GEOLOGY OF THE BAHAMAS

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INTRODUCTION

Geographical background

The Commonwealth of The Bahamas comprises the majority of an extensive archipelago of carbonate islands and shallow banks in the western North Atlantic Ocean (21° to 27°30' N and 69° to 80°30' W) (Fig. 3A-1). The southeastern portion of the same archipelago consists of the Turks and Caicos Islands (British West Indies), and the submerged Mouchoir, Silver, and Navidad banks. This chapter concerns the Bahamas, but the geology of the Turks and Caicos is similar (Wanless et al., 1989). The Bahamian archipelago covers 300,000 km², of which 136,000 km² is shallow bank, and 11,400 km² is subaerial land (Meyerhoff and Hatten, 1974). The banks are generally less than 10 m deep and are bounded by near-vertical declivities into very deep water. The Bahamas consists of 29 land masses referred to as islands, 661 cays (pronounced “keys”, generally minor islands), and 2,387 rocks (Albury, 1975). [The term “island” is used in this chapter to refer to both formal islands and cays.] The islands are predominantly low lying, and the topography is dominated by colianite (dune) ridges that extend up to 30 m on most, but not all, major islands. The highest elevation (63 m) occurs on Cat Island (Fig. 3A-1).

In the northwest, the archipelago consists of scattered islands on two large banks, Great Bahama Bank and Little Bahama Bank. Great Bahama Bank is embayed by two deep troughs: Tongue of the Ocean (TOTO) in the center (1400–2000 m), and Exuma Sound to the east (1700–2000 m). Little Bahama Bank is separated from Great Bahama Bank by Northwest and Northeast Providence Channels. To the southeast, the islands are on small banks that are separated by deep water (2000 m to > 4800 m). In many cases, the islands that occupy these banks encompass most of the bank area (e.g., Great Inagua Island, Figure 3A-1). In the Bahamas, only Cay Sal Bank (Fig. 3A-1) lacks significant islands.

Climatically, the Bahamas range from subtropical temperate in the north to semi-arid in the south. For example, on Grand Bahama Island the average temperature is 18°C (January) to 28°C (July), and average annual rainfall is 1355 mm, whereas on Great Inagua Island average temperature is 23.5°C (January) to 28.5°C (July), and annual rainfall averages only 700 mm (Sealey, 1990). Historical accounts indicate that Bahamian islands were once heavily vegetated with mixed tropical broadleaf coppice including mahogany. Today, the northern islands are largely covered by pine barrens with palmetto, but there are regions of limited broadleaf coppice. South of New Providence Island, the coppice becomes less dense, and tree
size declines as the climate becomes drier; on many islands, xeric vegetation and scrub dominate (Sealey, 1990).

The Bahamas lie within the zone of the northeast trade winds, and that has resulted in the preferential occurrence of islands on the eastern (windward) side of most banks. Trade winds have influenced the position and shape of many of the topographic ridges, but some eolianite ridges are aligned with other wind directions, especially that of the seasonal westerlies associated with fronts from the North American continent. Some islands have high ridges largely limited to the windward side (e.g., Andros Island), but most do not show such pronounced asymmetry.

Marine waters of the Bahamas average 18°C (winter) to 28°C (summer). The north equatorial current (Antilles Current) delivers water to the banks from the southeast. The current diverges and flows northwestward along the eastern margin of the archipelago at 0.6–0.8 kn, and northwestward through Old Bahama Channel south of Great Bahama Bank at 0.9 kn. The Bahamas are bounded on the west by

Fig. 3A-1. Map of the Bahamas and adjacent region showing the bank margins and most of the locations mentioned in the text. Locations not labeled on the map include: Conception Island located northwest of Rum Cay; Lee Stocking Island in the Exuma chain slightly north of Great Exuma; Joulter Cays just north of North Andros Island; Little San Salvador Island between the north end of Cat Island and the south end of Eleuthera; Schooner Cays slightly north of the northwestern projection at the south end of Eleuthera; and West Plana Cay between Mayaguana and Acklins islands.
the Florida Strait and Gulf Stream which flows at $\sim 2.5 - 2.8$ kn. (Data from the Hydrographic Chart of The Commonwealth of The Bahamas, first edition, 1977.) The Bahamas are a wholly carbonate province because these currents effectively isolate them from terrigenous sediment from the Greater Antilles and North America (Fig. 3A-I).

**Historical background**

The Bahamas are known as the site of Christopher Columbus' first landfall in the "New World", in 1492. Today, it is generally agreed that San Salvador Island (formerly known as Watling's Island, and called Guanahani by the native Lucayans) was most likely that landfall. According to Columbus' log, when he sailed from the island of his first landfall, many islands could be seen to the southwest. Interestingly, one can see many "islands" from San Salvador when atmospheric conditions are favorable. These "islands" are in fact the refracted images of hills on Rum Cay and Conception Island that lie 35 km and 54 km respectively to the southwest of San Salvador (Fig. 3A-I) (see Carew et al., 1995).

Following Columbus' voyage, exploitation and disease brought by the Spanish resulted in the rapid extinction of the native Lucayan and Arawak peoples, probably within just 25 years. The Bahama islands remained largely uninhabited for the following century and a half, until British adventurers began sparse settlement of the area in the mid-1600s. Much piracy occurred in the Bahamas, and that provoked Spanish raids in the area until the early 1700s, when the British began to exert some control on the governance of the archipelago.

When the American colonies won their independence, some British loyalists from the southeastern United States chose to leave and settle in the British-held Bahamas. Because the size of the land grant from the Crown depended on family size, including slaves, plantation owners moved their families and slaves to the Bahamas with the intention of re-establishing their plantations there. During this time of British influence, additional African slaves were brought in to work the plantations. Unfortunately, the soils of the Bahamas could not support long-term production of cotton or other large-scale farming, and the plantations soon began to fail. The Bahamas languished under British inattention, and most of the plantation owners ultimately left the Bahamas. The former slaves, freed by British government decree in 1834, were left behind, and the current population is composed largely of their descendants.

In the late eighteenth and early nineteenth centuries, the economy of these islands was based on agriculture, privateering, and wrecking. Following that, the Bahamas went through modest boom times and intervening relatively hard times. As examples, significant boom times resulted from gun-running to the Confederacy during the U. S. Civil War, and rum-running during Prohibition. Tourism began to flourish when wealthy Americans vacationed in the Bahamas during Prohibition, and it has grown into the dominant industry of the Bahamas today. In 1973, the Bahamas gained independence from Great Britain, but remained a British Commonwealth nation.
GEOLOGIC OVERVIEW

The geologic literature on the Bahamas is voluminous, and because of space limitations no attempt is being made herein to cover all of the relevant issues or references. We draw attention to many papers that contain extensive bibliographies, and we encourage any reader that wishes to become fully cognizant of the relevant literature on the Bahamas to consult those works. Recently, the Geological Society of America published Special Paper 300 (Curran and White, 1995, editors) on the geology of the Bahamas and Bermuda, and the references contained in the papers in that volume constitute an extensive bibliography. In addition, there is a large body of important work that documents the development of recent ideas concerning the geology of Bahamian islands, which is published in the Geology of the Bahamas Symposium Proceedings volumes and field trip guidebooks, as well as other publications, of the Bahamian Field Station on San Salvador Island, Bahamas.

Geologic research in the Bahamas dates back to the mid-nineteenth century (see summaries by Meyerhoff and Hatten, 1974, and Sealey, 1991). Amongst the earliest work is that of Captain Nelson who worked on Bermuda in the early 1830s, and was later assigned to the Bahamas. It was Nelson who first recognized the similarities between Bermuda (q.v., Ch. 2) and the Bahamas, and he recognized that eolian deposits dominate both island groups. It is particularly interesting that Nelson’s 1853 paper on the geology of the Bahamas was read to the Geological Society of London by none other than Sir Charles Lyell.

Other early views of the Bahamas held that they were the coral and carbonate-mantled northern portion of a mountain range that once stretched from Central America through the Greater Antilles to Florida; another interpretation suggested that the Bahamas were the result of delta-like deposition of the Gulf Stream that buried the northern extension of the eastern Caribbean mountain range mentioned above. Still other workers saw the Bahamas as entirely derived from corals, and that many of the islands were uplifted coral atolls (Sealey, 1991).

There is also lively discussion in the early literature about the apparent relative changes in sea level that can be discerned from the geological record of the Bahamas. Nelson contended that there was no evidence for either elevation or subsidence of the Bahamas. However, somewhat later views required no less than ~100 m of subsidence to account for the depths of ocean blue holes, and Shattuck and Miller (1905) called for repeated relatively rapid elevation and subsidence of the Bahamas. Field et al. (1931) appear to have been the first to make a connection between the seemingly disparate data and the changes in sea level associated with Pleistocene glaciation. In that sense, Field’s work was the start of the modern view of the geological development of the Bahamas.

Much of the post-1930s geologic research in the Bahamas has focused on tectonic evolution, modern carbonate depositional environments, and subsurface stratigraphy and bank evolution. Recently the terrestrial geology of some Bahamian islands has received considerable attention, and that is the primary focus of this chapter.
Tectonic evolution

There was much debate in the 1970s and 1980s about the early geologic evolution of the Bahamas. A major question concerned the nature of the crust that underlies the 5–10 km of predominantly shallow-water carbonates of the Bahamas (i.e., continental vs. volcanic vs. oceanic; e.g., Dietz et al., 1970; Lynts, 1970; Uchupi et al., 1971; Meyerhoff and Hatten, 1974; Mullins and Lynts, 1977; Sheridan et al., 1988; and references therein). Another question concerned the origin of the deep channels and the segmentation of the Bahamas into separate isolated banks. One school of thought held that the deep channels and banks began as grabens and horsts respectively, reflecting direct structural control (e.g., Ball, 1967a; Mullins and Lynts, 1977; Sheridan et al., 1988, and references therein). A second school of thought (e.g., Dietz et al., 1970) held that the channels and banks reflect a long-term dynamic equilibrium between normal depositional (i.e., shallow-water carbonate accumulation that keeps pace with subsidence) and erosional processes (i.e., turbidity currents that deepen and carve channels). A third school of thought held that the present channel and bank configuration evolved since the Late Cretaceous, and that formerly there was one large unsegmented bank, or “megabank” (e.g., Meyerhoff and Hatten, 1974; Schlager and Ginsburg, 1981; Sheridan et al., 1988, and references therein). Sheridan et al. (1988) summarize the diverse results of previous geologic and geophysical research concerning the tectonic evolution of the Bahamas, and they present a revised geologic history for this area, a brief synopsis of which follows.

The crust underlying the carbonates of the Bahamas was a product of the processes associated with rifting of Pangea and the opening of the North Atlantic basin in the late Middle Jurassic. The basement rocks in the northwestern Bahamas, under the Florida Straits, the Northwest Providence Channel, and the northernmost Tongue of the Ocean (TOTO) is “intermediate” or “transitional” rift crust, composed of tilted fault blocks of Jurassic volcaniclastics. Southeast of that region, the Bahamas are underlain by oceanic crust. The nature of the crust in the transition area between the Bahamas and Cuba remains poorly defined.

Development of the thick carbonate banks began in the Late Jurassic, and those carbonates formed a “megabank” that included the west Florida shelf, the Florida Platform, the Bahama Platform, and the Blake Plateau (Emery and Uchupi, 1972; Meyerhoff and Hatten, 1974). Although there is some evidence that deep-water reentrants penetrated the “megabank” in the Early Cretaceous, in most places that have been studied, the present channels and basins appear to be underlain by Lower Cretaceous shallow-water carbonates, which implies the absence of these deep areas at that time. The current deep-water channels and basins of the Bahamas appear to have existed in approximately their present positions since at least the Late Cretaceous (see Sheridan et al., 1988, and references therein).

Post-Cretaceous faulting that resulted in > 500 m displacement and tilting of blocks on the otherwise passive Atlantic margin has been attributed to interaction between the Caribbean and North American plates during the Late Cretaceous/Tertiary Cuban and Antillan orogenies. The orientations of the margins of the Bahama Banks are consistent with left-lateral wrench faulting caused by the oblique
Subduction of the North American plate under the Caribbean plate near Cuba (Sheridan et al., 1988, and references therein).

Subsurface stratigraphy

The Tertiary history of the Bahama Banks is dominated by intervals of aggradation and progradation in response to sea-level change and variations in banktop sediment production (e.g., Eberli and Ginsburg, 1987; Wilber et al., 1990; Hine et al., 1981a; Wilson and Roberts, 1992; Milliman et al., 1993). The Tertiary evolution of the Bahamas is discussed in greater detail by Melim and Masaferro in Chapter 3C. A brief discussion follows.

The subsurface stratigraphy of the Bahamas has been studied using seismic refraction, seismic reflection, magnetics, and gravity (see review by Sheridan et al., 1988); more recently, the geology and geophysics of Great Bahama Bank has been the subject of intensive seismic investigation (e.g., Eberli and Ginsburg, 1987, 1989).

In addition, the subsurface stratigraphy of the Bahamas has been studied via deep and shallow drilling. Prior to the recent University of Miami Bahamas Drilling Project, some results of which are summarized by Melim and Masaferro in Chapter 3C, the lithology of the deep subsurface of the Bahamas was known from four deep wells drilled on Andros Island, Cay Sal, Long Island, and Great Isaac. Limestone, dolostone, and evaporites were recovered in those wells. The Cay Sal and Great Isaac wells penetrated Upper Jurassic carbonates at slightly greater than 5 km depth, and the Andros Island and Long Island wells ended in Lower Cretaceous dolostone (Meyerhoff and Hatten, 1974; Sheridan et al., 1988; and references therein).

Numerous shallow boreholes also have been drilled at a variety of locations in the Bahamas, including: Crooked Island, Mayaguana Island, Great Inagua Island, Hogsty Reef, Grand Bahama Island, Great Abaco Island; Andros Island, Eleuthera Island, San Salvador Island, and New Providence Island (e.g., Meyerhoff and Hatten, 1974; Supko, 1977; Beach and Ginsburg, 1980; Pierson and Shinn, 1985; Aurell et al., 1995). An apparently important stratigraphic conclusion reached by study of such shallow subsurface rocks was the recognition that, at the margins of Great Bahama Bank, there is a transition from Pliocene skeletal and reefal facies to Quaternary oolites and eolianites (Beach and Ginsburg, 1980). It has been suggested that this transition may be related to the onset of northern hemisphere glaciation and more frequent glacioeustatic changes (Schlager and Ginsburg, 1981).

Some shallow coring has indicated that Pleistocene-Holocene sediments are about 24 m thick on Little Bahama Bank and as much as 40 m thick on Great Bahama Bank (Beach and Ginsburg, 1980). It has been suggested that such data may reflect differential subsidence among the individual banks of the Bahamas (Schlager and Ginsburg, 1981), and Sheridan et al. (1988) argue that it is plausible that differential subsidence has continued into the Holocene; however, recent study of exposed coral reefs and flank margin caves in the Bahamas indicates that the entire archipelago appears to have behaved similarly (no more than 1–2 m subsidence per 100 ky) for at least the last 300 ky (Carew and Mylroie, 1995a,b). Also, the thickness of the
Quaternary sediment package does not vary systematically across the Bahamas (e.g., Cant and Weech, 1986).

Modern depositional systems

The lithofacies of the modern Bahama banks have been used as models for the interpretation of ancient carbonates (e.g., Bathurst, 1975). Classic work on the sediments of the Bahama banks includes that of Illing (1954), Purdy (1963), Ball (1967b), Enos (1974), Gebelein (1976), Hine et al. (1981b), among many others. At the large scale, four major shallow-marine lithofacies (coralgal, ooid, grapestone, and lime mud) have been recognized in the Bahamas (see Milliman, 1974; Bathurst, 1975; Tucker and Wright, 1990; and references therein). Intertidal and supratidal lithofacies of the Bahamas have also been intensively studied. In particular, western Andros Island has provided much information on the dynamics of micritic tidal flat deposition (see Shinn et al., 1969; Bathurst, 1975; Hardie and Shinn, 1986; Tucker and Wright, 1990; and references therein). While those studies have yielded a general understanding of the large-scale facies mosaic, such as that of the Great Bahama Bank (Fig. 3A-2), the reader should be cognizant of the fact that there is much
greater variability in sediment type and facies distribution than is suggested by such
generalizations. Wide variability in accumulation, depositional style, and sediment
type on the Bahama banks results from differences in orientation to currents and
winds that influence the physical energy of various areas.

A wide variety of stromatolite development has been reported from the Bahamas.
Forms include very large (>2 m) subtidal stromatolites (Dravis, 1983; Dill et al.,
1986; Shapiro et al., 1995, and references therein), small coastal and subtidal stromatolites
(Pentecost, 1989), intertidal stromatolites (Reid and Browne, 1991), and
stromatolites in hypersaline lakes (Neumann et al., 1989). Bahamian stromatolites
generally occur where rapid currents (Dill et al., 1986; Shapiro et al., 1995) or
hypersalinity (Neumann et al., 1989) prevent grazing by macrofauna. Rapid
cementation has also been invoked as an important factor in stromatolite development
(Reid and Browne, 1991).

Surficial geology

The surficial geology of Bahamian islands has recently been studied with in-
creasing detail (e.g., Titus, 1980; Garrett and Gould, 1984; Carew and Mylroie, 1985,
ture of the surficial geology of most Bahamian islands is the occurrence of large
eolianite ridges. The original interpretation of the origin of these deposits held that
exposed banktop sediments were reworked into regressive sequences during sea-level
fall (e.g., Titus, 1980), or during stillstand and regression (Garrett and Gould, 1984).
Detailed work on San Salvador Island led to the realization that eolianite ridges
form during all phases of a sea-level highstand, and that those deposited during the
transgressive phase are often the most substantial accumulations (Carew and My-
vroie, 1985, 1995a, and references therein). The detailed discussion of this deposi-
tional model presented in Carew and Mylroie (1995a) is summarized in this chapter,
and is extensively cited as a source for additional citations to the relevant literature.
[Kindler and Hearty give an account of the constructional architecture of Bahamian
islands in Chapter 3B of this book. — Eds.]

GEOMORPHOLOGY OF BAHAMIAN ISLANDS

Landscapes

The Bahama islands exhibit a largely constructional landscape; that is, the
landforms have been created by accumulation of biogenic and authigenic carbonate
sediment deposited by currents, waves, and winds. All major islands in the Bahamas
are dominated by two landforms: eolianite ridges that commonly rise up to 30 m
above sea level (Fig. 3A-3), and lowlands composed of marine and terrestrial
deposits. Most Bahamian islands are dominated by Pleistocene rocks, with a lesser
amount of Holocene rocks, generally on island fringes. Analysis of the landforms on
San Salvador Island has shown that the island comprises 2.6% beach, 4.5% Ho-
Holocene rocks, 22% lakes and tidal creeks, 21% eolianite ridges, and 49% lowlands (Wilson et al., 1995). Because the lowlands consist primarily of intertidal and subtidal deposits including fossil reefs that have radiometric ages that indicate formation during the last interglacial (oxygen isotope substage 5e, ~125 ka), Wilson et al. (1995) referred to them as the Sangamon Terrace.

In the interior of Bahamian islands, topographic lows that extend below sea level, especially inter-dune swales, commonly contain lakes that are usually marine to hypersaline. Surface streams are absent. All land above 7 m elevation consists of eolian deposits, but land below 7 m elevation is a mixture of marine and terrestrial (incl. lacustrine) lithofacies. Pleistocene rocks are covered with a red micritic calcrete or terra rossa paleosol (Carew and Mylroie, 1991) unless it has subsequently been removed by erosion. On the other hand, Holocene rocks lack a well-developed calcrete or terra rossa paleosol, but a thin micritized crust sometimes occurs.

Although most of the landscapes in the Bahamas are largely of Pleistocene origin, a few Bahamian islands such as Joulter Cays and Schooner Cays are entirely Holocene. These Holocene islands are hardly more than exposed shoals, and they are only 100's of m long and wide, only 1.5–2.5 m high, and consist of intertidal and back-beach dune facies that are at the same elevations as sediments being currently laid down in similar depositional environments (e.g., Budd, 1988; Budd and Land, 1989; Halley and Harris, 1979; Harris, 1983; Strasser and Davaud, 1986). These Holocene deposits are up to 10.7 m thick (Budd, 1988). Cementation is vertically and laterally variable, but where it occurs, it is minimal and dominated by vadose freshwater meniscus cements, with occasional marine cements (e.g., Strasser and Davaud, 1986; Budd, 1988). The greatest degree of cementation in these islands is usually found beneath the water table (e.g., Budd, 1988), as is also true of the Holocene deposits on larger islands (e.g., McClain et al., 1992). While many of these
Holocene islands are primarily oolitic, subaerially exposed Holocene stillstand-phase deposits on Bahamian islands are usually peloidal and bioclastic.

**Karst processes**

The subsurface hydrology of the Bahamian Archipelago is complex. In Chapter 4, Whitaker and Smart describe in detail the complexities of the freshwater lens, its flow dynamics, and its chemistry in Bahamian islands, and their Case Study concerns the Bahamian blue holes. The discussion presented herein focuses on karst that is observable in the subaerial environment.

Dissolution of the carbonates of the Bahama islands has produced a karst landscape that is superimposed on the overall constructional landscape (Mylroie and Carew, 1995; Mylroie, et al., 1995a,b; and references therein). The four major categories of karst features of the Bahamas are: karren, depressions, caves, and blue holes. Karren are centimeter- to meter-scale features of dissolutional sculpturing of carbonate bedrock. Karren tends to be jagged on exposed rock surfaces, but smooth and curvilinear on soil-mantled surfaces. Small dissolution tubes carry water away from the karren. This entire zone of karren, small tubes, and soil is called the epikarst, which usually extends downward from the surface for tens of centimeters to a meter or more. A special type of karren, often called coastal phytokarst, but more properly termed biokarst (Viles, 1988), commonly occurs on coastal rocks affected by sea spray.

The large closed-contour depressions seen on Bahamian topographic maps typically are depositional lows, rather than the product of dissolution. Many extend below sea level, and they are commonly occupied by lakes of varying salinities (typically normal marine to hypersaline), depending on climate, season, lake size, and whether there are cave conduits or blue holes that connect them to the sea.

There are four common types of caves developed in Pleistocene rocks in the Bahamas: pit caves, flank margin caves, banana holes, and lake drains. Pit caves are vertical shafts that conduct water from the epikarst through the vadose zone to the water table (Mylroie and Carew, 1995; Mylroie et al., 1995b). Flank margin caves are subhorizontal voids produced in the discharging margin of a freshwater lens (Mylroie and Carew, 1995; Mylroie et al., 1995b). During the last interglacial sea-level highstand (∼125 ka), the Bahama islands consisted only of eolian ridges, each of which had its own small freshwater lens. The zone of vadose/phreatic freshwater mixing at the top of the lens, and the freshwater/marine phreatic mixing zone at the base of the lens are known to be environments where enhanced dissolution is likely to occur (James and Choquette, 1984; Mylroie and Carew, 1995; and references therein); so, at the lens margin where those two zones are superimposed, there is even greater potential for dissolution (Mylroie and Carew, 1995, and references therein). At the end of the last interglacial, these caves were abandoned as sea level and the freshwater lens fell. These caves commonly can be entered today through erosionally produced entrances along the flanks of many eolian ridges.

Banana holes are ovoid depressions found in the Sangamon Terrace terrain of the Bahamas (Harris et al., 1995; Wilson et al., 1995). They are commonly a few meters
deep and up to 10 m wide. The walls vary from sloping sides, to near vertical or overhung. Some banana holes are connected to adjacent roofed chambers. Like flank margin caves, these voids developed during the last interglacial, but they formed just beneath the surface of a shallow freshwater lens rather than at the lens margin. At the end of the last interglacial, these caves were drained. Subsequent roof collapse coupled with karren development on the exposed walls accounts for the variety of wall morphologies that are seen.

Lake drains are conduits that transmit tidally influenced water into and out of some lakes in the Bahamas (Mylroie et al., 1995b). The presence of these drains allows sufficient seawater to enter the lakes so that they maintain normal marine salinity where hypersaline conditions would otherwise develop. As these conduits are below present sea level, and are commonly too small for divers to enter, their morphology and origins are poorly understood.

Blue holes have been defined as, “...subsurface voids that are developed in carbonate banks and islands; are open to the earth’s surface; contain tidally influenced waters of fresh, marine, or mixed chemistry; extend below sea level for a majority of their depth; and may provide access to submerged cave passages” (Mylroie et al., 1995a, p. 231). Blue holes are further subdivided into ocean holes which open directly into the present marine environment, and inland blue holes that contain water of a variety of salinities (Mylroie et al., 1995a, and references therein; see also the Case Study of Chapter 4.).

Flank margin caves and banana holes are good indicators of past sea-level position because they form at the margin, or at the top, of a freshwater lens, respectively. They also developed very rapidly, in the 10–15 ky duration of the substage 5e sea-level highstand (Mylroie and Carew, 1995; Mylroie, et al., 1995b). Although the majority of the flank margin caves are developed in eolianites deposited prior to the interglacial associated with substage 5e (which formed the host islands in which these caves developed), banana holes and some flank margin caves are developed in carbonates deposited during substage 5e. These latter caves must have developed in transgressive or stillstand-phase deposits, during the regression from the acme of the last interglacial sea-level highstand (substage 5e). Flank margin caves and banana holes that are accessible today in the subaerial environment developed during the substage 5e highstand. Any flank margin caves or banana holes that formed during earlier highstands (pre-5e) are now below present sea level as a result of either a lower highstand position (relative to present) at the time of their formation, or subsequent isostatic subsidence of the Bahamas (Carew and Mylroie, 1995b).

Coastal processes

The coasts of Bahamian islands consist largely of rocky cliffs and sand beaches (Fig. 3A-4; see also 3A.12), but in some locales (such as the west coast of Andros Island) the lee sides may be flanked by tidal flats (Fig. 3A-5). Where coastal dynamics favor erosional processes, there are eroding Pleistocene and Holocene rocky cliffs, some of which have bioerosion notches (e.g., Salt Pond, Long Island)
Fig. 3A-4. Photograph of Grotto Beach on San Salvador Island illustrating the typical Bahamian island coastline consisting of rocky cliffs and sand beaches.

Throughout the Bahamas, there are numerous reentrants in the sides of Pleistocene eolianite ridges that have been considered to be fossil bioerosion notches formed during substage 5e. These reentrants are now recognized to be the eroded remnants of flank margin caves that have been largely removed by erosional retreat of the hillside that once contained them (Mylroie and Carew, 1991) (Fig. 3A-6B). The implications of this new interpretation are important because surface lowering of a few meters per 100 ky, which is in agreement with reported modern carbonate denudation rates (e.g., Ford and Williams, 1989, Tables 4-3 and 4-6), is sufficient to account for the several meters of hillside erosion necessary to reduce some flank margin caves to just the curving back wall. Such erosion would completely remove any bioerosion notches that had been on a hillside. Interpretation of these reentrants as “pristine” fossil bioerosion notches, which has been used to support a scenario that postulates extremely rapid sea-level fall at the end of the last interglacial (Neumann and Hearty, 1996), is incompatible with the interpretation that these reentrants are the eroded remnants of flank margin caves.

Tidal channels and creeks penetrate the shorelines of many islands, and there, tidal delta deposits may occur (e.g., Pigeon Creek, San Salvador Island; Deep Creek, South Andros Island). [The term “creek” in the Bahamas is derived from the British
usage, and it refers to estuaries and restricted marine embayments, not surface streams.] Progradational strandplains have developed where there has been substantial deposition during the Holocene (Fig. 3A-7) (Garrett and Gould, 1984; Strasser and Davaud, 1986; Andersen and Boardman, 1989; Mitchell et al., 1989; Wallis et al., 1991; Carney et al., 1993, and references therein). An ever-changing distribution of depositional and erosional effects on the shorelines of Bahamian islands is the result of changes in offshore features such as reefs and shoals. Both depositional and erosional coastal features in the Bahamas show evidence of changing conditions that have occurred in a short time (<10² y), and with a nearly static sea-level position.

Particularly in the southeastern Bahamas, there is little evidence of cliffing of outcrops on the margins of lakes. Unlike the hillslopes above lake level, the original constructional slopes of dune swales that are the sites of the lakes are only slightly modified by coastal processes active at lake margins. This is because of the generally arid conditions that create hypersaline lakes wherein virtually no dissolution or bioerosion occurs (Mylroie et al., 1995b, and references therein).
Fig. 3A-6. Photographs of modern and Pleistocene cliff-line notches. (A) A modern coastal biocorision notch at Salt Pond, Long Island. (B) A notch in an inland cliff, San Salvador Island. Although the notch has the appearance of a coastal biocorision notch, it is actually the remnant of the interior wall of a highly eroded flank margin cave. Note person in the center background for scale.
QUATERNARY EVOLUTION OF BAHAMIAN ISLANDS

Overview

The Quaternary evolution of the Bahamian islands has been controlled largely by glacioeustatic sea-level fluctuations that affected both the deposition and subsequent alteration of the carbonate sediments. The quantitative record of Quaternary glacioeustasy is inferred largely from the deep-sea oxygen isotope record (e.g., Shackleton and Opdyke, 1973; Imbrie et al., 1984; Chappell and Shackleton, 1986; Shackleton, 1987) which is calibrated, in part, by the raised coral-reef terraces of Barbados and New Guinea. The shallow depth of the Bahamian bank margins (generally ~10 m) and the slow subsidence of the Bahamas during the Quaternary
dictate that sea level must be within ~10 m of its present position before banktops begin to flood and the subtidal "carbonate factory" produces abundant sediment. The isotope-derived sea-level curve (Fig. 3A-8) indicates that, for much of the Quaternary, sea level was at least 10 m below present datum (0.1 per mil c in $\delta^{18}O$ is equivalent to a 10 m change in sea level, Fairbanks and Matthews, 1989). In terms of Bahamian island evolution, therefore, the Quaternary has consisted primarily of long periods ($10^5$ years) of island emergence and subaerial diagenesis punctuated by short intervals ($10^4$ years) of submergence and substantial carbonate sediment production. Some carbonate sediment produced during banktop flooding remains in the shallow-marine environment, but much is exported off the bank into deeper water (e.g., Droxler et al., 1988; Boardman and Neumann, 1986; and references therein), or is reworked into beach sediments and subaerial dunes (Carew and Mylroie, 1995a). Terra rossa paleosols and eroded (karst) surfaces are formed largely during lowstands, but they develop at all times exposed surfaces (Carew and Mylroie, 1985; 1995a).

Prior to the present sea-level highstand (stage 1), the most recent pre-Quaternary highstand occurred during the last interglacial (Sangamon interglacial; substages 5a and 5c, ~125 ka). Numerous lines of evidence indicate that at that time sea level surpassed its present elevation by up to 6 m. During that highstand, significant deposits of terrestrial (eolianites and lacustrine deposits) and marine beachface through subtidal deposits. Bahamian islands at that time consisted only of some of the existing eolianite ridges (e.g., Fig. 3A-9).

There are no known marine deposits that correlate with substages 3a or 5a. Elevation of sea-level highstands prior to the stage 5 interglacial (e.g., stages 7, 9, and 11) are less well known relative to stage 5 elevations, but the oxygen isotope record indicates that during the Pleistocene, sea level elevation was unlikely to have been higher than it was during stage 5 (Shackleton, 1987). The slow subsidence of
Bahamas and inferred sea-level history of the Quaternary suggest that pre-5e marine deposits are unlikely to be exposed on Bahamian islands today (Carew and Mylroie, 1995b). It is possible, however, for pre-stage 5 eolianites to be exposed on Bahamian islands today because eolian sediments may be deposited more than 30 m above sea level at the time of deposition. These eolianites, however, have been subject to continuous erosion since their deposition unless they are buried and protected by deposits of younger highstands (Figs. 3A-10, 3A-11). Correlation of such deposits with a particular highstand (isotope stage) is difficult because of the spatially patchy nature of their accumulation. Moreover, sequential eolian deposits are often situated
A Lowstand Phase: The carbonate platform is exposed, there is no carbonate deposition on the platform surface. Pedogenesis and karst processes dominate, leading to development of terra-rossa paleosols.

B Transgressive Phase: The platform surface is partially flooded, carbonate sediment production initiates. Transgressive eolianites are deposited.

C Stillstand Phase: Reefs grow up to sea level, and lagoons fill with carbonate sediment. The transgressive eolianites undergo attack by coastal processes.

D Regressive Phase: Sea level falls, decreasing the amount of submerged bank. Former subtidal deposits are re-mobilized, entombing other subtidal features and providing source material for regressive eolianites.
Potential onlap/overlap between eolianites deposited during separate sea-level highstands (e.g., substage 5e and earlier) and intervening terra rossa paleosols that accumulate largely when sea level is below the banktops. Note the various views of these relationships afforded by flank margin caves, pit caves, and road cuts or cliff exposures. (From Carew and Mylroie, 1995a.)

lateral to one another — not necessarily atop one another, as is the more common situation among other sediment facies.

Depositional model

Unraveling the geologic history of surficial deposits in the Bahamas is made difficult by the discontinuous nature of surficial sedimentation, the variable geometry of eolian deposits, the relatively sparse exposures, and the lack of material suitable for high-precision age determinations. Despite these difficulties, the surficial geology of Bahamian islands can be placed into a conceptual framework when the products and processes of subsidence, sea-level change, subaerial diagenesis, and carbonate sedimentation are integrated. Detailed studies of the surficial geology over many...
years permits us to delineate four stages, or phases, of island development in the Bahamas: transgressive phase, stillstand phase, regressive phase, and a lowstand phase (Fig. 3A-10) (see Carew and Mylroie, 1995a for a more thorough discussion).

Transgressive phase. In the early stages of banktop flooding by rising sea level, substantial subtidal sediment is produced, transported by waves to beaches, and then into dunes (Boardman et al., 1987). Formation of ooids and coated-grains is common during this phase (Carew and Mylroie, 1985, 1995a, and references therein; Hearty and Kindler, 1993); and ooid production must have occurred largely along the shoreface, such as reported by Lloyd et al. (1987) at the Turks and Caicos Islands and Ward and Brady (1973) along the Yucatan coastline.

Carbonate dunes do not develop far from, or migrate away from, their beach sources (Bretz, 1960; Carew and Mylroie, 1985, 1995a); so, as shoreline processes are driven inland by rising sea level, they “bulldoze” large amounts of sediment into high arcuate dune ridges that are commonly nucleated on and extend laterally (catenary) from high grounds remaining from previous highstand deposits (Carew, 1983; Garrett and Gould, 1984) (Fig. 3A-12). The beaches and dunes are composed of new allochems plus reworked allochems (particularly from eolianites) formed earlier in the same highstand (Andersen and Boardman, 1989), but it is rare to

Fig. 3A-12. (A) Aerial photograph of a catenary eolianite ridge developed between two preexisting high grounds that acted as nucleation points, San Salvador Island. The ridge, bordered by a sand beach, extends southward from the rocky headland of Crab Cay to Almgreen Cay. (B) Aerial photograph of a comma-shaped eolianite ridge that is catenary on a rocky headland (The Bluff, San Salvador Island) at the north. This ridge is the same one seen in Fig. 3A-14.
encounter clearly identifiable reworked allochems from earlier highstands (Carew and Mylroie, 1995a). Because transgressive-phase dunes lie close to the shoreline for the duration of the highstand, they are subjected to the combined effects of sea spray and meteoric precipitation that promote rapid freshwater vadose (meniscus style) cementation, with occasional traces of marine cement (e.g., Halley and Harris, 1979; Strasser and Davaud, 1986; White, 1995).

Today on numerous Bahamian islands, because of continued rise of sea level since their emplacement, transgressive-phase Holocene eolianites have been subjected to marine erosion that has formed sea cliffs up to 20 m high (some of which contain sea caves) and subaerial and subtidal wave-cut benches, some of which are now colonized by corals and other taxa (Fig. 3A-13) (Carew and Mylroie, 1995a). In some places, beach progradation seaward of these eroded Holocene eolianites has produced inland cliffs (Fig. 3A-14). Eolianite deposition and marine erosion during a single highstand can be detected by the lack of a terra rossa paleosol between the transgressive-phase eolianite and later features (e.g., corals on a wave-cut bench, boulder rubble in a sea cave, regressive-phase eolianite). Truncated eolianite bedding covered by a terra rossa paleosol or calcrete indicates either: (1)

Fig. 3A-13. Photograph showing corals growing on a wave-cut platform carved into a Holocene transgressive-phase eolianite of the North Point Member on High Cay, San Salvador Island. In the background and right is the highly eroded transgressive-phase eolianite. Circular colonies of Acropora palmata in the foreground and center are nearly 4 m in diameter.
Fig. 3A-14. Photograph showing view to the northwest of an eroded transgressive-phase Holocene eolianite ridge of the North Point Member, and talus that has accumulated at the base of the cliff line at Snow Bay, San Salvador Island. The windward half of the dune was eroded away by wave activity, and then apparent changes in coastal dynamics have led to accumulation of a sand beach seaward of the eroded eolianite ridge.

deposition and wave erosion during a single highstand, thus, a transgressive-phase eolianite (e.g., Fig. 3A-15A); or (2) deposition during one highstand, erosion on a subsequent highstand, and paleosol development during an ensuing lowstand (e.g., Fig. 3A-15B) (Carew and Mylroie, 1995a).

Holocene transgressive-phase eolianites have relatively few plant trace fossils, termed vegemorphs (Carew and Mylroie, 1995a), but they exhibit spectacular fine-scale (<1 mm) bedding such as sandflow, grainfall, and climbing wind-ripple cross laminae (e.g., White and Curran, 1988; Caputo, 1995) (Fig. 3A-16). Development of such laminae requires unobstructed windward slopes and lee slip faces, so they are not seen in the well-vegetated modern (stillstand phase) dunes in the Bahamas. Similar sedimentary architecture is also found among Pleistocene eolianites, especially those identified as transgressive phase (Caputo, 1993, 1995; and references therein). The transgressive-phase eolianites of the Bahama islands were probably sparsely vegetated because plant taxa adapted to mobile sand would have largely disappeared throughout the Bahamas during the preceding 100+ ky lowstand; hence, colonization would require recruitment from the North American mainland or Caribbean islands that do not have steep bank margins (Godfrey, pers. comm. 1994; Carew and Mylroie, 1995a). Direct analogs to these Holocene rocks and sediments can be seen in Pleistocene rocks (see Table 3A-1).
Fig. 3A-15. Diagrams illustrating the different temporal relationships that may occur where fossil reef deposits are seen to overlie a truncated eolianite. (A) Stratigraphic relationships at High Cay, South Andros Island, where corals are situated upon a wave-cut bench that was carved into the transgressive-phase French Bay Member later in the same highstand; this relationship is the same as that shown in Fig. 3A-13. (B) Stratigraphic relationships at Grotto Beach, San Salvador Island, where reefal deposits of the Grotto Beach Formation (substage 5e) are on an erosion surface developed on an eolianite of the pre-5e Owl’s Hole Formation. (From Carew and Mylroie, 1995a.)

Stillstand phase. Using modern geological conditions as a guide, the scenario for the stillstand phase is as follows. During the acme of the highstand, when sea level remains relatively stable (e.g., the last 2–3 ky of the Holocene), carbonate sediment production remains high, reef growth catches up, and lagoons fill because of the quieter conditions behind reefs and transgressive-phase eolianite ridges (Carew and Mylroic, 1995a, and references therein). Much of the marine record on the islands is probably deposited during the stillstand phase. This is probably also a time of significant off-bank transport of bank-derived sediment, particularly early in the stillstand, before reefs become barriers to off-bank transport, and lagoons are filling. On the islands, strandplains and beaches develop, prograde into the subtidal, and
Fig. 3A-16. (A) Photograph of well-preserved fine-scale laminae in a Holocene transgressive-phase eolianite (North Point Member of the Rice Bay Formation, on Long Island). (B) Close-up view of the fine-scale laminae.

entomb subtidal deposits. Many stillstand-phase progradational deposits may be indistinguishable from regressive-phase deposits. Today in the Bahamas, shoreline deposits composed of lithified Holocene calcarenite blocks entombed in penecontemporaneous sand are common (Fig. 3A-17A). These facies indicate shoreline
progradation and lithification followed by erosion to generate the blocks, and subsequent progradation that entombs the blocks. Similar deposits also occur in Pleistocene rocks (Fig. 3A-17B) (Carew and Mylroie, 1995a).

Table 3A-1

Characteristics associated with the transgressive (T), stillstand (S), and regressive (R) phases of the Quaternary depositional cycle

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>T</th>
<th>S</th>
<th>R</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eolian bedding preservation</td>
<td>fine scale</td>
<td>partially to highly disrupted</td>
<td>highly disrupted (esp. upper part)</td>
</tr>
<tr>
<td>Vegemorphs</td>
<td>few</td>
<td>abundant</td>
<td>extensive</td>
</tr>
<tr>
<td>Sea caves</td>
<td>penecontemporary</td>
<td>rare</td>
<td>none penecontemporary</td>
</tr>
<tr>
<td>Cliffling and boulder talus deposits</td>
<td>penecontemporary in beach and eolian facies</td>
<td>penecontemporary in back-beach to intertidal</td>
<td>penecontemporary cliffling</td>
</tr>
<tr>
<td>Protosols</td>
<td>uncommon</td>
<td>common</td>
<td>common</td>
</tr>
<tr>
<td>Corals</td>
<td>on penecontemporary wave-eroded benches</td>
<td>not found on penecontemporary benches</td>
<td>no penecontemporary benches</td>
</tr>
<tr>
<td>Facies relationships</td>
<td>eolian facies dominant, onlapped by S and R deposits</td>
<td>marine facies abundant, shallowing-upward sequences</td>
<td>eolian facies dominant, often overstepping marine facies</td>
</tr>
<tr>
<td>Environments represented in exposed rocks</td>
<td>predominantly eolian, occasional beach facies</td>
<td>eolian, marine, strandplain, lacustrine, tidal deltas</td>
<td>predominantly eolian</td>
</tr>
</tbody>
</table>
During the stillstand phase, heavily vegetated coastal dunes develop (as they have in the Holocene), and protosols accumulate on transgressive-phase eolianites and in other locales (Carew and Mylroie, 1995a). Flood- and ebb-tidal delta deposits develop at passes between islands, and prograde at the mouths of some tidal creeks (e.g., Pigeon Creek, San Salvador Island). Inter-dune swales may contain lakes with ostracod and molluscan assemblages (Hagey and Mylroie, 1995; Noble et al., 1995; Teeter and Quick, 1990).

Regressive phase. Although we have no modern analog for the regressive phase, the following scenario can be inferred from the Pleistocene record. When sea level falls in response to renewed continental glaciation, beaches and their associated facies retreat toward the bank margin, and regressive-phase beach and dune deposits bury portions of the stillstand-phase marine deposits. As the shallow subtidal area is lessened, sediment production is reduced, but previously deposited subtidal sediment (including reefs) may be remobilized as the zone of shoreline processes retreats through them and removes some, or all, of the subtidal record. As the shorelines approach the bank margins, there may be a large pulse of bank-derived sediment.

Fig. 3A-18. Photograph showing a calcarenite protosol that forms a horizontal protrusion in the center of this outcrop of regressive-phase eolianite of the Cockburn Town Member (Grotto Beach Formation) at The Bluff, San Salvador Island. (Photo previously published in Carew and Mylroie, 1995a.)
delivered to the deep environments off bank. Peloidal and bioclastic allochems will be important constituents of regressive-phase deposits where shoreline processes “chew-up” reefs and other subtidal deposits. Some of that sediment is reworked into regressive-phase dunes that may bury subtidal deposits that survive the passage through the retreating coastal zone (Carew and Mylroie, 1995a). Protosols commonly develop between times of major dune-building events during the stillstand and regressive phases (Fig. 3A-18; Table 3A-1). Regressive-phase dunes are likely to be well vegetated, and to bury vegetation, so regressive-phase eolianites commonly contain abundant vegemorphs and typically lack fine-scale bedding. Spectacular vegemorphs, often with abundant fossil pulmonate snails, are especially noted in the upper several meters of Pleistocene regressive-phase eolianites, where buried vegetation and roots provided preferred pathways for descending meteoric water (Fig. 3A-19) (Table 3A-1). Regressive-phase eolianites are occasionally seen to overlie fossil reefs (Fig. 3A-20). These regressive-phase eolianites generally should not be subjected to substantial wave erosion, as transgressive-phase eolianites are, but they may experience wave erosion during succeeding sea-level highstands.

Fig. 3A-19. Photograph showing spectacular development of vegemorphs below the terra rossa paleosol that caps an exposure of regressive-phase eolianite of the Cockburn Town Member (Grotto Beach Formation) at Crab Cay, San Salvador Island. Such spectacular vegemorphs are usually associated with regressive-phase eolianites. (Photo courtesy of Jim Teeter.)
Fig. 3A-20. Facies of the Cockburn Town Member seen at The Gulf, San Salvador Island, where a regressive-phase eolianite and a calcarenite protosol overlie a substage 5e reef-rubble deposit that was probably torn up by wave action when sea level fell past an unprotected (not buried) reef during the regression from the stillstand of the 5e sea-level highstand. (From Carew and Mylroie, 1995a.)

Lowstand phase. Once sea level falls more than 10 m below its present position, the Bahama banks are largely subaerially exposed. From then until sea level again rises above -10 m, only subaerial diagenetic processes and products (e.g., terra rossa paleosol development, pedogenesis; dissolution, karstification; and cementation) occur on the banks/islands. As previously discussed, such exposure has been about an order of magnitude longer than the time that the banks have been flooded. For further details about Bahamian paleosols and karst see the discussion elsewhere in this chapter, and in Carew and Mylroie (1991, 1995a), Boardman et al. (1995), Foss and Bain (1995), Mylroie and Carew (1995), Mylroie et al. (1995b).

STRATIGRAPHY OF BAHAMIAN ISLANDS

Overview

Stratigraphic studies of the surficial deposits of the Bahamas have used a variety of geologic evidence to support various stratigraphic interpretations. The major types of evidence commonly used include products of depositional processes (e.g., sedimentary structures, landforms, facies distribution), products of subaerial diagenesis (e.g., soil formation, dissolution-precipitation of limestone), fossil content, geochronologic determinations, and predictions made from modern analogs. Each technique has strengths and weaknesses (Table 3A-2). Detailed interpretations of the depositional/erosional history of a Bahamian island may be attempted through the integration of as many of these lines of evidence as possible (e.g., Garrett and Gould,
<table>
<thead>
<tr>
<th>Method</th>
<th>Utility</th>
<th>Difficulties</th>
<th>Relevant references</th>
</tr>
</thead>
<tbody>
<tr>
<td>Paleosols</td>
<td>May mark stratigraphic boundaries.</td>
<td>May be misinterpreted.</td>
<td>20, 5, 7</td>
</tr>
<tr>
<td>Terra rossa</td>
<td>May separate deposits from different highstands.</td>
<td>Cave fills.</td>
<td>5</td>
</tr>
<tr>
<td>Calcarenite protosol</td>
<td>Identify pauses in deposition during highstand.</td>
<td>Composite paleosols that represent more than one highstand/lowstand. Penetrative calcrete. May be misinterpreted as surfaces that separate different highstands.</td>
<td>5, 10, 11</td>
</tr>
<tr>
<td>Petrology</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Allochems</td>
<td>May aid in identifying depositional environment or stratigraphic unit.</td>
<td>Extreme lateral variability; lack of time dependency.</td>
<td>30, 7, 21, 22</td>
</tr>
<tr>
<td>Cements and diagenesis</td>
<td>May be clues to depositional and post-depositional environment.</td>
<td>Extreme lateral variability and complex overprinting</td>
<td>28, 35, 7</td>
</tr>
<tr>
<td>Sedimentary structures</td>
<td>May aid in identifying depositional environment.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Herringbone cross-bedding</td>
<td>Indicates subtidal deposits.</td>
<td></td>
<td>7</td>
</tr>
<tr>
<td>Fenestral porosity</td>
<td>May indicate intertidal deposits.</td>
<td>Also known from eolianites and other locales.</td>
<td>31, 32, 2</td>
</tr>
<tr>
<td>Paleontology</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fossil coral</td>
<td>May indicate subtidal deposits.</td>
<td>Valid only in situ.</td>
<td>37, 7, 8</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pristine reefs indicate rapid burial, not regression per se.</td>
<td>37, 16, 7</td>
</tr>
<tr>
<td>Method</td>
<td>Utility</td>
<td>Difficulties</td>
<td>Relevant References</td>
</tr>
<tr>
<td>-------------------------</td>
<td>--------------------------------------------------------------------------</td>
<td>------------------------------------------------------------------------------</td>
<td>---------------------</td>
</tr>
<tr>
<td>Trace fossils</td>
<td>Can identify terrestrial, intertidal, and subtidal facies.</td>
<td>Must be cognizant of appropriate traces.</td>
<td>13, 14, 36, 11</td>
</tr>
<tr>
<td>Cerion</td>
<td>May be useful to identify stratigraphic units.</td>
<td>Common lack of morphologic distinction between units; individual islands differ.</td>
<td>15, 19, 23</td>
</tr>
<tr>
<td>Marine, lake, or terrestrial shells</td>
<td>May indicate marine, lake, or terrestrial deposits.</td>
<td>Hermit crabs and birds may dislocate shells.</td>
<td>17</td>
</tr>
<tr>
<td>Geochronology</td>
<td>May identify times of deposition.</td>
<td>Variable reliability among methods.</td>
<td>7</td>
</tr>
<tr>
<td>Carbon-14</td>
<td>Reliable for Holocene.</td>
<td>Yields allochem ages, not time of deposition.</td>
<td>1, 3, 7</td>
</tr>
<tr>
<td>Uranium-series</td>
<td>Useful for fossil coral and speleothems.</td>
<td>Alpha-count vs TIMS.</td>
<td>12, 7</td>
</tr>
<tr>
<td>Paleomagnetics</td>
<td>May be useful to distinguish between terra rossa paleosols.</td>
<td>Need unaltered material.</td>
<td>12, 7</td>
</tr>
<tr>
<td>Amino acid racemization</td>
<td>May help distinguish among units.</td>
<td>Young rock precludes reversals, so record of secular variation only, so resolution is difficult.</td>
<td>27, 7</td>
</tr>
<tr>
<td>Cerion</td>
<td>Could be useful for deposits and paleosols.</td>
<td>Correlation with other data is often poor.</td>
<td>4, 18, 7</td>
</tr>
<tr>
<td>Whole-rock</td>
<td>May help distinguish among deposits.</td>
<td>Data is commonly unreliable.</td>
<td>24, 6, 7, 9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Correlation with other data is often poor; yields composite allochem ages.</td>
<td></td>
</tr>
<tr>
<td>Geomorphology</td>
<td>Correlation of age and degree of karst development is not substantiated.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>------------------------</td>
<td>------------------------------------------------------------------------</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karst</td>
<td>Correlation of age and degree of karst development is not substantiated.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Morphostratigraphy</td>
<td>Field evidence is commonly contrary to hypothesis. Variable elevation of Quaternary sea levels scrambles relationships.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Holocene Comparisons</td>
<td>Holocene not a complete cycle; regressive phase has not yet occurred.</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

References: 1, Andersen and Boardman (1989); 2, Bain and Kindler (1994); 3, Boardman et al. (1989); 4, Carew and Mylroie (1987); 5, Carew and Mylroie (1991); 6, Carew and Mylroie (1994b); 7, Carew and Mylroie (1995a); 8, Carew and Mylroie (1995b); 9, Carew and Mylroie (1995c); 10, Carew et al. (1992); 11, Carew et al. (1996); 12, Chen et al. (1991); 13, Curran (1984); 14, Curran and White (1991); 15, Garrett and Gould (1984); 16, Greenstein and Moffat (1996); 17, Hagey and Mylroie (1995); 18, Hearty and Kindler (1993); 19, Hearty et al. (1993); 20, James (1972); 21, Kindler and Hearty (1995); 22, Kindler and Hearty (1996); 23, Marcy et al. (1993); 24, Mirecki et al. (1993); 25, Mylroie and Carew (1995); 26, Mylroie et al. (1995b); 27, Panuska et al. (1995); 28, Pelle and Boardman (1989); 29, Rossinsky et al. (1992); 30, Schwabe et al. (1993); 31, Shinn (1967); 32, Shinn (1983); 33, Titus (1980); 34, Titus (1987); 35, White (1995); 36, White and Curran (1993); 37, White and Curran (1995).
Table 3A-3
Comparison of stratigraphies proposed for Bahamian islands

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<tr>
<td></td>
<td></td>
<td></td>
<td>Recent Ls</td>
<td>R.B. Fm</td>
<td>Unnamed</td>
<td>R.B. Fm</td>
<td>R.B. Fm</td>
<td></td>
</tr>
<tr>
<td>HOLO-</td>
<td>Cene</td>
<td>Recent U, C</td>
<td>H.B. Mbr</td>
<td>E.B. Mbr</td>
<td>H.B. Mbr</td>
<td>H.B. Mbr</td>
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<td></td>
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<td></td>
<td>N.P. Mbr</td>
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<td>N.P. Mbr</td>
<td></td>
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<tr>
<td>P</td>
<td>3</td>
<td>A</td>
<td>n.r.</td>
<td>G.L. Oolite</td>
<td>n.r.</td>
<td></td>
<td>**</td>
<td>n.r.</td>
</tr>
<tr>
<td>L</td>
<td>5a</td>
<td>Y</td>
<td>G.H. Ls</td>
<td>G.B. Fm</td>
<td>D.H. Ls</td>
<td>A.C. Fm2</td>
<td>**</td>
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<td>E</td>
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<td>A</td>
<td>D.H. Mbr</td>
<td></td>
<td>Upper Mbr</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>I</td>
<td></td>
<td>N</td>
<td>D.H. Mbr</td>
<td></td>
<td>Lower Mbr</td>
<td></td>
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<td>N</td>
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<td>S</td>
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<td>Unit II (Stage 7)</td>
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<td>Unit I (Stage 9?)</td>
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Abbrev: A.C., Almgreen Cay; C.T., Cockburn Town; D.H., Dixon Hill; E.B., East Bay; Fe.B., Fernandez Bay; Fr.B., French Bay; F.H., Fortune Hill; G.B., Grotto Beach; G.H., Grahams Harbour; G.L., Granny Lake; H.B., Hanna Bay; N.P. North Point; O.H., Owl's Hole; R.B., Rice Bay. n.r., no unit recognized in this position.

1Titus denoted his G.H. Ls and G.B. Ls only as Pleistocene, and made no correlation with oxygen isotope stages or absolute ages.


** No positively identifiable deposits at this position.
1984; Hearty and Kindler, 1993; Kindler and Hearty, 1996). However, because of the complexities created by the spatially patchy nature of deposition during highstands, differential erosion during lowstands, variability of amount and location of pre-existing high ground, and differences in sea level within and among highstands, such reconstructions are inherently interpretive — and possibly debatable (although provocative), or even wrong. On the other hand, our goal has been to develop a lithostratigraphic column for the Quaternary limestones of Bahamian islands (e.g., Carew and Mylroie, 1985, 1995a) that conforms to the Code of Stratigraphic Nomenclature (NACSN, 1983) and is based on criteria that can be utilized in the field, not only by ourselves but by others.

Beach and Ginsburg (1980) assigned all late Pliocene through Quaternary rocks in the Bahamas to the Lucayan Limestone (see Table 3A-3). The base of the Lucayan Limestone was defined biostratigraphically as coincident with the upper limit of the coral *Stylophora affinis* and a diagnostic molluscan assemblage equivalent to the Bowden Formation in Jamaica (Beach and Ginsburg, 1980; McNeill et al., 1988). The top of the Lucayan was defined as the present-day discontinuity surface, which is the land surface on Bahamian islands, and is recognized seismically beneath Holocene subtidal deposits on the submerged banks (Beach and Ginsburg, 1980). The thickness of the Lucayan Limestone is known to vary from about 43 m on Andros Island to as little as 10.5 m on Mayaguana Island (e.g., Cant and Weech, 1986). Magnetostratigraphic study of a core from San Salvador Island has suggested an age of 2.6–2.7 Ma (late Pliocene) for the base of the Lucayan Limestone (McNeill et al., 1988).

Studies of the surface geology of San Salvador Island led to abandonment of the term Lucayan Limestone for surficial rocks, because it was possible to recognize a more detailed stratigraphy. Moreover, the Lucayan as defined, mistakenly placed Holocene transgressive- and stillstand-phase rocks exposed on Bahamian islands within the Lucayan Limestone, while assigning currently-subtidal Holocene deposits to the post-Lucayan.

The first proposed stratigraphic column for the exposed rocks of a Bahamian island was that of Titus (1980) (see Table 3A-3). He interpreted the rocks of San Salvador Island as Pleistocene deposits that were laid down during sea-level regression from highstands. He made no suggestion concerning when in the Pleistocene they were deposited, and he indicated only that those units rested on pre-Pleistocene biomicrite.

In 1984, Garrett and Gould proposed phases of deposition for New Providence Island, but they did not tie the phases to a precise chronology or stratigraphy. The following year, we (Carew and Mylroie, 1985) proposed a revision to the stratigraphy of San Salvador (see Table 3A-3) because we recognized that: (1) much of the rock that Titus assigned to the Grahams Harbour Limestone, defined by Titus (1980), is Holocene rather than Pleistocene; (2) the rock cited as the type section for the Grahams Harbour Limestone does not correlate with the majority of rock assigned to that unit; (3) there is an older eolianite beneath Titus’s Grotto Beach Limestone at its type locality and elsewhere; and (4) substantial portions of the rock record on San Salvador were deposited during the transgressive and stillstand phases of sea-level highstands, rather than only during the regression.
Later, Titus (1987) revised his stratigraphy to accommodate then-current information (see Table 3A-3), and later we revised our stratigraphy because amino acid racemization (AAR) data had been utilized (inappropriately) to define parts of our previous stratigraphic column (see Carew et al., 1992). Using morphostratigraphy and whole-rock AAR data, Hearty and Kindler (1993) proposed two additional formation-rank stratigraphic units and several members; more recently Kindler and Hearty (1996) proposed a total of eight units based on AAR data and petrology (see Table 3A-3). The units proposed by Hearty and Kindler (1993) were time-stratigraphic (hence interpretive) units, not lithostratigraphic units, despite their use of formal stratigraphic terminology (see discussion in Carew and Mylroie, 1994b, 1995a,c and Hearty and Kindler, 1994). More recently, Kindler and Hearty (1996) substituted a number designation for the proposed units of their interpretive history of Bahamian islands. [In Chap. 3B, Kindler and Hearty explicitly use a time-stratigraphic scheme. — Eds.]

Recently, it has been suggested (Kindler and Hearty, 1996) that it is possible to identify deposits formed during separate highstands of sea level, and derive the position of sea level at the time of deposition (relative to present sea level) based upon allochem composition of the rocks. Although there are some generalities that seem to apply to some of the surficial rocks of Bahamian islands, reliance on such criteria as grain composition is, we believe, inappropriately simplistic. Differences in allochem character are likely to represent differences, or changes, in source area during a single highstand, rather than deposition during different highstands and sea-level positions. On San Salvador Island, for example, the Holocene transgressive-phase eolianites (deposited when sea level was at least several meters below its present position) often contain abundant superficial ooids and coated grains, but they become progressively more peloidal/bioclastic up-section (Carney et al., 1996; White, 1995). This change may be related to the growth of reefs up to wave base, and subsequent change in lagoon dynamics.

Our stratigraphic column of the Bahamian islands consists of three major lithostratigraphic units (Carew and Mylroie, 1995a) (see Fig. 3A-21). As each of these units is a depositional package that is (or will be) bounded by unconformities, that largely represent times of low sea level, they are also allostratigraphic units (NACSN, 1983). As this stratigraphy was initially developed on San Salvador Island, the nomenclature refers to locales there, and all type locations are on San Salvador; however, the stratigraphy is applicable throughout the Bahamas, and has been used by us and other geologists on many other Bahamian islands (e.g., Andersen and Boardman, 1989; Curran and White, 1991; White and Curran, 1993, 1995; Carew and Mylroie, 1995a, and references therein; Kindler, 1995). A brief discussion of Bahamian stratigraphy follows; for a more thorough treatment see Carew and Mylroie (1995a).
Fig. 3A-21. Lithostratigraphic column for the Bahama islands. In the field, individual units are not necessarily seen stacked atop one another, but are often found lateral to one another. The thin stippled and black layers are terra rossa paleosols, and they separate deposits formed during separate sea-level highstands. Where there are no intervening deposits such terra rossa paleosols represent the total time of one or more complete glacioeustatic sea-level cycles. (From Carew and Mylroie, 1995a.)

Nomenclature

Owl’s Hole Formation. The oldest rocks exposed on Bahamian islands are assigned to the Owl’s Hole Formation. By definition, the Owl’s Hole Formation consists of eolianite that is capped by a terra rossa paleosol that can be shown to be overlain by either a highly oolitic eolianite that is itself capped by a second terra rossa paleosol, or by subtidal deposits (Carew and Mylroie, 1995a). The age of the interglacial sea-level highstand(s) during which these eolianites were deposited has not been conclusively established, but based on plausible isostatic subsidence rates, and the late Quaternary glacioeustatic history (Fig. 3A-8), they most likely represent one or more of the interglacial highstands associated with oxygen isotope stages 7 (~220 ka), 9 (~320 ka), or 11 (~410 ka) (Carew and Mylroie, 1995a). According to Kindler and Hearty (1995), eolianite deposits from two separate pre-5e interglacial highstands can be identified in exposures at the Cliffs section on Eleuthera Island.
In nearly all cases, Owl's Hole eolianites consist of fossiliferous pelsparites and peloidal biosparites (fossiliferous and peloidal grainstones) (Carew and Mylroie, 1995a; Kindler and Hearty, 1995), but oolitic rocks are also known from this unit, for example, on New Providence Island (Schwabe et al., 1993; Hearty and Kindler, 1995). Owl's Hole rocks are often extensively micritized at the exposed surface, but portions remain relatively weakly cemented. From detailed study of the wall rock of many caves in the eolianite ridges of several Bahamian islands, and the outcrops of the ridges themselves, it has recently been shown that Owl's Hole rocks underlie many of the large Pleistocene eolianite ridges, and form more of the landscape of Bahamian islands than was previously thought (Schwabe et al., 1993; Carew and Mylroie, 1995a; Kindler and Hearty, 1995; and references therein).

**Grotto Beach Formation.** The most widespread depositional package exposed on Bahamian islands is the Grotto Beach Formation (Fig. 3A-21). It comprises eolianites and beach-face to subtidal marine limestones that, at places, can be subdivided into two members. The formation is capped by a terra rossa paleosol, except where it has been removed by later erosion. The Grotto Beach Formation contains exposed subtidal facies that are up to 5 m above modern sea level on numerous Bahamian islands, which is consistent with deposition during the substage-5e sea-level highstand (~132–119 ka, Chen et al., 1991; ~131–114 ka, Szabo et al., 1994; Carew and Mylroie, 1995b). Throughout the Bahamas, the transgressive-phase and some stillstand-phase eolianites of the Grotto Beach Formation are characterized by the abundant (up to 90% of the allochems) well-developed ooids that are similar to those seen at Joulter Cays today (Fig. 3A-22) (Carew and Mylroie, 1995a; Kindler and Hearty, 1995). Most of the subtidal facies and the regressive-phase eolianites of the Grotto Beach Formation are dominantly peloidal or bioclastic, but they com-

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**Fig. 3A-22.** Photograph of thin section showing typical ooids from an eolianite of the Grotto Beach Formation on South Andros Island. Field of view is ~1.8 mm. (Photo previously published in Carew and Mylroie, 1995a.)
monly contain ooids, except where they are close to a source of bioclastic debris (Carew and Mylroie, 1995a). Also, in some localities, such as on North and South Andros, there are abundant late Pleistocene oolitic subtidal shoal/beach deposits. Kindler and Hearty (1996) have suggested that oolitic deposits on the present major islands imply that sea level was above the current datum at the time of deposition. If one projects such an interpretation to a future sea-level highstand that is a few meters lower than present, then the Holocene Joulter Cays oolitic deposits would be a part of the Andros Island geology, and they would be incorrectly interpreted to represent sea-level conditions higher than present. Furthermore, in our experience, ooids seem to be more common in transgressive-phase deposits, which are typically developed at a sea-level position below the acme of a highstand (see discussion of the French Bay and North Point members). Perhaps ooids are so abundant in rocks of the Grotto Beach Formation because that highstand (substage 5e) submerged the platform for a longer time and reached a higher elevation than stage 1 sea level; as a result, those deposits are disproportionately represented in the rocks exposed above sea level today.

**French Bay Member.** The French Bay Member comprises the transgressive-phase eolianites through beach facies of the Grotto Beach Formation (Carew and Mylroie, 1995a). These rocks are predominantly fine to medium oosparites (oolitic grainstones) that exhibit grain fall, grain flow, and climbing wind-ripple laminae, and limited vegemorph development (Table 3A-1). Additional evidence that these rocks were deposited during the transgressive phase include outcrops containing: (1) a fossil sea cave containing boulder rubble, (2) cliff-line paleotalus deposits, and (3) outcrops of overlying regressive-phase eolianites (Carew and Mylroie, 1985, 1995a). The French Bay Member can be recognized on many Bahamian islands (e.g., High Cay off South Andros Island; West Plana Cay; the Exuma islands; San Salvador Island). At all these places, fossil corals lie on a wave-cut surface carved into French Bay eolianites with no intervening paleosol, or evidence of an eroded paleosol (Fig. 3A-5A). Identical relationships can be seen today on some Holocene transgressive-phase eolianites (e.g., Fig. 3A-13).

**Cockburn Town Member.** The Cockburn Town Member comprises the subtidal and stillstand- through regressive-phase beach and eolian deposits of the Grotto Beach Formation (Carew and Mylroie, 1995a). Subtidal deposits extend up to ~5 m above current sea level, and commonly grade upward into, or are entombed by, stillstand- and regressive-phase beach and dune deposits (Carew and Mylroie, 1985, 1995a, b; White and Curran, 1995; Carew et al., 1996). The marine subtidal deposits of the Cockburn Town Member are recognized in the field by features such as herring-bone cross bedding, asymmetrical ripples (Fig. 3A-23), abundant fossil marine molluscs, corals and marine trace fossils (e.g., Ophiomorpha, see Fig. 3A-24), and by coral reefs. Curran and White (1985) provided a detailed map and cross section illustrating facies changes at Cockburn Town fossil reef on San Salvador Island, and White (1989) described and illustrated the Sue Point fossil reef (see Fig. 3A-25). The near pristine preservation of many fossil reefs in the Bahamas
(Fig. 3A-25) indicates that they were catastrophically buried before the regression at the termination of the 5e highstand (Carew and Mylroie, 1995a, and references therein; Greenstein and Moffat, 1996). At the shoreline cliffs at Clifton on New Providence Island, subtidal shoal deposits can be seen to grade upward to beach facies (Garrett and Gould, 1984; Carew et al., 1992; Carew and Mylroie, 1995a; Carew et al., 1996). Precise mass-spectrometric $^{234}\text{U}/^{230}\text{Th}$ ages from fossil coral reefs on San Salvador and Great Inagua islands indicate that the substage-5e highstand lasted from about 132 to 119 ka (Chen et al., 1991). Data from in situ fossil coral reefs throughout the Bahamas are consistent with deposition during only that highstand (Carew and Mylroie, 1995b). White and Curran (1995) have suggested that there may have been a minor short-lived depression of sea level to at least
the position of current sea level during the 5e highstand, and TIMS $^{234}\text{U}/^{230}\text{Th}$ dates from corals above and below the purported erosion surface indicate that the low may have lasted no more than $10^3$ years centered at about 125 ka.

The stillstand through regressive-phase beach facies and eolianites are also assigned to the Cockburn Town Member because there is an unbroken gradation from marine to eolian rocks at many outcrops, and no terra rossa paleosol separates the marine and eolian facies (see Carew and Mylroie, 1995a). Eolianites of the Cockburn Town Member exhibit some, or all, of the following (Table 3A-1): disrupted internal bedding, calcarenite protosols, abundant vegemorphs, beach-face breccia facies, and eolianites overstepping fossil reefs; for examples, see Carew and Mylroie (1995a). Cockburn Town eolianites are commonly capped by elaborate paleosols with vadose pisolites, complex caliche/calcrete crusts, and abundant fossil pulmonate snails (mostly Cerion); unlike eolianites of the French Bay Member, eolianites of the Cockburn Town Member lack evidence of wave attack coeval with the highstand during which the dunes formed (Carew and Mylroie, 1995a). Subtidal shoal, lagoonal, and ebb-tidal delta deposits of the Cockburn Town Member occur up to ~5 m above present sea level on many Bahamian islands (e.g., Garrett and Gould, 1984; Titus, 1987; Carew et al., 1992, 1996; Carew and Mylroie, 1995a; Hagey and Mylroie, 1995; Noble et al., 1995; White and Curran, 1995; and references therein).

Formerly, we assigned some of the Grotto Beach Formation to a separate member (Dixon Hill Member) that was erroneously thought to have been deposited
Fig. 3A-25. Photograph of in situ Acropora palmata at Sue Point fossil reef, San Salvador Island. This superb preservation of elk horn coral in current-oriented growth position (inclined seaward) in the Cockburn Town Member of the Grotto Beach Formation indicates that it was protected by rapid burial before sea-level regression. (Photo previously published in Carew and Mylroie, 1995a.)

in association with substage 5a (Carew and Mylroie, 1985). We eliminated that member from our stratigraphy in 1992. For a more detailed discussion of this issue see Hearty and Kindler (1993, 1994) and Carew and Mylroie (1985, 1994a,b, 1995a,c).

Rice Bay Formation. The Holocene Rice Bay Formation comprises all rocks above the paleosol that caps the Grotto Beach Formation (Fig. 3A-21) (Carew and Mylroie, 1985, 1995a). Throughout the Bahamas, the Rice Bay Formation consists of eolianites and beach facies rocks that have been deposited during the transgressive and stillstand phases of the current sea-level highstand (stage 1). In places, two members can be recognized by differences in bedding character, allochem composition, and their position relative to current sea level (Carew and Mylroie, 1985, 1995a). Although there is some incipient development of thin calcretes (<1 mm) on some transgressive-phase eolianites of the Rice Bay Formation, terra rossa paleosols are absent on Rice Bay rocks. However, calcarenite protosols are currently forming in coastal areas and in swales between and on transgressive-phase eolianites.
Unlike the commonly oolitic beach and dune facies of the Grotto Beach Formation, rocks of the Rice Bay Formation are characterized by: (1) a generally low abundance of ooids (usually less than 25%, rarely up to 50%), especially high in the section; (2) the superficial nature and small size of those ooids (i.e., only a few laminae); (3) dominance of peloids and bioclasts, especially in the Hanna Bay Member; (4) limited diagenetic micritization; and (5) generally weak, meniscus, low-Mg calcite cements (Carew and Mylroie, 1985, 1995a). Superficial-ooid production occurred during the early phase of the Holocene transgression of the San Salvador platform, but in most places ooid production seems to have ceased by ~3 ky B.P. At Joulter Cays, Schooner Cays, and elsewhere, there are abundant well-developed Holocene ooids, but more generally, the Rice Bay Formation lacks such ooids, even where ooid shoals are present offshore (e.g., east coast of South Andros Island).

North Point Member. The North Point Member comprises the transgressive-phase eolianites (Table 3A-1; Figs. 3A-3, 3A-13) of the Rice Bay Formation. These rocks are commonly peloidal, but superficial ooids are common low in the section. Most rocks of the North Point Member have meniscus calcite cement, but in coastal outcrops there is occasional marine cement (Carew and Mylroie, 1995a, and references therein; White, 1995). At depth, these deposits are commonly uncemented. These eolianites were deposited when sea level was lower than at present, as indicated by steeply dipping foreset beds that continue at least 2 m below current sea level (Carew and Mylroie, 1985, 1995a). They are known on many Bahamian islands, but the most extensive deposits of this member that we have seen are found on Long Island.

Based solely on stratigraphic relationships, it was suggested that the North Point Member is <10 ky old (Carew and Mylroie, 1985). Radiocarbon ages obtained from whole-rock samples of North Point Member rocks range from 6.1 to 3.7 ky B.P., and average about 5 ky B.P. (Carew and Mylroie, 1995a; Boardman et al., 1987; Boardman et al., 1989). Apparently, significant sand was produced and incorporated into the North Point Member at a time that corresponds to the inflection on the Bahamian sea-level curve (see Boardman et al., 1989, Fig. 1; or Carew and Mylroie, 1995a, Fig. 17) that indicates the change to a slower rate of sea-level rise during the past ~4 ky.

Hanna Bay Member. The Hanna Bay Member comprises the stillstand-phase beach and eolian facies of the Rice Bay Formation. This member was initially limited to currently lithified rocks (Carew and Mylroie, 1985), but currently unlithified Holocene sediments and future regressive-phase deposits are now also considered to be part of the Hanna Bay Member, in similar fashion to the Cockburn Town Member of the Grotto Beach Formation (Carew and Mylroie, 1995a). This member consists largely of peloidal/bioclastic grainstones (except in ooid areas such as Joulter Cays) with predominantly meniscus low-Mg calcite cements. These rocks were deposited in equilibrium with current sea level; that is, lithified intertidal and beach facies of this member occur at the same elevation as corresponding facies of the modern beaches (Carew and Mylroie, 1985, 1995a). Radiocarbon ages of whole-
rock samples of the Hanna Bay Member range from ~0.3 to 3.2 ky B.P., and generally they are less than 2.5 ky B.P. (Boardman et al., 1987; Carew and Mylroie, 1995a). Rocks of the Hanna Bay Member are known on nearly all Bahamian islands and cays.

CONCLUDING REMARKS

The Quaternary depositional history of the shallow banks and islands of the Bahamas has been controlled principally by the glacioeustatic sea-level changes associated with glaciation and deglaciation of the continents. Significant production of carbonate allochems and mud occurred only when highstands of sea level flooded the bank tops (above ~10 m). As a result, the sedimentary record on Bahamian islands consists of packages of transgressive-phase, stillstand-phase, and regressive-phase deposits that were produced during the highest (interglacial) stands of Quaternary sea level. Between those relatively short depositional intervals, only subaerial erosion and fallout of atmospheric dust occurred on the platforms. Soils that would become terra rossa paleosols thus developed on the exposed surfaces, and now usually intervene between deposits of successive interglacials.

As a result of the glacioeustatic control of limestone deposition in the Bahamas, the lithostratigraphic units of Bahamian islands are also allostratigraphic units that are usually bounded by terra rossa paleosols. Because of the current high elevation of sea level, and the slow isostatic subsidence (1–2 m per 100 ky), the only marine subtidal deposits exposed on Bahamian islands are those deposited during oxygen isotope Substage 5e (~125 ka). Besides those subtidal rocks, eolianites possibly deposited during oxygen isotope stages 11 (~410 ka), 9 (~320 ka), 7 (~220 ka), and beach facies through eolianites of stages 5 (~125 ka), and 1 (present) comprise the surficial rocks of the islands of the Bahamas. Based upon physical stratigraphy, the rocks of the Bahamian islands can be divided into three major units: the middle Pleistocene Owl's Hole Formation, the overlying late Pleistocene Grotto Beach Formation, and the Holocene Rice Bay Formation. The formal stratigraphy that was first developed on San Salvador Island (Carew and Mylroie, 1985, 1995a) is applicable to all other Bahamian islands known to us.

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REFERENCES


134 J.L. CAREW AND J.E. MYLROIE


Neumann, A.C. and Hearty, P.J., 1996. Rapid sea-level changes at the close of the last interglacial (substage 5e) recorded in Bahamian island geology. Geology, 24: 775–778.


