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Geology of New Providence Island, Bahamas: A Field Trip Guide

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16th Symposium on the Geology of the Bahamas
and Other Carbonate Regions

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COVER PHOTOGRAPH: Caves Point, north shore of New Providence Island, Bahamas. The image summarizes the geologic history of the Late Quaternary in the Bahamas: deposition of an eolianite, and subsequent dissolutional modification. Jim Carew stands in the entrance of Cave Point East Cave, a flank margin cave breached by hillside retreat of the enclosing dune. This photograph, taken in 1988, contrasts sharply with what is seen today at this site (Stop One of the field trip), as documentation of the developmental pressures facing natural systems in the Bahamas.

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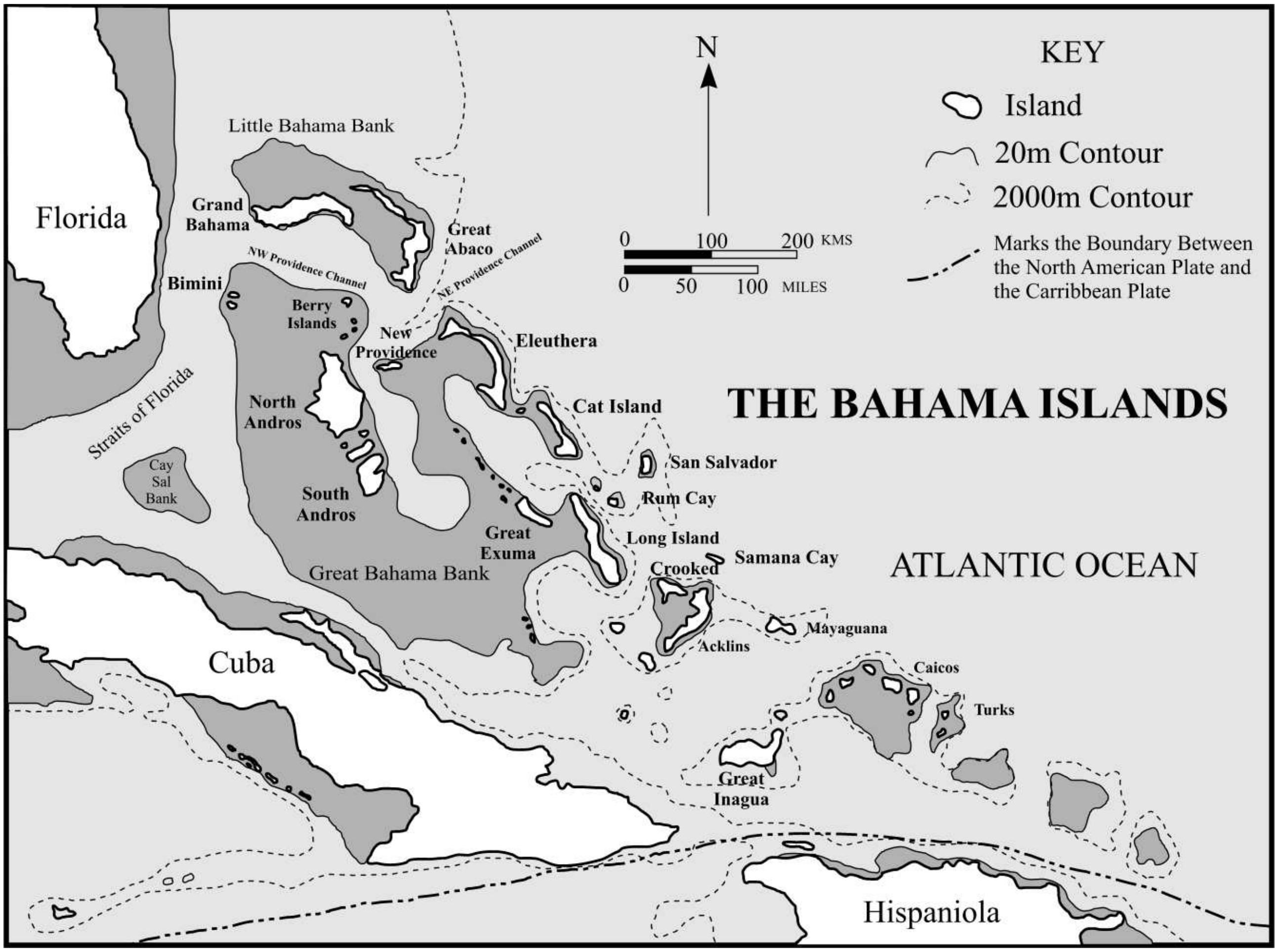

Figure 1. *Map of the Bahamian Archipelago, showing the location of San Salvador Island and New Providence Island. From Walker, 2006.*

INTRODUCTION

New Providence Island, Latitude N24^{o} 38’ 00” to N25^{o} 05’ 00” and Longitude W77^{o} 15’ 00” to W77^{o} 55’ 00”, consists of 208 km^2 of land (Sealey, 1990). The history of Nassau and New Providence Island is long and complex, and just a few highlights are presented here. The island was initially inhabited by native peoples who called themselves Lucayans, and who were essentially exterminated within decades after Columbus’ landing in the Bahamas in 1492. Europeans settled into New Providence by 1670 with at least 300 persons; in that year the Lords Proprietors of the Carolinas were granted a royal patent for the Bahamas (Albury, 1975). The Bahamas, and Nassau in particular, became a haven for pirate activities in the 17th century, resulting in the city being sacked and burned in 1703 by a combined French and Spanish force in an attempt to eliminate the problem (Albury, 1975). New Providence and the rest of the Bahamas changed dramatically in 1783 when the Loyalists, displaced British from the American Revolutionary war, settled in the Bahamas, bringing their slaves with them, and were governed centrally from Nassau (Sealey, 1990). The British government abolished slavery in 1833, with the final manumission in the Bahamas occurring in 1838 (Albury, 1975). Population levels exploded on New Providence after WWII as the tourist economy began to develop, but the major contributing factor was a steep decline in the death rate (Sealey, 1990). Independence for the Bahamas was celebrated in Nassau on 9 July, 1973. Today Nassau has more than 248,000 people, containing more than three-quarters of the nation’s present estimated total population of 316,182 https://www.cia.gov/library/publications/the-world-factbook/geos/bf.html).

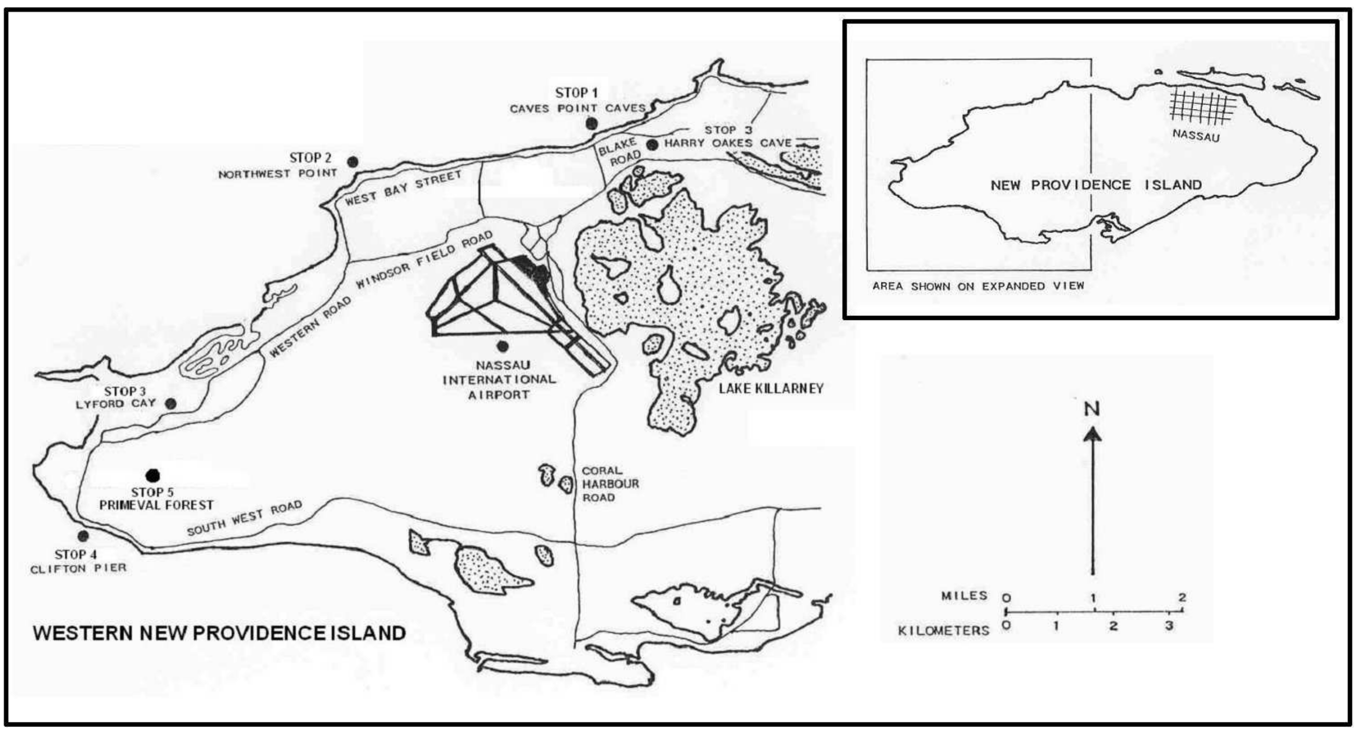

Figure 2. *Map of western New Providence, showing the major landmarks, primary roads and tracks, and field trip stop locations (solid circles).*

PROCEDURES AND GUIDELINES

This guidebook is an update of one published in 1992 for the Sixth Symposium on the Geology of the Bahamas, and a subsequent version updated for the Eighth Symposium in 1996 (Carew et al., 1992; 1996). The purpose of this field guide is to show members of the geologic community some very interesting Late Quaternary carbonate geology on the west side of New Providence Island, Bahamas. New Providence Island hosts the city of Nassau, capital of The Bahamas. The developed nature of the island offers advantages and disadvantages to geologists. The advantages are all the amenities of civilization, and all the quarries, roads, and road cuts that lead to easy access and observation. The disadvantages are all the obstacles of modern civilization, including traffic. In the interests of time, traffic, and economy of movement, this field trip has been set up to focus on the less congested western side of New Providence 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 New Providence Island, and all field trip participants are urged to treat the areas we visit with the utmost respect.

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

It is illegal to undertake scientific research in the Bahamas without a government research permit.

A section on the geology of the Bahamas, and a section on island karst processes, have been included in this field guide for those field trip participants who are unfamiliar with the Bahamas. Those sections follow immediately. For all of you "old hands", you can skip these sections and proceed directly to the **Field Trip Stops** heading on page 25, to see what we turned up during our studies this year. The official name for the country we are in is "The Bahamas", but "Bahamas" will be used in this field guide, as experience has taught us that readers get confused when "The" is used; as they assume a new sentence has begun. Despite recent attempted changes (since revoked) in the stratigraphic code, the term "Quaternary" will be used in this field guide. The depositional events of the Bahamas are tied to Quaternary sea-level position. As a result, the Marine Isotope Stage or MIS (formally the Oxygen Isotope Stage or OIS) designation for Quaternary sea-level events is utilized in this field guide. The term "carbonates" is used in the text to include the common mineralogies of calcite, aragonite, and dolomite. The Late Quaternary carbonate units of the Bahamas have significant aragonite. Because of their young age the original depositional mineralogy of the allochems has been preserved, and inversion to calcite has not proceeded to a conclusion as in most older rocks. Dolomite is rare in the subaerially-exposed carbonate rocks.

The Gerace Research Centre has undergone a series of name changes since its founding in the early 1970's. What began as the College Center of the Finger Lakes field station, or CCFL, became in the late 1980's the Bahamian Field Station. In 2002, in honor of its founder, Dr. Donald T. Gerace, the field station was renamed the Gerace Research Center as a 30th anniversary recognition. In 2007, as a result of closer oversight by the College of the Bahamas, the word “Center” in the title became “Centre”, the British spelling. Understanding the evolution of the field station name is important for sorting out publications from the field station cited in the literature. The field station website is now www.geraceresearchcentre.com.

REGIONAL SETTING

Morphology of the Bahamas

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 modern topography of the Bahamas is characterized by two distinct realms: the shallow-water banks and the deep-water troughs. The shallow areas are composed of several flat-topped, steep-sided carbonate platforms that have been sites of carbonate deposition since at least Cretaceous time, resulting in a minimum sedimentary rock thickness of 5.4 km (Meyerhoff and Hatten, 1974) and perhaps as much as 10 km (Uchupi et al., 1971). Great Bahama Bank, the largest of these banks, extends continuously over more than 400 km from N to S and 250 km from W to E. The relief of the banks is low and most submerged areas are covered with less than 10 m of water (Newell et al., 1960). Small islands cap the banks mainly on their windward, eastern margins. Their area sums up to about 11,400 km^2 (Meyerhoff and Hatten, 1974). Channels and re-entrants, together with the peri-platform ocean, make up the deep-water realm. In Exuma Sound and Tongue of the Ocean, the two embayments that cut into Great Bahama Bank (Figure 1), water depths exceed 1000 m. These deep-water areas are separated from the banks by steeply dipping slopes that generally show higher angles on the eastern (windward) flanks of the platforms. Below the platform edge (40-60 m), the slopes dip almost vertically down to depths of around 135-145 m ("the wall"; Grammer and Ginsburg, 1992), where the slope angle decreases again. Where facing the open Atlantic Ocean (Bahama Escarpment), the slopes reach depths greater than 4000 m. The northwestern Bahama islands are isolated landmasses that project above sea level from Little Bahama Bank and Great Bahama Bank. To the southeast, beginning in the area of San Salvador, the Bahamas comprise small isolated platforms capped by islands that cover a significant portion of the available platform area.

Origin and tectonic setting of the Bahamas

Early workers assumed a continental basement for the Bahamas. However, this hypothesis was questioned by the first plate-tectonic reconstructions of the area (e.g. Bullard et al., 1965) where the Bahamas entirely overlap the African continent. Dietz et al. (1970) therefore suggested an oceanic composition for the underlying crust. Based on seismic, gravimetric and paleomagnetic data, Meyerhoff and Hatten (1974) nevertheless demonstrated the continental nature of the Bahamian substrate. Their hypothesis was later confirmed by Mullins and Lynts (1977) who resolved the overlap problem by a pre-rift reconstruction involving rotation of the region, thus obtaining a perfect fit with Africa. It is now admitted that the archipelago rests on the stretched continental crust of the North American plate.

The striking present-day configuration of banks and troughs of the Bahamas was also the subject of a long-lasting controversy, from which three main theories have evolved. This peculiar topography was initially interpreted as an inherited subaerial drainage pattern (Hess, 1960). Later, Mullins and Lynts (1977) put forward the "graben hypothesis", which explains this morphology as resulting from an initial horst and graben pattern consistent with continental rifting during the opening of the Atlantic Ocean. Finally, several authors (Meyerhoff and Hatten, 1974; Schlager and Ginsburg, 1981; Sheridan et al., 1988) proposed the "megabank hypothesis", which holds that the modern Bahamas are a segmented remnant of a much larger and continuous Mesozoic carbonate platform. Recent work by Eberli and Ginsburg (1987), Mullins and Hine (1989), and Manfrino and Ginsburg (2001) have demonstrated that the Bahama banks are undergoing both depositional progradation and erosional segmentation.

Most studies (e.g., Uchupi et al., 1971; Mullins and Lynts, 1977; Carew and Mylroie, 1995a) praise the tectonic stability, or slow subsidence of the Bahamas archipelago, which makes it an excellent gauge for Quaternary sea-level studies. This may well be true for the northern Bahamas, but less so for the southern islands (see Masaferro et al., 1999), some of which stand less than 100 km to the north of the oblique convergence zone between the North American and the Caribbean tectonic plates (Dolan et al., 1998). Significant earthquakes have been felt on Great Inagua (the southernmost Bahamian island) as recently as 1957 (Pierson, 1982), and the uneven elevation of last interglacial marine deposits on this island suggests it could have been tilted (Kindler et al., 2007). Mayaguana Island also displays evidence of an uplift event (Kindler et al., 2011). However, geoid changes due to deglaciation and mantle plasticity could also be involved.

SUBAERIAL GEOLOGY OF THE BAHAMAS

Research on the ancient limestones forming the Bahamas islands always lagged behind that concerned with modern platform sediments. Bahamian cays were long thought to essentially consist of oolitic eolianites formed during the last interglacial period (e.g., Newell and Rigby, 1957; Bathurst, 1975). This notion was however refuted when older carbonates of skeletal composition and marine origin were discovered onshore (Garrett and Gould, 1984; Carew and Mylroie, 1985; Kindler and Hearty, 1995). The first major paper on island geology was published less than 30 years ago (Garrett and Gould, 1984). Since about that time, biannual geological symposia, organized at the Bahamian Field Station (now the Gerace Research Centre) on San Salvador, favoured the exchange of ideas on Bahamian island geology. Building on earlier work by Titus (1980; 1983), Carew and Mylroie (1985) proposed a three-part stratigraphic model comprising Holocene, Sangamonian (MIS 5e) and pre-Sangamonian lithologic units, essentially based on observations from San Salvador. This model was challenged by Hearty and Kindler (1993) from work on this and other islands, which resulted in a heated debate regarding the number of stratigraphic units, the occurrence of marine deposits predating the last interglacial, and methodological approaches (e.g., Carew and Mylroie 1994; Hearty and Kindler 1994). These conflicting studies nonetheless agreed that the exposed rocks of the Bahamas were all of mid- to late-Quaternary age (Hearty and Kaufman, 2009; Mylroie, 2008). This basic view was recently invalidated by Kindler et al. (2008; 2011) who found exposed upper Miocene, Pliocene and lower Pleistocene strata on Mayaguana Island. However, this island is the sole example to date of pre-middle Pleistocene surficial outcrops in the Bahamas, and as noted earlier, may represent a unique local uplift event.

Depositional models

Carbonate sedimentation and glacioeustasy

The Bahamas Islands consist of vertically stacked and/or laterally juxtaposed carbonate units, mostly eolianites, separated by thin red to brown layers generally interpreted as paleosols (Figure 3). One fundamental principle of Bahamian island geology, derived from a glacio-eustatic model initially developed from observations in Bermuda (Bretz, 1960; Land et al., 1967; Vacher and Rowe, 1997), is that carbonate deposits accumulate during interglacial highstands of sea level, when the platform tops are flooded by a shallow water layer (e.g., Titus, 1983; Carew & Mylroie, 1995b; 2001). This model remains certainly valid for the Quaternary, but the recent discovery of elevated marine deposits of Miocene and Pliocene age on Mayaguana suggests that, on some banks and during these time intervals, carbonate production and deposition occurred during lowstands, but stalled during highstands that were much higher than present sea level (Godefroid, 2012; Kindler et al., 2011).

Significance of paleosols.

The paleosols of the Bahamas have been classified in various ways (e.g., Boardman et al., 1995; Nawratil de Bono, 2008), but two main types are commonly recognized: *terra-rossa* paleosols and calcarenite protosols (Carew and Mylroie, 1991). The former term refers to red to brown earthy material and associated pedogenic structures such as vegemorphs (rhizoliths), micritic crusts, and disarticulated blocks of parent material. The term protosol, coined by Vacher and Hearty (1989) from observations on Bermuda, designates white to tan, unstructured sandy layers, locally containing abundant fossil pulmonate snails. Another basic concept of Bahamian island geology holds that *terra-rossa* paleosols develop during prolonged episodes of subaerial exposure characteristic of glacial periods, whereas calcarenite protosols can form at any time during an interglacial, and correspond to brief pauses in carbonate sedimentation (Carew and Mylroie, 1995b). *Terra-rossa* paleosols thus seem to have a stratigraphic significance, but should be used with caution when reconstructing the stratigraphic record of the Bahamas islands because they can bifurcate, merge, or be piped down in karstic conduits (Carew and Mylroie, 1991). The paleosol question is indeed complex. Godefroid et al. (2010) recently demonstrated that some clay-rich breccia layers interpreted so far as Wisconsinan paleosols actually represent colluvial deposits of

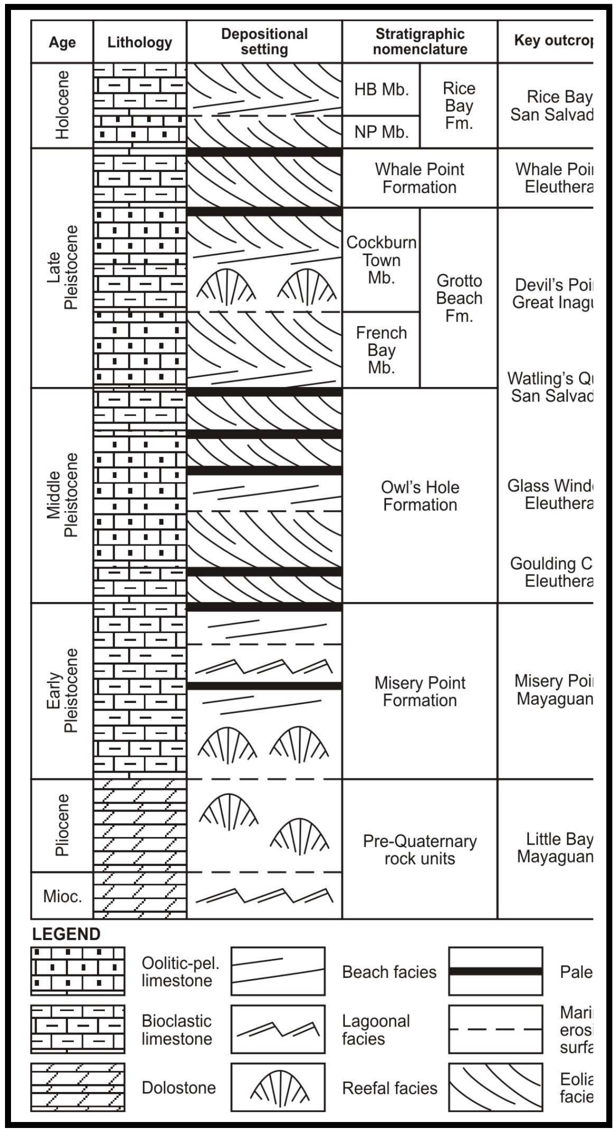

Figure 3. *Synthetic stratigraphic column of the Bahama Islands (Mb. = Member; Fm. = Formation; Qua. = Quarry; pel. = peloidal).*

Holocene age. Confirming earlier reports (Hearty and Kindler, 1995; Kindler and Hearty, 1997; Hearty 1998), recent research further shows that paleosols of intermediate maturity can be found between carbonate units deposited during distinctive interstadials of the same interglacial.

Deposition phases during one sea-level cycle

Carew and Mylroie (1995b; 1997) proposed an empirical model of island development that can be used for deciphering the stratigraphy of the Bahamas archipelago. According to these authors, carbonate packages deposited during interglacial periods comprise three parts: a transgressive phase, a stillstand phase, and a regressive phase. These phases each contain subtidal, intertidal and eolian sediments, the latter ones are generally the most extensively represented. Deposits of each phase present distinctive sedimentological characteristics (e.g., preservation of fine-scale eolian structures, abundance of plant remains; occurrence of interstratified protosols (Carew and Mylroie, 1997) that facilitate their recognition in the field. The authors of this section considers this model as valuable, but would like to emphasize that the history of sea-level fluctuations during one single interglacial (e.g., MIS 5e) can be more complex than a simple rise, stillstand and fall (Neumann and Hearty, 1996; Carew and Mylroie, 1999; Thompson and Goldstein, 2005; Blanchon et al., 2009), and thus generate a more intricate stratigraphic record. In addition, geological evidence from Eleuthera clearly shows that some of the deposits presumably correlated with the MIS 5e regression, and used as case examples in the presented model (Carew and Mylroie, 1995b, 1997), actually correspond to a younger sea-level highstand event, probably during MIS 5a; a MIS 5a sea level position to modern elevations has been demonstrated from Mallorca Island, Spain (Dorale et al., 2010).

Comparing highstands between archipelagos, or even within an extended archipelago such as the Bahamas, assumes a consistent glacioeustatic record across the globe. Recent work on the rheologic behavior of the mantle in response to ice loading and unloading, known as Glacial Isostatic Adjustment (GIA), indicates that glacioeustasy can work in a unique local pattern, even distant from ice sheets, making comparisons difficult (Spada 2012). For example, extrapolating the Dorale et al. (2010) MIS 5a sea-level position at modern levels from Mallorca in the Mediterranean basin to the Bahamas is especially problematic. The mid-Holocene sea-level highstand, well recognized in the western Pacific as 1 to 2 m above current levels (e.g., Miklavič et al., 2012), is not well documented in the Atlantic basin because of significant differences in the thickness and temperature-controlled rheology of the oceanic crust in those two locations.

Morphostratigraphic principles

Several authors (e.g., Titus, 1980; Garrett and Gould, 1984; Hearty and Kindler, 1993) have used the morphostratigraphic principles of lateral accretion (Itzhaki, 1961; Vacher 1973) and of catenary growth (Garrett and Gould, 1984; Carew and Mylroie, 1997) to unravel the sequence of deposition exposed on Bahamian islands. The former principle states that, on a prograding shoreline, deposits become younger seaward. In contrast, the latter one asserts that catenary ridges are younger than their anchoring headlands. Carew and Mylroie (1995b) have discussed the limitations of this method pertaining to the varying height of sea level at times of deposition and the composite nature of some ridges. It follows that the morphostratigraphic approach using topographic maps or air photographs can be helpful to get a first approximation of depositional sequences, but it has to be complemented by detailed field, sedimentological and petrographic studies.

Dating methods

^{14}C dating

Radiocarbon dating has been used to unravel the Bahamian stratigraphic record spanning the latest Pleistocene and the Holocene. Analyses were performed on whole-rock samples collected from eolianites (Carew and Mylroie, 1987) or beachrock (Kindler and Bain, 1993), shell fragments of terrestrial snails (Hearty and Kaufman, 2009), and micritic crusts (Aalto and Dill, 1996; Godefroid et al. 2010). Providing whole-rock samples have been leached to reduce the possibility of contamination by diagenetic cement, obtained ages reflect the average time of *formation* of constituent grains. The actual time of *deposition* of a specific rock body may thus be constrained by the age of constituent allochems and that of contained fossil shells such as *Cerion* land snails (Hearty and Kaufman, 2009). For Holocene eolianites, the technique has been particularly useful, as cement volumes are miniscule, and the time window between allochem formation and their subsequent eolian deposition is necessarily small.

U-series dating

In the Bahamas, U-series methods have been frequently employed to obtain ages from coral specimens (Neumann and Moore, 1975; Carew and Mylroie, 1987; Chen et al., 1991; Kindler et al., 2007), and, less commonly, from speleothem (Carew and Mylroie, 1987; Li et al., 1989) and whole-rock samples (Muhs and Bush, 1987). Thanks to the advent of thermal ionization mass spectrometry (Edwards et al. 1987), the error margins on last-interglacial ages have now been narrowed down to less than 1000 yrs. The moot point of this method, however, is that coral specimens may not necessarily represent closed geochemical systems after death (Fruijtier et al., 2000). Precipitation of a late cement phase in the coral porosity (Kindler et al., 2007), or migration of radio-nuclides following subtle diagenetic modifications (e.g., aragonite recrystallization) of the analyzed samples may produce erroneous ages. Such difficulties might nonetheless be overcome by analyzing many samples from the same outcrop, and even from the same coral, to check the reproducibility of the method, and also by applying an open-system age model (Thompson et al., 2003). The most recent dates for MIS 5e in the Bahamas, are now 124 to 115 ka (Thompson et al., 2011), 25% shorter as well as younger than the previously accepted time of 131 to 119 ka of Chen et al. (1991).

Amino-acid racemization (AAR) dating

This technique relies on the slow *post-mortem* interconversion (racemization) of L-amino acids, within indigenous proteins preserved in the analyzed sample, to increasing proportions of their respective D-configurations until an equilibrium mixture of D- and L-amino acids is attained. The rate of this chemical reaction depends on both time and ambient temperature within the geological deposit. If the temperature history of a region is known, D/L ratios can thus provide an estimation of the sample age. Several authors (e.g., Carew 1983; Hearty and Kindler, 1993; Goodfriend and Gould, 1996; Hearty and Kaufman, 2000, 2009; Hearty, 2010) have analyzed the amino-acid composition of marine shells, coral fragments, land snails (*Cerion* sp.), and whole-rock samples to decipher the stratigraphic record of the Bahamas and/or to constrain evolutionary trends. These attempts, in particular the use of whole-rock samples, resulted in a heated and somewhat sterile debate on the reliability of the method (see summary and references in Carew and Mylroie, 1997). The authors of this paragraph agree with Carew and Mylroie (1995b) that amino-acid racemization dating should not be used singlehandedly to define Bahamian stratigraphic units. However, the consistency of D/L values and ^{14}C ages measured on *Cerion* shells of Holocene age (Goodfriend and Gould, 1996; Hearty and Kaufman, 2009), and the fact that over 90% of whole-rock alloisoleucine/isoleucine (A/I) ratios obtained from Bermuda and the Bahamas in the past 20 years (Hearty et al., 1992; Hearty and Kaufman, 2000) are consistent with the physical stratigraphy of these regions, both indicate that the AAR method has some validity.

Paleontology

Due to the widespread occurrence of the terrestrial gastropod *Cerion* (Figure 4) in paleosols and eolianites, several attempts have been made to use paleontological criteria for establishing the stratigraphy of the Bahamas islands. These efforts have only been partly successful because of the very large number (~600) of *Cerion* species (e.g. http://invertebrates.si.edu/cerion), and the influence of geographical and ecological factors on the snail morphology (Rose, 1983; Gould, 1997; Hearty, 2010). Based on species recognition, Garrett and Gould (1984) identified three sequential *Cerion* fauna in the dunal deposits of New Providence Island, characterizing the pre-Sangamonian, the Sangamonian (MIS 5e), and the Holocene, respectively. Unfortunately, this biozonation is only valid for New Providence: Pleistocene deposits on Inagua include a different *Cerion* fauna (Goodfriend and Gould, 1996), and the fossil specimens found on San Salvador are indistinguishable from modern species occurring in the same place (Gould, 1997).

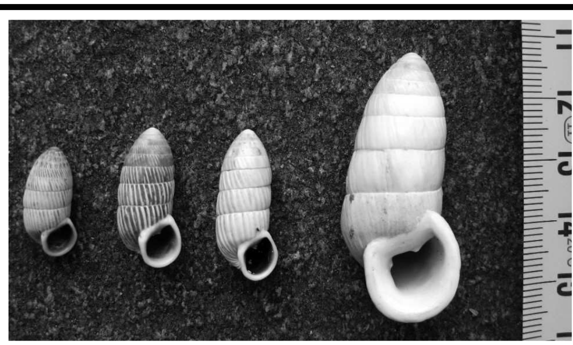

Figure 4. *Modern Cerion sp. snails collected from Mayaguana and Inagua Islands. Note variability in size color, and rib relief (P. Kindler photo).*

Approaches based on morphometric and statistical methods (Hearty et al., 1993; Fronabarger et al., 1997; Hearty and Schellenberg, 2008; Hearty, 2010) produced conflicting results, but further demonstrated that changes in the morphology of *Cerion* shells are mostly related to geographical and ecological factors (moisture, temperature, wind, population density). Interestingly, Steve Gould was a participant on the 1996 New Providence field trip (Carew et al., 1996), and at Stop One of that trip (Stop Two of this trip), he located a *Cerion* in the Holocene eolianites that overlie a MIS 5e fossil reef, and identified it, based on his experience, as being solely from the Holocene on New Providence. Later, at Stop 2 (Carew et al., 1996), he was shown similar *Cerion* from an Owl's Hole (pre-MIS 5e) outcrop, and Gould promptly explained that he had been wrong earlier that day. Such has been the frustration with *Cerion* in the Bahamas.

Magnetostratigraphy

Several researchers have tried to use paleomagnetic criteria to date and correlate Bahamian eolianites and paleosols. These studies were essentially conducted on San Salvador and Eleuthera, and obtained results have to be viewed with guarded optimism (Panuska et al., 1995). Early evidence for a reversed polarity in a paleosol on San Salvador (Carew and Mylroie, 1985) was later proved erroneous (Stowers et al., 1989; Carew and Mylroie, 1995b). Based on secular variation of paleomagnetic directions (i.e., inclination and declination), Panuska et al. (1999) identified three distinct paleosols on San Salvador (Fernandez Bay, Gaulin Cay, and Sandy Point Pits magnetotypes), but acknowledged that some paleomagnetic data were at odds with the geological mapping. On Eleuthera, distinctive, vertically stacked paleosols, separated by metyers-thick eolianites, yielded similar paleomagnetic directions, suggesting either an unknown and complex pedogenic process or a failure of the method on this island (Panuska, 2004). Clearly, paleomagnetic criteria should not be used to supersede stratigraphic evidence (Panuska et al., 1995).

Sr-isotope stratigraphy

The Sr-isotope ratio in seawater depends on the varying proportions of the heavier isotope (^{87}Sr) originating from fluvial input of dissolved continental material, and of the lighter isotope (^{86}Sr) derived from hydrothermal leaching of basalts at mid-ocean ridges. Sr can substitute for Ca in the crystal lattice of marine carbonates. The ^{87}Sr/^{86}Sr ratio in marine carbonates reflects that of seawater during precipitation, and thus provides a numerical age for these minerals when compared with the Sr-isotope evolution of global seawater (Howarth and McArthur, 1997; McArthur et al. 2001). In the case of dolomite, the Sr may partly, or wholly, originate from the dolomitizing fluid, which, in the Bahamas, is essentially seawater (Whitaker et al., 1994). In such a case, obtained ages correspond to the youngest possible age of the precursor sediments or the oldest possible age of dolomitization (Vahrenkamp et al., 1988). Sr-isotope stratigraphy has long been used to date the subsurface rocks of the Bahamas (Swart et al., 1987; Vahrenkamp et al. 1988; 1991), but has only recently been applied to decipher island stratigraphy (Kindler et al., 2008; 2011). This method is most useful to distinguish Miocene, Pliocene, and lower to middle Pleistocene rocks, but is not sensitive enough to resolve stratigraphic details in the past 200 ka.

Electron Spin Resonance

Electron spin resonance, or ESR, is an age-dating technique that measures the cumulative damage done over time to crystal lattice structure in minerals. The damage is a result of the radiation load experienced by the sample. A reconnaissance study (Deely et al., 2011; 2012) examined both marine and terrestrial molluscan fossils from San Salvador Island. Fossil corals *in situ* from the Cockburn Town fossil reef on San Salvador Island gave an age of 127 +/- 6 to 138 +/- 10 ka, consistent with the existing U/Th dates. However, two fossil corals from a rubble facies behind and below the reef gave ages of 190+/-17 ka and 223 +/- 18 ka, consistent with a MIS 7 origin. Samples of the bivalve *Codakia* gave a broad range across MIS 5, from 82 +/- 10 ka to 114 +/- 11 ka. *Cerion* samples yielded ages 49 +/- 6 ka to 75 +/- 8 ka, an age equivalent to MIS 5a and younger. Because of diagenetic questions, and assumptions about dosage exposure, these ages can be considered only in general terms. On one hand, they seem to indicate that four sea-level events are determinable on San Salvador: MIS 7, 5e, 5c, and 5a. On the other hand, as a result of the unknowns, the samples can be considered to be all MIS 5e, as they are in proper relative stratigraphic order: coral rubble on which lie *in situ* corals, in turn overlain by shallow lagoonal facies in turn overlain by regressive eolianites. Work is in progress to see if this technique can be refined, and obtain reproduceable results.

Bahamian stratigraphic units

In this section, we combine the stratigraphic observations made by various teams of researchers on almost all Bahamas islands (see Figure 3), and try to reconcile views that are not as divergent as previously published (e.g., Carew and Mylroie, 1997; Kindler and Hearty, 1997).

Pre-Quaternary rocks

Four lithostratigraphic units, briefly described below, were identified on Mayaguana, and are new additions to the stratigraphic record of the Bahamas islands (Kindler et al., 2011, Godefroid, 2012). They mostly comprise peritidal carbonates that formed during interglacial episodes of platform flooding, separated by karstic surfaces and paleosols corresponding to exposure during glacial intervals.

The Mayaguana Formation: This unit forms one small exposure at the western end of a km-sized rocky headland near Little Bay, a small embayment on the north coast of Mayaguana, at elevations from present sea level up to +1 m. It consists of fine-grained, hard limestone containing a rich assemblage of larger benthic foraminifers including *Miogypsina globulina* and *Miolepidocyclina burdigalensis* (Figure 5), indicative of a shallow, high-energy, peri-reefal environment. This limestone yielded $^{87}Sr/^{86}Sr$ ratios averaging 0.708546 ± 0.00001, indicating an early Miocene age (18.4-18.7 Ma), which corroborates the biostratigraphic age (Burdigalian) obtained from the foraminiferal assemblage.

The Little Bay Formation: Exposed at both ends of the Little Bay headland, at elevations from sea level up to +1.5 m, this unit consists of fine-grained, hard dolostone, displaying multidirectional cross-bedding and ripple laminations. Petrographic analysis reveals a dense microsucrosic dolomite, preserving no evidence for any primary rock fabric. Sedimentary structures suggest that this unit was deposited in a high-energy tidal environment. Sr-isotope analyses gave an average $^{87}Sr/^{86}Sr$ ratio of 0.708988 ± 0.000025, indicating a late Miocene age (5.59-6.81 Ma). Similar dolostone has been described on San Salvador at depths exceeding 140 m below the island surface (Supko, 1977).

The Timber Bay Formation: This unit crops out at several locations along the north coast of Mayaguana at elevations from sea level up to +3 m. It consists of hard, partly dolomitized, coral/algal boundstone and rudstone, with a bioclastic grainstone matrix. Allochems were mimically replaced by cryptocrystalline dolomite, whereas coarser dolomite precipitated in intra- and intergranular pores. The occurrence of coral/algal build-ups and its overall poorly stratified nature relate to a shallow and energetic reefal setting. The average $^{87}Sr/^{86}Sr$ ratio of collected samples is 0.709067 ± 0.000011 corresponding to a middle Pliocene age (2.12-4.09 Ma).

The Misery Point Formation: This unit is best exposed on a 1 km long, up to 15 m high, rocky shore segment, the Misery Point cliff, but also occurs at four other locations along the northern coast of the island. It includes three vertically stacked shallowing-up sequences corresponding to lithostratigraphic units of lower rank (members) and consisting of pervasively weathered carbonates, separated and capped by paleosols. Sedimentary

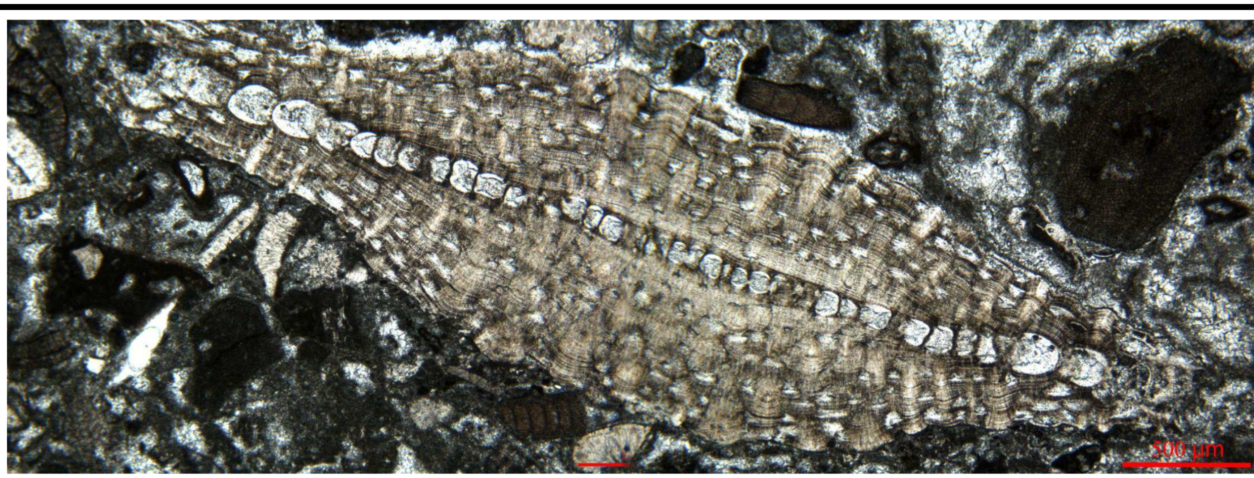

Figure 5. *Equatorial section of Miolepidocyclina burdigalensis.*

structures and faunal content suggest these sequences were deposited by marine processes in reefal, peri-reefal, and beach settings, respectively. $^{87}Sr/^{86}Sr$ ratios from these carbonates give a range between 0.709126 ± 0.00008 and 0.709138 ± 0.00015 corresponding to an early Pleistocene age (~1.00 and 1.19 Ma). These deposits have not been identified on the surface of any other Bahamian islands, but could possibly be found at the base of high sea cliffs in northern Eleuthera (Kindler et al., 2011).

The Owl's Hole Formation

The Owl's Hole Formation was originally defined on San Salvador (Carew and Mylroie, 1985) to designate a predominantly bioclastic eolianite (Stowers et al., 1989; Hearty and Kindler, 1993) exposed in a deep dissolution pit in the SW portion of that island. Similar deposits were later found on most Bahamian islands and shown to be much more extensive than previously thought (Schwabe et al., 1993; Carew and Mylroie, 1995; Kindler and Hearty, 1995; 1996). Initially presented to describe all rock units blow MIS 5e (Sangamon), we propose here to regroup under this label all rock bodies occurring above lower Pleistocene deposits and below the Grotto Beach Formation (see next section). In addition to the bioclastic eolianite previously mentioned, which should be downgraded to member status, this formation would include several paleosol-capped, moderately weathered calcarenite units of various petrographic composition (oölitic-peloidal, bioclastic-peloidal, bioclastic) deposited in shallow-marine and terrestrial settings. These units can be observed in vertical superposition in northern Eleuthera (Kindler and Hearty, 1997; Hearty, 1998; Panuska et al., 2002), but they occur also in lateral juxtaposition on most Bahamian islands. Marine facies associated with this formation have been identified on New Providence (Garrett and Gould, 1984; Hearty and Kindler 1997), Eleuthera (Hearty and Kindler, 1995; Hearty, 1998; Hearty et al., 1999; Kindler and Hearty, 2000), and Great Inagua (Kindler et al., 2007). Tentative U-Th dating of whole-rock samples collected from the Owl's Hole Formation gave ages between ~200 ka (Muhs and Bush, 1987) and ~316 ka (Kindler et al., 2007). Paleomagnetic analyses revealed a normal polarity (Stowers et al., 1989; Panuska, 2004). The A/I ratios measured from these calcarenites range from 0.87 to 0.55 (Hearty and Kaufman, 2000; Hearty, 2010), suggesting a middle Pleistocene age. This age is further supported by two $^{87}Sr/^{86}Sr$ values obtained from exposures of this formation on Mayaguana and Great Inagua that average 0.709158 (0.28-0.88 Ma; Kindler et al., 2011). The Owl's Hole Formation thus comprises deposits formed during interglacial highstands of the middle Pleistocene, spanning MIS 7 to possibly 15 (Hearty and Kaufman, 2000; Hearty 2010). It is identifiable by its position below the Grotto Beach Formation, its modest degree of diagenetic alteration compared to older units, and the predominance of bioclastic-peloidal eolianites.

The Grotto Beach Formation

First identified on San Salvador (Titus, 1980; 1983), this formation is the most conspicuous stratigraphic unit in the Bahamas. It forms the highest elevation of the archipelago, a value of 63 m at Mt Alvernia on Cat Island. It includes all deposits found above the Owl's Hole Formation and below the Rice Bay Formation (see next section), and comprises two distinctive members: the French Bay and the Cockburn Town Members.

The French Bay Member (Carew and Mylroie, 1985) is made of a well-preserved oölite commonly showing pristine sedimentary structures typical of the subtidal, intertidal, and predominantly eolian depositional environments (e.g., Caputo, 1995; Kindler and Hine, 2008). It has been attributed to the early part of the last interglacial period (Marine

Isotope Substage 5e), based on its stratigraphic position relative to the radiometrically dated Cockburn Town Member (see next section) and whole-rock A/I ratios averaging 0.48 throughout the Bahamas islands (Hearty and Kaufman, 2000). The French Bay Member has been interpreted as the transgressive-phase eolianite of the Grotto Beach Formation (Carew and Mylroie, 1985; 1995b; 1997), and alternatively, as representing an early highstand phase during MIS 5e (Hearty and Kindler, 1993; Kindler and Hearty, 1996; Hearty and Kaufman, 2000).

The Cockburn Town Member (Carew and Mylroie, 1985) mainly comprises coral boundstone/rudstone and shell-rich floatstone, corresponding to ancient reefal and lagoonal settings, respectively, and associated eolianites. The latter facies commonly drapes the interior of many Bahamian islands such as on Mayaguana, Great Inagua, and New Providence. U-Th ages obtained from well-preserved coral specimens allow correlation with MIS 5e. Older ages have been reported from one site in Great Inagua (Kindler et al., 2007), suggesting that some reef portions could span two interglacial periods. These ages, however, appear to be questionable (B. Thompson, pers. comm., 2007). Thompson et al. (2011) have done a detailed analysis of subtle diagenetic effects on fossil corals that can influence U/Th ratios, such that they believe the age of the Cockburn Town Member (MIS 5e) corals in the Bahamas is actually 124 to 115 ka (as previously noted). A/I ratios measured on samples collected from the rudstone matrices average 0.40 (Hearty and Kaufman, 2000). In many localities (e.g., at the type section of the Grotto Beach Formation on San Salvador), predominantly oölitic calcarenites, displaying a shallowing-upward succession of facies from subtidal to eolian, conformably overlie the aforementioned reefal facies. These rocks were lumped with regressive phase eolianites in the Cockburn Town Member, because of the difficulty in the field of differentiating progradational deposits formed during a sea-level stillstand from regressive facies produced by an actual sea-level decline (Carew and Mylroie, 1985; 1995b; 1997). Other authors (Hearty and Kindler, 1993; Kindler and Hearty, 1996; Hearty and Kaufman, 2000) regard these deposits as a separate entity corresponding to a late MIS 5e sea-level highstand. Considering the complex sea-level history of the last interglacial (Neumann and Hearty, 1996; Thompson and Goldstein, 2005; Blanchon et al., 2009), the details of stratigraphic subdivision remain an open question.

<u>The Whale Point Formation</u>

The Whale Point Formation (new name) includes vegemorph-rich, paleosol-capped bioclastic eolianites overlying or juxtaposed to the Grotto Beach Formation. The occurrence of a paleosol and/or a calcrete at the base and at the top of this unit indicates that it post-dates the Grotto Beach Formation, but predates the Rice Bay Formation (see next section). A/I ratios obtained from whole-rock samples collected from these calcarenites average 0.31 (Hearty and Kaufman, 2000), suggesting a correlation with MIS 5a. This unit was initially identified on San Salvador and named the Almgreen Cay Formation by Hearty and Kindler (1993). This addition to the Bahamian stratigraphic record known at that time (Carew and Mylroie, 1985) generated much controversy in the subsequent decade (e.g. Carew and Mylroie, 1994; Hearty and Kindler, 1994; Panuska et al., 2002; Mylroie et al., 2008) mostly because, on San Salvador, stratigraphic relationships are not straightforward and this new formation was essentially defined on the basis of petrologic, geomorphologic, and AAR data. By contrast, on Eleuthera, the Whale Point and the Grotto Beach Formations occur in vertical superposition and are separated by a fairly mature paleosol that probably developed over a few 10 ka (Kindler and Hearty, 1995; Hearty and Kindler 1995; Hearty, 1998; Brentini, 2008). We chose to rename this unit the Whale Point Formation because it is at Whale Point, in northern Eleuthera, that the stratigraphic relationships are most clear. The definitive attribution of isolated exposures of paleosol-capped, vegemorph-rich, bioclastic eolianites to the Whale Point Formation or to the regressive-phase deposits of the Grotto Beach Formation is a task awaiting the next generation of Bahamian geologists.

The Rice Bay Formation

The Rice Bay Formation (Carew and Mylroie, 1985) may overlie any paleosol-capped stratigraphic unit (usually the Grotto Beach Formation), but is only covered by a mm-thin micritic crust, a protosol, or a modern soil. This unit comprises two members, the North Point and the Hanna Bay Members, both defined on San Salvador (Carew and Mylroie, 1985), that can be vertically stacked, as on Eleuthera, or laterally juxtaposed, as at the type section.

The North Point Member (Carew and Mylroie, 1985) consists of transgressive-phase oölitic to peloidal/bioclastic eolianites displaying pristine small-scale structures such as grain fall, grain flow, and climbing wind-ripple laminae (White and Curran, 1988; Kindler 1992; 1995), that extend down to, and below, modern sea level. These oölites differ petrographically from those forming the French Bay Member by the occurrence of mostly thinly-coated grains and more abundant peloids and bioclasts (Kindler and Hearty, 1996). Most recent ^{14}C data obtained from whole-rock samples and *Cerion* sp. shells suggest that these sediments were formed on the shallow platforms around 6.2 ka and deposited on islands over the next 1000 years or less (Hearty and Kaufman, 2009).

The Hanna Bay Member (Carew and Mylroie, 1985) is represented by predominatly bioclastic calcarenites deposited in a beach-dune environment, usually in equilibrium with modern sea level (Davaud and Strasser, 1984; Carew and Mylroie, 1985). Locally low-angle planar beach beds including phreatic cements, and correlated with the Hanna Bay Member, have been observed up to 2 m above present sea level (Kindler et al., 2011). Recently obtained ^{14}C ages (Hearty and Kaufman, 2009) show that these sediments were generated between 4.7 and 3.8 ka ago, and deposited subsequently on the islands. A model accounting for the depositional break between the two members of the Rice Bay Formation and their distinctive petrographic composition has been presented by Kindler (1992; 1995).

Field work in the Bahamas is greatly assisted if the basic stratigraphy can be worked out on site. Such real time study makes the typical geochronological methods of determining stratigraphic units inoperative. Figure 6 is the simplified geologic column for the Bahamas, eliminating the pre-middle Pleistocene units of Mayaguana, and those demonstrated solely on Eleuthera - the subdivisions of the Owl's Hole Formation and the Whale Point Formation. The remaining units can be found on most Bahamian islands, and with careful field work, can be resolved without the use of geochronologic tools. Figure 7 is a flow chart that is useful throughout the Bahamas to figure out what unit is present in outcrop. It is not definitive for some cases, especially Owl's Hole eolianites, and certainly not for the Mayaguana pre-Pleistocene outcrops. The simplistic Rice Bay – Grotto Beach – Owl's Hole stratigraphy of the Bahamas developed on San Salvador Island has been successfully used on many Bahamian islands, and has been applied by a series of field guides produced over the years by the GRC. The simplified stratigraphic column of Figure 6 has been utilized on New Providence Island (Carew et al., 1996), South Andros Island (Carew et al., 1998), Eleuthera Island (Panuska et al., 2002), Long Island (Curran et al., 2004), Cat Island (Mylroie et al., 2006) and Rum Cay (Mylroie et al., 2008a). The model has also been used suc-

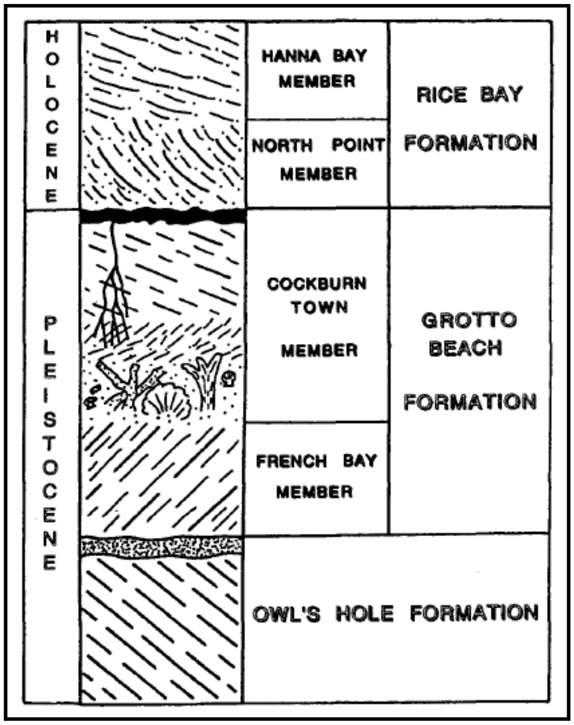

Figure 6. *Simplified stratigraphic column for the Bahamas.*

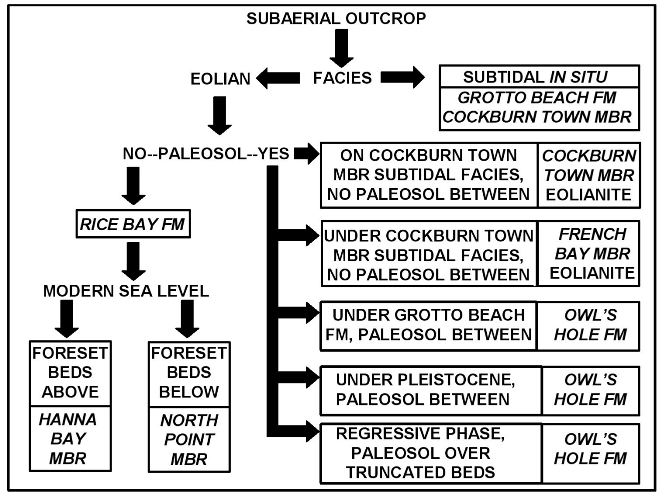

Figure 7. *Flow chart to apply field evidence toward making a decision about placement of an outcrop into the simplified Bahamian stratigraphy of Figure 6.*

cessfully on Abaco Island (Walker et al., 2008). However, Figure 6 remains a simplification, as fieldwork, especially on Eleuthera, as noted earlier, has shown that there are subtleties and complexities that the simplified geologic column does not address. And there is always the unexpected new discovery, such as those on Mayaguana. Figure 3 is the definitive stratigraphy for the Bahamas.

KARST, CAVES, AND LANDFORM DEVELOPMENT IN THE BAHAMAS

The Bahama Islands have landscapes that are dominated by original depositional features, and are only slightly modified by subsequent dissolutional (karst) processes. The high porosity of the limestones that form the islands results in rapid infiltration of meteoric water and the absence of surface streams and related erosional features such as valleys and channels. Classic surface karst landforms such as sinking fresh-water streams, blind valleys, and dissolution sinkholes are absent. The karst features of the Bahamas fall into four main categories: karren, depressions, caves, and blue holes.

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

Table 1: *Components of the Carbonate Island Karst Model (Jenson et al., 2006).*

Distinct geomorphic types that distinguish carbonate islands from one another	Common attributes that distinguish island karst from karst of interior settings
Simple Carbonate Islands: Non-carbonate rocks remain below the zone of fresh-water influence, and recharge is exclusively autogenic. (Fig. 1A)	The karst is **eogenetic,** i.e., it has developed in carbonate rocks that are young and have not been buried beyond the range of meteoric diagenesis.
Carbonate Cover Islands: Non-carbonate basement rocks deflect percolating water and partition the fresh-water lens. (Fig. 1B)	**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.
Composite Islands: Non-carbonate basement exposed at the surface, and allogenic recharge is delivered to insurgences at the contacts. (Fig. 1C)	**Glacio-eustatic sea-level fluctuations** impose dissolutional and diagenetic diagenetic imprints reflecting the vertical migration of the lens.
Complex Islands: Interfingering of carbonate/non-carbonate facies and faulting combine to produce complex aquifer features. (Fig. 1D)	**Tectonic uplift and subsidence** overprints the glacio-eustatic imprints with additional dissolutional and diagenetic imprints, as well as structural modifications.

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 Bahamas, being primarily Mid to Late Quaternary in age in a non-tectonic environment, display this eogenetic character extremely well (small portions of Mayaguana Island being an exception).

The Fresh-Water Lens

In any essentially homogeneous body of rock like that of the carbonates forming the Bahamian islands, the fresh-water lens floats on underlying, denser seawater that permeates the subsurface. The model for the ideal behavior of such water masses is the Dupuit-Ghyben-Herzberg model (Vacher, 1988). In reality, variations in rock permeability and other factors result in distortion of the ideal lens shape in the Bahamas (e.g. Wallis et al., 1991). Nonetheless, the Dupuit-Ghyben-Herzberg model serves as a useful first approximation of the relationship between the fresh water and underlying marine groundwater in an island. Figure 8 is a cartoon showing how the fresh-water lens is configured.

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 fresh-water 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 relatively recent past. We can therefore enter dry caves today throughout the Bahamas. In contrast, the blue holes of the Bahamas lead into caves that are currently flooded by seawater. These blue holes represent the cumulative dissolution and

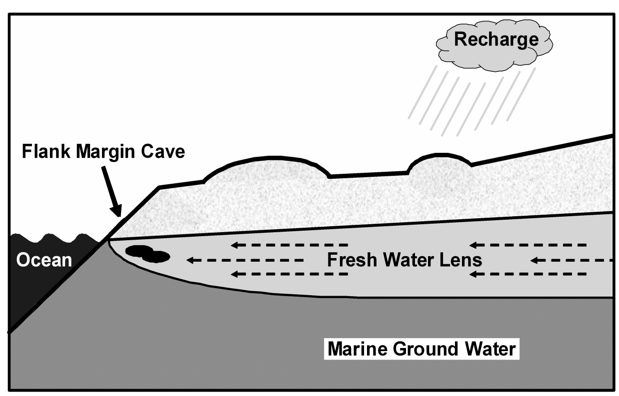

Figure 8. *Cartoon of the fresh-water lens in a carbonate island. Dashed arrows indicate flow within the lens.*

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. Today, Bahamian caves that formed above modern sea-level elevation prior to MIS 5e time lie at or below modern sea level owing to isostatic subsidence of the platforms (Carew and Mylroie, 1995b); however some data can be marshaled that call this interpretation into question (Lascu, 2005) and earlier highstands (e.g. MIS 11) may be involved. Taking isostatic subsidence into account, sea level was high enough to produce the observed subaerial caves for a maximum of about 9,000 years of the MIS 5e time period (or less, given the possible fluctuations of that highstand). 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 seen today represent development during a very short time period in association with small freshwater lenses and with minimal overprinting by later events. Any model that attempts to explain development of these dissolutional caves must operate under extremely tight constraints of time and space.

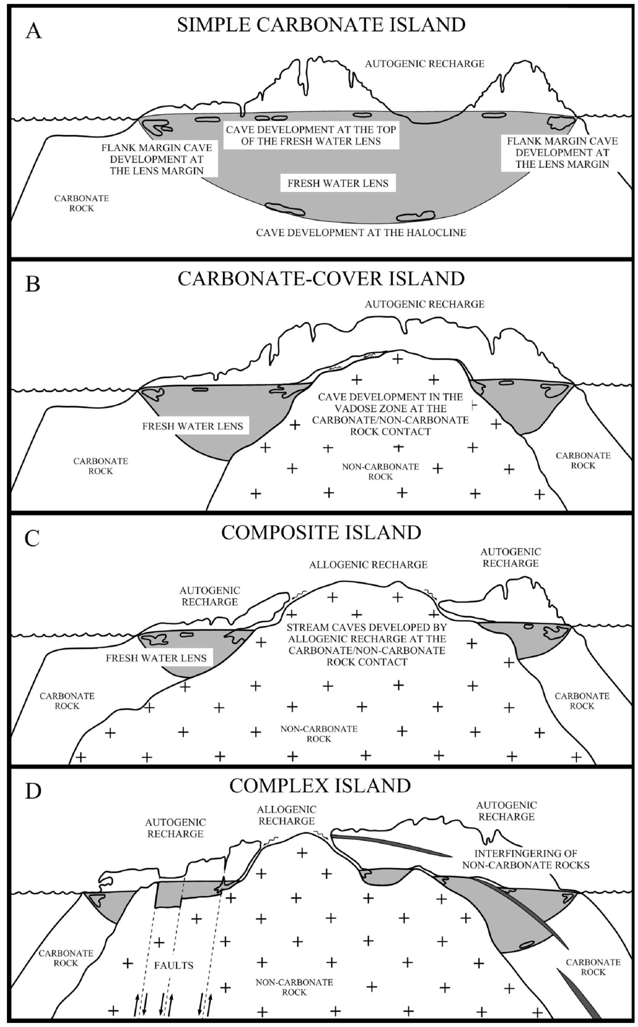

Figure 9. *The Carbonate Island Karst Model, or CIKM, catalogues islands into four forms. Many islands do not fall into strict categories but may show different features in different areas; see discussion in text. From Jenson et al., 2006.*

Carbonate Island Karst Model (CIKM)

The Bahamas were the starting point for the development of what has become the Carbonate Island Karst Model (Jenson et al., 2006) or CIKM. The salient points of this model are shown in Table 1 and Figure 9. The key aspects are that cave and karst development in islands in eogenetic carbonate rocks is very different from continental interiors, where most karst research has traditionally been done. Basically, karst development under the CIKM is controlled by the youthful age of the eogenetic rock involved (almost always Cenozoic, commonly Quaternary), the dissolutional aggressivity provided by mixing of freshwater and seawater, the elevated mircrobial activity associated with the mixing zone, and the change in sea level created by glacio-eustasy and tectonics. Island structure, especially as regards carbonate and non-carbonate rocks, is also crucial (Figure 9).

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 that karst differ**s** little from what would be found in a tropical, continental interior such as in Belize, or Vietnam. In the Bahamas, tectonics and non-carbonate rocks are not a factor, and application of the CIKM is greatly simplified. The Bahamas are solely S*imple Carbonate Islands* (Figure 9).

Karren

Karren are dissolutional sculpturing at the centimeter to meter scale found on exposed and soil-covered carbonate rocks of all ages and types. A variety of etched and fretted surfaces develop; on exposed surfaces this etching is sharp and jagged, whereas on soil-mantled or recently exhumed surfaces it tends to be smooth and curvilinear. Taboroši et al. (2004) reviewed the various terms and mechanisms proposed over the years to describe and explain karren on carbonate islands and coasts, and they provide updated interpretations. The term *eogenetic karren* has been applied by them to these unique etching and dissolution of surface carbonates on eogenetic carbonate islands and coasts. This term replaces the traditional "phytokarst" of Folk et al. (1973) or "biokarst" of Viles (1988). Eogenetic coastal karren is common in coastal areas wetted by sea spray. While biology plays an important part in this karren development, it is the eogenetic nature of the rock that leads to the spectacular morphology. The large amount of primary porosity, the differing character of the allochems, and the variability of the cements of eogenetic carbonate rocks create an environment that easily hosts endolithic boring algae. The biologic activity of those algae, coupled with invertebrate grazers and inorganic mixing dissolution, carve the eogenetic rock in jagged and irregular shapes. Those shapes are reinforced by the pre-existing high variability of the allochem-dominated structure of the host rock. Recrystallized telogenetic rocks, in coastal environments, lack this allochem domination and produce much more subdued karren.

In inland settings, karren are part of the epikarst, the weathered zone on the limestone surface. The epikarst consists of soil and loose blocks of limestone, which over a vertical distance of less than a meter grades into a dissolutionally etched and fretted surface containing numerous small holes and tubes. These holes and tubes in turn rapidly consolidate into widely-spaced discrete flow paths that transmit water downward into the rock mass (see pit caves below). On slopes, the soil and loose rock mantle is commonly removed, and the bedrock karren, tubes and holes are fully exposed.

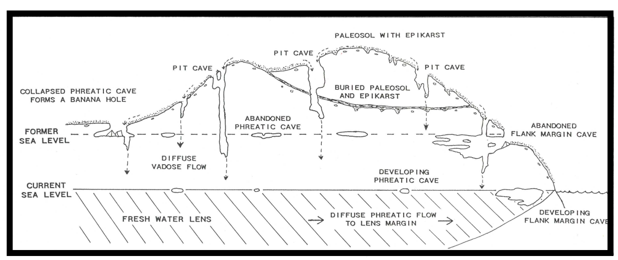

Figure 10. *Cartoon representation of the three main cave types found on a simple carbonate island: pit caves, flank margin caves, and banana holes. Changing sea level moves the fresh-water lens and its dissolutional environments vertically. From Mylroie and Carew, 1995.*

Depressions

Depressions are large basins completely enclosed by surrounding topography. In the Bahamas, large closed depressions, as viewed on topographic maps or on air photographs, represent constructional depressions maintained by subsequent dissolutional processes. These depressions have not been excavated by karst dissolution. The influence of climate on depression development in Atlantic Quaternary carbonate islands, (wet Bermuda versus dry San Salvador) is reviewed by Mylroie et al. (1995a). These large constructional depressions originate as swales between eolianite ridges or as fossil lagoons. Depressions that extend below sea level contain lakes with salinities ranging from fresh to hypersaline, depending on the water budget of a given island, lake surface area, and subsurface hydrology. On San Salvador, yearly evapotranspiration exceeds precipitation and the lakes suffer a net loss of water. This upcones marine water from depth, and further evaporative losses make the lakes hypersaline (Davis and Johnson, 1989). The fresh-water lens is consequently partitioned into smaller bodies underlying the remaining dry land. Some of these constructional depressions are linked to caves (lake drains, see below) that connect to the sea; tidal pumping causes exchange of waters that keeps some of these lakes at marine salinity. Lake margins are little modified by dissolutional processes when lake waters are hypersaline, as on San Salvador, but on wetter Quaternary eolian islands, such as Bermuda, depressions with fresh water enlarge laterally by dissolution (Mylroie et al., 1995a).

CAVES

There are four main types of caves in the Bahamas: pit caves, flank margin caves, banana holes, and lake drains. The first three cave types are well understood, but lake drains remain something of a mystery. Figure 10 is an idealized representation of the major cave features found in the Bahamas, excepting lake drains.

Pit Caves

Pit caves are found throughout the Bahamas, sometimes in very dense clusters, and occasionally at the tops of hills. As the name suggests, these are vertical, or near-vertical, shafts that typically drop 5 to 10 m (Figure 11), often descending in a stair-step fashion, with occasional small chambers. They rarely intersect other cave types, or penetrate to the fresh-water lens. Their walls show classic vertical grooves formed by supercritical laminar flow of descending vadose water. During major rain events, they can be observed to efficiently collect water from the epikarst and conduct it downwards as vadose fast-flow routes. Their high density in places was initially thought, based on water budget considerations, to indicate much higher rainfall conditions at a past time. The high pit cave density is now understood to reflect competition and piracy among pit caves, such that some lose their recharge to "upstream" competitors (Harris et al., 1995). These caves can be complex as a result of this competition, which commonly leads to intersection of pit caves by one another. Pit caves form independently of sea level and fresh-water lens position, and can form in any exposed carbonate rock on an island. They are absent from Holocene rocks, but penetrate a variety of Pleistocene eolianites.

Flank Margin Caves

The largest of the caves that develop in simple carbonate islands form at the distal margin of the fresh water lens, under the flank of the enclosing landmass, so they are called *flank margin caves* (Mylroie and Carew, 1990) (Figure 12). These large caves are commonly entered where a hillside has been breached by erosion, or a portion of the cave roof has

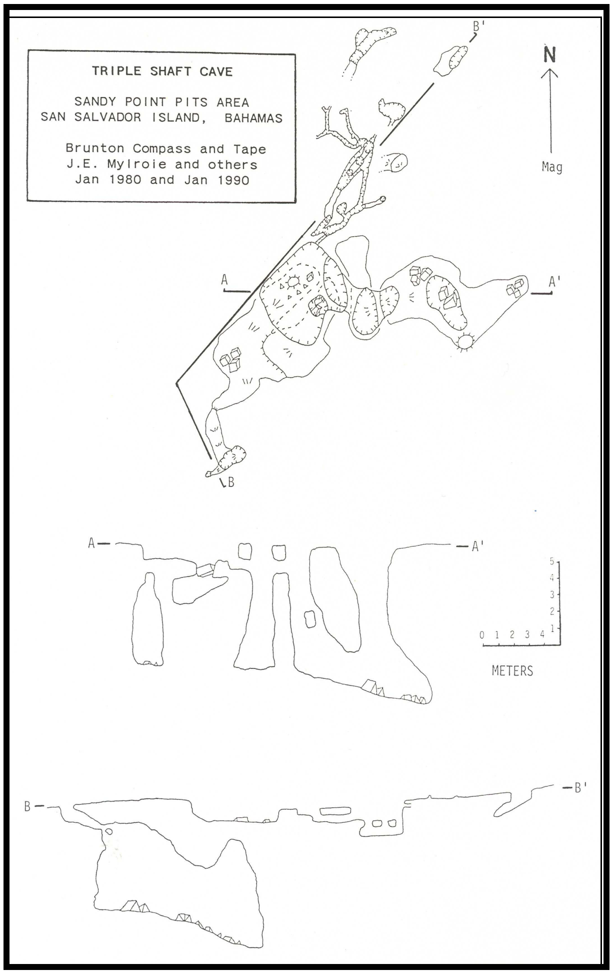

Figure 11. *Triple Shaft Cave, Sandy Point Pits, San Salvador Island, a typical Bahamian pit cave. Note the shallow epikarst water collection system, and the interconnected shaft system. From Harris et al., 1995.*

collapsed. The caves are commonly found at 1 to 7 m elevation, which is in agreement with the position of at least one earlier sea level highstand during the Quaternary, the last interglacial associated with MIS 5e, which lasted from 124 to 115 ka (Thompson et al., 2011). However, examples exist at higher elevations, such as St Francis Grotto on Cat Island (Mylroie et al., 2006) or Conch Bar Cave on Middle Caicos (Smart et al., 2008) and no single explanation encompasses them all. The vast majority of flank margin caves show that sea level reached 6 m higher than at present. Given that the Bahamas are tectonically stable, only a glacioeustatic sea-level highstand could have elevated the fresh-water lens above modern sea level, and so placed the fresh-water lens at, and slightly above, that +6 m elevation. The resulting cave morphology is predictable and consistent: large chambers near the edge of the hill containing the cave, numerous ramifying passages near the back of the cave, and many cross-links and connections. Cave chambers are wider than they are high, with curvilinear and cuspate margins. Remnant bedrock pillars are common. Passages heading inland commonly end in blank bedrock walls. As important as what the caves contain is what they do not contain: no turbulent flow markings such as wall scallops, no stream-laid sediments, no sinking stream or spring entrances. To explain these caves, the flank margin cave model was developed to interpret the size, shape, position and configuration of the caves (Mylroie and Carew, 1990; 1995). The caves develop in the distal margin of the fresh-water lens, just under the flank of the enclosing landmass (Figure 10). At the distal margin of the fresh-water lens, the mixing environment of the vadose input to the water table is superimposed on the mixing environment of the fresh-water lens with underlying marine water, thereby increasing dissolution beyond what either environment could do alone. Additionally, the lens cross section decreases at the lens margin, so flow velocities increase, transporting reactants in, and products out, faster than elsewhere in the lens (Raeisi and Mylroie, 1995). Finally, both the top of the lens, and the halocline, are density interfaces that can trap organic material. Oxidation and microbial consumption of the organics creates CO_2 that can drive more dissolution; excess organics can create anoxic conditions and drive H_2S-mediated dissolution. The H_2S model appears supported by ^{34}S analysis of intergranular gypsum from flank margin caves on San Salvador, which showed depletion values (-24) associated with biomediation of

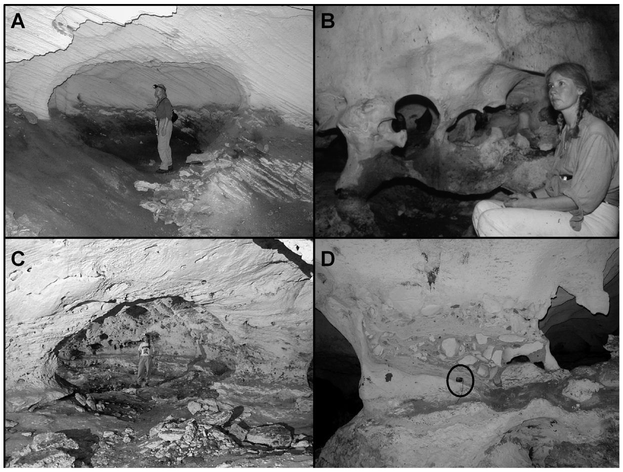

Figure 12. *Images from Bahamian flank margin caves. A, Hamiltons Cave, Long Island, showing horizontal dissolutional roof cutting eolian beds, brown horizon is where guano was mined; B, phreatic dissolutional fretwork, Harry Oakes Cave, New Providence; C, Chambers, pockets, and vugs, Ten Bay Cave, Eleuthera; D, paleokarst infill cut smooth by dissolution, 1702 Cave, Crooked Island (compass in circle 8 cm long).*

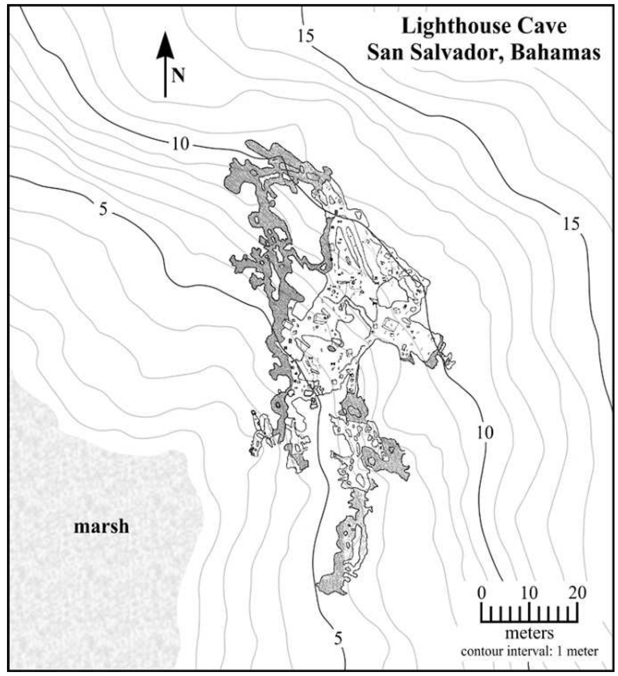

Figure 13. *Map of Lighthouse Cave, San Salvador Island. The flank margin cave has one large central chamber with interconnected and ramifying dead-end passages. The cave follows the flank of the hill contour. Gray areas contain standing marine water today.*

sulfur in anoxic zones (Bottrell et al., 1993). There has been debate in the literature regarding the relative importance of the three mechanisms (e.g., Schwabe and Carew, 2006), but recently it has been suggested that the hydrologic condition of rapid lens-margin discharge alone, and not the geochemistry, is the critical factor (Moore et al., 2007).

Flank margin caves, as described earlier, have a variety of morphological features, all consistent with a non-turbulent, phreatic origin. The oval or globular chambers that are oriented parallel to the longitudinal trend of, and just under the flank of, the ridge in which they have formed, indicate the importance of the lens margin (Figure 13). Small radiating tubes extend from these large chambers into the ridge interior where they end abruptly or pinch out. These passages represent individual diffuse flow paths that delivered fresh water into the flank margin mixing area. The abrupt end of these passages reflects the position of the mixing front when sea level fell and the caves became subaerial. One final interesting aspect of flank margin caves is that they occur on scales from small chambers up to immense caves without loss of their general morphology or position with respect to the land surface. Size-rank plots reveal that the caves self-select in three size groups based on growth of the initial voids, amalgamation of voids, and then amalgamation of clusters of voids (Roth et al., 2006). Computer modeling of void genesis creates the same plot, but adds a fourth group of tiny caves that were ignored by cave surveyors (Labourdette et al., 2007).

The general morphology of flank margin caves is similar to that of other caves formed under different mixed-water conditions (Mylroie and Carew, 1995), such as in the Guadalupe caves of New Mexico. This pattern of globular chambers, maze-like passage connections, thin wall partitions, and dead-end passages are called spongework or ramiform caves Palmer, 1991), typical of development under *hypogenic* conditions, decoupled from direct surface hydrology. Complete discussion of the fresh-water lens, cave development, and the flank margin model can be found in Mylroie and Carew (1990; 1995), and Mylroie and Mylroie (2007).

Banana Holes

Abundant, but smaller caves that develop in the flat plains of the Bahamas, away from the modern lens margin are called *banana holes*. Banana holes are circular to oval chambers 5 to 10 m in diameter, and 1 to 3 m high, with phreatic morphologies but lacking the size and passage ramifications found in flank margin caves (Figure 14). They are located 1 to 7 m above sea level, but laterally they are well away from where the lens margin would have been, under current topography, with sea level at that elevation. They are entered where their ceilings have collapsed, or rarely where a pit cave has intersected them. They can be found in dense concentrations, up to 3000/km^2 (Harris et al., 1995). Occasionally, a collapsed banana hole has a connection with an adjacent, uncollapsed banana hole. In the Bahamas, they commonly develop in the broad, low-elevation regions that make up significant parts of these islands. As a result, their roofs are thin and prone to collapse. Once collapsed, banana holes collect soil and vegetative debris, as well as water and are favored for the growing of specialty crops, such as bananas, which is how their name was derived (Harris et al., 1995). The initial interpretation (Harris et al., 1995) was that their smaller size relative to flank margin caves was a result of their position away from the higher flow velocities of the lens margin. That they existed at all seemed to indicate that vadose-phreatic mixing of fresh water, and biogeochemical activity, are still important in carbonate dissolution. Recently, it has been proposed that banana holes are found preferentially in subtidal units overlain by beach and dune sands (Mylroie et al., 2008; Infante et al., 2011), and formed syndepositionally as lagoons were infilled by prograding sands, and the fresh-water lens followed. The small size of these features reflects the transient time they were in the lens margin before the progradation continued seaward; they are immature flank margin caves.

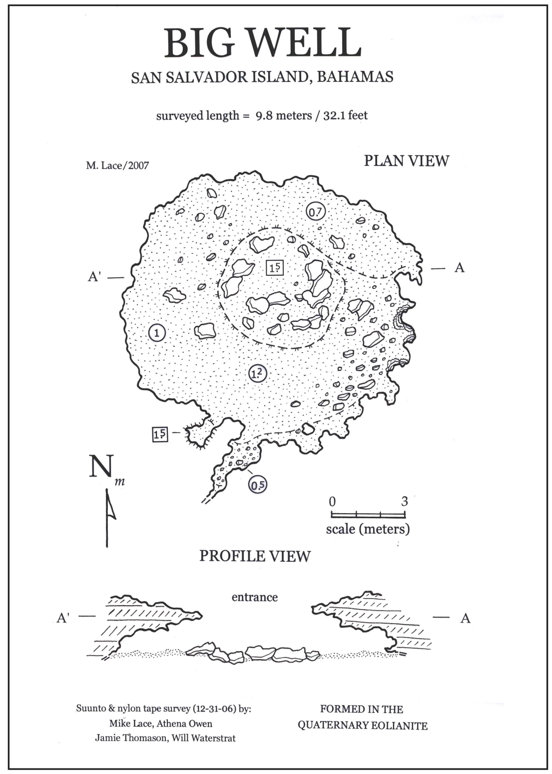

Figure 14. *Map of Big Well Cave, San Salvador Island, a typical banana hole with one low, wide oval chamber. Note the thin roof, which promotes collapse. Cartography by M. Lace.*

Lake Drains

Lake drains are a contrived term used to describe small, un-enterable conduits that carry water into and out of lakes in a pattern related to tides. Their presence results in lakes that commonly maintain marine salinity despite climatic conditions that might favor freshwater, such as in the northwestern Bahamas, or hypersaline water, such as in the more arid southeastern Bahamas. Some lakes are large, and their lake drains are small, such that the conduit flow from the sea is not sufficient to greatly modify the lake salinity. Unlike flank margin caves and banana holes, lake drains are true conduits. Their mode of origin and configuration are poorly understood, but their influence can be important on lake geochemistry and as a result, lake ecology (Davis and Johnson, 1989). They remain cryptic features.

Depressions Revisited

In most telogenetic carbonate settings found in continental interiors, the smaller closed contour depressions, sinkholes (American terminology) or dolines (European terminology), form quite differently than those in eogenetic carbonate islands such as the Bahamas. Sinkholes form by three main methods: 1) carbonate dissolution by focused epikarst flow to create small, saucer-shaped depressions; 2) collapse and subsidence of a thick soft sediment or soil mantle into a small dissolutional opening; and 3) collapse of the bedrock roof of a dissolutional chamber. In telogenetic continental carbonates, the longer time span available results in both a thick residual soil mantle, and more time for bedrock dissolution. The telogenetic character of the bedrock results in epikarst flow focusing to joint intersections and related flow pathways. Sinkholes formed by dissolution and soil subsidence are very common. In contrast, cave passages are relatively rare (less than 1% of the porosity), so collapses into bedrock voids are also relatively rare. In the Bahamas and related eogenetic environments, there are no thick soils to subside downward, and dissolutional openings, even large ones like pit caves, open flush to the land surface. There has been little time since bedrock deposition for dissolution of saucer-shaped sinkholes. In contrast, because flank margin caves and banana holes are extremely common, often with relatively thin bedrock roofs, collapse sinkholes are common, and they are the dominant small depression type in the Bahamas. The rule of thumb in the Bahamas for depressions is: large ones are constructional, and small ones are cave collapses.

Blue Holes

Famous karst features of the Bahamas are the blue holes. 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. (1995b). A different approach to defining and describing blue holes can be found in Schwabe and Carew (2006). Blue holes are defined here 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., 1995b, 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” (Mylroie et al., 1995b, p. 225). The most common alternative use of the term “blue hole” is to describe large and deep karst springs in continental interiors (Mylroie et al., 1995b). 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

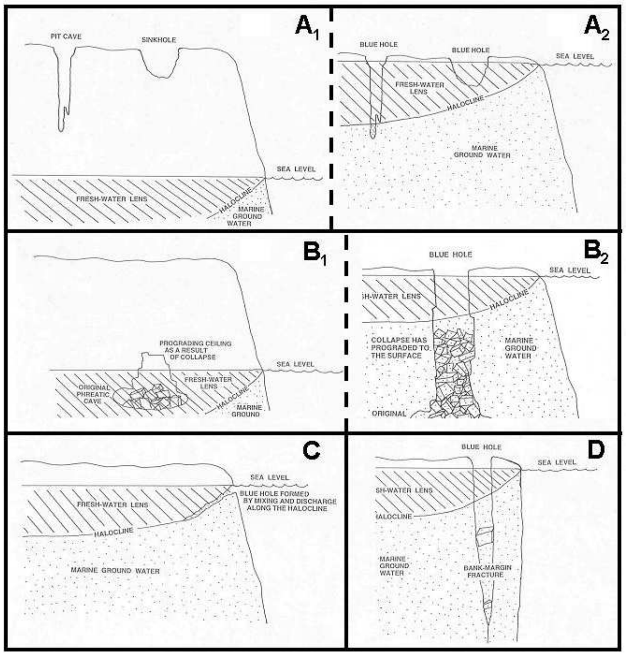

Figure 15. *Cartoon of the polygenetic origin of blue holes: A_1 and A_2, pit caves formed at low sea levels are flooded by sea level rise; B_1 and B_2, dissolutional voids formed at lower sea levels prograde upwards by collapse; C, dissolution along halocline (note image is vertically exaggerated at least 10x); D, bank margin collapse creates water-filled fractures. From Mylroie et al., 1995b.*

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 (Palmer, 1985).

Blue holes are flooded karst features of polygenetic origin. They have developed in a variety of ways (Figure 15). 1) Pit caves that formed during sea-level lowstands are flooded during sea-level highstands to form some blue holes. 2) In places, failure of the steep margins of the Bahama banks produces bank-margin fractures. Wide gaps along these fractures can result in blue holes (Palmer, 1986; Carew and Mylroie, 1989a; Carew et al., 1992). 3) Collapse of large dissolutional voids at depth that were produced during sea-level lowstands or pre-Pleistocene highstands can prograde to the surface to form blue holes. Blue holes may open laterally into horizontal cave systems at a variety of depths. Some of these connections may be random intersection of caves formed by other means. Recently, it has been proposed that blue holes out on the banks have become infilled with Holocene sediments, and the exhalation of diffuse flow water from these buried structures by tides is responsible for the enigmatic whitings of the Bahama Banks (Larson and Mylroie, 2012).

Pseudokarst

Pseudokarst means literally “false karst”. A good example would be kettle holes in glacial sediments, or lava tubes in volcanic terrain. In the Bahamas, with 100% carbonate outcrops, how does one get pseudokarst? There are actually two types of pseudokarst in the Bahamas: tafoni and sea caves. Tafoni are hollows and undercuts that form in rock cliffs as a result of a variety of subaerial weathering processes independent of dissolution. They can resemble breached flank margin caves. The term has many definitions, and the reader is referred to Owen (2007) for a complete discussion of the term, its definition, and the processes that make tafoni in the Bahamas. Tafoni are abundant on cliffs in eolianite facies, as cliffing exposes the poorly-cemented interior to subaerial processes such as wind and drying, while the vertical face limits meteoric cementation. Marine cliffs, because of cementation by sea spray, don’t form tafoni unless the sea cliff is very high, or the result of episodic storm events, such that sea spray is a limited factor. Collapse of sides of an eolian ridge into a cave or blue hole creates a vertical cliff in which tafoni readily form. Tafoni are found in Holocene Rice Bay Formation eolianites, indicating that they can form quickly. Small tafoni are found in road cuts and quarries, even in sawn blocks of eolianite used in buildings, which documents the rapidity of the process. Tafoni in protected Holocene eolianite sea cliffs are the same size as those in protected coastal Pleistocene eolianites, indicating that it was Holocene sea level that created the cliffing to form the tafoni. Descriptions of tafoni on San Salvador can be found in Owen (2007). Relict tafoni formed high in MIS 5e sea cliffs are also described from Abaco (Walker, 2006; Walker et al., 2008).

Sea caves are the result of wave action on a rocky coast. Sea caves are found in all rock types, and most literature describes them as forming at major structural weaknesses such as faults, dikes, and lithologic contacts (Waterstrat, 2007; Waterstrat et al., 2010). In the Bahamas, the Quaternary eolianites don't have faults, dikes, or lithologic contacts, but sea caves are abundant, implicating wave energy patterns, and not rock characteristics, as the major control of the spacing and size of the resulting sea caves. Detailed descriptions of sea caves on San Salvador can be found in Waterstrat (2007).

It is important to be able to differentiate tafoni from sea caves, and both from flank margin caves. Both sea caves and flank margin caves are a measure of sea level at their time of formation, but only flank margin caves provide information on the nature of the fresh-water lens that formed them. Sea caves and flank margin caves found in island interiors from a past, higher sea level event(s) provide information on denudation rates. If an open chamber is found at the base of an eolianite ridge, and it is identified as a sea cave, then erosive denudation has been minimal. If the chamber is a flank margin cave, then enough hillside has been eroded away to breach into the cave. Mis-identifying sea caves as flank margin caves, or vice versa, can create problems in interpreting denudation rates and paleohydrology. Tafoni are not restricted to sea level, and so misidentifying a tafoni as a paleo sea cave or a breached flank margin cave will result in an incorrect sea level interpretation. Tafoni misidentified as breached flank margin caves also will lead to incorrect paleohydrology interpretations. Owen (2007) and Waterstrat et al. (2010) provide a quantitative methodology using area to perimeter ratios of cave maps, and entrance width to maximum width ratios, to differentiate the three cave types, with varying degrees of success.

Island Area Versus Cave Development

Cave divers report large and long conduit cave systems in the Bahama banks at depths averaging 20 m or deeper (Palmer, 1985). Such active conduit caves are not found in the Bahamas today at or near modern sea level. At a sea level position of -20 m, the Bahama Banks would be mostly exposed, and most current islands would become significantly larger, or join into a group of mega-islands. Today, 11,400 km^2 is the sum of the island area; if the banks were exposed, that area would increase more than an order of magnitude to 136,000 km^2 (Meyerhoff & Hatten, 1974). As an island becomes larger, its perimeter increases linearly, but its area, and hence meteoric catchment, increases by the square. Mylroie and Mylroie (2007) suggest that at some point, as islands enlarge, diffuse flow in the fresh-water lens to the island perimeter cannot adequately address the water budget, and that conduit flow initiates. In contrast, flank margin caves can be found in islands as small as a few hundred meters long and a few tens of meters wide. The cave diver data indicate that at some size threshold, the diffuse flow that makes flank margin caves becomes too inefficient, and turbulent conduit flow initiates. Such long conduit caves have been also found at 20 m depth on Bermuda (Mylroie et al. 1995a), where a drop of 20 m would expose the large lagoon north of the present island.

The area to perimeter ratio of islands may create conditions that favor conduit flow for larger islands. Blue holes, especially the large, deep circular ones, are most likely the result of progradational collapse from deep-seated voids. Some of these voids may relate to large conduits developed during Quaternary glacioeustatic lowstands, and as a result, most deep blue holes end at about -100 m. Deeper blue holes such as Deans Blue Hole at 200 m, may relate to older, deeper voids such as those described by Meyerhoff & Hatten (1974). Such deep voids may have formed in a shallower environment, and subsided with the platform to their present location. In any event, progradational collapse allows blue holes to integrate cave systems at a variety of levels without requiring a dissolutional mechanism for that integration. Dissolutional voids may have initiated the collapse process, but the linkage of shallow and deep cave systems, as well as with the surface, is a non-dissolutional action. Therefore blue holes represent an old, laterally and vertically complex environment that can penetrate a variety of depths, with a variety of organic loadings, and a variety of water chemistries. This complexity has been implicated in the taxonomic diversity of blue holes (Mylroie and Mylroie, 2011)

Karst Summary

Island karst as seen in the Bahamas is dominated by sea-level position, the fresh-water lens, and the eogenetic nature of the carbonate rocks. The caves and karst features that form on islands differ dramatically from similar features in continental interiors. Because island karst occurs in or near the carbonate depositional environment, preservation of such karst as paleokarst is more likely than preservation of karst features in continental situations.

FIELD TRIP STOPS

Stop One: Caves Point

This stop is a well-known location as it is labeled on the New Providence topographic sheet 1, just east of the intersection of Blake Road with West Bay Street . Recently, the area has undergone much tourist development, such that the cave, once on a lonely hillside, is now surrounded by buildings (see cover photo and Figure 16 for a 1988 view). Two of the best-known caves in the Bahamas are found here (Mylroie et al., 1991). There are two main caves that are easily seen from the road. They are not currently connected, but they may have been when they formed. Erosion of the hill side in which they developed has removed their outer chamber walls and exposed their interiors. The entrance chamber floors are about 3 meters above sea level.

1. Caves Point East Cave - This cave is basically a large, single chamber. The entrance is 21 m wide and averages 3 m high, and is oriented east-west. The chamber extends 16 m to the south (Figure 17). It has a bedrock floor that rises to the rear of the cave, and the ceiling is breached by two dissolution pits. The cave has a simpler plan than those of most Bahamian flank margin caves, in that it lacks a series of complex tubes and passages radiating from the main chamber.

2. Caves Point West Cave - This cave is similar to its eastern neighbor, except that the large, broad entrance chamber has passages off the back that lead to a large inner chamber. It is conceivable that such a chamber also lies behind the East Cave, as the flank margin model does not necessitate that all chambers be connected. The West Cave is a simple cave with two large chambers that overlap and are connected by a few openings that are a few meters in diameter (Figure 17). The entrance is

Figure 16. *Caves Point West Cave, in 1988 (see cover for Caves Point East Cave).*

23 m wide and the east-west oriented outer chamber is an average of 3 m high. The entrance is the breached wall of a room that extends to the south about 9 m. To the west are a several small chambers. The floor of the entrance chamber slopes upward. Three holes in the floor of the chamber, and two holes at the back of the chamber lead downward to a second, inner chamber that lies under and to the south of the entrance chamber. This room, which extends to the south, is 22 m long and 12 m wide. The ceiling is an average of 5 m high. There is a low wide chamber developed farther to the west. The floor of this passage floods during high tide. Both chambers have smooth, phreatic dissolution surfaces because they initially developed below the water table. A series of steps cut into the stone lead from the northernmost part of the lower chamber southward and up to the top of a collapse-block in the center of the room. The whole setting is interesting and leads the imagination to thoughts of pirates and buried treasure.

The large open entrances to both caves make them appear, at first glance, to be sea caves formed during a past higher sea level. This is especially true for the East Cave. However, close examination reveals flowstone and dissolved rock surfaces that are characteristic of cave development below a water table followed by a drop of the water level and the deposition of secondary subaerial calcite (flowstone). The features of the inner chamber of the West Cave could only have formed by

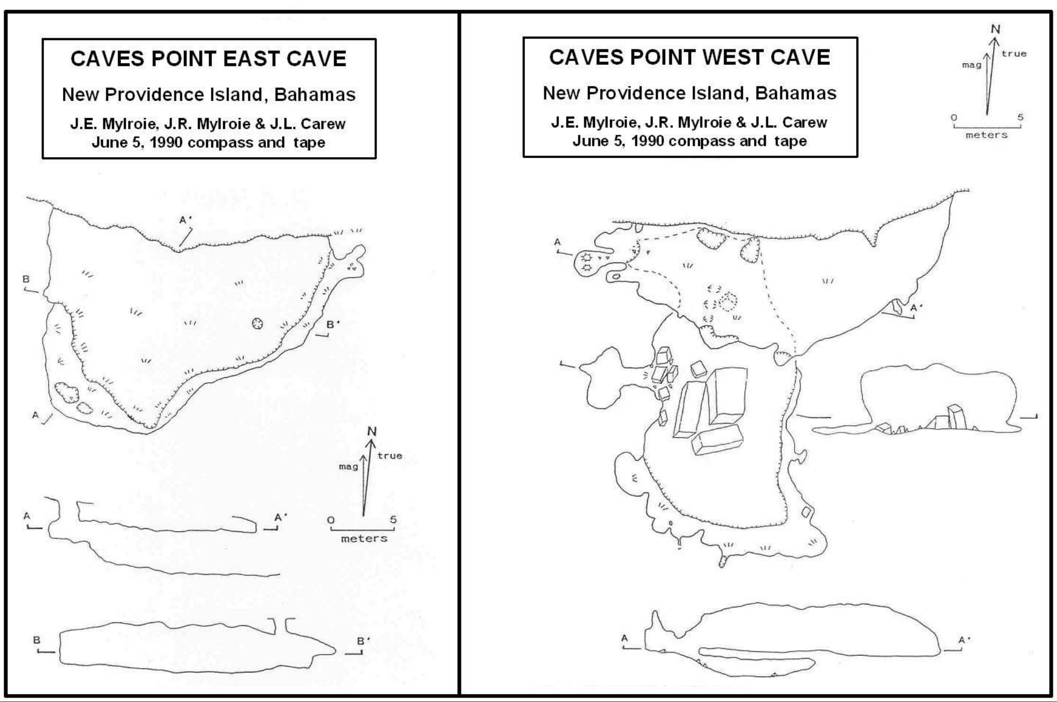

Figure 17. *Map of the Caves Point East and West Caves (Mylroie et al. 1991).*

chemical dissolution. Both caves lack the many small tubes and interconnections found in most flank margin caves, but the overall form fits the model.

Stop Two: Northwest Point

Northwest Point, as its name implies, is at the northwest extremity of New Providence Island, just east of where the island coast turns to the southwest. How much can be seen here depends on the amount of sand on the beach, which varies from season to season.

The height of the outcrop is about 2 m; the top 1.5 m is a back-beach deposit of the Holocene Rice Bay Formation (Figure 18). This peloidal deposit lies unconformably on a red to orange micritic crust, which represents a *terra-rossa* paleosol developed on Pleistocene rocks. The elevation of these back beach facies, in agreement with current sea level, and the lack of a paleosol overlying them indicates that they belong to the Hanna Bay Member. The radiocarbon age of the allochems in the deposit is 2,737 +/-80 yBP, which is consistent with this field interpretation. Beneath the paleosol are numerous fossil corals. While an absolute date is not available from this specific site, the presence of well preserved, *in situ* fossil corals indicates that the unit formed during MIS 5e approximately 120,000 years ago (Carew and Mylroie, 1995b), which places it in the Cockburn Town Member of the Grotto Beach Formation. The fossil reef can be followed laterally along the outcrop for a few tens of meters before it is replaced by a shelly calcarenite.

The presence of the paleosol directly on the fossil reef yields some clues as to the denudation rate of the limestone surface during the 100,000+ years that it was subaerially exposed prior to the deposition of the overlying eolianite. Fossil corals have not been found in

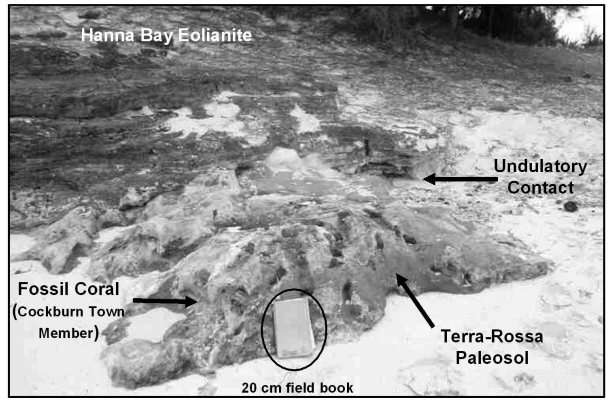

Figure 18. *Northwest Point, showing Holocene Hanna Bay eolianites overlying a late Pleistocene Cockburn Town Member fossil reef, with a terra-rossa paleosol in between.*

the Bahamas above 4 m, so the outcrop represents a possible maximum of 3.5 m of denudation of fossil reef since the end of MIS 5e about 115,000 years ago, and the deposition of the eolianite about 2,700 years ago, a maximum denudation rate of ~3 m/100 ka. The rate could be higher if the micritic portion of the *terra-rossa* paleosol, once formed, acted to protect the underlying rock. Dissolutional denudation on carbonate islands is often overlooked when assigning sea-level positions to exposed fossil reefs. On Guam, Miklavic et al. (2012) demonstrated that a minimum of 5 m, and a maximum of 8 m, had denuded from a MIS 5e fossil reef (a minimum rate of ~5 m/100 ka).

Stop Three: Lyford Cay Roadcut

Exit the bus and examine the roadcut. Be wary of passing cars, and treat the grounds with respect. The Lyford Cay roadcut is a beautiful example of eolianite packages separated by paleosols, which has been examined by numerous geologists in the past 25 years or so (Garrett and Gould, 1984; Muhs et al., 1990; Carew and Mylroie, 1991; Carew et al., 1996; Hearty and Kindler, 1997; Reid, 2010).

The lowermost rock body consists of bioclastic-peloidal grainstone (up to 62% bioclasts) (Figure 19) (Hearty and Kindler, 1997), whereas the other units are predominantly composed of oöids and peloids (up to 80% oöids, and no more than 6% bioclasts (Figure 20) (Hearty and Kindler, 1997). There are six sedimentary packages recognized here that are wholly or predominantly eolianites. Based on the presence of fenestral porosity and low-angle planar bedding, Garrett and Gould (1984) and Reid (2010) interpreted two of these packages as eolianites that grade downward into beach facies and suggested that sea level was around +10 m above present during deposition of these units. Fenestrae or keystone vugs are usually interpreted as diagnostic of an intertidal setting (Dunham, 1970). However fenestral porosity associated with eolian sedimentary structures has been observed up to elevations of 18 to 40 m above msl in Bermuda (Kindler and Strasser, 2000), on Providenciales (Caicos Platform; Wanless and Dravis, 1989), and in the Bahamas (Bain, 1985; Kindler, 1991). Two conflicting interpretations have been proposed to explain these peculiar occurrences: heavy rainfall (Bain and Kindler, 1994; Kindler and Strasser 2000,2002) and giant-wave action (Wanless and Dravis, 1989; Hearty et al., 2002) which, in any case, show that keystone vugs are not always a reliable indicator of intertidal deposition and thus of past sea levels. The lack of an early generation of fibrous marine cement in these fenestrae-rich grainstones (Figure 20) supports a non-intertidal depositional setting for these beds.

A recent study of the geology of New Providence (Reid, 2010) illustrated outcrops at a new NNW-SSE roadcut through an ~1 kilometer-wide arcuate dune ridge (referred to as Serenity Middle) near Lyford Cay. That ridge is catenary between Northwest Point and the western end of Lyford Cay, and comprises three superposed ridges (Reid, 2010). Reid (2010) reports that the rocks at that road cut are primarily peloidal grainstones.

The paleosols separating the eolianite units at Lyford Cay are not all of the same type. Some

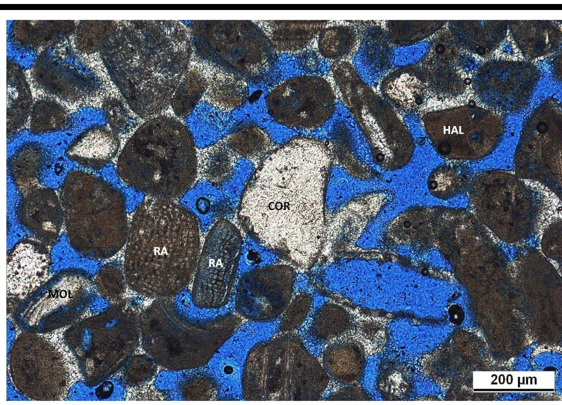

Figure 19. *Sample NP 49. Microfacies of the lowermost eolianite (Owl's Hole Formation) at Lyford Cay. Note the predominance of bioclastic particles many of which are partly micritized (HAL = Halimeda fragment, COR = coral debris, RA = red-algal clast, MOL = mollusk fragment*

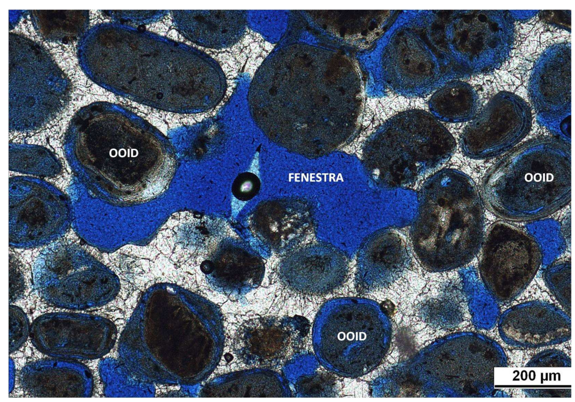

Figure 20. *Sample NP 54. Microfacies of fenestrae-rich oölitic-peloidal limestone in the Grotto Beach formation. Note the absence of early fibrous cement that would indicate a marine diagenetic setting.*

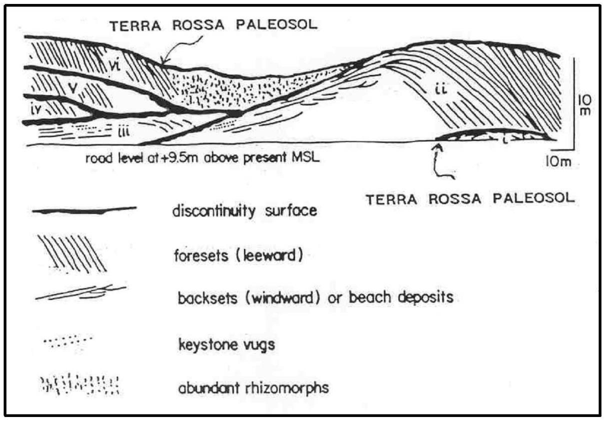

Figure 21. *Roadcut section at Lyford Cay, displaying eolianites and paleosols in a stacked sequence. See text for details. Modified from Garret and Gould, 1984.*

paleosols are calcarenite protosols, and some are *terra-rossa* paleosols (Carew and Mylroie, 1991). *Terra-rossa* paleosols and calcarenite protosols commonly form boundaries between deposits of Quaternary carbonates, especially eolianites. The term *terra-rossa* paleosol refers to the red to orange-brown *terra rossa*, and associated soil-related structures such as vegemorphs (rhizomorphs, rhizocretions), hard micritic calcrete crusts, vadose pisolites (soil glaebules), and disarticulated blocks of weathered parent material. These paleosols represent long-term exposure of the parent material to surficial weathering, and the micritic calcrete is similar to caliches found in other semi-arid environments. They also contain mineral components (e.g., hydrous iron oxides, quartz, etc.) derived from atmospheric dust (Muhs et al., 1987, 1990). Such paleosols are common in Bermuda and in the Bahamas, and they developed primarily at times when sea level was low (below -10 m) and the platforms were emergent (Bretz, 1960; Land, et al., 1967; Carew and Mylroie, 1991; Vacher and Harmon, 1987; Vacher and Hearty, 1989; Nawratil de Bono, 2008). During those times no new carbonate sediment accumulated on the platforms, because banktops act as “carbonate factories” only when they are inundated by high sea levels (Carew and Mylroie, 1997).

Calcarenite protosols are composed of organics and eolian-transported sediment captured by the baffling effect of vegetation on carbonate land surfaces (Harmon et al., 1983). These white to pale red/orange, unstructured, and poorly indurated buried soils commonly contain abundant fossil pulmonate snails (*Cerion*). They are generally concordant with the depositional layering of the underlying unit, which is usually a back-beach or dune deposit. Calcarenite protosols develop during periods when the carbonate banks are at least partially flooded, and carbonate sediment is being produced. They represent pauses in the accumulation of the larger depositional packages within which they are contained, as opposed to major breaks between separate sea-level highstands and their associated sediment packages. On Bermuda, these features were initially called "soils of accretion" (Sayles, 1931). These soils were later termed "accretionary soils" (Land et al., 1967; Harmon et al., 1983; Vacher and Harmon, 1987), but more recently they have been termed calcarenite protosols (Vacher and Hearty, 1989; Carew and Mylroie, 1997).

An example of the importance of distinguishing between *terra-rossa* paleosols and calcarenite protosols is provided by this stop at Lyford Cay. Here, Garrett and Gould (1984) described a package of six sedimentary units (their units i to vi) separated by discontinuity surfaces marked by paleosols (Figure 21). However, close examination of the outcrop shows that the lowest unit, Garrett and Gould's unit **i**, is separated from the upper units by a *terra-rossa* paleosol (Carew and Mylroie, 1991); whereas the surfaces separating their units **ii** through **vi** are calcarenite protosols. The entire sequence is capped by a second *terra-rossa* paleosol. Garrett and Gould (1984) placed the entire sequence within one depositional phase (their phase 2), whereas we suggest that unit **i** was deposited during a glacioeustatic highstand of sea level, then a fall of sea level resulted in a substantial interval of non-deposition and the development of that lower *terra-rossa* paleosol. Subsequently, sea level again rose onto the platform top and episodic eolian deposition produced alternating eolianite and calcarenite protosol sequences. Then, a second sea level fall produced the conditions for development of the upper *terra-rossa* paleosol.

Uranium/thorium analysis of oöids (Muhs et al. 1990) and whole-rock amino-acid data (Hearty and Kindler, 1997) in this sequence at Lyford Cay support Carew and Mylroie's (1991) field interpretation. U-Th ages from Unit **i** range between 197 and 212 ka (Muhs et al., 1990) suggesting a correlation with MIS 7, whereas the alloisoleucine/isoleucine (A/I) ratio of 0.674 obtained by Hearty and Kindler (1997) is more consistent with a stage 9 or 11 age. Such an older age is most likely, because it is generally accepted today that the MIS 7 highstand was well below today's high sea level (MIS 1) (Lisiecki and Raymo, 2005). Units **ii** through **vi** were deposited during MIS 5 (specifically substage 5e), as confirmed by AAR (Hearty and Kindler, 1997) and U-Th data (Muhs et al., 1990), although one age obtained by these authors correlates better with substage 5c (106 ± 2 ka). These data place the upper unit (Garret and Gould's **ii-vi**) in the Grotto Beach Formation (Carew and Mylroie, 1995b, 1997), and the underlying unit (Garrett and Gould's **i**) in the Owl's Hole Formation. Recognition of the difference between *terra-rossa* paleosols and calcarenite protosols is therefore important when relating carbonate sediment packages to their corresponding sea-level highstand events. This outcrop further illustrates the difficulty in using morphostratigraphy for geologic mapping, for in the absence of the road cut the ridge would have been incorrectly mapped as a single eolianite. See also the inside of the back cover.

Re-board the bus.

Lunch Stop – Clifton Heritage National Park

Our lunch stop site will be the picnic area of this newly established national park opened in 2008. Located off Southwest Road adjacent to the industrial area of Clifton Pier, this beautiful park is on the site of a former Loyalist-era plantation, established in the late 18th century, and includes ruins of numerous plantation structures as well as other important pre-Loyalist archeological sites. Today the park offers multiple recreational and educational activities from camping to bird watching to archaeological and nature educational tours. Future plans are ambitious and can be found on the park website:
http://www.bahamascliftonheritagepark.org.
As time permits, one might want to take a short stroll to visit the "Sacred Space" site and view the intriguing figures carved into *Casurina* tree trunks (Figure 22) by the Bahamian sculptor Antonius Roberts to "mark the triumph of hope and determination to conserve our heritage." Another point of interest are the plantation-era stone steps cut into the limestones of the Clifton cliffs and leading to the remains of an old stone wharf (Figure 23). These cliffs reveal the best exposures of the Upper Pleistocene MIS 5e sequence on New Providence, as discussed in the following Stop 4 description.

Stop Four: Clifton Pier

The sea-cliff exposures and associated anthropogenic cuts in the rock along the southwest coast of New Providence (Fig. 23) provide a classic vantage point on the island for close examination of Upper Pleistocene (MIS 5e) shallow subtidal to beach grainstone beds. Because deep water lies immediately offshore, the Clifton Pier area today is heavily developed with docks for transport and storage of petroleum products and other shipping and

Figure 22. *Lunch stop, Clifton Heritage National Park. Three of the figures carved into Casurina tree trunks at the "Sacred Space" site.*

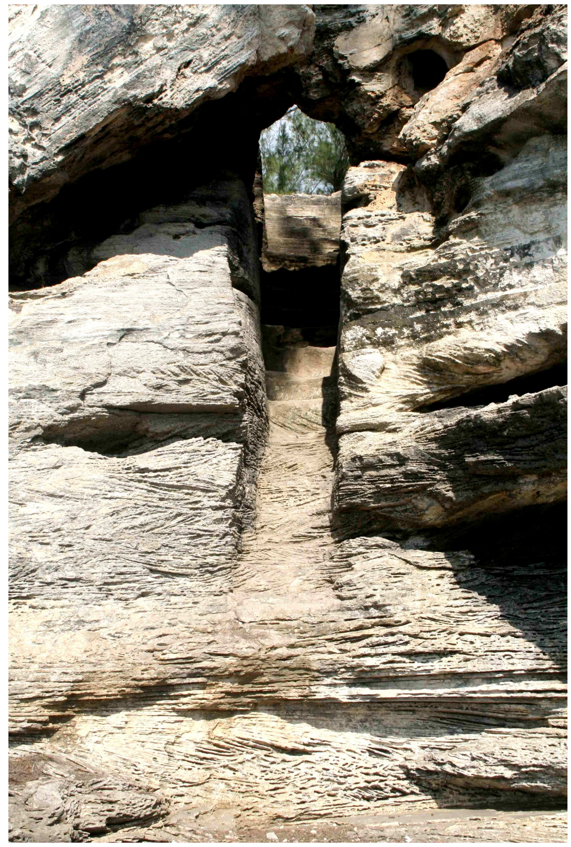

Figure 23. *Lunch stop, Clifton Heritage National Park. A looking-upward view of the stone steps carved into limestone strata of the Clifton cliffs. Well-defined cross-bedding is characteristic of these strata. The steps lead to a small stone wharf that was active during the Loyalist period and maybe earlier.*

industrial activities (Figure 24). As a result, the area now is far from a natural state, and much of the adjacent coast is not easily accessible, particularly for a large group. We will visit the old Clifton Pier dock site (N 25°00'15"; W 77°32'38"). ***Please be careful and mindful of others in our group when climbing on the cliffs and moving along the coast.***

These cliffs have been investigated by a number of geologists in the past, including the seminal studies of Ball (1967) and Garrett and Gould (1984), and more recent investigations by Aurell et al. (1995), Hearty and Kindler (1997), and Reid (2010). This site also was visited in 1996 as part of a Bahamian Field Station symposium field trip (Carew et al., 1996). In preparation for that trip, Curran and White (1996) constructed a stratigraphic section adjacent to the east side of the concrete dock at Clifton Pier (Figure 25). With a total thickness of over 6 m, this section reveals much about the vertical facies sequence, bedding styles, and physical and biogenic sedimentary structures

Figure 24. *Clifton Pier. View looking southeast toward the docks and other industrial structures along the cliffs today.*

typical of Bahamian subtidal to beach transition sequences.

The rocks consist mostly of peloidal-skeletal (bivalves, gastropods, green and red algae, serpulids, and benthic foraminifers) grainstones with variable oöid content. The lowest outcropping beds laterally adjacent to Clifton Pier commonly contain up to meter-scale boulders in trough cross-bedded to planar-bedded grainstones (Figure 26). These boulders have planar lamination and are of similar composition to the host rock, suggesting that they formed by erosion of near-contemporaneous, lithified beach deposits in an energetic shallow subtidal zone. The lower units of the cliffs are mostly overlain by finer-grained peloidal-skeletal grainstones that display well-formed trough cross-beds, including sets of herring-bone cross-beds that decrease in thickness upward (Figure 27). The distinctive, conical trace fossil *Conichnus conicus* is common in these units and occurs up to 1 to 2.5 m above mean high water (MHW; Figure 25). At least one horizon of rounded cobbles, very possibly of eroded beachrock origin, occurs above the *C. conicus* zone. The upper part of the Clifton Pier section consists of more trough and planar (tabular) cross-bedded strata that are overlain by a beach

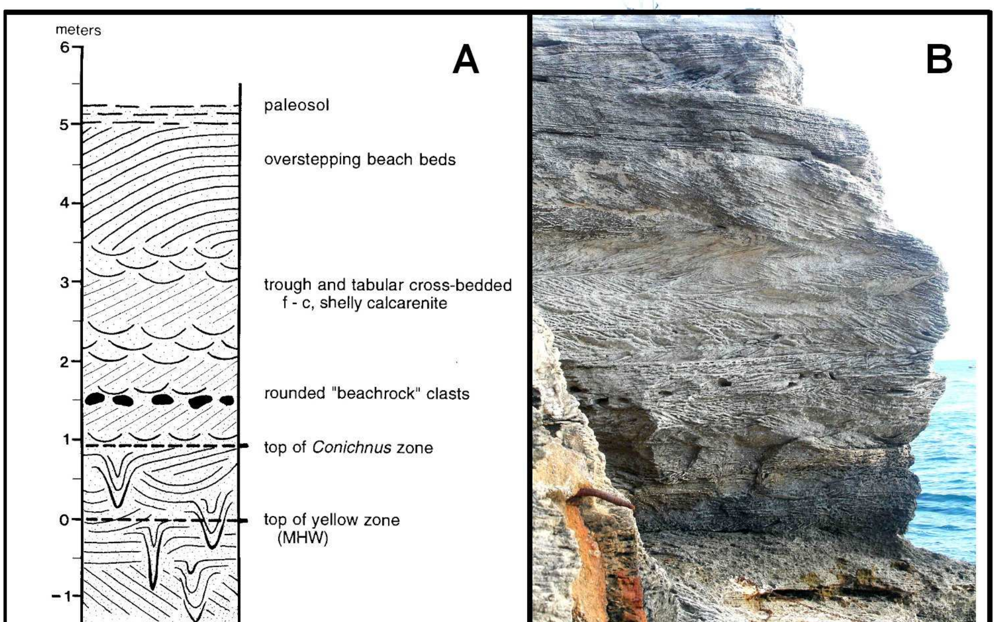

Figure 25. *Clifton Pier. A) Stratigraphic column for the Clifton Pier section constructed by Curran and White (1996). B) Recent view of this section, located immediately adjacent to the east side of the concrete dock.*

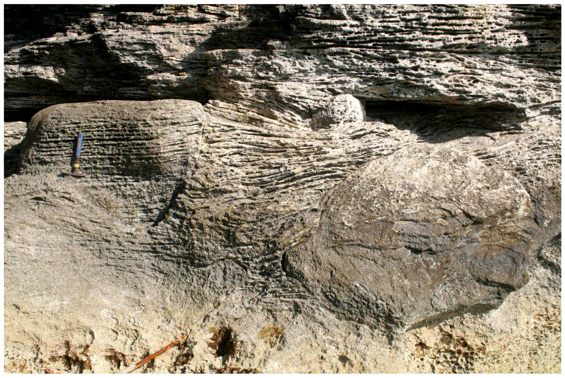

Figure 26. *Clifton Pier. Boulders are common within the lowermost strata exposed laterally from the dock area. This photo was taken near the rock steps within the Park. Note the large boulder composed of well-laminated grainstone in the right foreground. Hammer for scale = 26 cm; it may also be resting on a boulder.*

Figure 27. *Clifton Pier. Close-up view of the trough and planar (tabular) cross-bedded, grainstone strata characteristic of the Clifton cliffs. Hammer = 26 cm.*

sequence. A good lateral-view sketch of the Clifton Pier section can be found in Aurell et al. (1995, Fig. 5).

Also prominent in the Clifton Pier cliffs are numerous deep, subvertical fractures that run subparallel to the shoreline and extend downward, below sea level. These fractures intersect with bedding planes to form large rectangular blocks that may be spalled off during storms (Sealey 1990). Formation of the fractures was very possibly related to platform margin failure as envisioned by Carew et al. (1998). The platform margin is proximal to the island at Clifton Pier, making this one of the better sites in the Bahamas to view margin-failure features on the land. The fissures are roofed over in many areas, producing caves of sometimes appreciable extent. The fissures often narrow down and become infilled with micritic *terra rossa* paleosol to form caliche dikes. Recent development in this area has obscured many of these features. There has been discussion (Daugherty, et al., 1987; Carew and Mylroie, 1989a; Mylroie et al., 1991) that such fractures in The Bahamas are related to failure of the platform margin. Such platform margin failure has been reported on the large scale in The Bahamas (Mullins and Hine, 1989). Clifton Pier is the one place on New Providence Island where the platform margin is proximal to the island landmass, making such margin-failure features visible on land.

The Clifton Pier sequence is capped by a thin *terra rossa* paleosol, indicating a pre-Holocene age. However, the rocks at Clifton Pier have not been dated by either U/Th or amino-acid racemization methods. A small *Siderastrea radians* coral colony (*in situ* ?) from the beds of this sequence at nearby Clifton Point yielded a U/Th age of 146$\pm$9 ka (Neumann and Moore, 1975), confirming that these rocks are late Pleistocene in age. U/Th dating methods have improved greatly in recent years, as reviewed on pages 8 herein. Using the methods of Thompson et al. (2011), it is very likely that if this coral passed today's high aragonite purity standard and was dated, it would yield a significantly younger age. Based on stratigraphic position, Aurell et al. (1995), Reid (2010), and earlier workers placed the Clifton Pier sequence within MIS 5e, the Last Interglacial. As such, these rocks can be assigned to the Grotto Beach Formation of Carew and Mylroie (1995b, 1997). Furthermore, their largely subtidal to beach character supports assignment to the Cockburn Town Member.

The stratigraphic section of Figure 25 represents a classic Bahamian shallowing-upward facies sequence, including high-energy shoreface, lower foreshore, foreshore (beach) and backshore deposits, with possibly some washover deposits in the uppermost interval. Aurell et al. (1995) interpreted the foreshore facies as having a progressive rise across the outcrop and suggesting that sedimentation occurred during a phase of rising sea level. Curran and White (1996) noted overstepping beach beds at the top of the sequence which they interpreted as marking an initial stage of MIS 5e regression, or a progradation. Based on the elevation of the beach facies, relative sea level during deposition of the Clifton Pier strata peaked at least 5 to 6 m above present. This matches the elevation of other similar Upper Pleistocene sequences elsewhere in the Bahamas, suggesting that New Providence Island is tectonically stable.

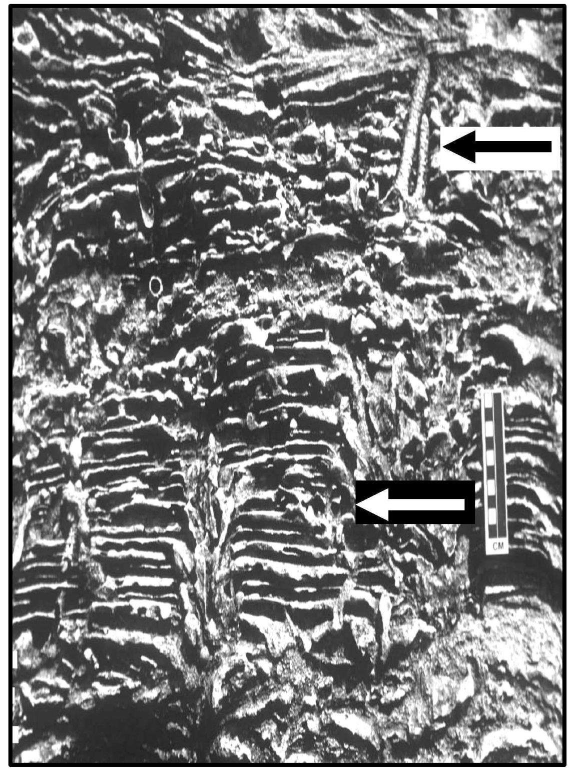

Figure 28. *Clifton Pier. Subtidal beds with an ichno-assemblage dominated by Conichnus conicus (white arrow) and Ophiomorpha (black arrow). Open circles are cuts through Ophiomorpha tunnels or shafts. Degree of bioturbation present here is higher than for most of the Clifton Pier sequence.*

The shallow subtidal facies at Clifton Pier contains an ichno-assemblage dominated by *Conichnus conicus* and *Ophiomorpha* **(Figure 28**), with rare *Skolithos linearis*, and horizonal, gently meandering burrows. This ichno-assemblage occurs elsewhere in the Bahamas and beyond, and its depositional significance was reviewed in some length by Curran (2007, p. 236-239). *C. conicus* is a large, conical, vertical, presumed animal burrow that exhibits nested, cone-in-cone laminations in longitudinal view and is circular to oval-shaped in horizontal cross-section. The structures commonly exhibit a well-lithified exterior rim with the top of the conical form filled by coarse, poorly laminated sediment (Figure 29). In this planar (tabular) and trough cross-bedded sequence, *C. conicus* specimens are abundant and robust, averaging 9 to 10 cm in diameter (Figure 30) and reaching 1.2 m in length. Most burrow lengths measured at Clifton Pier represent minimums because the full length of any specimen only rarely is revealed in outcrop.

Conichnus previously has been interpreted as the dwelling burrow or resting trace of an

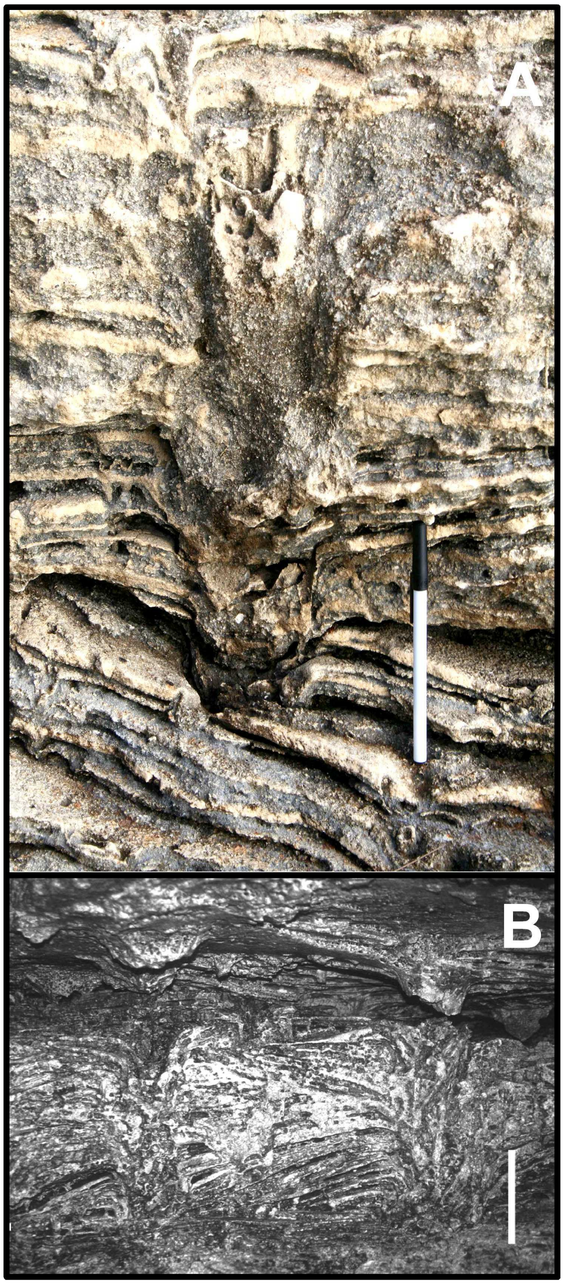

Figure 29. *Clifton Pier. A) Well developed Conichnus conicus specimen. Open circle just below the base of the pen is an Ophiomorpha shaft. Pen = 15 cm. B) C. conicus specimens exposed in a small sea cave just above mean high water level. Note the tops of several C. conicus specimens (lithified infill sediment) preserved in the roof of the cave. Scale bar = 12 cm.*

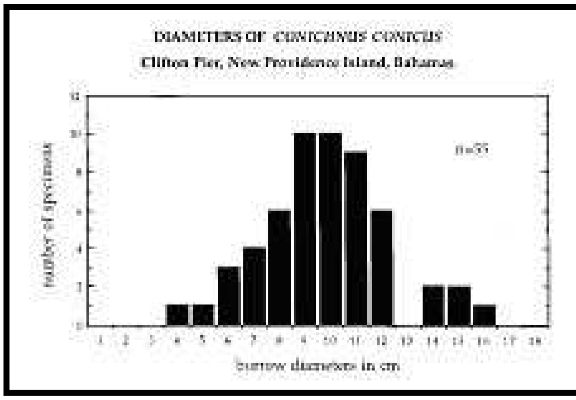

Figure 30. *Clifton Pier. Histogram of the distribution of burrow diameters for 55 specimens of Conichnus conicus measured near the Clifton Pier section.*

anemone-like organism (Pemberton et al., 1988), and Shinn (1968) made a convincing case for origin by the escape upward burrowing of a sea anemone by observing burrowing activity of the modern anemone *Phyllactis conguilegia* in oölitic sands of the Bahamas Banks. Specimens were collected and aquarium experiments conducted to record the burrowing activities of this species. With repeated applications of layers of sediment, *P. conguilegia* specimens produced biogenic structures closely comparable in form to those found at Clifton Pier Based on this analogy, the nested, cone-in-cone *C. conicus* structures at Clifton Pier can be interpreted as representing the dwelling burrows and escape structures of sea anemones attempting to keep pace with pulses of rapid sedimentation under shoaling sediment conditions. In this scenario, *Ophiomorpha* and *Skolithos linearis* were subordinate in occurrence to *Conichnus conicus* owing to active shoaling. However, caution is warranted with regard to the anemone burrow interpretation. In a study titled '*Conical structures, trace fossils or not?,*' Buck and Goldring (2003) showed that some similar conical structures can have a mechanical origin.

Interpretation of *C. conicus* as an anemone escape burrow remains viable, but additional field observations and experiments with live anemones definitely are needed for better understanding of the significance of these structures.

Ophiomorpha also is present in the subtidal beds at Clifton Pier (Figure 29), but its occurrence is clearly subordinate to *Conichnus conicus*. *Ophiomorpha* specimens here are morphologically very similar to the most common forms of *Ophiomorpha* previously described by Curran (2007, and earlier references cited therein) from other areas in the Bahamas, except that the Clifton Pier forms tend to be less robust (outside burrow diameters of 1.5 cm or less as opposed to 2 to 3 cm diameter average elsewhere). Figure 31 shows a well-developed *Ophiomorpha* burrow system with multiple branches and shaft and tunnel segments that are well-lined and have the pelleted exterior and smooth interior surfaces characteristic of this ichnotaxon. The planar (tabular) and trough cross-bedded sediments that hosted *Ophiomorpha* at Clifton Pier were deposited in an energetic environment, and the relatively mobile, shoaling bottom conditions likely suppressed dense burrowing by callianassid shrimp, the trace-maker organism for *Ophiomorpha*. Ball (1967) originally interpreted the Clifton Pier sequence as representing deposition by a southwesterly accreting spit. This general interpretation was supported by the Aurell et al. (1995) study of cores from the Clifton Pier area. Their summary diagram (Aurell et al., Fig. 8) shows an open-water, burrowed peloidal-skeletal facies of a subtidal sand flat on a semi-protected bank overlain by the shallowing-upward sequence observed in the Clifton Pier outcrops.

Figure 31. *Clifton Pier. Complex Ophiomorpha burrow exhibiting multiple shaft and tunnel branches in well-laminated grainstone. Outside diameter of shafts = 1.5 cm.*

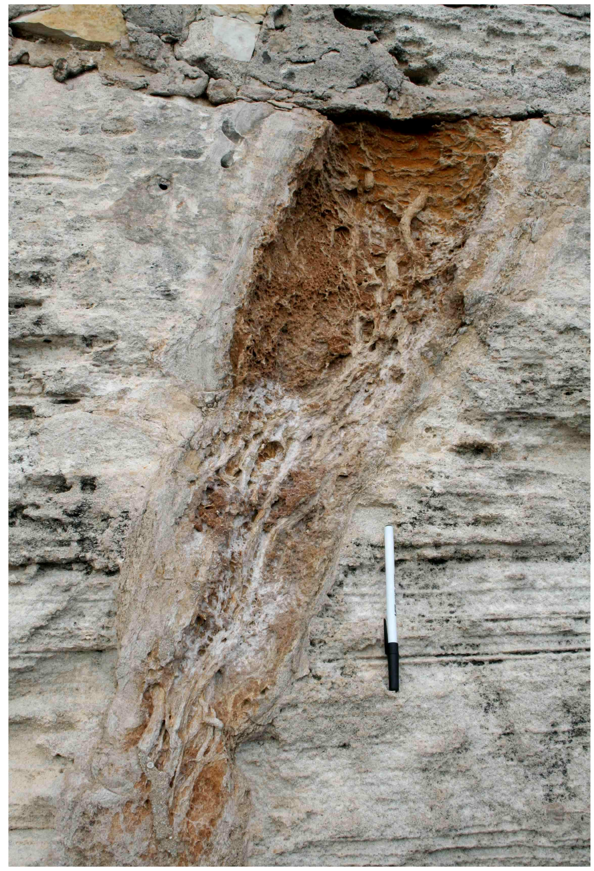

Figure 32. *Clifton Pier. Obliquely vertical dissolution tube filled with terra rossa paleosol material and rhizomorphs. This structure is at the top of the Clifton Pier section, and the capping material is concrete. Pen = 15 cm.*

Figure 33. *Clifton Pier. A large and complex, infilled dissolutional cave near the top of the Clifton Cliffs section. Hammer = 26 cm.*

A variety of interesting and complex dissolutional karst features also are found cutting into the upper-most Pleistocene beds of the Clifton Pier sequence and exposed on the vertical faces of the cliffs. Figure 32 illustrates an obliquely trending, dissolutional tube filled with *terra rossa* paleosol material and well-formed rhizomorphs (vegemorphs, rhizoliths), and Figure 33 shows a dissolutional cave that was sequentially infilled with paleosol sediment, lithified, and subsequently exposed by lateral cliff erosion or human excavation during development of the area. The lateral development of the void is characteristic of banana hole genesis, but a large pit cave could display this pattern, if the cut was off-line from the central axis of the pit.

Re-board the bus for travel to our final field trip stop.

Stop Five: Primeval Forest

The Primeval Forest, a new Bahamian national park, is located on the west side of New Providence Island, Bahamas (Figure 2). The small site, 3.5 hectares (8.6 acres), contains a rare high-canopy forest that has escaped development. The park was named for the presence of old forest and plant life. The karst features of the park made the area impossible to use for agriculture, which saved the mature plant community. The relatively unspoiled nature of the site was instrumental in motivating the creation of the park by the Bahamas National Trust (BNT) to preserve the landscape and its biological contents. The park trends from highest elevation in the northwest, on the side of an eolian ridge, down to a low plateau or bench to the southeast, which in turn becomes a flat lowland. The bench portion of the park is riddled with numerous pits, caves, natural bridges, sinkholes, and banana holes. The park is undergoing preparation for receiving large numbers of visitors, with various signs and diagrams for interpretation of the park's features. We will traverse through the park and enter one notable karst feature, to view both the dissolutional morphology, and the depositional facies present.

Karst survey work for the BNT was done in the spring of 2006 (Mylroie et al., 2008b). A total of 122 surface stations were set to catalogue 102 prominent karst features, ~29 per hectare (Figure 34). An unexpected outcome of the survey was the degree of connectivity of karst features. These features occurred at a variety of elevations as phreatic tubes, but followed the strike (NE-SW) of a low bench on the side of the eolianite ridge. The bench drops to the SE into a series of low karst valleys that are the result of the coalescing collapse of pits, caves and banana holes. The final unexpected discovery was the herring-bone cross-bedding, and the trace fossils *Ophiomorpha* sp. and *Conichnus conicus* in pit and cave walls (Figures 35 and 36). Based on current understanding, these subtidal indicators identify the rocks as the Cockburn Town Member of the Grotto Beach Formation, late Pleistocene age. The phreatic passages seen in the caves are

Figure 34. *Surface survey of the Primeval Forest karst features.*

Figure 35. *Outcrop in a cave showing herring-bone cross-bedding and perhaps Conichnus conicus. Pencil 15 cm for scale.*

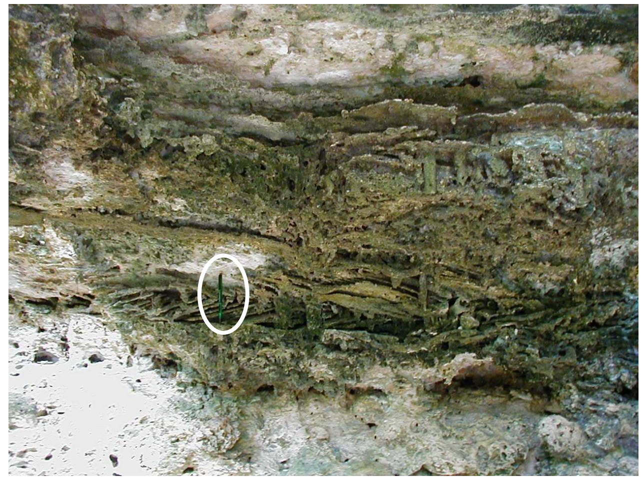

Figure 36. *Outcrop in a cave showing Ophiomorpha, Conichnus conicus, and herring-bone cross-bedding. Pencil in white oval 15 cm long for scale.*

syndepositional, as both the rock and the fresh-water lens position were produced by the last interglacial sea-level highstand, MIS 5e, ~125 ka.

The "A" survey--a cave system that was originally thought to be two small, interconnected banana holes, turned out to be a complex cave. This survey was initiated in the first natural bridge cave on the main trail from the parking area into the Primeval Forest. What was originally thought to be two interconnected banana holes ended up being part of a single cave system that continued for 125 cave survey stations before all components were finally linked to the survey (Figures 37 and 38). Connectivity of this level is highly unusual and has not been described anywhere else in the Bahamas.

The caves and karst features of the Primeval Forest are similar to what is found throughout the Bahamas, and the high density of the features, their interconnectivity and complexity are unique. There are three main features present: banana holes, pit caves, and sinkholes. The sinkholes in the Primeval Forest have formed mostly by cave collapse. Pit caves are vertical pathways dissolved out of the rock by water descending from the surface into the subsurface (Figure 39). In the Bahamas, they are commonly 2 to 10 m deep, as seen here. Banana holes are the surface expression of voids dissolved beneath the water table and later expressed at the surface by collapse. In that regard, they are a special case of sinkhole, in that their vertical bedrock walls and overhanging ledges reveal their origin as an underground void. Banana holes are considered phreatic in origin, that is, they developed below the water table, and as explained earlier, are now recognized to be immature flank margin caves. Pit caves are considered vadose in origin as they develop above the water table. The phreatic voids of the banana holes have been revealed by roof collapse, and are dry today because they formed approximately 120,000 years ago, when sea level and the water table were higher.

The history of the formation of the rocks and karst features at the Primeval Forest is an example of compression of geologic events,

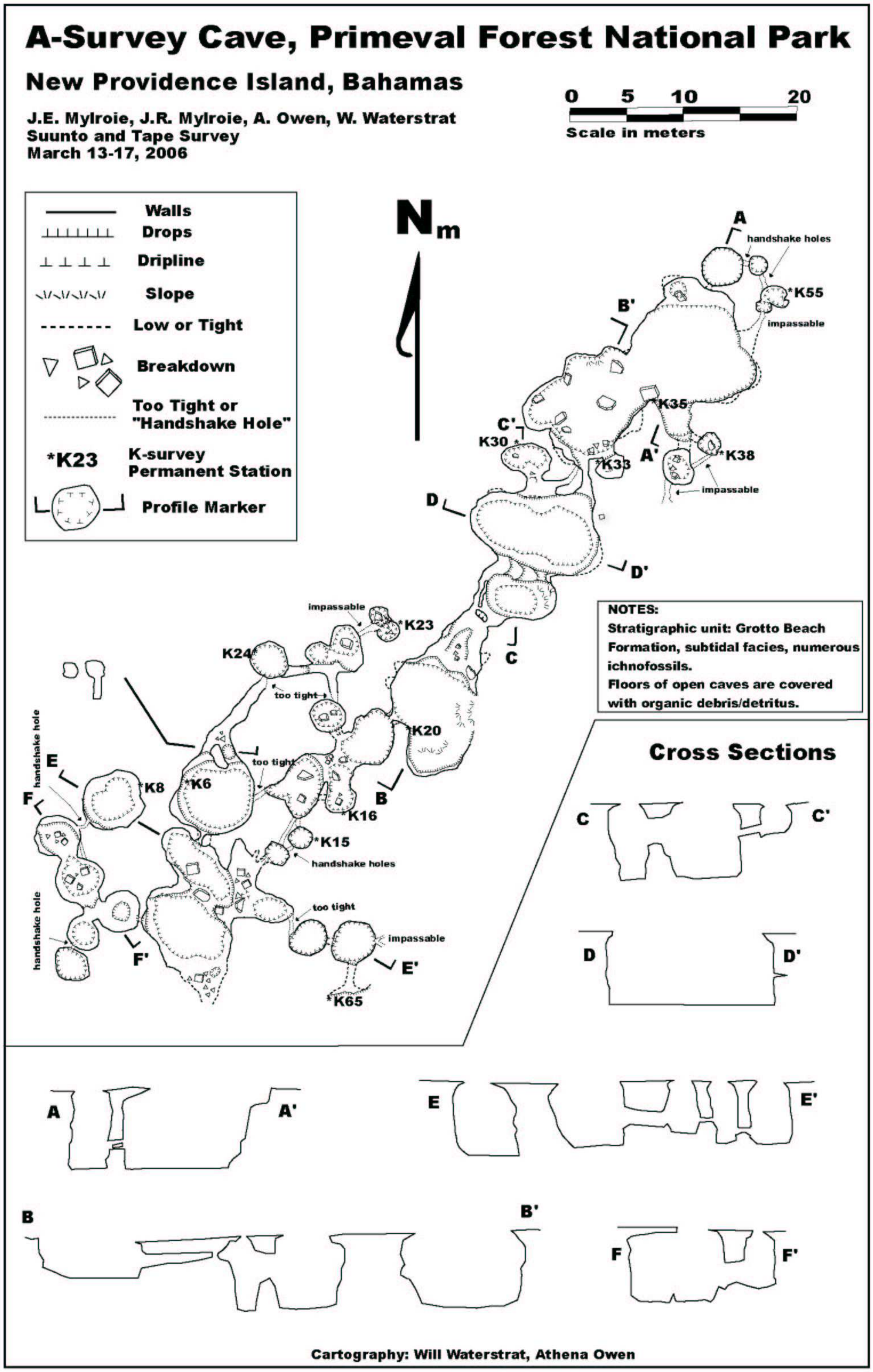

Figure 37. *Map of A-Survey Cave, which proved to be unexpectedly complex.*

Figure 38. *Surveying a tight connection in A-Survey Cave.*

Figure 40. *Trucnated cave fragment in an open karst valley.*

wherein the karst features formed at or near the same time as the rock formations, instead of long afterward. The rock units were deposited when sea level was about 6 m higher than present, about 120,000 years ago. A broad shallow lagoon existed at this site, with waves creating the oscillation ripple marks that resulted in herringbone cross bedding. Organisms living in this shallow lagoon left the trace fossils seen in Figures 35 and 36. The lagoon was gradually filled in by prograding beach sands, and subsequently by eolianites,

Figure 39. *A pit cave in the Primeval Forest, one of many that dot the hillside.*

that allowed the water table, or fresh-water lens, to sequentially invade the entire rock sequence (Figure 41). This nearly simultaneous deposition and subsequent invasion by a fresh-water lens that resulted in formation of small flank margin caves is called *syndepositional* cave development (Mylroie and Mylroie, 2009). A study of banana holes on San Salvador Island has just been completed (Infante et al., 2011), which has confirmed this model for banana hole formation as a migrating flank margin cave forming environment.

The landscape underwent a progressive erosional evolution. When sea level fell after MIS 5e, the banana holes began to undergo collapse, vadose pit caves formed in the newly created rock. The land surface overall was progressively degraded by dissolutional denudation. As a result, broad open areas and

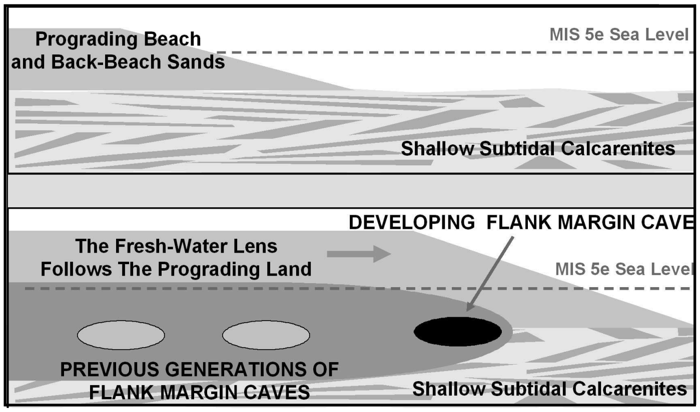

Figure 41. *Beach and back-beach dune sands prograding into a shallow lagoon create conditions for a fresh-water lens to form. Within that lens, small flak margin caves (banana holes) form, which become abandoned as progradation continues and the lens margin advances.*

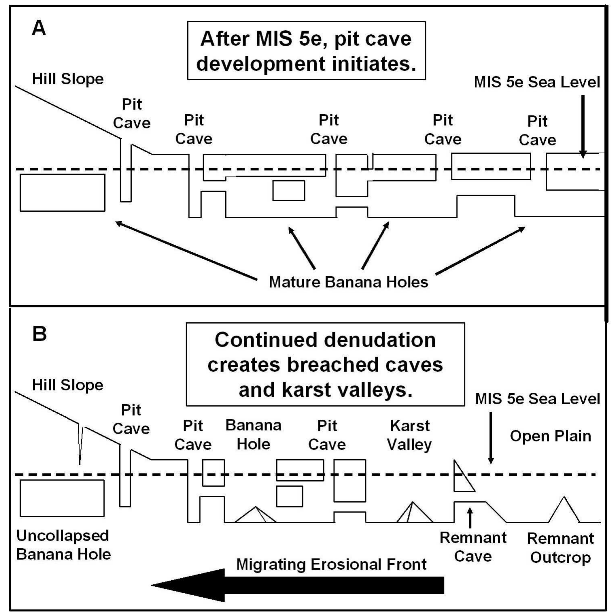

Figure 42. *After sea-level fall at the end of MIS 5e, surficial karst processes and karst denudation overprint the caves, and an erosional progression marched upslope.*

depressions formed, making large sinkholes, called karst valleys (Figures 34, 40 and 42) The karst valleys trend northeast to southwest. There has been over 100,000 years since the banana hole formation began, and erosion over that time has allowed an erosional continuum to form, as seen in Figure 42. Few cave or karst features remain to the southeast, where the land is a low, undulating plain. The karst valleys with remnant caves are found toward the northwest. Farther to the northwest, there are intact and interconnected pit caves and banana holes. A few isolated pit caves are found still farther northwest and upslope in the eolianites. Banana hole voids, if they exist, have not yet collapsed in this area. About two kilometers south of this park, at Clifton (Stop Four), the cliffs show the same sequence of herringbone cross bedding and trace fossils seen at the Primeval Forest, but they occur over a greater vertical extent, indicating the presence of deeper lagoonal water in that location.

ACKNOWLEDGMENTS

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Calcarenite Protosol at Lyford Cay, New Providence Island, Bahamas. Note the white, unstructured nature of the paleosol, the crab chelae above the hammer, and the numerous *Cerion*, with a fine example to the right of the hammer.

Land use issues in karst: collapse risk and pollution, New Providence Island, Bahamas

Inner chamber of Caves Point East Cave, showing the tangential intersection of the inner chamber with the outer chamber, typical of flank margin cave development.