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Geology of Long Island, Bahamas: A Field Trip Guide

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Geology of Long Island, Bahamas: A Field Trip Guide

by

H. Allen Curran, John E. Mylroie, Douglas W. Gamble, Mark A. Wilson, R. Laurence Davis, Neil E. Sealey, and Vincent J. Voegeli

12th Symposium on the Geology of the Bahamas and Other Carbonate Regions

Gerace Research Center
San Salvador, Bahamas
2004
COVER PHOTO: The Holocene/late Pleistocene contact is sharp and well exposed at many places along the rocky windward (east) coast of Long Island. These cliffs are located immediately south of Coral Gardens (Stop 5). Here transgressive-phase carbonate eolianite beds of the North Point Member of the Rice Bay Formation overlie regressive-phase eolianite beds of the Cockburn Town Member of the Grotto Beach Formation. View is looking north toward Coral Gardens. Photo by Al Curran.
Figure 1. Index map for the Bahama Archipelago and locations of Long Island and San Salvador (modified from Curran and White, 1995).
INTRODUCTION

The purpose of this field guide is to describe and illustrate some of the well exposed and most interesting geological sites on the north end of Long Island, Bahamas. Long Island is one of the Bahamian “Family Islands” or “Out Islands,” and it has a population of about 4,000 inhabitants. The north end of the island lies about 270 km southeast Nassau, the capital city of The Bahamas, located on New Providence Island (Figure 1). Life on scenic Long Island is tranquil and community oriented. The island has a past history of salt production by the now defunct Diamond Salt Company. Sponging still operates on a small scale, and tourism and its related activities is growing.

Originally named Fernandina by Christopher Columbus, in honor of his Spanish king mentor, the island takes its present name from its geographic form. It truly is a long island, extending for about 120 km in length, but with a width of no more than 7 km at the widest point (Figure 2). The late Quaternary, all-carbonates geology of Long Island is varied and spectacular. This field trip is restricted to the north end of Long Island, given that only part of one day is available for our geologic reconnaissance. We hope that this field guide and trip will pique interest in the geology of Long Island and serve as a catalyst for future geologic investigations over the greater extent of the island.

Many of the sites we will visit are on private or restricted lands and special permission is necessary for entry to some of these areas. For these reasons, it may not be possible for someone who is using a copy of this field guide at a later date to follow precisely in our footsteps. Last-minute changes owing to weather conditions or time constraints may result in our field trip taking alternate routes and visiting a slightly different selection of sites than presented in the text. We are guests on Long Island, and all field trip participants are urged to treat the areas we visit with the utmost respect. It is illegal to undertake scientific research in The Bahamas without a research permit.

PLEASE DO NOT MAKE SAMPLE COLLECTIONS AT ANY STOP WITHOUT ASKING THE FIELD TRIP LEADERS IN ADVANCE. WE REQUEST THE SAME RESPECT FOR PRESERVATION OF THESE GEOLOGIC SITES FROM ALL SUBSEQUENT USERS OF THIS GUIDEBOOK.
REGIONAL SETTING

The Bahama Islands comprise a 1,000 km long portion of a NW-SE trending archipelago that extends from Little Bahama Bank off the coast of Florida to Great Inagua Island, just off the coast of Cuba (Figure 1). The archipelago extends farther southeast as the Turks and Caicos Islands, a separate political entity, and terminates with Silver Bank and Navidad Bank. The northwestern Bahama islands are isolated landmasses that project above sea level from two large carbonate platforms, Little Bahama Bank and Great Bahama Bank. To the southeast, beginning in the area of San Salvador Island, The Bahamas comprise small isolated platforms capped by islands that cover a significant portion of the available platform area. The Bahamian platforms have been sites of carbonate deposition since at least Cretaceous time, resulting in a minimum sedimentary cover thickness of 5.4 km (Meyerhoff and Hatten, 1974) and perhaps as much as 10 km (Uchupi et al., 1971). The large platforms to the northwest are dissected by deep channels and troughs (Figure 1), whereas the isolated platforms of the southeastern Bahamas largely are surrounded by deep water. Water depths on the platforms are generally less than 10 meters.

The origin of the Bahama platforms has been the subject of much debate, from which two main theories have evolved. Mullins and Lynts (1977) proposed a "graben" hypothesis, which explains the current configuration of the Bahama Archipelago as the result of plate tectonic motion that produced the opening of the Atlantic Ocean in the Mesozoic. The pattern of banks, troughs and basins is explained as resulting from an initial horst and graben pattern consistent with continental rifting. The competing theory is the "megabank" hypothesis (Meyerhoff and Hatten, 1974; Sheridan et al., 1981; Ladd and Sheridan, 1987), which holds that the modern Bahamas are a segmented remnant of a much larger and continuous Mesozoic carbonate platform. Recent work by Eberli and Ginsberg (1987), Mullins and Hine (1989; 1990), Melim and Masaferro (1997), and Manfrino and Ginsberg (2001) has demonstrated that the Bahama banks are undergoing both depositional progradation and erosional segmentation.


SUBAERIAL GEOLOGY OF THE BAHAMAS

The exposed rocks of The Bahamas are all late Quaternary carbonates, dominated by subtidal facies at low elevations and by eolianites at elevations above 6 m. Paleosols can occur at all elevations. The glacio-eustatic sea-level changes during Quaternary time alternately have flooded and exposed the Bahamian platforms, subjecting them to cycles of carbonate deposition and dissolution, respectively. Significant carbonate deposition has occurred in the past only when the platforms are flooded, as is increasingly the case today.

The carbonate sequences of the Bahamas can be viewed as individual packages deposited on each sea level highstand, separated by erosional unconformities (usually marked by paleosols) produced by each sea-level lowstand (Carew and Mylroie, 1995a; 1997). Each depositional package consists of three parts: a transgressive phase, a stillstand phase, and a...
regressive phase. These phases each contain a subtidal, intertidal, and eolianite component. Holocene sea level is sufficiently high so that the only marine deposits exposed on land today are those associated with the high stillstand phase of oxygen isotope substage 5e, about 130,000 to 119,000 years ago (Chen et al., 1991). At its maximum, sea level was about 6 m higher than at present. The transgressive and regressive marine deposits of substage 5e are below modern sea level, and the stillstand subtidal deposits of sea-level highstands prior to those of substage 5e also are not visible. Given isostatic subsidence rates of 1-2 m per 100,000 years (Mullins and Lynts, 1977; Carew and Mylroie, 1995b), earlier highstands were either not high enough, as for stage 7, or if high enough, occurred too long ago, as for stage 9 and earlier, to generate deposits above modern sea level. In contrast, eolianites form topographic highs that extend well above past and modern sea levels, so eolianites from several highstands are widely exposed on Bahamian islands.

During transgression, carbonate sediments are deposited when bank tops flood, and beach sediments are continually remobilized by the bulldozing action of the advancing sea. Large dunes are formed, which may be subsequently attacked by wave action as sea level continues to rise. Only the largest or most favorably positioned transgressive eolianites survive the rise of sea level to a maximum. During the stillstand of a sea-level high, subtidal and intertidal facies are deposited, but as the system reaches equilibrium, eolianite production is probably less than during transgression. During regression stillstand subtidal deposits are reworked by beach processes, and substantial eolianite packages can be formed. These regressive eolianites are abandoned by falling sea level. As sea level moves off the platform, erosional forces take over and soils are produced that will eventually be preserved as paleosols. It is important to recognize that during the Quaternary, the Bahamas have been in a sea-level lowstand condition as a result of glacioeustasy for about 85 to 90% of the time. The Quaternary carbonate units seen exposed in the Bahamas today represent deposition during that small fraction of the time when sea levels were high enough to flood the banks and turn on the carbonate sediment factory.

In the Bahamas, the most complete sequence of deposits representing a transgression, stillstand, and regression cycle is the depositional package formed during the oxygen isotope substage 5e event. Older packages are incomplete, for the reasons given earlier, and the Holocene package does not, as yet, contain a true regressive phase (although it does contain progradational regressive deposits).

A general model for the development of the stratigraphy of Bahamian islands was first proposed by Carew and Mylroie in 1985. This model was developed using San Salvador Island as the specific example, but it has been successfully transported to other islands by Carew and Mylroie (1989b) and by other workers (Wilbur, 1987, 1991; Kindler, 1995). The model has been modified with the accumulation of new data (Carew and Mylroie, 1989b; 1991; 1995a; 1997). A stratigraphy developed from this model is presented in Figure 3, and it is the basis for the stratigraphic
assignments and descriptions given in this field guide. This stratigraphy is based on field relationships, and does not require the use of geochronological tools, although it subsequently has been substantiated by a number of geochronologic methods (Carew and Mylroie, 1987). A spirited debate developed in the 1990s about Bahamian stratigraphy (see Carew and Mylroie, 1997, and references therein) centered on the reliability of amino acid racemization (AAR) analyses for making stratigraphic subdivisions in the absence of evidence provided by field relationships.

The eolianite packages older than oxygen isotope substage 5e were initially lumped together as the Owl's Hole Formation, although it was recognized at the time that this unit probably contained eolianites from a number of pre-substage 5e sea-level events (Carew and Mylroie, 1985). AAR data were subsequently used to subdivide the Owl’s Hole into multiple units on San Salvador Island (Hearty and Kindler, 1993). While AAR data were considered controversial, it was later established on Eleuthera Island that subdivisions of the Owl’s Hole could be demonstrated in the field (Kindler and Hearty, 1995; Panuska et al., 2002). Paleomagnetic analysis of the secular variation in paleosols also indicated that the Owl’s Hole could be subdivided into an upper and lower unit on San Salvador Island (Panuska et al., 1999). However, secular variation analysis is not field evidence, and there are no subtidal units in this formation. The eolianites of this unit are predominantly bioclastic, and ooids are extremely rare. The Owl's Hole Formation is usually recognized in the field by its relationship to overlying deposits. Efforts to subdivide units in the Bahamas by petrologic methods have been attempted (e.g. Kindler and Hearty, 1996; 1997), but demonstrated petrologic variability among single Pleistocene units, as well as in Holocene units, indicates that this technique is not reliable.

Overlying the Owl's Hole Formation, and separated from it by a paleosol or other erosion surface, is the Grotto Beach Formation. This formation was deposited during oxygen isotope substage 5e, and it consists of two members.

The older French Bay Member is a transgressive eolianite (Carew and Mylroie, 1985; 1997). In some places, transgressive eolianites are marked by an erosional platform on which later stillstand fossil corals are found (Halley et al., 1991; Carew and Mylroie, 1995a). The Cockburn Town Member is a complex array of stillstand subtidal and intertidal facies overlain by regressive eolianites. In earlier versions of the Carew and Mylroie stratigraphic model (Carew and Mylroie, 1985; 1989b), the Grotto Beach Formation also contained the Dixon Hill Member, thought to represent an eolianite deposited during oxygen isotope substage 5a about 85,000 years ago. This member, based solely on amino acid racemization data, proved incorrect (Carew et al., 1984), and it was subsequently eliminated from the stratigraphy (Carew and Mylroie, 1995a; 1997). During Grotto Beach time, ooids were produced in great numbers, and the vast majority of eolianites in the Grotto Beach Formation are either oolitic or contain appreciable ooids.

Another debate concerning Bahamian geology involved the existence of oxygen isotope substage 5a eolianites (see Carew and Mylroie, 1997; Kindler and Hearty, 1997 and references therein). As noted above, AAR data were used to create a substage 5a eolian unit in 1985 (Carew and Mylroie, 1985), but that unit was dropped when field work demonstrated that the unit in question was older than substage 5e (Carew and Mylroie, 1997), and laboratory tests indicated use of the land snail Cerion for AAR work was extremely unreliable (Mirecki et al., 1993). However, use of whole rock AAR analysis was argued to avoid this problem and was used to identify substage 5a eolianites on a number of Bahamian islands (Kindler and Hearty, 1997 and references therein), including Long Island. No field evidence has been established to demonstrate, one way or the other, that substage 5a units exist in the Bahamas. However, AAR data have been used to identify substage 5a eolian units on Bermuda (Vacher and Hearty, 1989).

Overlying the Grotto Beach Formation and separated from it by a paleosol or other
erosion surface are the rocks of the Rice Bay Formation, deposited during Holocene time. The Rice Bay Formation also is divided into two members, based on their depositional history relative to Holocene sea level. The North Point Member consists entirely of eolianites, whose foreset beds can be followed at least 2 m below modern sea level in some places. Whole rock carbon-14 measurements from the North Point Member indicate ages centered around 5,000 yBP. Laterally adjacent, but rarely in an overlying position, is the younger Hanna Bay Member. This unit consists of intertidal facies and eolianites deposited in equilibrium with modern sea level. The eolianites have a radiocarbon age centered around 3,000 yBP, but some beach rock has ages as young as 400 yBP (Carew and Mylroie, 1987). While weakly developed ooids have been reported from the early stages of North Point Member deposition (Carney and Boardman, 1991), the Rice Bay Formation exposures on San Salvador are predominantly peloidal and bioclastic. The North Point Member is now being attacked by wave erosion. Sea caves, talus, and coral communities on wave-cut eolianite benches of the North Point Member exist, as mentioned earlier for the rock record of the French Bay Member of the Grotto Beach Formation

**GEOLOGY OF LONG ISLAND**

Although field work in 1990 demonstrated the many aspects of the karst geology of Long Island (Mylroie et al., 1991), systematic study of the overall carbonate geology of the island has not yet been done. This field guide is the first published use of a Bahamian geologic model on Long Island. The field trip stops will show examples of the Rice Bay Formation, including an excellent exposure of the Hanna Bay Member/North Point Member contact, not commonly reported elsewhere in the Bahamas. The Rice Bay Formation/Grotto Beach Formation contact is well developed and frequently exposed on Long Island, with the North Point Member generally being the Holocene unit overlying Pleistocene rocks. The Pleistocene unit in most such cases of Holocene/Pleistocene contact is believed to be the regressive-phase eolianites of the Cockburn Town Member of the Grotto Beach Formation. Identification of regressive- versus transgressive-phase eolianites can be determined following criteria established by Carew and Mylroie (1995; 2001), as shown in Table 1.

The Grotto Beach Formation/Owl's Hole Formation contact has not been located, and the existence of Owl's Hole rocks is only indirectly demonstrated by virtue of hosting mature karst features. The trip stops will follow the stratigraphic column of Figure 3 downward, starting with Holocene units and then moving to the late Pleistocene units. A brief one-day field trip simply cannot cover the geology of the entire island, but we think that the stops selected for this trip do offer a representative overview of the Quaternary geology of Long Island. Furthermore, we hope that these stops will spur comment and discussion and will demonstrate how the stratigraphic model presented earlier can be applied to Long Island.

**TABLE 1. PHASES OF DEPOSITION AND DIAGNOSTIC CHARACTERS**

<table>
<thead>
<tr>
<th>Phase of Deposition</th>
<th>Transgressive Phase</th>
<th>Stillstand Phase</th>
<th>Regressive Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fine-scale eolian bedding</td>
<td>Disrupted eolian bedding</td>
<td>Disrupted eolian bedding</td>
<td></td>
</tr>
<tr>
<td>Few vegemorphs</td>
<td>Abundant vegemorphs</td>
<td>Extensive vegemorphs</td>
<td></td>
</tr>
<tr>
<td>Penecontemporary cliffting and boulder paleotalus</td>
<td>Penecontemporary notching of beach and intertidal facies, and beach-face breccia facies</td>
<td>Lack of penecontemporary wave erosion</td>
<td></td>
</tr>
<tr>
<td>Penecontemporary sea caves</td>
<td>Rare sea caves</td>
<td>Lack of sea caves</td>
<td></td>
</tr>
<tr>
<td>Corals on wave-eroded benches</td>
<td>No corals on eroded benches</td>
<td>No penecontemporary benches</td>
<td></td>
</tr>
<tr>
<td>Lack of protosols</td>
<td>Protosols common</td>
<td>Protosols common</td>
<td></td>
</tr>
<tr>
<td>On lapped still-stand or regressive-phase deposits</td>
<td>Marine facies abundant</td>
<td>Commonly peloidal/bioclastic</td>
<td></td>
</tr>
<tr>
<td>Predominantly eolianites, marine deposits rare</td>
<td>Ebb-tidal delta, lacustrine, and strand plain deposits</td>
<td>Eolianites overstepping marine deposits</td>
<td></td>
</tr>
</tbody>
</table>
Carbonate Island Karst Model

<table>
<thead>
<tr>
<th>Common attributes that distinguish Island karst from karst of interior settings</th>
<th>Distinct geomorphic types that distinguish carbonate islands from one another</th>
</tr>
</thead>
<tbody>
<tr>
<td>The karst is eogenetic.</td>
<td>Simple Carbonate Islands: Non-carbonate rocks remain below the zone of fresh-water influence, and recharge is exclusively autogenic. (Fig. 4a)</td>
</tr>
<tr>
<td>Dissolution is enhanced at the surface, bottom, and margin of the freshwater lens by mixing of waters and trapping of organic materials at these boundaries.</td>
<td>Carbonate Cover Islands: Non-carbonate basement rocks deflect percolating water and partition the fresh-water lens. (Fig. 4b)</td>
</tr>
<tr>
<td>Glacio-eustatic sea-level fluctuations impose dissolutional and diagenetic imprints reflecting the vertical migration of the lens.</td>
<td>Composite Islands: Non-carbonate basement exposed at the surface, and allogenic recharge is delivered to insurgences on the contacts. (Fig. 4c)</td>
</tr>
<tr>
<td>Tectonic uplift and subsidence overprints the glacio-eustatic imprints with additional dissolutional and diagenetic imprints, as well as structural modifications</td>
<td>Complex Islands: Interfingering of carbonate/non-carbonate facies and faulting combine to produce complex aquifer features. (Fig. 4d)</td>
</tr>
</tbody>
</table>

Table 2. The Carbonate Island Karst Model (CIKM). Modified from Vacher and Mylroie, 2002.

FRESHWATER LENS HYDROLOGY AND KARST PROCESSES

In any essentially homogeneous body of rock like that of the carbonates forming the Bahamian islands, the freshwater lens floats on underlying, denser seawater that permeates the subsurface. The model for the ideal behavior of such water masses is the Ghyben-Herzberg-Dupuit model. In reality, variations in rock permeability and other factors result in distortion of the ideal lens shape (e.g. Vacher and Bengtsson, 1989). Nonetheless, the Ghyben-Herzberg-Dupuit model serves as a useful first approximation of the relationship between the freshwater and underlying marine groundwater in an island.

During past higher stands of the sea, the fresh groundwater lens in each island was as high or higher than it is today. Beneath the surface of those past freshwater lenses, within the limestone rock of the islands, caves were produced by dissolution. Each time sea level fell, the caves became abandoned and dry. Under today's climatic conditions the Earth is warm and sea level is relatively high, but not quite as high as at some times in the past. We can therefore enter dry caves today throughout The Bahamas. In contrast, the blue holes of The Bahamas lead into caves that are flooded by seawater. These blue holes represent the cumulative dissolution and collapse that has occurred during many sea-level fluctuations. The complexity of cave passages found in blue holes is the result of overprinting of repeated marine, freshwater, and subaerial conditions throughout Quaternary time. Conversely, the presently dry caves of The Bahamas formed during the relatively short time periods of the late Pleistocene when sea level was higher than at present. Bahamian caves that formed above modern sea-level elevation prior to oxygen isotope substage 5e time today lie below modern sea level owing to isostatic subsidence of the platforms (Carew and Mylroie, 1995b). Taking isostatic subsidence into account, sea level was high enough to produce the observed subaerial caves for a maximum of about 11,000 years of the oxygen isotope substage 5e time period. In addition, during that sea-level highstand, only the eolian ridges and a few beach and shoal deposits stood above sea level, and island size in The Bahamas was dramatically reduced compared to that of today's islands. As a consequence, freshwater lens volumes and discharges were comparably reduced. An end result of this scenario is recognition that dry Bahamian caves represent development during a very short time period within small freshwater lenses and with minimal overprinting by later events. Any model that attempts to explain development of these caves must operate under these tight constraints of time and space.
The Bahamas were the starting point for the development of what has become the Carbonate Island Karst Model (CIKM). The salient points of this model are shown in Table 2 and Figure 4. The key aspects are that cave and karst development in carbonate islands is very different from that found in carbonate rocks of continental interiors, where most such research has been done. Basically, karst development under the CIKM is controlled by the youthful age of the rock involved (almost always Cenozoic, commonly Quaternary), the dissolutorial aggressivity provided by mixing of freshwater and seawater, and the change in sea level created by glacio-eustasy and tectonics. Island configuration, especially as regards carbonate and non-carbonate rocks (Figure 4) is also crucial. This last item has little meaning in The Bahamas, as these islands are 100% carbonates.

The youthfulness of Bahamian carbonate rocks creates different water-flow dynamics than are found in the dense, diagenetically-mature carbonates of continental interiors. The Bahamas exemplify what has been described as eogenetic karst, defined by Vacher and Mylroie (2002, p. 183) as “the land surface evolving on, and the pore system developing in, rocks undergoing eogenetic, meteoric diagenesis.” The term “eogenetic” was taken from Choquette and Pray (1970, p. 215) who defined the three time-porosity stages of carbonate rock evolution: “the time of early burial as eogenetic, the time of deeper burial as mesogenetic, and the late stage associated with long-buried carbonates as telogenetic.” Eogenetic carbonate rocks have not been extensively compacted or cemented and retain much of their primary depositional porosity. Most carbonate islands, and almost all carbonate islands found in tropical or subtropical locations, are made up of eogenetic limestones (Late Cenozoic) that were deposited proximal to the setting in which they presently occur. The term eogenetic karren has been applied to the unique etching and dissolution of surface carbonates on carbonate islands and coasts (Taborosi et al., in press), which encompasses inland forms as well as the traditional “phytokerst” of Folk, et al. (1973). Taborosi et al. (in press) also review the various terms and mechanisms proposed over the years to describe and explain such karren, providing updated interpretations.

Island karst has been defined as that which forms under the constraints of the CIKM, whereas karst that develops in the interior of islands, removed from CIKM controls, is karst on islands (Vacher and Mylroie, 2002). For example, the caves and karst found in The Bahamas is island karst, but the cockpit karst of Jamaica, or the Mogote karst of Cuba and Puerto Rico, is karst on islands, as it differs little from what would be found in a tropical, continental interior karst.
The term "blue hole" has been used in a variety of ways. A complete review of the history of blue hole studies, and the various uses of the term, can be found in Mylroie, et al. (1995). Blue holes are defined as: "subsurface voids that are developed in carbonate banks and islands; are open to the earth's surface; contain tidally-influenced waters of fresh, marine, or mixed chemistry; extend below sea level for a majority of their depth; and may provide access to submerged cave passages." (Mylroie, et al. 1995, p. 225). Blue holes are found in two settings: ocean holes open directly into the present marine environment and contain marine water, usually with tidal flow; inland blue holes are isolated by present topography from marine conditions, and open directly onto the land surface or into an isolated pond or lake, and contain tidally-influenced water of a variety of chemistries from fresh to marine. The most common alternative use of the term “blue hole” is to describe large and deep karst springs (Mylroie et al., 1995).

In the northwestern Bahamas, blue holes with depths in the 100-125 m range are common, and it was thought that their depth was limited by the position of the lowest glacial sea-level lowstand, which was about 125 m below present sea level. However, exploratory wells commonly intersect voids below that depth (e.g. depths of 21 to 4082 m; the deepest of these voids was large enough to accept 2,430 m of broken drill pipe [Meyerhoff and Hatten, 1974]).

Dean’s Blue Hole on Long Island is known to be over 200 m deep, ending in a vast chamber (Wilson, 1994). Blue holes commonly lead into major horizontal cave systems, such as Lucayan Caverns on Grand Bahama Island, and Conch Blue Hole on North Andros Island.

This field trip will visit a single blue hole of relatively average characteristics (Stop 7). Dean’s Blue Hole is too far south, and too remote, for access by this field trip. Blue holes are varied and interesting, and there has been much discussion about their origin, definition, and exploration. Such discussions likely will continue during this field trip and at the 12th Geology Symposium. The field trip also will visit a flank margin cave, which developed in the discharging margin of a past, higher elevation freshwater lens, just under the flank of the enclosing eolian ridge (hence the name “flank margin cave”). Eogenetic karren will be seen at a number of field trip stops, both inland and coastal, and will be briefly presented and discussed. Long Island has some of the largest known flank margin caves in the Bahamas, such as Salt Pond Cave and Hamilton’s Cave. These caves exhibit large chambers, some of which are intact (Figure 5) and some that are segmented (Figure 6). A complete discussion of these caves, and others is given in Mylroie et al. (1991).
DESCRIPTIONS OF THE FIELD TRIP STOPS
(All field trip stop locations appear on the maps of Figures 7 and 8 and on the map of the inside back cover of this guidebook)

Stop 1 – Holocene Rocky Coast Exposures:
From the airport, drive east to the coast along Ocean View Drive, using the first east-heading road north of the airport, and proceed south almost 2 km to where the road makes a sharp bend to the right (west). Here a less-used side road joins from the south at right angles. Follow this side road for a short distance (~75-100 m) to its end at a small cul-de-sac; a sign “Beaches” will be seen and a narrow path leads down to the coast. Move along the path in single file, and try not to bunch up. Just before the path turns and becomes steep leading down to the coast, there is a good view to the west across the island and to the south along the coast. The relief developed in the fossil-dune topography here is quite impressive. Take care along the steep portion of the path, and make every effort not to loosen debris that will be a hazard to those below.

After reaching the beach, move south to the area of cliff exposures. There is a seaward projection of the cliffs here, and, if the tide is high, easy passage around this point likely will not be possible. In any case, we will not be walking south of this small point. Exploration of the point will reveal the presence of a small sea cave here, with a partial filling of well-cemented breccia. This entire section of windward coast is one of significant interest for its rock exposures and also as a natural laboratory for observation of the dynamics of both physical and biological Bahamian-style coastal processes.

You may have noticed that there was no sign of any significant paleosol layer on the outcrop surfaces at the top of the cliffs or along the pathway coming down to the beach. Following the Carew and Mylroie stratigraphy (Figure 4), this indicates that the rock exposures here are of Holocene age. Given that some of the dip directions are seaward and these dipping beds extend below sea level (another good example will be seen a bit farther to the north), these eolianite beds can be assigned to the North Point Member of the Rice Bay Formation. In detail, the lamination of the beds and their other mesoscale features are similar to those described by White and Curran (1988) from North Point on San Salvador, the type area of the North Point Member. However, the lowermost beds of the outcrop area shown in Figure 9 are atypical in
that they do not show strong eolian bedding, displaying instead a general lack of bedding and a spongioform texture. Furthermore, these beds contain distinctive vertical and long cylindrical holes of several to 15 cm or more in diameter, and the beds erode to create an overall pillar-like form (Figure 10). A similar example of this North Point Member facies will be seen at Stop 3, and further discussion is presented with that stop description. The details of the subenvironment and processes of deposition represented by these beds have not yet been fully developed and should provide a point for lively discussion during the field trip.

Now head north along the coast and around a rocky headland. Here a series of planar beds can be seen dipping asymptotically toward modern sea level (Figure 11). Again, the lack of a paleosol on this surface indicates that the rock is Holocene in age, and its congruency with modern sea level identifies the unit as belonging to the Hanna Bay Member of the Rice Bay Formation. Ripples are well preserved (Figure 12), along with other mesoscale bedding features. This surface also commonly exhibits troughs with lithified breccia-block fill (Figure 13). The modern processes of coastal erosion and deposition that can be observed along this section of coast provide good explanations for many of the features found preserved within this Hanna Bay Member surface.

Continuing north a few tens of meters, it can be seen that the Hanna Bay Member overlies another unit dipping landward, with the planar Hanna Bay beds lying unconformably on truncated eolianite beds. There is no paleosol at the unconformity, and the beds of the lower unit dip below modern sea level. The lack of a paleosol on top of the lower unit is again an indication of Holocene age, and the dip of the eolian beds below sea level confirms identification of the North Point Member of the Rice Bay Formation. The lower Hanna Bay beds contain clasts of North Point material as a rubble layer at the contact. A short distance north from this point, a field of boulders, mostly from the Holocene units, lies at the base of the Holocene rock cliffs. Farther north, the unconformity
Figure 11. Extensive exposure of the Holocene planar beds of the Hanna Bay Member at Stop 1. Photo by Mark Wilson.

Figure 12. Wind ripples preserved in the planar Hanna Bay Member beds, Stop 1. Scale = 15 cm. Photo by John Mylroie.

Figure 13. (Above) Breccia-block fill in troughs of the Hanna Bay Member beds, Stop 1. Pen = 15 cm. Photo by Al Curran.

Figure 14. (Right) North end of Stop 1, with seaward-dipping North Point Member beds in foreground and boulder field and Holocene/Pleistocene contact in background. Photo by John Mylroie.

Figure 15. Numerous specimens of fossil Cerion preserved in the paleosol at Stop 2. Scale = 15 cm. Photo by John Mylroie.

Figure 16. Root casts or rhizomorphs at Stop 2. Scale = 15 cm. Photo by John Mylroie.
between the overlying North Point Member and the underlying Grotto Beach Formation (probably the Cockburn Town Member) can be seen. The paleosol separating the Holocene from the late Pleistocene is obvious at this point. This reach of coast can be traversed, but it is not suitable for group transit (Figure 14). Take pictures, and from this point we will backtrack a short distance and follow the narrow trail west up and over the crumbly rock cliffs and dunes to the road where we will re-board the bus.

Stop 2 – Cerion Graveyard: After re-boarding the bus, we will travel north about 200 m to another “Beaches” sign. Follow the path down to the beach and rocky coast. This stretch of coast consists of late Pleistocene exposures. A house high up on a cliff to the south can be seen here.

Immediately after leaving the trail and heading south, a depression bordered by a deep red paleosol will be encountered. Within this depression are thousands of well-preserved fossil specimens of a large species of the land snail Cerion (Figure 15), along with numerous root casts (Figure 16).

Cerion was the favorite research organism of the late Stephen Jay Gould. Along with his students, Gould spent over 20 years in field work and laboratory study in order to reassess the patterns of geographic and temporal variation of Cerion throughout the Bahamas. A Natural History magazine essay by Gould (1996) related in fascinating detail how the distinctive morphologic variation within the genus Cerion could have been used to determine the island of Columbus’ first landfall in the Bahamas if only he had returned and archived in Spain a small collection of Cerion shells. Gould and Glenn Goodfriend, his former student (also recently deceased), did field work together on Long Island in this very area shortly before their respective untimely deaths. They were studying Cerion shell-form variation from specimens collected from a series of pits excavated in the un lithified dune sands along the windward coast of the island. Goodfriend was to have presented a paper on this work at the 11th Geology Symposium (Goodfriend et al., 2002), but he was unable to attend owing to illness at that time. A thorough review of the significance of Cerion in The Bahamas was published by Steven Jay Gould in the proceeding volume for the 8th Geology Symposium (Gould, 1997), which he attended as keynote speaker.

The rocky headland to the south is a continuation of the headland that was north of Stop 1. The Holocene/Pleistocene unconformity is again obvious here. The abundant root casts associated with the paleosol at this outcrop (variously called rhizomorphs, rhizcretions, or vegemorphs) are indicative of regressive-phase eolianites (Table 1). The underlying eolianites are most likely Cockburn Town Member of the Grotto Beach Formation, but they possibly could be older regressive dunes associated with the Owl’s Hole Formation. Below the house to the south, the paleosol-encrusted rocks form a small headland. Wave erosion has stripped out weaker rock from beneath the paleosol, forming a paleosol cave, a type of sea cave. Return to the bus and re-board.

Stop 3 – Stella Maris Cabana Beach: The bus will continue north for about 300 m to the Stella Maris beach where the ocean-side pool and cabanas are located. Exit the bus, walk to the coast, and stroll south for a few meters.

This late Pleistocene outcrop area demonstrates inversion of topography. Here
there are numerous little pillars, from 10 cm to 1 m high. All are capped by dish-shaped crusts of resistant paleosol (Figure 17). Prior to the Holocene rise in sea level, this region was an epikarst or karren surface, with numerous small dissolution pits. These pits collected soil, which hardened into a resistant paleosol. As waves initially commenced eroding this area ~3,000 years ago, the surface began to be stripped away. The paleosol crust was thickest and most resistant at the bottom of the small pits, and so withstood the erosional stripping of the surface around them. As a result, the topography is inverted, and what was the bottom of small pits now are the tops of small pillars. A number of large boulders also are present in this area. Several of the bigger ones exhibit a thick rind of vermetid-snail encrustation, along with the remains or traces of other encrusters, indicating that the boulders once were subtidal or intertidal and have been transported and deposited by wave action.

Head back north across the cabana beach and move past the boulder rip-rap to where true outcrop begins. This exposure presents a great close-up view of the late Pleistocene/Holocene contact, as shown in Figure 18A and in the stratigraphic column of Figure 19. Notice that the late Pleistocene beds contain numerous fossil Cerion shells identical to those found at Stop 2.
developed vertical, cylindrical holes (Figure 20). These cylindrical holes have inside diameters of up to 12 cm and lengths of up to 1.3 m. They are very similar to structures reported in equivalent Holocene beds on Lee Stocking Island (White and Curran, 1993; Kindler, 1995). These cylindrical holes do not appear to be solutional in origin. White and Curran (1993, p. 186-187) speculated that the Lee Stocking Island structures represented palm tree-trunk molds. They also reported finding palm frond imprints in eolianite talus blocks on Lee Stocking Island, and they cited supporting evidence for this interpretation from similar structures found on Bermuda and interpreted as palm tree molds. The size, close spacing, and host rock characteristics of the structures here and those on Lee Stocking Island are similar.

However, one real difference is that the structures on Lee Stocking Island occur in the top layers of North Point Member-equivalent beds whereas here they are at the base of the section. An initial interpretation for Units 3 and 4 of this section is that Unit 3 represents a protosol layer upon which a coastal coppice palm community developed. Sands from the leeward sides of encroaching dunes intercalated the palm-dominated community, with the trunks of the palms molded by sand. Rapid cementation of the sands by meteoric waters and decay of the palm trunks created the vertical cylindrical structures and spongioform texture of Unit 4. Beds lying above Unit 4 have typical eolianite lamination and contain numerous rhizomorphs and represent the continued migration of dunes over the coastal palm community. Problems with this interpretation are that there apparently are no root ball structures preserved from the palm trees, and there is little or no evidence of preservation of other types of woody plants that presumably would have co-existed with the palm trees. Further study is warranted for these unusual and distinctive vertical structures and the sedimentary units containing them.

This outcrop extends farther north across the Stella Maris property, and some interesting exposures are present in the man-made cuts of
the saltwater wading pool and beyond. We will not have time today to investigate these exposures, so save them for your next trip to Long Island. Return to the bus and re-board. Depending on the time, we will make our lunch stop next or immediately after Stop 4.

**Stop 4 - Stella Maris Party Cave:** From Stop 3, continue north to the second left-hand turn heading west and uphill (the first left-hand turn is not a through street). Go uphill and take the second left, and about 100 m to the south, on the west side of the road, is a sign marked “Party Cave” and a place to pull off the road. Exit the bus and follow the paved path downhill to the south, and continue as it hooks left (west) and leads to a large cave entrance.

While on the path, note the many dissolution holes and pits in the rock on either side of the path. This is the epikarst or surface karren, the result of meteoric water dissolving the upper few meters of rock. There is debate about how much of this rock exposure is natural, and how much is the result of soil loss through agricultural practices (mostly slash and burn). Clearly the rock looks like Swiss cheese, and gives the impression of high porosity at depth. But the greater part of this porosity disappears with in a meter or two of the surface. This can be observed in the cave, as the ceiling is largely solid without nearly as many holes as were seen on the surface a few meters above. The epikarst is a thin but very permeable region that acts to create a perched aquifer. Vadose flow is by percolation, or by moving laterally and descending through vadose fast-flow routes, or pit caves. A few of these features can be seen entering the cave ceiling.

The “Party Cave” hosts the weekly Stella Maris Resort barbecue. In 1990, it was mapped as Stella Maris Cave (Figure 21). When first entering the cave, a small boat is seen, and it serves as the party bar (Figure 22).

This is a classic flank margin cave, with a single globular main chamber, along with short passages off the sides and rear that quickly end. The cave was a mixing chamber, and water entered and exited as diffuse flow, while fresh water/sea water mixing created substantial dissolution at the lens margin. The position of the cave in the distal margin of the fresh-water lens means that it formed under the flank of the
Figure 23. Holocene North Point Member beds overlying late Pleistocene Cockburn Town Member rocks at Stop 5. Photo by John Mylroie.

Figure 24. Fossil Diploria clivosa coral colony in subtidal beds of the Cockburn Town Member, Stop 5. Scale = 15 cm. Photo by John Mylroie.

Figure 25. Beds of the CTM Member subtidal facies overlying truncated eolianite beds, probably representing the French Bay Member, Stop 5. Photo by John Mylroie.

Figure 26. Overview of the blue hole, Stop 6. Photo by John Mylroie.

Figure 27. The Pleistocene eolianite ridge of Stop 7. Photo by John Mylroie.

Figure 28. Exposure of Holocene North Point Member eolianite, viewed from the coast on the east side of Columbus Monument hill. Photo by John Mylroie.
land mass. With present sea-level conditions, the cave is dry, and the cave's position made it vulnerable to hill slope and scarp retreat, which have breached the cave to the surface. Without this secondary erosion, the cave would have no natural entrance.

Flank margin caves are excellent indicators of sea-level position, as the freshwater lens is controlled by the location of sea level. As in most caves, secondary subaerial calcite deposits, primarily stalagmites, are found, and they can be used for age-dating by the U/Th method. These data give limits as to when the cave could have formed (e.g. Carew and Mylroie, 1995b). Stable isotope analysis also can provide information on past climate conditions.

Because the Stella Maris Cave had to have formed on a previous glacio-eustatic sea-level highstand, the eolianite the cave is developed in had to be already present. For many reasons (e.g. Carew and Mylroie, 1995b), the cave most likely formed during oxygen isotope substage 5e. The eolianite is therefore probably Owl's Hole Formation, although it could be French Bay Member of the Grotto Beach Formation. In the latter case, the eolianite would have formed on the transgression, the freshwater lens would have entered the eolianite as sea level reached its peak, pushing the freshwater lens into the dune, and forming the cave by dissolution. Exit the cave and return along the path to re-board the bus.

Stop 5 - Coral Gardens: From the "Party Cave", reverse direction, return to Ocean View Drive, and turn left (north) onto the drive. Follow the coast road north for about 1 km; after about 2/3rds of a km the road will angle inland to the northwest. Go past one turn to the right, take the second turn to the right, and head northeast back towards the coast. The road ends at a T-intersection. Exit the bus, and follow the trail down to the coast to the Coral Gardens (a fringing offshore reef).

The rocky bench here is much like what was seen at Stops 1, 2, and 3. Exposed are Holocene dunes (North Point Member) overlying late Pleistocene rocks (Figure 23 and the guidebook cover photo). In this case, it is clear that these rocks are Cockburn Town Member subtidal facies and regressive-phase eolianites. At the shoreline cliff, in isolated localities, are fossil coral colonies (Figure 24); West Indian top shells are also present, along with other molluscan fossils. Subtidal facies above modern sea level in the Bahamas are virtually always indicative of the Cockburn Town Member of the Grotto Beach Formation. A few meters to the south, the subtidal units can be seen to overlie a truncated eolianite (Figure 25). The absence of a paleosol at this locality suggests that the lower eolianite is the French Bay Member of the Grotto Beach Formation, but the wave action that planed off the eolian surface possibly could have removed a paleosol, in which case the lower unit would be Owl's Hole Formation eolianite. Return to the bus and re-board.

Stop 6 - Blue Hole: The bus will head a short distance north (about 150 m) and take a right turn onto a dead end road. From the dead end, a trail leads down slope to the edge of a large blue hole (Figure 26). This trail is steep in places, and space at the bottom is small, so maneuver carefully to fit everyone in.

This is a large blue hole surrounded by low rock cliffs. The cliffs are highest to the north side, and the deepest point is below the highest cliff. The salinity is normal marine. This blue hole has many fish, and also turtles, which indicates a connection to the sea, although the turtles likely have been introduced. As noted in the Introduction, blue holes are polygenetic, forming from flooding by glacio-eustasy of vadose pit caves, by collapse of deep-seated dissolutional chambers, or by bank-margin failure. The large size and circular shape of this blue hole indicate that it most likely formed by collapse of a deeper void, with subsequent retreat of the bedrock perimeter. Blue hole water chemistry can be complex, from fresh to marine to hypersaline. Blue hole biota can be similarly complex, especially in terms of microbial activity. Scramble back up the trail, and re-board the bus.
Figure 29. View looking east from the top of Columbus Monument. The ridge across the channel is Holocene North Point Member eolianite. Photo by John Mylroie.

Figure 30. The Columbus Monument in 1990. The inscription on the plaque reads “Dedicated to the gentle, peaceful, and happy aboriginal people of Long Island, the Lucayans, and to the arrival of Christopher Columbus on Oct. 17, 1492.”

Stop 7 – Pleistocene Eolianite Ridge: Return to Ocean View Drive, and continue west to the west coast, turning right (north) on the Government Main Road. Head north about 1 km, and stop at a road cut into an eolianite ridge (Figure 27). WATCH OUT FOR TRAFFIC.

This eolianite shows classic rollover of the beds from south to north. The road cut exhibits dissolution pits, both small and large, in the epikarst at the surface. The outcrop also indicates how root casts and bulldozer grooves can confuse the observer about the true orientation of the bedding, in this case dipping left to right. Note the wall at the top of the cut. It was built into the pits. These pits show how paleosol can preferentially collect and harden, generating the resistant layer that created the inverted topography, as seen at Stop 3. Eolianites like this are difficult to place in a stratigraphy when they occur in isolated locations such as this stop. For this reason, much effort has been invested in using AAR, secular variation, and petrologic studies to assign ages to these otherwise problematic carbonate rocks, but results have yielded mixed success. Accordingly, they are commonly mapped as “undifferentiated Pleistocene.” Re-board the bus.

Stop 8 – Columbus Monument: Continue north on the Government Main Road for 7 km, passing through the communities of Burnt Ground, Glenton Settlement, and into Seymour Settlement. As the road goes steeply uphill and right (east) to enter Seymour Settlement, a dirt road heads north to North End and the Columbus Monument. After a little over 2 km, the road ends at North End. Leave the bus, and walk to the coast at the right. From here, a beautiful exposure of North Point Member eolianite can be viewed immediately to the north (Figure 28). Retrace your steps to the steep trail leading to the top of the dune and the Monument.

From the dune top, looking east, one can see across a channel to a ridge of North Point eolianite (Figure 29). Beyond, in the distance, is a second ridge which is made up of late Pleistocene eolianites, most likely regressive-
phase beds of the Cockburn Town Member. Looking west, the Columbus monument stands forlorn. In 1990, it looked much better (Figure 30). Farther to the west, lower eolianites with a paleosol surface can be seen (Figure 31); this is the Pleistocene foundation upon which the North Point Member eolianites rest. The significance of this outcrop is that it gives evidence of the prodigious allochem production that occurred as the Bahama banks were initially flooded coming out of the last glaciation, as the carbonate factory was turned on. All of the North Point Member eolianite seen here was produced in at most 2,000 years. Return to the bus and re-board.

Stop 9 – Stella Maris’ Santa Maria Beach Reserve: Return to the main paved road in Seymour Settlement. Drive a short distance south (about 0.75 km) and make a right turn to the west on the paved road to Cape Santa Maria. Follow this road to the T-intersection with the Cape Santa Maria main road and turn left. Continue south on the Cape Santa Maria road for about 1.25 km; a sign “Beach Reserve” on the right side marks the Stella Maris property. Turn right on the sand road heading toward the beach and park.

Cape Santa Maria is a spit-like feature that has developed in later Holocene time with rising sea level. The beaches of the Cape are among the most beautiful found anywhere in The Bahamas, and the shallow waters normally are as clear and delightful as seawater can be. Excellent ripple patterns and other shallow-water bedforms lie offshore. Unfortunately, we will not have time today for a snorkel swim.

An extensive field of low, variably lithified fossil dunes lies on either side of the sand access road and extends to the west across the Cape Santa Maria main road. From the bus, walk to the north and stroll amongst the low dunes (Fig. 32). Notice that no paleosol is present, indicating that these dunes are Holocene in age. As one moves west toward the beach, the dunes begin to level out a bit. On the backside of the beach, lithified horizontal beds are present, representing a beach backshore environment with deposition near or at present sea level. This indicates that these dune and backshore beds should be assigned to the Hanna Bay Member of the Rice Bay Formation. The overall setting here appears to be very similar to that of Sandy Hook, the Holocene strand plain on San Salvador (Carney et al., 1996), and to the Holocene beach-dune complex of the Joulters Cays off the north end of North Andros Island (Boardman and Camey, 2000). Carney et al. (1996) determined that the strand plain on San Salvador developed by shoreline progradation of a beach-dune complex over the past few thousand years. The main elements of their model of strand plain formation very likely could be applied to Cape Santa Maria, although no detailed study of this area has yet been undertaken.

Bedding within the low fossil dunes is well displayed in many areas, particularly on the lee slopes of the dunes. Modern snail shells and land crab parts are abundant; note that the Cerion shells present here are smaller than those seen at the earlier stops on the east side of Long Island, attesting to the morphologic plasticity of this remarkable land snail. The vegetation, while not thick, is that of a Bahamian coastal coppice. Small palm trees are common in this area, and their trunks have diameters comparable to those of the vertical, cylindrical holes preserved in the Holocene dune beds seen at Stops 1 and 3.

Move from the fossil dunes over to the modern beach, and walk a short distance to the
Figure 33. Holocene Hanna Bay Member backshore beds cropping out along the Santa Maria beach. Photo by Al Curran.

north toward the rock outcrops (Figure 33). If the tide is low enough, a good vertical-surface view of these exposures will be possible. The horizontal beds represent deposition in a beach uppermost foreshore to backshore setting, very much like that described for similar beds of the Hanna Bay Member on Lee Stocking Island (White and Curran, 1993) and San Salvador (Curran and White, 1987, 1999, 2001). Beds with well-developed bubble porosity are common. Specimens of the trace fossil *Psilonichnus upsilon*, constructed by the ghost crab *Ocypode quadrata*, are moderately common and indicative of a backshore environment (Curran and White 1987, 1991). The overlying beds display low dip angles and represent transition to a dunal environment. Trace fossils formed by insects occur in these beds, notably specimens of the stellate burrow (Fig. 34), interpreted by Curran and White (1999, 2001) to have been formed by sweat bees. Some small cluster burrows, similar to those described by Curran and White (1999, 2001), also occur in these rocks, along with numerous root molds and some rhizomorphs.

This stop ends our field trip on Long Island. Return to the bus and re-board. We will retrace our route to the main road and head south to the airport. After a brief stop to use the facilities, we will board the planes and fly to San Salvador Island for the start of the 12th Geology Symposium.

CONCLUDING REMARKS

The purpose of this field trip to Long Island has been to present to participants attending the 12th Geology Symposium an introduction to Bahamian geology and an overview of some of the more important and interesting geologic features of the north end of Long Island. The stops selected and their order of presentation were arranged to provide efficiency of transportation, to meet permission requirements, and to exhibit the best snapshot possible of the available geology. We hope that the geology of the field stops visited today will generate lively and productive discussion throughout the time of this Symposium and beyond. The field trip leaders welcome your comments and criticisms. Please share your thoughts with us during the Symposium.

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Notes
Locations of field trip stops described in this guidebook. Cartography by Doug Gamble.