, at approximately 6 kilometers (4 miles) north-northeast of the south-dipping Nortecubana fault (and approximately 640 kilometers [400 miles] from the Turkey Point Units 6 & 7 site), suggests that this earthquake could have occurred on another fault. Due to uncertainties in the locations of the 1914

earthquake as well as the fault, this does not preclude the 1914 earthquake from having occurred on the Nortecubana fault. No focal mechanism or depth determination for this earthquake is available with which to help identify the causative fault. It is unlikely that an earthquake of this magnitude would have ruptured to surface of the ocean floor but, even if it had, bathymetric data are insufficient to assess the presence of a submarine fault scarp and no detailed submarine paleoseismic studies are available for the region. Thus, it is not possible to definitively state whether the 1914 earthquake occurred on the Nortecubana or another fault.

The submarine Nortecubana fault typically does not appear on regional surface geologic maps. For example, the Nortecubana fault is not shown on Perez-Othon and Yarmoliuk's (Reference 848) 1:500,000 scale geologic map, Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic maps, and the 1:2,000,000 scale geologic map from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3). This fault, however, is shown on regional tectonic compilations and other maps. For example, Pushcharovskiy et al.'s (Reference 847) 1:500,000 scale tectonic map of Cuba shows the Nortecubana fault as an unnamed, discontinuous, dashed line north of Cuba. The 1:2,000,000 scale neotectonic and lineament maps from the Nuevo Atlas Nacional de Cuba (Reference 944, plates III.2.4-8 and III.3.1-11) show but do not label the Nortecubana fault as solid and dashed lines, respectively. Because they do not label faults by name, it is not clear whether the Nortecubana fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Mesozoic age for an unnamed fault in the vicinity of the Nortecubana fault (Figure 2.5.1-369).

Oriente Fault Zone

The most seismically active region of Cuba today is the Oriente fault zone, located offshore south of eastern Cuba (Figures 2.5.1-229, 2.5.1-247, 2.5.1-251, and 2.5.1-368 Sheet 3). This left-lateral fault system is part of the active North America-Caribbean Plate boundary and connects the Cayman Trough spreading center to the Septentrional fault (Figure 2.5.1-202). Geodetic data indicate that between 8 and 13 millimeters/year of slip are accommodated on this structure; hence it is classified as a capable tectonic source. For further discussion, see Subsections 2.5.1.1.2.3.1.2, 2.5.2.4.4.3.2.2, and 2.5.2.4.4.3.2.3.

Sierra de Jatibonico fault

The Sierra de Jatibonico fault is a 1-2 km-wide zone that parallels the trend of Cuba along its 450-km length. Both Khudoley (Reference 910) and Hatten et al. (Reference 911) describe the fault as being vertical at the surface but gradually flattening at depth, reaching a minimum dip of 55°S. Hatten et al. (Reference 911) state that there is a component of right-lateral displacement along the fault, whereas Pardo (Reference 439) only cites the throw of 1500 meters.

There are no studies that document fault activity or seismicity along the Sierra de Jatibonico fault zone. Mapping by Hatten et al. (Reference 911) shows that the fault juxtaposes the Zueleta and Remedios units, each of which is capped by middle Eocene sediments containing high-angle faults with a component of right-lateral slip. Pardo (Reference 439) also dates the youngest units in the Jatibonico belt (approximately the same as the Remedios unit from Hatten et al. [Reference 911]) as middle Eocene. Assuming that faulting within each unit was contemporaneous with movement on the Sierra de Jatibonico fault, that provides a minimum age of last activity.

Pinar Fault

The Pinar fault is a northeast-striking, steeply southeast-dipping fault in western Cuba (Figures 2.5.1-247, 2.5.1-251, 2.5.1-289, and 2.5.1-368 Sheet 1). As mapped by Tait (Reference 448) and shown on Figure 2.5.1-368 Sheet 1, the Pinar fault is located, at its nearest point, approximately 330 kilometers (205 miles) from the Turkey Point Units 6 & 7 site. As mapped by Garcia et al. (Reference 489), the Pinar fault is approximately 320 kilometers (200 miles) southwest of the site at its nearest point. As mapped by Cotilla-Rodríguez et al. (Reference 494), the Pinar fault

is approximately 360 kilometers (225 miles) southwest of the site at its nearest point. Rosencrantz (Reference 529) maps a series of offshore faults along the eastern Yucatan Platform and tentatively indicates they could be the offshore southwestern extension of the Pinar fault.

The Sierra del Rosario in western Cuba displays a prominent and fairly linear southeast-facing mountain front, suggesting the possibility of recent or ongoing uplift associated with the Pinar fault. There are, however, conflicting opinions in the literature regarding whether the Pinar fault is active. Garcia et al. (Reference 489) note the Pinar fault is grossly expressed as a prominent escarpment and suggest the Pinar fault "was reactivated in the Neogene-Quaternary" and may have produced the January 23, 1880 $M_w$ 6.13 earthquake (Figure 2.5.1-368 Sheet 1). Cotilla-Rodríguez et al. (Reference 494) describe the Pinar fault as having "very nice relief expression" but conclude it is "inactive." Cotilla-Rodríguez et al. (Reference 494) provide no evidence in support of their assessment but suggest that the 1880 earthquake instead occurred on the subsurface Guane fault, which is subparallel to the Pinar fault and is located in the Los Palacios basin to the southeast (Figures 2.5.1-368 Sheet 1). Cotilla-Rodriguez and Cordoba-Barba (Reference 943) cite historical accounts of the severity and distribution of earthquake-related damage as evidence that the January 23, 1880 earthquake occurred on the Guane fault instead of the Pinar fault. Cotilla-Rodriguez and Cordoba-Barba (Reference 943) conclude that the Pinar fault "is not the seismogenetic element of the January 23, 1880 earthquake" and that it is "subordinate to" the Guane fault. Gordon et al. (Reference 697) describe multiple phases of deformation in western Cuba in general and on the Pinar fault in particular. Gordon et al. (Reference 697) are unable to constrain the upper bound of the age of most-recent deformation on the Pinar fault "because lower Miocene rocks were the youngest rocks from which observations were made."

The Phase 2 earthquake catalog indicates that a $M_w$ 6.13 earthquake occurred on January 23, 1880 in western Cuba in the vicinity of the Pinar and Guane faults (Figure 2.5.1-368 Sheet 1). The epicenter of this poorly located, pre-instrumental earthquake is approximately 11 kilometers (7 miles) south of the trace of the steeply southeast-dipping Pinar fault and approximately 8 kilometers (5 miles) north of the Guane fault. As Garcia et al. (Reference 489) suggest, however, locational uncertainties for historical earthquakes in Cuba could be on the order of 15 to 20 kilometers (9 to 12 miles) or more. Based on available information, it is not possible to definitively state whether the 1880 earthquake occurred on the Guane fault, the Pinar fault, or another fault in the region. No focal mechanism or depth determination for the 1880 earthquake is available with which to help identify the causative fault. Moreover, no paleoseismic trench studies or detailed tectonic geomorphic assessments are available for the Pinar fault, Guane fault, or other faults in the region. The Phase 2 earthquake catalog indicates generally sparse to moderate seismicity in the vicinity of the Pinar fault (Figure 2.5.1-368 Sheet 1). There does not appear to be an alignment of epicenters along the Pinar fault, but rather scattered small-magnitude earthquakes appear distributed throughout western Cuba both north and south of the fault in the Sierra del Rosario mountains and the Palacios Basin. The Phase 2 earthquake catalog indicates that additional minor- to moderate-magnitude ($M_w$ 4 to 5.1) earthquakes occurred in western Cuba near the Pinar and Guane faults in 1896, 1937, 1944, and 1957 (Figure 2.5.1-368 Sheet 1).

The Pinar fault is depicted on many regional scale maps of Cuba, including numerous maps in the Nuevo Atlas Nacional de Cuba (Reference 944) and Pushcharovskiy's (Reference 847) 1:500,000 scale tectonic map of Cuba. Available geologic mapping at scales between 1:250,000 and 1:1,000,000 is consistent with an active Pinar fault. These data do not, however, require that the Pinar fault is active. Generally, there is a lack of young deposits mapped along the Pinar fault with which to assess the age of its most-recent slip. Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic mapping shows an unnamed fault in the vicinity of the Pinar fault that, along most of its length, juxtaposes Jurassic-age limestones of the Arroyo Cangre and San Cayetano formations on the northwest against Paleogene-age deposits on the southeast. This map shows the southernmost 5 kilometers (3 miles) of the fault as a dashed line that juxtaposes Jurassic limestone on the northwest against upper Pliocene to lower Pleistocene undifferentiated alluvial and marine

deposits, which may constitute evidence for activity. Along strike immediately to the south near Playa de Galafre on Cuba's southern coast, however, the fault is covered by the same upper Pliocene to lower Pleistocene unit with no apparent deformation (Reference 846). Along the central portion of the fault near Pinar del Rio, Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic mapping shows an approximately 6-kilometer-long (4-mile-long) section where weakly cemented upper Pliocene-lower Pleistocene undifferentiated alluvial and marine deposits on the southeast are fault-juxtaposed against middle Jurassic Arroyo Cangre Formation on the northwest. This map relationship may indicate that the Plio-Pleistocene deposits are faulted. Alternatively, the Plio-Pleistocene deposits may have been deposited against preexisting topography along the fault, and therefore possibly post-date the age of most recent faulting. Based on the crude scale of mapping, it is unclear which of these alternative interpretations is correct.

Perez-Othon and Yarmoliuk (Reference 848) present geologic mapping of Cuba at a scale of 1:500,000. Their map does not include fault names but shows a fault in the vicinity of the Pinar fault that generally juxtaposes Jurassic-age rocks on the northwest against Eocene to Miocene rocks on the southeast. Near Pinar del Rio, they map a small patch of Pliocene-to-Pleistocene-age conglomerates that apparently are correlative with Pushcharovskiy et al.'s (Reference 846) upper Pliocene to lower Pleistocene undifferentiated alluvial and marine deposits in the same area and described above. According to Perez-Othon and Yarmoliuk's (Reference 848) mapping, and unlike Pushcharovskiy et al.'s (Reference 846) mapping, these Plio-Pleistocene deposits extend very close to, but are not in contact with, the fault. Instead, Perez-Othon and Yarmoliuk (Reference 848) show Jurassic-age limestone in fault contact with Eocene-age rocks in this area. Farther to the northeast near Los Palacios, Perez-Othon and Yarmoliuk (Reference 848) show an approximately 2- to 4-kilometer-long (1- to 2-mile-long) stretch along the central section of the fault where Quaternary alluvial deposits are juxtaposed against Jurassic carbonate rocks. The resolution of Perez-Othon and Yarmoliuk's (1985) (Reference 848) mapping is insufficient to determine whether these Quaternary alluvial deposits are faulted or if they were deposited against preexisting topography along the fault, and therefore possibly post-date the age of most-recent faulting. As an inset to their geologic map, Perez-Othon and Yarmoliuk (Reference 848) provide an additional map that shows their estimates of fault ages in Cuba. On their inset map of fault ages in Cuba, Perez-Othon and Yarmoliuk (Reference 848) assign a Neogene-Quaternary age to a northeast-striking fault that is presumed to be the Pinar fault (the inset map does not include fault names). Despite this Neogene-Quaternary age on the inset map, their 1:500,000 scale geologic map shows unnamed northwest-striking faults, to which they assign a Paleogene age on their inset map, as offsetting the younger Pinar fault.

The Nuevo Atlas Nacional de Cuba includes a 1:1,000,000 scale geologic map of Cuba (Reference 944, plate III.1.2-3). No fault names appear on this map, but a fault in the vicinity of the Pinar fault is shown as juxtaposing Jurassic carbonate rocks on the northwest against Miocene and older rocks on the southeast. Due to the crude scale at which this map is presented, however, it is not possible to constrain with certainty the age of faulting. This atlas also includes a 1:2,000,000 scale neotectonic map of Cuba (Reference 944, plate III.2.4-8) that defines “zones of maximum neotectonic gradient” and classifies them as “moderate,” “intense,” or “very intense.” Only the modern plate boundary offshore southern Cuba is classified as “very intense” in this scheme. No fault names appear on this map, but a fault in the vicinity of the Pinar fault is shown in an “intense” zone.

Surcubana Fault

At its nearest distance, the Surcubana fault as mapped by Cotilla-Rodriguez et al. (Reference 494) is located approximately 370 kilometers (230 miles) from the site (Figures 2.5.1-368 Sheet 1, 2.5.1-368 Sheet 2, and 2.5.1-368 Sheet 3). Cotilla-Rodriguez et al. (Reference 494) do not include the Surcubana fault in their list of twelve “seismoactive” faults in Cuba and this fault generally is not described by other studies of faulting in Cuba (References 439, 489, and 786).

In the Phase 2 earthquake catalog, seismicity is sparse along and near the Surcubana fault, with only a dozen or so earthquakes located within approximately 30 kilometers (20 miles) of the more than

800-kilometer-long (500-mile-long) trace (Figures 2.5.1-368 Sheet 1, 2.5.1-368 Sheet 2, and 2.5.1-368 Sheet 3). Of these earthquakes, all are low to moderate magnitude and most are located at the southeastern end of the fault near the active plate boundary and may instead be associated with the Oriente fault. The closest earthquakes to the central and western sections of the Surcubana fault from the Phase 2 earthquake catalog are located at approximately 81° west longitude (Figures 2.5.1-368 Sheet 1 and 2.5.1-368 Sheet 2). The first of these is located approximately 8 kilometers (5 miles) north of the trace and occurred on March 27, 1964 with $M_w$ 3.7. The second is located approximately 5 kilometers (3 miles) south of the trace and occurred on October 22, 2005 with $M_w$ 3.8. Because they do not label faults by name, it is not clear whether the Surcubana fault is depicted on Perez-Othon and Yarmoliuk's (Reference 848) inset map of fault ages in Cuba, but they indicate a Mesozoic age for an unnamed fault in the vicinity of the Surcubana fault (Figure 2.5.1-369).

Like the Nortecubana fault, the submarine Surcubana fault typically does not appear on regional surface geologic maps. For example, the Surcubana fault is not shown on Pushcharovskiy et al.'s (Reference 846) 1:250,000 scale geologic maps, and the 1:2,000,000 scale geologic map from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.1.2-3). This fault is shown on regional tectonic compilations and other maps. For example, Pushcharovskiy et al.'s (Reference 847) 1:500,000 scale tectonic map of Cuba shows the Surcubana fault as an unnamed, discontinuous, dashed line south of Cuba. The 1:2,000,000 scale neotectonic map from the Nuevo Atlas Nacional de Cuba (Reference 944, plate III.2.4-8) shows, but does not label, the Surcubana fault as a solid line. The lineament map from the same atlas (Reference 944, plate III.3.1-11) shows but does not label the Surcubana fault as discontinuous and dashed lines.

Other Cuban Structures

Numerous other tectonic structures exist on the island of Cuba. Some of these are limited in extent, unstudied, or unnamed. These include the Punta Alegre fault, folds along the northern edge of Cuba, and many short, unnamed northeast- and northwest-striking faults. The Punta Alegre fault was discovered by logging repeated strata in oil wells just offshore north-central Cuba (Figures 2.5.1-247 and 2.5.1-290). This fault is not imaged with seismic data, but postulated from well data. It is depicted with a vertical dip, but its orientation and extent are unknown (Reference 501).

Eocene and older strata along the northern edge of Cuba are deformed in a series of anticlines and synclines typically associated with underlying thrust faults (Figures 2.5.1-252 and 2.5.1-282). Because these folds are covered by undeformed Miocene and younger strata, they are pre-Miocene in age, and probably formed during the Eocene collision of the Greater Antilles Arc with the Bahama Platform.

Many short (<10 kilometers [<6.2 miles] in length) northeast- and northwest-striking faults, with undetermined sense of slip, do cut strata as young as middle Miocene throughout the island of Cuba. Where younger units (such as Plio-Pleistocene) overlie these same structures, they are consistently unfaulted. This suggests that these short faults are pre-Quaternary in age. Many of these faults do not intersect units younger than Miocene, so the faulting on these structures can only be described as Miocene or younger. These structures may be correlated with post-early Miocene normal faults and cross-cutting strike-slip faults described in outcrops in western Cuba (Reference 697).

In summary, many faults have been mapped on the island of Cuba. Aside from the Oriente fault, most of these faults were active during the Cretaceous to Eocene, associated with subduction of the Bahama Platform beneath the Greater Antilles Arc of Cuba and the subsequent southward migration of the plate boundary to its present position south of Cuba (Figure 2.5.1-250). However, only a few detailed studies of the most recent timing of faulting are available, and conflicting age assessments exist for many of the regional structures (Table 2.5.1-204). The available data indicate that the Oriente fault system, located offshore directly south of Cuba, should be characterized as a capable tectonic source. Aside from the Oriente fault, no clear evidence for Pleistocene or younger faulting is available for any of the other regional tectonic structures on Cuba, and none of these faults are

adequately characterized with late Quaternary slip rate or recurrence of large earthquakes. The scales of available geologic mapping (1:250,000 and 1:500,000; References 846, 847, and 848) do not provide sufficient detail to adequately assess whether or not individual faults in Cuba can be classified as capable tectonic structures.

Additionally, elevated marine terraces were identified along the northern coast of Cuba as early as the late 19th century (Reference 912). Recent studies of the marine terraces along the north coast of Cuba, especially for the stretch between Matanzas and Havana, are summarized below. Subsection 2.5.1.1.1.2.3 provides a description of the Quaternary deposits and surfaces in the Matanzas region, including the Pleistocene-age Terraza de Seboruco surface west of Matanzas Bay. Ducloz (Reference 915) suggests that the elevated marine terraces along Cuba's north coast likely formed as the result of both fluctuations in sea level and epeirogenic uplift (Table 2.5.1-208). Ducloz (Reference 915) suggests that reactivation of a regional scale anticline may be partly responsible for formation of the terrace surfaces near Matanzas.

Similarly, Shanzer et al. (Reference 923) identify three Pleistocene-age marine terraces in the Matanzas-Havana region. Shanzer et al. (Reference 923) correlate segments of the Pleistocene-age Terraza de Seboruco between Matanzas and Havana and suggest that this terrace is approximately 1.5 to 3 meters (4.9 to 9.8 feet) lower at Havana than at Matanzas. Shanzer et al. (Reference 923) do not consider erosion of the terrace surface to explain the difference in elevation between Havana and Matanzas. Shanzer et al. (Reference 923) postulate that this difference in elevation may be the result of differential tectonic uplift, but they do not suggest what structure or structures may be responsible for this postulated tectonic uplift.

Toscano et al. (Reference 925) also observe that the Terraza de Seboruco in the Matanzas area is just a few meters above mean sea level, similar to the elevation of other Substage 5e reef deposits throughout "stable" portions of the Caribbean, and therefore can be explained solely by changes in sea level. Toscano et al. (Reference 925) conclude, "no obvious tectonic uplift is indicated for this time frame along the northern margin of Cuba."

Pedoja et al. (Reference 920) investigate late Quaternary coastlines worldwide and observe minor uplift relative to sea level of approximately 0.2 millimeter per year, even along passive margins, outpacing eustatic sea level decreases by a factor of four. Pedoja et al. (Reference 920) suggest that the decreasing number of subduction zones since the Late Cretaceous, coupled with relatively constant ridge length, has resulted in an increase in the average magnitude of compressive stress in the lithosphere. They argue that this average increase in compressive stress has produced low rates of uplift even along passive margins, as observed in their widespread measurements of uplifted continental margins. The measurements specific to Cuba suggest that the Substage 5e terrace in the Matanzas area (i.e., the Terraza de Seboruco) has been uplifted at an average rate that ranges from approximately 0.00 to 0.04 millimeter per year over the last approximately 122 ka (Reference 920).

### Seismicity of Cuba

Maps of instrumental and pre-instrumental epicenters for Cuba show that seismicity can be separated into two zones: (a) the very active plate boundary region, including the east Oriente fault zone along Cuba's southern coast, and (b) the remainder of the island away from the active plate boundary region, which exhibits low to moderate levels of seismic activity (Figures 2.5.1-267, 2.5.2-220, and 2.5.2-221). Regarding (b) above, along the north coast of Cuba between Havana and Matanzas, the Phase 2 earthquake catalog indicates sparse minor- to light-magnitude seismicity. It is possible that these earthquakes occurred on faults partially responsible for uplift of the marine terraces along Cuba's north coast in the site region. However, the association of the uplift of these terraces and earthquakes with individual faults in northern Cuba is uncertain. Based on the Phase 2 earthquake catalog, earthquakes do not appear to be aligned along faults in the Matanzas-Havana region. In addition, there are no known focal mechanisms available for these earthquakes that would

help to constrain the causative fault or faults nor is there sufficient data to correlate uplift of marine terraces with these individual faults in northern Cuba.

It is possible that the elevations above modern sea level of marine terraces along Cuba's north coast in the site region are partially the result of tectonic uplift (References 915 and 923). The Terraza de Seboruco is the only terrace in northern Cuba for which radiometric age control is available. There is not sufficient data on this or other marine terraces in northern Cuba to assess the implications for active faulting. As discussed in Subsection 2.5.1.1.1.2.3, Toscano et al.'s (Reference 925) U-Th analysis of corals collected from the Terraza de Seboruco indicates that tectonic uplift is not required to explain the present elevation of this Substage 5e terrace. Instead, they conclude that the elevation of this terrace surface is consistent with other Substage 5e terraces in other tectonically stable regions of the Caribbean and that global fluctuations in sea level, not tectonic uplift, are responsible for the Terraza de Seboruco's present elevation above modern sea level. Likewise, Pedoja et al.'s (Reference 920) global study suggests that the elevation of the Terraza de Seboruco is consistent with the elevations of other Substage 5e terraces in tectonically stable regions worldwide.

Based on studies by Toscano et al. (Reference 925) and Pedoja et al. (Reference 920), active faulting is not required to explain the elevation of the Terraza de Seboruco along Cuba's north coast in the site region. However, observations of the Terraza de Seboruco cannot necessarily be used to preclude possible strike-slip faulting in the site region. As shown by the Phase 2 earthquake catalog, only sparse minor-to light-magnitude seismicity is observed along Cuba's northern coast between Havana and Matanzas. It is possible that at least some of these earthquakes occurred on the faults mapped in the region. However, in the absence of well-located hypocenters and focal mechanisms, these earthquakes cannot be definitively attributed to a particular fault or faults.

The east Oriente fault zone is an active plate boundary, with seismic activity concentrated on the Cabo Cruz Basin and the Santiago deformed belt. Focal mechanisms from the Cabo Cruz area show consistent east-northeast to west-southwest oriented normal faulting, indicative of an active pull-apart basin. In the Cabo Cruz Basin, all hypocenters are less then 30 kilometers (19 miles) deep. The Santiago deformed belt mechanisms show a combination of northwest-directed underthrusting and east-west left-lateral strike-slip, consistent with a bi-modal transpressive regime (Reference 504). In the Santiago deformed belt, thrust mechanisms occur between depths of 30 and 60 kilometers (19 and 37 miles), while the strike-slip mechanisms are shallower.

According to the Phase 2 earthquake catalog (Subsection 2.5.2.1.3), eight approximately $M_w$ 6.8 to 7.5 events (in August 1578, February 1678, June 1766, August 1852, February 1917, February 1932, August 1947, and May 1992) probably occurred offshore southern Cuba, likely in the Cabo Cruz Basin and/or the Santiago deformed belt (Figure 2.5.2-214).

Figures 2.5.2-201 and 2.5.2-210 show that although Cuba is now part of the North America Plate, the central and western portions of the island away from the active plate boundary region exhibit a moderate level of seismicity that is higher than that observed in Florida. Figures 2.5.2-215 and 2.5.2-216 show that microseismicity is distributed roughly evenly throughout this zone, but with a tendency for epicenters to be located to the southeast part of the island. Activity between the Nipe fault and the east Oriente fault zone appears denser than on the rest of the island (Figure 2.5.2-215). This may partially be a detection effect, however, since a denser concentration of seismograph stations exists in this region (Reference 505).

Reported earthquakes in central and western Cuba away from the active plate boundary region typically are of low to moderate magnitude. Two of the largest earthquakes in this region occurred in January 1880 (MMI VIII and magnitude 6.0 to 6.6) near the Pinar fault in western Cuba, and February 1914 ($M_w$ 6.2) offshore northeastern Cuba near the Nortecubana fault (Reference 494) (Figure 2.5.2-214). However, there is no direct evidence that these earthquakes occurred on the

Pinar and the Nortecubana faults. The Phase 2 earthquake catalog (see Subsection 2.5.2.1.3) indicates $M_W$ 6.13 and 6.29 for the 1880 and 1914 earthquakes, respectively.

#### 2.5.1.1.2 Geology beyond the Site Region

This subsection addresses the geologic and seismic data/information on structures outside the 200-mile (320-kilometer) radius of the Units 6 & 7 site region that may be relevant to evaluating geologic hazards to the Units 6 & 7 site. The geologic hazards specifically include seismic hazards evaluated in the PSHA of Subsection 2.5.2 and tsunami hazards discussed in Subsection 2.5.1.1.5 and evaluated in Subsection 2.4.6. This subsection includes a description of the physiography, stratigraphy, structure, and seismicity of portions of the North America Plate and portions of the Caribbean Plate near its boundary with the North America Plate. Due to their remote distance from the Units 6 & 7 site, features of the Caribbean-South America Plate boundary are not discussed in this subsection.

##### 2.5.1.1.2.1 Geology of the Southeastern North America Plate Geologic Provinces

The following subsections describe physiography, stratigraphy, structures, and seismicity of the southeastern North America and northern Caribbean plates.

###### 2.5.1.1.2.1.1 Geology of the Gulf of Mexico

**Physiography of the Gulf of Mexico**

The Gulf of Mexico is a semi-enclosed, small ocean basin located at the southeastern corner of the North America Plate that covers an area of more than 1.5 million kilometers$^2$ with a maximum water depth of approximately 3700 meters (12,100 feet). The Gulf of Mexico is a sedimentary basin that consists of thick accumulations of detrital sediments and massive carbonates that have been affected by salt tectonics. Mesozoic to Cenozoic sediments accumulated within the expanding and subsiding basin. Following thermal subsidence, the basin continued to subside due to lithostatic loading, eventually attaining a stratigraphic sequence comprising nearly 15,000 meters (49,200 feet) of evaporites overlain by prograding clastic deltaic and turbidite deposits interbedded with organic rich shales and pelagic carbonates. In the northern, southern, and eastern portions of the Gulf of Mexico, the broad continental shelf is up to 170 kilometers (106 miles) wide. In the western portion, the continental shelf east of Mexico is less than 13 kilometers (8 miles) wide in some places. The physiography of the Gulf of Mexico Basin has been controlled by processes such as subsidence, carbonate platform development, eustatic changes in sea level, salt diapirism, oceanic currents, gravity slumping, and density flows (turbidites) (References 506 and 507).

Antoine (Reference 508) divides the Gulf of Mexico Basin into seven provinces based on morphology. Bryant et al. (Reference 506) divide the Gulf of Mexico into more detailed physiographic provinces based on bathymetry and topographical features (Figure 2.5.1-292). Counterclockwise along the Gulf Coast from Florida to the Yucatan Peninsula, these provinces include the following: Florida Straits, including the Pourtales Escarpment; Florida Plain; Florida Middle Ground, West Florida Shelf, and West Florida Terrace (together known as the Florida Platform in Subsection 2.5.1.1.1); DeSoto Slope and Canyon; Mississippi Alabama Shelf; Mississippi Canyon; Mississippi Fan; Texas-Louisiana Shelf; Texas-Louisiana Slope; Rio Grande Slope; East Mexico Shelf; East Mexico Slope; Western Gulf Rise; Veracruz Tongue; Campeche Knolls; Bay of Campeche; Campeche Canyon; Sigsbee Abyssal Plain, the Yucatan Shelf and Campeche Escarpment; Campeche Terrace; and Yucatan Channel.

Water enters the Gulf of Mexico through the Yucatan Channel, circulates as the Loop Current, and exits through the Straits of Florida, eventually forming the Gulf Stream. Portions of the Loop Current often break away forming eddies or 'gyres' that affect regional current patterns. Smaller wind driven and tidal currents are created in near shore environments.

Drainage into the Gulf of Mexico is extensive and includes 20 major river systems (>150 rivers) covering over 3.8 million kilometers$^2$ of the continental United States (Reference 510). Annual freshwater inflow to the Gulf of Mexico is approximately 10.6 x $10^{11}$ meters$^3$ per year (280 trillion gallons). Eighty-five percent of this flow comes from the United States, with 64 percent originating from the Mississippi River alone. Additional freshwater inputs originate in Mexico, the Yucatan Peninsula, and Cuba.

### Stratigraphy of the Gulf of Mexico

The basement beneath the Gulf of Mexico is characterized by a regional unconformity that separates pre- and syn-rift rocks from overlying Lower Jurassic to Recent lithologies that reflect the tectonic history of the southeastern North America Plate and its boundary with the Caribbean Plate. Because the Gulf of Mexico has been subsiding continually since the Pangean rifting event, it contains the most complete sequence of strata that represent nearly 150 m.y. of uninterrupted geologic history.

Based on seismic reflection profiles, the Gulf of Mexico includes a deep zone that contains normal-thickness oceanic crust or “thin” oceanic crust. This crust was created in the Late Jurassic through Early Cretaceous along two seafloor spreading segments (References 511 and 512), a larger southwest-northeast oriented spreading center beneath the abyssal plain north of the Campeche Escarpment, and a shorter northwest-southeast oriented spreading center that lies just east of and parallels the Florida Escarpment (Figure 2.5.1-214). The normal-thickness oceanic crust produced by the southwest-northeast spreading center is about 400 kilometers (250 miles) wide. The normal-thickness oceanic crust produced by the northwest-southeast spreading center is narrower and possibly younger (Reference 410). This normal oceanic crust in the Gulf of Mexico Basin is generally 5 to 6 kilometers thick and is characterized by refraction velocities of 6.8 to 7.2 kilometers/second, probably corresponding to oceanic layer 3 found in most normal ocean basins (Reference 410). Sawyer et al. (Reference 410) could not distinguish oceanic layer 2 in the Gulf of Mexico because of its deep burial and lack of density contrast between it and the compacted clastic and carbonate sediments that overlie it. However, Sawyer et al. (Reference 410) note that a layer identified on their seismic reflection profile (Figure 2.5.1-293) most likely includes this layer and the carbonates. The top of this interval is the mid-Cretaceous sequence boundary that occurs throughout the basin and is interpreted as the top of the oceanic crust.

Surrounding the area of normal oceanic crust is an area of transitional crust (Figures 2.5.1-238, 2.5.1-239, 2.5.1-240, 2.5.1-241, and 2.5.1-242). This area flanks the basin on all sides and occupies narrow belts to the east and west with a wider region to the south and a broad zone to the north (Figure 2.5.1-238). Based on limited refraction data, a prominent, high-amplitude, basinward-dipping reflector/unconformity is interpreted to be the top of the crust. Over much of the area, the surface is relatively smooth (probably erosional), although in places it is offset by small faults. The crustal thickness ranges from 8 to 15 kilometers (5 to 9 miles) with velocities of 6.4 to 6.8 kilometers/second. The surface also truncates a thick older sedimentary sequence (Late Triassic to Early Jurassic syn-rift deposits) (Reference 410). The transitional crust is unconformably overlain by, and shows onlap relationships with, the Middle Jurassic salt and Upper Jurassic and Cretaceous sediments. In the southeastern Gulf of Mexico, the top of the thin transitional crust rises to shallow depths over a northeast-southwest-trending basement arch. The arch area is characterized by Mesozoic-faulted blocky basement (Figure 2.5.1-294). In the eastern Gulf of Mexico beneath the West Florida Basin, the top of the thin transitional crust consists of thick salt and sediments and is seen along the western part of the West Florida Basin (Reference 410).

The thick transitional crust in the Gulf of Mexico generally lies landward of the thin transitional crust. Based on seismic reflection data, thick transitional crust has a thickness from about 20 kilometers (12 miles) up to normal continental crust thickness of about 35 to 40 kilometers (22 to 25 miles) (Figure 2.5.1-238). The crust is characterized by relatively shallow, well-defined basement highs with intervening lows. The high areas overlie crust with thickness close to normal continental crust, while the lows overlie thinner crust, probably extended continental crust (Reference 429). The typical thick

transitional crust is seen best in the northeastern Gulf of Mexico Basin (Reference 410) (Figures 2.5.1-240 and 2.5.1-242).

The southeastern portion of the Gulf of Mexico, closest to the Units 6 & 7 site, is located north of Cuba between the Campeche and Florida Escarpments (Figure 2.5.1-210). The seafloor is shallower than in the Gulf of Mexico Basin proper and is characterized by erosional channels (the Straits of Florida and the Yucatan Strait) and large knolls (i.e., Pinar del Rio Knoll, Catoche Knoll, and the Jordan Knoll) (Figure 2.5.1-210). Based on a seismic stratigraphic analysis combined with DSDP drilling data, Schlager et al. (Reference 794) determine that the southeastern Gulf is underlain by rifted and attenuated transitional crust covered by a thick sedimentary section of pre-mid-Cretaceous rocks.

The Late Cretaceous-Cenozoic cover is relatively thin over most of the area, but it thickens to the south towards Cuba. The pre-mid Cretaceous section probably reflects an overall transition upward from nonmarine to shallow marine and then deep marine deposits as the basin subsided. The sedimentary sequences overlying the basement as seen from DSDP Leg 77 cores are grouped into five units: a Late Triassic-postulated Early Jurassic rift basin (TJ); a widespread postulated Jurassic nonmarine to shallow marine unit (J1); a more restricted postulated Late Jurassic shallow to deep marine unit (J2); a widespread Early Cretaceous unit (EK); and a Late Cretaceous-Cenozoic unit (KC) (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

According to Schlager et al. (Reference 794), there are no drilling data for the pre-Cretaceous history of the southeastern Gulf. However, a scenario for the pre-Cretaceous history can be discussed on the basis of interpretation of seismic data and regional comparisons. The basement is approximately early Paleozoic (500 Ma) and consists of metamorphic rocks (such as phyllite and gneiss-amphibolite) intruded by early Mesozoic (160 to 190 Ma) basic dikes and sills. In some places the basement contains some low-amplitude reflections, seen as broad uplifts and basins to the south and north. High relief tilted Mesozoic fault-blocks are in the central part of the southeastern part of the Gulf of Mexico (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

Unit TJ is a Late Triassic-Early Jurassic rift sequence consisting of southeast-dipping parallel reflections filling a northeast-southwest-trending graben system. Unit TJ onlaps the basement and is truncated by prominent unconformity. The unit is probably composed of nonmarine sediments and volcanics (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

Unit J1 is a widespread unit in the south with a relatively uniform thickness of several kilometers, and shows high-amplitude and discontinuous seismic character. The unit onlaps broad basement highs and is undeformed in the southeastern part of the Gulf of Mexico except where the J1 unit is downdropped along prominent northwest-southeast graben system and along the broad trough north of Cuba. Northward, the unit fills half-grabens between tilted fault blocks. The seismic character to the north suggests non-marine synrift sediments such as alluvial fans, lacustrine deposits, volcanics, and evaporites. The upper part of the unit in the south may be shallow marine platform with the lower part nonmarine (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

Unit J2 consists of uniform, variable-amplitude, continuous reflections. The unit is widespread over most of the area, deformed in depressions between horsts and absent on high-standing blocks to the west. The seismic reflection data suggest deep marine deposition in the central part of the southeastern portion of the Gulf of Mexico. Possible low relief shelf margins in places are suggestive of transition to shallow marine conditions around the periphery. Unit J2 most likely represents a major marine transgression, concurrent with the establishment of the seaway (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

Unit EK is widespread throughout the north and eastern portions of the Gulf of Mexico nearest to the Straits of Florida. It is thin or absent to the south and west because of nondeposition on high-standing

areas and post mid-Cretaceous erosion. Unit EK has a thickness of up to 2 kilometers (1.2 miles) and thickens to the east along the base of the Florida Escarpment. The unit is a deep-water carbonate whose main source of carbonate supply was the Florida Platform and planktonic carbonate production (Reference 794) (Figures 2.5.1-242, 2.5.1-241, and 2.5.1-295).

Toward the south of the Gulf of Mexico, the lower part of the late Cretaceous-Cenozoic KC unit forms thick wedges of clastic sediment originating from Cuba. The upper part of the KC unit forms a thin blanket with internal unconformities. To the north, the unit thins then thickens into the Gulf of Mexico Basin (Reference 794) (Figures 2.5.1-241, 2.5.1-242, and 2.5.1-295).

#### Structures of the Gulf of Mexico

The deep basin of the Gulf of Mexico is draped by several kilometers of generally undisturbed Cretaceous to Quaternary sedimentary strata (Figures 2.5.1-241, 2.5.1-240, and 2.5.1-242). Normal faulting and volcanic activity associated with the opening of the Gulf of Mexico Basin was widespread and ended in the Jurassic to Cretaceous (References 368 and 849). In the southeastern Gulf of Mexico, between the Yucatan and Florida Platforms, undisturbed Cretaceous and younger strata cover the Mesozoic normal faults cutting the basement (e.g., Figures 2.5.1-293, 2.5.1-294, and 2.5.1-295). Strike-slip structures exposed in eastern Mexico, and proposed offshore, along the western Gulf of Mexico accommodated the opening of the Gulf of Mexico in the Jurassic (Reference 849). The Gulf of Mexico Quaternary strata are disturbed along the northern Gulf of Mexico coast, from the Florida Panhandle west to Texas, where aseismic gravity-driven growth faults extend the thick fluvial-deltaic sedimentary sections into the Gulf of Mexico Basin (Reference 430). However, because the Florida Platform remained a site of carbonate deposition and lacks a thick clastic section, the eastern Gulf of Mexico adjacent to the Florida Platform is not a site of growth faulting (Reference 513). Some normal faults near the western edge of the Florida Platform accommodated extension during the opening of the Gulf of Mexico in Jurassic and early Cretaceous periods (Figure 2.5.1-264).

#### Seismicity of the Gulf of Mexico

The Phase 1 earthquake catalog (Subsection 2.5.2.1.2) indicates that the Gulf of Mexico is characterized by low seismicity rates (Figure 2.5.1-267). According to the Phase 1 earthquake catalog, the two largest earthquakes in the Gulf of Mexico are the September 10, 2006, Emb 5.90 and February 10, 2006, Emb 5.58 earthquakes. Subsection 2.5.2.4.3.1 provides additional discussion regarding these two earthquakes. The overall seismicity pattern within the Gulf of Mexico shows no correlation with geologic or tectonic features (Subsection 2.5.2.3).

### 2.5.1.1.2.1.2 Geology of the Yucatan Platform

#### Physiography of the Yucatan Platform

The Yucatan Platform comprises the emergent portion of the Yucatan Peninsula; the broad, shallow carbonate platform that extends mostly north and west of the peninsula; a narrow, deeper water carbonate terrace (the Campeche Bank) that rims the shallow carbonate platform; and a steep walled escarpment that transitions from the Yucatan Platform to the Gulf of Mexico and Yucatan Basin abyssal plains. From east to west at its widest point, the continental shelf of the Yucatan Platform extends about 675 kilometers (420 miles), from the western Gulf Coast of Mexico to the southwestern tip of Cuba. From north to south, the Yucatan Platform extends about 1000 kilometers (620 miles) into the Gulf of Mexico from the western end of the Cayman Trough.

The Yucatan Platform has been the site of limestone and evaporite deposition since the Early Cretaceous. Currently, living reefs and biohermal mounds, wave cut terraces, and small-scale karst features dominate the topography of the area. The broad shelf is surrounded on three sides by the steep Campeche Escarpment that plunges as much as 3600 meters (11,800 feet) from the shelf edge to the Gulf of Mexico floor. The escarpment, with a slope of up to 35°, is broken by the

Campeche Terrace and a number of box canyons. The terrace is at a depth of approximately 1000 meters (3300 meters), with a width of 200 kilometers (124 miles) and an average slope of 5°. The Campeche Terrace is analogous to the Blake Plateau, being a drowned portion of an active carbonate platform. Small canyons and large-scale slumping interrupt the linearity of the Campeche Escarpment (Reference 506). The largest canyon, the Catoche Tongue, parallels the north-northeast oriented transform fault margin on the eastern side of the Yucatan Platform (Figure 2.5.1-210).

The Yucatan Platform formed as the result of reef building and upward growth by slow accumulation of carbonate sediments and evaporites. The growth has kept pace or exceeded subsidence since the Early Cretaceous (Reference 506). The Campeche Escarpment, like the Florida Escarpment, represents the eroded margins of the Early Cretaceous carbonate platform. Due to the presence of carbonate talus deposits in localized areas along its base, the Campeche Escarpment has undergone erosion and retreat (Reference 725). The Florida Escarpment has retreated as much as 8 kilometers (5 miles), and the Blake Escarpment has retreated as much as 20 kilometers (18 miles) since the mid-Cretaceous (Reference 725). According to Buffler et al. (Reference 515), these Early Cretaceous margins are interpreted to have been established along a regional tectonic hinge zone separating thick transitional crust from thin transitional crust.

### Stratigraphy of the Yucatan Platform

The Yucatan Peninsula is the exposed part of the Yucatan Platform. It is comprised primarily of Cretaceous carbonate platform rocks overlying a Paleozoic crystalline basement. The basement beneath the Yucatan Platform and the possible relationship between the Yucatan Platform and the Florida and Bahama Platforms are described in Subsection 2.5.1.1.1.1.1.2 (as part of the discussion of carbonate platforms: growth, shut downs, and crashes).

A 180- to 200-kilometer (112- to 125-mile) wide impact structure, the Chicxulub crater (which formed 65.55 ± 0.05 Ma [$^{40}Ar/^{39}Ar$ dating of glassy melt rocks/tektites] at the K/Pg boundary [References 516, 517, and 518]), forms the northwest margin of the peninsula (see discussion of “Oceanic and Atmospheric Reorganization and Extinction Event” in Subsection 2.5.1.1). The deeply buried Chicxulub Crater is located partly onshore and offshore of the northwest part of the Yucatan Peninsula, near the town of Chicxulub. The Chicxulub crater is overlain by up to 1.5 kilometers (4920 feet) of Tertiary sediments (Reference 519). The geomorphology of the peninsula is characterized by a relatively smooth platform with elevations that range between 25 and 35 meters (82 to 115 feet) broken by rounded karstic depressions (locally known as “senates”) and uninterrupted by stream valleys (Reference 520).

Pleistocene reef limestones, lagoonal packstone-wackestones, strandline grainstones, and calcretes are exposed in quarries and low sea cliffs along the Caribbean coast of the Yucatan Peninsula from the northern cape to Tulum. These shallow-marine and subaerial limestones are similar in elevation, sedimentology, stratigraphy, and age to similar limestones found on Isla Cozumel. The Isla Cozumel consists of caliche facies in Upper Pleistocene limestones. Sub-Caliche I facies consist of coralline wackestone and molluscan wackestone. Super-Caliche I facies consist of coral-reef facies and skeletal and oolitic grainstone-packstone and burrowed skeletal grainstone-packstone. Holocene eolianites were deposited along the northeastern shoreline that is adjacent to the narrow ramp, but these are absent south of Isla Cancun, where the margins of the peninsula and offshore platforms are steep (Reference 521).

The correlative Upper Pleistocene limestones reflect the same history of late Quaternary sea-level fluctuation for the eastern Yucatan coast and the offshore carbonate islands. The similar elevations of these age-equivalent rocks also suggest there has been little or no differential structural movement along this portion of the Yucatan continental margin for at least the past 200 k.y. As seen from the similarity of the elevations of Upper Pleistocene limestone of Yucatan and those of oxygen isotopic substage 5e, limestones in stable areas of the Caribbean, there has been no significant subsidence or uplift of the eastern Yucatan Peninsula after mid-Pleistocene (Reference 521).

### Structures of the Yucatan Platform

At its nearest point, the Yucatan Platform province lies about 370 miles (600 kilometers) west-southwest of the Units 6 & 7 site. Bedrock structure of the Yucatan Platform is constrained by surface geologic mapping (compilation in Reference 492) and gravity and magnetic studies (Reference 522), which indicate the platform comprises denser basement rock with a cover of Cretaceous through Oligocene strata. The basement of the Yucatan Platform exhibits an undulating and irregular surface consisting of a variety of pre-Late Paleozoic igneous, metamorphic, and sedimentary terranes often overprinted with a Late Paleozoic metamorphic age. The metamorphic signature represents the assemblage of Pangea (Reference 522).

Alvarado-Omana (Reference 522) speculates that the undulating and irregular basement surface, modeled using gravity and magnetic data, could be due to either: (a) crustal thinning and stretching associated with North America-South America rifting between which the Yucatan block was situated in the Jurassic; or (b) density differences between northern and southern Yucatan Platform basement rock (0.05 g/cc greater in the northern portion). Alvarado-Omana (Reference 522) suggests that density differences could result from the possibility that the Yucatan block comprises a series of “micro-continental” blocks that surround the Yucatan Platform. A possible explanation for the density differences could be that the Jurassic rift basins of the northern Yucatan Platform (Reference 523) contain higher-density material than the cover of Cretaceous and younger strata overlying the basement of more southerly portions.

The northern Yucatan Platform region was uplifted, accommodated mostly by normal faulting along its northwestern margin during the Late Triassic, concurrent with opening of the Gulf of Mexico (References 522 and 524). This process created the Campeche Escarpment that delineates and extends along the northwestern margin of the platform. French and Schenk (Reference 492) mapped several normal faults along a small portion of the Campeche Escarpment, but little is known of their age; they could be gravitational due to the steepness of the escarpment. No seismicity from the Phase 1 or Phase 2 earthquake catalogs (Subsection 2.5.2.1) is associated or coincides with those faults. Uplift continued into the Middle Jurassic, followed by subsidence due to onlapping deposition of the Lower Cretaceous Carbonate Platform over the dense Paleozoic basement.

Pindell et al. (Reference 523) propose the Yucatan Platform underwent two episodes of counter-clockwise rotation. The first episode involved 10 to 15° of rotation from Late Triassic through Late Jurassic, associated with North America-South America continental rifting. The second involved an additional 30 to 35° of rotation from Late Jurassic though earliest Cretaceous associated with later stages of oceanic spreading in the Gulf of Mexico. Rotations are constrained by alignment of the Jurassic rift basins of the northern Yucatan Platform with those of North America (Reference 523). These rotations are not, however, directly associated with deformation of the Yucatan Platform.

From Early Cretaceous through Late Cretaceous (Maastrichtian) time, the platform existed as a relatively passive margin (References 522, 525, and 523). Beginning in the Masstrichtian, the Caribbean Plate passed along the eastern margin of the Yucatan Platform. Regional stress fields transitioned from those related to oblique sinistral convergence from Late Cretaceous (Maastrichtian) through late Paleocene time, to oblique sinistral extension from late Paleocene through Middle Eocene time (Reference 525) (Figure 2.5.1-297). The active margin of the eastern Yucatan Platform represented the North America-Caribbean Plate boundary. Several normal faults are mapped by French and Schenk (Reference 492) parallel, and 50 to 75 miles (80 to 120 kilometers) west of the former plate boundary. These structures have been described as offshore extensions of the Pinar fault in Cuba (e.g., Reference 529). No seismicity from the Phase 1 or Phase 2 earthquake catalogs (Subsection 2.5.2.1) are coincident or associated with those faults. After passage of the Caribbean Plate, the eastern margin of the Yucatan Platform became passive as sinistral faulting associated with northeastern motion of the Caribbean Plate shifted to the Oriente fault and the adjacent Yucatan Basin sutured to the North America Plate.

At the extreme southeastern corner of the Yucatan Platform, offshore Belize, Lara (Reference 819) mapped a series of Cretaceous to Eocene left-lateral transtensional faults and a set of Pliocene high-angle normal faults using seismic reflection data. The earlier structures reflect the Cretaceous-Eocene strike-slip boundary as the Greater Antilles Arc moved past the southeastern Yucatan Platform. The youngest structures may be influenced by the Cayman trough rifting to the east (Reference 819).

#### Seismicity of the Yucatan Platform

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates that earthquakes in the Yucatan Platform are small to moderate in magnitude and concentrated towards the south near the Polochic fault (Figure 2.5.1-202) and to the southeast near the southwest extension of the Nortecubana fault (Figure 2.5.1-267). The proximity of these earthquakes to these features suggests possible association of seismicity with the active Polochic-Motagua fault system, and possible crustal weakness associated with the southwest extension of the Nortecubana fault. Aside from these two possible examples, the overall seismicity pattern within the Yucatan Platform shows no correlation with geologic or tectonic features (Subsection 2.5.2.3).

### 2.5.1.1.2.1.3 Geology of the Yucatan Basin

#### Physiography of the Yucatan Basin

The major physiographic features of the Yucatan Basin include the abyssal plain, occupying the western and northern half of the province, which gives way southward to faulted bank areas and culminates on the southern boundary of the province with the shallow water Cayman Ridge and its emergent Cayman Islands (Figure 2.5.1-210).

The Yucatan Basin lies between the Yucatan Peninsula and Cuba and the east-northeast-trending Cayman Ridge. On the west, the fault-controlled Yucatan Strait separates the Yucatan Platform from a narrow strip of carbonate platform at the western margin of the Yucatan Basin. To the north and northeast, the Yucatan Basin and its ridges dip beneath the Cuba margin along a sediment-filled trench (Reference 529). To the south, the Yucatan Basin is separated from the Cayman Trench by the Oriente fault system (Figures 2.5.1-202, and 2.5.1-210).

The Yucatan Basin itself is separated into a deeper (4000 to 4600 meters or 13,000 to 15,000 feet) northwestern part containing the Yucatan Plain and a shallower (2000 to 3500 meters or 6500 to 11,500 feet) southeastern part that is dominated by ridges (the Cayman Ridge on the south and the more subdued Camaguey Ridges to the northeast) that strike northeast across the basin. Linear, sediment-filled basins lie between the Cayman and Camaguey Ridges (Reference 526) (Figure 2.5.1-296).

The Cayman Ridge trends west-southwest from the Sierra Maestro of southern Cuba to within 100 kilometers (60 miles) of the base of the Honduras continental slope where it disappears beneath thick sediment cover in the Yucatan Basin. Over much of its length, a double ridge crest separates small perched basins, valleys, or flats (References 526 and 499). The Cayman Islands; Misteriosa, Pickle, and Rosario Banks; and some isolated algal reefs lie near or above sea level on top of the Cayman Ridge (References 527 and 528) (Figure 2.5.1-296).

#### Stratigraphy of the Yucatan Basin

Surficial pelagic-hemipelagic sediments of the Yucatan Basin consist of foraminifera- and pteropod-rich chalk marl oozes and marl clays. Chalk oozes predominate on the elevated southeastern portion of the basin. Marl oozes predominate within the turbidite-lutite sequences of the Yucatan Basin, reflecting influx of sediments from terrigenous sources. Turbidites consist of a heterogeneous series of terrigenous sands, muds, and carbonate sands (Reference 526).

The Belize Fan feeds terrigenous sands and muds into the southwest area of the Yucatan Basin abyssal plain. The primary sediment sources for the Belize Fan are the Polochic, Motagua, Chamelecon, and Ulua rivers that flow from the mountains of Guatemala and Honduras. The rivers converge at the head of the Belize and Motagua Fans. The Yucatan Basin Slope gradients reverse in its eastern extension, leading upslope toward the mouth of the Cauto River. The Cauto River drains much of the Sierra Madre Oriental of Cuba. A well-developed drainage network funnels pelagic carbonate sediments into the Yucatan Plain from the shallower portion of the Yucatan Basin. In addition, carbonates are also brought in from the continental and island slopes of Yucatan and Cuba via canyons (Reference 526).

Based on seismic reflection data, including extensive multi-channel data, Rosencrantz (Reference 529) concludes that the Yucatan Basin is underlain by crust of complicated internal structure, composed of oceanic crust of two different origins plus continental crust, distributed across the former North America-Caribbean Plate boundary between the Yucatan Platform and the Yucatan Basin. Rosencrantz (Reference 529) identifies three distinct crustal types or blocks. The first crustal type underlies the western flank of the basin and includes metasediments lithologically similar to Paleozoic continental rocks found at depth across the Yucatan Platform. This crustal type is postulated to represent the offshore continuation of the adjacent Yucatan Platform. The possible relationship between the crust of the western flank of the basin and that of the Yucatan Platform and the Florida and Bahama Platforms are described in Subsection 2.5.1.1.1.2.1.2. The second includes the topographically heterogeneous areas of the eastern two thirds of the basin (including the Cayman Rise, Cayman Ridge, and Camaguey Trench) and is dominated by a subsided volcanic rise or arc resting on probably oceanic crust of pre-Tertiary age. The eastern edge of the rise and adjacent basins dip northeast beneath the Cuban margin along the sediment filled Camaguey Trench. The third type of crust occupies a rectangular deep area within the western third of the basin. Available evidence indicates that this crust is oceanic and represents a large, mature pull-apart basin set within a wide paleo-transform zone between the western platform and the eastern oceanic basin (Reference 529). The oceanic crust was produced by back-arc spreading behind the Cuban Arc (References 210 and 526).

Seismic reflection profiles and regional gravity interpretations suggest that the crust beneath the deep north-central and western parts of the Yucatan Basin is oceanic, but that crust thickens southward to more than 20 kilometers (12 miles) beneath the Cayman Ridge (Reference 529). K/Ar cooling ages of volcanic, metavolcanic, and granodiorite rocks dredged from the southern wall of the Cayman Ridge indicate ages of 59 to 69 Ma. This suggests that the thicker crust represents a buried Late Cretaceous island arc resting on Late Cretaceous or older crust (Reference 528). Lewis et al. (References 810 and 811) analyzed Nd-Sr and Pb isotope ratios of arc-related calc-alkaline granitoids and volcanic rocks from the western part of the Cayman Ridge and indicate that these rocks were intruded into continental crust. This confirms that crustal rocks of the western Cayman Ridge are the rifted eastern extension of the continental Maya block of Belize, Mexico, and Guatemala, as has been suggested previously (Reference 815) (see related discussion in Subsection 2.5.1.1.2.2.2).

Inferred oceanic crust from the deep western part of the basin appears to be younger (Late Cretaceous to Eocene) on the basis of heat flow (Reference 530) and depth-to-basement measurements (References 526 and 222). Pindell et al. (Reference 525) use ages of pull-apart basin faults offsetting age-dated sediments in the surrounding region to estimate an age for the initiation of rifting as late Middle Eocene, or about 45 Ma, in this portion of the basin (Reference 529).

Based on multichannel seismic reflection lines across the basin, Rosencrants (Reference 529) finds that Yucatan Basin abyssal sediments are mostly undisturbed, indicating that the basin has been tectonically quiescent since spreading ceased in the Late Eocene (see Subsection 2.5.1.1.3 for geologic history). The basement relief at the southern portion of the Yucatan Basin has the

appearance of tilted fault blocks, which suggests the possibility that distension, rifting, and foundering of preexisting crust occurred during the opening of the basin (Reference 526).

### Structures of the Yucatan Basin

At its nearest point, the Yucatan Basin lies 260 miles (420 kilometers) southwest of the Units 6 & 7 site. Its convex-northwest margin represents a portion of the former sinistral transform and oblique-convergent margin of the Caribbean Plate. Its relatively linear southern margin is defined by the east-northeast striking Cayman Trough and sinistral strike-slip western Oriente fault system (Reference 492) (Figure 2.5.1-229).

Structure within the Yucatan Basin is limited to Eocene and older basement rocks that are overlain by relatively undeformed post-Eocene cover of oceanic sediments (References 530, 529, and 525). Deformation of sedimentary cover over basement rocks mostly is due to gravitational adjustments, such as slumping over the pervasively steep and irregular basement surface, and exhibits little to no deformation related to late Cenozoic tectonics of the current plate boundary (Reference 529).

The origin of basement structure in the Yucatan Basin is associated with Late Cretaceous (Maastrichtian) through Late Eocene east- and northeast-directed subduction of the proto-Caribbean ocean crust beneath the Caribbean Plate. During this time, the Caribbean Plate passed between the bottleneck formed between the Yucatan Platform on the North America Plate to the north, and the South America Plate (Figure 2.5.1-297). Beginning at 72 Ma, motion of the northwestern portion of the plate, the Escambray terrane, was directed northwest while the remainder of the plate was directed northeast. These motions imparted stresses, causing sinistral oblique subduction of the Yucatan Platform and proto-Caribbean oceanic crust beneath the Escambray terrane that persisted until 56 Ma. Beginning at 56 Ma, rollback of the proto-Caribbean crust caused counter-clockwise rotation of the Yucatan Platform and redirection of the Escambray terrane vector to the northeast, subparallel to the vector of the remainder of the Caribbean Plate (References 525 and 523). The redirection of the Escambray Terrace vector during the Eocene formed a pull-apart basin bound by sinistral normal faults that leaked new ocean crust and marked the North America-Caribbean Plate boundary. During this time, the La Trocha and Trans Basin faults developed within the Yucatan Basin and the Oriente fault developed along the southern margin of the basin to accommodate the differentially directed vectors between the Escambray terrane and the remainder of the Caribbean Plate (References 529 and 525) (Figure 2.5.1-297). A consequence of this model for the opening of the Yucatan basin is that the northeast-striking faults (such as the Pinar, La Trocha, and Nipe faults) would have initiated as mainly normal fault structures. However, available kinematic data on the Pinar, for example, indicate it is a left-lateral strike-slip structure (Reference 697), and therefore, support a different opening style (e.g., Reference 639).

The primary structures within the Yucatan Basin are the Trans Basin fault and faults associated with the pull-apart structure formed during the Late Cretaceous (Maastrichtian) through Middle Eocene opening of the Yucatan Basin. The La Trocha fault strikes east-northeast in Cuba, within the Greater Antilles deformed belt province, and continues southwest as the Trans Basin fault across the Yucatan Basin (Figure 2.5.1-286). The Trans Basin fault is identified in four fault-normal seismic reflection profiles and diagrams from Rosencrantz (Reference 529). Rosencrantz (Reference 529) interprets about 50 kilometers (31 miles) of displacement along the fault, estimated from onshore geologic relations of the La Trocha fault and offshore offset of a graben by the Trans Basin fault. Displacement along these faults occurred during the latest Paleocene through Middle Eocene. Also during this time, the crustal block east of the sinistral Trans Basin fault was subducted beneath Cuba along the presently inactive Camaguey Trench (Reference 529). The Camaguey Trench delineates the boundary between the Greater Antilles deformed belt and Yucatan Basin provinces of French and Schenk (Reference 492), and terminates to the west at the La Trocha-Trans Basin fault. Subduction along the Camaguey Trench is thought to have been active either during the Cretaceous as a part of the Cuban Arc, or during the Eocene as a back thrust behind the Cuban Arc. Seismic reflection profiles and diagrams in Rosencrantz (Reference 529) indicate that the trench is presently buried by

several kilometers of undeformed oceanic sediments. There is no stratigraphic or geomorphic evidence for any activity along the Trans Basin fault since the Middle Eocene. However, the onshore La Trocha fault (in the Greater Antilles deformed belt geologic province) is considered Pliocene-Quaternary seismoactive by Cotilla-Rodríguez et al. (Reference 494), who correlate five macroseismic events with the fault. Additionally, only two Phase 2 earthquake catalog earthquakes of $M_w \geq 7$ are located within the Yucatan Basin, one of which ($M_w$ 7.7) is located well within the province margins and nearly coincident with the Trans Basin fault mapped by Rosencrantz (Reference 529). Five other earthquakes ($M_w$ 3 to 4.6) from the Phase 2 earthquake catalog (Subsection 2.5.2.1) lie within close proximity of the Trans Basin fault, suggesting it may have some seismogenic potential within the Yucatan Basin.

The pull-apart structure and associated faults (Figure 2.5.1-297) as the “Eocene Ocean” accommodated about 350 kilometers (217 miles) of cumulative oblique sinistral extension between the Caribbean and North America plates between the Late Paleocene to Middle Eocene (References 529 and 525). A cluster of 15 historical earthquakes ($M_w$ 3.5 to 6.4) from the Phase 2 earthquake catalog (Subsection 2.5.2.1) occurs in the southwest corner of the Yucatan Basin. The cluster is coincident with the Eocene pull-apart structure and associated faults, and likely represents seismogenic reactivation of the faults due to far-field stresses caused by the Oriente fault that lies 5 to 60 miles (8 to 100 kilometers) to the south.

#### Seismicity of the Yucatan Basin

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates moderately abundant earthquakes within the Yucatan Basin (Figure 2.5.1-267). The preponderance of these is concentrated at the margins of the basin near the Oriente fault near southwestern Cuba and near the west end of the Swan Islands fault zone. Additionally, the Phase 2 earthquake catalog indicates moderately abundant earthquakes that range from $M_w$ 3.1 to 7.5 in the eastern corner of the Yucatan Basin. These events likely are associated with far-field stress in normal faults striking parallel to the Oriente fault, located approximately 15 to 80 miles (25 to 130 kilometers) south in the Cayman Trough (Reference 529). Several additional earthquakes occur in interior to the province, about 100 kilometers (60 miles) or more from known active faults.

### 2.5.1.1.2.1.4 Geology of the Charleston, South Carolina, Seismic Zone

#### Physiography of the Charleston, South Carolina, Seismic Zone

The Charleston, South Carolina, seismic zone is located along the Atlantic coast of South Carolina, within the Coastal Plain geologic province. Elevations range from sea level in the southeast map area to 114 feet (35 meters) in the northwest, reflecting a gentle net regional slope to the southeast of about 2.8 feet/mile (5 0.53 meters/kilometer). Locally, steep bluffs along major rivers may expose a few feet of Tertiary sediment. Elsewhere, the Charleston region is covered by a ubiquitous blanket of lower Pleistocene to Holocene sand and clay that obscures the distributional pattern of underlying Tertiary stratigraphic units.

Landsat imagery and topographic maps of the South Carolina coastal plain indicate that the courses of the Santee, Black, Lynches, and Pee Dee rivers and the Caw Caw Swamp are noticeably curved toward the north-northeast along a 15-kilometer (9-mile) wide, 200-kilometer (125-mile) long zone from south-southwest of Summerville, South Carolina to just east of Florence, South Carolina (Reference 533) (Figure 2.5.1-298). Other river anomalies observed within the zone include incised channels, changes in river patterns, and convex-upward longitudinal profiles (References 533 and 534). While these anomalies may indicate a lithologic boundary formed by a paleo-shoreline, the trend of the zone of anomalies does not parallel the trends of other paleo-shorelines. Marple and Talwani (References 533 and 534) conclude that this zone is likely due to tectonic deformation.

### Stratigraphy of the Charleston, South Carolina, Seismic Zone

The Coastal Plain sediments in Georgia and South Carolina mostly consist of unlithified sediments interbedded with lesser quantities of weakly lithified to indurated sedimentary rocks (Reference 775). Lithologies include stratified sand, clay, limestone, and gravel. These units dip gently seaward and range in age from Late Cretaceous to Recent. The sedimentary sequence thickens from 0 feet at the Fall Line to more than 3962 feet (1219 meters) at the coast (Reference 536). Regionally, rocks and sediments dip and thicken toward the southeast, but dips and thicknesses vary owing to the presence of a number of arches and embayments within the province (Figure 2.5.1-299).

The shallow subsurface Tertiary stratigraphy of the greater Charleston, South Carolina region reflects the tectonic development and setting of the region over the past 34 m.y. Upper Eocene and Oligocene stratigraphic horizons show a net regional dip toward the southwest or south, whereas Miocene and Pliocene horizons show a shift to net regional dips toward the southeast (Reference 775).

A number of localized areas show persistent net upward or downward motion attributed to Tertiary crustal adjustments (Reference 534).

### Structures of the Charleston, South Carolina, Seismic Zone

The August 31, 1886, earthquake that occurred near Charleston, South Carolina, 500 miles (800 kilometers) north of the Units 6 & 7 site, is the largest historical earthquake in the eastern United States. The event produced MMI X shaking in the epicentral area (Figure 2.5.2-212) and was felt as far away as Chicago (Reference 538).

As a result of this earthquake and the relatively high seismic risk in the Charleston area, government agencies funded numerous investigations to identify the source of the earthquake and the recurrence history of large magnitude events in the region. Because no primary tectonic surface deformation was identified with the 1886 event, a combination of geology, geomorphology, and instrumental seismicity data have been used to suggest several different faults (East Coast fault system, Woodstock fault, and Ashley River fault) as the source for Charleston seismicity. However, the source of the 1886 earthquake has not been definitively attributed to any particular fault.

### Seismicity of the Charleston, South Carolina, Seismic Zone

Seismicity data in the Charleston, South Carolina region include historical accounts of the large 1886 Charleston earthquake (Phase 1 earthquake catalog Emb 6.75), instrumental records of low-magnitude events, and paleoliquefaction studies describing the occurrence of large prehistoric earthquakes in coastal South Carolina.

Estimates of the magnitude of the 1886 Charleston earthquake generally are in the high-6 to mid-7 range. For example, Martin and Clough (Reference 537) base their $M_w$ 7 to 7.5 estimate on a geotechnical assessment of liquefaction features produced by the 1886 earthquake. Johnston (Reference 538) estimated a $M_w$ 7.3 ± 0.26 for the 1886 Charleston event, based on an isoseismal area regression accounting for eastern North America anelastic attenuation. More recently, Bakun and Hopper (Reference 539) indicate a best estimate of $M_w$ 6.9, with a 95 percent confidence level corresponding to a range of $M_w$ 6.4 to 7.1. Bakun et al. (Reference 758) indicate that the 1886 Charleston earthquake was felt as far south as Key West, Florida with Modified Mercalli Intensity (MMI) III. Additionally, five felt reports indicate MMI III to IV in the Tampa-St. Petersburg-Fort Meade, Florida area (Reference 758). One felt report from Fowey Rocks Lighthouse in Biscayne Bay, Florida indicates MMI IV for the 1886 Charleston earthquake.

Based on local seismic networks, three zones of elevated microseismic activity have been identified in the greater Charleston area. These include the Middleton Place-Summerville, Bowman, and Adams Run seismic zones. The Middleton Place-Summerville seismic zone is an area of elevated

microseismic activity located approximately 12 miles (20 kilometers) northwest of Charleston (References 540, 541, 542, 543, and 544). Between 1980 and 1991, 58 events with duration magnitude (Md) 0.8 to 3.3 were recorded in a 7- by 9-mile (11- by 14 kilometer) area, with hypocentral depths ranging from approximately 1 to 7 miles (0.5 to 11 kilometers) (Reference 542). Seven events from this zone are listed in the Phase 1 catalog, with Emb values ranging from 3.30 to 3.51. The elevated seismic activity of the Middleton Place-Summerville seismic zone has been attributed to stress concentrations associated with the intersection of the postulated Ashley River and Woodstock faults (References 545, 542, 546, and 543). Some investigators speculate that the 1886 Charleston earthquake occurred within this zone (References 539, 546, and 544). The Bowman seismic zone is located approximately 50 miles (80 kilometers) northwest of Charleston, South Carolina, outside of the meizoseismal area of the 1886 Charleston earthquake. The Bowman seismic zone is identified on the basis of a series of local magnitude ($M_L$) $3 < M_L < 4$ (Emb 3.14 to 4.28 in the Phase 1 earthquake catalog) earthquakes that occurred between 1971 and 1974 (References 540 and 547). The Adams Run seismic zone, located within the meizoseismal area of the 1886 Charleston earthquake, is identified on the basis of four magnitude <2.5 earthquakes (not listed in the Phase 1 earthquake catalog), three of which occurred in a two-day period in December 1977 (Reference 544). Bollinger et al. (Reference 540) downplay the significance of the Adams Run seismic zone, noting that, in spite of increased instrumentation, no additional events were detected after October 1979.

Liquefaction features are recognized in the geologic record throughout coastal South Carolina and are attributed to both the 1886 Charleston and earlier moderate- to large-magnitude earthquakes that occurred in the region since mid-Holocene time (e.g., References 548, 549, 550, 551, and 552). Paleoliquefaction features predating the 1886 Charleston earthquake are found throughout coastal South Carolina. The spatial distribution and ages of paleoliquefaction features in coastal South Carolina constrain possible locations and recurrence rates for large earthquakes (References 548, 549, 550, 551, and 552). Talwani and Schaeffer (Reference 553) combined previously published data with their own studies of paleoliquefaction features in the South Carolina coastal region to derive possible earthquake recurrence histories for the region. Talwani and Schaeffer (Reference 553) describe two alternative paleo-earthquake scenarios that include both moderate (approximately $M_W$ 6+) and large (approximately $M_W$ 7+) earthquakes (Table 2.5.2-215), and they estimate a 500- to 1000-year recurrence of large earthquakes in the Charleston region since mid- to late-Holocene time, with a preferred estimate of approximately 550 years.

#### 2.5.1.1.2.2 Geology of the Caribbean Plate Provinces

This subsection includes a description of the physiography, stratigraphy, structure, and seismicity of portions of the Caribbean Plate near its boundary with the North America Plate. Due to their remote distance from the Units 6 & 7 site, features of the Caribbean-South America Plate boundary are not discussed in this subsection.

##### 2.5.1.1.2.2.1 Geology of the Cayman Trough

**Physiography of the Cayman Trough**

The Cayman Trough (Figures 2.5.1-210 and 2.5.1-202) is an elongated deep basin, oriented west-southwest to east-northeast, that extends 1600 kilometers (1000 miles) from the Windward Passage between Cuba and Hispaniola to the Gulf of Honduras. Images from the long-range side-scan sonar instrument Geological LOng-Range Inclined Asdic (GLORIA) elucidate the morphology of the walls and floor of the trough. The rectangular basin is bounded to the north and south by steep scarps that locally rise more than 5000 meters (16,400 feet) from the basin floor. These scarps are or have been active transform faults (Swan Islands fault zone to the north and the Oriente fault zone to the south) during the development of the basin. The greatest depth, 6800 meters (22,300 feet), occurs adjacent to the north wall between Grand Cayman Island and Cuba. The northern boundary of the basin, south of the Yucatan Basin and the Cayman Ridge, marks the

boundary of the Caribbean and North America Plates. Note that the terminology “Cayman Ridge” refers to the line of islands and shoals that include the Cayman Island chain. This reflects normal usage in Caribbean literature but is distinct from the terminology used in French and Schenk (Reference 492), who use the term, contrary to other geologic literature, to designate the north portion of the northern Nicaraguan Rise.

The Cayman Trough has three morphologic areas. On the western third of the trough, a relatively flat abyssal plain lies at a depth of about 5000 meters (16,400 feet). The central third of the trough lies at a depth of about 5500 meters (18,000 feet) and includes an active spreading center characterized by north-south-trending ridges. The eastern third of the trough is an abyssal basin that lies at a depths of between 4000 and 6800 meters (13,100 to 22,300 feet) and exposes the tops of older southeast- to northwest-trending ridges. The change in ridge orientation between the eastern and central portions of the trough records a change in spreading direction. The history of relative motion recorded in the crust accreted at this spreading center both outlines the age and duration of tectonic events along the northern boundary of the Caribbean and provides a measure of constraint over the relative motions between the Caribbean Plate and surrounding plates (Reference 222).

The active spreading center in the central region represents younger rocks with older rocks to the east and west (Reference 554). This north-south spreading axis is very short, 150 kilometers (90 miles) long and 30 kilometers (19 miles) wide. The rift valley is deep (5500 meters [18,000 feet] average depth with a maximum depth of 6000 meters [20,000 feet]) and is flanked by rift mountains with peaks of 2500 meters (8200 feet) deep. The average strike of the spreading zone is about 080° (Reference 499). As the North America Plate moves westward relative to the Caribbean, a continual opening takes place at the spreading center, which is filled with upwelling mafic asthenospheric material and hardens to form new oceanic crust (References 555, 528, and 499). Ten to fifty meters (33 to 164 feet) of the spreading axis is characterized as a series of volcanic ridges, cones, and depressions in a 2 to 3 kilometers (1.2 to 1.9 miles) wide belt that parallels the valley walls (Reference 555). The rift valley walls rise abruptly from the edge of the rift valley and consist of a series of fault escarpments and ledges that form inward facing steps a few meters to tens of meters in relief. Subsequent erosion and the formation of talus ramps have modified the small-scale morphology to a minor extent (Reference 527).

### Stratigraphy of the Cayman Trough

The composition and age of the rock units cropping out in the Cayman Trough are derived from 80 dredge hauls on Duke University's research vessel *Eastward* during 1971, 1972, and 1973 and 94 sampling stations from the research vessels *Knorr* in 1976 and *Oceanus* in 1977, in addition to geophysical data. In general, the Cayman Ridge and northern Nicaraguan Rise are composed of metamorphic, plutonic, volcanic, sedimentary, and carbonate rock units. The trench floor is composed of mafic and ultramafic rocks (References 528 and 555). The Cayman Trough has four distinctive morphotectonic regions: eastern Cayman Trough, Cayman Ridge, northern Nicaraguan Rise (Subsection 2.5.1.1.2.2.5.1), and the mid-Cayman spreading center.

The eastern Cayman Trough covers the area south of the Sierra Maestra (Cuba) and consists of granodiorites, tonalites, and basalts that exhibit various degrees of alteration and metamorphism. Limestone, large manganese nodules, thick manganese plates, and coral were sampled at shallower depths between the Sierra Maestra and Jamaica (Reference 528).

The diverse rocks along the Cayman Ridge (Figure 2.5.1-202) are located in the area west of the Sierra Maestra. In the deepest part of the ridge, metamorphic and plutonic rocks with lesser amounts of volcaniclastics, volcanics, and late Cretaceous and late Paleocene shallow-water carbonates crop out (>2500 meters or >8200 feet). Along the western end of the ridge, amphibolites, gneisses, and micaceous schists were retrieved from the dredge samples. However, the predominant rock type recovered was a medium to coarse-grained hypautomorphic-granular granodiorite. Greenschist grade metamorphism and cataclastic textures occur frequently in the plutonic rocks; these are similar

to those found south of the Sierra Maestra (Reference 528). Extrusive rocks range from basalt to aplitic rhyolite, but the majority are andesites or dacites. The colors of the volcanic rocks are purple to reddish due to enrichment of oxidized mafic and opaque minerals. The pyroclastics exhibit virtoclastic textures with various degrees of alteration and devitrification. Some of the tuffs are intercalated with microfossil bearing carbonates and clays. The sedimentary rocks appear as outcrops along the ridge and consist of volcanic breccia, conglomerate, arenites, and argillite that are composed mostly of igneous fragments with small amounts of mineral, clastic, metamorphic, and biogenic clasts in clay matrices. Also present are nonvolcanic argillite, graywacke, arkose, and conglomerate. Lastly, the carbonate constituents range in age from Miocene to Pleistocene and are generally micritic, planktonic oozes with some reef limestones with abundant shallow-water biologic material such as coquina, sponges, coral, algal balls, sea biscuits, echinoids, and mollusk shells (Reference 528).

The mid-Cayman spreading center has a crustal sequence identical to the mid-oceanic ridge (Reference 528). Dredging samples consisted of serpentinite (with minerals of, orthopyroxene, clinopyroxene, and spinel) and probably pseudomorphs after olivine. This is indicative of mineral assemblages that are stable in the mantle at depths of 25 to 70 kilometers (15 to 45 miles). The samples indicate that they were crystallized from a melt in the crust or very shallow mantle (Reference 558). The Oriente fault yielded dredge samples consisting primarily of serpentinite and serpentinized peridotite with minor quantities of graywacke and basalt. Serpentinized peridotite and coarse gabbro were dredged from the walls of the mid-Cayman spreading center. Dolerite and basalt were retrieved on outcrops higher along the escarpments. Lesser amounts of metavolcanics, metasediments, marble, and limestone were sampled from the dredge hauls. The amount of carbonate dredged from the mid-Cayman spreading center was slight and difficult to identify as in situ on the top of the ridges. Most are micritic limestones with pelagic forams and minor angular fragments of plagioclase, clinopyroxene, chlorite, iddingsite, amphibole, and epidote. It is possible that these limestones formed at depth within the trench as seen from a lack of shallow-water fossils and granitic detritus (Reference 528).

### Structures of the Cayman Trough

The Cayman Trough comprises a central north-northwest-trending spreading axis, with strike-slip faults extending both east and west from its southern terminus and a strike-slip fault extending east from its northern terminus (Figure 2.5.1-202). Extending east from the northern end of the spreading axis is the left-lateral Oriente fault, which connects with the Septentrional fault on the island of Hispaniola. From the southern end of the spreading axis, the Swan Islands fault extends to the west, eventually linking with the Motagua fault in Guatemala and Honduras.

To the east of the southern end of the spreading axis, the Walton, Duanvale, and Enriquillo-Plantain Garden faults extend eastward through Jamaica to Hispaniola. The submarine portions of these structures were mapped with the aid of the SeaMARC II sidescan instrument (Reference 559). The spreading axis itself is offset by a short discontinuity. Seismicity indicates this is a left-lateral strike-slip fault (Reference 499). The Oriente fault is described in detail in Subsections 2.5.1.1.1.3.2.4 and 2.5.1.1.2.3.1.2.

The Swan Islands transform includes continuous bathymetric lineaments defined by small scarps, furrows, sag structures, or en echelon folds and fissures offshore (Reference 559). The Swan Islands are formed by a right-step in the left-lateral fault, which creates a restraining bend, and the islands rise about 5000 meters (16,400 feet) relative to the seafloor in the adjacent portions of the trough. The overlapping segments of the fault, known as the East and West Swan Islands faults, overlap west of the Swan Islands and come to within 12 kilometers (7.5 miles) of each other (Reference 559). Analyses of magnetic anomalies in the seafloor indicate that this fault has been active since sometime between 50 and 30 Ma (Reference 499). Detailed information regarding this structure, which is an active tectonic fault and a seismic source zone, is found in Subsection 2.5.1.1.2.3.1.1.

The Walton fault extends for about 185 miles (300 kilometers) eastward from the southern end of the mid-Cayman spreading center to northwestern Jamaica (Reference 766). Slip is transferred from the Walton fault across the island of Jamaica through a broad restraining bend that includes the east-west striking Duanvale, Rio Minho-Crawle River, South Coast, and Plantain Garden faults (Reference 503). The geometry of the Walton-Duanvale fault is more complicated than the Swan Island transform, with pull-apart and pop-up structures intersecting its sinuous trace (Reference 559). The Walton-Duanvale fault probably developed in the late Miocene (Reference 559). The topography of Jamaica results from the complicated interaction of the Walton-Duanvale fault system and the Enriquillo-Plantain Garden fault system (Figure 2.5.1-300). This entire system is described in more detail in Subsection 2.5.1.1.2.3.2.3.

#### Seismicity of the Cayman Trough

The Cayman Trough includes major active plate boundary structures, including the spreading axis and the Swan Islands, Enriquillo-Plantain Garden, Walton-Duanvale, and Oriente faults. The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates abundant large earthquakes in this area (Figure 2.5.1-267). These earthquakes are concentrated along these major active plate boundary structures and seismicity mapping of the region clearly identifies the gross fault structure of the Cayman Trough (e.g., Reference 813) (Figure 2.5.1-267). Subsection 2.5.1.1.2.3.1 provides additional discussion regarding the active tectonic structures of the Cayman Trough and associated instrumental and historical seismicity.

### 2.5.1.1.2.2.2 Geology of the Southeastern Greater Antilles

The Greater Antilles are a group of Caribbean islands comprised of Jamaica, Cuba, Hispaniola, Puerto Rico, and the Cayman Islands. Due to its location relative to the Units 6 & 7 site, Subsections 2.5.1.1.1.1.3 and 2.5.1.1.1.2.3 include descriptions of Cuba in some detail. This subsection describes the physiography, stratigraphy, structures, and seismicity of the islands of Jamaica, Hispaniola, and Puerto Rico. The Cayman Islands are discussed in Subsection 2.5.1.1.2.1.3 as part of the Cayman Ridge of the Yucatan Basin.

Mattson (Reference 804) and Pindell and Barrett (Reference 219) propose that Cretaceous igneous rocks of the Greater Antilles islands of Cuba, the Cayman Ridge, Hispaniola, and Puerto Rico originated in an intra-oceanic island arc, with northeast-dipping subduction, bounding one edge of a proto-Caribbean Sea (Figure 2.5.1-347, part B). In these models, attempted subduction of a Pacific-derived oceanic plateau (the Caribbean ocean plateau) caused the Greater Antilles Arc to reverse its polarity to south-southwest-dipping subduction. The arc then migrated to the north-northeast, consuming the Jurassic to Early Cretaceous proto-Caribbean ocean crust. Based on lithologic types and metamorphic rock ages in Cuba, Dominican Republic, and Puerto Rico, Mattson (Reference 804) suggests that the arc polarity reversal occurred during the latest Early Cretaceous (120-130 Ma) and that renewed subduction began during the early Late Cretaceous (110 Ma) and ended by middle Late Cretaceous (85 Ma). Mattson (Reference 804) notes that volcanism ceased by 85 Ma. Pindell and Barrett (Reference 219) note that obducted ophiolites (the Bermeja Complex of Puerto Rico) were accreted to the south side of the island before about 95 Ma (the middle Late Cretaceous or Campanian time), suggesting north-dipping subduction. Pindell and Barrett (Reference 219) also note that after about 80 Ma (middle Late Cretaceous or Santonian to Campanian time) ophiolitic complexes were emplaced on the north side of the arc, suggesting south-dipping subduction. Draper et al. (Reference 808) cite new structural data from central Hispaniola, suggesting that a mid-Cretaceous orogenic event resulted in the obduction of peridotites onto the early Great Antilles Arc in the late Early Cretaceous (Aptian-Albian). Draper et al. (Reference 808) and Draper and Barros (Reference 834) also note that this event is synchronous with chemical changes of the arc magmas in Hispaniola, Puerto Rico, and central Cuba, and thus, both may be related to the postulated Greater Antilles Arc polarity reversal.

The geologic evidence used in these early models to support Cretaceous subduction polarity reversal is the present-day out crop of older (Jurassic-Early Cretaceous [?]) high pressure/low temperature metamorphic rocks along the southern flank of arc rocks in Cuba and Puerto Rico and younger (Late Cretaceous-early Tertiary [?]) high pressure/low temperature metamorphic rocks along the northern flank of arc rocks in Cuba and Hispaniola (Reference 833). Nearly 30 years after Mattson's work (Reference 804), the basic model for the development of the Greater Antilles Arc has been tested and is still the most accepted model of early development of the Greater Antilles Arc and the Caribbean Plate (Reference 807).

The Cretaceous-Eocene island arc rocks of the northeastern Caribbean can be subdivided into a basal Late Jurassic (?) to Early Cretaceous primitive island arc (PIA) suite and an overlying Late Cretaceous-Oligocene calc-alkaline (CA) rock suite (Reference 568) (Figure 2.5.1-301). Pindell and Barrett (Reference 219) consider intermediate and calc-alkaline plutons, lavas, and tuffs as evidence of subduction. They find that the period over which each arc was volcanically or magmatically active correlates approximately with the period of active subduction. Calc-alkaline arc activity in the northeastern Caribbean terminated in Eocene-Oligocene time by collision of the arc with the Bahama carbonate platform (Reference 219).

Although the local stratigraphy and structure of arc rocks of the islands of Greater Antilles is complex, a striking correlation exists between Late Cretaceous-Eocene volcanic arc-related lithologies and intercalated siliciclastic and carbonate deposits. The Cretaceous-Paleogene histories of island arc development in Cuba, the Cayman Ridge, Hispaniola, and Puerto Rico are similar, suggesting that these islands belonged originally to the same arc system. According to Pindell and Barret (Reference 219), westernmost and north-central Cuba may be continental and unrelated to the Greater Antilles Arc and the island of Jamaica, part of the Greater Antilles island group, may be part of a different volcanic arc (the Chortis arc of Meschede and Frisch [Reference 856] or the Nicaragua-Jamaica Arc of Pindell and Barrett [Reference 219]). Pindell and Barrett (Reference 219) assert that the Nicaragua-Jamaica Arc included pre-Mesozoic continental crust in Jamaica and the northern Nicaraguan Rise, those areas may be genetically related to the Chortis arc of southern Guatemala, Honduras, northern Nicaragua and propose that the western Nicaraguan Rise. In contrast, Mann et al. (Reference 814) propose that the crust underlying the northern Nicaraguan Rise and Jamaica is not continental but of volcanic island arc origin.

#### 2.5.1.1.2.2.2.1 Geology of Jamaica

### Physiography of Jamaica

Jamaica is the third largest of the Greater Antillean islands and lies at the edge of the seismically active plate boundary between the North America and Caribbean Plates (References 560 and 493). The island is approximately 130 kilometers (80 miles) long and 80 kilometers (50 miles) wide, with a total area of 10,991 kilometers$^2$. It is the emergent part of the eastern apex of the Nicaraguan Rise and is separated from the North America Plate by the Cayman Trough. Over 60 percent of the surface outcrop is limestone that has been extensively karstified (Reference 217).

The physiography of Jamaica resembles the other islands of the Greater Antilles, with its mountains, limestone plateaus, and steep seaward slopes rising abruptly from a coastal plain that in most places is extremely narrow. The Blue Mountains (maximum elevation 7388 feet [2250 meters]) begin near the east end of the island and parallel the northeast coast for about a third of its length. The Blue Mountains represent the eroded core of an ancient volcanic arc, once much more extensive. Over the western two-thirds of the island and partly encircling the Blue Mountains is a plateau of white limestone that arches gently down to the north and south. Another ancient volcanic core exposed by erosion of this plateau forms several small chains (maximum elevation 3165 feet or 965 meters) with deeply cut flanks that parallel the axis of the Blue Mountains (Reference 217).

The tropical to subtropical climate of Jamaica results in the deep weathering of volcanic sediments that underlie the Blue Mountains. This weathering forms deep residuals soils that are highly susceptible to both rainfall and earthquake-induced landslides.

### Stratigraphy of Jamaica

Jamaica is composed of Cretaceous and Tertiary rocks (Figures 2.5.1-302 and 2.5.1-303) exposed in blocks and belts across the island. The blocks of Jamaica are Cretaceous in age. Fault-bounded belts of younger (Tertiary) rocks flank and separate the blocks (Reference 805) (see the Structures subsection for Jamaica). Pliocene and Quaternary rocks, found mostly around the coast, consist of patch reef (carbonate) sediments with some subaerial to submarine fanglomerates. Late Pleistocene through Holocene sediments are neritic and form a series of raised marine terraces (Reference 217).

Three main structural blocks and three belts (morphotectonic units) have been identified in Jamaica (Reference 217). The blocks, from west to east, are the Hanover, Clarendon, and Blue Mountain blocks. These three blocks are separated by two northwest-trending graben structures; the Montpelier-Newmarket belt separates the Hanover and Clarendon blocks, and the Wagwater belt separates the Clarendon and Blue Mountain blocks. The North Coast belt is an east-west-trending unit that abuts the northern edge of the central Clarendon block.

The Hanover, Clarendon, and the Blue Mountain blocks consist of Early to Late Cretaceous (Albian to Maastrichtian) volcanic, volcaniclastic, and plutonic assemblages with some minor limestones. The stratigraphy of these older rocks is different for each block due to lateral variations in rock types deposited in small basins of the Cretaceous island-arc system.

The Hanover block contains only Late Cretaceous rocks exposed in four inliers: the Lucea, Jerusalem Mountain, Green Island, and Grange inliers. An inlier is an area or group of older rocks surrounded by young rocks (Reference 202). The Lucea inlier contains a 4000-meter (13,100-foot) thick sequence of shales, sandstones, and minor limestones ranging from late Santonian to early Campanian in age. An important feature in these rocks is a submarine canyon complex consisting of conglomerate channel fill that cuts across and disturbs the underlying shales and sands (Reference 217). Other structural units of the Lucea inlier contain sequences of clastic deposits and minor limestones, including channelized sands of Santonian age (Reference 217). The Green Island, Grange, and Jerusalem Mountain inliers contain lithologies similar to the Lucea inlier but are younger. Lithologies include a Late Cretaceous (Maastrichtian) arenaceous red bed sequence with rudist limestones and red fluvial sandstones and conglomerates (Reference 217).

The Clarendon block forms the central part of the island and contains Early to Late Cretaceous rocks that range in age from pre-Barremian to possible late Maastrichtian. The rocks are exposed in five main inliers (Lazaretto, Benbow, Central, St. Ann's Great River, and the Maldon and Calton Hill) and several minor ones (the Above Rocks, Sunderland, and Marchmont). Amphibolites occur in the Lazaretto inlier of the extreme southeastern part of the Clarendon block. The Benbow inlier contains the oldest sedimentary rocks in the island including over 4000 meters (13,100 feet) of volcanogenic conglomerates, sandstones, volcanic flows, and rudistid limestone. The Central inlier contains the most complete sequence of the Clarendon block and contains Upper Cretaceous igneous rocks and volcaniclastic deposits intercalated with rudistid limestone layers. The oldest rocks are volcaniclastic conglomerates, overlain by intercalated limestones and shales. Volcanic formations containing epiclastic sandstones and conglomerates interbedded with andesite flows unconformably overlie the shales. Volcanically derived siltstones overlie the volcanic formations and are interbedded with limestones. The top of the sequence consists of red volcanogenic and fluvial deposits, some containing pumice fragments in addition to ignimbrite flows. The Above Rocks inlier, the eastern part of the Central inlier, is dominated by granitoid rocks intruded into siliceous sedimentary rocks. The St. Ann's Great River, Sunderland, Calton Hill, Maldon, and Marchmont inliers are in the northern and northwestern parts of the Clarendon block and, unlike the eastern inliers, are devoid of the volcanic rocks. The St. Ann's Great River inlier contains shales, sandstones, and conglomerates of early

Coniacian to late Campanian age that are unconformably overlain by Eocene sediments. The Sunderland, Calton Hill, and Marchmont inliers contain conglomerates and shales of the Sunderland Formation (Santonian to Campanian age). Red Maastrichtian sandstones and conglomerates occur in the southern region of these inliers (Reference 217).

The Blue Mountain block, consisting of the Blue Mountain and Sunning Hill inliers, occupies the eastern third of the island and contains Campanian to Maastrichtian volcanic rock and volcanogenic sedimentary rocks with major limestone horizons (Figures 2.5.1-302 and 2.5.1-303). The Blue Mountain inlier contains a thick sequence of interbedded andesitic tuff, flows, and volcanogenic conglomerates as well as contemporaneous pelagic limestones underlain by a chert-basalt-gabbro ophiolitic complex, granitoid intrusives, and regionally metamorphosed rocks, including mafic blueschists, greenschists, and amphibolite-facies rocks (References 217 and 806).

The younger Montpelier-Newmarket zone, Wagwater belt, and North Coast belt consist of Paleocene clastic rocks, later Tertiary carbonate rocks, and late Tertiary to Quaternary carbonate rocks that are occasionally intercalated with clastic sequences. From oldest to youngest, the three belts include the Wagwater, Yellow Limestone, White Limestone, and Coastal Groups (Figures 2.5.1-302 and 2.5.1-303). Each group consists of shallow water facies (lagoonal to shelf edge) and deep-water faces. The Wagwater Group consists of a lower section of red conglomerates, a middle section of interbedded conglomerates and thinly bedded sandstones and shales, and an upper section of dacitic volcanics and minor basalts, interbedded with clastic rocks. The Wagwater Group strata range in age from Early Paleocene to lower Eocene and attain a total thickness of approximately 7000 meters (23,000 feet) (Reference 217). Gypsum occurs at several places and reaches a maximum thickness of 60 meters (200 feet). The Yellow Limestone Group is Lower to Middle Eocene in age and the White Limestone Group is Middle Eocene to Late Miocene in age. Together the Yellow and White Limestone Groups represent over 2750 meters (9000 feet) of lagoonal, shelf edge, and deep-water carbonates. The overlying Coastal Group is Pliocene to Pleistocene in age and consists of shallow water lagoonal and patch reef sediments with some subaerial to submarine fanglomerates.

According to Westcott and Etheridge (Reference 561) in Lewis and Draper (Reference 217), the clastic rocks of the Wagwater Group represent fan-delta and proximal to distal submarine-fan deposits. Erosion of the Cretaceous volcanic rocks supplied source material to several fan systems, which developed at the steep margin of the Wagwater belt basin. By late Early Eocene time, a general marine transgression submerged the entire island and led to the deposition of thick limestones. Volcanic activity concluded by the early Middle Eocene and was followed by a period of relative tectonic quiescence until the Middle Miocene. During this time, thicknesses of up to 2750 meters (9000 feet) of the Yellow and White Limestone Groups accumulated. The depositional environments of the Yellow and White Limestone Groups have been determined from a combined study of fauna and lithology to indicate that from Paleocene to middle Eocene time Jamaica experienced an island-wide marine transgression, with rapid subsidence of the North Coast, Wagwater, and Montpelier-Newmarket belts and the southern Hanover block (Reference 217). The Blue Mountain and Clarendon blocks subsided more slowly. North and east of the Wagwater fault bounding the western margin of the Wagwater belt, Coastal Group sediments consist of deep-water facies. In the Late Miocene to Pliocene, emergence occurred to subaerially expose both shallow-water and deep-water limestones. Some of this emergence was probably due to a eustatic sea-level drop, although much of it may have been due to tectonic causes (Reference 217). After the Aftonian interglacial, most sediments are neritic and the late Pleistocene geology is expressed mainly as a series of raised marine terraces (Reference 217). Alluvium is confined to interior valleys, river floodplains, and the coastal margins.

### Structures of Jamaica

The island of Jamaica occupies the northeastern tip of the Nicaraguan Rise, the eastern part of the Chortis block. Lewis and Draper (Reference 217) discuss structures seen on Jamaica in terms of the

tectonic development of the island. The oldest rocks on the island are in the Blue Mountains province and are attributed to an early to mid-Cretaceous west-dipping island arc. By the late Cretaceous, subduction had shifted southeast of the island and magmatism for the most part ceased.

During the Paleocene to middle Eocene, the presently observed northwest-southeast oriented block-and-graben structures were created. The northwest-striking faults shown in Figure 2.5.1-300 initiated during this time period. These include the Montpelier-New Market fault zone, Santa Cruz fault, and Spur Tree fault. Several hypotheses have been presented to explain the northeast-southwest extension that gives rise to these rift structures, but none are conclusive (Reference 217).

After a 30 m.y. period of quiescence and submersion, the left-lateral transcurrent regime active today was established in the late Miocene. The role of Jamaica as a restraining bend in the Caribbean-Gonâve plate boundary was initiated at this time, and west-northwest- to east-southeast-striking left-lateral faults overprinted the earlier fault pattern. Mapped faults on the island are shown in Figure 2.5.1-300. To the north, the Duanvale fault zone extends from the north-central part of the island to the west through Montego Bay and connects to the Walton fault (Figure 2.5.1-202). To the east, the Plantain Garden fault extends to the east and connects to the Enriquillo fault in and west of Haiti. In the center of the island this deformation zone is expressed as the Rio Minho-Crawle River fault zone. The South Coast fault along the southwest coast also reflects the current stress regime.

Lewis and Draper (Reference 217) point out that some of Jamaica's early Tertiary normal faults may be reactivated as thrust faults, a prime example being the Blue Mountains fault. Oblique folding and tilting of beds has accompanied the current compressional stress field, along generally north-south-trending axes. Mann and Burke (Reference 859) describe the Wagwater belt, which comprises the western part of the Blue Mountains physiographic province, as an intra-arc inverted basin structure. They suggest the belt initially formed as a basin parallel to, and along the axis of, the Greater Antilles Arc in the early Paleocene through early Eocene. With initiation of the Cayman Trough, the stress regime changed to transpressive, as the region became a restraining bend in a strike-slip system. The basin was thus uplifted (inverted) to form the present-day physiography, and normal faults originally associated with basin development are reactivated as thrust faults (Reference 849).

#### Seismicity of Jamaica

Jamaica has experienced 13 earthquakes of MMI VII and greater since the mid 1600s. The most severe was the $M_W$ 7.75 (Phase 2 earthquake catalog) 1692 Port Royal earthquake near Kingston (Figure 2.5.2-214), which submerged the town and killed a quarter of its inhabitants. A MMI IX event in 1907 (Phase 2 earthquake catalog $M_W$ 6.64) in the same region caused 1000 fatalities (Reference 563). The pattern of present-day microseismicity indicates the most intense activity is in the Blue Mountains region, the topographically highest region of the island. Focal mechanisms show a mixture of thrust and strike-slip mechanisms, consistent with transpression due to northeast-southwest compression (Reference 503). Aside from the Blue Mountains region, associations between seismicity and faults are not clear (Figure 2.5.1-267).

### 2.5.1.1.2.2.2.2 Geology of Hispaniola

#### Physiography of Hispaniola

Hispaniola is a mountainous island about 660 kilometers (410 miles) long and 260 kilometers (160 miles) wide. Haiti occupies the western part of the island and the Dominican Republic the eastern part. Hispaniola has many features in common with the islands of Cuba and Jamaica to the west and Puerto Rico to the east. Hispaniola is the second largest island of the Greater Antilles Deformed Belt, a Cretaceous-early Tertiary island arc that stretches from Cuba to Puerto Rico and the British Virgin

Islands. Hispaniola is separated from Cuba to the northwest by the Windward Passage, 4000 meters (13,100 feet) deep; from Jamaica to the west-southwest by the Jamaica Passage, 3000 meters (9800 feet) deep; to the east from Puerto Rico by the Mona Passage, 460 meters (1500 feet) deep; and to the north from the Bahama Banks by the Old Bahama Channel (coincident with the northwestern portion of the Puerto Rico trench) 4300 meters (14,000 feet) deep.

Hispaniola is located at the convergence of five physiographic-structural trends in the northern Caribbean: the main axis of the Greater Antilles Deformed Belt, the Cayman Trench, the Nicaraguan Rise, the Beata Ridge, and the Bahamas-Cuba intersection (Reference 217) (Figures 2.5.1-202 and 2.5.1-210). The significance of Hispaniola's location with respect to these trends is discussed later in this subsection.

Hispaniola has four nearly parallel west-northwest-trending mountain ranges, separated by three relatively narrow, alluvial-filled, longitudinal structural depressions (Reference 796) (Figure 2.5.1-304). At the northeastern end of the island, the Cordillera Septentrional is bounded on the southwest by the Cibao Valley. Southwest of the Cibao Valley lies the Massif du Nord-Cordillera Central and the Sierra de Seibo. These, in turn, are separated on the southwest by the Central Plateau-San Juan Valley. Southwest of the Central Plateau-San Juan Valley lies the Montagnes Noire and the Sierra de Neiba and its northwest extension, the Matheaux-Trou d'Eau. These central mountainous regions are bordered on the southwest by the Enriquillo Graben. South and east of the Enriquillo Graben lies the Sierra de Bahoruco trending west into the Massif de la Selle and continuing into the Massif de la Hotte. In general, the mountain ranges in the northern part of Hispaniola trend about N 40-50° W, oblique to the main axis of the island. This trend is parallel to the structural grain of central and eastern Cuba. However, the mountains ranges in the southwestern part of the island (the Massif de la Hotte and the Massif de la Selle of the Southern Peninsula) have an east-west trend, which is parallel to the axis of Hispaniola and the Greater Antilles as a whole (Reference 217).

Coastal plains also occur on the island of Hispaniola and are most extensive on the southeast coast of the Dominican Republic. In Haiti, where the mountains frequently stretch to the shoreline, the area of coastal plain is relatively small. Raised coral reef terraces, all of Quaternary age, are found in a number of localities along the coast, indicating that local uplift of up to several hundred meters continued at least well into the Pleistocene (Reference 796).

### Stratigraphy of Hispaniola

Hispaniola consists of an agglomeration of twelve tectonic terranes or zones as indicated on Figure 2.5.1-305 and Table 2.5.1-206, namely:

1. Samaná
2. Puerto Plata-Pedro Garicía-Río San Juan
3. Altamira
4. Oro
5. Seibo
6. Tortue-Amina-Maimon
7. Loma Caribe-Tavera
8. Duarte
9. Tireo

10. Triois Rivières-Peralta

11. Presqu'île du North-Ouest-Neiba

12. Selle-Hotte-Bahoruco

One of the terranes (Selle-Hotte-Bahoruco) is a fragment of oceanic plateau terrane that crops out over the southern one-third of the island. As described in Subsection 2.5.1.1, the buoyancy of oceanic plateau crust makes it unlikely that this crustal fragment was accreted to Hispaniola during the development of the island-arc terranes. Eleven of the terranes are fragments of island-arc terranes that crop out over the northern two-thirds of the island. The eleven island-arc terranes, which range in age from Early Cretaceous to late Eocene, can be classified on the basis of lithologic associations, geochemistry, and structure as: (a) fragments of oceanic crust on which the island arcs were built, (b) fragments of the forearc/accretionary prism of an island arc, (c) fragments of the volcano-plutonic part of an island arc, and (d) a fragment of a back-arc basin (Reference 566) (Table 2.5.1-206). All twelve tectonic terranes generally have elongated shapes and are bounded by high angle strike-slip or reverse faults (Figure 2.5.1-302). Several of the terrane boundaries are either completely or partially covered by 1- to 6-kilometer (0.6- to 3.7-mile) thick, late Miocene to Recent clastic and carbonate sedimentary basins (Reference 564). The carbonates of the Seibo terrane began to form in the Early Cretaceous (Aptian to Albian) (Reference 566) while carbonates of the Selle-Hotte-Bahoruco terrane formed in a gradually deepening marine environment in the Paleocene to Miocene. The clastic sedimentary basins formed and filled during a transpressional phase of terrane docking (References 217 and 566).

The tectonic terranes of Hispaniola can also be divided on the basis of their deformational characteristics, as indicated in Table 2.5.1-206. Three types of terranes are identified:

- Stratigraphic terrane
- Metamorphic terrane
- Disrupted terrane

Seven of the twelve island-arc terranes are stratigraphic terranes, which are characterized by coherent sequences of strata in which depositional relations between successive lithologic units can be demonstrated (Reference 566). These seven stratigraphic terranes can be further subdivided into: (a) five fragments of the volcano-plutonic part of an island (these fragments are composed dominantly of volcanic and associated sedimentary rocks and the underlying plutonic roots of the island arc), (Tireo, Seibo, Oro, Presqu'île du Nord-Ouest-Neiba, Altamira); (b) one fragment of a back-arc basin characterized by mainly deep-marine turbiditic rocks of submarine fan facies (Trois Rivières-Peralta); and (c) one fragment of an oceanic plateau characterized by thick sequences of pillow basalts and gabbros with overlying deep-sea sedimentary deposits (Selle-Hotte-Bahoruco) (Reference 566) (Table 2.5.1-206).

Three of the twelve terranes are metamorphic terranes, characterized by rocks metamorphosed to a high enough grade that original minerals, stratigraphic features, and stratigraphic relationships are obscured. The Samaná metamorphic terrane is a fragment of the forearc/accretionary prism of an island arc, the Duarte metamorphic terrane is a fragment of ocean floor including seamounts, and the Tortue-Amina-Maimon metamorphic terrane is a fragment of a volcano-plutonic part of an island arc (Reference 566) (Table 2.5.1-206).

Two of the twelve terranes, the Puerto Plata-Pedro García-Río San Juan and Loma Caribe-Tavera terranes, are disrupted terranes. Disrupted terranes are characterized by brittle deformation that obscures the depositional relations between successive lithologic units. The Puerto Plata-Pedro

García-Río San Juan terrane consists of blocks of heterogeneous lithology and age set in a matrix of serpentinite (Reference 566) (Table 2.5.1-206).

These various basement terranes of Hispaniola were left-laterally translated from points of origin to the west along faults associated with the Cayman spreading center. Mann et al. (Reference 566) note that translation of terranes along strike-slip faults can act to disperse terranes or to accrete them. In Hispaniola, Mann et al. (Reference 566) propose that the numerous terranes accreted over time because of offsets, or “restraining bends,” in the controlling faults since the Miocene. Furthermore, the effect of deformation within the restraining bend in Hispaniola was absorbed by: (a) uplift and erosion of lower crustal rocks (e.g., Duarte, Loma Caribe-Tavera, and Tortue-Amina-Maimon terranes of central Hispaniola); (b) large-scale rotation of terranes about vertical axes (e.g., post-Eocene counterclockwise rotation of Tireo terrane); (c) large-scale underthrusting of one terrane beneath another (e.g., Selle-Hotte-Bahoruco terrane beneath island-arc terranes of central Hispaniola); and (d) splaying of the strike-slip fault into several different strands at the restraining bends (e.g., the Oriente fault splays into the Bahamas Channel, Camu, and Septentrional faults in northern Hispaniola) (Reference 219). The end result is the geologic history of adjacent terranes is often quite distinct and difficult to unravel (Reference 904).

Mann et al. (Reference 566) postulate that many of the terrane boundaries separating island-arc and oceanic plateau terranes were reactivated as oblique-slip faults after active subduction ceased following the collision between Hispaniola and the Bahama Platform. Based on the complex structural relations found in the field, Mann et al. (Reference 566) conclude that Early Miocene to Recent transpression at the Hispaniola restraining bend (or convergent segment of the east-west-striking North America-Caribbean strike-slip plate boundary) produced ten morphotectonic zones that correspond to the major mountain ranges and intervening clastic sedimentary basins of Hispaniola (References 566 and 217). The boundaries of each of the zones are generally well-defined topographic escarpments or lineaments, and each zone has geologic characteristics that distinguish it from its neighboring zones. In general, morphologic boundaries between the ten morphotectonic zones correspond well to major differences in the rock types of tectonic terranes because of Neogene reactivation of major crustal faults separating tectonic terranes (Reference 566). Correlation between the ten morphotectonic zones and the twelve tectonic terranes of Hispaniola is described in Table 2.5.1-205.

### Structures of Hispaniola

Hispaniola comprises an amalgamation of several terranes, reflecting a long and complex geologic and tectonic history. Present-day structures have been imprinted on these terranes that reflect the current role of Hispaniola as a microplate between the North America plate to the north and the Caribbean Plate to the south. In addition, this microplate has been proposed to consist of two parts: the Gonâve microplate to the west of central Hispaniola, and the El Seibo microplate to the east (Figure 2.5.1-202).

Because the major tectonic structures associated with Hispaniola have been described elsewhere, brief descriptions are provided below with cross-references to more detailed descriptions elsewhere. These features are shown in Figure 2.5.2-214.

1. Septentrional fault: Described in detail in Subsection 2.5.1.1.2.3.2.1. This left-lateral strike-slip fault is the dominant plate boundary between Hispaniola and the North America Plate, and separates the Cordillera Septentrional-Samana Peninsula from the Cibao Valley (Figure 2.5.1-202). Slip rates from trenching and GPS studies range from 6 to 12 millimeters/year in the Cibao Valley region, decreasing to the east.

2. North Hispaniola subduction zone (NHSZ): Described in detail in Subsection 2.5.1.1.2.3.2.2. This south-dipping thrust fault merges with the Puerto Rico Trench (subduction zone) in the Mona Passage region to the east of the island (Figure 2.5.1-202).

3. Enriquillo-Plantain Garden fault zone: Described in detail in Subsection 2.5.1.1.2.3.2.3. This left-lateral strike-slip fault forms the boundary between Hispaniola and the Caribbean Plate in the western part of the island (Figure 2.5.1-202). The slip rate has been estimated to be about 8 millimeters/year.

4. Muertos Trough: Described in detail in Subsection 2.5.1.1.2.3.3. This north-dipping subduction zone accommodates north-south compression between Hispaniola and the Caribbean Plate (Figures 2.5.2-214 and 2.5.1-202).

5. Mona Passage extensional zone: Described in detail in Subsection 2.5.1.1.2.3.3. This is a zone of about 5 millimeter/year of east-west extension between Hispaniola and Puerto Rico (Figures 2.5.1-210 and 2.5.1-202).

**Seismicity of Hispaniola**

Descriptions of the historically significant and largest earthquakes associated with the active structural features of Hispaniola are briefly described below, along with cross-references to more detailed descriptions.

- Septentrional fault (Subsection 2.5.1.1.2.3.2.1). The 1842 $M_w$ 8.20 (Phase 2 earthquake catalog) earthquake occurred on the western part of the fault. Farther east in the Cibao Valley, paleoseismic trenching studies on the Septentrional fault indicate that the most recent surface faulting event occurred about 1200 A.D. (Reference 570). Damaging historic earthquakes also occurred in the vicinity in 1562, 1783, 1887, and 1897 (Phase 2 earthquake catalog $M_w$ 7.23, 6.13, 7.93, and 7.03, respectively). (There is some uncertainty regarding the date of the 16$^{th}$ century earthquake. The Phase 2 earthquake catalog indicates that it occurred in 1562, whereas Scherer [Reference 571] indicates 1564). Due to the proximity of the Northern Hispaniola subduction zone, however, some or all of these may have occurred on that feature.

- Northern Hispaniola subduction zone (Subsection 2.5.1.1.2.3.2.2). Large earthquakes occurred on this feature in 1946, 1948, and 1953. The largest of these occurred in 1946, approached magnitude 8 (Phase 2 earthquake catalog $M_w$ 7.90), and caused loss of life and extensive damage. As discussed in the previous paragraph, it is possible that four other large historic earthquakes occurred on this feature.

- Enriquillo-Plantain Garden fault (Subsection 2.5.1.1.2.3.2.3). Large, damaging earthquakes occurred on the Enriquillo-Plantain Garden fault in 1751, 1770, and 2010. Magnitudes of the 1751 and 1770 earthquakes are $M_w$ 6.83 and 7.53, respectively (Phase 2 earthquake catalog). The destructive January 12, 2010, $M_w$ 7.0 (Reference 572) earthquake near Port-au-Prince, Haiti occurred after completion of the Phase 2 catalog and is therefore not included. Subsection 2.5.1.1.2.3.2.3 provides additional information regarding the January 12, 2010, earthquake.

- Muertos Trough (Subsection 2.5.1.1.2.3.3). A $M_w$ 7.28 earthquake near the western end of the Muertos Trough occurred in 1751 (Figure 2.5.2-214).

- Mona Passage extensional zone (MPEZ) (Subsection 2.5.1.1.2.3.3). In 1918, a $M_w$ 7.30 earthquake located in the Mona Passage generated a tsunami and ground shaking that caused extensive damage to coastal communities of northwest Puerto Rico. Abundant low to moderate magnitude seismicity is currently occurring in the Passage (Reference 573).

#### 2.5.1.1.2.2.2.3 Geology of Puerto Rico

### Physiography of Puerto Rico

The island of Puerto Rico is the smallest and easternmost island of the Greater Antilles. In addition to the principal island, the Commonwealth of Puerto Rico includes the islands of Vieques, Culebra, Culebrita, Palomino (the Spanish Virgin Islands), Mona, Monito, and various other isolated islands. Deep ocean waters fringe Puerto Rico. The Mona Passage, which separates the island from Hispaniola to the west, is about 75 miles (120 kilometers) wide and more that 3300 feet (1000 meters) deep. The 28,000-foot (8500-meter) deep Puerto Rico Trench parallels the north coast of Puerto Rico. The Muertos Trough, more than 18,000 feet (5500 meters) deep, parallels the south coast of Puerto Rico.

Puerto Rico can be divided into three major physiographic provinces (Reference 881):

- The Upland province
- The Northern Karst province
- The Coastal Plains province

The topography of the Upland (or Interior) province reflects primarily the effects of erosion on a structurally complex sequence of many kinds of igneous and sedimentary rocks. The Northern Karst province shows the effects of limestone dissolution. The Coastal Plains province is an area predominantly of deposition (Reference 881).

The Upland province is formed by a central mountain chain commonly known as the Cordillera Central, extending across the interior of the island from east to west. This mountain chain includes the La Cordillera Central, La Sierra de Cayey, La Sierra de Luquillo, and La Sierra Bermeja. These rocks of the Upland province have been uplifted and erosionally dissected to form an asymmetric mountain range in which the southern slopes dip more steeply than the northern slopes. The mountains of the Cordillera Central in this province rise to more than 1300 meters (4300 feet) above sea level (Reference 574).

The Northern Karst province is a limestone region that reflects an advanced stage of limestone dissolution and contains extensive zones of mogotes (also known as haystack hills), sinkholes, caves, limestone cliffs, and other karst features.

The Coastal Lowlands province extends 13 to 19 kilometers (8 to 12 miles) toward the south from the northern coast and 3 to 13 kilometers (2 to 8 miles) toward the north from the southern coast of Puerto Rico. A series of smaller Coastal Lowlands valleys lie perpendicular to the western and eastern coasts of Puerto Rico. This area was originally formed by the erosion of the interior mountains.

### Stratigraphy of Puerto Rico

Puerto Rico can be divided into three east-west-trending terranes: a Central Igneous Zone of island arc volcanic strata flanked to the north and south by younger carbonate strata covered by alluvium (the Northern Carbonate and Southern Carbonate Zones) (Figure 2.5.1-307). The most prominent feature of the three geologic terranes in the Puerto Rico-Virgin Islands area is a large, east-west-trending arch. The generally undeformed carbonate strata in the northern limb of the arch exhibit a smoother, more uniform dip than the folded and faulted carbonate strata exposed in the steeper, southern limb. The volcanic island arc basement rocks on Puerto Rico are exposed in the core of the arch (Reference 809).

The three terranes are discussed in the following paragraphs.

Central Igneous Zone of Island Arc Strata

The Central Igneous Zone of island arc volcanic strata in Puerto Rico, with rocks ranging in age from Early Cretaceous (Aptian) to Eocene and dating from about 120 to 45 Ma, represents one of the longest oceanic arc sequences preserved in the world (Reference 809). The strata rest on an unusually thick crust, reaching a maximum in the northeast of about 30 kilometers (Reference 809). Donnelly et al. (Reference 568) suggest that much of this represents underplating by arc-related plutonic bodies, rather than accumulations of material produced through volcanism.

Detailed, systematic geologic mapping indicates that post-volcanic sedimentary platform deposits consisting of limestone and detrital materials ring the island and cover extensive parts of the arc platform. However, representative strata of the entire volcanic arc sequence are exposed. Based on over 50 years of detailed stratigraphic study by the USGS and others, various researchers (References 217, 809, and 880) recognize three distinct igneous provinces across the island. The provinces are (a) a northeast igneous (or volcanic) province, separated from the central province by the Cerro Mula fault (shown as the North fault zone in Figure 2.5.1-307); (b) a central igneous (or volcanic) province, dominated by volcanic debris accumulated during sequential development of five east-west-oriented volcanic belts; and (c) a southwestern igneous (or volcanic) province, with a northwest-southeast-trending boundary of uncertain origin, containing remnants of two sequential island arc volcanic belts of Late Cretaceous (Campanian-Maastrichtian) and Eocene age.

In addition to the island arc volcanics, the oceanic Sierra Bermeja Complex crops out in the southwestern corner of the island. This complex consists of a tectonic mélange of partly serpentinized ultramafic rocks representing the lithospheric upper mantle originally composed of spinel-bearing peridotites (References 809 and 836). The mélange incorporates rafts, blocks, and boulder-sized clasts of Early Jurassic to Late Cretaceous pelagic sediments (Mariquita Chert), including radiolarian chert of Pacific provenance (References 804 and 820). The mélange also contains later volcanogenic strata (Cajul Basalt) and amphibolites (Las Palmas amphibolite mélange) of probable Early Jurassic age, representing pre-island arc oceanic crust (Reference 880).

Volcanic strata preserved in the central igneous province are subdivided into five major volcanic phases on the basis of stratigraphic and geochemical relations. Basalts evolved progressively from early primitive island arc tholeiites, to calc-alkaline basalts, and finally to incompatible element-enriched shoshonite basalts. Following a hiatus, calc-alkaline volcanism resumed. Correlative strata in the northeast province display a more restricted compositional range from early island arc tholeiites to calc-alkaline basalts. Volcanic strata of dominantly calc-alkaline affinities in the last phases of volcanism in both the northeastern and central provinces are chemically identical. Jolly et al. (Reference 809) infer that, by the mid-Late Cretaceous (mid-Santonian), the northeast and central blocks were tectonically juxtaposed by strike-slip movement along the Cerro Mula fault, the principal strand of the North Puerto Rico fault zone. The last phases of volcanism are represented in the west by a sequential pair of subparallel island arc belts of Campanian-Maastrichtian and late Paleocene–Eocene age, accompanied by extensive flanking sedimentary basins. A hiatus between the last two phases of volcanism, representing both a period of erosion and a nonvolcanic interval, persisted across the entire island from uppermost Late Cretaceous (Maastrichtian) through the Early Paleocene (Reference 809).

In the southwestern igneous province, island arc strata date from about 85 Ma (mid-Late Cretaceous), and the basement is inferred to have been transported into the active volcanic zone of the island arc simultaneously with left-lateral displacement along the Cerro Mula fault (Reference 809).

Jolly et al. (Reference 809) indicate that no consensus has developed regarding the polarity and tectonic history of subduction during generation of the Greater Antilles Arc. As discussed in

Subsection 2.5.1.1.2.2.2, Mattson (Reference 804) and Pindell and Barrett (Reference 219) propose that Cretaceous igneous rocks of the Greater Antilles islands of Cuba, the Cayman Ridge, Hispaniola, and Puerto Rico originated in an intra-oceanic island arc, with northeast-dipping subduction, bounding one edge of a proto-Caribbean Sea (Figure 2.5.1-347, part B). Structural fabric data from central Hispaniola suggest a reversal from east- to west-dipping subduction occurred early in arc history during the mid-Cretaceous (Aptian to Albian time). It has been suggested that initial subduction was from the west until arrival of a buoyant oceanic basalt plateau, the Caribbean Plateau, which developed in the Pacific basin at about 88 Ma, forcing a reversal in polarity of subduction between 105 and 55 Ma (Reference 833). Alternatively, the early arc might have formed along the margin of the Caribbean Plateau and advanced eastward locked with the plateau accompanied by west-dipping subduction of the proto-Caribbean (Atlantic) plate throughout arc history (Reference 809).

Northern Zone Carbonate Zone

In northern and western Puerto Rico, the siliciclastic San Sebastian Formation forms the base for the lower Oligocene to lower Pliocene Puerto Rico-Virgin Islands carbonate platform. Based on single-channel and multichannel seismic reflection lines, van Gestel et al. (Reference 670) evaluated the regional stratigraphy and structure of the platform. The platform covers an area of 18,000 kilometers$^2$ and extends from the eastern Dominican Republic on the island of Hispaniola, west of Puerto Rico, to the Virgin Islands, east of Puerto Rico. The continuity and similarity of facies across the platform indicate a remarkable stability over this area for a period of almost 35 million years (Reference 670). Where onshore platform rocks have been studied in detail in northern Puerto Rico (Reference 881) and southern Puerto Rico (Reference 882), they indicate deposition at sea level with minor periods of subsidence in the early Pliocene (Reference 670).

The carbonate strata in northern Puerto Rico include an uninterrupted sequence of generally undeformed Late Oligocene to Middle Miocene Lares, Cibao, Los Puertos, and Aymamon formations. The carbonate lithologies reflect a general shallowing of sea level, punctuated by two rapid episodes of deepening near the Oligocene/Miocene boundary and two more in the Middle Miocene. An erosional unconformity breaks the sequence from the Middle to Late Miocene, after which carbonate deposition continued from Late Miocene to Late Paleocene with the Quebradillas Formation (Reference 881). Outcrop data for the northern Puerto Rico carbonates show a 4° dipping package of homogeneous carbonate layers unconformably overlying Cretaceous-Eocene arc basement rocks (Reference 670). Alluvium is sparsely intercalated with the shallow carbonate strata and overlies the carbonate strata exposed at the surface on the northeastern side of the island.

Two amphitheater-shaped escarpments, A and B, are seen in the GLORIA images on the lower slope north of Puerto Rico. Amphitheater A is about 60 kilometers (37 miles) across and up to 2250 meters (7380 feet) deeper than the surrounding seafloor; amphitheater B is smaller and is about 30 kilometers (19 miles) across and 1500 meters (5000 feet) deep. Based on seismic reflection profiles, an estimated 1500 kilometers$^3$ of sedimentary section have been removed from the larger amphitheater, and a system of canyons has formed. The interior of amphitheater B appears to have an irregular, high backscatter surface with no canyons; more recent slumping may have occurred in the smaller amphitheater. The implication is that modification of the amphitheaters may be an ongoing, presently active process (Reference 576).

Southern Carbonate Zone

The carbonate strata in southern Puerto Rico are about the same age as the carbonate units in northern Puerto Rico. The Ponce and the Juana Diaz formations, Early Oligocene to Early Miocene and Middle to Late Miocene, respectively, have been mapped in southern Puerto Rico (References 809 and 881). The carbonate section of southern Puerto Rico has a maximum thickness of 500 meters (1600 feet) (Reference 670). The strata generally dip more steeply (10°) southward toward the Muertos Trough and show more faulting and folding both in outcrops (Reference 882) and

in seismic reflection profiles than the carbonate strata on the north coast (Reference 670). Alluvium is sparsely intercalated with the shallow carbonate strata and overlies the carbonate strata exposed at the surface on the southeastern side of the island.

### Structures of Puerto Rico

As described in the preceding section, Puerto Rico can be divided into three terranes: the Central Igneous zone and the Northern and Southern Carbonate zones (Figure 2.5.1-307). Whereas the Northern Carbonate zone is essentially undeformed, the Southern Carbonate zone contains steeply dipping faults. Folding is prominent only in the southwest Igneous province (Reference 217). The Central Igneous zone (Figure 2.5.1-307) is separated by major northwest-striking faults as seen in Figure 2.5.1-307. The island is also traversed by the North fault zone to the north, and the South fault zone to the south (also referred to as the Great North Puerto Rico fault zone and Great South Puerto Rico fault zone). Both North and South fault zones exhibit unknown but large amounts of left-lateral slip. Based on unfaulted Oligocene to Miocene age strata that overlie these structures, activity on these features apparently occurred before the Middle Miocene (Reference 217), and they are not considered to be seismogenic (Reference 577).

Trenching studies on the South Lajas fault, a 30-kilometer (19-mile) long east-west-striking fault in southwestern Puerto Rico, revealed two surface faulting events in the past 7000 years (Reference 578). This is the only Holocene fault currently documented on the island.

### Seismicity of Puerto Rico

Local seismograph networks have been operated in Puerto Rico since the mid-1970s. Early results (Reference 587) show shallow seismicity beneath the island, and a south-dipping plane of seismicity associated with the subducting North America Plate extending to depths of about 150 kilometers (93 miles) (Figure 2.5.1-309). Later studies confirm this pattern (e.g., Reference 588). Crustal seismicity on the island of Puerto Rico is sparse, consisting of low- to moderate-magnitude (magnitude ≤5) activity (References 573 and 577). Seismicity appears to be more dense in the southwestern part of the island, where Huerfano et al. (Reference 573) interpret a pattern of northwest-southeast transtension. Relocations of seismicity in this area suggest that most of this seismic activity is associated with the Muertos subduction zone (Reference 569). An earthquake of approximately magnitude 6 was felt in southwest Puerto Rico in 1670 (Reference 579). This event may have occurred on one of the MPEZ faults (Subsection 2.5.1.1.2.2.4) to the west, on the Muertos subduction zone, or an unidentified fault in western Puerto Rico.

The island of Puerto Rico is surrounded and underlain by seismogenic features that have caused damaging earthquakes in historical times. Because these are described elsewhere, only cross-references are provided below:

- Puerto Rico Trench (Subsection 2.5.1.1.2.2.3). In 1787, a $M_w$ 8.03 (Phase 2 earthquake catalog) earthquake occurred in the vicinity of the Puerto Rico Trench.

- Muertos Trough (Subsection 2.5.1.1.2.2.4). A $M_w$ 7.28 earthquake near the western end of the Muertos Trough in 1751 (Figure 2.5.2-214).

- Mona Passage (Subsection 2.5.1.1.2.2.4). In 1918, a $M_w$ 7.30 earthquake located in the Mona Passage (Phase 2 earthquake catalog) generated a tsunami and ground shaking that caused extensive damage to coastal communities of northwest Puerto Rico. Abundant low to moderate magnitude seismicity is currently seen in the Passage (Reference 573).

#### 2.5.1.1.2.2.3 Geology of the Puerto Rico Trench

### Physiography of the Puerto Rico Trench

The Puerto Rico Trench is the surface manifestation of the Puerto Rico subduction zone (PRSZ). The trench itself is an unusual feature, being the deepest point in the Atlantic Ocean (>8 kilometers or >5 miles deep) and exhibiting the lowest free-air gravity anomaly on earth (Reference 581). It lies about 120 kilometers (75 miles) north of Puerto Rico and is about 1750 kilometers (5700 feet) long and 100 kilometers (62 miles) wide. It is located where the North America Plate is subducting under the Caribbean Plate. The subduction is highly oblique (10 to 20°) to the trench axis with a large component of left-lateral strike-slip motion. The trench is also characterized by a large negative free-air gravity anomaly, -380 mGal, which indicates the presence of an active downward force (Reference 581). This gravity anomaly is located 50 kilometers (31 miles) south of the trench with a water depth of 7950 meters (26,000 feet). A carbonate platform that is tilted strongly to the north provides evidence for extreme vertical tectonism in the region. Starting in the Late Oligocene, the platform strata were deposited as a thick, flat-lying sequence on top of Cretaceous to Paleocene arc rocks. At 3.5 Ma, the carbonate platform was tilted by 4° toward the trench over a period of less than 40,000 years (Reference 582). The northern edge of the carbonate platform is at a depth of 4000 meters (13,000 feet), and its reconstructed elevation on land in Puerto Rico is at +1300 meters (4300 feet) (Reference 582).

The physiographic and structural features of the trench were imaged in 2002 to 2003 using the SeaBeam 2112 multibeam system by the National Oceanic and Atmospheric Administration (NOAA). Backscatter mosaic images derived from the multibeam bathymetry data aided in interpretation (Reference 582). The bathymetric data obtained by NOAA illustrate in great detail the northern edge of tilted carbonate platform and southern edge on land. The images also show thrust faults, normal faults, strike-slip faults, the head scarp of slope failures, debris toes, fissures in the seafloor, a pull-apart basin, and the location of a probable extinct mud volcano. In addition, photographic images of the sea floor, obtained by the USGS, show that slabs of limestone (70 kilometers or 43 miles wide) have broken off and slid into the trench (Reference 582) (Subsection 2.5.1.1.5 contains a discussion of submarine landslides associated with the Puerto Rico Trench).

### Stratigraphy of the Puerto Rico Trench

The Puerto Rico Trench can be divided into a western and eastern part at about 65 to 66° W. The western part includes the deepest part of the trench and is associated with the most oblique convergence. This part is 10- to 15-kilometers (6 to 10 miles) wide and 8300- to 8340-meters (27,200- to 27,400-feet) deep relative to mean sea level. The trench floor is flat and covered by pelagic sediments. Seismic profiles show the western part of the trench to be underlain by rotated blocks of the North America Plate that indicate trench subsidence (Reference 582). The trench floor narrows to the west and abruptly shallows to 4700 meters (15,400 feet) as it turns into the Hispaniola Trench, where convergence is more perpendicular. The eastern part of the trench is shallower by 700 meters (2300 feet) and more rugged than the deep western part. In the eastern section, the subducting North America Plate is observed in seismic lines to be broken into blocks that are not rotated (Reference 582).

The basin plain in the floor of the Puerto Rico Trench provides an example of a turbidite deposit resulting from a gravity flow event of regional derivation (Reference 583). The largest correlatable coarse layer within piston coring range on this basin plain extends for at least 300 kilometers (190 miles) with maximum thicknesses of close to 200 centimeters (6.6 feet). Although small by Hatteras or Sohm Abyssal Plain standards (see discussion of megasedimentary events in Subsection 2.5.1.1), this turbidite represents a sizeable volume of material to be derived from relatively small source areas. The volume of this flow, called the Giant Turbidite, was first estimated by Connolly and Ewing (Reference 584) to be 30 kilometers$^3$ (7.2 miles$^3$), but much more detailed coring reveals a more likely volume of 2 kilometers$^3$ (0.5 miles$^3$) (Reference 583). Apparently a turbidity current was

produced by a very large seismic event, perhaps affecting the islands of Puerto Rico and Hispaniola and the Virgin Platform simultaneously. Material flowing into the western end of the trench most likely was derived from Hispaniola, and material at the eastern end of the trench came from the slope of the Virgin Platform. However, the bulk of the sediment of the Giant Turbidite almost certainly was derived from the insular margin of Puerto Rico (Reference 315).

### Structures of the Puerto Rico Trench

The island of Puerto Rico is located within an approximately 250-kilometer wide deformation zone associated with the northern Caribbean-North America plate boundary. Deformation within this zone largely is controlled by left-lateral strike-slip faulting (Reference 858). A 535-kilometer (330-mile) long fault is located 10 to 15 kilometers (6 to 10 miles) south of the Puerto Rico trench and passes through rounded hills that form the accretionary prism. The fault is interpreted to accommodate left-lateral motion because it is apparently associated with a left-stepping pull-apart depression. Seismic reflection data show that the steeply dipping fault penetrates 5 kilometers (3 miles) through the accretionary sediments before terminating in the subduction interface. Part of this fault trace was first identified as a weak lineament on a GLORIA backscatter image and was named the Northern Puerto Rico Slope fault zone (NPRSFZ) (or “Bunce fault”) (Reference 670) (Figure 2.5.1-308). The NPRSFZ ends at the western end of the Puerto Rico Trench in several splays and appears to be the only active strike-slip fault. Its proximity to the trench suggests that slip along the subduction interface is oblique. Another fault closer to Puerto Rico, the South Puerto Rico Slope fault zone (SPRSFZ), has no clear bathymetric expression (Reference 582).

The NPRSFZ is deflected southward at 65° W, perhaps due to stress by the oblique subduction of a localized topographic ridge on the North America Plate known as the Main Ridge (Figure 2.5.1-308). Ten Brink et al. (Reference 582) suggest that the Main Ridge is underlain by a subducted ridge of seamounts because its axis is perpendicular to the observed abyssal-hill grain of the subducting North America Plate. Ten Brink et al. (Reference 582) also suggest that the resistance to subduction of the buoyant Main Ridge has resulted in the formation of local tectonic structures, including thrust and strike-slip faults and a reentrant in the trench axis.

A fault trace at the western edge of the Puerto Rico Trench is interpreted by ten Brink et al. (Reference 582) to be the eastern end of the Septentrional fault (Figure 2.5.1-308). The fault ends abruptly in a 1000-meter- (3280-feet-) deep circular depression 25 kilometers (15 miles) west of the Mona Rift. The Mona Rift consists of three en echelon depressions with depths that range from 5000 to 8150 meters (16,400, to 26,700 feet), which cut the carbonate platform and extend almost to the NPRSFZ (labeled “Bunce fault” in Figure 2.5.1-308) (References 582 and 585). The rift accommodates east-west extension between Hispaniola and Puerto Rico (Reference 585). A large slump failure along the western wall of the upper rift basin may be related to the 1918 earthquake and tsunami (Reference 319).

New multibeam bathymetry of the entire Puerto Rico trench reveals numerous retrograde slope failures at various scales at the edge of the carbonate platform north of Puerto Rico and the Virgin Islands (Reference 887). This, together with the fact that the edge of the carbonate platform is steeper than most continental slopes, indicates a higher potential for run-up, possibly as much as 20 meters (66 feet), than along many other U.S. coasts (Reference 887). The tilted carbonate platform of Puerto Rico provides evidence for extreme vertical tectonism in the region. The carbonates were horizontally deposited over Cretaceous to Paleocene arc rocks starting in the Late Oligocene. Then, at 3.5 million years before present, the carbonate platform was tilted by 4° toward the trench over a time period of less than 40,000 years (Reference 582), such that its northern edge is at a depth of 4000 meters (13,100 feet) and its reconstructed elevation on land in Puerto Rico is at +1300 meters (+4300 feet) (Reference 887). The precariously perched carbonate platform contributes slumped material made of carbonate blocks that fail, at least in initial stages, as a coherent rock mass.

Two semicircular escarpments, 30 to 50 kilometers (20 to 30 miles) across are mapped along the northern edge of the carbonate platform at a distance of 35 to 50 kilometers north of Puerto Rico. The bathymetry and side-scan images indicated that the semicircular escarpments were shaped by continuous retrograde slumping of smaller segments. Fissures near the edge of the carbonate platform indicate that the slumping process is ongoing (Reference 582).

### Seismicity of the Puerto Rico Trench

LaForge and McCann (Reference 577) model the PRSZ as two segments: (a) a shallowly dipping segment that ranges in depth from 10 to 40 kilometers (6 to 25 miles) and (b) a steeper portion extending to 130 kilometers (80 miles) depth (Figure 2.5.1-310). LaForge and McCann (Reference 577) also distinguish between an eastern and western Puerto Rico subduction zone based on the location of the impingement of the Bahama Bank on the trench at about 66.8° W longitude (Figure 2.5.1-310). This is due to denser seismicity in the western part, likely related to resistance of the buoyant Bahama Bank to subduction and therefore tighter seismic coupling in this area. Similarly, Mueller et al. (Reference 589) divide the Puerto Rico subduction zone into eastern and western portions, and model magnitude 7.9 earthquakes in the eastern section with return periods of 190 years, and magnitude 8.0 events in the western section with return periods of 200 years. LaForge and McCann (Reference 577) use both maximum moment and exponential models in the two zones, allowing for a wider range of magnitudes. However, the rates of the largest magnitude events are similar to Mueller et al. (Reference 589), on the order of several hundred years.

Seismicity of magnitude <7 is abundant in the Puerto Rico subduction zone, but only two events have exceeded magnitude 7 in the 500-year historical record. McCann (Reference 600) suggests that a magnitude 8 to 8.25 interface earthquake occurred on this segment in 1787 (Phase 2 catalog $M_W$ 8.03), rupturing from roughly Mona Canyon on the west to the Main Ridge on the east. This earthquake caused widespread damage on the island. In 1943, a magnitude 7.8 earthquake (Phase 2 catalog $M_W$ 7.60) ruptured an approximately 80-kilometer (50-mile) wide section of the subduction zone across Mona Canyon, and on the basis of a focal mechanism, it was judged to have occurred on the shallow interface (Reference 591).

#### 2.5.1.1.2.2.4 Geology of the Muertos Trough/Mona Passage

### Physiography of the Muertos Trough/Mona Passage

The Muertos Trough is an east-west-trending depression, which is slightly concave to the north. The trough is 650 kilometers (400 miles) long and runs from the Beata Ridge in the west to the insular slope of the Aves Ridge in the east (Figures 2.5.1-202 and 2.5.1-311). The water depths of the trough are greater than -5550 meters (-18,000 feet). The Muertos Trough consists of elongated, narrow, sub-parallel ridges with the seaward slope steeper than the landward slope. The accreted pelagic sediments are from the foreland region and the turbiditic sediments are mostly derived from Hispaniola and Puerto Rico. The turbidity currents may form the deep canyons whereas the rivers carry the suspended material from onshore areas to the Muertos Margin (Reference 592).

### Stratigraphy of the Muertos Trough/Mona Passage

The A” and B” seismic reflector horizons of the Venezuelan Basin gently dip to the north beneath the turbidite fill of the Muertos Trough and continue beneath the insular slope of the Muertos Trough. The insular slope that runs parallel to the trough is formed by an east-west deformed belt (Reference 592).

The axial slope of the Muertos Trough becomes deeper from east to west with a maximum depth of about -5580 meters. The trough is marked by a smooth seafloor with approximately 0° slope. The trough is characterized by a series of wedges of smooth, closely spaced, and subparallel reflectors with high seismic reflectivity. Core samples taken in 2005 indicate that the trough seafloor comprise different sources of sediments, including interbedded turbiditic and pelagic sediments, that are

underlain by homogeneous carbonate pelagic mudstones and siltstones from the Venezuelan Basin. The turbidite wedge is separated from the Venezuelan Basin layers by a basal unconformity (Reference 592).

The toe of the insular slope, which runs parallel to the east-west deformed belt, defines the northern boundary of the Muertos Trough. In the northern boundary between the Venezuelan Basin and the toe of the insular slope, there is high lateral variability in morphological features such as turbidite trough fill, onlap, detachment, deformation front, basal unconformity, anticlinal ridge, and incipient slope basin that appear located forward of the main deformation front. In the eastern part of the northern boundary there is no distinct morphological trough, which might possibly be due to higher sediment supply. In this eastern part of this boundary, the turbidite wedge is wider and thicker and continues southwards. Fault escarpments that separate the flat seafloor of the trough wedge and the northward slope of the Venezuelan Basin are located where the insular slope meets the Venezuelan Basin at the southern margin of the Muertos Trough (Reference 592).

The western segment is an elongated flat area that is the deepest part of the Muertos Trough. It is confined to the south by escarpments that are subparallel to the deformation front, which forms a structural ponded basin. The width of the confined trough is variable; however, it becomes narrower and shallower eastward. The eastern segment is a smooth, gentle bathymetric undulation without a distinct morphological trough (as seen in the western segment). The eastern trough segment does not consist of normal faults in the outer wall of the trough (Reference 592).

The escarpments in the western segment of the Muertos Trough are the result of normal faults that affect the sedimentary cover of the A” and B” reflector horizons. From seismic reflection profiles, at least 20 of these normal fault scarps are observable in the Venezuelan Basin near the trough and beneath the turbidite wedge in the western segment. In the eastern segment of the Muertos Trough, the single anticline located in the main deformation front is forming a small ridge that is sub-parallel to the front. This anticline has an elongated shape due to the activity of a propagating blind thrust that is folding the trough fill material. This thrust is the result of the propagation of the detachment surface toward the turbidite wedge; however, horizontal turbidite layers bury the thrust, which is interpreted to suggest a low rate of recent activity (Reference 592).

Granja-Bruña et al. (Reference 592) divide the Muertos Trough into three east-west-trending provinces: the lower slope, the middle slope, and the upper slope. The lower slope is at the base of the insular slope from the toe of the deformation front to the convex slope break. The middle slope is from the convex slope break to the concave slope break; however, in many places the concave slope break is not well defined in the bathymetry data due to a higher sedimentation rate and lower deformation rate (thrusting activity). When the sedimentation rate is faster than the thrusting activity, slope basins are completely filled, which then forms the terraces. This is seen as a smooth bathymetric profile with shallower horizontal and smooth downward-dipping sedimentary reflectors. Another characteristic of the middle slope is the imbricate structure, which is similar to the lower slope, but extended slope deposits bury the structure. The upper slope is located between the concave slope break and the edge of the carbonate platform (at the top of the island arc consisting of Hispaniola and Puerto Rico). It is characterized by talus, the steeply sloping area between the carbonate platform and the terrace deposits located at the base of the steep slope (the material is derived from the carbonate platform), and by the presence of terraces with gentle seaward slopes. Important sedimentary processes such as mass movement, gravity flows, slumping, and sliding define the talus area. The mass movement and gravity flows show a smoother bathymetry. The slumped areas are sometimes aligned with the escarpment and ridges (Reference 592).

### Structures of the Muertos Trough/Mona Passage

The Muertos Trough forms the boundary between the Caribbean Plate to the south, the Hispaniola microplate to the northwest, and the Puerto Rico-Virgin Islands (PRVC) microplate to the northeast.

These two microplates are separated by the MPEZ, a region of east-west extension (Figure 2.5.1-202).

South of the island of Puerto Rico, the North Caribbean deformed belt comprises two primary features: the Muertos thrust belt (labeled “LMDB” on Figure 2.5.1-327) and the Anegada passage (Figures 2.5.1-210, 2.5.1-308, and 2.5.1-328). The Muertos Trough is the ocean floor manifestation of the Muertos subduction zone. It is about 5 kilometers (about 3 miles) deep near central Hispaniola, becoming shallower toward the east, reaching a depth of 4 kilometers (2.5 miles) at the longitude of eastern Puerto Rico, where the bathymetric feature disappears. The Muertos thrust belt (Figure 2.5.1-311) appears to be an accretionary wedge structure (Reference 593). Based on GPS measurements, Jansma et al. (Reference 594) calculate an average compressive relative motion of 2.4 millimeters/year between southwestern Puerto Rico and stable Caribbean Plate, in a west-southwest direction. This plate boundary thus accommodates largely left-lateral relative motion, with a north-south compressive component (Figure 2.5.1-311).

### Seismicity of the Muertos Trough/Mona Passage

Seismicity in the vicinity of the Muertos Trough appears to be more dense to the west (Reference 595). A great earthquake of estimated magnitude 8.0 (Phase 2 earthquake catalog $M_w$ 7.28) occurred on the Muertos thrust belt in 1751 (Reference 596) (Figure 2.5.2-214). A $M_S$ (surface-wave magnitude) 6.7 (Phase 2 earthquake catalog $M_w$ 6.70) earthquake that occurred in 1984 in the western Muertos thrust belt displays a thrust mechanism consistent with the direction of relative plate motion and location of the subducting Caribbean Plate (Reference 595). However, no moderate to large earthquakes recorded in the historical record appear on the Muertos subduction zone east of Hispaniola. A possible exception is an event of approximately magnitude 6 that was felt in southwest Puerto Rico in 1670 (Reference 579). However, an origin of this earthquake on one of the MPEZ faults or an unidentified fault in western Puerto Rico is equally likely.

LaForge and McCann (Reference 577) assigned slip rates to the Muertos thrust belt of 1.2 and 0.6 millimeters/year to sections west and east of 67° W, respectively, based on constraints from GPS measurements and historic seismicity. This corresponds to return periods of a few thousand years for earthquakes in the $M_w$ 7.8 to 8.2 range, and should be considered conservative values. Mueller et al. (Reference 589) does not consider the Muertos thrust belt to be an active feature, citing lack of positive evidence. However, McCann (Reference 597) performed a joint hypocenter-velocity model inversion using local earthquakes and identifies well-defined active seismicity on the Muertos thrust belt beneath Puerto Rico. The sense of motion on the Muertos thrust belt beneath Puerto Rico is a subject of controversy. Despite the highly oblique relative plate motion, no strike-slip faults are seen on land or in the accretionary prism, which are typical of such plate boundary environments (e.g., the Septentrional fault of northern Hispaniola). On this basis ten Brink et al. (Reference 593) suggest that all motion on the Muertos thrust belt is due to compressive stresses transmitted from the Puerto Rico thrust belt to the north. Until focal mechanisms from well-located earthquakes on the Muertos thrust belt are available, this question will remain unanswered. In summary, while the Muertos subduction zone appears to be an active feature beneath Puerto Rico, its seismic potential remains enigmatic.

At about 65° W, bathymetric expression of the Muertos Trough disappears, and the North Caribbean deformed belt is expressed as the Anegada Passage (Figures 2.5.1-210, 2.5.1-308, and 2.5.1-328). The Anegada Passage is underlain by a late Neogene complex of extensional basins and intervening ridges in the northeastern Caribbean. It cuts the older Antillean Arc Platform, from the Puerto Rico-Lesser Antilles Trench in the northeast to the Muertos Trough in the southwest. It is an east-northeast-striking extensional zone approximately 50 kilometers (31 miles) wide, which separates the Puerto Rico-Virgin Islands microplate from the Caribbean Plate to the south (Reference 598). Several deep basins, including the Virgin Islands and Whiting Basin (Figures 2.5.1-312 and 2.5.1-313), were formed between 11 and 4.5 Ma as the Puerto Rico-Virgin Islands microplate underwent approximately 20° of counterclockwise rotation. This rotation is postulated to be due to the impingement of the Bahama Bank on the northwest corner of Puerto Rico

(e.g., Reference 599). Although no rotation has been noted during the last few million years, active deformation in the Anegada Passage basins is indicated by abundant seismicity, including a tsunamigenic event with an estimated magnitude of 7.3 (Phase 2 earthquake catalog $M_w$ 7.50) that occurred in 1867 (Reference 600).

The Investigator fault (Figure 2.5.1-312) cuts the slope between the Puerto Rico Island Platform and the Muertos Trough, and exhibits north-south extension that increases from west to east. Based on orientation and bathymetric expression, LaForge and McCann (Reference 577) divide the Investigator fault into west and east segments (Figure 2.5.1-312). They estimate slip rates of 0.8 and 1.5 millimeters/year on the west and east segments, respectively.

#### 2.5.1.1.2.2.5 Geology of the Nicaraguan Rise

The Nicaraguan Rise (or Plateau) is a major submarine crustal feature that extends northeast across the Caribbean Sea from the coast of Honduras and Nicaragua to northeast of Jamaica, where it intersects the southwestern part of the Southern Peninsula of Haiti. The Nicaraguan Rise covers an area of some 413,000 kilometers$^2$ (159,500 miles$^2$) (Figure 2.5.1-210). Little is known about its structure and lithological composition, and it is probably the least understood major crustal feature in the Caribbean (References 810 and 811).

The broad shelf area of the Nicaraguan Rise to the northeast of the land areas of Honduras and Nicaragua and extending to Jamaica (the upper Nicaraguan Rise of Reference 526) is here termed the northern Nicaragua Rise (Figure 2.5.1-210). The southern boundary of the northern Nicaraguan Rise is the Pedro fault (or Fracture) zone. The southern Nicaraguan Rise extends from the Pedro fault (or Fracture) zone to the Hess Escarpment (Figure 2.5.1-319).

Morphologically the northern Nicaraguan Rise is characterized by a series of carbonate banks and shelves separated by channels and basins that have evolved from a continuous carbonate “megabank” established over basement highs (References 300, 864, and 601). In essence, the submarine shelf area of the northern Nicaraguan Rise is a topographic extension of the Precambrian-Paleozoic continental Chortis block (References 850, 851, 852, and 319). For this reason Meyerhof (Reference 853) and others maintained that a considerable part of the Nicaraguan Rise must be underlain by Pre-Mesozoic continental crust. New information, however, indicates that the Chortis block is not compositionally homogeneous (Reference 812) and that most of the basement rock of the northern Nicaraguan Rise is not of continental composition but consists of island arc crust and is likely to be of similar composition to the island of Jamaica near the northern end of the rise (References 601, 528, 811, and 217). With the exception of the northern Honduran borderlands, no rocks older than Cretaceous in age are known on Jamaica or have been reported from any part of the Nicaraguan Rise (Reference 812).

The Nicaraguan Rise represents a broad carbonate platform that formed over an calc-alkaline island arc basement to the north and over block-faulted oceanic plateau crust to the south (Reference 219). The San Pedro fracture zone (Figure 2.5.1-319) represents the boundary between these two basement types, separating the northern Nicaraguan Rise and the southern Nicaraguan Rise. The area underlain by calc-alkaline island arc includes Jamaica, the shallow banks, and the intervening deeps of the northern Nicaraguan Rise, and is bounded on the north by the Cayman Trough. The southern Nicaraguan Rise is separated from the Colombian Basin to the southeast by the Hess Escarpment (Reference 219) (Figure 2.5.1-319).

The carbonate platform of the northern and southern Nicaraguan Rise was drowned by the Miocene carbonate crash (see discussion of carbonate platforms: growth, shut downs, and crashes in Subsection 2.5.1.1.1.1.1.2). Typical sediments found in the middle/upper Miocene carbonate crash interval (9.6 to 13.5 Ma) are micritic nannofossil chalk and clayey nannofossil chalk. The drowning interval is equivalent to the lithologic Unit I found at Site 1000 of ODP Leg 165 (Reference 299). The

Unit I carbonate platform sediments display the high sedimentation and accumulation rates averaging 47.0 meters/m.y. (4.5 to 7.5 g/cm$^2$/k.y.), the mass accumulation rates were calculated from the sedimentation rates. The mass accumulation rates results for the noncarbonated portion increased steadily from the bottom of the section to a peak at approximately 380 meters below sea floor (mbsf) (approximately 11 Ma) and then declined upsection to a low at approximately 230 mbsf (approximately 6.5 Ma). Carbonate mass accumulation rates basically parallel the noncarbonated mass accumulation rates record except at approximately 450 and 380 mbsf (12.8 to 10.8 Ma) where the mass accumulation rates converge, a sharp peak centered at approximately 180 mbsf in the uppermost Miocene section (approximately 5.5 Ma) and a broader peak in the Pliocene centered at 100 mbsf (approximately 3.5 Ma). The bulk mass accumulation rate record is dominated by the carbonate component and mostly follows the trends of carbonate mass accumulation rates (Reference 299).

The high-carbonate mass accumulation rates at Site 1000 generally indicate the proximity to a periplatform environment, where pelagic settling is mixed with other fine sediments derived from the surrounding banks. The highest carbonate mass accumulation rates are out of sync with turbidite occurrence, which can be interpreted to reflect increased pelagic input during the lower middle Miocene; it is also consistent with an increase in primary productivity. Non-carbonate mass accumulation rates at Site 1000 show an increase from the base of the cored interval throughout the late middle Miocene with peaks in the lower and middle Miocene (Reference 300).

#### 2.5.1.1.2.2.5.1 Geology of the Northern Nicaraguan Rise

##### Physiography of the Northern Nicaraguan Rise

The carbonate banks and reef shoals that are part of the northern Nicaraguan Rise are the Pedro Bank, Thunder Knoll, Rosalind Bank, Serranilla Bank, and Alice Shoal (Figure 2.5.1-314). These carbonate banks, knolls, and shoals are separated by four northeast-trending channels or troughs that range in depth from less than 400 meters to 1500 meters (from less than 1300 feet to 4900 feet). The channels deepen towards their ends but in most cases merge with canyons that lead down to the Pedro Escarpment or down into the Cayman Trough. Between the southern end of the Cayman Trough and the northern part of the northern Nicaraguan Rise, there is a broad boundary that rises to 500 meters (1640 feet). South of the Pedro Bank, the channel is floored by a plain at 1300 to 1400 meters (4300 to 4600 feet) depth. Linear depressions occur along the base of the Pedro Escarpment. The Jamaican Plain occupies one of these depressions. The line of the Pedro Escarpment and the Jamaican Plain is interrupted by the Banco Nuevo Ridge (References 555, 526, 558, 499, and 300).

##### Stratigraphy of the Northern Nicaraguan Rise

The rocks recovered from the north side of the northern Nicaraguan Rise are similar to those from the Cayman Ridge. Schistose metamorphics and plutonic rocks are absent from the stratigraphic section or might not have been sampled. However, breccias, wackes, and arenites contain detrital material indicative of granitic and metamorphic sources. Most of the metamorphics are low-grade greenschists. Shearing and cataclasis are evident in the rocks. An example is a quartzite that was composed of fragmental, recrystallized quartz in a polycrystalline quartz groundmass with minor opaque bands and degrees of recrystallization of fossiliferous micritic carbonates (Reference 528).

Interbedded arenites and graywackes with lesser amounts of argillite and carbonates are the most abundant sedimentary rock types along the northern Nicaraguan Rise. A reddish brown color on some of the breccias and wackes is due to oxidation of iron oxides and red-brown clay (Reference 252) (see discussion of marine red beds associated with LIPs in Subsection 2.5.1.1). Clastic carbonates and sedimentary rocks with carbonate cement and tuffs and tuffaceous clastic rocks were also recovered (Reference 528).

Dredging on the Walton Basin; the Pedro, Rosalind, and Diriangen Channels; and the northern part of Rosalind Bank recovered neritic limestone samples (consisting mostly of corals [*Montastrea costata*, *Stylophora* cf. *imperatoris*, and *Porites trinitatis*], green algae, and the benthic foraminifer, *Miogypsina gunteri*). The fossiliferous assemblages yielded an early Miocene age (22 to 20 Ma). High resolution seismic profiles in the interbank channels across the northern Nicaraguan Rise reveal that the basin and channel subseafloor consists of a series of foundered, faulted, and folded shallow carbonate banks and barrier reefs. These carbonate banks and barrier reef materials, possibly as young as Early/Middle (?) Miocene, were buried under a relatively recent periplatform sedimentary cover. The top of the neritic carbonate layer is marked by a major unconformity (Reference 236).

The northern Nicaraguan Rise was continuously covered by shallow carbonate banks and barrier reefs. Partial foundering of these banks and reefs occurred during the Middle Miocene and possibly as early as late Early Miocene (References 236 and 853). Foundering of the reefs and banks of the northern Nicaraguan Rise might have been the direct consequence of the initiation of the Caribbean Current and the development and strengthening of the North Atlantic Western Boundary Current (Figure 2.5.1-213) in the middle Miocene. (Reference 236).

### Structures of the Northern Nicaraguan Rise

The Nicaraguan Rise is bounded by the Cayman Trough to the north and by the Hess Escarpment to the south (References 601 and 602) (Figure 2.5.1-314). The Hess Escarpment extends for 1000 kilometers (620 miles) in a southwesterly direction and forms a divide between the Colombian Basin to the south and the Nicaraguan Rise to the north. It is a linear northeast-trending escarpment of highly variable relief (100 to 3000 meters (330 to 9800 feet), facing the Colombian Basin (Reference 526). Locally, an undeformed onlap sequence is seen over the escarpment and has a possible age of Late Cretaceous to Recent. The Hess Escarpment appears to form a major crustal boundary that separates blocks with different Neogene fault styles and basement characteristics. To the north of the southwestern end of the escarpment Neogene and possibly Quaternary north-south striking normal faults form a series of horsts and grabens (Reference 493).

The Pedro fault zone (Figure 2.5.1-317) divides the northern Nicaraguan Rise from the southern Nicaraguan Rise. Arden (Reference 601) describes oil industry wells from the Nicaraguan Rise that encountered plutonic rocks of Late Cretaceous and early Cenozoic age that are unconformably overlain by Cenozoic carbonate banks of the Nicaraguan Rise.

The western half of the northern Nicaraguan Rise is dominated by complex basement structural rises and normal faults compiled by Case and Holcombe (Reference 480) from private industry data. These faults have no consistent direction and range from <20 kilometers (<12 miles) to approximately 100 kilometers (62 miles) long. Rogers et al. (Reference 603) relate this faulting to the Colon fold-thrust belt of eastern Honduras, which records a Late Cretaceous shortening event due in part to the suturing of the Siuna terrane to the eastern Chortis terrane in the Late Cretaceous. They recognize thrust faulting and normal faulting in this area of the northern Nicaraguan Rise as starting in the late Cretaceous (post-80 Ma) and continuing into the Eocene, but ending by the beginning of the Oligocene.

The Eastern half of the northern Nicaraguan Rise contains many fewer identified faults, with the majority of these faults concentrated on the north near the Cayman Ridge province and in the south near the Pedro fault zone (Reference 480).

### Seismicity of the Northern Nicaraguan Rise

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates moderately sparse seismicity in the northern Nicaragua Rise. Magnitudes of these events range from approximately $M_W$ 3 to 7, with all but one event less than $M_W$ 6.0 (Phase 2 earthquake catalog) (Figure 2.5.1-267). The majority of the events are located proximal to the Cayman Ridge. Earthquakes south of the Cayman Ridge may

have occurred on the Cayman Ridge, but are mislocated, or may be correctly located and are due to stress effects near the Cayman Ridge. The Phase 2 earthquake catalog extends south to 15° N latitude (Figure 2.5.2-201) and does not cover the southern half of the northern Nicaraguan Rise.

#### 2.5.1.1.2.2.5.2 Geology of the Southern Nicaraguan Rise

##### Physiography of the Southern Nicaraguan Rise

The southern (or lower) Nicaraguan Rise appears to be a thickened oceanic crustal block bounded on the northwest by the Pedro Escarpment, on the southeast by the Hess Escarpment, on the northeast by the Morant Trough, and on the southwest by the San Andres Trough. The Hess Escarpment and other rift valleys and escarpments with the same northeast trend, occur across the southern Nicaraguan Rise (Figure 2.5.1-319). Scattered volcanic cones rise above the floor of the rise (e.g., La Providencia and San Andres Islands). Overall, the rise lies at a water depth of 2000 to 4000 meters (6500 to 13,100 feet), with depth increasing generally to the southeast. Based on multichannel seismic refraction data, the crust has been regarded as oceanic in origin, similar to the crust in the Colombian Basin to the south (References 526 and 604).

##### Stratigraphy of the Southern Nicaraguan Rise

The Caribbean crust in the southern Nicaraguan Rise area has been penetrated by drilling during DSDP Leg 15 (Site 152) and ODP Leg 165 (Site 1001) (Figure 2.5.1-211). ODP Site 1001 is located on the Hess Escarpment and is approximately 40 kilometers (25 miles) west-southwest of DSDP Site 152 (Figure 2.5.1-211). Seismic reflection data obtained from DSDP Leg 15 suggest that most of the deposits of the southern Nicaraguan Rise are uniformly pelagic and not characteristic of shallow-water deposits. Detailed lithologic descriptions are available for drill cores from both DSDP Site 152 and ODP Site 1001 (e.g., References 299, 604, and 606). The following provides a representative description of lithologies from ODP Site 1001.

According to Sigurdsson et al. (Reference 299), core recovered at ODP Site 1001 consists of four lithologic units. The basaltic basement (Unit IV) is radiometrically dated at about 77 Ma (mid-Campanian). Unit IV consists of a succession of 12 formations that likely represent individual pillow lavas and sheet flows. The margins are often highly vesiculated and glassy. The igneous basement is overlain by three sedimentary units. Based on fossil assemblages, the lowermost sedimentary unit, Unit III, is a Late Cretaceous sedimentary section of calcareous limestone and claystone with interbedded foraminiferal rich sand layers and ash layers that are thicker and more frequent in the lower part of Unit III. Unit II generally consists of calcareous chalk with foraminifers to mixed sedimentary rock with clay, and is interbedded with chert and volcanic ash layers and, near the bottom, is more clay rich with thin interbeds of foraminiferal rich sand layers. Based on fossil assemblages, Unit II corresponds with the Paleocene-Eocene section. The uppermost sedimentary unit, Unit I, generally consists of clayey nannofossil sediment to clayey nannofossil ooze with foraminifers, showing highly variable carbonate contents and magnetic susceptibility throughout the column (Reference 299). Based on fossil assemblages, Unit 1 corresponds to the Miocene-Pleistocene section.

The Cretaceous/Tertiary boundary interval was recognized in core recovered at ODP Site 1001 (Holes A and B); several clay rich units between the basal Paleocene and upper Maastrichtian limestones were also recovered. A 1.7- to 4.0-centimeter (0.7- to 1.6-inch) thick light gray, highly indurated limestone of earliest Paleocene age overlies the clay rich strata constituting the bulk of the recovered boundary deposit. The topmost layer of the boundary deposits is a 3.5-centimeter (1.4-inch) thick massive clay. This unit contains rare grains of shocked quartz and overlies a 3.5-centimeter (1.4-inch) thick smectitic claystone with dark green spherules. The base of the boundary deposit is a 1- to 2-centimeter (0.4- to 0.8-inch) thick smectitic clay layer with shaly cleavage. In addition to these three clay layers, two pieces of polymict micro-breccia were recovered consisting of angular clasts (<6 millimeters [<0.2 inches]) of claystone and limestone in an

unconsolidated matrix of smectitic clay. The total boundary deposit has an inferred thickness of approximately 25 centimeters (9.8 inches) (Reference 299).

According to Sinton et al. (Reference 606), $^{40}Ar$-$^{39}Ar$ incremental heating experiments of the basalts recovered on the southern Nicaraguan Rise in the vicinity of ODP Site 1001 and DSDP Site 152 indicate that the youngest period of volcanism occurred at about 81 Ma. Electron microprobe analyses show that the basalts are tholeiitic and generally similar to mid-ocean ridge basalts in composition. The comparatively low incompatible element concentrations (at the same MgO concentrations) in the ODP Site 1001 glass may signify derivation from either a more depleted mantle source or higher degrees of partial melting. The volcanism at this site is part of the continuing widespread submarine volcanism in the region that postdates the initial 90-Ma eruptions of the Caribbean oceanic plateau.

#### Structures of the Southern Nicaraguan Rise

The southern Nicaraguan Rise is a deep region of highly variable relief with rare scattered small carbonate banks, separated from the Colombian Basin in the south by the Hess Escarpment and separated from the northern Nicaraguan Rise by the Pedro fault zone (Reference 300) (Figure 2.5.1-317). Faulting in the southern Nicaraguan Rise ranges from <20 kilometers (<12.4 miles) to over 100 kilometers (62 miles) long and is dominated by a general west-southwest to east-northeast direction. It is comprised primarily of normal faults (Reference 480). Holcombe et al. (Reference 526) describe evidence for young faulting and volcanism within seismically active rifts imaged by marine seismic reflection profiles from the southern Nicaraguan Rise, and propose that diffuse east-to-west rifting of the rise occurs in response to sinistral shear along its bounding escarpments.

#### Seismicity of the Southern Nicaraguan Rise

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates sparse seismicity within the southern Nicaraguan Rise, all of which are $M_W \leq 5.5$. These earthquakes are primarily located in the northern portion of the southern Nicaraguan Rise near the Cayman Ridge, the Cayman Trough, and the southernmost extent of the Greater Antilles deformed belt. This seismicity, therefore, is likely related to its proximity to these active tectonic features. The Phase 2 earthquake catalog extends south to 15° N latitude (Figure 2.5.2-201) and does not cover the southern one-third of the southern Nicaraguan Rise.

### 2.5.1.1.2.2.6 Geology of the Colombian Basin

#### Physiography of the Colombian Basin

The center of the Caribbean Plate is divided into the Colombian and Venezuelan Basins separated by a north-south topographic high, the Beata Ridge (Figure 2.5.1-210) (Subsection 2.5.1.1.2.2.8). The basins are covered by flat-lying sediments and irregularities in the topography of the basement are attributed to volcanic features. The Colombian Basin is bounded by the southern Caribbean deformed belt to the south, the North Panama deformed belt to the west, and the Hess Escarpment to the north, a prominent, 1000 kilometers (620 miles) long bathymetric lineament. The southern Nicaraguan Rise (Subsection 2.5.1.1.2.2.5.2) and the Cayman Trough (Subsection 2.5.1.1.2.2.1) are located to the northwest of the Hess Escarpment (Reference 606). The North Panama and the South Caribbean deformed belts are underlain by thick sections of folded Cretaceous and Cenozoic sedimentary deposits. The deformed belts merge in the Gulf of Uraba where they form a V-shaped embayment. The western margin of the Colombian Basin is a narrow (10 to 20 kilometers or 6 to 12 miles) continental shelf offshore of Costa Rica and Panama. The eastern margin of the basin is defined by scarps related to normal- or oblique-slip faulting on the western side of the Beata Ridge (Reference 607).

## Stratigraphy of the Colombian Basin

In the Colombian Basin, recent faults strike northwest and bound the Mono Rise. A major unconformity suggests that uplift along current active northwest striking fault zones bounding the Mono Rise began in middle Miocene time (Reference 493).

The Magdalena Fan and the Colombian Plain dominate the sea-bottom morphology of the eastern half of the 3000 to 4000 meters (9800 to 13,100 feet) deep Colombian Basin. The Costa Rica Fan and Panama Plain occupy the southwestern extremity of the basin. Relief ranges from about zero to a few tens of meters; higher relief is associated with the Mono Rise, uplifted fault blocks, and channels on the fans. The dominant sediment source for the Colombian Plain and Magdalena Fan is the Rio Magdalena that drains from the Colombian Andes. The Costa Rica Fan's sediment source is from the rivers of eastern Honduras and the Central American mountains. The Panama Plain's sediment source is from the west in addition to the Rio Atrato in Colombia. Channels that are from the Panama Plain that lead into the Colombian Plain provide a pathway for Central American sediments to reach the center of the Colombian Basin (Reference 526).

Only the upper Miocene-Recent sediments have been drilled in the Colombian Basin. DSDP Site 154 (11° 0.5.11'N, 80° 22.75'W) was drilled on the Panama Outer Ridge (Figure 2.5.1-211). Sediments consisted of 153 meters (500 feet) of Pliocene and younger pelagic deposits that had constituents mainly composed of foram-bearing nanno-fossil marl. These pelagic deposits overlie a Pliocene and Miocene terrigenous sequence of deposits, which have calcareous, ash-bearing clay interspersed with black beds of pyrite and ash, containing turbidites. DSDP Site 502, Mono Rise (11° 29.42'N, 79° 22.78'W), consisted of cored material that was similar to Site 154; however, no turbidites were present in the calcareous clays of the lower unit (Reference 526) (Figure 2.5.1-211).

Seismic reflection records show that 1 to 3 seconds or about 1 to 4 kilometers (0.6 to 2.5 miles) of strata overlie an irregular oceanic crust in the central and western parts of the Colombian Basin; the strata consists of turbidite sequences, pelagic and hemipelagic deposits. There are two main reflector horizons, the A” and the B”. The A” reflector horizon coincides with the top of the Upper Cretaceous-middle Eocene siliceous pelagic carbonates, whereas the B” reflector horizon correlates with the top of the Upper Cretaceous basalt sill/flow complex. In the northeastern most Colombian Basin, the A” and B” reflectors extend beneath the Colombian Plain turbidites that are adjacent to the Beata Ridge and Hess Escarpment and are locally present around basement structural highs and within the central part of the basin (Reference 526).

Bowland (Reference 607) delineates five seismic stratigraphic units; they are from CB5 (oldest) to CB1 (youngest). Unit CB5 is mostly a sheet-drape deposit consisting of pelagic limestones, chalks, and clays (deposited in an open marine environment) that lies directly on igneous basement and is restricted to structural high areas of the oceanic plateau. The sequence is about 0.3 seconds or 0.5 kilometers (0.3 mile) thick over the Mono Rise and regionally thins to basement lows adjacent to the rise. Thinning of the unit to the west is caused most likely by transition through the carbonate compensation depth and/or erosion in a strong bottom current regime (References 607 and 526).

Unit CB4 is restricted to the highest areas of the Colombian Plateau and has a maximum thickness of about 0.8 seconds or 0.9 kilometers (0.6 mile) in the depression next to the North Panama deformed belt and at the crest of the Mono Rise. The seismic facies are hummocky-mounded to chaotic. This might be due to internal deformation of unconsolidated sediment. The sediments consist of upper Miocene siliceous microfossils and calcareous clay composed of poorly crystallized montmorillonite-smectite that might have a southern Central American province (References 607 and 526).

Units CB3 and CB2 are mid-Eocene to late Miocene in age and consist of unconfined turbidity-flow deposits interbedded with hemipelagic and pelagic layers. The turbidites are probably volcaniclastic related to volcanism north and west of the Colombian Basin during the Tertiary. Eocene and

Oligocene limestone and marl occur on the Nicaraguan Rise, which suggests that carbonate-clast turbidites may also be present (References 607 and 526).

Unit CB1 consists of a pelagic sequence on the Mono Rise and on the uplifted Site 154 fault block and gravity-flow deposits elsewhere. The unit includes the younger sediment wedge beneath the Panama Plain and the younger fan sequence underlying the Costa Rica Fan (late Miocene to Holocene in age) (Reference 607).

The crustal layers within the western Colombian Basin have velocities within the range of normal oceanic crust; however, the crustal thickness varies from near normal to more than twice the average for typical oceanic crust. Typical ocean crustal thickness is about 7 kilometers (4 miles). The top of the crust consists of ridges and basins and is at least 18 kilometers (11 miles) below the Mono Rise. Compressional wave velocities within the uppermost basement average about 4.6 kilometers/second. The basement of the western Colombian Basin, including the Mono Rise, exhibits a smooth upper surface and occasionally is stratified as indicated by well defined internal reflectors. Farther east towards the Magdalena Fan the reflectors are absent and the oceanic crustal thickness is about 8.5 kilometers (5.3 miles). Reflectors within the eastern foundation of the Mono Rise overlap the rough basement which indicates that the rise may be younger. Heat flow unit (hfu) averages 1.57 hfu in the western basin but only 1.6 hfu east of the rise (Reference 526).

In an effort to explain the thickness of oceanic plateau crust and corresponding greater depth to the Moho (up to 16 kilometers or 10 miles below sea level), various researchers proposed that the Caribbean was an area of extensive intrusion by primary basaltic magma (Reference 608). The A" and B" reflector horizons show up on seismic profiles in the Colombian and Venezuelan Basins, and samples can be obtained from on land sections in Costa Rica, Colombia, and Curaçao. On DSDP Leg 15, reflector horizon B" was sampled at five drill sites with recovery of only about 15 meters (50 feet) of basement. The samples consist of basalt and diabase whose mineralogy and geochemical characteristics are distinct from those of the typical mid-ocean ridge basalt (Reference 605). This discovery led to the recognition of a Coniacian to early Campanian flood basalt event within the Caribbean. The flood basalt extends for 600,000 kilometers$^2$ (232,000 miles$^2$) and is exceptionally thick (up to 20 kilometers or 12 miles). The top of the plateau is the widespread smooth B' seismic reflector (Reference 609).

Radiometric ages indicate that the Caribbean Plateau formed during at least two major magmatic events, the first at about 90 to 88 Ma and the second at about 76 to 72 Ma (Reference 610). Revillon et al. (Reference 611) uses petrographic and geochemical data to demonstrate that the magmas produced during the different episodes have very similar petrological and chemical compositions. These data indicate that all the magmas came from a mantle source of similar composition and that the conditions under which they formed were reproduced at least three times from the Cretaceous into the Tertiary. Evidence to support fractional melts (in the spinel stability field) is the uniform, flat rare earth element patterns found in the gabbros and dolerites that were derived from an isotopically depleted source (Reference 611).

Revillon et al. (Reference 611) dated several samples by the $^{40}Ar/^{39}Ar$ method, either on whole rocks or separated plagioclases. Most samples have ages between 80 and 75 Ma, which are consistent with previous ages within the province, but a subordinate intrusive phase occurred at about 55 Ma (Reference 611).

### Structures of the Colombian Basin

The Colombian Basin primarily comprises a depositional basin, with sediments ranging in age from Late Cretaceous to Eocene (Reference 607). The Colombian Basin is underlain by the oceanic plateau type crust, which has been dated to 69 to 139 Ma (Reference 245). This 70 m.y. period of continuous igneous activity is in sharp contrast to other data that indicate two major pulses (at 92 to 88 Ma and 76 to 72 Ma) of igneous activity created the Caribbean oceanic plateau (see CLIP

discussion in Subsection 2.5.1.1). It is recognized as normal oceanic crust in thickness but the crust is overlain by nearly 2 kilometers (1.2 miles) of sediment (Reference 612). Bowland and Rosencrantz (Reference 613) used seismic reflection data to interpret that the eastern margin of the Colombian Basin is defined by scarps related to normal- or oblique-slip faults associated with the western Beata Ridge.

Bowland (Reference 607) describes a fault-bounded block adjacent to the Hess Escarpment and west of the Mono Rise, likely uplifted in Miocene to Holocene time. This block has a positive free-air gravity signature (Reference 614) and is aligned with several faults that extend to the southwest and displace basement and overlying sediments (References 766 and 613).

Bowland and Rosencrantz (Reference 613) recognize a zone of closely spaced normal faults and faults associated with a horst that displaces basement at least 500 meters (1600 feet) in the Colombian Basin. They also recognize a zone of normal faults that disrupts basement where the Mono Rise encounters the North Panama deformed belt and small-offset normal faults that displace basement on the southwestern flank of Mono Rise. Normal faults on the Colombian Plateau may be the result of thermal contraction and differential subsidence of laterally heterogeneous crust.

#### Seismicity of the Colombian Basin

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates sparse seismicity within the Colombian Basin with $M_W < 6$. These earthquakes are located in the northeastern portion of the Colombian Basin near the Beata Ridge and the southern extension of the Greater Antilles deformed belt. The Phase 2 earthquake catalog does not cover the southern two-thirds to three-quarters of the Colombian Basin province.

### 2.5.1.1.2.2.7 Geology of the Venezuelan Basin

#### Physiography of the Venezuelan Basin

The southern portion of the Caribbean Plate includes the Colombian and Venezuelan Basins separated by a north-south topographic high, the Beata Ridge (Figures 2.5.1-202 and 2.5.1-210) (Subsection 2.5.1.1.2.2.8). The basins are covered by flat-lying sediments. Irregularities in the topography of the basement are attributed to volcanic features and structural offsets. The Venezuelan Basin is bounded on the west by the Beata Ridge, on the north by the Muertos Trough (Subsection 2.5.1.1.2.2.4), on the east by the Aves Ridge, and on the south by the south Caribbean marginal fault. At the south Caribbean marginal fault, the Venezuelan Basin is obliquely subducted to the east-southeast beneath the continental South America Plate (Reference 615).

The Venezuelan Basin is floored by oceanic crust that lies at water depths of between 3 and 5 kilometers (2 and 3 miles). The topography of the basin is subdued.

#### Stratigraphy of the Venezuelan Basin

Venezuelan Basin is underlain by igneous oceanic crust throughout, marked by the B” seismic horizon. The seismic stratigraphy to the level of B” is derived from data collected at DSDP Sites 146 and 149 (Figure 2.5.1-211) and later data collected at ODP Site 165. In the western region, the B” horizon is a smooth surface, whereas in the eastern part the B” horizon has the rough surface (References 616 and 617). Northeast-trending magnetic anomalies in the basin have been interpreted as reflecting crustal accretion at a spreading ridge, between Late Jurassic and mid-Early Cretaceous (127 and 155 Ma) (Reference 618). Prior to the mid-Late Cretaceous (Senonian at ~88 Ma), widespread and rapid eruption of basaltic flows began in concert with extensional deformation of the Caribbean crust. Thick volcanic wedges characterized by divergent reflectors that are observed along the boundary that separates rough from smooth oceanic crust are coincident with an abrupt shallowing of the Moho and appear to be bounded by a large, northwest-dipping fault system (Reference 255).

The outer margins of the basin are dominated by thick, turbidite-filled abyssal plains, which have not been penetrated by deep-sea drilling (Reference 526). On the other hand, DSDP drill sites in the interior of the basin have recovered a thick succession of pelagic sediments (DSDP Sites 29, 146, 149, and 150) (References 619 and 620). Upper Cretaceous limestone and marls containing basaltic ash overlie the igneous basement, but at DSDP Site 153 the Upper Cretaceous sediments include carbonaceous clays, which imply euxinic conditions and restricted circulation during early evolution of the basin. Paleocene limestones and clays are overlain by lower Eocene cherts and hard siliceous limestone, which mark seismic horizon A". Miocene to Oligocene deposits are foraminiferal-nannofossil chalks and clays. Holocene to Miocene deposits are foraminiferal-nannofossil chalk oozes, marl oozes, and clays.

The stratigraphy to the level of B" reflector horizon of the Venezuelan Basin is derived from data collected at DSDP Sites 146 and 149 (Figure 2.5.1-211). The holes were nearly continuously cored and provide a 762-meter (2500-feet) composite section that represents pelagic sediments. No major unconformities were found and, as a result, most of the foraminiferal and nannofossil biostratigraphic zones and several of the radiolarian zones were identified. Recent to lower Miocene deposits are foraminiferal-nannofossil chalk oozes, marl oozes, and clays. Lower Miocene to lower Eocene deposits are radiolarian-nannofossil chalks and oozes thick in volcanic material. Underlying the middle Tertiary sediments is a lower Eocene (?)-Paleocene (?) section of chert associated with limestone (Reference 526).

In the deepest part of the Venezuelan Basin, the B" surface is rough compared to areas where the B" surface is smooth, requiring the distinction between rough B" and smooth B". The smooth B" may represent the older proto-Atlantic Plate. The overlying finely laminated sequence was designated A," corresponding to older than Middle Eocene (approximately 50 Ma) and younger than Senonian (approximately 88 Ma) consolidated cherts and chalks. The A" to B" sequence varies in thickness across the Caribbean, up to a maximum of 600 to 800 meters (2000 to 2600 feet), with the thickest sequence roughly coincident with areas of rough basement in the Venezuelan A" to B" for rough B" areas (Reference 253). Similarly, Driscoll and Diebold (Reference 253) note that the hemipelagic-pelagic sediment sequence above A" displays a pronounced increase in thickness across the rough-smooth B" boundary in the Venezuelan Basin. This increase thickness is interpreted to have been caused by a hiatus or non-deposition toward the northwest and away from the depositional center of the Venezuelan Basin.

Using multi-channel seismic reflection data, other maker horizons were identified. Leroy and Mauffret (Reference 621) recognized that B" is sometimes overlain by a thin layer, 2V, interpreted to be original oceanic crust overlain by a thin volcanic layer. Below B", an intra basement reflector (sub-B?) marks the top of original oceanic crust that is sandwiched between an upper volcanic layer and lower underplated material. This underplated layer that forms a very thick layer, 3V, beneath the Beata and Nicaragua volcanic plateaus, is attributed to the presence of magnesian-rich rocks (picrites or ultramafic cumulates). The upper part of layer 3V is gabbroic and outcrops on the Beata Ridge. A highly reflective horizon (R) is located at the top of this layer.

Horizon A" was shown to be overlain by reflector eM, with Horizon A" representing the boundary between unconsolidated Early Miocene to Eocene oozes and consolidated Lower Eocene cherts and chalks-ooze and Early to Middle Miocene calcareous ooze. According to James (Reference 608), DSDP/ODP drilling showed that A" marks the top of a middle-Eocene chert-limestone section below unconsolidated sediments.

Horizon B" is smooth over the Caribbean Plateau and rough in areas of the Caribbean underlain by normal oceanic crust. Smooth B" ties to 90 to 88 Ma basalts sampled by drilling (Reference 605) and these are interpreted to indicate voluminous plateau volcanism over a short period. As discussed in Subsection 2.5.1.1.2.2.6, radiometric ages have identified at least two major magmatic events responsible for the production of the Caribbean Plateau, the first and largest at about 90 to 88 Ma

and the second at about 76 to 72 Ma (Reference 610). Rough B" has never been penetrated by drilling. The rough B" profile is seen in the southeastern Venezuelan and western Colombian Basins (References 613 and 255) and is thought to represent "normal" thickness of the proto-Atlantic oceanic crust.

#### Structures of the Venezuelan Basin

The Venezuelan Basin consists of thicker than normal oceanic crust of Jurassic age that was thickened by emplacement of dikes and sills in Jurassic to early Cretaceous time, and then intruded by sills and flows in the mid- to late-Cretaceous (References 623 and 624). Faults and monoclines of Miocene and younger age are seen in the basin interior, indicative of minor internal deformation (References 623 and 625).

#### Seismicity of the Venezuelan Basin

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) shows scattered, sparse seismicity of $M_W \leq 4$ in the northwestern part of the Venezuelan Basin. The Phase 2 earthquake catalog extends south to 15° N latitude (Figure 2.5.2-201) and does not cover the southern one-half portion of the Venezuelan Basin.

### 2.5.1.1.2.2.8 Geology of the Beata Ridge

#### Physiography of the Beata Ridge

The 2000-meter (6600-feet) deep Beata Ridge is a prominent topographic structure that trends south southwest from Cape Beata in Hispaniola and divides the 4000- to 5000-meter (13,100- to 16,400-foot) deep Colombian and Venezuelan Basins (Figure 2.5.1-210). It is 450 kilometers (280 miles) long and up to 300 kilometers (186 miles) wide, with a highly asymmetrical east-west profile due to a steep (15 to 25°) escarpment to the west that rises 2500 meters (8200 feet) above the Colombia Abyssal Plain, and a gentler slope to the east to the Venezuelan Basin (References 625, 778, and 628).

#### Stratigraphy of the Beata Ridge

In general, dredge material from the Beata Ridge consists of igneous rocks, holocrystalline basalts, and dolerites (Reference 626). Three discrete units are identified at DSDP Site 151 (Figure 2.5.1-211). Unit I consists of Tertiary pelagic sediments rich in carbonate faunal assemblages. Only fragments of the Paleocene and Eocene sequence are present. Three meters (10 feet) of basalt were recovered, but the contact with the overlying sediments was not recovered. Unit II is the hard ground that marks an unconformity between the Paleocene sediments and the overlying Santonian age sediments. Unit III is characterized by foraminiferal sands, volcanics, and carbonaceous clays of Santonian age and is capped by a siliceous hard ground (Reference 605). Magmatic samples that were collected during 12 selected dives (NB-04 to NB-16) that were distributed from north to south of the ridge. The samples consist of gabbro and dolerite that formed relatively continuous massive outcrops or boulders up to a few tens of centimeters across in talus. Based on subtle differences in structure, these rock units are interpreted as a sequence of sills. Some of the outcrops show concentric spheroidal forms; this alteration was superimposed on an earlier phase of sea floor alteration. Volcanic rocks are rare, but where present always formed pillowed lava flows. Basalts were observed at the base of the escarpment below outcrops of gabbro and dolerite (Reference 611).

The deepest dredge located at the base of the escarpment, 4100 meters (13,450 feet), contains deeply weathered rocks that are completely altered to clay, zeolite, and limonite phases. Nine dredge hauls contain igneous rocks in various states of weathering. They were distributed at depths ranging from 4000 meters to 2300 meters (13,100 to 7550 feet). The majority of the samples in all nine dredges are holocrystalline with textures ranging from ophitic to glomeroporphyritic. Several samples

found in dredge hauls 10, 12, and 31 have a porphyritic texture but the groundmass has a hemihyaline texture and is composed of a mixture of palagonite, acicular plagioclase, and opaque oxides (Reference 626).

Numerous dredged samples of basalt from the Beata Ridge were radiometrically dated (feldspars and whole rock) at 64 to 65 Ma. Several samples contain olivine or pseudomorphs after olivine, which might represent the eruption of linear intrusive bodies associated with block faulting of the Beata Ridge. The correlation of these bodies to reflectors A" and B" east and west of Site 151 (located off the Beata Ridge) would indicate a date of at least late Cretaceous (Reference 629).

### Structures of the Beata Ridge

The Beata Ridge (Figure 2.5.1-210) extends from south-central Hispaniola on the north to the Aruba Gap at about 14° N to the south (Figure 2.5.1-316). It is a roughly triangular shaped region, about 200 kilometers (124 miles) north to south, and about 200 kilometers east to west at 14° N. The northern tip of the triangle is on land and comprises the Bahoruco Peninsula (for location, see morphotectonic zone 7 of Figure 2.5.1-305) of south-central Hispaniola, the southwestern corner of the triangle is DSDP Site 151 Ridge (a north-south ridge northwest of the Aruba Gap), and the southeastern corner of the triangle is the Beata Plateau (Figures 2.5.1-210 and 2.5.1-316). Relief generally decreases from the north, which is above sea level, to 4 kilometers (2.5 miles) below sea level to the south where it ends in the Aruba Gap. The northern termination also coincides with the eastern end of the Enriquillo-Plantain Garden fault and the western end of the Muertos Trough.

Mauffret and Leroy (Reference 630) present a detailed tectonic analysis of this feature, based on multi-channel seismic surveys, DSDP results, bathymetry from a Seabeam (SEACARIB I) survey, and focal mechanism studies of one earthquake. Because this reference appears to be the most comprehensive analysis to date, it provides the source for the summary below, unless otherwise stated.

The Beata Ridge consists of unusually thick oceanic crust (about 20 kilometers or 12 miles), formed by underplating of normal oceanic crust in the late Cretaceous, creating an oceanic volcanic plateau with subsequent transpression and uplift in the mid-Miocene. A petrologic analysis of dredged rocks identified three episodes of emplacement, one at 80 Ma, one at 76 Ma, and the last at 55 Ma (Reference 611). These authors propose that the first two episodes are related to original formation of the CLIP (see discussion of "Large Igneous Province (LIP) Events" in Subsection 2.5.1.1) over probably more than one hotspot in the Pacific, and the third is due to later localized crustal thinning with contemporaneous magma emplacement.

Sub-elements within the Beata Ridge are (from north to south): the Tairona Ridge, the DSDP Site 151 Ridge, the Taino Ridge, and the Beata Plateau (Figure 2.5.1-316). It is bordered by the Colombian Basin and Haiti subbasin to the west, the Dominican subbasin and Venezuelan Basin to the east, and the Aruba Gap to the south. The west side of the Beata Ridge forms a relatively steep escarpment, with northeast-southwest oriented right-lateral strike-slip faults strongly suggested between Tairona Ridge and the Bahoruco Peninsula, and between DSDP 151 Ridge and Tairona Ridge. The Beata Ridge decreases in elevation from west to east, with the east side showing evidence for west-verging thrust faults. This indicates that the ridge is overriding the Venezuelan Basin. An east-west seismic line across the Taino Ridge shows evidence for initial east-west normal faulting, followed by later thrust faulting in the opposite direction on the same feature (Reference 630).

A tectonic model for the Beata Ridge and its relationship to surrounding elements of the Caribbean region is shown in Figure 2.5.1-317. Sheet 1 shows the proposed configuration in the early Miocene. The role of the Beata Ridge then was to accommodate differential motion between the Colombia and Venezuela microplates via southwest-dipping thrust faults. Sheet 2 shows proposed relations at present. The ridge still accommodates Colombia-Venezuela microplate differential motion (the Colombia microplate moving eastward faster than the Venezuela microplate), but due to the

counterclockwise rotation of the Venezuela microplate, the deformation is partitioned into strong transpression (manifested largely as northeast-southwest strike-slip faults) on the west side and thrust faulting, as the Venezuela microplate is being overridden, on the east side. Since the early Miocene, closure between the North and South America plates has caused the north end of Beata Ridge to collide with the Hispaniola microplate. On the south end, the 40-kilometer (25-mile) wide Aruba Gap accommodates the differential motion via the Pecos fault zone, a transpressive zone exhibiting strike-slip and reverse faulting. The Euler pole for this system is placed just south of the south end of Beata Ridge (Figure 2.5.1-317), consistent with the increasing deformation, and topography, of the Ridge from south to north (Reference 630).

**Seismicity of the Beata Ridge**

The Phase 2 earthquake catalog (Subsection 2.5.2.1.3) indicates sparse seismicity in the vicinity of the Beata Ridge, the largest earthquake having $M_w$ 4.8. Seismicity from 1900 to 1994 (Reference 631) shows one earthquake near the southern end of the Bahoruco Peninsula (see morphotectonic zone 7 of Figure 2.5.1-305) of a magnitude $M_w$ 4, and a $M_w$ 5.8 earthquake (Reference 632) near the south end of the Taino Ridge. A focal mechanism for this earthquake (Reference 633) shows northeast-southwest directed thrust faulting, consistent with the tectonic model shown in Figure 2.5.1-317. The Beata Ridge is believed to be an oceanic spreading ridge that was active 80 to 55 Ma comprising unusually thick (20 kilometers or 12 miles) oceanic crust (Reference 611). As such, it may provide a zone of weakness in the crust and thus generate small to moderate earthquakes. The extent of the Phase 2 earthquake catalog is south to 15° N latitude (Figure 2.5.2-201), and does not cover the southern one-third to three-quarters of the Beata Ridge.

#### 2.5.1.1.2.3 Active Tectonic Structures of the Northern Caribbean Plate

Active tectonic structures on the southeastern North America Plate are described in Subsections 2.5.1.1.1.3 and 2.5.1.1.2.1. This subsection describes the active tectonic structure of the northern Caribbean Plate. The structures are grouped as single faults, fault systems, or spreading centers. Some faults and fault systems are transforms and one is a subduction zone. This following discussion emphasizes tectonic elements that are either (a) capable of generating large to great earthquakes (i.e., M [magnitude] approximately 7.5 or greater) and/or (b) within the 200-mile radius site region.

The Caribbean Plate is presently moving relative to the North America Plate at a rate of approximately 20 millimeters/year along an azimuth of roughly 075° (References 502, 635, and 636). Cuba was transferred to the North America Plate in the early to mid-Tertiary, and thus is not directly involved in the plate boundary tectonics, except along its southern coast. In the Caribbean-North America Plate boundary region, the relative plate motion is accommodated by the mid-Cayman spreading center and several subvertical, left-lateral transform faults extending from offshore of the northern coast of Honduras eastward through the Cayman Trough and through Jamaica and Hispaniola. The Cayman spreading center is located southwest of the Cayman Islands and is characterized by a north-south-trending axis of spreading with an average rate of approximately 15 millimeters/year since approximately 25 to 30 Ma (Reference 222). West of the Cayman Trough, Caribbean-North America Plate motion is accommodated offshore on the left-lateral Swan Islands fault (Figure 2.5.1-202). East of the Cayman Trough, on Hispaniola, the orientation of the plate-bounding structures changes and motion is partitioned between strike-slip faults (e.g., Septentrional and Enriquillo faults), minor oblique-reverse faults, and subduction on thrust faults (e.g., Northern Hispaniola thrust fault) (References 637, 638, and 639). East of Hispaniola, the Caribbean-North America Plate boundary becomes an oblique subduction zone or zones at the Puerto Rico Trench and Muertos Trough, and finally a more pure dip-slip west-dipping subduction zone in the Lesser Antilles.

The kinematics of crustal deformation and faulting in Cuba are poorly understood. Geodetic data show that the current plate boundary is mostly south of Cuba along the Oriente and

Enriquillo-Plantain Garden faults and that modern deformation rates across Cuba are likely <0.1 inch (3 millimeters) per year relative to North America (References 502 and 503). Some strike-slip faults have been mapped on Cuba, but none are adequately characterized with late Quaternary slip rates or timing or recurrence of large earthquakes (Reference 494). The Oriente and Enriquillo-Plantain Garden faults are active left-lateral strike-slip faults associated with the North America-Caribbean Plate boundary.

The Oriente fault zone is a left-lateral transform fault extending from the northern tip of the Mid-Cayman spreading center 500 miles (800 kilometers) to the southeastern tip of Cuba. The remainder of North America-Caribbean Plate motion that is not accommodated along the southern Cayman Trough boundary, or approximately 8 to 13 millimeters/year, is attributed to this fault. Again, variation in historical seismicity and geometry of the Oriente fault warrants its division into eastern and western segments. The largest historical earthquakes on the western Oriente fault are the 1992 $M_w$ 6.8 to 7.0 event (Phase 2 earthquake catalog $M_w$ 6.80) and a magnitude 7.0 to 7.1 (Phase 2 earthquake catalog $M_w$ 7.20) earthquake that occurred off of the southwestern tip of Cuba (References 640, 489, and 641). The eastern Oriente fault along southern Cuba is characterized by more intense seismic activity and focal mechanisms indicating strike-slip, oblique, and reverse mechanisms (References 504 and 640). The largest historical earthquake on the eastern Oriente fault is the June 1766 $M_w$ 7.53 earthquake (Subsection 2.5.2.1.3).

The Septentrional fault is a left-lateral strike-slip fault that extends for roughly 400 miles (640 kilometers) west from the Mona Passage to the Windward Passage, where it merges with the Oriente fault (References 840 and 637). Strain is partitioned on this structure and on the gently south-dipping Northern Hispaniola thrust fault (References 591 and 643). The best estimate of a slip rate for the fault is 6 to 12 millimeters/year (References 636, 570, and 643), and it has been suggested that large historical earthquakes ($M_w$ 7.75 to 8.0) occurred on this structure (Reference 641).

The Northern Hispaniola fault is an east-west-striking, north-directed thrust system. Geodetic data indicate a deformation rate of 5 millimeters/year on this structure (Reference 358). Historical seismic events of up to $M_s$ 8.1 (Phase 2 earthquake catalog $M_w$ 7.90) have been attributed to a shallowly south-dipping thrust fault plane (Reference 591). Variations in seismicity and crustal structure along strike indicate the fault is segmented and best described by a more seismically active eastern segment and a quieter western segment that roots at the Septentrional fault.

The Swan Islands, Walton-Duanvale, and Enriquillo-Plantain Garden fault systems are left-lateral strike-slip faults associated with the mid-Cayman spreading center, which collectively form the southern margin of the Cayman Trough. The estimated slip rate for the system is approximately 8 millimeters/year (References 503 and 502). Slip is transferred more than 600 miles (970 kilometers) across these structures (causing a restraining bend in Jamaica) and eventually feeds into the Muertos Trough. The Jamaican restraining bend is interpreted as a boundary between a western portion of the system (the Walton-Duanvale fault) and an eastern portion (Enriquillo-Plantain Garden fault). Multiple historical events of magnitude approximately 7.5 have ruptured on the Enriquillo fault (Reference 641). The Swan Islands fault system is a left-lateral oceanic transform extending 450 miles (720 kilometers) west of the mid-Cayman spreading center. Geodetic data indicate that essentially the entire 18 to 20 millimeters/year Caribbean-North America Plate motion is accommodated on the Swan Islands fault system (References 502 and 635). An historical earthquake with an estimated magnitude of 8.3 (Phase 2 earthquake catalog $M_w$ 7.69) is attributed to the western portion of the Swan Islands fault system (Reference 641).

#### 2.5.1.1.2.3.1 Cayman Trough Tectonic Structures

The Cayman Trough comprises a central north-northwest-trending spreading axis, with strike-slip faults extending both east and west from its southern terminus and a strike-slip fault extending east from its northern terminus (Figure 2.5.1-202). Extending east from the northern end of the spreading

axis is the left-lateral Oriente fault, which connects with the Septentrional fault on the island of Hispaniola. From the southern end of the spreading axis, the Swan Islands fault extends to the west, eventually linking with the Motagua fault in Honduras. To the east of the southern end of the spreading axis, Walton fault, Duanvale fault and Enriquillo-Plantain garden fault extend eastward through Jamaica to Hispaniola. The submarine portions of these structures were mapped with a sidescan instrument (Reference 559). The spreading axis itself is offset by a short discontinuity. Seismicity indicates this is a left-lateral strike-slip fault (Reference 499). The Oriente fault is described in detail in Subsections 2.5.1.1.1.3.2.4 and 2.5.1.1.2.3.1.2.

The Cayman Trough tectonic structures include two major fault systems, the western and eastern segments of the Swan Islands fault and the western and eastern segments of the Oriente fault. The two fault systems are described in the following subsections.

#### 2.5.1.1.2.3.1.1 Swan Islands Fault

The Swan Islands fault is a left-lateral oceanic transform fault that extends from the southern tip of the mid-Cayman spreading center westward for roughly 450 miles (720 kilometers) where it merges with the onshore Polochic-Motagua fault system of Central America (Figure 2.5.1-202). West of the mid-Cayman spreading center, the northern margin of the Cayman Trough does not appear to accommodate significant left lateral relative plate motion; essentially the entire 18 to 20 millimeters/year North America-Caribbean Plate motion is accommodated on the Swan Islands fault (Reference 502).

Interpretation of high-resolution sea-floor bathymetry suggests the Swan Islands fault consists of several faults that locally form restraining and releasing geometries (References 563 and 655). West of the Swan Islands, the Swan Islands fault is expressed on the sea floor as a relatively continuous lineament. The thickened crust associated with the emergent Swan Islands is associated with a roughly 20-mile (32-kilometer) wide right step-over that forms a restraining geometry and a probable segmentation point for rupture propagation. Surrounding and east of the Swan Islands, the fault consists of one or more sections of about 60 to 120 miles (100 to 200 kilometers) in length to the eastern termination at the mid-Cayman Trough. Here, the crust of the mid-Cayman Trough that bounds the fault to the north is about 3.5 miles (5.5 kilometers) thick based on gravity (Reference 635).

McCann and Pennington (Reference 560) and McCann (Reference 641) note a large earthquake that occurred in August 1856 off the northern coast of Honduras may have ruptured the western portion of the Swan Islands fault. The estimated magnitude for this event is about M 8.3 (Reference 641), based on descriptions of the event summarized by Osiecki (Reference 645) (Table 2.5.2-221). Earlier accounts of a similar great event off the northern Honduran coast suggest the possibility of a prior magnitude of approximately 8 (Phase 2 earthquake catalog $M_W$ 7.69) earthquake on the western Swan Islands fault in 1539 (Reference 641). The probability of at least one great historical earthquake on the western Swan Islands fault suggests the fault is fully coupled. The eastern section of the fault between the Swan Islands and the mid-Cayman spreading center is not associated with large historical earthquakes.

#### 2.5.1.1.2.3.1.2 Oriente Fault

The Oriente fault is a left-lateral transform fault that forms the northern boundary of the Gonâve microplate and extends for more than 500 miles (800 kilometers) from the southeastern tip of Cuba westward to the northern tip of the mid-Cayman spreading center (References 632, 840, 559, and 844) (Figure 2.5.2-214). To the east, the Oriente fault connects with the Septentrional fault in the Windward Passage. Slip-rate on the Oriente fault is estimated at 8 and 13 millimeters/year, with a best estimate of 11 millimeters/year. This estimate is based upon subtracting the approximate 7 to 11 millimeters/year rate of Gonâve-Caribbean relative motion measured in Jamaica (Reference 503)

and Haiti from the entire 18 to 20 millimeters/year North America-Caribbean Plate motion (Reference 502).

The structural complexity and historical seismicity of the Oriente fault changes character along strike and forms the basis of a division into western and eastern sections (Figure 2.5.2-214). The western Oriente fault extends from the mid- Cayman spreading center to the southern tip of Cuba and the offshore Cabo Cruz Basin. This section of the fault is characterized by a simple, linearly continuous expression on the seafloor trending almost exactly parallel to relative Caribbean-North America Plate motion (References 840, 502, and 559). Seismicity on the western Oriente fault is less frequent than on other areas of the plate boundary, including on the eastern Oriente fault. Most seismicity has been localized in the Cabo Cruz pull-apart basin, which is associated with left-lateral strike-slip-normal oblique motion (References 504 and 640). The largest historical earthquakes on the western Oriente fault are the May 1992 magnitude 6.8 to 7.0 (Phase 2 earthquake catalog $M_W$ 6.80) earthquake on the Cabo Cruz Basin and the February 1917 M 7.0 to 7.1 (Phase 2 earthquake catalog $M_W$ 7.20) earthquake that occurred offshore the southern tip of Cuba (References 640, 641, and 489). A magnitude 6.2 earthquake in 1962 (Phase 2 earthquake catalog $M_W$ 6.29) on the western Oriente fault adjacent to the Cayman spreading center is the largest historical event west of the Cabo Cruz Basin and reveals pure left-lateral strike-slip motion (Reference 640). It is unclear if the low seismicity rate on the western Oriente fault west of the Cabo Cruz Basin indicates it is fully locked, or if it is mostly unlocked and sliding at a relatively uniform rate. As mentioned previously, the crust of the Cayman Trough that constitutes the southern block of the Oriente fault is anomalously thin (2 to 6 kilometers or 1 to 4 miles) for distances up to 200 miles (320 kilometers) or more from the mid-Cayman spreading center (Reference 844), which probably limits the seismogenic thickness of the western Oriente fault. A low coupling of the western Oriente fault west of the Cabo Cruz Basin would be consistent with oceanic transform faults worldwide, for which up to 95 percent of total slip is released aseismically (Reference 843).

The eastern Oriente fault extends along southern Cuba and is characterized by a zone that includes: (a) segmented, discontinuous, and probably vertical strike-slip faults and (b) more continuous, steeply north-dipping faults of the Santiago deformed belt south of the strike-slip faults (Reference 840). The eastern Oriente fault is characterized by more intense seismic activity than the western Oriente fault (Figure 2.5.2-215), with focal mechanisms indicating strike-slip, oblique, and reverse mechanisms (References 504 and 640). Seismicity depths reach 70 kilometers (45 miles) beneath southern Cuba associated with the Santiago deformed belt, indicating a thick seismogenic crust that contrasts with the thin crust of the western Oriente fault (Reference 504). The seismic moment release of historical large earthquakes is consistent with the approximately 11 millimeters/year slip rate on the Oriente fault determined by GPS (References 840 and 843), indicating that the plate interface there is fully locked (Reference 643).

#### 2.5.1.1.2.3.2 Greater Antilles Deformed Belt Faults

While the previous sections describe tectonics of individual components of the Greater Antilles deformed belt, a number of recent studies have attempted to use GPS and other geophysical information to infer seismic hazards for the region as a whole by integrating these observations into a regional, self-consistent model.

Dixon et al. (Reference 780), using campaign GPS measurements over a ten year period (1986 to 1995), find that the North America-Caribbean relative motion was about 21 millimeters/year, twice the NUVEL-1A rate deduced from global plate rate inversions (References 649 and 650). Using elastic strain accumulation models for the Northern Hispaniola fault, Septentrional fault, and Enriquillo fault, they inverted for slip on these features. Their results indicate 4 ± 3 millimeters/year on the Northern Hispaniola fault, 8 ± 3 millimeters/year on the Septentrional fault, and 8 ± 3 millimeters/year on the Enriquillo fault. It is important to note that they assumed no aseismic slip and neglected any postseismic deformation effects, which may have affected northeast Hispaniola after the 1946

earthquake. These effects, however, are unlikely to perturb the results by more than 1 to 2 millimeters/year (Reference 651). These results were found to be consistent with a broader study encompassing the northern North America-Caribbean plate boundary region (Reference 502). Calais et al. (Reference 358) performed a similar study for Hispaniola and calculate 5.2 ± 2 millimeters/year for the Northern Hispaniola fault, 12.8 ± 2.5 millimeters/year for the Septentrional fault, and 9.0 ± 9.0 millimeters/year for the Enriquillo fault.

An update of this analysis, expanded to the northeastern Caribbean from western Hispaniola to the central Lesser Antilles, was presented by Manaker et al. (Reference 643). They also modeled coupling ratios (if a fault slips aseismically this is 0, if fully locked it is 1), which estimate how much motion is translated into earthquakes. Fault slip rates are shown in (Figure 2.5.1-318). Rates on the Northern Hispaniola fault are 5 to 6 millimeters/year, 8 ± 5 millimeters/year on the Septentrional fault, and 7 ± 2 millimeters/year on the Enriquillo fault. These are consistent with the estimates mentioned above. The coupling ratios (Figure 2.5.1-318) show the Septentrional fault to be tightly coupled to the west, with a decrease to the east. This is consistent with the concept that the impingement of the Bahama Bank on the Caribbean Plate gives rise to high coupling and high seismicity, and that to the east subduction of normal oceanic crust decreases coupling and consequently reduces the seismic hazard (e.g., Reference 577).

Ali et al. (Reference 652) modeled Coulomb stress changes in the northeastern Caribbean due to the occurrence of 12 historic earthquakes, including effects of postseismic viscoelastic relaxation. These stress changes were then interpolated to three-dimensional representations of the major faults. The authors suggest that the 1751 event on the eastern Enriquillo fault was “encouraged” by the >0.1 MPa stress increase caused by the 1751 Muertos Trough earthquake, and that the east-to-west progression of earthquakes on the Northern Hispaniola fault was “encouraged” by loading resulting from each previous large event. The results quantify the concept of stress building up on a fault over time, and that stresses were high on the Enriquillo fault prior to the January 12, 2010, earthquake.

#### 2.5.1.1.2.3.2.1 Septentrional Fault

The Septentrional fault extends approximately 600 kilometers (370 miles), from the Mona Passage on the east, to the northwestern tip of Hispaniola (Figure 2.5.1-202). On the west side it merges with the Oriente fault at about 74° W, where the plate boundary changes orientation from west-northwest to east-northeast. On the east side it has been observed to end in a circular depression about 25 kilometers (15.5 miles) west of Mona Canyon (Reference 582). This location is shown in Figure 2.5.1-320. As global plate motions were developed and those in the Caribbean became known (e.g., References 653 and 654), the importance of the Septentrional fault as a major plate boundary component was recognized (e.g., References 655 and 493).

Mann et al. (Reference 779) present results of paleoseismic and geomorphic studies of the Septentrional fault. As shown in Figure 2.5.1-321, they divide the fault into three sections, the western Septentrional fault system (identified as “western SFS” in Figure 2.5.1-321), the central Septentrional fault system, and the eastern Septentrional fault system. In the eastern area the fault parallels the southern shore of the Samana Peninsula, with the submarine trace of the fault lying about 2 kilometers (1.2 miles) south of the mountain front to the north (Reference 657). The central portion lies within the heavily populated Cibao Valley, and is marked by a 100-kilometer (62-mile) long trace on the valley floor. The scarp relief ranges from 1.1 to 11.3 meters (3.6 to 37 feet), with alternating facing directions. In the western section the fault bifurcates, with the southern section continuing through the western Cibao Valley and intersecting the coastline at the town of Pepillo Salcedo, and the northern section cutting through the northern Cordillera and intersecting the coastline at the town of Monte Cristi. The northern section is well exposed, but exhibits no evidence of Quaternary activity. The southern section, which is probably the more active trace (because it merges with the Oriente fault offshore to the west), is largely obscured due to recent fluvial sedimentation and erosion.

Early trenching studies near Santiago in the Cibao Valley (Reference 658) concluded that the most recent surface faulting event in this part of the fault occurred at least 430 years ago, as of 1993, and probably more than 730 years before 1993. Prentice et al. (Reference 658) estimate a slip rate of between 5 and 9 millimeters/year, based on estimates of the total plate boundary rate estimates at the time, which ranged from 12 to 37 millimeters/year (References 649 and 660, respectively). They conclude that about 3.5 meters (11.5 feet) of strain had accumulated on the fault since the last rupture.

Prentice et al. (Reference 570) present results and interpretations of all geomorphic and paleoseismic investigations of the Septentrional fault. They conclude that the slip rate on the central Septentrional fault is 6 to 12 millimeters/year, and that the last surface faulting event occurred about 800 years ago. This equates to strain accumulation between 5 and 10 meters, implying a potential earthquake in the magnitude 7.5 to 8 range (Reference 662).

Large historic events in northern Hispaniola occurred in 1564, 1783, 1842, 1887, and 1897 (Reference 571): all produced strong shaking in the Cibao Valley. The fact that all surface rupture identified by Prentice et al. (Reference 570) predated these events means that either: (a) any or all occurred on the Northern Hispaniola fault or unidentified structures in the northern Hispaniola region, or (b) any or all occurred on the Septentrional fault, but were deep enough not to produce surface rupture. The latter is a distinct possibility, given the lack of surface rupture during the recent highly destructive $M_w$ 7.0 Haiti earthquake (Reference 572) and the lack of knowledge regarding how deep the Septentrional fault extends.

#### 2.5.1.1.2.3.2.2 Northern Hispaniola Fault

The North Hispaniola fault is the south-dipping plate boundary between the North America Plate and the island of Hispaniola. The left-lateral strike-slip Septentrional fault forms the other component of this boundary, and is discussed in Subsection 2.5.1.1.2.3.2.1. The eastern boundary of the North Hispaniola fault coincides with the western end of the Puerto Rico Trench and the eastern boundary of the contact between the Bahama Platform and Hispaniola (Figure 2.5.1-322). The western boundary is not as clear, but appears to be between 73° W and 74° W (Figures 2.5.2-202 and 2.5.1-323), where it merges with the Nortecubana fault and ceases to function as the modern plate boundary.

Early results from GPS measurements indicated 21 ± 1 millimeters/year relative motion between southern Hispaniola and stable North America, about twice the estimate from global plate motion models (Reference 780). A southward decrease in velocities was noted, and combined with elastic strain models, results in estimates of 4.3 ± 3 millimeters/year on the North Hispaniola fault (Figure 2.5.1-324). Other estimates are 4 millimeters/year (Reference 663), 5.2 ± millimeters/year (Reference 358) (Figure 2.5.1-311), 12.8 millimeters/year (Reference 664), and 5 to 6 millimeters/year (Reference 643). The relative motion is highly oblique, almost parallel, to the North Hispaniola fault.

Because the Septentrional fault is estimated to slip at a rate of 6 to 12 millimeters/year (Reference 570), most of the North America–Hispaniola relative plate motion is taken up on that feature. Figure 2.5.1-325 shows a kinematic diagram of the relationship between the two structures.

A number of significant historical earthquakes have occurred on the North Hispaniola fault, including the 1943, 1946, 1953, and 2003 earthquakes (Figure 2.5.1-326). All had thrust mechanisms consistent with subduction of the North America Plate beneath Hispaniola. These earthquakes are described below.

The 1943 event has been studied by a number of authors, and magnitude estimates range from $M_S$ 7.5 to $M_S$ 7.8 (Phase 2 earthquake catalog $M_w$ 7.60). It has been associated with postulated high

friction on the North Hispaniola fault due to the presence of the Mona block, the subducted southeast portion of the Bahama Bank (Reference 591).

The August 4, 1946, earthquake ruptured an approximately 195 by 95 kilometer (121 by 69 mile) section of the Northern Hispaniola fault near northeastern Hispaniola (References 591 and 638) (Figure 2.5.1-326). Dolan and Wald's body-waveform inversions (Reference 591) yield a focal mechanism for the 1946 earthquake that indicates rupture occurred either on a shallowly south-dipping plane that strikes 085°, or a steeply northeast-dipping plane that strikes 110°. Dolan and Wald (Reference 591) prefer the shallowly south-dipping plane, which is consistent with subduction of the North America Plate. Magnitude estimates range from $M_s$ 7.8 (Reference 666) to $M_s$ 8.1 (Reference 665) (Phase 2 earthquake catalog $M_w$ 7.90). A tsunami generated by this event was responsible for about 100 deaths (Reference 857). An aftershock of approximately $M_s$ 7.3 in 1948 appears to extend the 1946 rupture zone downdip and to the northwest (Reference 638) (Figure 2.5.1-326). An additional earthquake of approximately $M_s$ 7.3 in 1953 (Phase 2 earthquake $M_w$ 6.93) extended the 1946 earthquake rupture zone to the northwest.

The rupture area of the 2003 $M_w$ 6.4 Puerto Plata earthquake (Phase 2 earthquake catalog $M_w$ 6.40) is shown in Figure 2.5.1-326 as the green area adjacent to the 1953 rupture area. The orange and blue stars and green circle denote epicentral locations from three different agencies. This event and its aftershocks were judged to have occurred on the North Hispaniola fault (Reference 638) (Figure 2.5.1-325).

Large destructive earthquakes in northern Hispaniola appear in the earlier historic record in 1564, 1842, and 1887. The 1564 event destroyed the towns of Santiago and La Vega (Reference 571). The causative structures of these earthquakes are not known, but the most likely candidates are the North Hispaniola fault and the Septentrional fault. The 1842 event has been associated with the Septentrional fault (Figure 2.5.1-327), but association with the North Hispaniola fault cannot be ruled out. This earthquake was probably in the magnitude 8 range (Phase 2 earthquake catalog $M_w$ 8.23), caused several thousand deaths, and generated a tsunami (References 571 and 666).

#### 2.5.1.1.2.3.2.3 Walton-Duanvale and Enriquillo-Plantain Garden Strike-Slip Fault System

The Walton-Duanvale, Plantain Garden, and Enriquillo faults are left-lateral strike-slip faults that collectively, from west to east, form the southern margin of the Cayman Trough and Gonâve microplate (Figure 2.5.1-202). The Walton fault extends for about 185 miles (300 kilometers) eastward from the southern end of the mid-Cayman spreading center to northwestern Jamaica (Reference 766). Slip is transferred from the Walton fault across the island of Jamaica through a broad restraining bend that includes the east-west striking Duanvale, Rio Minho-Crawle River, South Coast, and Plantain Garden faults (Reference 503). The Plantain Garden fault continues eastward and connects with the Enriquillo fault offshore southwestern Haiti. The Enriquillo-Plantain Garden strike-slip fault zone extends for about 375 miles (600 kilometers) from southeastern Jamaica to south-central Hispaniola and terminates eastward in the southern Dominican Republic east of Lake Enriquillo (Reference 383). There, slip apparently is transferred in a complex manner onto the Muertos Trough.

Several large earthquakes (magnitude 6.5 and greater) have struck the Port-au-Prince region of Haiti in the past (Table 2.5.2-221). These earthquakes are attributed to movement on the east-west oriented Enriquillo fault (Figure 2.5.2-214), a major tectonic element with a long history of deformation and slip (Subsection 2.5.2.4.4.3.2.10). The 1751 M 7.5 earthquake occurred near Port-au-Prince, Haiti, and the 1770 M 7.5 earthquake was located further to the west of Port-au-Prince on the Enriquillo fault. The Enriquillo fault ruptured again in a large earthquake still farther west in April 1860 (M 6.7) was accompanied by a tsunami.

On January 12, 2010, the $M_w$ 7.0 Haiti earthquake struck the Port-au-Prince region of Haiti causing significant damage and many casualties throughout the city. The earthquake epicenter set by USGS was 18.457° N, 72.533° W, which places the earthquake 25 kilometers (15 miles) west south-west of Port-au-Prince on or near the Enriquillo fault and 1125 kilometers (700 miles) southwest of Miami, Florida. Although the lack of local seismograph station data makes the precise earthquake location and depth somewhat uncertain, the focal depth has been estimated to be 13 kilometers (8 miles). The focal mechanism solution for the main shock indicates a left-lateral oblique-slip motion on an east-west oriented fault, which is consistent with the earthquakes that have occurred as left-lateral strike-slip faulting within the Enriquillo fault zone. This active fault accommodates a slip rate (and weights, as used in the estimate of its contribution to the PSHA of Subsection 2.5.2.4) of 6 [0.2], 8 [0.6], and 10 [0.2] millimeters/year, accounting for nearly half the overall movement between the Caribbean and North America plates (Table 2.5.2-217).

The USGS finite-fault model for the $M_w$ 7.0 Haiti earthquake (Reference 667) shows the surface area of the causative fault that ruptured is quite compact with a down-dip extent of approximately 20 kilometers (12 miles) and a length of approximately 50 kilometers (31 miles). Therefore, the implied fault-surface area available for seismogenic rupture of the causative fault is about 1000 kilometers$^2$. For an average slip rate of approximately 8 millimeters/year and assumed shear modulus of $3.0 \times 10^{11}$ dyne-cm, the rate of increase in the seismic moment ($M_0$) is about $2.4 \times 10^{24}$ dyne-cm/year. The seismic moment deficit that has accumulated since the 1770 earthquake (240 years) is $5.8 \times 10^{26}$ dyne-cm, equivalent to an unrelieved elastic strain that could release in a moment magnitude of about M 7.1 to 7.2 based on a standard moment-magnitude relation (Reference 668). Thus, the Port-au-Prince region of Haiti has a well-documented history of large earthquakes, and the historical pattern of earthquakes indicates that an earthquake of magnitude 7.0 or larger could strike southern Haiti near Port-au-Prince at any time.

The maximum magnitude (Mmax) probability distribution [and weights] for the Enriquillo fault for the PSHA described in Subsection 2.5.2.4 was considered to be $M_w$ 7.5 [0.2], 7.7 [0.6], and 7.9 [0.2] (Table 2.5.2-217). These values are based on rupture dimensions of about 120 to 250 kilometers (75 to 155 miles) long (from mapping described in Subsection 2.5.2.4.4.3.2.10) and 15 to about 18 kilometers (9 to about 11 miles) wide. Thus, the Mmax distribution used in the PSHA is comparable to the upper estimates of historical earthquakes attributed to this fault source. Based on these interpretations of magnitudes, the Mmax probability distribution was used to capture the uncertainty in the magnitude range of the largest historical earthquakes. The highest weight was given to $M_w$ 7.7 to support a source model whereby the Enriquillo-Plantain Garden strike-slip fault zone is fully coupled. This information shows that the 2010 $M_w$ 7.0 Haiti earthquake was expectable and completely within the magnitude and recurrence assessments incorporated in the PSHA.

#### 2.5.1.1.2.3.3 Muertos Trough and Mona Passage Extensional Zone

##### Muertos Trough

The Muertos Trough is a 300-mile (480-kilometer) long linear feature defined prominently in the bathymetry off the southern shores of the Dominican Republic and Puerto Rico (Figure 2.5.1-202) and a prominent north-dipping trend in seismicity (References 669 and 595). The structure accommodates underthrusting of the Caribbean Plate beneath the Puerto Rico microplate, which is situated between the Muertos Trough and the Puerto Rico-Northern Hispaniola subduction zone. The Muertos Trough ruptured in October 1751 in a great earthquake with an estimated magnitude 8.0 (Reference 596) (Phase 2 earthquake catalog $M_w$ 7.28).

##### Mona Passage

Mona Passage is the oceanic geographic feature that separates the islands of Puerto Rico and Hispaniola, and is about 150 kilometers 92-miles) east-west, and 50 kilometers (31 miles) north-south. The Mona Passage extensional zone (MPEZ) incorporates this oceanic part and

extends about 50 kilometers (31 miles) into eastern Hispaniola, and is thought to include the southwestern corner of Puerto Rico (Figure 2.5.1-210, sheet 2). Structurally, the MPEZ is part of the Puerto Rico-Virgin Islands microplate. Van Gestel et al. (Reference 670) describe it as a symmetric arch of the carbonate platform, with gently dipping north and south flanks superimposed by mainly north striking, but also northwest-southeast-striking, oriented normal faults.

Figure 2.5.1-328 shows a more detailed view of the bathymetry of the MPEZ. Three rift features can be seen: Mona Canyon on the north limb, and Yuma Basin and Cabo Rojo Rifts on the south limb. Mona Canyon was the site of a $M_w$ 7.2 earthquake in 1918 ($M_w$ 7.30 in the Phase 2 earthquake catalog) that caused severe ground shaking and a tsunami that affected northwest Puerto Rico. One hundred and sixteen deaths were recorded, and damage estimates approach $25,000,000 (Reference 671). Reid and Taber (Reference 672) postulated that Mona Canyon was associated with the earthquake. Tsunami modeling by McCann (Reference 671) successfully matched the observed effects on land with the rupture of a fault on the eastern wall of the canyon. Later studies (Reference 673) present evidence that the tsunami was caused by a landslide, located within the canyon, which was triggered by the earthquake.

Based on seven years of GPS measurements in Hispaniola and Puerto Rico, Calais et al. (Reference 358) measured 5 ± 3 millimeters/year of extension oriented east-northeast to west-southwest across the MPEZ. They postulate this was due to the impingement of Hispaniola on the Bahama Bank. The Puerto Rico-Virgin Islands microplate, being less impeded, moves to the east (relative to fixed North America) at a faster rate, which causes extension in the MPEZ. LaForge and McCann (Reference 577) identify 26 faults in the MPEZ, including two faults in western Puerto Rico (Cerro Goden and South Lajas faults), and two faults associated with Mona Canyon (Figure 2.5.1-320). Slip rates are estimated by projecting the above horizontal extension rate onto these faults. Mondziel et al. (Reference 585) estimate a late Pliocene to Present extension rate of 0.4 millimeters/year across Mona Canyon. Assuming a fault dip of 64°, this agrees with the LaForge and McCann (Reference 577) estimate of 0.92 millimeters/year. This equates to a return period of 1900 years for a $M_w$ 7.2 event, which is the largest magnitude postulated by Mondziel et al. (Reference 585) for the MPEZ faults.

The South Lajas fault in southwest Puerto Rico (Figure 2.5.1-320) is considered to be part of the MPEZ. Paleoseismic investigations reveal the occurrence of two surface-faulting events in the past 7000 years (References 570 and 578). LaForge and McCann (Reference 577) estimate a slip rate of 0.51 millimeters/year for this fault.

Mueller et al. (Reference 589) model the MPEZ hazard by uniformly distributing the 5 millimeters/year extension throughout the zone. Their maximum magnitude estimate for the MPEZ is $M_w$ 7.2 to 7.4.

The NPRSFZ and Bowin fault (Figure 2.5.1-320) are not well expressed in the seafloor bathymetry, but are considered to be strike-slip faults taking up some portion of the near arc-parallel plate motion (Reference 581). The NPRSFZ has a width of only a few kilometers due to its proximity to the Puerto Rico Trench (Reference 591), and therefore is not considered by LaForge and McCann (Reference 577) and Mueller et al. (Reference 589) to be a seismic source. The Bowin fault appears to be a possible eastward extension of the Septentrional fault. LaForge and McCann (Reference 577) and Mueller et al. (Reference 589) assign it a rate of 1 millimeters/year. LaForge and McCann (Reference 577) estimate a maximum magnitude of $M_w$ 7.3; Mueller et al. (Reference 589) estimate $M_w$ 7.6.

#### 2.5.1.1.2.3.4 Puerto Rico Trench

The Puerto Rico Trench is the bathymetric manifestation of the Puerto Rico subduction zone (PRSZ). The trench itself is an unusual feature, being the deepest point in the Atlantic Ocean (8+ kilometers

or 5+ miles deep) and exhibiting the highest negative free-air gravity anomaly on earth (Reference 675). With the advent of plate tectonics in the 1960s, it was recognized as a plate boundary, and early attempts were made to estimate the relative plate motion between Puerto Rico and the North America Plate (References 676 and 632).

Figure 2.5.1-322 shows the regional tectonic elements associated with the Puerto Rico Trench. The North America Plate is colliding with the Puerto Rico-Virgin Islands microplate at a highly oblique angle. Although relative plate motions were first deduced from global plate motion models, the deployment of GPS networks has permitted refined estimates. Recent GPS studies indicate a relative convergence between the North America Plate and the Puerto Rico microplate of 16.9 ± 1 millimeters/year in a west southwest direction (References 594 and 358). This is intermediate between values of 11 millimeters/year (Reference 649) and 37 millimeters/year (Reference 660), which were previously estimated on the bases of global plate motion models and the length of the downgoing slab, respectively.

Local seismograph networks have been operated in Puerto Rico since the mid-1970s. Early results (Reference 587) showed shallow seismicity beneath the island and a north-dipping plane of seismicity associated with the subducting North America Plate extending to depths of about 150 kilometers (90 miles) (Figure 2.5.1-309). Later studies confirm this pattern (Reference 573).

Seismicity on the deeper portion of the subduction zone is persistent. Shepherd et al. (Reference 677) list two magnitude 6 events that occurred between 64° W and 67° W, deeper than 50 kilometers (31 miles), from 1900 to 2001. An earthquake in 1844 (Phase 2 earthquake catalog $M_W$ 6.40) may have been a moderate to large intra-slab event, with moderate shaking and low-level damage reported uniformly throughout the central-eastern part of the island. The event was not reported to have been felt in Hispaniola, but was claimed to have been felt in St. Thomas and Guadeloupe (Reference 641). Based on the 1900 to 2001 observations, LaForge and McCann (Reference 577) estimate a return period of 84 years for $M_W$ 6.0 and above, with an upper bound of $M_W$ 7.5.

LaForge and McCann (Reference 577) distinguish between an eastern and western PRSZ based on the location of the impingement of the Bahama Bank on the trench at about 66.8° W longitude (Figure 2.5.1-310). This is due to denser seismicity in the western part, which likely results from resistance of the buoyant Bahama Bank to subduction and therefore tighter seismic coupling. LaForge and McCann (Reference 577) use 80 percent coupling in the western part and 20 percent in the eastern part. Manaker et al. (Reference 643) also find low coupling, less than 50 percent, for the plate interface in this region, based on an integrated GPS strain model for the northeast Caribbean. Grindlay et al. (Reference 679) attribute this low coupling to the old age of the subducting crust and to the collapse of the northern island platform due to passage of the Bahama Bank from east to west with accompanying tectonic erosion of the upper plate.

Between the Puerto Rico Trench and the north coast of the island, the general pattern of seismicity with depth less than 30 kilometers (19 miles) (thus likely to be associated with the plate interface) shows the heaviest activity near the Main Ridge, moderate activity west of the island, and sparsest activity north of the island (Reference 588). McCann (Reference 588) proposes that this feature is a non-buoyant structure on the downgoing North America Plate, which would function as a barrier to rupture propagation along the plate interface from the west (assuming the Bahama Bank intersection with the Puerto Rico Trench is the western barrier), thus limiting the size of maximum earthquakes affecting northern Puerto Rico to $M_W$ 8 to 8.4.

Doser et al. (Reference 681) present focal mechanisms for the larger historical events in this region (Figure 2.5.1-249). Figure 2.5.1-249 shows focal mechanisms for the region northwest and northeast of the island. With few exceptions these show the expected pattern of west southwest-directed thrust faulting, consistent with the relative plate motion. Slip vectors in Figure 2.5.1-249 show consistency

with this pattern, with some apparent partitioning between more northerly and more easterly vectors in the upper left of the figure. The lack of moderate-sized earthquakes directly north of the central and eastern shore of the island is illustrated in Figure 2.5.1-249.

Seismicity of $M_w < 7$ is abundant in the PRSZ, but only two events have exceeded $M_w$ 7 in the 500 year historical record. McCann (Reference 600) suggests that a $M_w$ 8 to 8.25 (Phase 2 earthquake catalog $M_w$ 8.03) interface event occurred on this segment in 1787, rupturing from roughly Mona Canyon on the west to the Main Ridge on the east. This event caused widespread damage on the island. In 1943, a magnitude 7.8 earthquake (Phase 2 earthquake catalog $M_w$ 7.60) ruptured an approximately 80-kilometer (50-mile) wide section of the subduction zone across Mona Canyon. On the basis of its focal mechanism, it is judged to have occurred on the shallow interface (Reference 591).

#### 2.5.1.1.2.3.5 Other Tectonic Elements

Other tectonic elements in northern North America-Caribbean Plate boundary region that are associated with seismicity and/or Cenozoic tectonics include: (a) the Nicaraguan Rise and Hess Escarpment, (b) the Beata Ridge, (c) the northern boundary of the Cayman Trough west of the mid-Cayman spreading center, and (d) the Yucatan Basin and the ancestral plate boundary zone/escarpment separating the Yucatan Basin from the Maya block/Yucatan Peninsula. All these features are probably capable of earthquakes of $M_w$ approximately 7, similar to the April 1941 $M_w$ 7 earthquake (Phase 2 earthquake catalog $M_w$ 7.03) on the Nicaraguan Rise southwest of Jamaica (References 641 and 640). However, all these features are distant from southern Florida (all greater than 200 miles or 320 kilometers), some greater than 400 miles (640 kilometers), all have low strain rates compared to the plate boundary faults (e.g., Reference 502), and all have comparable or lesser magnitude potential than plate boundary faults or source areas (i.e., Cuba) that are closer to Units 6 & 7 site.

### 2.5.1.1.3 Geologic Evolution of the Site Region and Beyond

The Units 6 & 7 site is located on the North America Plate, approximately 400 miles (640 kilometers) north of the Caribbean Plate boundary. This subsection provides an overview of the major tectonic and geologic events that occurred in the site region and beyond during the past few hundred million years, with an emphasis on those events that currently are expressed in the tectonic features and geology of the site region. Figure 2.5.1-329 presents a summary of these tectonic and geologic events in the evolution of the North America/Caribbean Plate boundary. Aside from buried Paleozoic and older basement rocks, the geology of the site region primarily comprises Mesozoic and younger transitional crust and overlying strata.

#### 2.5.1.1.3.1 Paleozoic Tectonic History

Because the site region (portions of the Florida Platform, Bahama Platform, and Cuba) was not a part of North America until the Permian Period, it has a different tectonic history. For completeness, the subsequent text includes a review of the tectonic history of southeastern North America, namely the events of the Appalachian orogenies, followed by a discussion of the evidence that exists for the tectonic history of the pre-Mesozoic basement of the site region.

**Southeastern North America**

Three primary mountain-building events affected the rocks of southeastern North America. The Taconic orogeny occurred in the Middle to Late Ordovician approximately 480 to 435 Ma as one or more terranes, perhaps microcontinents, and/or volcanic island arcs collided with the eastern margin of Laurentia via subduction on an east-dipping subduction zone. 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 (References 682 and 683). The Neoacadian, or the younger portion of the Acadian orogeny exhibited in the southern Appalachians, included unconformities in foreland strata, plutonism, limited migmatization, and some faulting (References 683 and 795). The final and most significant tectonic event of the southern Appalachians was the Alleghany orogeny, during which Gondwana (Africa, Florida Platform basement, and South America) collided with Laurentia, and the intervening Rheic Ocean was consumed in the Permian. This significantly shortened the previously-accreted terranes and translated them westward across the eastern margin of North America (Reference 683). Metamorphism, plutonism, and faulting in the Piedmont of the Carolinas and Georgia accompanied the uplift of the mountain range from New England to Alabama (Reference 683).

### The Florida Platform

The collision of Africa with North America during the late Paleozoic occurred along a buried Alleghanian suture known as the Suwannee suture, the Suwannee-Wiggins suture, or the South Georgia suture (Reference 342). This suture represents the boundary between the crust of Laurentia to the north and the crust of Africa or Gondwana to the south (Reference 684). Multiple versions of this roughly east-west striking structure have been hypothesized, all of which occur generally buried beneath the coastal plain of southern Alabama and Georgia subparallel to the Brunswick magnetic anomaly and just north of the elongate Triassic South Georgia Rift system (References 342, 344, and 685) (Figure 2.5.1-229).

Limited information is available regarding the Precambrian to Paleozoic evolution of the site region, located south of the Suwannee suture. The sources of this information are: (a) borings from the Suwannee terrane in central and northern Florida, 100 miles (160 kilometers) north of the site; (b) the allochthonous Socorro complex on northern Cuba, 200 miles (320 kilometers) south of the site; and (c) borings from transitional crust of the southeastern Gulf of Mexico, 250 miles (400 kilometers) southwest of the site. No Precambrian or Paleozoic components have been encountered in the few borings on the Bahama Platform.

Rocks drilled in the Suwannee terrane in central Florida include low-grade, felsic metavolcanics of the Osceola volcanic complex; the undeformed Osceola Granite; a suite of high-grade metamorphic rocks, such as gneiss and amphibolite, belonging to the St. Lucie complex; and a succession of generally undeformed Paleozoic sedimentary rocks. These units, described in detail in Subsection 2.5.1.1.1.2.1.1, indicate that the Suwannee terrane experienced Late Proterozoic to Early Carboniferous plutonism, volcanism, and high-grade metamorphism, followed by tectonic quiescence during the lower Ordovician through Devonian.

Using correlations with rocks in West Africa, it appears that the Gondwana-derived Florida Platform basement (the Suwannee terrane) experienced tectonism from 650 to 500 Ma, a quiescent Paleozoic, and only localized thrusting or metamorphism during the Late Paleozoic amalgamation of Pangea (References 686 and 337), in contrast to the significant Paleozoic deformation of multiple orogenies experienced in Laurentia (North American craton).

### The Gulf of Mexico and Northern Cuba

Two borings for the DSDP in the southeastern Gulf of Mexico encountered basement rock (Reference 687). The samples are located approximately 250 miles (400 kilometers) southwest of the Units 6 & 7 site. Sixty-six feet (thirty meters) of phyllite were found at the bottom of one boring, samples from which yielded whole rock $^{40}Ar/^{39}Ar$ ages of approximately 450 to 500 Ma. Samples from the other boring indicate that mylonitic gneiss and amphibolite were intruded by several generations of diabase dikes. Hornblende from the amphibolites yielded $^{40}Ar/^{39}Ar$ ages of approximately 500 Ma, while biotite from a gneiss yielded a 350 Ma age. The dikes have whole-rock ages that range from 190 Ma to 163 Ma (Reference 793). These results confirm that the transitional crust sampled in the basement of the southeastern Gulf of Mexico (and probably the southern Florida

Platform and western Bahamas) consists of pan-African crust intruded by Jurassic diabase (Reference 410).

In northeastern Cuba, the Socorro complex consists of a metasedimentary unit of marble, quartzite, and mylonitized granite. The entire complex occupies a very small area, and structural interpretations indicate that it was thrust northward along with the surrounding Cretaceous arc rocks. Biotite separated from the marbles yield an $^{40}Ar/^{39}Ar$ age of approximately 900 Ma with the plateau indicating a reheating event at 60 Ma. Two discordant conventional U-Pb zircon fractions from the granite yield a Jurassic lower intercept age and a 900 Ma upper intercept age. These data were interpreted to reflect a 900 Ma (possibly Grenville) metamorphic event and a Paleogene heating event that probably reflects the emplacement of the Greater Antilles volcanic arc. The 900 Ma age may indicate that the Socorro complex originated from the Yucatan, Chortis, or other Central American block (References 689 and 442) (Figure 2.5.1-206).

#### 2.5.1.1.3.2 Mesozoic Tectonic History

During the Mesozoic, major tectonic events occurred in the site region and beyond, including the opening of the Gulf of Mexico, the development of the Caribbean Plate, and the opening of the Atlantic Ocean. This subsection presents a generalized tectonic history of the site region and beyond, with emphasis on the Gulf of Mexico and north-central Caribbean region. In summary, the super-continent Pangea began to rift apart in the Late Triassic-Early Jurassic Periods, moving North America away from conjoined South America and Africa. This led to the widespread development of a series of Triassic-Jurassic rift basins along the eastern margin of Laurentia, from New England to southern Georgia (e.g., Reference 471). Both the Atlantic Ocean Basin and the Gulf of Mexico opened, and the continental crustal extension during this time is reflected in a few buried early Mesozoic normal faults within the site region.

It should be noted that details of the interpretation may vary from author to author, but this summary represents current general ideas. Iturralde-Vinent and Lidiak (Reference 690) and Giunta et al. (Reference 691) discuss current research directions for future clarification of Caribbean Plate tectonic history. For example, details of magmatic, metamorphic, and stratigraphic events suggest that the Great Antilles Arc may have comprised more than one arc.

The discussion presented here favors the “Pacific origin” reconstruction of Caribbean Plate tectonic history, which assumes that the plate originated in the present-day Pacific Ocean to the southwest of Central America, and migrated eastward to fill the gap between the diverging North and South America plates. An opposing “in situ” reconstruction has been presented (e.g., Reference 608), which postulates that the present Caribbean is the result of simple northwest-southeast extension between the North and South America plates. The “Pacific origin” reconstruction appears to be favored by most researchers at present, and thus is presented here.

#### Formation of the Gulf of Mexico

Figure 2.5.1-206, modified from Pindell and Kenan (Reference 696), shows proposed relations between North America, South America, and Africa in the early Jurassic. The Suwannee suture marks the closing of the proto-Atlantic ocean and collision of combined Africa and South America (Figures 2.5.1-205 and 2.5.1-204). To the north this suture is paralleled by the Alleghanian deformation front, which occurred from middle Carboniferous (Late Mississippian) through early Permian time (Figure 2.5.1-204). The Yucatan block lay in what is now the Gulf of Mexico, and the Chortis block is seen at the southern tip of what is now North America. Note that some reconstructions interpret a portion of southern Florida originated to the west, in what is now the eastern Gulf of Mexico, moving to its current position via a hypothetical Jurassic transform (such as the Bahamas Fracture Zone) (References 460 and 212). However, more recent reconstructions of western Pangea indicate a continuous rift system connecting the Atlantic and Gulf of Mexico, but located south of the southern margin of the Florida and Bahama Platforms (Reference 692).

At about 165 Ma, North America began to separate from (the combined) South America and Africa. Spreading ridges and new oceanic crust began to form between the eastern U.S. seaboard and western Africa, and between the Gulf Coast and Mexico and northwestern South America.

Throughout the late Jurassic and early Cretaceous, the Yucatan block became detached from North and South America, isolated by a seaway to the southwest, the Proto-Caribbean Seaway to the southeast, and the Gulf of Mexico to the north (Figure 2.5.1-206). The latter is the only one of these in existence today. Also at this time, the Bahama Platform, a region of thicker-than-oceanic crust upon which an up to 10-kilometer (6.2-mile) thick limestone (carbonate) platform developed. This feature has behaved in a continental-like manner in term of its resistance to subduction. The opening of the Gulf of Mexico and counterclockwise rotation of the Yucatan block occurred, about a pole of rotation near south Florida (Figure 2.5.1-206).

A detailed chronology of the opening of the Gulf of Mexico is presented by Bird et al. (Reference 511). The majority of the development of the Gulf of Mexico rifting began at 160 Ma. Bird et al. (Reference 511) estimate 42° of counterclockwise rotation, over a period of 20 m.y., consistent with estimates from other workers, and a completion of the formation of the Gulf of Mexico by 140 Ma (lower early Cretaceous). This includes the following major events:

1. Development of a terrestrial rift valley between the Yucatan block and the present Gulf Coast.

2. Periodic flooding of seawater from adjacent seaways, leaving behind massive salt deposits. This occurred before the oceanic spreading center developed.

3. Initiation of ocean-floor spreading. Development of a plume near the center of the rift zone left behind high-density rocks on either side of the spreading ridge. These features are not visible on the ocean floor, but are readily seen on gravity anomaly maps. The salt deposits were separated by the spreading ridge, into the Sigsbee salt to the north, and the Campeche salt to the south. Both deposits have since flowed toward the center of the Gulf of Mexico.

4. Contemporaneous with event 3, the East Mexican transform fault developed along the coast of Mexico to accommodate the asymmetric opening of the Gulf of Mexico.

The Atlantic Ocean Basin begun to open at an earlier date, moving Africa (Gondwana) away from Laurentia via the same rift system that continued into the Gulf of Mexico. Following widespread continental rifting, seafloor spreading began in the Atlantic Ocean at ~185 Ma (Reference 421). The most detailed mapping of the seafloor indicates that oceanic crust is located northeast of the Bahama Platform and east of the Blake Spur magnetic anomaly, though its been speculated that oceanic crust may exist as far west as the East Coast magnetic anomaly north of latitude approximately 29° N (Reference 466). Spreading centers may have led to the development of the East Coast magnetic anomaly and the Blake Spur magnetic anomaly (e.g., Reference 409).

The majority of the above outlined tectonic events occurred well outside the site region during the Triassic and Jurassic. However, evidence of the regional tectonic history is seen in the two primary phenomena on the Florida and Bahama Platforms: (a) the intrusion of late Triassic and early Jurassic rift-related magmatic products, such as basalts and rhyolites (e.g., Reference 463), and (b) the existence of normal faults buried by Cretaceous and younger strata (e.g., Reference 307). Volcanics sampled in subsurface southern Florida indicate a mantle-derived source (Reference 694), and those in central and northern Florida share characteristics with rocks in the Gondwana-derived Carolina terrane (Reference 695). The compaction and differential subsidence of the sedimentary section deposited over a faulted basement topography led to the development of arches and lows on the Florida Platform, such as the Peninsular Arch and the Sarasota Arch (Reference 413, see Subsection 2.5.1.1.1.3.2.1). Extensional thinning of the basement and intrusion of rift-related magma led to the 'transitional' nature of the crust beneath the Florida and Bahama Platforms. The deposition

of thick Cretaceous carbonates on the Florida and Bahama Platforms and in parts of Cuba indicate that the site region was a slowly subsiding shallow carbonate platform by that time, and that these rift-related tectonic events had ceased (References 441 and 413).

### The Cretaceous Greater Antilles Arc

During the Late Cretaceous, the approach of the Greater Antilles Arc collided with the Florida and Bahama Platforms. This process eventually led to the transfer of material now found on Cuba from the Caribbean to the North America Plate.

In the Early Aptian, North and South America drifted farther apart (Figure 2.5.1-206). The eastern Bahama Platform is postulated to be created by a “leaky” transform fault or “hot spot” along the proto-mid-Atlantic Ridge (Reference 696). The creation of the Gulf of Mexico and accretion of the Yucatan block to North America are complete.

During the late Cretaceous, the Greater Antilles Arc was initiated, and the Caribbean plate developed. In detail:

1. In the Aptian, a central section of subducting Farallon Plate between North and South America switched polarity (the dip of subducted oceanic crust switched to west), and started moving into the proto-Caribbean Seaway. The eastern boundary of the new Caribbean Plate was the Greater Antilles Arc (Figure 2.5.1-206).

2. Behind it, the Central American Arc was initiated, accommodating most of the Farallon-North America Plate motion. The new Caribbean Plate lay between the two arcs, as it does today (Figure 2.5.1-206).

3. Collision of the Greater Antilles Arc with southeast Yucatan accreted continental material to the western Greater Antilles Arc, indicated by detrital mica ages from western Cuba (Reference 442) (Figures 2.5.1-206 and 2.5.1-250).

After the Greater Antilles Arc collided with the Yucatan Platform, it moved northward into the proto-Caribbean seaway and approached the Bahama Platform (Figure 2.5.1-250). The southeast Yucatan margin was then a strike-slip boundary, and at the Paleocene (60 Ma), the arc collided with the Bahama Platform. The Yucatan Basin formed behind the Greater Antilles Arc, and the Chortis block (Figure 2.5.1-206) moved to the east, forming present day southern Guatemala, Honduras, and northern Nicaragua.

#### 2.5.1.1.3.3 Tertiary Tectonic History

### Paleocene - Miocene Greater Antilles Arc Collision with Bahama Platform

By early Cenozoic time, the Greater Antilles Arc had moved northwards, consuming proto-Caribbean oceanic crust along the way, far enough to make contact with the Bahama Platform, which initiated an arc-continent collision. This event led to the following effects, which gave rise to the configuration of the present-day northern Caribbean:

1. Greater Antilles Arc collided with Bahama Platform. In western Cuba, the Greater Antilles Arc reached the Bahama Platform, and because the Bahama Platform crust was buoyant, subduction ceased at the northwestern end of the arc (Reference 697). As a result, arc volcanism ceased, ophiolites and Bahama Platform carbonates were obducted and transferred northward, and the arc massif collapsed and began to erode. The collision initiated in western Cuba in late Paleocene to early Eocene time, and continued eastward (Reference 220). The Cretaceous arc and ophiolite sequence now exposed on Cuba was

thrust onto Cuba and transported northward over the Bahama Platform along the Domingo thrust fault and other structures (Reference 439).

2. Rotation of Caribbean Plate movement to the east. As further northward movement of the plate was impeded, the direction of movement rotated to the east where less resistance was encountered; i.e., the North America oceanic plate to the east of the Bahama Platform. The change in plate direction resulted in the creation of a set of left-lateral strike-slip faults that progressively shifted position from northwest to southeast (References 219 and 697). These are, respectively, the Pinar, La Trocha, Camaguey, and Nipe faults (Figure 2.5.1-247). These faults represent paleo-plate boundaries as the landmass on the northwest side became attached to the North America Plate. This process occurred over a period of 40 m.y., from Late Paleocene (60 Ma) to early Miocene (20 Ma) (Figure 2.5.1-250). The last, and currently active, fault in the progression is the Oriente fault.

Within the site region, the effects of the collision of the Greater Antilles arc with the Bahama Platform led to the development of faulting in northern Cuba and in the Straits of Florida during the Eocene time. The Walkers Cay fault and Santaren anticline were also active at this time, and deformation or later reactivation may have occurred in the Miocene on all of these structures and may have continued into the Quaternary on the Walkers Cay, Santaren anticline, and faults in Cuba.

### Opening of the Cayman Spreading Center and Trough

The Cayman Trough is postulated to have initiated from a pull-apart basin at about 49 Ma, or early Eocene (Reference 499). This is the result of the left-step in the left-lateral North America-Caribbean Plate boundary (Reference 655). Magnetic stripes on the ocean floor indicate that the Cayman Trough began generating oceanic crust at approximately 44 Ma (References 460 and 222). This signified the end of significant Caribbean-North America Plate boundary motion being accommodated in northern and central Cuba. A Paleogene arc along the Cayman Ridge (a submarine ridge parallel to and north of the Cayman Trough) has been postulated by Sigurdsson et al. (Reference 299) to be the remnants of a volcanic arc that would have formed behind the Greater Antilles Arc, after its passage to the north. However, arguments against this interpretation have been presented by Leroy et al. (Reference 499).

By expansion of the Cayman Trough, development of the Oriente fault as the northern bounding transform fault, the Walton fault as the southeastern bounding fault, and the Swan Islands fault as the southwestern bounding fault occurred. According to Leroy et al. (Reference 499), Hispaniola was connected to Cuba until about early Oligocene (30 m.y.), after which motion on the Oriente fault separated the two. Detailed analysis of depth and age relations of the magnetic reversals in the Cayman Trough indicates that it was spreading at 20 to 30 millimeters/year before 26 Ma, but spreading slowed to a rate of less than 15 millimeters per year after 26 Ma (Reference 222). Sometime after the Late Miocene the Oriente fault connected with the Septentrional fault of northern Hispaniola, which developed as a result of highly oblique left-lateral North America-Caribbean Plate motion. After the Late Miocene, the Enriquillo-Plantain Garden fault extended eastward into southern Hispaniola. It now extends halfway into that island, where it terminates against the Beata Ridge and Muertos subduction zone. Sykes et al. (Reference 660) suggest that spreading rates in the Cayman Trough slowed at 2.4 Ma because the Enriquillo-Plantain Garden fault system became the new plate boundary (References 699 and 655).

### Convergence between North America and South America Plates

Sometime in the mid-Tertiary (probably early Miocene, about 15 Ma [Reference 593]) minor north-south convergence between the North America and South America Plates initiated. Evidence for this includes northward migration of the Beata Ridge and development of the Muertos subduction zone. The Beata Ridge is now “docked” onto south-central Hispaniola, and coincides with the eastern

termination of the Enriquillo-Plantain Garden fault and western termination of the Muertos subduction zone.

During the Miocene, clastic sediments of the Hawthorn group were deposited over parts of the Florida Platform, while southernmost Florida and the Bahama Platform continued to be sites of carbonate deposition (References 368 and 393). The deposition of the Hawthorn clastics is important because it reflects that the channels connecting the Gulf of Mexico and Atlantic Ocean that formally prevented the progradation of Appalachian-derived clastics were finally overwhelmed by Appalachian-derived sediment (References 257, 234, 473, and 396). These channels were located in northern Florida and southern Georgia, and existed as the Suwannee Straits in the late Cretaceous to Paleocene and Gulf Trough during the Eocene to Oligocene (References 257, 234, and 473). This Miocene phenomenon, progradation of clastics across the Florida Platform after the Suwannee channel system was filled, was possibly influenced by increased clastic supply from the Appalachians and increased flow through the Straits of Florida (References 221 and 396).

#### 2.5.1.1.3.4 Quaternary Tectonic History

Within the site region, the Quaternary Period is characterized by sedimentary deposition in both marine and terrestrial environments. On the Florida Platform, the Pleistocene Anastasia Formation and the Miami Limestone were deposited. The Miami Limestone grades into the Key Largo Limestone, which is a shallow shelf-margin coral reef deposit. Within the submerged areas of the Straits of Florida and the Bahama Banks, Neogene sedimentation is dominated by basinal carbonates and slope deposits of peri-platform oozes intercalcated with turbidites and often controlled by ocean current activity and sea level changes (Reference 228).

Faults within the Straits of Florida, the Santaren anticline, the Walkers Cay fault, and faults in Cuba were all active in the Tertiary (Figure 2.5.1-229). The Santaren anticline, the Walkers Cay fault, and faults in Cuba may have experienced continued tectonic activity into the Quaternary period.

Present day tectonic features of the northern Caribbean region are shown in (Figure 2.5.1-202). The Nortecubana fault system, sometimes interpreted as a suture between the northwestern Caribbean Plate and the North America Plate, is more aptly described as the fold-and-thrust belt from the collision. The collision-and-suture process proceeded from northwest to southeast, beginning at 60 Ma and ending at 40 Ma. The portions of Cuba within the site region are far (>300 kilometers) from the active plate boundary between the Caribbean and North American plates and exhibit low to moderate seismicity rates.

West of 71° W, the Cayman Trough separates the current Caribbean-North America Plate boundary into two subparallel, predominantly left-lateral strike-slip features, the Oriente and Septentrional faults on the north, and the Swan Islands-Walton-Duanvale-Enriquillo-Plantain Garden fault system on the south. These accommodate a relative plate motion of about 20 millimeters/year, which appears to be about equally divided between the two features (References 358, 652, and 643).

#### 2.5.1.1.4 Contemporary Stress Regime

Three types of forces are generally responsible for the stress in the lithosphere:

- Gravitational body forces or buoyancy forces, such as the ridge-push force resulting from hot, positively buoyant young oceanic lithosphere near the ridge against the older, colder, less buoyant lithosphere away from the ridge (Reference 700). This force is transmitted by the elastic strength of the lithosphere into the continental interior.

- Shear and compressive stresses transmitted across plate boundaries (such as strike-slip faults or subduction zones).

- Shear tractions acting on the base of the lithosphere from relative flow of the underlying asthenospheric mantle.

### Contemporary Stress Regime within the Site Region

The Earth Science Teams (ESTs) that participated in the EPRI (Reference 456) evaluation of intra-plate stress concluded that tectonic stress in the CEUS region is primarily 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 is ridge-push force associated with the mid-Atlantic Ridge, transmitted to the interior of the North America Plate by the elastic strength of the lithosphere. Other potential forces acting on the North America Plate were judged to be less significant in contributing to the magnitude and orientation of the maximum compressive principal stress.

In general, the ESTs focused on evaluating the state of stress in the mid-continent and Atlantic seaboard regions, for which stress indicator data were relatively abundant. Fewer stress indicator data were available for the Gulf of Mexico, Gulf Coastal Plain, and Florida Peninsula, and thus these areas received less scrutiny in the EPRI (Reference 456) studies. Since 1986, an international effort to collate and evaluate stress indicator data culminated in publication of a new *World Stress Map* (Reference 702) that has been periodically updated (Reference 703). Plate-scale trends in the orientations of principal stresses were assessed qualitatively based on analysis of high-quality data (Reference 704), and previous delineations of regional stress provinces were refined (Reference 705). Statistical analyses of stress indicators confirmed that the trajectory of the maximum principal compressive stress is uniform across broad continental regions at a high level of confidence (Reference 706). 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 orientation of compressive forces acting on the North America Plate from the mid-Atlantic Ridge (Reference 704).

According to the continental United States stress map of Zoback and Zoback (Reference 707), most of the CEUS is in the mid-plate stress province, which displays a consistent northeast-southwest maximum compressive stress orientation. However, coastal regions of Texas, Louisiana, Mississippi, Alabama, and northwestern Florida can exhibit down-to-the-gulf growth faulting. Hence, this area has been designated as the Gulf Coast stress province, characterized by northeast-southwest to north-northeast to south-southwest horizontal tension (Reference 702). The boundary between the mid-plate and Gulf Coast stress provinces terminates in the northern Florida Peninsula, but there is a lack of stress data from areas near the Florida Peninsula and most of Cuba. Because the southern Florida Peninsula doesn't exhibit the geologic features (such as salt-rooted normal faults) associated with the Gulf Coast stress province, the site region is generally interpreted to be part of the mid-plate stress province (Reference 705) (Figure 2.5.1-330).

The mid-plate stress province may exhibit reverse or strike-slip faulting under east-northeast- to west-southwest- to northwest-southeast-oriented compressive stress. This region extends from an approximately north-south-oriented line through Texas, Colorado, Wyoming, and Montana to the Atlantic margin and potentially into the Atlantic Ocean Basin. Within this province, the orientation of maximum compressive stress is generally parallel to plate velocity direction (Reference 702). Richardson and Reding (Reference 646) conclude that the observed northeast-southwest trend of principal stress in the mid-plate stress province of the central and eastern United States (CEUS) dominantly reflects ridge-push body forces associated with the mid-Atlantic Ridge. They estimated the magnitude of these forces to be approximately $2x10^{12}$ to $3x10^{12}$ N/m (Newton/meter), or 2.9 to 4.4x10 pounds per square inch (psi), (i.e., the total vertically integrated force acting on a column of lithosphere 3.28 feet wide), which corresponds to average equivalent stresses of approximately 40 to 60 MPa (megaPascals), or 5800 to 8700 psi, distributed across a 30-mile (~48-kilometer) thick elastic plate. Humphreys and Coblentz (Reference 647) evaluated the contribution of shear tractions on the base of the North American lithosphere to intra-continental stress and conclude 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 Ocean Basin.

Research on the state of stress in the continental United States since publication of the EPRI (Reference 456) studies confirms observations that stress in the CEUS is characterized by relatively uniform northeast-southwest compression. Few new data have been reported since the EPRI (Reference 456) study that better determine the orientations and relative magnitudes of the principal stresses in the site region. Given that the current interpretation of the orientation of principal stress is similar to that adopted in EPRI (Reference 456) a reevaluation 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 site region and site area 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. The mid-plate stress province is the most likely characterization of the tectonic stress at the site region and site area (Figure 2.5.1-330).

### Contemporary Stress Regime in the North America-Caribbean Plate Boundary Region

The Caribbean Plate is presently moving relative to the North America Plate at a rate of approximately 18 to 20 millimeters/year along an azimuth of roughly 075° (References 502, 635, and 636). In the Cuba and Caribbean-North America Plate boundary region, the relative plate motion is accommodated by the mid-Cayman spreading center and several subvertical, left-lateral transform faults extending from offshore of the northern coast of Honduras eastward through the Cayman Trough and through the islands of Jamaica and Hispaniola. The Cayman spreading center itself is located southwest of the Cayman Islands and is characterized by a north-south-trending axis of spreading with an average rate of approximately 15 millimeters/year since approximately 25 to 30 Ma (Reference 222). West of the Cayman Trough, Caribbean-North America Plate motion is accommodated offshore on the left-lateral Swan Islands fault (Figure 2.5.1-202). East of the Cayman Trough, on Hispaniola, the orientation of the plate-bounding structures changes and motion is partitioned between strike-slip faults (e.g., Septentrional and Enriquillo faults), minor oblique-reverse faults, and subduction on low-angle thrust faults (e.g., Northern Hispaniola thrust fault) (References 637, 638, and 639). East of Hispaniola, the Caribbean-North America Plate boundary becomes an oblique subduction zone or zones at the Puerto Rico Trench and Muertos Trough, and finally a more pure dip-slip west-dipping subduction zone in the Lesser Antilles. At the longitude of Puerto Rico, south-dipping subduction of North American crust at the Puerto Rico Trench and north-dipping subduction of Caribbean crust at the Muertos Trough accommodate relative plate motion with a highly oblique sense of shear (Figure 2.5.2-214).

Hypocenters and focal mechanisms of historical earthquakes provide information on fault geometry, crustal thickness, and fault kinematics throughout the Cuba and northern North America-Caribbean Plate boundary region. The kinematics of crustal deformation and faulting in Cuba are poorly understood. Geodetic data show that the current plate boundary is mostly south of Cuba along the Oriente and Enriquillo-Plantain Garden faults and that modern deformation rates across Cuba are likely <0.1 inch (3 millimeters) per year relative to North America (References 502 and 503). The Cayman Trough and western Hispaniola are characterized by shallow (crustal) seismicity with most focal mechanisms consistent with left-lateral strike-slip on east-west striking vertical faults (References 560, 632, and 640). Shallow seismicity in Jamaica is associated with strike-slip, oblique, and reverse focal mechanisms accommodating left-lateral shear across east-west and west-northwest-striking faults (Reference 503). Seismicity along the Oriente fault changes along strike south of Cuba as focal mechanisms show strike-slip and oblique-normal movement near the Cabo Cruz Basin and strike-slip to north-dipping reverse motion trends associated with the Santiago deformed belt (Reference 504). Shallow- to intermediate-depth seismicity defines south-dipping planes associated with the Northern Hispaniola fault and Puerto Rico Trench, with shallow focal mechanisms consistent with underthrusting of North American crust beneath the Caribbean on gently

dipping planes (Reference 591). A north-dipping zone of seismicity from shallow to intermediate crustal depths is associated with the Muertos Trough (Reference 591).

#### 2.5.1.1.5 Tsunami Geologic Hazard Assessment

This subsection provides geologic support for discussions in Subsection 2.4.6, Probable Maximum Tsunami Hazards. An extensive review of available scientific literature produced no positive evidence for Quaternary seismically induced or landslide-generated tsunami deposits within the 200-mile radius of the Units 6 & 7 site region. The literature does provide sedimentary evidence for Pliocene to Recent submarine mass movements on the Florida-Hatteras shelf and elsewhere in the western Atlantic (e.g., References 315, 316, 317, and 318). Literature also provides sedimentary evidence for Neogene and older submarine mass movements in the Florida Straits, the Bahama Platform, the northern coast of Cuba, the southeastern Gulf of Mexico, and the Yucatan Basin (References 422, 476, 727, 738, 740, and 742). Extensive geologic and historic literature is available documenting tsunami events and submarine mass movement in the Caribbean Basin (e.g., References 582, 681, 737, 738, and 739). The sedimentary and historic observation records that support a tsunami geologic hazard assessment are discussed in the following subsection.

According to Tappin (Reference 729), all forms of submarine mass movements are potentially tsunamigenic, yet there is a paucity of data relating submarine failures to tsunami generation and the physics of the process is still not well understood. Extensive retreat of the escarpments defining the edges of the Yucatan and Bahama carbonate platforms have been proposed by numerous researchers (e.g., References 305, 308, 794, and 726). However, the relevance of the processes of carbonate platform retreat to the Turkey Point site cannot be established because no stratigraphic evidence has been found to link escarpment retreat to any tsunami-like events along the southern coast of Florida. Wide-spread evidence for Miocene gravity flows in channels and troughs of the Bahama Platform has been documented (e.g., References 422, 727, and 728). Again, the relevance of seismically-induced Miocene gravity flows to the Turkey Point site region has not been established because no stratigraphic evidence has been found to link gravity flow deposits on the Bahama Platform to any tsunami-like events along the southern coast of Florida. Submarine landslides and the associated volumes of displaced seawater, whether triggered by a seismic event or another type of event (e.g., meteorite, volcanic activity, gravitation loading, or gas hydrate decomposition) are the likely cause of most Caribbean tsunamis and resulting tsunami deposits. The scientific literature does not address a means of discriminating between seismically-induced tsunami deposits and non-seismic tsunami deposits, even when there is a close relationship in time between a seismic event and a tsunami. The following describes the known characteristics of tsunami deposits, how they would be distinguished from hurricane deposits, and identifies possible locations conducive to deposition and preservation of tsunami deposits in the Turkey Point site region.

Tuttle et al. (Reference 889) distinguish tsunami from storm surge deposits, based on a comparison of deposits from the 1929 Grand Banks tsunami and the 1991 Halloween storm. The 1929 Grand Banks tsunami was caused by an earthquake-generated landslide that left chaotic deposits on the Burin Peninsula of Newfoundland. The Halloween storm caused sand and pebble overwash of barrier beaches and seawalls, and extensive damage along the New England coast, including Martha's Vineyard off the southern coast of Massachusetts. By 2004, researchers also had closely examined the character and extent of tsunami deposits generated by the 1755 Lisbon earthquake and the 1960 Chilean earthquake. As noted by Tuttle et al. (Reference 889), the challenge of discriminating between the two types of deposits was that both tsunami and storm surge processes result in the onshore transport and re-deposition of sediments. Tuttle et al. (Reference 889) conclude that four discriminators (included verbatim below) could be used to distinguish between tsunami and storm deposits:

1. Tsunami deposits exhibit sedimentary characteristics consistent with landward transport and deposition of sediment by only a few energetic surges, under turbulent and/or laminar flow

conditions, over a period of minutes to hours; whereas characteristics of storm deposits are consistent with landward transport and deposition of sediment by many more, less energetic surges, under primarily laminar flow conditions, during a period of hours to days.

2. Both tsunami and storm deposits contain mixtures of diatoms indicative of an offshore or bayward source, but tsunami deposits are more likely to contain broken valves and benthic marine diatoms.

3. Biostratigraphic assemblages of sections in which tsunami deposits occur are likely to indicate abrupt and long-lasting changes to the ecosystem coincident with tsunami inundations.

4. Tsunami deposits occur in landscape positions, including landward of tidal ponds, that are not expected for storm deposits.

Similarly, Morton et al. (Reference 890) characterize the distinction between tsunami and storm deposits as being related to differences in the hydrodynamics and sediment-sorting processes during transport. Tsunami deposition results from a few high-velocity, long-period waves that entrain sediment from the shoreface, beach, and landward erosion zone. Tsunamis can have flow depths greater than 10 meters (33 feet), transport sediment primarily in suspension, and distribute the load over a broad region where sediment falls out of suspension when flow decelerates. In contrast, storm inundation generally is gradual and prolonged, consisting of many waves that erode beaches and dunes with no significant overland return flow until after the main flooding. Storm flow depths are commonly <3 meters (9.8 feet), sediment is transported primarily as bed load by traction, and the load is deposited within a zone relatively close to the beach (Reference 890). A schematic of typical tsunami and storm deposits is shown in Figure 2.5.1-348.

As noted by Dawson and Stewart (Reference 891), hurricane deposits are quite different from tsunami deposits. For example, Scoffin and Hendry (Reference 892) use coral rubble stratigraphy on Jamaican reefs to identify past hurricane activity, while Perry (Reference 893) use storm-induced coral rubble in reef facies from Barbados to identify episodes of past hurricane activity. Similarly, in coastal Alabama, a series of hurricanes during historical time resulted in the deposition of multiple sand layers in low-lying coastal wetlands, but never as extensive as tsunami sediment sheets. By contrast, the overwash fans along the New England coastline used by Donnelly et al. (Reference 894) to reconstruct a 700-year record of hurricane activity have analogues with similar tsunami deposited fans (References 895, 896, and 897).

A literature review indicates that storm surges result in the deposition of discontinuous sedimentary units and that tsunamis, in contrast to storm surges, generally result in deposition of sediment sheets, often continuous over relatively wide areas (Figure 2.5.1-348). The tsunami units are also deposited a considerable distances inland. For example, sediment sheets produced by the 1755 tsunami in Algarve, Portugal occur in excess of 1 kilometer (0.6 mile) inland (Reference 898).

Part of the difficulty in understanding the characteristics of tsunami deposits is due to a lack of knowledge of offshore hydraulics and sediment dynamics during modern tsunami events. Whereas the majority of the literature concerned with tsunami sedimentation has focused attention on the coastal zone, relatively little attention has been given to tsunami depositional processes both in the near-shore and offshore (Reference 891). This is because tsunami sediments are readily identifiable and easy to study along coastlines. But each incoming tsunami wave is associated with strong backwash flow from the coast into the sea. This highlights the strong possibility that sediments picked up by tsunamis may also drape the sea floor, a consequence of the cumulative depositional effect of each backwash flow associated with the train of tsunami waves (Reference 899). During this phase, pulses of tsunami backwash may generate turbidity currents that move seaward towards the abyssal zone via submarine gullies and canyons (Reference 900).

Current tsunami research is focused on identifying potential onshore and offshore tsunami deposits as well as discriminating between types of tsunami deposits. Onshore deposits from submarine landslide-generated tsunamis probably could not be discriminated from earthquake-generated tsunamis except by careful radiometric age correlations between causative events, such as has been done with the pre-historic Storegga landslide (Reference 901). Clearly, bolide impact deposits are often associated with inclusion of impact ejecta (impact debris), including shocked minerals and impact spherules, depending on the proximity of the impact site to the deposits (References 320, 902, and 903). Volcanic collapse-generated tsunami deposits generally include a significant component of airfall tephra (Reference 731).

The boring logs from the subsurface investigations of the Units 6 & 7 site are described in detail in Subsection 2.5.1.2. The logs indicate that geologic conditions are uniform across the site (Figure 2.5.1-232) and show no evidence of interruption by a tsunami-like event. The vegetated depressions and surficial drainage channel areas were targeted with inclined borings and surficial soils sediment (muck) sampling during the supplemental field investigation (References 995 and 996). This supplemental program was an effort to further examine the characteristics of these surficial deposits. The sediment record described in surficial samples from the site provided no direct evidence for deposits or sedimentary structures that could be interpreted as evidence for high-energy depositional events (e.g., hurricane or tsunami landfalls). That is, no storm beds, tsunamigenic deposits (upward fining clastic sequences), peaks in sand content (sand sheets), or erosive surfaces were identified in any borings at the site.

The following describes available evidence for potential landslide tsunami sources along the southeastern Atlantic margin of North America, the western edge of the Florida Escarpment, the eastern edge of the Blake Plateau, the eastern edge of the Bahama Platform, across the Straits of Florida, and along the northern coast of Cuba. The most likely source for an earthquake-generated tsunami that is close enough to possibly affect the Units 6 & 7 site is the Puerto Rico Trench. Modeling of that source is discussed later in this subsection.

### Potential Central or Western Atlantic Tsunami Sources

In the early 2000s, several professional publications (References 730 and 731) predicted the effects of a mega-tsunamis caused by underwater volcanic edifice collapse in the Canary Islands. As a result, U.S. and Caribbean scientists began reexamining physiographic evidence for large, undersea landslides in the detailed topographic data from the GLORIA side-scan imagery and other remote sensing data and submersible observations within the U.S. Exclusive Economic Zone (EEZ) and in the greater Caribbean region (Reference 732).

As noted in Subsection 2.5.1.1, megasedimentary events are recognized throughout the world's ocean basins. Pilkey (Reference 315) identifies a “megaturbidite” that he calls the Black Shell Turbidite in the Hatteras Abyssal Plain, north of the Blake Plateau. A “megaturbidite” is an underwater landslide that moves great volumes of sedimentary material downslope, in the process displacing large volumes of water and likely causing a tsunami at the water surface. The Black Shell Turbidite is at least 100 kilometers$^3$ (24 miles$^3$) in volume and perhaps double that. Based on radiocarbon dates of the youngest shell fragments incorporated by the megaturbidite, the event occurred about 16,900 years ago (Reference 316). According to Pilkey (Reference 315), major events such as the occurrence of the Black Shell Turbidite should not be assumed to be restricted to times of lowered sea level. The 1929 Grand Banks submarine landslide (Reference 317) may have involved as much as 400 kilometers$^3$ (96 miles$^3$) of material. This slump resulted in a turbidity current that traveled 500 kilometers (311 miles) to the Sohm Abyssal Plain, but the full areal extent of the resulting turbidite is unknown (Reference 315). The tsunami waves associated with the Grand Banks landslide had amplitudes of 3 to 8 meters (10 to 26 feet) and run-up of up to 13 meters (43 feet) along the Burin Peninsula, Newfoundland. Waves crossing the Atlantic Ocean were recorded on the coasts of Portugal and the Azores and visually observed along the coasts of Canada and in Bermuda, and recorded on tidal gauges as far south as Charleston, South Carolina (Reference 318).

The role that salt diapirism and methane gas hydrate decomposition may play in landslide potential near the Blake Ridge is discussed in some detail in Subsection 2.5.1.1.1.1.1.3.

Contrary to widely held views based on studies of ancient rocks, basin plains are not necessarily distal portions of fans dominated by thin and relatively fine sediments (Reference 734). On the contrary, basin plains such as the Hatteras and Sohm Abyssal Plains are distinguished by the thickest and coarsest sands of any off-shelf sedimentary environments (Reference 315). This finding suggests that these large turbidite deposits are relatively common and capable of moving significant quantities of near-shore materials (coarse sands) across the continental shelf and slope and far onto the abyssal plain. According to Fine et al. (Reference 318), the Grand Banks landslide carried mud and sand eastward up to 1000 kilometers (620 miles) at estimated speeds of 60 to 100 kilometers/hour (37 to 62 miles/hour).

Based on data in the National Geophysical Data Center (NGDC) database, during the past 200 years a total of six tsunamis have been recorded in the Gulf of Mexico and East Coast States (Reference 735). Three of these tsunamis were generated in the Caribbean, two were related to magnitude 7+ earthquakes along the Atlantic coastline, and one reported tsunami in the Mid-Atlantic states may be related to an underwater explosion or landslide. In the NGDC database, as of June 2010, there are five documented run-up events listed for the Florida Peninsula. The run-up events are listed in the Table 2.5.1-207. The NGDC database does not include the 1946 Dominican Republic tsunami with a possible run-up of 10 feet (3 meters) at Daytona Beach, Florida, because the exact measure of the tidal gage run-up at that location cannot be verified from the original cited reference (Reference 736).

### Potential Puerto Rico Trench Tsunami Sources

The Puerto Rico Trench is the deepest part of the Atlantic Ocean, with water depths exceeding 8400 meters (27,600 feet) (Reference 582). Large landslide escarpments have been mapped on the seafloor north of Puerto Rico although their ages are unknown. Seismic stratigraphy of the landslide slopes appears to indicate that massive carbonate blocks slide coherently. The failure of coherent blocks on a steep slope appears to cause the high tsunami run-up associated with Puerto Rican tsunamis (Reference 582).

The October 11, 1918, $M_W$ 7.2 (see Subsection 2.5.2.1) earthquake in the Mona Passage between Hispaniola and Puerto Rico generated a local tsunami along the western coast of Puerto Rico that claimed approximately 140 lives (References 681, 737, 738, and 739). A tsunami with run-up heights reaching 6 meters (20 feet) followed the earthquake causing extensive damage along the western and northern coasts of Puerto Rico, especially to those villages established in a flood plain (Reference 739). High-resolution bathymetry and seismic reflection lines in the Mona Passage show a fresh submarine landslide 15 kilometers (9 miles) northwest of Rinćon in northwestern Puerto Rico and in the vicinity of the first published earthquake epicenter. The landslide area is approximately 76,000 meters$^2$ (830,000 feet$^2$) and probably displaced a total water volume of 10,000 meters$^3$ (353,000 feet$^3$). The landslide's headscarp is at a water depth of 1200 meters (3900 feet), with the debris flow extending to a water depth of 4200 meters (13,800 feet) (Reference 738). This submarine landslide is now believed to be the primary cause of the 1918 tsunami (Reference 738).

### Potential Bahama Platform and Straits of Florida Tsunami Sources

The Ocean Drilling Program (ODP) drilled four holes up to 447 meters (1500 feet) on the Bahama Platform in the Straits of Florida immediately south of the Units 6 & 7 site (Site 626, Figure 2.5.1-211). Holes 626C and 626D provided significant results as described below (Reference 740). Three stratigraphic units, numbered I through III from the surface down, were identified.

- Unit I is 122 meters (400 feet) thick and consists of skeletal carbonate sands comprising planktonic foraminifers and neritic skeletal grains from the platforms, middle Miocene to Pleistocene in age.

- Unit II is 48 meters (158 feet) thick and consists of muddy lime rubble, graded rubble and sand, and muddy sand, interpreted as debris flows and turbidites with intercalations of pelagic sediment, middle Miocene in age.

- Unit III is 277 meters (900 feet) thick and consists of skeletal carbonate sands as in Unit I with numerous intercalations of lithified layers (skeletal grainstones and packstones), of late Oligocene to middle Miocene in age.

Units I and III are interpreted as contourite deposits of the Gulf Stream, which sweeps the drill site with bottom velocities of 20 to 40 centimeters/second (8 to 16 inches/second) (Reference 741). In general, the contourite deposits formed in the lower velocity zones along margins or beneath the core of higher-energy currents, where flow velocities are low enough to induce deposition but yet high enough to contain a high suspended load that would not be present in the absence of the current. If there is a strong or concentrated nepheloid layer, (a layer of water above the ocean floor that contains significant amounts of suspended sediment (Reference 905)), a rapid deposition of sediment will occur, forming a contourite/turbidity deposit (Reference 906). Unit II interrupts the contourite record in an impressive way. During a four million year interval in the middle Miocene, debris flows and turbidites were emplaced too rapidly to be reworked by the bottom currents. ODP scientists reviewing the stratigraphic relationships hypothesize that the large debris flow and turbidites of Unit II represent material that had accumulated on top of the carbonate banks at a marine high stand and that the material became unstable as sea level fell (Reference 740). Debris flows and associated turbidites, the size of Unit II, might not be expected to occur in today's environment of rising sea levels. The present current regime is different from that in the Miocene because sea level has been generally rising since the end of the Wisconsinan glacial stage (References 907 and 908).

The Unit II gravity flows are synchronous with the “Abaco episode” (associated with the Jacksonville Fracture Zone shown in Figure 2.5.1-229) in the western North Atlantic Ocean. The “Abaco episode” is represented by gravity-flow deposits spanning most of the Miocene, with sedimentation-rate peaks in the middle Miocene as described by Reference 727. The gravity flow material points to sources on the adjacent Bahama Platform, on its flanks, and on the floor of the Straits of Florida. Several other sediment gravity-flow events occurred in the region during the same period. Lower to middle Miocene gravity-flow deposits were cored at Sites 627 and 628 (Figure 2.5.1-211) and large middle Miocene slumps were identified from seismic profiles north of Little Bahama Bank near Sites 627 and 628 (Reference 476). However, these deposits differ in scale and lithology from those at Site 626. The Great Abaco Member of the Blake Ridge Formation in the Blake-Bahamas Basin (for location see Figure 2.5.1-214), penetrated at DSDP Sites 391 (Reference 422) and 534 (Reference 728), contains gravity-flow deposits that span most of the Miocene, with sedimentation-rate peaks in the middle Miocene. Off the west coast of Florida, Mullins et al. (Reference 305) document a middle Miocene slide scar that resulted from the failure of a 120-kilometer (93-mile) length of the western margin of the Florida carbonate platform. The timing of these flow events suggests the possibility of a common paleotectonic cause.

While there is clear evidence of past submarine landslides near the Florida Peninsula, the stratigraphic record, especially from drill cores, is incomplete for use in evaluating the aerial extent of landslide effects. However, based on recent bathymetric data and for PMT purposes, a potential landslide-induced tsunami is discussed in Subsection 2.4.6.1.3.

### Potential West Florida Escarpment Tsunami Sources

Information on the potential West Florida Escarpment sources is discussed in Subsection 2.4.6.1.2, Submarine Landslides in the Gulf of Mexico.

### Potential Northern Coast of Cuba Tsunami Sources

Subsequent to the 2004 Indian Ocean tsunami, Cuban geologists began reexamining the historical, geologic, and seismic records of Cuba to evaluate potential tsunami hazards. Iturralde-Vinent (Reference 742) summarizes the current understanding of tsunami hazards in Cuba with a simple graphic (Figure 2.5.1-345). The graphic indicates large marine boulders deposited on the southern coast of Cuba, possibly by tsunamis, on the extreme southwestern coast, on the Isla de la Juventud, and along the seismically active southeastern coast of Cuba. Iturralde-Vinent (Reference 742) also identifies a significant coastal area of northwestern Cuba as a zone of potential tsunami hazards, with evidence of medium size carbonate boulders emplaced by waves on coastal terraces. A hazard zone for 3-meter (10-feet) high tsunamis is located on the northern coast of Cuba, between the cities of Havana and Matanzas (Figure 2.5.1-345).

Based on a lack of field evidence for tsunami deposits in southern Florida, it appears that the southern Florida coastline is generally protected from potential tsunami events by the broad expanse of shallow water of the Straits of Florida, by the steep Atlantic-facing escarpments represented by the Blake Plateau and the Bahama Platform, and by the steep Gulf of Mexico-facing escarpment of the Florida Escarpment. Knight (Reference 743) provides initial modeling of tsunami impacts to the Gulf of Mexico and southern Atlantic Coast from earthquake sources within the Gulf and the Caribbean regions. The two-dimensional depth averaged model developed at the University of Alaska, Fairbanks (Reference 744) is used to propagate initial disturbances to all points along the U.S. Gulf of Mexico and Atlantic coasts. Four initial sea level disturbances were created using Okada's formulas (Reference 745) in conjunction with associated hypothetical earthquakes. The model earthquakes do not necessarily correspond to expected magnitude, likelihood of rupture, or precise location on known thrust faults. They were chosen in part to excite various ocean basins and to present worst-case conditions. The results indicate that sources outside the Gulf of Mexico are not expected to create tsunamis threatening the Gulf Coast. The results also indicate that, due to significant energy losses from bottom friction through the shallower portions of the Straits of Florida and Caribbean region, both Gulf of Mexico and Atlantic coasts, including the Units 6 & 7 site, are well shielded from the large model (Puerto Rico Trench) Caribbean source (Reference 743). Subsection 2.4.6 contains a more detailed discussion of tsunami modeling specific to the Units 6 & 7 site.

In their report to the NRC, the Atlantic and Gulf of Mexico Tsunami Hazard Assessment Group and U.S. Geological Survey (Reference 746, p. 72) note “[w]e believe the reason why there are no reports from the 1755 tsunami [from the Lisbon earthquake] in southern Florida could be attributed to the northern Bahama Banks, which may have acted as a barrier to that area.”

## 2.5.1.2 Site Geology

The Units 6 & 7 site is located within the Southern Slope subprovince of the Atlantic Coastal Plain physiographic province. The geology of the site (Figure 2.5.1-331) was and is influenced by sea level fluctuations, processes of carbonate and clastic deposition, and erosion. The Paleogene (early Tertiary) is dominated by the deposition of carbonate rocks while the Neogene (late Tertiary) is influenced by the deposition of quartz sands, silts, and clays (Reference 287). Deposition of carbonate rock resumed again during the Pleistocene. Within the site area the dominant rock types are limestones of the late Oligocene Arcadia Formation, the late Oligocene- to early Miocene sands and silts of the Peace River and Tamiami formations, and the Pleistocene fossiliferous limestone of the Fort Thompson Formation, the Key Largo Limestone, and the Miami Limestone (Figure 2.5.1-332). Minor units of alluvial soils, organic muck/peat, and silt cover the surface. During

the Pleistocene, worldwide glaciation and fluctuating sea levels influenced the geology in the site vicinity (Subsection 2.5.1.1.1.1.1.1). Drops in sea level caused by growth of glaciers increased Florida's land area significantly, which led to increased erosion and clastic deposition. Warm interglacial periods resulted in a rise in sea level and an increase in nutrient-rich waters leading to an increase in carbonate build-up (Reference 287). The geology within the site area is dominated by flat, planar bedding in late Pleistocene and older units. No tectonic structures have been identified within the site area (Subsections 2.5.1.2.3 and 2.5.1.2.4). Subsections 2.5.1.2.2, 2.5.1.2.3, 2.5.1.2.4 and 2.5.3 describe karst-related vegetated solution depressions (Figure 2.5.1-333).

#### 2.5.1.2.1 Site Physiography and Geomorphology

The Units 6 & 7 site is located within Miami-Dade County, Florida, approximately 25 miles (40 kilometers) south of Miami, 8 miles (13 miles) east of Florida City, and 9 miles (14 kilometers) southeast of Homestead, Florida. The site area is located within the Southern Slope subprovince of the Florida Platform (a partly submerged peninsula of the continental shelf) within the Atlantic Coastal Plain physiographic province (Figure 2.5.1-217). The south Florida area is a broad, gently sloping plain with poor drainage; most of the area is below the piezometric surface in saltwater marshes and swamps overlain by peat. The Units 6 & 7 site is bordered on the east by Biscayne Bay, on the west by Florida City and Homestead, on the south by Key Largo, and on the north by Miami. There are numerous canals to the west within an Everglades mitigation bank. The physiographic features bordering the plant property are the Everglades, Florida Keys, and the Atlantic Continental Slope (Figure 2.5.1-217).

The surface geology at the site area is characterized by organic muck (peaty soil) and Miami Limestone (Figures 2.5.1-331 and 2.5.1-334). The organic muck/peat (as described in Subsection 2.5.1.2.2) is the dominant surficial sediment type, whereas the Miami Limestone is exposed in the northern and western parts of the site area (Figures 2.5.1-335 and 2.5.1-334). The Miami Limestone is a marine carbonate consisting predominantly of oolitic facies of white to gray limestone with fossils (mollusks, bryozoans, and corals). The overlying organic muck located in the site area is a light gray to dark gray to pale brown sapric muck (strongly decomposed organic peaty soil) with trace amounts of shell fragments that have reaction to hydrochloric acid. The muck/peat varies in thickness across the site from 2 to 11 feet (0.6 to 3.4 meters) (References 708 and 996).

The site is at or near sea level with an existing elevation of -3.2 to 0.8 feet (NAVD 88). The site generally is flat and uniform throughout with the exception of vegetated depressions, as seen in Figures 2.5.1-333 and 2.5.3-202. The vegetated depressions are dissolution features within the Miami Limestone, described in Subsections 2.5.1.2.2, 2.5.1.2.3 and 2.5.3.

#### 2.5.1.2.2 Site Area Stratigraphy

As part of the Units 6 & 7 site characterization program, subsurface information was collected from 97 geotechnical borings, 22 separate groundwater borings, and 6 cone penetration tests (CPTs). Of the 97 geotechnical borings drilled, 37 are located within the boundary of the Unit 6 power block (600-series borings) and 36 are located within the boundary of the Unit 7 power block (700-series borings) (Figure 2.5.1-336). Subsection 2.5.4 contains a more detailed description of the comprehensive geotechnical investigation employed to characterize the subsurface of the site. The rock core descriptions on the boring logs described in Subsection 2.5.4 (References 708 and 995) and are based on the carbonate classification system described by Dunham (Reference 709) that is commonly used in Florida. The geologic formations encountered in the geotechnical exploration were identified in the field. The upper and lower Fort Thompson Formation identified on the boring logs are reinterpreted in this subsection as the Key Largo Limestone and Fort Thompson Formation. This is based on a broad review of geologic publications, the predominance of coralline structure in the Key Largo Limestone and moldic porosity in the lower Fort Thompson Formation.

Of the 97 geotechnical borings drilled and sampled as part of the Units 6 & 7 site investigations, only 6 were advanced to a depth greater than 290 feet (88 meters) below ground surface (bgs): B-701 was advanced to a depth of 615.5 feet (187.6 meters) bgs, B-601 was advanced to a depth of 419.2 feet (127.8 meters), R-6-1b was advanced to a depth of 464.1 feet (141.5 meters), and R-7-1 was advanced to depth of 459.4 feet (140.0 meters). The remaining 2 (R-6-2 and R-7-2) were advanced to 360 and 370 feet (109.7 and 112.8 meters) respectively, for pressuremeter testing. The remaining 91 borings ranged in depth from 100 to 290 feet (30 to 88 meters) bgs with a median of approximately 128 feet (39 meters) bgs. The Units 6 & 7 subsurface investigation obtained detailed information about the near-surface geologic characteristics and composition of sediments underlying the site. Information gathered from the regional investigation coupled with specific data obtained from borings that were drilled as part of the subsurface investigation were used to develop the site stratigraphic column presented in Figure 2.5.1-332.

Geophysical logs were obtained for 10 of the 88 borings of the initial site investigation. A suite of nine different geophysical logs was prepared for each of the ten borings in which geophysical logging was accomplished. Natural gamma logs were recorded as part of the electric log suite and as a correlation tool with the caliper log. Gamma-ray logs are used to identify lithology, with gamma counts of shale, silt, and clay generally higher (moving to the right) because clays adsorb radioactive particles more readily than other materials. A spontaneous potential (SP) log was also taken to identify lithology, but SP is not as sensitive to changes in lithology as the natural gamma curves. Three different resistivity logs were taken to record the resistivity of the formation at various intervals away from the boring wall and to track the effects of the drilling fluid at different levels. These three logs are also used to identify rock type with sandy units moving the curve to the right and clays moving the curve to the left. A caliper log was taken to record changes in the diameter of each borehole. Suspension shear (S) and compression (P) wave velocity logs were completed in each of the ten designated borings. Finally, an acoustic televiewer log was taken to provide a visual inspection of the interior walls of the boring. The key at the top of each log identifies each of the curves. In the supplemental site investigation, geophysical tests were performed in four of the nine geotechnical borings that were drilled. Tests included: four deviation surveys, two P-S Suspension, two Acoustic Televiewer, two caliper and two Natural Gamma (Reference 995). A more detailed description of the down-hole geophysical logging is available in References 708 and 995.

Figures 2.5.1-331, 2.5.1-335, and 2.5.1-334 show the geology of the site vicinity, site area, and site. The site stratigraphic column (Figure 2.5.1-332) presents the lithologic and hydrostratigraphic units encountered during the site subsurface investigation. Subsection 2.4.12 describes the hydrogeologic units in more detail. These strata are described below as they occur from the ground surface to depth beneath the site. Most borings drilled for the site subsurface investigation penetrate the Miami Limestone, Key Largo Limestone, and Fort Thompson Formation to a depth of over 100 feet (30 meters). Forty deeper borings penetrated into the underlying Tamiami Formation at approximately 115 feet (35 meters) bgs and continued to a depth of around 150 feet (46 meters) bgs; sixteen of these borings continued to depths in excess of 215 feet (66 meters) bgs and penetrated the Peace River Formation of the Hawthorn Group. Borings B-701 (DH), R-6-1b, and R-7-1 encountered the top of the Arcadia Formation, which has an average elevation of –454.8 feet (138.6 meters). This stratum is not fully penetrated in borings B-701(DH), R-6-1b, and R-7-1, with the bottom of boring at El. –617 feet (188 meters), –464.1 feet (141.5 meters), and –459.4 feet (140.0 meters), respectively. The description and characteristics of the geologic units encountered in the site investigation are described below. Boring logs are included in References 708 and 995.

The Holocene section at the Turkey Point Units 6 & 7 site is classified as marl, wetland soils belonging to the saprist (muck) group, and peat. Surficial deposits in the relatively flat areas outside the vegetated depressions at the Turkey Point site were variably characterized as either marls (clay and elastic silt), organic-rich elastic silt, or peat sediments (Reference 996). The surficial layers within the vegetated surface depressions at the Turkey Point site were characterized as peat (Reference 996). Laminated surficial deposits found mostly outside vegetated depressions are

interpreted to have likely resulted from cyclical changes in oxidation-reduction conditions and/or chemical equilibria, in which dark colored, organic-rich sediments were likely deposited under flooded, low-oxygen (reducing) conditions, and light colored, carbonate-rich (HCl reactive) laminae were likely deposited under open marsh, shallow water (and thereby less anaerobic) conditions (Reference 996). In coastal Florida wetlands, marl deposition is typically associated with freshwater conditions. Within the Turkey Point site cores, evidence for historic freshwater conditions is provided by the presence of intact specimens of *Planorbella* spp., a freshwater gastropod (Reference 996).

Saprist soils are generally defined as those in which two-thirds or more of the material is decomposed, and less than one-third of plant fibers are identifiable (Reference 276). Eighty-eight borings were drilled and sampled (standard penetration test [SPT] samples in soil, continuous coring in rock) as part of the Turkey Point Units 6 & 7 initial subsurface investigation (Reference 708). During the supplementary investigation, a total of nine borings are drilled (Reference 995), with three of the borings inclined towards surface depressions. In addition, surficial “muck” deposit (soft, surficial soil, and sediment layers) samples are collected at nine locations (Reference 996). Drilling and sampling locations are shown on Figure 2.5.1-378. The description of the Holocene section (i.e., “muck” deposits [soft, surficial soil, and sediment layers]) in the soil borings across the Units 6 & 7 site (References 708, 995, and 996) includes the thickness, color, hardness, and the presence of organics, silt, roots, and shell fragment contents. The surficial deposits were sampled at the site every 2.5 feet (0.8 meter) using the SPT geotechnical sampling method during the initial site investigation (Reference 708) and sampled continuously in 2013 using a McCauley Sampler (Reference 996). The muck soils are classified under the Unified Soil Classification System in accordance with ASTM D2488-06. Modifiers such as trace (<5 percent), few (5 to 10 percent), little (15 to 25 percent), some (30 to 45 percent) and mostly (50 to 100 percent) were used to provide an estimate of the percentage of gravel, sand and fines (silt or clay size particles), or other materials such as organics (muck) or shells. In general, the thickness of the surficial deposits ranges from 2 to approximately 11 feet (0.6 to 3.4 meters). Muck is observed in the geotechnical borings and the multichannel analysis of surface waves (MASW) survey data across the site. The surficial deposits are thicker in the areas of the surficial dissolution features (vegetated depressions), filled entirely with peat (Figures 2.5.4-229 and 2.5.4-230). Color ranges from black to light gray, dark grayish brown to light brownish gray, and dark olive brown to light olive brown. Mottled coloration is also noted in the muck. The consistency of the muck is very soft-to-soft. Peat, with fibrous internal structure, is identified within organic soils in 13 of the site borings (B-614, B-625, B-626, B-702, B-715, B-725, B-727, B-729, R-6-1a, R-6-1a-A, R-6-1b, R-6-2, and R-7-4) as well as in all borings sampled continuously using the McCauley Sampler (Reference 996). The organic content of the muck (elastic silt and organic-rich silt) was estimated to vary from 2.9 to 30.3 percent, with an average of 10.6 percent (Reference 996).

Only one sample from boring B-601 (DH) contains mostly silt. Trace to some sand is noted in three borings: B-617, B-623, and B-723. Neither the sand nor the silt can be correlated across the site as continuous stratigraphic units. However, fine-grained calcareous material, marl, appears to overlie the muck in 10 borings: B-736, B-738, B-802, B-810, B-812, B-813, M-6-1b, M-6-2a, M-7-1a, and M-7-2c. This marl-like material is described as a fat clay to sandy fat clay and elastic silt that is light/dark gray to greenish gray, very soft, moist to wet, with some fine grained sand and strong hydrochloric reaction (References 708 and 996). Where present, the marl/elastic silts represent the uppermost surficial sediments. This type of marl deposit forms when the ground surface is flooded for several months each year in the summer followed by a number of dry months during the winter (hydroperiod). During the hydroperiod, the microalgae (periphyton) grow on the surface water. The precipitation of the microalgae from the calcium bicarbonate saturated water creates marl (Reference 909).

It is important to note that the surficial deposits described at the Turkey Point site generally correspond to the surficial sediment sequences described within other coastal wetland systems adjacent to Biscayne Bay (References 397, 997, and 998). Moreover, the thickness of the surficial

sediments at the Turkey Point site is similar to those reported for other coastal wetland locations in southern Florida and the greater circum-Caribbean, including the Bahamas and Bermuda (Reference 996).

The bedrock surface throughout the site consists of the Pleistocene Miami Limestone overlain by muck/peat. At the site, the Miami Limestone is a white, porous, sometimes sandy, fossiliferous, oolitic limestone (grainstone) with vugs that are typically oriented in either the horizontal or the near-vertical direction. The formation is mostly soft to medium hard throughout, but typically very hard at the base. The top of the Miami Limestone was generally encountered at a depth of 3 to 11 feet (0.9 to 3.4 meters) bgs. The Miami Limestone is approximately 25 feet (7.6 meters) thick beneath the site.

The Pleistocene Key Largo Limestone underlies the Miami Limestone at Units 6 & 7 site. The contact between the Miami Limestone and the Key Largo Limestone is generally an irregular gradational contact primarily inferred from changes in hardness and oolite content. Based on previous investigations by others (e.g., Reference 710), the Key Largo Limestone was initially identified and logged as the upper Fort Thompson Formation. Subsequent investigation, including a review of recent publications (e.g., References 373 and 711) and a reexamination of the rock core, indicate that the coralline limestone facies should be identified as the Key Largo Limestone, not the upper Fort Thompson Formation. The Key Largo Limestone is a coralline limestone characterized by the presence of vuggy porosity with a high degree of interconnectivity. The coralline vugs within the Key Largo Limestone typically exhibit evidence of precipitation of secondary minerals (i.e., calcite). The contact between the Key Largo Limestone and the underlying Fort Thompson Formation has been identified at the site as a marker layer of dark gray, fine-grained limestone occurring at the base of the Key Largo Limestone. The dark gray limestone encountered in most of the site borings is generally 2 feet (0.6 meter) or more thick and often exhibits a sharp color change from light to dark gray at its base marking the transition from the Key Largo Limestone to the Fort Thompson Formation.

The Key Largo Limestone is a fossil coral reef that is believed to have formed in a complex of shallow-water, shelf-margin reefs and associated deposits along a topographic break during the last interglacial period (Reference 406). The Key Largo Limestone is exposed at the surface in the Florida Keys from Soldier Key on the northeast to Newfound Harbor Keys near Big Pine Key on the southwest. At the site, the Key Largo Limestone is generally encountered at a depth of 23 to 35 feet (7 to 11 meters) to bgs and is approximately 22 feet (7 meters) thick.

The Pleistocene Fort Thompson Formation directly underlies the Key Largo Limestone. The Fort Thompson Formation is generally a sandy limestone with zones of uncemented sand interbeds, some vugs, and zones of moldic porosity after gastropod and/or bivalve shell molds and casts. Overall, the vugs and molds within the Fort Thompson Formation create a secondary porosity with a lower degree of interconnectivity than the vugs within the Key Largo Limestone. The top of the Fort Thompson Formation is generally encountered at a depth between 44 and 54 feet (13 to 16 meters) bgs and has a thickness of approximately 66 feet (20 meters) at the site.

The Pliocene Tamiami Formation directly underlies the Fort Thompson Formation. The contact between the Tamiami Formation and the Fort Thompson Formation is an inferred contact picked as the bottom of the last lens of competent limestone encountered. The placement of this inferred contact in each boring was primarily determined from core recoveries and drill rates. The Tamiami Formation is a poorly defined lithostratigraphic unit containing a wide range of mixed carbonate-siliciclastic lithologies. The Tamiami Formation generally consists of well-sorted, silty sand, but locally it is interlayered with clayey sand, silt, and clean clay. The top of the Tamiami Formation is generally encountered at a depth between 102 and 125 feet (31 to 38 meters) bgs with an average thickness of 95 feet (29 meters) at the site.

The Miocene-Pliocene age Peace River Formation of the Hawthorn Group directly underlies the Tamiami Formation. The Peace River Formation comprises interbedded sands, clays, and carbonates. The contact between the Peace River Formation and the Tamiami Formation is inferred based an increase in activity on the gamma ray log (References 708 and 995). The Peace River Formation is penetrated in only the 14 deepest borings and generally consists of well-sorted, silty sand down to approximately 455 feet (139 meters) bgs. The top of the Peace River Formation is encountered at a depth between 206 and 223 feet (63 and 68 meters) bgs with an average thickness of 242 feet (74 meters).

The Oligocene-Miocene age Arcadia Formation of the Hawthorn Group underlies the Peace River Formation. The Arcadia Formation consists of carbonate rock ranging from packstone to wackestone to mudstone, with a few isolated lenses of silty sand. The Arcadia Formation varies in hardness from soft (friable) to hard (well indurated), with colors ranging from pale yellow to greenish gray. The Arcadia Formation is fossiliferous with shell molds and casts of bivalves and gastropods found in some locations within the core. The unit is capped by a gray, hard, indurated dolostone/grainstone with sugary texture containing a few gastropod shell molds and casts. The Arcadia Formation is encountered in borings B-701 DH, R-6-1b, and R-7-1. The top of the Arcadia Formation is encountered at approximately El. –455 feet (–139 meters). This stratum is not fully penetrated in borings B-701(DH), R-6-1b, and R-7-1, with the bottom of boring at El. –617 feet, (–188 meters), –464.1 feet (–141.5 meters), and –459.4 feet (–140.0 meters), respectively. The thickness of the Arcadia Formation at the site exceeds 161 feet (49 meters).

Four geologic cross sections, two isopach (thickness) maps, two structure contour maps, and a site geologic map were prepared from the information obtained from the site subsurface investigation (Figure 2.5.1-334). Geologic cross section A-A' (Figures 2.5.1-338 and 2.5.1-386) extends east-west through the power blocks and 13 borings, including the four deepest borings, B-601, B-701, R-7-1, and R-6-1b. Cross section B-B' (Figures 2.5.1-339 and 2.5.1-387) extends west-east through the southern edge of the site and eight borings, the deepest at 46.6 meters (153 feet) bgs. Cross section C-C' (Figures 2.5.1-340 and 2.5.1-388) extends diagonally northwest-southeast through the entire site and passes through the western power block. Cross section C-C' includes eight borings including two of the deepest borings, B-701(DH), at a depth of 187.6 meters (615.5 feet) bgs, and R-7-1, at a depth of 140 meters (459 feet) bgs. Cross section D-D' (Figures 2.5.1-341 and 2.5.1-389) also extends diagonally northwest-southeast through the entire site but passes through the eastern power block. Cross section D-D' includes six borings; the deepest at a depth of 215 feet (66 meters) bgs.

The locations of the surface traces of the cross sections are shown on Figures 2.5.1-342 and 2.5.1-344 (isopach maps of the Key Largo Limestone and the Fort Thompson Formation, respectively) and Figures 2.5.1-349 and 2.5.1-343 (structure contour maps of the top of the Key Largo Limestone and the Fort Thompson Formation, respectively). Two versions of each of the four cross-sections are provided. Cross-sections in the first set (Figures 2.5.1-336, 2.5.1-339, 2.5.1-340, and 2.5.1-341) are truncated at the elevation of -61 meters (-200 feet) NAVD 88 and depict the subsurface stratigraphy with a vertical exaggeration of 12 to 1. Figures 2.5.1-386, 2.5.1-388, and 2.5.1-389 depict a thicker section of the subsurface stratigraphy on the same cross-sections with a vertical exaggeration of only 4 to 1.

The cross sections indicate that geologic contacts beneath the site are relatively flat and undeformed. This stratigraphy reflects the environment of deposition and subsequent erosion of the paleosurface. The flat and undeformed nature of the geologic contacts is reflected in the isopach maps of the Key Largo Limestone (Figure 2.5.1-342) and the Fort Thompson Formation (Figure 2.5.1-344) that indicate a relatively uniform thickness across the site with no abrupt changes. The structure contour maps of the top of the Key Largo Limestone (Figure 2.5.1-349) and the Fort Thompson Formation (Figure 2.5.1-343) show a relatively flat paleosurface. Boring logs and descriptions of the lithology are included in the geotechnical data reports in References 708 and 995. Section 5.3 of Appendix 2.5AA provides a discussion of the isopach and structure contour maps and reasons for

concluding that they provide no strong evidence for the presence of large collapse features in the Key Largo Limestone or Fort Thompson Formation at the site.

#### 2.5.1.2.3 Site Area Structural Geology

This subsection provides a review of the structural setting from published maps and literature and the Units 3 & 4 UFSAR (Reference 712), which is supplemented by new information from the 2008 geologic mapping and exploration program performed as part of this investigation (Reference 708), the supplemental field investigation (Reference 995), and sampling performed in surficial muck deposits using a McCauley Sampler (Reference 996). The site lies on the stable Florida carbonate platform; no faults or folds are mapped at the surface within more than 25 miles (40 kilometers), and the nearest fault identified by Cunningham (Reference 999) is buried by Pliocene and younger strata (Figures 2.5.1-331 and 2.5.1-393). New data include geologic mapping and bedding attitudes interpreted from lithologic contacts in boreholes. Taken together, these data indicate generally flat, planar bedding in Pleistocene and older units and an absence of geologic structures within the site area.

The site area geologic and surficial maps and cross sections (Figures 2.5.1-335, 2.5.1-334, 2.5.1-337, 2.5.1-338, 2.5.1-339, 2.5.1-340, and 2.5.1-341) present basic structural information. Figure 2.5.1-343 is a structure contour map of the top of the late Pleistocene Fort Thompson Formation. Isopach maps of the Key Largo Limestone and the Fort Thompson Formation are also shown in Figures 2.5.1-342 and 2.5.1-344. Geologic field reconnaissance was performed to verify general structural interpretations presented in literature describing southern Florida and observations of that work are presented herein. The reconnaissance effort generally increased with increasing proximity to the site.

The entire site area was inspected using both pre-construction (1940 black-and-white stereo pairs) and more recent (1999 color infrared and 2004 true color) 1:40,000-scale aerial photography acquired from the USGS and the FDEP. The 1940s aerial photographs showed a considerably greater unaffected surface area and, therefore, a greater number of surface features allowing for identification of a greater number of lineaments than more recent images. Several dominant lineaments are interpreted based on the 1940s aerial photographs, in which the trace and alignments of these features, as well as changes in the orientation of the traces of these features, allowed for the identification of numerous lineaments. Bearings were measured for lineaments without consideration of type of surface feature that formed the lineament. All identified surface lineaments were annotated with straight lines for simplicity. Three main lineament orientations and two subsidiary orientations are observed. The three main lineament orientations are east-west, northeast-southwest, and northwest-southeast with two subsidiary orientations of east-northeast-west-southwest and north-northeast-south-southeast. The most recent study of lineaments in south Florida was done by the U.S. Army Corps of Engineers (USACE) (Reference 345) in which they identified lineaments on Landsat imagery, and compared the lineaments from the Landsat data with published geologic structural data in order to provide interpretations of possible correlations to known regional hydrogeological trends and anomalies. The intent of this study was to help identify structural features that may impact hydraulic parameters of the Floridan Aquifer System. The study concludes that the lineaments are systematically oriented in four principal groups of external stresses in the west-northwest, north-northwest, north-northeast, and east-northeast directions. These are similar to the orientations identified at the site. The study states that numerous investigations have shown that most of the fractures and faults initially identified as lineaments are vertical or near-vertical zones of fracture concentrations that formed predominately by Earth tides, tectonic forces, diagenetic, or weathering processes (Reference 345). The documented fracture orientations in Florida supported the initial assumption that the lineaments identified in the Turkey Point site area were associated with fractures in the subsurface.

Interpretations of the photographs did not identify any structural features to be studied further within the site area. Field reconnaissance within the site area included detailed geologic mapping of the site and inspection of available outcrops of Miami Limestone along the banks of the cooling canals. The late Pleistocene Miami Limestone underlies the entire site area, and is mapped at the surface throughout large portions of southern Florida (Figure 2.5.1-331). However, this unit rarely crops out and is often covered by recent unconsolidated soil or other deposits (Figure 2.5.1-337). The portions of the site that have not been disturbed by the construction of cooling canals are covered by a thin (roughly 2– to 11–foot or 0.6– to 3.4–meter thick) veneer of organic-rich mud, peat and silt, generally referred to as "muck" (Figure 2.5.1-334). Field reconnaissance, a review and interpretation of aerial photography, a review of published literature, and an analysis of the results of the subsurface exploration (Reference 708) did not reveal any evidence for tectonic deformation within the site vicinity or site area. No folds, fractures, faults, or geomorphic features indicative of faulting, or other tectonic structures have been observed or mapped (Figures 2.5.1-331, 2.5.1-335, 2.5.1-334, 2.5.1-337, 2.5.1-338, 2.5.1-339, 2.5.1-340, and 2.5.1-341) in the site vicinity or site area.

Regional structural information from borings across southern Florida indicates that Cretaceous to Pleistocene strata are generally flat-lying or have shallow dips (<1°) that likely reflect paleotopography (References 257, 713, and 714) (Figures 2.5.1-259 and 2.5.1-260). For example, the base of the Fort Thompson Formation has a dip of 0.06° to the southeast in the vicinity of the existing cooling canals (from Reference 715). Data presented in Figures 2.5.1-338, 2.5.1-339, 2.5.1-340, and 2.5.1-341 and Reference 708 confirm that planar, undisturbed bedding persists beneath the site. Based upon local boring data, vertical relief of several feet is found in the contact of the Miami Limestone with the underlying Key Largo Limestone in the site vicinity. However, upon examination during the field reconnaissance, this relief is considered to be a primary sedimentary feature related to the reef origin of the Key Largo Limestone and not due to tectonic or non-tectonic deformation. Field reconnaissance, a review and interpretation of aerial photography, a review of published literature, and an analysis of the results of the subsurface exploration (Reference 708) indicate that the Miami Limestone and older strata are consistently oriented and are not measurably offset or deformed by faulting within the site area (Figures 2.5.1-335, 2.5.1-334, 2.5.1-337, 2.5.1-338, 2.5.1-339, 2.5.1-340, and 2.5.1-341).

Additionally, previous site and regional investigations (References 712 and 715) have identified no systematic jointing patterns within bedrock underlying the site area. Geologic field reconnaissance also included inspection of aerial imagery for lineaments and possible hazards within the site area. Lineaments and local alignments of vegetated depressions were identified in the site area (Figure 2.5.3-202). Field reconnaissance, a review and interpretation of aerial photography, a review of published literature, and an analysis of the results of the subsurface exploration (References 708, 995, and 996) found no geomorphic evidence to suggest differential uplift across any of the lineaments or any structural or stratigraphic evidence to suggest lateral displacement across any of the lineaments. These lineaments do not correlate with any potential folds, faults, or other tectonic or geologic structures within the site area (see description in Subsection 2.5.3.2 and Figures 2.5.1-335, 2.5.1-334, 2.5.1-337, 2.5.1-338, 2.5.1-339, 2.5.1-340, 2.5.1-341, and 2.5.1-342).

#### 2.5.1.2.4 Site Geologic Hazards

Examination of the Units 6 & 7 site area has provided no evidence of active tectonic features, no known tsunami deposits, no evidence for seismically induced paleoliquefaction features or other indicators of paleoseismic activity, and no known sinkholes in the underlying karst terrane. No piercement-type salt domes are located within the site vicinity. The nearest salt dome is approximately 220 miles (350 kilometers) southeast of Units 6 & 7 site along Cuba's northern coast (Reference 430). Evaluation of seismic hazards in the site region and beyond is discussed in this subsection and in Subsections 2.5.2 and 2.5.3. This subsection provides further discussion of efforts to identify dissolution features at the site area as well as any tsunami-related features within the site region.

Twichell et al. (Reference 320) conclude that large landslides related to the upward migration of salt along normal faults may be the cause of increased activity of small earthquakes in this area, and these earthquakes along with oversteepening of the sea floor due to salt movement could lead to repeated slope failures.

### Dissolution Features

Subsection 2.5.1.1.1.1.1.1 describes the three main types of sinkholes common to Florida as defined by Sinclair and Stewart (Reference 264). Also, as described in Subsection 2.5.1.1.1.1.1.1, the FGS classifies sinkhole occurrences into four type areas, Area I through Area IV (Figure 2.5.1-222).

The Units 6 & 7 site is located in Area I, where sinkholes, if they occur, are typically solution sinkholes. Their gradual development is explained by the relatively slow rate at which limestone dissolution and removal occurs. The maximum rate of wall retreat in limestone is estimated to be on the order of 0.04 inches (1.02 millimeters) per year (Reference 716). The maximum potential dissolution for the Lower Suwannee River Basin in Florida, which is located approximately 300 miles (480 kilometers) northwest of the site, is calculated to be less than 0.002 inches (0.05 millimeters) per year (Reference 717). Thus, active dissolution of limestone at the site is not considered to present a geologic hazard.

While large cavities and collapses are not expected in Area I, which includes the Units 6 & 7 site, localized factors can influence whether or not cavities and the potential for collapse exist. Catastrophic collapse is not known to occur in southern Florida. In addition to the fact that cavities in this area are generally small, a stress mechanism, such as a rising or falling water table, is necessary to trigger the collapse of overburden into preexisting cavities that may have taken long periods of time to form (Reference 719). Such a natural triggering mechanism is not prevalent in south Florida because the water table is generally very close to the surface with little fluctuation.

According to Renken et al. (Reference 721), sinkholes and caves have been found in Miami-Dade County only along the Atlantic Coastal Ridge, where limestone is present at a relatively high elevation, extending southward from south Miami toward Everglades National Park (Figure 2.5.1-217). The Atlantic Coastal Ridge is up to 15 feet (4.5 meters) high as it trends north-northeast to south-southwest into the site vicinity (Reference 265). Further discussion of the Atlantic coastal ridge is found in Subsection 2.5.1.1.1.1. Parker et al. (Reference 722) state that the Miami Limestone and Fort Thompson Formation have significant permeability and solution features that have created turbulent flow conditions in some wells. According to Cunningham et al. (Reference 723), topographic relief related to karst dissolution is well documented in the Lake Belt area of north-central Miami-Dade County approximately 17 miles (27 kilometers) northwest of Units 6 & 7. No major sinkholes have been reported within the site vicinity, with the closest reported sinkhole near Ives Estates on the very northern edge of Miami-Dade County, being about 40 miles (63 kilometers) to the northeast of the site (Reference 1021). These features are discussed in greater detail in Subsection 2.5.1.1.1.1.1.1.

An FGS investigation (Reference 724) concludes that most of Miami-Dade County is underlain by limestone containing solution cavities.

Zones of secondary porosity have formed in limestone beneath the site where microkarst features have developed (Subsections 2.4.12.1.3.1 and 2.5.1.2.4). These zones of secondary porosity provide areas of preferential groundwater flow. The microkarst features are thought to have formed by solution enlargement of sedimentary structures when fresh groundwater formerly flowed from inland areas, mixed with seawater, and facilitated dissolution as it flowed through the zone to the sea. The zones of secondary porosity can be subdivided into two categories: touching-vug porosity and moldic porosity.

The two zones of secondary porosity were identified at the site following review of the geophysical logs, the geotechnical boring logs, and the shear wave velocity logs. In general, the zones of secondary porosity were identified based on increases in borehole diameter on the caliper logs, darkened areas on the acoustic televiewer images, typically lower P-S wave velocity values, rod drops, and in the case of touching-vug porosity, loss of drilling fluid circulation. Figures 2.5.1-351, 2.5.1-352, and 2.5.1-353 show the approximate locations of the two zones of secondary porosity on three example-boring logs, B-604 (DH), B-608 (DH), and B-710 G (DH), and their locations at the Turkey Point Units 6 & 7 site are shown on Figures 2.5.1-228 and 2.5.4-202. Figures 2.5.1-351, 2.5.1-352, and 2.5.1-353 were compiled using the lithology, caliper, natural gamma, acoustic televiewer, and velocity (*Vs* and *Vp)* logs.

Recent studies by Cunningham et al. (References 404 and 723) suggest vuggy porosity is common within the Biscayne Aquifer (Miami Limestone, Key Largo Limestone, and Fort Thompson Formation) and that typical solution features associated with the touching-vug porosity include solution-enlarged fossil molds up to pebble size, molds of burrows or roots, irregular vugs surrounding casts of burrows or roots, and bedding plane vugs. Cunningham et al. (Reference 404) show images of vugs in the Miami Limestone and Fort Thompson Formation, with cavernous vugs approximately 4 feet in height (Figure 2.5.1-385). The results of extensive site investigation for Turkey Point Units 6 & 7 (Subsections 2.5.1.2.2 and 2.5.4.2.2, and Reference 708) offer no evidence that karstification of the area has developed cavernous limestone with the potential for collapse and formation of sinkholes.

Touching-vug porosity occurs on the site within the approximate depth interval of 20 to 35 feet (Figures 2.5.1-351, 2.5.1-352, and 2.5.1-353) near the contact of the Miami Limestone and Key Largo Limestone (the “upper zone” of secondary porosity discussed in Subsection 2.4.12.1.4). Because the elevation of ground surface at the site is approximately 0 feet NAVD 88 (Reference 708), this depth interval corresponds approximately to -20 to -35 feet NAVD 88. The origin of this porosity is solution enlargement of burrows, inter-burrow vugs, moldic fossils, root molds, and vugs between root casts (References 404, 723, and 969). These structures are sufficiently numerous and closely spaced to form a laterally continuous zone of interconnected voids. Results of drilling and coring within the zone of touching-vug porosity during the site subsurface investigation have shown the feature to be laterally persistent, generally of centimeter scale, with very few indications of possible larger voids such as a rod drop.

Moldic porosity occurs in pockets within the approximate depth interval of 60 to 75 feet (-60 to -75 feet NAVD 88) (Figures 2.5.1-351, 2.5.1-352, and 2.5.1-353) in the Fort Thompson Formation and forms the “lower zone” of secondary porosity discussed in Subsection 2.4.12.1.4. The origin of this feature is preferential dissolution of fossil shells and other organic structures rather than the matrix rock within which they are contained, resulting in void spaces of generally centimeter scale within molds of the structures (References 404, 723, and 969). Results of drilling and coring within the zone of moldic porosity during the site subsurface investigation have shown the feature not to be laterally persistent but occurring in isolated sandy pockets with very few indications of possible larger voids such as a rod drop.

As seen from the cores taken during the subsurface investigation and photographs of the cores (References 708 and 995), the potential origin of the touching-vug porosity within the upper zone is associated with original reef structure and, based on Cunningham et al. (References 404 and 723), solution enlargement. The potential origin of the lower zone of secondary porosity is moldic porosity resulting from dissolution on in situ bivalve shells.

As further discussed in Appendix 2.5AA, dissolution of the limestone in the upper zone of secondary porosity likely occurred during the Wisconsinan glacial stage of the Pleistocene Epoch when sea level was lower than during the preceding interglacial stages when the Miami Limestone and Key Largo Limestone were formed (Figures 2.5.1-372 and 2.5.1-373) and fresh groundwater from the Everglades mixed with seawater and discharged through the zone to the sea. The coralline vugs

within the Key Largo Limestone typically exhibit evidence of precipitation of secondary minerals such as calcite (Subsection 2.5.1.2.2). This finding suggests that the environment within the upper zone of secondary porosity is currently one dominated by calcite recrystallization rather than solution. The position of the freshwater/saltwater interface is approximately 6 miles (9.6 kilometers) inland from the site (Figure 2.4.12-207), groundwater within the zone of touching-vug porosity is saline (Tables 2.4.12-210 and 2.4.12-211), the long- term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.78 foot (0.2 meter) per century (Subsection 2.4.5), and there is no freshwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote further dissolution of the limestone within the zone of touching-vug porosity does not now exist and development of large underground caverns with the potential for collapse is not likely within this upper zone of secondary porosity. Further, this zone will be completely removed during excavation of the nuclear island foundations (Subsection 2.5.4.5.1).

Dissolution of the limestone in this zone of secondary porosity likely occurred during the early to mid-Pleistocene Epoch when sea level fluctuated to a level lower than when the Fort Thompson Limestone was formed and fresh groundwater from inland areas discharged through the formation toward the sea. As noted previously, the position of the freshwater/saltwater interface is approximately 6 miles (9.6 kilometers) inland from the site (Figure 2.4.12-207), groundwater within the zone of moldic porosity is saline (Tables 2.4.12-210 and 2.4.12-211), the long-term sea level rise trend at Miami Beach, Florida, as estimated based on data from 1931 to 1981, is 0.78 foot (0.2 meter) per century (Subsection 2.4.5), and there is no freshwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote further dissolution of the limestone within the zone of moldic porosity does not now exist and development of large underground caverns with the potential for collapse is not likely within this lower zone of secondary porosity.

While touching-vug and moldic porosity similar to that noted by Cunningham et al. (References 404 and 723) and Lucia (Reference 969) occur at the Turkey Point Units 6 & 7 site, it should be noted that only occasional small rod drops were noted during the site investigation (References 708 and 995) (Subsections 2.5.1.2.4, 2.5.4.1.2.1 and 2.5.4.4.5.5). A “rod drop” occurs when, while drilling, the bit encounters a relatively soft zone or void and the drill head and rod string suddenly advances at a rate much faster than the rate when drilling the overlying more competent material. A rod drop can also occur during an SPT when the weight of the string of drill rods is sufficient to advance the SPT sampler at the bottom of the borehole without additional blows of the sampling hammer. The occurrence of a rod drop indicates the presence of very soft or very loose material, which can be interpreted as void or cavity infill or as inter-bedded materials with substantially different hardness or compactness. Alternatively, a rod drop could indicate that the drill or sampler might have penetrated a cavity that is only partially filled with soft or loose material.

Groundwater levels monitored in onsite observation wells indicate a consistent site-wide upward vertical flow potential within the Biscayne Aquifer (Table 2.4.12-204). The geotechnical logs of the boreholes in which the rod drops occurred indicate that, except for the two drops that occurred in the Miami Limestone, the drops occurred as the drill or sampler advanced from relatively competent rock into a more sandy zone. In this situation, the upward hydrostatic head within the aquifer may have caused an upward blowout of the sand into the borehole when the confining layer above the sand was breached. The rod drops may have occurred not because the drill or sampler encountered very soft or very loose material indicative of void infill, but because liquefaction of the sand in the blowout zone reduced its bearing capacity to less than the down-pressure on the drill or the weight of the rod string.

The evaluation of all data (References 708 and 995) indicate that outside the vegetated depressions and drainages (in vertical borings), a total of 20.1 feet of interpreted tool drops (due to voids and/or voids filled with soft sediments) are observed, in a total of 7918.4 feet cored, for a 0.3 percent of the total cored in 93 borings. Individual drops in the vertical borings range from 0.4 feet to 4 feet (1.5 feet max within the Unit 6 & 7 building footprints). Results from the site investigations (References 708

and 995), show that interpreted tool drops are found more often under the vegetated depressions and drainages. In the three inclined borings, a total of 15.2 feet of tool drops are observed, in a total of 356.4 feet cored, for a 4.3 percent of the total cored length. Individual drops in the inclined borings range from 0.3 feet to 2.5 feet. Boring locations with interpreted tool drops, among all sampling locations, are shown in Figure 2.5.1-378. The maximum length of interpreted tool drop (due to voids and/or voids filled with soft sediments) is limited to 1.5 feet within the Unit 6 & 7 building footprints, and the frequency of encountering an interpreted tool drop is less than 0.5 percent site-wide. These statistics are based on the drilling conducted during both, the initial and supplemental site investigations (References 708 and 995).

Cavities observed during rock core operations were relatively small. The overall data collected during the Units 6 & 7 subsurface investigations are consistent with a communication with the FGS, which indicates that dissolution present in the site area is generally considered to be microkarst with numerous small cavities. This information is consistent with investigations by Cunningham (References 404 and 723) in the Biscayne Aquifer in southeastern Florida.

An investigation of small surface depressions identified within the site (Figure 2.5.1-333) and site area is discussed in Subsection 2.5.3. The UFSAR for Turkey Point Units 3 & 4 concludes that "[s]uch depressions are not sinkholes associated with collapse above an underground solution channel, but rather potholes, which are surficial erosion or solution features" (Reference 712). These solution potholes are not expected to form large voids beneath the surface that would pose a hazard to the site (Reference 264).

An integrated geophysical survey focused on the Units 6 & 7 power block area and the small surface depressions identified within the site is discussed in Subsection 2.5.4.4.5. Based on an integrated interpretation of the boring data (Subsection 2.5.4.1.2.1) and the integrated site geophysical survey data, there is no apparent evidence for sinkhole hazards or for the potential of surface collapse due to the presence of large underground openings. The origin and significance of the surface depressions as well as the interpretation of the geophysical survey data are discussed further in Appendix 2.5AA. The locations of the vegetated depressions correlate well with results of the geophysical surveys (Figures 2.5.4-223 and 2.5.4-228). The presence of peat within the vegetated depressions, as well as the soft zones within the Miami Limestone (indicated by relatively low SPT "N" values recorded in logs of soil borings drilled on the geophysical survey lines), correlates well with low-gravity anomalies.

The MASW data indicate that the vegetated depressions at the site are underlain by continuous Key Largo Formation (Figures 2.5.4-227 and 2.5.4-241). These two figures show MASW data along survey lines 9 and 10 that intersect at a prominent vegetated depression. Within the limits of survey resolution, the microgravity data do not indicate the presence of large subsurface voids. To address uncertainties in the resolution of the geophysical data away from survey lines and at depth beneath the foundation, a microgravity survey will be conducted at the base of the Unit 6 and Unit 7 nuclear island excavations (Subsection 2.5.4.4.5).

What can be interpreted as karst or sinkhole-like features similar to the small surface depressions on site have been noted in aerial photographs of the nearby portion of Biscayne Bay (Appendix 2.5AA). The bay has been modified and dredged and has an average water depth that ranges from 6 to 13 feet (Reference 991). Assuming the water level in the bay is 0 feet NAVD 88, the bottom of the bay ranges in elevation from approximately -6 to -13 feet NAVD 88. According to Reich et al. (Reference 992), sediments overlying bedrock in the bay range in thickness from less than 6 inches to 30 feet. Using this information and the elevations of the bottom of the bay, the elevation of the bedrock surface within which the "karst/sinkhole-like features" occur on the floor of the bay (or alternatively the "vegetated depressions," "local depressions," and "potholes" described in Subsection 2.5.3) ranges from -6.5 to -43 feet NAVD 88. The upper zone of secondary porosity within the Biscayne Aquifer is located near the contact of the Miami Limestone and Key Largo Limestone at

an approximate elevation of -28 feet NAVD 88 (Subsection 2.5.1.2.4). The lower zone of secondary porosity is located within the Fort Thompson Formation at an approximate elevation of -65 feet NAVD 88 (Subsection 2.5.1.2.4). Based on site stratigraphic data (Subsection 2.5.1.2.2), the units are relatively flat and it appears that the upper zone of secondary porosity at the Turkey Point Units 6 & 7 site occurs within the stratigraphic interval within which the "karst/sinkhole-like features" occur on the floor of Biscayne Bay. That level is the stratigraphic interval of the Miami Limestone and Key Largo Limestone (Figure 2.5.1-332). Results of the site subsurface investigation (References 708, 995, and 996) have demonstrated the absence of large solution cavities at this stratigraphic interval on the site.

While the touching-vug porosity exhibited in the upper zone of secondary porosity and the "karst/sinkhole-like features" on the bottom of Biscayne Bay may be in the same stratigraphic interval, the formation of these dissolution features is somewhat different. Dissolution features such as vugs are typically post-depositional features that occur in a freshwater phreatic system in which groundwater has filled open spaces and causes dissolution. The "karst/sinkhole-like features" on the bottom of the bay appear to be paleo-dissolution features that formed during the Wisconsinan (most recent) glacial stage of the Pleistocene when sea level was approximately 100 meters (328 feet) lower than the modern ocean (Reference 262) and at an elevation favorable for dissolution by rainwater of subaerial limestone in what is now the bay. More information on the development of the "karst/sinkhole-like features" on the bottom of Biscayne Bay is provided in Appendix 2.5AA and in the following paragraph, together with a summary of the evolution of the bay.

The process of limestone deposition in Florida was variable during the Pleistocene Epoch due to fluctuations in glacial runoff and the corresponding sea level. The Sangamon interglacial corresponds to the Q5e interglacial stage that occurred between approximately 125,000 and 75,000 thousand years ago (Reference 928). During this time, sea level rose globally and in Florida resulted in an increase in marine carbonate deposition. Sea level was approximately 20 feet higher than today (References 993 and 994) and covered the entire Florida peninsula south of Lake Okeechobee (Reference 994). The marine sediments (i.e., the Miami Limestone and Key Largo Limestone) that accumulated during the Sangamon and the previous interglacial high sea level stands (Reference 928) were lithified and their depositional morphology preserved. Two elongated sediment ridges that formed the Key Largo Ridge and the Atlantic Coastal Ridge resulted in the limestone basin that is now filled by Biscayne Bay, Card Sound, and Barnes Sound.

During lower sea level stands of the Wisconsinan glacial stage, the Florida platform became emergent (sea level was approximately 100 meters (328 feet) lower than today) and the sea floor of Biscayne Bay was exposed (Reference 262). The exposed sea floor of the Bay was altered by rainwater. Dissolution, re-precipitation, and vegetative soil formation cemented the calcareous surface and slowly produced a very hard reddish limestone "soil crust" over the surface. Carbonate dissolution resulting from infiltration of rain water produced solution holes and pipes into the underlying limestone and solution-hole drainage, in particular dendritic drainage patterns, developed on the limestone of Biscayne Bay and its vicinity, including the Turkey Point Units 6 & 7 site. This process of surface dissolution ended in Biscayne Bay when sea level rose and flooded the Bay but continued on emergent areas, including the Turkey Point Units 6 & 7 site. The depositional morphology and paleo-dissolution morphology resulting from the Sangamon interglacial high sea level stand and Wisconsinan glacial low sea level stand are preserved on the sea floor of Biscayne Bay (References 993 and 994).

It should also be noted that the information related to submarine sinkholes, paleokarst collapses, and sag structures in the site vicinity described in Subsection 2.5.1.1.1.1.1.1 suggests that while dissolution features are present, most are not currently active. Active dissolution is thus likely to be limited at Turkey Point Units 6 & 7, as is the potential for deformation due to collapses within existing (i.e., "paleo") dissolution features.

For example, the observed collapse structures at Jewfish Creek/Lake Surprise (Reference 1016) and Key Largo (Reference 959) appear to have occurred in the Pleistocene, coincident with sea level lowstands, and thus are not particularly relevant analogs for active (or future) surface collapse at (or near) the Turkey Point Units 6 & 7 site. Although substantial in scale and extent, the numerous sag features described in Cunningham and Walker (Reference 958) and various others (References 1013, 1014, and 1015) similarly provide no evidence for post-Pliocene deformation. It seems likely then that comparable collapses within similar features, if present below Turkey Point Units 6 & 7, have already occurred (and are now stabilized).

Importantly, it should be noted that isopach and structure contour maps for the Pleistocene Key Largo Limestone and Fort Thompson Formation at Turkey Point Units 6 & 7 provide no evidence for co-located or similarly oriented closed-contour depressions that would indicate surface subsidence (Figures 2.5.1-342, 2.5.1-343, 2.5.1-344, and 2.5.1-349). Moreover, the few closed-contour depressions that have been mapped are generally not deeper than 0.3 to 0.6 meters (1 to 2 feet). Consequently, these data also suggest that large collapse structures do not underlie the site or have stabilized (if present).

The position of the freshwater/saltwater interface is approximately 6 miles inland from the bay in the vicinity of the site (Figure 2.4.12-207), groundwater beneath the site is saline (Tables 2.4.12-210 and 2.4.12-211), sea level is rising (Subsection 2.4.5.2.2.1), and there is no fresh groundwater shoreline flow near the site. Therefore, a freshwater/saltwater mixing zone that would promote further dissolution of the limestone underlying the Turkey Point Units 6 & 7 site or the dissolution features on the floor of Biscayne Bay does not exist. These features on the floor of Biscayne Bay do not appear to have the capacity for development of large underground caverns with the potential for collapse and formation of sinkholes.

#### Volcanism

Based on the absence of Holocene volcanic features in the site region, no volcanic activity is anticipated in the site vicinity. The closest active volcano is found in on Utila Island off the northwest coast of Honduras, about 750 miles (1200 kilometers) southwest of the Units 6 & 7 site. Active volcanoes are also found in the Lesser Antilles. Soufrière Hills on the island of Monserrat is the closest currently active volcano, located about 1300 miles (2100 kilometers) from the Units 6 & 7 site. Pliocene-Quaternary basaltic flows from volcanic structures are well documented in southern Haiti immediately north of the Saumatre Lake and in the Dominican Republic north of the San Juan Valley. These rocks have been dated at 1 ± 0.5 Ma by K/Ar methods and may be younger (References 816 and 818).

### 2.5.1.2.5 Site Area Engineering Geology Evaluation

Subsection 2.5.1.2.5 addresses engineering soil properties and behavior of foundation materials in Subsection 2.5.1.2.5.1; zones of alteration, weathering, dissolution, and structural weakness in Subsection 2.5.1.2.5.2; prior earthquake effects in Subsection 2.5.1.2.5.3; and effects of human activities in Subsection 2.5.1.2.5.4. Further discussions on earthquake effects are found in Subsection 2.5.2. Additional discussions on soil properties and foundation materials are in Subsection 2.5.4.

#### 2.5.1.2.5.1 Engineering Soil Properties and Behavior of Foundation Materials

Engineering soil properties, including index properties, static and dynamic strength, and plasticity and compressibility are described in Subsection 2.5.4. The foundation bearing strata will be evaluated and geologically mapped as the subgrade excavation is completed to confirm that the observed properties are consistent with those used in the design and that any deformation features discovered during construction do not have the potential to compromise the safety of the plant. The NRC staff will be notified when Seismic Category I excavations are open for construction.

#### 2.5.1.2.5.2 Zones of Alteration, Weathering, Dissolution, and Structural Weakness

Field reconnaissance, a review and interpretation of aerial photography, a review of published literature, and an analysis of the results of the subsurface exploration (References 708, 995, and 996) found no unusual zones of alteration, weathering profiles, or structural weakness in the surface or subsurface. Subsection 2.5.3 describes an investigation of small surface depressions in the site area that are due to surficial limestone dissolution.

In addition, Subsection 2.5.4.4.5 describes an integrated geophysical survey that focused on the Units 6 & 7 power block areas to further evaluate the potential for carbonate dissolution features occurring at the site. Any noted desiccation, dissolution, weathering zones, joints, or fractures will be evaluated and mapped as the subgrade excavation is completed to confirm that the mapped characteristics, such as fracture frequency, are consistent with the borehole data used in the design.

#### 2.5.1.2.5.3 Prior Earthquake Effects

A ground and aerial field reconnaissance investigation, as well as a literature review was conducted to identify potential earthquake-related deformation at the site. These investigations included a review of aerial photography to evaluate anomalous features including depressions, topographic highs, and lineaments. The geologic and geomorphic study found no evidence for active folding or faulting or other past earthquake activity. No features were identified during this investigation that may be related to earthquake-induced ground shaking, including liquefaction-related sand blows or lateral spread fractures. The field reconnaissance augmented and verified aspects of previous geologic maps and publications by the USGS, the FGS, and other researchers. These investigations have recognized no geomorphic, stratigraphic, or other features indicating recent tectonic deformation within the site vicinity. In addition to this field reconnaissance, a review and interpretation of aerial photography, a review of published literature, and an analysis of the results of the subsurface exploration (References 708 and 995) (Figures 2.5.1-335, 2.5.1-334, 2.5.1-337, 2.5.1-338, 2.5.1-339, 2.5.1-340, and 2.5.1-341) that were found no evidence of past earthquake activity or liquefaction-related features within the site area or site vicinity.

#### 2.5.1.2.5.4 Effects of Human Activities

The anthropogenic effects in southeastern Florida of urban development, water mining, limestone mining, oil and gas development, agriculture, and construction of drainage canals have affected the regional groundwater table and associated saltwater intrusion. For example, about 208 Mgal/d of treated municipal sewage was injected into the lower Floridan aquifer, at or below the Boulder Zone, during 1988 (References 747 and 748). There are no indications that the groundwater table has been affected at the site due to those activities. Subsection 2.4.12 contains a more detailed description of the groundwater characteristics.

No oil, gas, or metallic mineral resources have been reported in the site area. There is no present mining or excavation of nonmetallic mineral resources within the site area. The closest quarrying activities are located 8 miles (13 kilometers) from the site. No oil or gas production activities occur within the site or site area. Some oil and gas exploration has been performed in southern Florida, and approximately six dry holes were drilled within the site vicinity (Reference 373).

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## Table 2.5.1-201
## Locations of DSDP and ODP Drill Sites Referenced in FSAR 2.5

| DSD/ ODP | Leg # | Site # | Location Name | Latitude | Longitude | Subsection # |
|---|---|---|---|---|---|---|
| DSDP | 4 | 29 | Beata Ridge/A" and B" Horizons | 14° 47.11'N | 69° 19.36'W | 2.5.1.1 |
| DSDP | 15 | 146 | Venezuelan Basin | 15° 06.99'N | 69° 22.67'W | 2.5.1.1.2.2.7 |
| DSDP | 15 | 149 | Venezuelan Basin | 15° 06.25'N | 69° 21.85'W | 2.5.1.1 and 2.5.1.1.2.2.7 |
| DSDP | 15 | 150 | Venezuelan Basin | 14° 30.69'N | 69° 21.35'W | 2.5.1.1.2.2.7 |
| DSDP | 15 | 151 | Beata Ridge | 15° 01.02'N | 73° 24.58'W | 2.5.1.1.2.2.7 and 2.5.1.1.2.2.8 |
| DSDP | 15 | 152 | Northern Nicaraguan Rise | 15° 52.72'N | 74° 36.47'W | 2.5.1.1.2.2.6 and 2.5.1.1.2.2.7 |
| DSDP | 15 | 153 | Aruba Gap | 13° 58.33'N | 72° 26.08'W | 2.5.1.1.2.2.7 and 2.5.1.1.2.2.8 |
| DSDP | 15 | 154 | Colombian Basin, drilled on Panama outer ridge | 11° 05.11'N | 80° 22.75'W | 2.5.1.1.2.2.5 |
| DSDP | 44 | 391 | Blake/Bahamas Basin (for location see Figure 2.5.1-214) | 28° 13.73'N | 75° 36.76'W | 2.5.1.2.4 |
| DSDP | 68 | 502 | Mono Rise/Colombian Basin | 11° 29.42'N | 79° 22.78'W | 2.5.1.1.2.2.6 |
| DSDP | 76 | 534 | Blake/Bahamas Basin | 28° 20.6'N | 75° 22.9'W | 2.5.1.2.4 |
| DSDP | 77 | 537 | Vicinity of the Catoche Knoll | 23° 56.01'N | 85° 27.62'W | 2.5.1 |
| DSDP | 77 | 538 | on top of the Catoche Knoll | 23° 50.98'N | 85° 10.26'W | 2.5.1 |
| ODP | 101 | 626 | Bahamas/Straits of Florida | 25° 35.08'N | 79° 32.73'W | 2.5.1.2.4 |
| ODP | 101 | 627 | Bahamas/Northern Slope, Little Bahama Bank | 27° 38.10'N | 78° 17.65'W | 2.5.1.2.4 |
| ODP | 101 | 628 | Bahamas/Little Bahama Bank, Mid-Slope | 27° 31.85'W | 78° 18.95'W | 2.5.1.2.4 |
| ODP | 165 | 999 | Kogi Rise in Colombian Basin | 12° 44.639'N | 78° 44.360'W | 2.5.1.1 and 2.5.1.1.2.2.7 |
| ODP | 165 | 1000 | Northern Nicaraguan Rise | 16° 33.223'N | 79° 52.044'W | 2.5.1.1.2.2.5 |
| ODP | 165 | 1001 | Southern Nicaraguan Rise/Hess Escarpment | 15° 45.427'N | 74° 54.627'W | 2.5.1.1.2.2.5.2 |

## Table 2.5.1-202
## K/Pg and Cenozoic Boundary Events Affecting the Caribbean, Gulf of Mexico, and Florida Regions

| Boundary Name | Boundary Event | Description |
|---|---|---|
| K/Pg (~65 Ma) | End-Cretaceous mass extinction event had widespread effects, including the production of toxic gases (NO, $NO_2$) and nitric acid ($HNO_3$), ejection of gases and dust into the stratosphere, destruction of stratospheric ozone, major wildfires that consumed most terrestrial biomass, and widespread evidence of solar radiation absorbing soot (Reference 870). | General consensus is that the K/Pg extinction event was caused by a 65.5 Ma bolide impact at the Chicxulub crater of the Yucatan Peninsula (Reference 518). The tsunami deposits associated with the impact event are found throughout the southern and southeastern U.S. and have been recovered from drill cores throughout the Gulf of Mexico, Cuba and the Caribbean (References 516 and 299). |
| P/E (~56 Ma) | The Paleocene to early Eocene boundary was a return to a "greenhouse" state that had occurred through most of the later Mesozoic. The entire Earth was much warmer than today on average, a condition that required efficient heat transport from the equators to the poles (Reference 871). Sedimentary oxygen isotope ratios indicate a greenhouse-related thermal maximum (Reference 872) possibly caused by one or a combination of events such as changing oceanic circulation patterns (e.g., Reference 873), continental slope failure due to increased current strengths (e.g., Reference 878), sea-level lowering (e.g., Reference 874), bolide impact (e.g., Reference 875), and explosive Caribbean volcanism (References 209 and 220). | Miller et al. (Reference 877) proposed that the early Eocene peak in global warmth and sea level was due not only to slightly higher ocean-crust production but also to a late Paleocene-early Eocene tectonic reorganization. The largest change in mid-ocean ridge length of the past 100 m.y. occurred approximately 60 to 50 Ma, associated with the opening of the Norwegian-Greenland Sea, a significant global reorganization of spreading ridges, and extrusion of 1 to 2 x $10^6$ kilometers$^3$ of basalts of the Brito-Arctic province (Reference 877). A late Paleocene to early Eocene sea-level rise coincides with this ridge-length increase, suggesting a causal relation. Miller et al. (Reference 877) suggest that this reorganization also increased $CO_2$ outgassing and caused global warming to an early Eocene maximum. |
| E/O (~33 Ma) | The E/O boundary is a period of climatic deterioration loosely called the "doubtless," a transition between greenhouse and icehouse conditions. Climate generally cooled progressively. The boundary is represented by changes in oceanic circulation patterns and in stable isotope composition in sediments. Changes were possibly caused by addition or withdrawal of greenhouse gases. At least three major bolide impacts also occurred at 35.5-36 Ma, but they caused no significant change in climate or extinctions (Reference 871) | The E/O boundary represents a decline of mean global temperatures by more than 10° C and was accompanied by expansion of Antarctic glaciation (Reference 883). Tecktites/microtecktites appear across the Caribbean Basin from the Chesapeake Bay and the Popigai bolide impacts (Reference 884). The pervasive Everglades unconformity is postulated to be due to erosion from the tsunami that followed the Chesapeake Bay impact (Reference 286). Some have suggested that $CO_2$ and other greenhouse gases were locked up in methane hydrates on the seafloor (Reference 871). The opening of a seaway between the South Tasman Rise and Antarctica was very close to the E/O boundary (Reference 885). |

## Table 2.5.1-203
## Florida's Marine Terraces, Elevations, and Probable Ages

| Terrace Name | Elevation Range (feet above MSL) | Notes | Probable Age[(a),(b)] |
|---|---|---|---|
| Silver Bluff | 1–10 | — | 0.043 Ma |
| Princess Anne[(c)] | 10–20 | — | 0.064 Ma |
| Pamlico | 10–25 | — | 0.095–0.145 Ma |
| Bethera[(d)]<br>Talbot[(b)] | 25–42 | Formed during pause in sea level retreat from 100-25 feet | 0.210 Ma<br>0.120[(b)]–0.227 Ma |
| Penholoway[(b)] | 42–70 | Formed during pause in sea level retreat from 100-25 feet | 0.393–0.408[(b)] Ma |
| Wicomico | 70–100 | Penholoway-Wicomico form single transgressive-regressive sequence | 0.393 Ma |
| Okefenokee[(d)]<br>Sunderland | 100–170 | Okefenokee and Sunderland terraces grouped by some authors | 0.763 Ma<br>1.430 Ma |
| Coharie | 170–215 | Coharie-Sunderland form single transgressive-regressive sequence | 1.650 Ma |
| Hazelhurst | 215–320 | — | 1.66 to 1.98 Ma |

(a) Probable age is a calculated from ΔH = kT (k = 0.135 x 10-3) with final correlation of high sea level data with deep-sea core stages (Reference 260). Age is given in millions of years before present (Ma).
(b) The approximate age is derived from modeling precipitation, karstification, isostatic uplift, and sea level rise (Reference 927).
(c) The Princess Anne terrace is not seen in Florida but is the ninth terrace that Ward (Reference 260) observes in South Carolina.
(d) Based on terrace recognized in southern Georgia; not recognized as a separate terrace in Florida in Reference 271.

Source: Modified from References 260, 271, and 927

## Table 2.5.1-204 (Sheet 1 of 2)
## Summary of Regional Fault Zones of Cuba

| Fault Name | Within Site Region (200 miles)? | Youngest Unit Offset (1989, Tectonic Map of Cuba, 1:500,000 scale)[a] | Youngest Unit Offset (1988, Geologic Map of Cuba, 1:250,000 scale)[b] | Assigned Age from Geologic Map of Cuba Inset (1985, 1:500,000 scale)[c] | Assigned Age(s) from Other Sources |
|---|---|---|---|---|---|
| Baconao | No | Eocene | Not mapped | Neogene-Quaternary | Active[d] |
| Camaguey | No | Not mapped | Eocene (portions dashed) | Paleogene | Active[d] |
| Cochinos | Yes | Not mapped | Not mapped | Not mapped | Active[d] |
| Cubitas | No | Pre-Cenozoic | Early Miocene (dashed) | Mesozoic | Pliocene-Quaternary[d]<br>Post-Middle Eocene[i] |
| Domingo | Yes | Eocene | Eocene | Not mapped | Inactive[d]<br>Late Eocene[e,f] |
| Guane | No | Not mapped | Not mapped | Paleogene | Active[d] |
| Habana -Cienfuegos | Yes | Not mapped | Not mapped | Paleogene | Active[d] |
| Hicacos | Yes | Not mapped | Miocene | Not mapped | Active[d] |
| La Trocha | No | Not mapped | Middle-Upper Miocene | Neogene-Quaternary | Pliocene-Quaternary[d]<br>Eocene[g] |
| Las Villas | Yes | Eocene | Eocene (portions dashed) | Mesozoic | Pliocene-Quaternary[d] |
| Nipe | No | Not mapped | Miocene (dashed) | Neogene-Quaternary | Active[d] (assessment of Cauto-Nipe fault) |
| Nortecubana | Yes | No data (no mapping offshore) | No data (no mapping offshore) | Mesozoic and Neogene-Quaternary | Inactive [d] |
| Oriente | No | No data (no mapping offshore) | No data (no mapping offshore) | Neogene-Quaternary | Active[d] (assessment of Bartlett-Caiman fault)<br>Active[j] |
| Pinar | Yes | Upper Pliocene-Lower Pleistocene (also covered by same unit) | Early-Middle Miocene | Neogene-Quaternary | Inactive[d] |
| Punta Allegre | No | No data (not mapped at surface) | No data (not mapped at surface) | No data (not mapped at surface) | Unassigned[h] |

## Table 2.5.1-204 (Sheet 2 of 2)
## Summary of Regional Fault Zones of Cuba

**Notes:**

(a) Mapa Tectonico de Cuba (Reference 847)
(b) Mapa Geologico de Cuba (Reference 846)
(c) Mapa Geologico de la Republica de Cuba (Reference 848)
(d) Cotilla-Rodríguez et al. (Reference 494)
(e) Iturralde-Vinent (Reference 440)
(f) Pardo (Reference 439)
(g) Leroy et al. (Reference 499)
(h) Ball (Reference 468)
(i) van Hinsbergen et al. (Reference 500)
(j) Mann et al. (Reference 493)

## Table 2.5.1-205
## Correlation of Morphotectonic Zones and Tectonic Terranes in Hispaniola

| | Morphotectonic Zone | Tectonic Terrane |
|---|---|---|
| Zone 1 | Old Bahama Trench (offshore) | — |
| Zone 2 | Cordillera Septentrional-Samaná Peninsula | Puerto Plata-Pedro García-Río San Juan; Samaná |
| Zone 3 | Cibao Valley | One or more of the three following terranes may be in the subsurface of Zone 3: Altamira; Tortue-Amina-Maimon; Seibo |
| Zone 4 | Massif du Nord-Cordillera Central | Tortue-Amina-Maimon; Loma Caribe-Tavera; Duarte; Tireo; Trois Revières-Peralta |
| Zone 5 | Northwestern-south-central zone (includes Plateau Central, San Juan Valley, Azua Plain, Sierra de Ocoa, Presqu'île du Nord-Ouest) | Presqu'île du Nord-Ouest-Neiba |
| Zone 6 | Cul-de-Sac Plain; Enriquillo Valley | Selle-Hotte-Bahoruco terrane appears to underlie most of the subsurface of Zone 6 |
| Zone 7 | Southern Peninsula; Massif de la Selle; Massif de la Hotte; Sierra de Bahoruco | Selle-Hotte-Bahoruco |
| Zone 8 | Eastern Peninsula; Cordillera Oriental; Seibo coastal plain | Seibo; Oro |
| Zone 9 | San Pedro Basin and north slope of the Muertos Trough | One or more of the following terranes may be in the subsurface of the San Pedro Basin: Loma Caribe-Tavera; Tortue-Amina-Maimon; Seibo |
| Zone 10 | Beata Ridge and Southern Peninsula | Selle-Hotte-Bahoruco |

Source: Reference 566

## Table 2.5.1-206
## Tectonic Interpretation of Terranes in Hispaniola

| Fragments of the Forearc / Accretionary Prism of an Island Arc | Fragments of the Volcano-Plutonic Part of an Island Arc | Fragments of Ocean Floor including Seamounts | Fragment of a Back Arc Basin | Fragment of an Oceanic Plateau |
|---|---|---|---|---|
| 1. Samaná | 3. Altamira | 7. Loma Caribe-Tavera | 10. Trois Rivières-Peralta | 12. Selle-Hotte- Bahoruco |
| 2. Puerto Plata-Pedro García-Río San Juan | 4. Oro | 8. Duarte | | |
| | 5. Seibo | | | |
| | 6. Tortue-Amina-Maimon | | | |
| | 9. Tireo | | | |
| | 11. Presqu'île du Nord-Ouest-Neiba | | | |

Note: Numbers for tectonic terranes correspond to tectonic terranes (zones) in Figure 2.5.1-305
Modified from: Reference 566

| Division of Terranes in Hispaniola Based on Deformational Characteristics[a] | | |
|---|---|---|
| Stratigraphic[b] | Metamorphic[c] | Disrupted[d] |
| 3. Altamira | 1. Samaná | 2. Puerto Plata-Pedro García-Río San Juan |
| 4. Oro | 6. Tortue-Amina-Maimon | 7. Loma Caribe-Tavera |
| 5. Seibo | 8. Duarte | |
| 9. Tireo | | |
| 11. Presqu'île du Nord-Ouest-Neiba | | |
| 12. Selle-Hotte- Bahoruco | | |
| 10. Trois Rivières-Peralta | | |

**Notes:**
(a) Numbers for tectonic terranes correspond to tectonic terranes (zones) in Figure 2.5.1-305
(b) Stratigraphic terranes are characterized by coherent sequences of strata in which depositional relations between successive lithologic units can be demonstrated.
(c) Metamorphic terranes are characterized by rocks metamorphosed to a high enough grade that original minerals, stratigraphic features, and stratigraphic relationships are obscured.
(d) Disrupted terrane are characterized by brittle deformation that obscures the depositional relations between successive lithologic units.

Source: Modified from Reference 566

**Table 2.5.1-207**
**NOAA NGDC Database Tsunami Run-Up Events**

| Date | Tsunami Source | Run-up Location(s) | Distance from Source(s) (Kilometers) | Maximum Water Height |
|---|---|---|---|---|
| 09/01/1886 | Charleston, SC EQ M 7.7 | Jacksonville, FL<br>Mayport, FL | 327<br>310 | No data (eye witness reports) |
| 08/04/1946 | Dominican Republic EQ M 8.1 | Daytona Beach, FL | 1648 | No data (tidal gage measurement) |
| 08/08/1946 | Dominican Republic EQ M 7.9 | Daytona Beach, FL | 1571 | No data (tidal gage measurement) |
| 07/03/1992 | Probably meteorologically induced –not true tsunami (several eye witness reports of offshore fireball) | Daytona Beach, FL | No data | 6 meters |
| | | New Smyrna Beach, FL | | 1.2 meters |
| | | Ormond Beach, FL | | 1.2 meters |
| | | St. Augustine, FL | | No data |
| 12/26/2004 | Sumatra EQ M 9.0 | Trident Pier, FL | 16,475 | 0.17 meters |

## Table 2.5.1-208
## Marine Terraces in the Matanzas Area of Northern Cuba (Sheet 1 of 2)

| Marine Terrace (Reference 915) | Elevation of Marine Terrace (Reference 915) | Geologic Stratum (References 915 and 921) | Depositional Environment (Reference 915) | Possible Geologic Age (Reference 915) | Possible Geologic Age (Reference 921) |
|---|---|---|---|---|---|
| | | | Start of emergence | Start of the Upper Miocene | |
| | | | Erosion Cycle (No. 1) | Upper Miocene | |
| | | | Uplift and buckling (env. 60m) | Pliocene (?) | |
| | | | Erosion Cycle (No. 2) | | |
| | | | Uplift (env. 80 m) | Pliocene | |
| | | | Erosion Cycle (No. 3) | | |
| | | | Uplift and folding (10 and 45 m) | | |
| La Rayonera | 25 and 51 m | | Erosion Cycle (No. 4) | | Pliocene-Pleistocene |
| | | | Uplift and folding (15 and 25 m) | | |
| Yucayo | 15 and 33 m | | Erosion epicycle (No. 1) | Pliocene (?) | Pliocene-Pleistocene |
| | | | Drop in sea level: uplift, very light folding (10m) | Pliocene (?) | |
| Puerto | 16 m | | Erosion epicycle (No. 2) | Start of the Illinoian Glaciation | |
| | | | Drop in sea level (11 and 13 m) | | |
| submarine terrace | No. 1 (-1 m) | | Erosion epicycle (No.3) | | |
| | | | Drop in sea level (1 m) | | |
| submarine terrace | No. 2 (-2 and -6 m) | | Erosion epicycle (No. 4) | | |
| | | | Drop in sea level (env. 130 m) | Illinoian Glaciation Maximum | |
| Continental Shelf terrace | | | Erosion epicycle (No. 5) | | |
| | | | Rise in eustatic sea level (11 m) | Sangamon Interglacial | |
| | | Jaimanitas Formation (Terraza de Seboruco), Rosario Terrace (continental alluvial terrace) | Formation of fringing reefs on uplifted alluvium deposits | | Pleistocene |

## Table 2.5.1-208
## Marine Terraces in the Matanzas Area of Northern Cuba (Sheet 2 of 2)

| Marine Terrace (Reference 915) | Elevation of Marine Terrace (Reference 915) | Geologic Stratum (References 915 and 921) | Depositional Environment (Reference 915) | Possible Geologic Age (Reference 915) | Possible Geologic Age (Reference 921) |
|---|---|---|---|---|---|
| Limits of the Terraza de Seboruco | +/- 8 m | | Drop in sea level (env. 12 m) | Start of the Wisconsinan Glaciation | |
| submarine terrace | No. 3 (-10 and -17 m) | | Erosion epicycle (No. 6) | | |
| | | | Drop in sea level (env. 10 m) | | |
| submarine terrace | No. 4 (-20 and -55 m) | | Erosion epicycle (No. 7) | | |
| | | | Drop in sea level (env. 110 m) | Wisconsinan Glaciation Maximum | |
| Limit of the Restart of the Continental Plate | | | Erosion epicycle (No. 8) | | |
| | | | Eustatic rise to present sea level | Flandrian Transgression | |
| | | Recent alluvium | Induation of river valleys | | |

Source: References 915 and 921

## Table 2.5.1-209
## Marine Terrace Sequences in Southern Florida

| Epoch | Litho-stratigraphic Unit | Marine Sequence Stratigraphic Unit | Radiometric Age Date (ka) | Sample Location | Depth/Elevation | MIS |
|---|---|---|---|---|---|---|
| Pleistocene | Key Largo Limestone/Miami Limestone | Q5e (youngest) | 130–121 | Windley Key, Upper Matecumbe Key and Key Largo | ~4.9 to 5.3 meters above sea level at Windley Key Quarry, water depths of ~16 and ~22 meters | 5e |
| | | Q5c<br>Q5a | 112.4 to 77.8 | Conch Reef, Looe Key, Carysfort Light area and Molasses Reef | water depth of -15.2 and -15.5 meters (Carysfort Light area) | 5c<br>5a |
| | | Q4b? | 230–220 | Long Key Quarry | ~0.7 to 3.5 meters above sea level | 7 |
| | Fort Thompson Formation | Q4a | 340–300 | Point Pleasant Core | NR | 9 |
| | | Q3<br>Q2<br>Q1 (oldest) | ***<br>***<br>*** | Grossman Ridge Rock Reef and Joe Ree Rock Reef | NR | 11<br>11<br>11? |

**Notes:**
"?" uncertainty
*** no reliable dates (Reference 928)
NR- denotes the elevations are not recorded in Reference 928
The Radiometric Age Date column is derived from Uranium-series ages ($^{234}U/^{238}U$) on corals and thermal ionization mass-spectrometric Uranium-Thorium (TIMS U-Th) dating.
The Depth Column is approximate.

Source: References 928, 929, 930, and 933

## Table 2.5.1-210 (Sheet 1 of 3)
## Coordinates for Karst Features Presented in Figures 2.5.1-390 and 2.5.1-391

| Name | Legend | Latitude[(a)] | Longitude[(a)] |
|---|---|---|---|
| FGS-SIR-86-004 | Surface Sinkhole/Subsidence Feature | 25° 59' 23.97" N | 80° 10' 59.85" W |
| FGS-SIR-86-003 | Surface Sinkhole/Subsidence Feature | 26° 05' 39.87" N | 80° 07' 28.38" W |
| FGS-SIR-86-001 | Surface Sinkhole/Subsidence Feature | 26° 10' 26.63" N | 80° 07' 56.95" W |
| FGS-SIR-86-002 | Surface Sinkhole/Subsidence Feature | 26°10' 26.91" N | 80° 08' 18.39" W |
| FGS-SIR-87-001 | Surface Sinkhole/Subsidence Feature | 25° 57' 45.20" N | 80° 09' 59.07" W |
| Unnamed | Submarine Sinkhole | 24° 15' 29.98" N | 81° 55' 05.99" W |
| NR-1 | Submarine Sinkhole | 24° 13' 53.98" N | 82° 18' 11.99" W |
| Miami | Submarine Sinkhole | 25° 51' 29.98" N | 80° 01' 53.99" W |
| Marathon South | Submarine Sinkhole | 24° 15' 11.98" N | 80° 54' 17.99" W |
| Marathon North | Submarine Sinkhole | 24° 15' 23.98" N | 80° 54' 05.99" W |
| Key Biscayne | Submarine Sinkhole | 25° 42' 11.98" N | 79° 58' 35.99" W |
| Jordan, west lobe | Submarine Sinkhole | 24° 16' 23.98" N | 81° 02' 11.99" W |
| Jordan, east lobe | Submarine Sinkhole | 24° 16' 23.98" N | 81° 01' 53.99" W |
| Jordan East | Submarine Sinkhole | 24° 16' 05.98" N | 80° 58' 53.99" W |
| Cay Sal Bank | Submarine Sinkhole | 23° 54' 59.98" N | 80° 19' 59.99" W |
| Calypso Port 2 | Submarine Sinkhole | 26° 12' 05.38" N | 79° 59' 10.79" W |
| Calypso Port 1 | Submarine Sinkhole | 26° 07' 29.38" N | 79° 56' 43.19" W |
| Devils Punch Bowl | Spring | 25° 44' 48.54" N | 80° 12' 22.74" W |
| Tequesta | Spring | 25° 37' 03.62" N | 80° 18' 05.19" W |
| Montgomery Botanical Center | Spring | 25° 39' 45.40" N | 80° 16' 42.05" W |
| Coconut Grove | Spring | 25° 43' 31.02" N | 80° 14' 14.81" W |
| Wanless-Tedesco | Spring | 25° 35' 37.98" N | 80° 18' 16.49" W |
| SW Biscayne Bay (Gonzalez, 2004) | Spring | 25° 25' 32.98" N | 80° 19' 11.49" W |
| SITE 39 | Spring | 25° 36' 46.68" N | 80° 18' 07.69" W |
| SITE 38 | Spring | 25° 36' 46.98" N | 80° 18' 07.69" W |
| Ricisak Spring | Spring | 25° 36' 23.18" N | 80° 18' 25.79" W |
| BBS42 | Spring | 25° 36' 55.28" N | 80° 18' 19.19" W |
| BBS41 | Spring | 25° 36' 55.88" N | 80° 18' 20.39" W |
| BBS40 | Spring | 25° 36' 55.88" N | 80° 18' 19.49" W |
| BBS37 | Spring | 25° 36' 55.48" N | 80° 18' 17.89" W |
| BBS36 | Spring | 25° 36' 55.48" N | 80° 18' 17.39" W |
| BBS35 | Spring | 25° 36' 57.38" N | 80° 18' 20.99" W |
| BBS34 | Spring | 25° 36' 00.08" N | 80° 18' 08.59" W |
| BBS33B | Spring | 25° 36' 57.58" N | 80° 18' 10.59" W |
| BBS33A | Spring | 25° 36' 57.68" N | 80° 18' 10.39" W |
| BBS32 | Spring | 25° 36' 59.88" N | 80° 18' 09.69" W |
| BBS31 | Spring | 25° 36' 57.78" N | 80° 18' 16.99" W |
| BBS30 | Spring | 25° 36' 57.78" N | 80° 18' 17.29" W |

## Table 2.5.1-210 (Sheet 2 of 3)
## Coordinates for Karst Features Presented in Figures 2.5.1-390 and 2.5.1-391

| Name | Legend | Latitude[a] | Longitude[a] |
|---|---|---|---|
| BBS29 | Spring | 25° 36' 57.38" N | 80° 18' 17.29" W |
| BBS28 | Spring | 25° 36' 56.88" N | 80° 18' 18.19" W |
| BBS27 | Spring | 25° 36' 57.08" N | 80° 18' 17.89" W |
| BBS26 | Spring | 25° 36' 21.98" N | 80° 18' 28.19" W |
| BBS22 | Spring | 25° 36' 20.48" N | 80° 18' 27.39" W |
| N1 | Seismic Sag or Collapse Feature | 25° 33' 48.94" N | 80° 16' 45.20" W |
| N7 | Seismic Sag or Collapse Feature | 25° 32' 17.59" N | 80° 11' 42.04" W |
| N5 | Seismic Sag or Collapse Feature | 25° 28' 38.47" N | 80° 13' 49.46" W |
| N2 | Seismic Sag or Collapse Feature | 25° 28' 14.92" N | 80° 15' 45.52" W |
| EW4 | Seismic Sag or Collapse Feature | 25° 31' 41.16" N | 80° 12' 56.56" W |
| EKE2 | Seismic Sag or Collapse Feature | 25° 29' 02.30" N | 80° 06' 46.83" W |
| EKE1 | Seismic Sag or Collapse Feature | 25° 29' 21.38" N | 80° 08' 54.91" W |
| BBN1 | Seismic Sag or Collapse Feature | 25° 31' 47.66" N | 80° 09' 08.80" W |
| Government Cut | Seismic Sag or Collapse Feature | 25° 46' 49.36" N | 80° 10' 28.61" W |
| Jewfish Creek | Seismic Sag or Collapse Feature | 25° 11' 02.20" N | 80° 23' 10.01" W |
| Key Largo | Seismic Sag or Collapse Feature | 25° 08' 37.39" N | 80° 17' 49.13" W |
| NNRCOM3D | Seismic Sag or Collapse Feature | 26° 07' 11.62" N | 80° 19' 58.29" W |
| NNRC-A | Seismic Sag or Collapse Feature | 26° 05' 37.77" N | 80° 13' 31.83" W |
| L-35A | Seismic Sag or Collapse Feature | 26° 10' 01.06" N | 80° 19' 13.24" W |
| I595 | Seismic Sag or Collapse Feature | 26° 05' 17.82" N | 80° 12' 33.24" W |
| C-9-E | Seismic Sag or Collapse Feature | 25° 57' 09.78" N | 80° 11' 33.54" W |
| C-9-D | Seismic Sag or Collapse Feature | 25° 57' 24.78" N | 80° 21' 02.36" W |
| C-9-C | Seismic Sag or Collapse Feature | 25° 57' 24.69" N | 80° 21' 52.84" W |
| C-9-B | Seismic Sag or Collapse Feature | 25° 57' 24.61" N | 80° 23' 24.22" W |
| C-9-A | Seismic Sag or Collapse Feature | 25° 57' 24.40" N | 80° 23' 39.13" W |
| C-11-D | Seismic Sag or Collapse Feature | 26° 04' 04.38" N | 80° 10' 58.11" W |
| C-11-C | Seismic Sag or Collapse Feature | 26° 04' 04.11" N | 80° 11' 25.52" W |
| C-11-B | Seismic Sag or Collapse Feature | 26° 03' 54.62" N | 80° 14' 42.71" W |
| C-11-A | Seismic Sag or Collapse Feature | 26° 03' 42.24" N | 80° 24' 48.46" W |
| C-1-A | Seismic Sag or Collapse Feature | 25° 33' 06.55" N | 80° 21' 04.01" W |
| C-1-B | Seismic Sag or Collapse Feature | 25° 32' 55.19" N | 80° 20' 51.34" W |
| C-1-C | Seismic Sag or Collapse Feature | 25° 32' 44.45" N | 80° 20' 34.72" W |
| Hillsboro Sag 7 | Seismic Sag or Collapse Feature | 26° 19' 40.74" N | 80° 06' 02.95" W |
| Hillsboro Sag 6 | Seismic Sag or Collapse Feature | 26° 19' 40.35" N | 80° 07' 02.48" W |
| Hillsboro Sag 5 | Seismic Sag or Collapse Feature | 26° 19' 39.96" N | 80° 08' 02.01" W |
| Hillsboro Sag 4 | Seismic Sag or Collapse Feature | 26° 19' 40.49" N | 80° 12' 13.84" W |
| Hillsboro Sag 3 | Seismic Sag or Collapse Feature | 26° 20' 20.06" N | 80° 14' 25.93" W |
| Hillsboro Sag 2 | Seismic Sag or Collapse Feature | 26° 20' 30.75" N | 80° 15' 04.86" W |
| Hillsboro Sag 1 | Seismic Sag or Collapse Feature | 26° 20' 46.70" N | 80° 15' 57.05" W |
| NNRW26APR | Seismic Sag or Collapse Feature | 26° 07' 57.06" N | 80° 23' 03.49" W |

## Table 2.5.1-210 (Sheet 3 of 3)
## Coordinates for Karst Features Presented in Figures 2.5.1-390 and 2.5.1-391

| Name | Legend | Latitude[(a)] | Longitude[(a)] |
|---|---|---|---|
| Weeping Rock Cave | Cave | 25° 37' 24.05" N | 80° 18' 30.11" W |
| Strawberry Fields Cave | Cave | 25° 34' 15.89" N | 80° 23' 12.11" W |
| Stink Vine Cave | Cave | 25° 41' 00.55" N | 80° 16' 36.51" W |
| Smathers Cave | Cave | 25° 40' 02.10" N | 80° 16' 54.09" W |
| Root Cave | Cave | 25° 40' 54.39" N | 80° 16' 32.83" W |
| Rim Cave | Cave | 25° 39' 51.27" N | 80° 16' 43.09" W |
| Razor Rock Cave | Cave | 25° 37' 09.37" N | 80° 18' 19.67" W |
| Pathos Cave | Cave | 25° 39' 53.67" N | 80° 16' 42.86" W |
| Palma Vista Cave | Cave | 25° 24' 14.32" N | 80° 39' 09.41" W |
| Owaissa-Bauer Cave | Cave | 25° 31' 19.39" N | 80° 28' 07.32" W |
| Old Cutler Road Cave | Cave | 25° 39' 56.07" N | 80° 16' 46.28" W |
| Matheson | Cave | 25° 40' 54.28" N | 80° 16' 30.46" W |
| Hurricane Cave | Cave | 25° 37' 17.81" N | 80° 18' 32.07" W |
| Frango Fringe Cave | Cave | 25° 36' 20.92" N | 80° 19' 07.70" W |
| Floating Rock Cave | Cave | 25° 37' 23.99" N | 80° 18' 31.28" W |
| Fat Sleeper Cave | Cave | 25° 37' 26.12" N | 80° 18' 23.57" W |
| Devastation Cave | Cave | 25° 37' 15.73" N | 80° 18' 30.21" W |
| Cutler Hammock Cave | Cave | 25° 36' 30.04" N | 80° 19' 06.21" W |
| Crystal Pool Cave | Cave | 25° 40' 02.82" N | 80° 16' 55.14" W |
| Whispering Pines Cave | Cave | 25° 35' 29.34" N | 80° 20' 00.18" W |
| Large Sinkhole | Cave | 25° 37' 19.42" N | 80° 18' 16.87" W |
| Joint Cave | Cave | 25° 37' 16.40" N | 80° 18' 16.15" W |
| Florea Unnamed 2 | Cave | 25° 37' 26.56" N | 80° 18' 22.38" W |
| Florea Unnamed 1 | Cave | 25° 37' 25.53" N | 80° 18' 25.33" W |
| FIU Cave | Cave | 25° 37' 24.75" N | 80° 18' 27.36" W |
| Creepy Crawly Cave | Cave | 25° 36' 24.03" N | 80° 19' 10.36" W |
| Coon Cave | Cave | 25° 36' 20.25" N | 80° 19' 10.48" W |

(a) Coordinates are provided in NAD83 Geographic Coordinate System (latitude and longitude) with degrees, minutes, seconds format.

Data from References 951, 955, 958, 959, 989, 999, 1000, 1003, 1004, 1013, 1015, 1016, 1017, and 1021.