Coastal Karst Development in Carbonate Rocks

4

John E. Mylroie

Abstract

Coastal karst cave development globally is biased towards the tropics and subtropics, where carbonate deposition is ongoing, and therefore carbonate coasts are common. The Carbonate Island Karst Model (CIKM) delineates the unique conditions that separate coastal karst from traditional karst areas of continental interiors. In these warm-water carbonate islands, diagenetically immature, or eogenetic carbonate rocks are host to a fresh-water lens that creates flank margin caves in a diffuse flow environment. Diagenetically mature, or telogenetic carbonates, can also host flank margin caves. Flank margin caves can form rapidly, as carbonate sediment is deposited, to produce syndepositional caves called banana holes. Flank margin caves can survive as open voids for millions of years, and as infilled diagnostic features for tens of millions of years. Vadose fast flow routes called pit caves form as a result of surface micritization to provide point recharge to the fresh-water lens. The presence of non-carbonate rocks can perch vadose flow, creating stream caves that terminate in the fresh-water lens. When sea level falls to create large exposed carbonate platforms, phreatic conduit flow develops to carry recharge to the platform periphery. Collapse of these conduits, as well as bank margin fracture, account for the majority of blue holes in carbonate platforms. Closed depressions in eogenetic carbonate islands are commonly constructional, relicts from variable carbonate deposition. The most common sinkhole type is the cave-collapse sinkhole. Morphometric analysis of flank margin caves supports cave origin as the amalgamation of individual chambers, provides evidence of denudation rates, and can differentiate flank margin caves from some pseudokarst cave types.

4.1 Introduction

An overview of cave types found in rocky coastlines is presented in Chap. 1. The unique morphology of surficial karst exposed on the coast is

J.E. Mylroie (✉)
Department of Geosciences, Mississippi State University, Mississippi State, MS 39762, USA
e-mail: mylroie@geosci.msstate.edu

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described in detail in Chap. 2. The hydrological and geochemical processes at work inside the fresh-water lens of carbonate islands and coasts is discussed in Chap. 3. This chapter will focus on karst cave development by dissolution in coastal carbonates.

Carbonate rocks, as presented here, represent three common mineral species: aragonite, calcite, and dolomite. Aragonite and calcite are both calcium carbonate, or $CaCO_3$, but with a different crystal structure, and these minerals form limestone. Aragonite is slightly more soluble than calcite in surface conditions. Most modern calcium carbonate is deposited as aragonite, so it is abundant in very young limestones worldwide. After deposition, aragonite inverts to calcite within tens to hundreds of thousands of years, depending on the nature of the environment and fluids in contact with the limestones, so most older limestones are made of calcite. Dolomite is a calcium and magnesium mixture, with a formal composition of $CaMg(CO_3)_2$, and can occur in a wide variety of forms (Machel 2004). Most dolomite in coastal settings is initially deposited as calcium carbonate, and is subsequently altered to dolomite by the process of dolomitization. Dolomite is theoretically as soluble as calcite, but in the real world it dissolves much more slowly. Calcite enriched in magnesium (18–20 % Mg), but well below the level of dolomite, is called high-Mg calcite and it is a common skeletal material in certain marine organisms, such as echinoderms and red algae. As with aragonite, and unlike dolomite, high-Mg calcite tends to dissolve readily or alter quickly to regular calcite. The carbonate metamorphic rock marble can be made from either limestone or dolomite. Evaporites, for the reasons given in Chap. 1, are extremely rare in coastal settings and will not be discussed here.

Karst caves in coastal settings are found in very young carbonate rocks composed mostly of aragonite and high-Mg calcite, as well as in older carbonates composed mostly of calcite. Dolomite hosts karst caves as well, but in coastal regions there is uncertainty regarding the timing of cave formation as regards the dolomitization of the limestones. Three options can be presented: (1) the caves develop in limestone, after which the limestone is dolomitized but the caves remain; (2) the limestone is dolomitized, and then the caves form; and (3) cave formation and dolomitization occur concurrently with cave dissolution (perhaps as a linked process, i.e. "syngenetic"). Caves have been found in dolomite on numerous islands (e.g. Barbados, Curacao, Isla de Mona), and they seem comparable in form and morphology as those found in limestones.

As noted above, as Cenozoic limestones get older, their original aragonite and high-Mg calcite mineralogy changes to a low-Mg calcite mineralogy, generally within a few hundred thousand years at the most. This alteration can occur even when the limestones involved have not been buried, or subjected to heat and pressure. As geologic time proceeds, limestones are commonly buried, and as a result begin to undergo a change in character called *diagenesis*. This process is gradual, but over the length of geologic time, it can significantly alter the nature of a limestone unit. The stages of maturation of a limestone were described by Choquette and Pray (1970) as subdivisions of the postdepositional evolution of carbonate porosity into three *time-porosity stages* conforming to the rock cycle. They defined "the time of early burial as *eogenetic*, the time of deeper burial as *mesogenetic*, and the late stage associated with erosion of long-buried carbonates as *telogenetic*" (Choquette and Pray 1970, p. 215). Vacher and Mylroie (2002, p. 183) defined eogenetic karst as: "the land surface evolving on, and the pore system developing in, rocks undergoing eogenetic, meteoric diagenesis." They further defined *telogenetic karst* as: "the karst developed on and within ancient rocks that are exposed after the porosity reduction of burial diagenesis." Islands can exhibit both types of karst depending on their geologic history, for example the Bahamas as an eogenetic karst example (Chap. 7), and coastal New Zealand as a telogenetic karst example (Chap. 17). The stages of limestone diagenetic maturation and the subsequent impact on fluid flow are presented in Fig. 4.1.

The vast majority of carbonate coasts and islands are in the tropics and subtropics, where limestone deposition is currently ongoing. As a

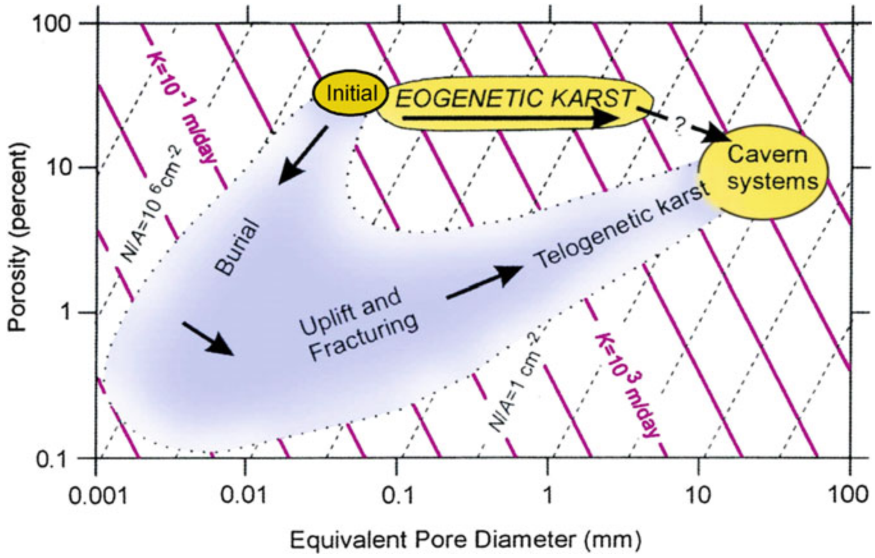

Fig. 4.1 The evolution of hydraulic conductivity in carbonate rocks (as modeled by straight tubes), showing the reduction in porosity and permeability with deep burial. Note that eogenetic rocks take a "short cut" to cavernous porosity (From Vacher and Mylroie 2002)

result, the limestones are considered eogenetic. They retain much of their original depositional porosity, and their allochems (organic and inorganically precipitated rock particles) are readily identifiable. As noted in Chap. 2, this eogenetic character of the rocks has a major influence on the nature of coastal karren. There is also a major impact on the development of karst caves. Vacher and Mylroie (2002) noted two important aspects of island karst development. First, one must differentiate between "karst on islands", and "island karst". The former describes karst features found on islands, almost always in the interior, where the karst landscapes resemble what is found in the interior of continents. Such karst is unaffected by the island setting, and lacks a signature associated with coastal conditions. Many features that the casual observer associates as island karst, such as mogotes in Puerto Rico or cockpit karst in Jamaica, are not restricted to the island setting at all. Those features are the result of the climate of the area, base level history, and the nature of the limestone and are identical to what would be found in a nearby continental setting, such as the interior of Belize. Island karst, on the other hand, refers to karst features that are linked to the coastal environment. These karst features are influenced by sea-level change and fresh water mixing with sea water, conditions not present in island or continental interiors.

Second, islands can be considered a special case of a carbonate coastline, which can be found along the edges of continents as well. South Florida and Yucatan would be examples of eogenetic carbonate coasts on continents (Chaps. 15 and 16), whereas the coast of the Adriatic Sea would constitute a telogenetic continental example (Chap. 17). Carbonate islands, especially small islands, offer limits in space that constrain models used to explain the karst features they contain. In tropical settings, these carbonate islands are commonly made up of eogenetic limestones, and in some cases such as the Bahamas, the limestones are only Middle Pleistocene to Holocene in age. This young age of the limestones provides limits on the time available to produce the observed karst features. Island water budgets are isolated and also limited, adding water budget constraints. The reason that much of the coastal cave and karst literature is focused on islands is because of the constraints offered by that setting. The Bahamas provide almost the ideal example (Chap. 7). The islands are all very young, with eogenetic limestones, and island size ranges from very small to very large. The small islands exist on both very small and on very large platforms, allowing island size versus platform size to be evaluated. They are in a quiescent tectonic setting, such that sea-level fluctuation is solely the result of glacioeustasy and long-term isostatic subsidence. Much of the early work on island karst was done in the Bahamas, as the simplicity of that location, and the time and spatial constraints in place, allowed easy recognition of the factors controlling the karst landforms. As other island groups were studied, many formed within more complex tectonic settings, the basic foundation factors could be determined despite the complexity of the site-specific situation. The result was the development of an island karst classification scheme.

4.2 The Carbonate Island Karst Model (CIKM) and Coastal Cave Development

The carbonate island karst model, or CIKM, is an attempt to characterize the levels of complexity found within carbonate islands, and by extension, continental carbonate coasts. The parameters that control the development of island karst were initially outlined by Mylroie and Vacher (1999) and codified as the Carbonate Island Karst Model, or CIKM, by that name after the initial study of Guam (Mylroie and Jenson 2000; Mylroie et al. 2001). As with any theoretical construct, the CIKM has been modified and evolved over the years as new data have emerged from many coastal karst areas (e.g. Mylroie and Mylroie 2007). The principles of the CIKM include:

1. Mixing of fresh and salt water at the boundaries of the fresh-water lens results in a localized area of preferential porosity and permeability development. Collection of organics at these boundaries may also enhance dissolution. The maximum dissolution occurs at the lens margin, where the water table and halocline mixing zones, and their associated organics, are superimposed, and where lens flow velocities are fastest.
2. Sea level change, which also moves the position of the fresh-water lens.
   (A) Glacioeustacy has moved sea level, and thus the fresh-water lens position, up and down more than 100 m throughout the Quaternary (note, for this chapter the glacioeustatic sea-level events will be identified by their Marine Isotope State, or "MIS"; see Fig. 1.5, Chap. 1).
   (B) Local tectonic movement, sediment compaction, and isostatic adjustment can cause overprinting of dissolutional and diagenetic features developed during different glacioeustatic events.
3. The karst is commonly eogenetic in that it has developed on rocks that are young and have never been buried below the zone of meteoric diagenesis; however telogenetic rocks do occur on carbonate islands and coasts.
4. Carbonate islands can be divided into four categories based on carbonate and non-carbonate relationships, and sea level (Fig. 4.2).
   (A) Simple Carbonate Island—Only carbonate rocks are present within the range of the fresh-water lens position (Fig. 4.2a). Meteoric catchment is entirely autogenic and flow within the fresh-water lens is controlled entirely by properties of the carbonate rock. The Bahamas are examples of simple carbonate islands.
   (B) Carbonate-Cover Island—Only carbonate rocks are exposed at the surface and the catchment is entirely autogenic (Fig. 4.2b). Non-carbonate rocks exist under the carbonate rocks and may partition and influence flow within the lens, including conduit flow at the carbonate and non-carbonate contact. Bermuda (at a glacioeustatic sea-level lowstand) is an example of a carbonate-cover island.
   (C) Composite Island—Both carbonate and non-carbonate rocks are exposed at the surface (Fig. 4.2c), allowing for allogenic and autogenic catchment. The lens is partitioned and conduit cave systems can develop at the contact of the carbonate

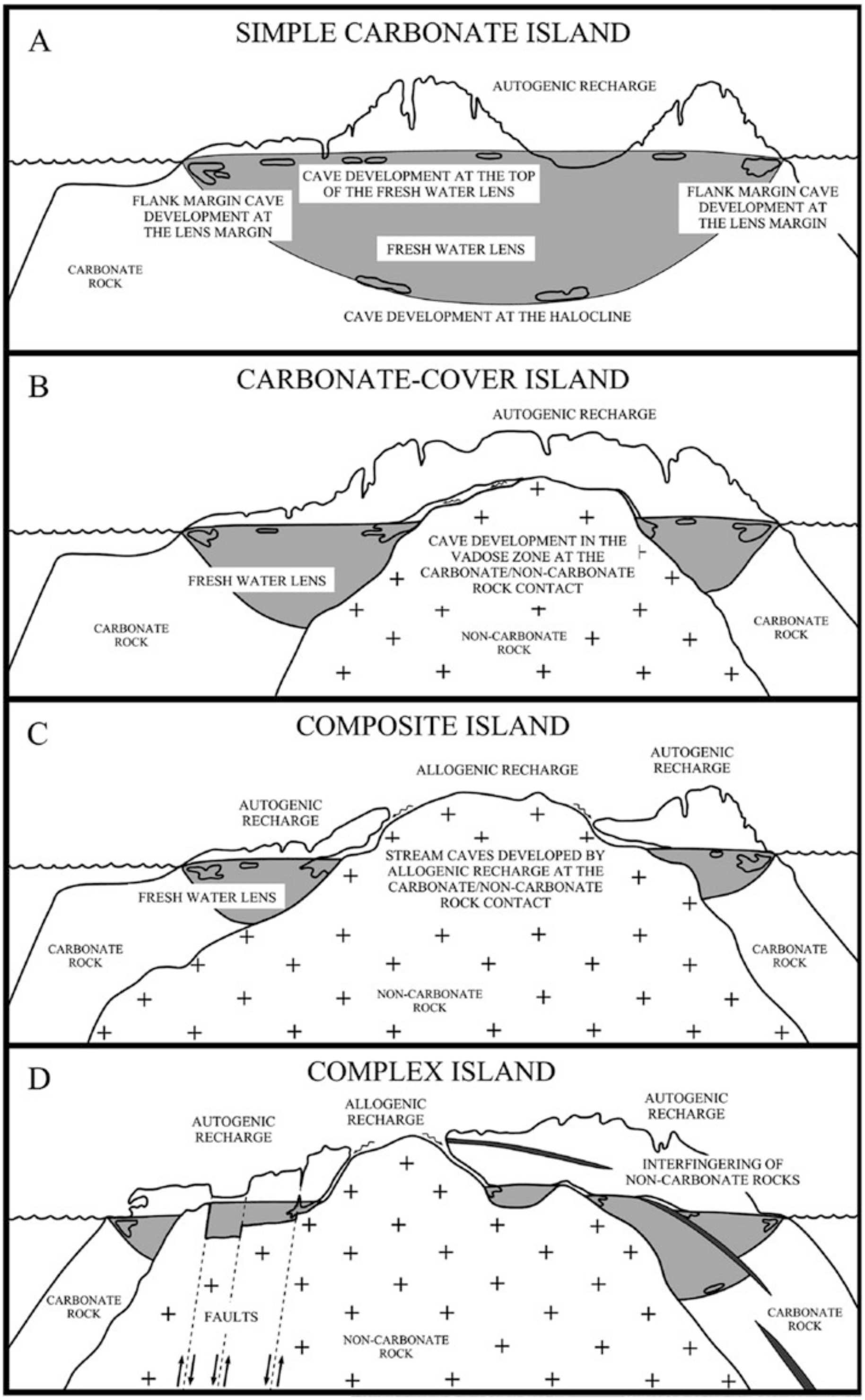

Fig. 4.2 The classification of island type within the CIKM, based on carbonate and non-carbonate rock interactions. Gradational conditions, or the subdivisions of portions of islands with respect to island type, are possible. (**a**) Simple Carbonate Island. (**b**) Carbonate Cover Island. (**c**) Composite Island. (**d**) Complex Island. See text for details

and non-carbonate rocks. Barbados and Guam are examples of composite islands.

(D) Complex Island—Carbonate and non-carbonate rocks are complexly interrelated by depositional relationships and/or faulting (Fig. 4.2d). Perching, isolation, and confining of the freshwater lens is possible. Saipan is an example of a complex island.

Autogenic recharge in Fig. 4.2 refers to collection of meteoric water directly on to the carbonate rock surface. As a result, the water is chemically modified in terms of its dissolutional potential as it descends by diffuse flow through the vadose zone to the top of the fresh-water lens. Allogenic recharge is collection of meteoric water on a non-carbonate surface, where the flow can concentrate into streams that flow laterally and then contact the carbonate rocks. The water enters the carbonate rocks as discrete point inputs that still retain most of their dissolutional potential. The difference between autogenic and allogenic catchment is both a geochemical and flow configuration difference.

The rigid classification of islands into one of these four categories is commonly not possible. Instead, regions or portions of islands can be classified according to the CIKM, which may assist in explaining the variety of cave and karst development that can occur on carbonate islands. Even the use of the term "carbonate island" needs care. An island can be primarily a non-carbonate rock, as in the Hawaiian Islands, and still contain isolated carbonate outcrops with caves and karst, for example the eolian calcarenites at Kahuku Point on the north end of Oahu, to which the CIKM can be applied. Chapter 12 presents another example of a volcanic island with minor carbonate outcrop.

While the allogenic catchment observed on composite and complex islands would appear to be karst on islands as opposed to island karst, in many islands the stream caves that developed from this allogenic flow discharge to the fresh-water lens, and not to the shoreline. This flow dynamic makes the allogenic catchment flow quite different from traditional stream caves found in continental interiors, or in the interior areas of large islands like Cuba or Jamaica.

4.3 Cave Development on Carbonate Islands

As Fig. 4.2 shows, a wide variety of dissolutional voids can develop on carbonate islands. To begin this discussion, the simplistic Bahama Islands will be examined. Chapter 7 is an overall perspective on cave development and distribution in the Bahamas, the discussion here will focus on the mechanisms involved.

4.3.1 Flank Margin Caves

In simple carbonate islands such as the Bahamas, extensive dry caves can be found that developed during a past glacioeustatic sea-level highstand. They are enterable today because while the present is a state of a high sea-level elevation, current sea level is not as high as earlier in the Quaternary (e.g. last interglacial or MIS 5e ∼124–115 ka, Thompson et al. 2011). Within the fresh-water lens (see Chap. 3 and figures therein, or refer to Fig. 4.2), dissolution is favored at specific sites. The top of the lens is where descending vadose water mixes with the phreatic water of the lens. These waters are both commonly saturated with $CaCO_3$, but at different initial conditions. As explained in Chap. 3, this mixing of waters results in enhanced dissolution of $CaCO_3$. Similar mixing dissolution occurs at the base of the fresh-water lens, where fresh water and marine water mix (called a *halocline* if the boundary is sharp, a *mixing zone* if the boundary is gradational in salinity). Of these two mixing zones, the fresh water mixing with marine water at the bottom of the lens appears more important, perhaps because the top of the lens is vulnerable to CO_2 degassing, which may counter-act the mixing dissolution environment there.

Both the top and bottom of the fresh-water lens are density boundaries. Organics that collect at these two boundaries can oxidize to produce excess CO_2 to drive dissolution. If organic loading is significant, the system may become anoxic, allowing a variety of unusual biogeochemical environments that can promote $CaCO_3$ dissolution. In blue holes, the large surface openings accept major amounts of organic debris, including macroscopic vegetative remains as well as particulate matter, which collect at the halocline. In those cases, anoxia at the halocline produces H_2S which drives dissolution (Bottrell et al. 1991). For regular vadose flow in carbonate aquifers, large organic material is filtered out, and only dissolved organic carbon (DOC) and small particulate matter can reach the fresh-water lens, although vadose fast-flow routes called pit caves can transport large organic particles downward at specific sites. In continental settings, it has been shown that DOC and fine particulate organics can percolate downward through the vadose zone and promote dissolution within the aquifer (Wood and Petritus 1984). In the Bahamas it has been shown that organics can migrate to the fresh-water lens, and through biomediation of sulfur enhance dissolution in flank margin caves (Bottrell et al. 1993).

The fresh-water lens integrates all the catchment it receives and discharges it to the ocean. The lens also thins as it approaches the coast. These two factors result in the fastest flow velocities in the lens being found at the distal margin of the lens. Chemical reactions that occur here will have the advantage in that reactants are brought in, and products removed, faster than elsewhere in the lens. The lens margin also brings the top and bottom of the lens into the same area, such that the mixing impact of each environment is added together, as is the geochemical activity of any included organics at those boundaries. As a result, the greatest amount of $CaCO_3$ dissolution should occur at the lens margin. Decades of field work in the Bahamas and in other localities (Mylroie and Mylroie 2007) has shown that the largest caves found on small, simple carbonate islands are found in this location. These caves are named *flank margin caves* because of the location in what would have been the margin of the lens, under the flank of the enclosing land mass. The term was developed in the late 1980s (Mylroie 1988; Mylroie and Carew 1990) based on work in the Bahamas, and has since been expanded to a variety of island types around the world, including Isla de Mona (Frank et al. 1998), the Mariana Islands (Jenson et al. 2006), Puerto Rico (Lace 2008), Barbados (Machel et al. 2011), the Adriatic coast (Otonicar et al. 2010), Australia (Mylroie and Mylroie 2009a), and New Zealand (Mylroie et al. 2008b). The dissolutional mechanism is extremely powerful, and cuts across a variety of primary and secondary features in caves, as shown in Fig. 4.3.

Flank margin caves develop as dissolutional voids within a diffuse flow regime. They form without entrances, and human access can only be later gained after other erosional processes, such as hillslope retreat or collapse, open them to the outside environment. They have a consistent passage pattern in part controlled by their size. Their size is in turn controlled by how long the fresh-water lens was stable at a single position. Small flank margin caves are commonly singles chambers with a few short side passages (Fig. 4.4). Larger caves are a collection of chambers, and the largest caves are amalgamations of chamber collections. The caves fit the *ramiform* and *spongework* classifications of A. Palmer (1991), but with specific asymmetries related to their position at the lens margin. First, the chambers nearest the lens margin are usually the largest (Fig. 4.4a). To the rear of these chambers are smaller passages that head inward to what was once the lens interior. As the original chamber enlarged, marine water invaded the chamber and the mixing front moved to the back, inland wall of the chamber as a series of individual fresh-water discharge points and hence mixing locations. These passages grew headward into the lens. The passages may have cross-connected to create a maze-like set of passages, but they all end in blank bedrock walls (Fig. 4.4b). These passage terminations are the position of the mixing front when sea level fell and the caves became senescent. If the lens was stable for a sufficient time, a second set of chambers may develop, parallel to the original set of chambers and to the lens margin, but interior to the initial row of chambers. Figure 4.5a shows this idea conceptually, Fig. 4.5b, c shows how it actually appears in a very large flank margin cave. Because the geochemically active portion of the lens margin does not penetrate deep into the island interior, flank margin caves can

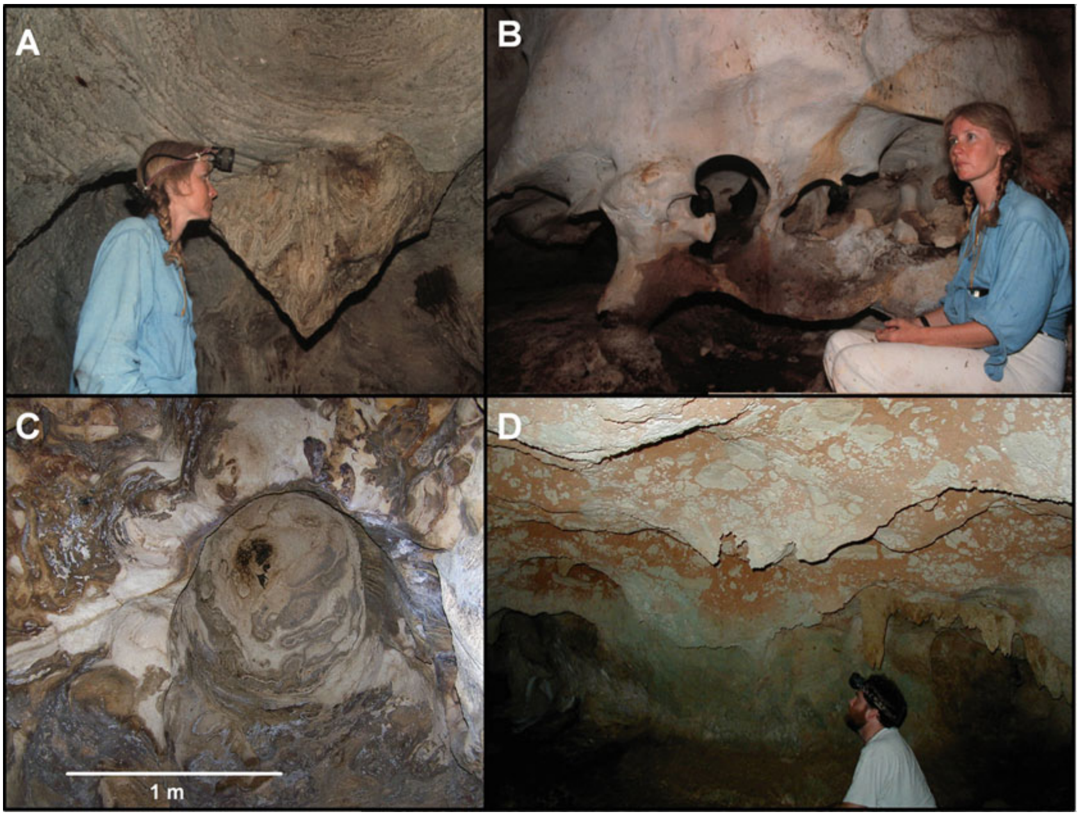

Fig. 4.3 Field examples of the power of dissolution in the fresh-water lens margin. (**a**) Smooth surface cut across both bedrock and flowstone, Hunts Cave, New Providence Island, Bahamas (indicating a phreatic-vadose-phreatic series of events). (**b**) Complex dissolutional structure, Harry Oakes Cave, New Providence, Bahamas. (**c**) Dissolutional surface cut smoothly through *Acropora Palmata* fossils and matrix, Suicide Cow Cave, Barbados. (**d**) Terra Rosa paleosol cut smoothly by dissolution, Red Roof Cave, Long Island, Bahamas

continue to grow in size only by amalgamating in a shore-parallel manner with other collections of chambers forming in the same way. Therefore the initially globular caves elongate along the lens margin by these connections (Fig. 4.5).

Flank margin caves contain a suite of phreatic dissolutional forms which all are consistent with slow flow. The features associated with turbulent stream flow, such as wall scallops or sediment bars and deposits, are absent in flank margin caves. The cave interiors are characterized by cuspate, curvilinear walls, isolated bedrock pillars, globular rooms, maze-like areas, and dead-end passages (Fig. 4.6). Another common feature are bell holes, vertical cylindrical holes in ceilings that can be meters high and up to a meter wide (Fig. 4.7). Their origin has been ascribed to three main options: condensation corrosion, bat activity, and convective phreatic flow. The debate has been significant, and the entire topic is discussed in Birmingham et al. (2010). We support the convective phreatic flow hypothesis and consider bell holes diagnostic of a diffuse flow regime that would not shear vertical convection cells in phreatic water.

Bermuda is presented in Fig. 4.2 as a carbonate cover island; however Bermuda has few flank margin caves in an environment that should have supported such cave development. Vacher and Mylroie (1991) and Mylroie et al. (1995a) explain this absence of flank margin caves as the result of the excess denudation the Bermuda eolianites have undergone. Comparing Bermuda to San Salvador, which are about the same size, are

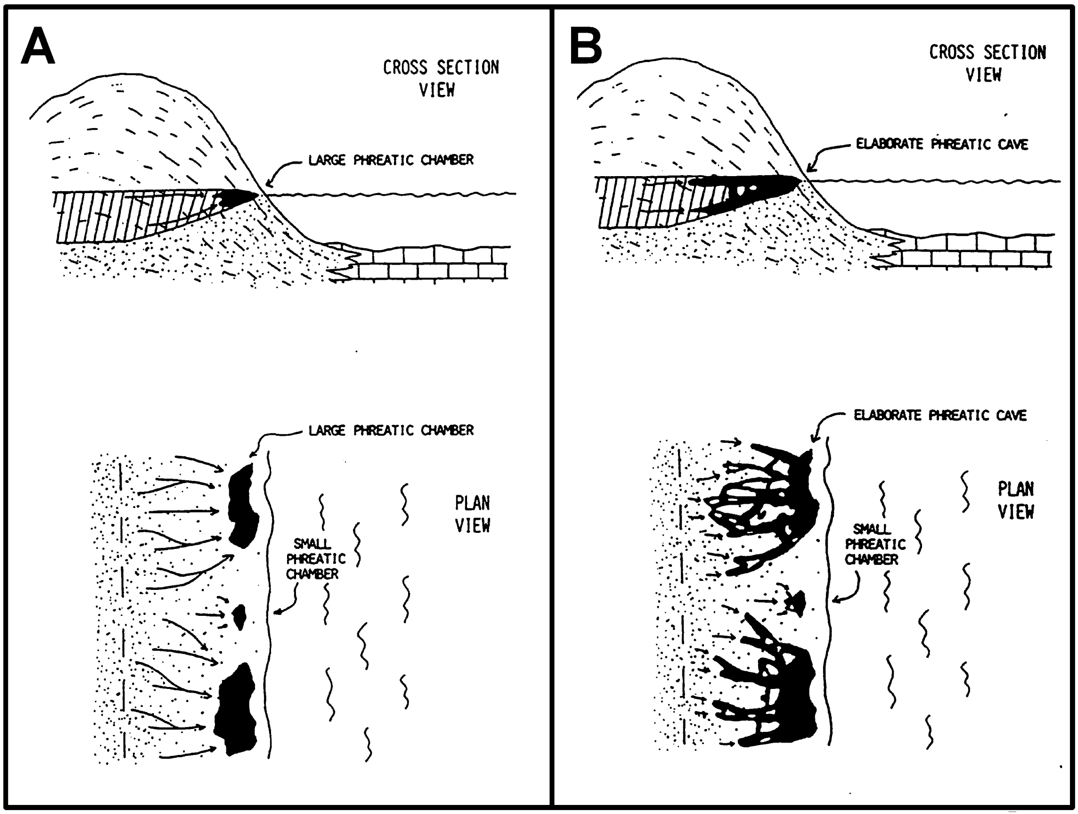

Fig. 4.4 Flank margin cave development. (**a**) Initial conditions, with chambers developing at the lens margin. (**b**) The fresh water lens retreats as cave porosity increases, moving the active mixing front to the back wall of the cave creating dead-end passages and maze-like interconnections (From Mylroie and Carew 1990)

dominated by Quaternary eolianites, and contain numerous interior water bodies, Mylroie et al. (1995a) argued that San Salvador Island is in a negative water budget environment, such that the interior water bodies are saline or hypersaline, and not capable of dissolving the lake coastlines. The negative water budget also retarded dissolutional denudation of the eolianites ridges, and the flank margin caves within them survive. On Bermuda, in contrast, there is a strongly positive water budget. The interior depressions are occupied by fresh water, unless they have been breached to the sea. During earlier sea-level highstands, more of the interior depressions were occupied by fresh water, which enlarged the lake coastlines by dissolution until those interior coastlines breached to the sea and were flooded by marine water. The eolian ridges underwent a higher degree of denudation, and existing flank margin caves were effectively removed. The large caves for which Bermuda is famous either formed as conduit caves when the platform was exposed by glacioeustasy, as in the Bahamas, or by progradational collapse of stream caves perched on the volcanic edifice that supports Bermuda. Other climatic implications for fresh-water lens configuration are explored in Mylroie and Carew 1995.

There are some caves in the Bahamas that fit the flank margin model almost perfectly in all aspects except for the elevation of where they are found, which is 10 m or more above current sea level. Osprey cave on Crooked Island, and Port Howe Sea Cave on Cat Island are examples of caves above 10 m. The higher passages in Hatchet Bay Cave, Eleuthera are also above 10 m elevation. But the crowning example is St. Francis Grotto, at 55 m elevation on Mount Alvernia,

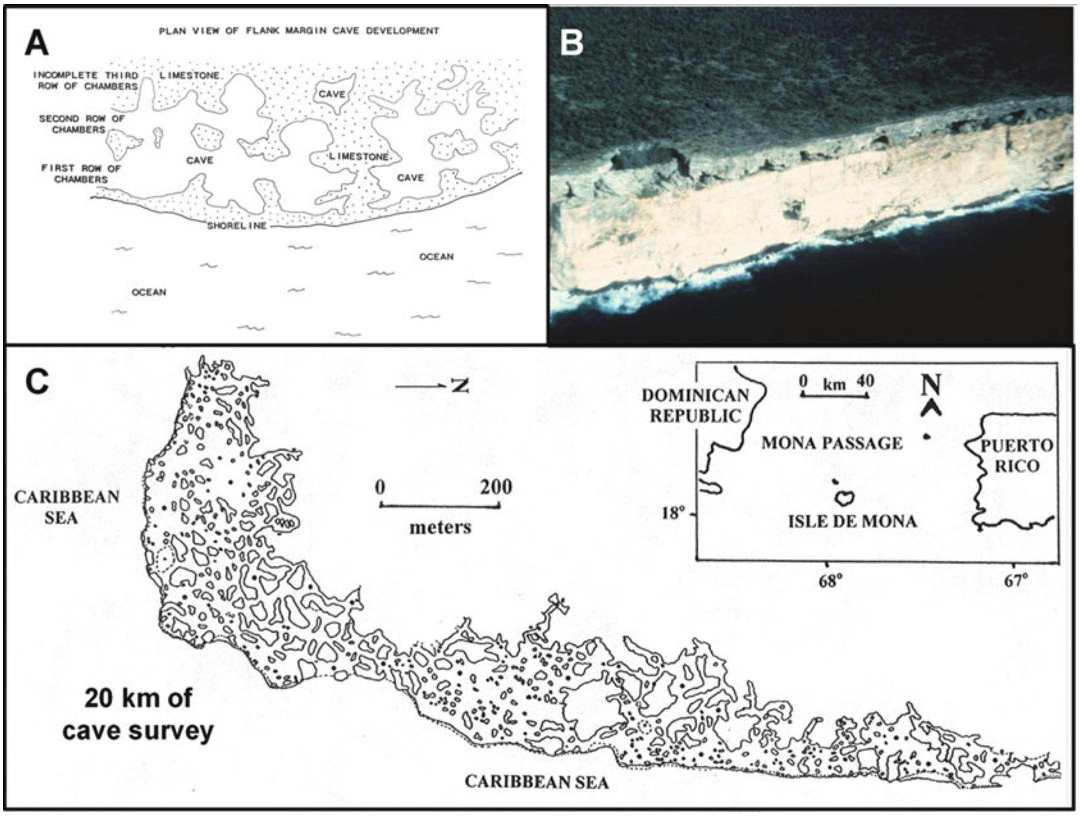

Fig. 4.5 The landward migration of the mixing front noted in Fig. 4.4b can occur as a step function. (**a**) Diagrammatic representation of shore-parallel cave chambers, with progressively less chamber linkage inland. (**b**) Coast of Isla de Mona, Puerto Rico, showing numerous cave entrances to Sistema Faro exposed by sea-cliff retreat. (**c**) Map of Sistema Faro (Cartography by M. Ohms, Isla de Mona Project), showing shore-parallel rows of passages and chambers. Note that the cave wraps around the easternmost margin of the island, and despite almost 20 km of mapped passage, extends inland less than 300 m

Cat Island (the highest point in the Bahamas at 63 m). The caves found at the +10 m elevation could perhaps be explained by accepting one of the higher values argued for the MIS 11 sea-level highstand about 440 ka (e.g. Hearty et al. 1999). But St Francis Grotto on Mount Alvernia is itself higher than any other topographic elevation in the rest of Bahamian Archipelago. Mylroie et al. (2006) attempted to explain the cave by the banana hole model of Harris et al. 1995, speculating that a thick terra rossa paleosol had perched a small water table, and the cave resulted from vadose and phreatic fresh-water mixing. As noted below in the banana hole section, the Harris et al. 1995 model may no longer apply. Cursory field work on upper Mount Alvernia did not locate the proposed perching paleosol either. The St Francis Grotto has classic phreatic dissolutional features, low, ovoid and globular rooms, and all the features described for flank margin caves. It has no pit cave features or morphologies. While most of the flank margin caves in the Bahamas found above +6 m elevation can be explained by proposing a sea level a few meters above the +6 m of MIS 5e, it would require melting almost all the ice on the planet to get sea level to +55 m elevation on Mount Alvernia in a tectonically stable environment like the Bahamas. These higher elevation caves remain a mystery, and perhaps a major thorn in the side of the flank margin model, which otherwise seems to explain phreatic coastal caves in the Bahamas so well.

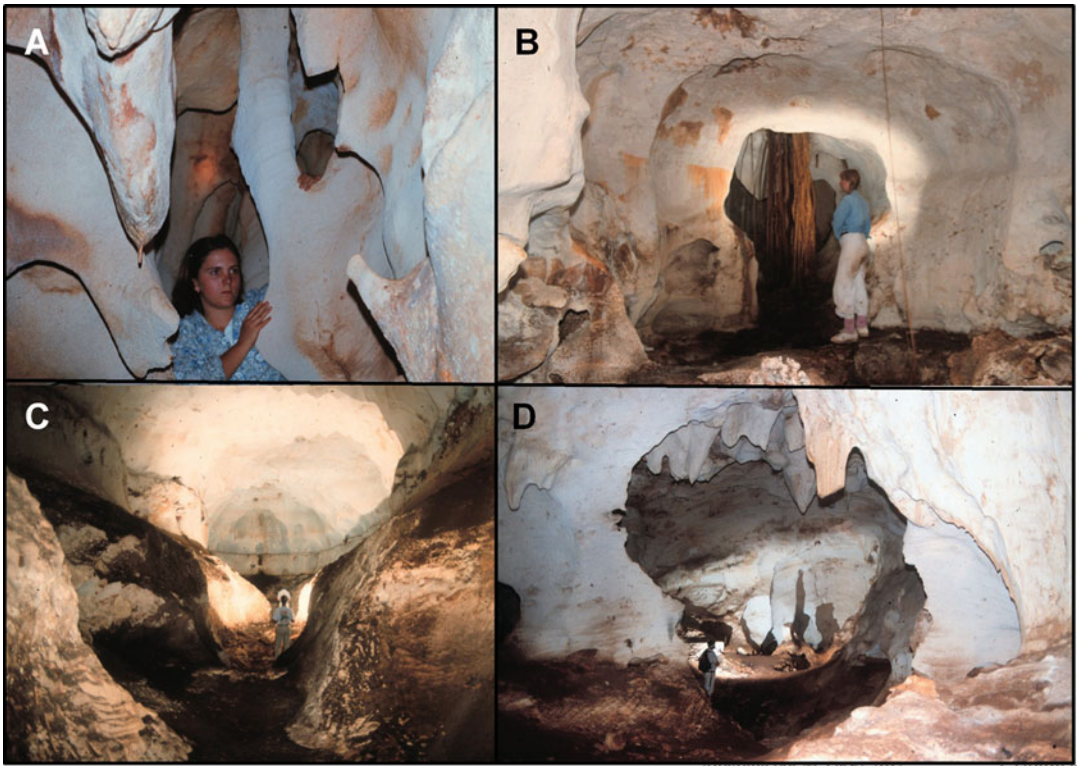

Fig. 4.6 Flank margin cave phreatic morphologies from the Bahamas. (**a**) Dissolutional sculpture cutting dipping eolianite beds, Lighthouse Cave, San Salvador. (**b**) Cuspate passage formed by linking globular chambers, Harry Oakes Cave, New Providence. (**c**) Large phreatic tube ending in a blank bedrock wall, Salt Pond Cave, Long Island. The blank wall indicates the position of the mixing front when sea level fell and the cave was abandoned by the fresh-water lens. (**d**) Large chamber, Hamilton's Cave, Long Island. The cave formed during the last interglacial (MIS 5e), a time span of 9,000 years

4.3.2 Telogenetic Flank Margin Caves

The eogenetic limestones common to many tropical islands and coasts allow significant matrix flow, especially as touching vug permeability develops (Vacher and Mylroie 2002). This eogenetic character has been presented within the CIKM as the reason for the mazy, ramiform and spongework shape of flank margin caves, and their origin as intersecting chambers and chamber collections within a coastal diffuse groundwater flow field. However, what happens when telogenetic rocks, with their almost non-existent matrix porosity, are placed in coastal situations? Shortly after the first presentation of the flank margin model in 1988 (Mylroie 1988), Proctor (1988) applied the model to Devonian limestones of Devon, southwest England, and indicated joints and bedding planes dominated the morphology of the resultant caves. Work in two disparate coastal areas with telogenetic carbonates, the Adriatic islands of Croatia and New Zealand (Chap. 17), showed that flank margin caves with passage characteristics of eogenetic limestones could be found in areas where the telogenetic rock offered a multitude of flow pathways, and in that manner imitated the flow pathways in eogenetic carbonates.

In Croatia, dense, hard Upper Cretaceous carbonates are locally mantled by a breccia interpreted to be a paleotalus. Otoničar et al. (2010) discovered flank margin caves preferentially developed in such paleotalus deposits in coastal settings, whereas adjacent Upper Cretaceous limestones had no caves (Chap. 17).

Fig. 4.7 Bell holes in the ceiling of Jumbey Hole, Acklins Island, Bahamas, looking straight up. Note the cylindrical shape, and the large height to width ratio

They hypothesized that the many flow pathways offered by the breccia allowed mixing to occur over a volume, as opposed to just along joints, fractures and bedding planes, and the resultant caves are typical in plan and morphology as those found in eogenetic island settings. The clasts in the paleotalus are Upper Cretaceous and telogenetic, but their re-deposition as a paleotalus created an appreciable diffuse flow field condition.

In New Zealand, Mylroie et al. (2008b) examined coastal carbonate outcrops along both North Island and South Island. Because of the extensive recent tectonism in New Zealand, Oligocene to Pliocene limestones found in the coastal setting have undergone deep burial and have been returned to the surface as dense, hard telogenetic limestones, despite their relatively young geologic age. In the Kaikoura area of northeastern South Island, intense deformation of the Paleocene Amuri Limestone has resulted in a highly fractured rock with cracks spaced ever few tens of centimeters (see Chap. 17). As a result, flank margin caves have formed with a morphology and pattern similar to that observed in eogenetic carbonates. On the west coast of South Island, near Punakaiki, the Oligocene Waitakere Limestone has a very dense joint spacing (Chap. 17); as a result, a series of small flank margin caves has formed along each joint. These voids are aligned horizontally, in accordance with the past fresh-water lens position, despite the dip of the beds being 20° to the southeast. In this case, the caves are not joined as in the first two telogenetic examples just presented, but the void abundance along a horizontal datum indicates a flank margin setting for cave origin.

The major difference between eogenetic and telogenetic limestones in coastal settings is the degree to which water flow can be spread out over a volume, and the degree to which it is restricted to fracture flow paths formed by joints, bedding planes, and faults. As seen in Devon, England, and Punakaiki South Island New Zealand, the lack of matrix flow creates cave voids spaced out along the fracture pattern, albeit with a horizontal datum control provided by the margin of the fresh-water lens at the time of void development. However, when the teleogenetic rock can be altered so as to create a closely spaced systems of openings (Croatian paleotalus) or fractures (Kaikoura fractured rock), then the flow mimics the diffuse flow regime of eogenetic rocks, and the flank margin caves that formed appear similar to those found in eogenetic carbonates.

On Mallorca, the largest of the Balearic islands in the coastal waters of Spain, Gines and Gines in Chap. 11 demonstrate the importance of facies control on cave pattern. In the reef facies, Vallegornera Cave is a series of globular chambers, but in the tight lagoonal mud facies, joint control is the only effective flow pathway and the cave pattern responds accordingly. In this case, eogenetic carbonate rocks, in the tight lagoonal facies, are imitating the lack of porous media flow routinely observed in telogenetic rocks.

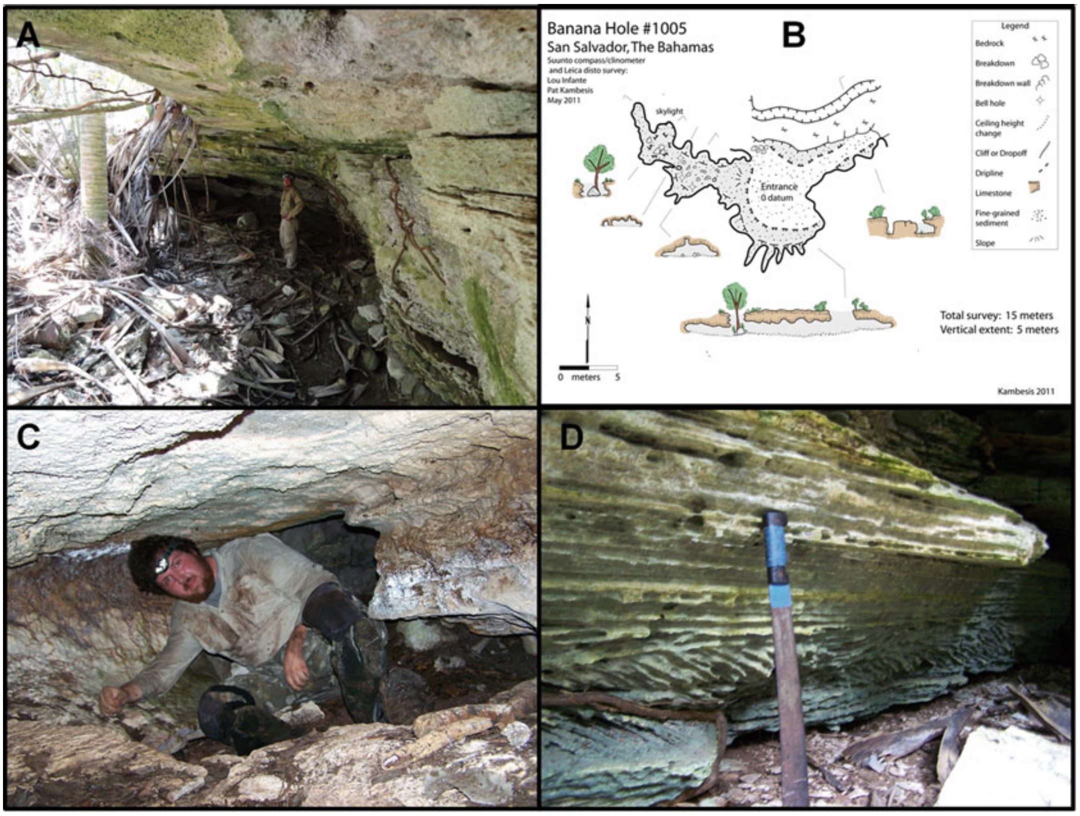

Fig. 4.8 Banana holes, San Salvador Island, Bahamas. (**a**) Large open banana hole. (**b**) Map of a typical banana hole, showing open and roofed sections. (**c**) Roofed section of a banana hole, with phreatic wall morphologies. (**d**) Wall rock of a banana hole, showing herring bone cross bedding, indicative of a subtidal depositional environment. The rock grades upward through beach to backbeach dune facies

4.3.3 Banana Holes

The Bahamas also contain another phreatic dissolution cave called the *banana hole*. The name is derived from native terminology for circular to oval bedrock openings in the land surface, meters to tens of meters across and a few meters deep (Fig. 4.8), that collect water and soil and are therefore excellent environments for growing specialty crops such as bananas (Harris et al. 1995). The surface openings commonly lead to an overhung chamber, the walls of which have phreatic dissolutional sculpturing. They are found in lowland plains in the Bahamas in large numbers. In many cases, there is no overhang and the walls are vertical to the surface. In other cases, a very small opening leads into an almost entirely intact cave. These features are usually a single oval to circular chamber, but connections to adjacent chambers, some which have completely intact ceilings, do occur. For many years the model proposed to explain banana holes (Harris et al. 1995) assumed vadose and phreatic fresh-water mixing at the top of the lens as the drive mechanism for dissolution. Given that most banana holes were far away from the current shoreline, the flank margin model using the discharging lens margin as the dissolutional mechanism was deemed inappropriate. The great abundance of banana holes in the Bahamas, and their relative scarcity in other carbonate settings, such as Isla de Mona or the Marianas Islands was attributed to the higher elevation of those islands, such that small voids at depth lacked the accommodation space to allow collapse to prograde tens of meters to the surface above. In the Bahamas, however, the banana holes are found in plains only 6–10 m above sea level, and the cave roofs are only a meter or two thick. These voids express quite easily by collapse. Because banana holes with intact roofs, partial roofs, and no roofs are all found in large numbers, their surface expression by collapse of a thin roof appears to be the appropriate mechanism.

Recently, several large-scale banana hole mapping projects in the Bahamas (Mylroie et al. 2008a; Infante et al. 2011) made an important observation: the banana holes were always located in subtidal facies (above modern sea level) that transitioned upward to beach, back beach and back-beach dune facies. In the Bahamas, subtidal facies above modern sea level are part of the Cockburn Town Member of the Grotto Beach Formation (Carew and Mylroie 1995, 1997), which correlates to the last interglacial sea-level highstand, MIS 5e approximately 124–115 ka (Thompson et al. 2011), which was ~6 m above modern sea level. These deposits were prograding strand plains during MIS 5e. The caves in these deposits must be younger than the rocks, but the only sea-level highstand that could have placed the fresh-water lens at the proper elevation (+6 m) was MIS 5e. In other words, the caves are *syndepositional*, forming in the rock immediately upon the rock's deposition (Mylroie and Mylroie 2009b). The banana holes are now seen as small, immature flank margin caves (Infante et al. 2011). They developed in the back-beach dune system shortly after the deposition of those carbonates during MIS 5e. As the strand plain continued prograding seaward (as a result of excess sediment supply), the fresh-water lens followed the advancing land, occupying the strand plain facies from subtidal units upward through the back-beach dune facies. As strand plain progradation continued, the initial flank margin caves were abandoned by the lens margin, and new ones began to form at the new lens margin position. With each stand plain progradational event, the sequence was repeated. Because the lens margin was migrating, void development was terminated at a given location quickly, and these immature flank margin cave chambers did not have time to become large, or to pervasively intersect their neighbors. This new hypothesis explains the occurrence of banana holes in the fossil strand plains, their distribution in what is today an inland setting, their development with thin roofs, and their small size (relative to regular flank margin caves), and their great abundance. This new interpretation avoids the controversy about the true dissolutional potential of the top of the fresh-water lens.

4.3.4 Pit Caves

Returning again to the Bahamas to exploit the time and space constraints available there, the discussion turns to pit caves. On many of the eolianite ridges of the Bahamas are vertical pits and shafts that extend from the surface downward as much as 10 m (Fig. 4.9). Many are above the position of past glacioeustatic sea-level highstands, which means they cannot be the result of dissolution in a past fresh-water lens. The caves contain vertical grooving and other indications of descending vadose flow. They may have a stair-step configuration. They rarely intersect flank margin caves at lower elevations. These caves are morphologically similar to classic vadose shafts that develop in the telogenetic karst areas of continental interiors (e.g. White 1988). Fieldwork (Harris et al. 1995) has shown that they can occur in large numbers in relatively small areas. This observation was initially thought to indicate that they could be a paleoclimatic indicator, and that their abundance was due to a higher rainfall in the past. However, Harris et al. (1995) studied these pit caves during major rainstorms, and were able to determine that they efficiently captured surface flow, but that they were also in competition with each other. The high abundance reflected shifting capture points over time, and abandonment of some pit caves, as opposed to concurrent pit cave development due to high precipitation. Pit cave complexes are found on other islands, for example, Isla de Mona (Chap. 9).

Pit caves are vadose fast flow routes that bypass the diffuse flow network and deliver water to depth in the carbonate outcrop quickly. On Guam,

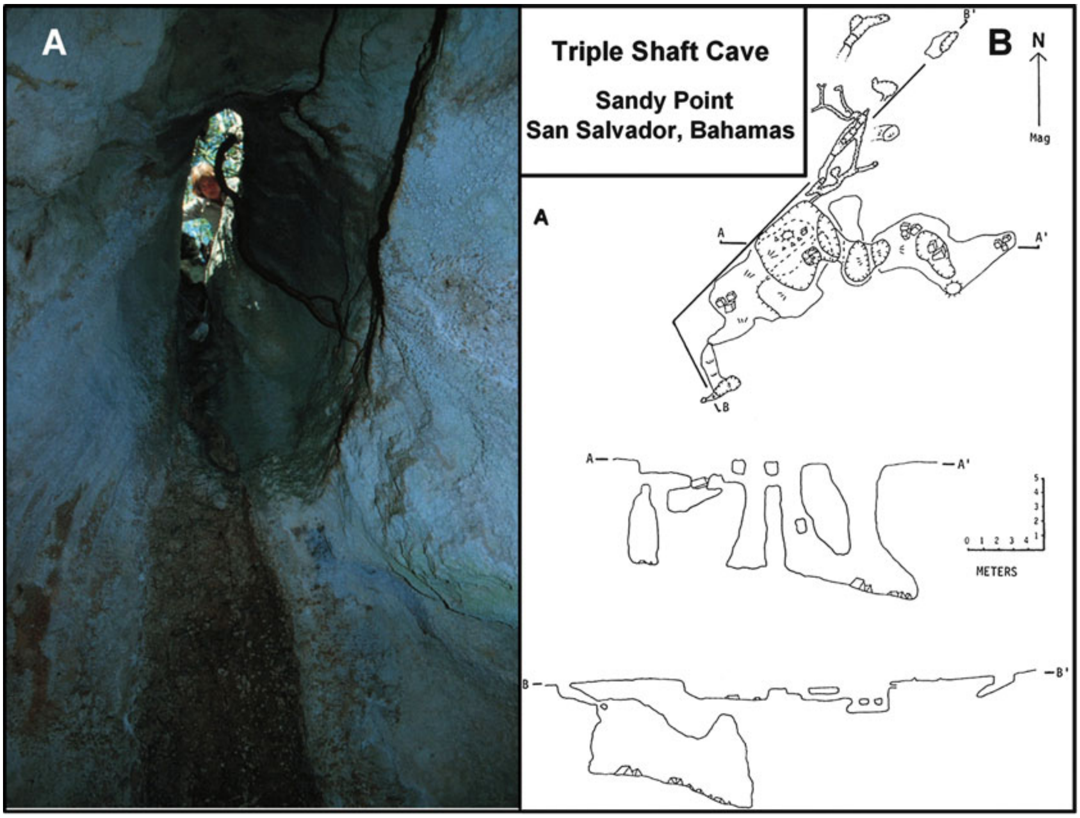

Fig. 4.9 Pit caves on San Salvador Island, Bahamas. (**a**) Typical pit cave morphology as a deep but constricted vadose pathway. (**b**) Map of Triple Shaft Cave, showing pit competition for recharge gathered on the epikarst

it has been estimated that 40 % of the recharge to the fresh-water lens is by these vadose fast-flow routes (Jocson et al. 2002); the other 60 % can take months to reach the lens by diffuse flow. These observations help explain why some wells respond quickly to meteoric events, while other wells do not. It depends on the proximity of the well to a vadose fast flow route, which can mound the top of the lens and affect a nearby well, but the mound tails off before reaching more distant wells.

Why do pit caves develop on young eogenetic carbonates, where diffuse flow rates are high? The Bahamian study of Harris et al. (1995) suggests that cementation and micritization of the exposed limestone surface by repeated wetting and drying events (similar to caliche formation) results in a calcrete crust that can perch some of the meteoric water which then exploits any weak points in that crust to create a pit cave. Even in the absence of a calcrete crust, precipitation events can apply water to the land surface faster than infiltration can accommodate the flow. Antecedent events can limit infiltration even during moderate precipitation events. In hindsight, overland flow should be expected, and flow would focus downward to any spot where some characteristic of the rock accepted greater flow amounts. The result is a pit cave. Pit caves tend to end in sandy floors. The degree to which the floor of the pit cave is the downward limit of dissolution, or merely a washed in debris pile is unknown. Pit floors have been excavated for a meter or two in a few locations, but a true bedrock floor was not reached. In Majors Cave, San Salvador Island, a pit cave on the ridge top was forced downward through small passages into the rear of the main chamber of the cave, one of the few places in the Bahamas where such a vertical connection can be made.

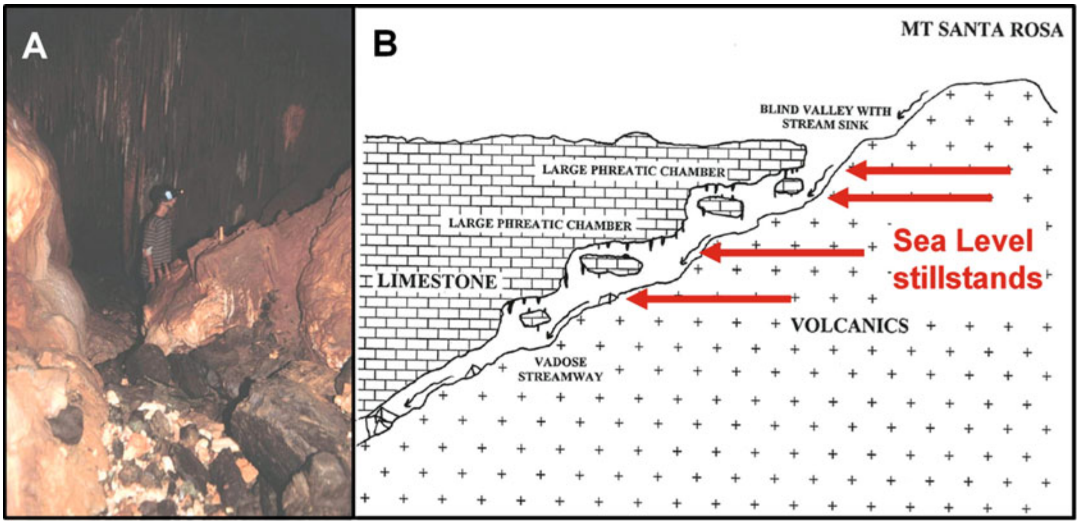

Fig. 4.10 Allogenic stream cave development, Guam, Marianas Archipelago. (**a**) Image of a cave passage developed at the limestone–basalt contact, Awesome Cave. (**b**) Cartoon cross section of Awesome Cave, showing stacked phreatic chambers undercut by vadose streamway. Each chamber represents a past fresh-water lens position

4.3.5 Allogenic Caves

Carbonate islands and coasts can form typical stream caves as found in continental interiors. Such caves are "karst on islands", as previously explained. However, in many cases, the development of a stream cave can occur in a manner consistent with "island karst". As shown in Fig. 4.2, when non-carbonate rock perches vadose flow, stream caves form. They can arise purely from autogenic catchment, as in carbonate-cover islands, or from allogenic catchment, as in composite or complex islands. In all cases, however, the opportunity may exist for the downstream end of these caves to end at the fresh-water lens, and create a vadose/phreatic mixing situation.

Because of glacioeustasy, during sea-level lowstands, such vadose stream caves can form downward to the new lens position. When an interglacial occurs, sea level and the fresh-water lens will rise, flooding the formally vadose passages. If the non-carbonate layer never reaches the coastline above any past lens position, then the discharge from the stream cave will always be into the lens. In some respects, the downstream end of the stream cave can be said to be hydraulically dammed, as it empties into a porous media aquifer (especially for eogenetic carbonates). As result, back flooding would be expected during high discharge events, as observed in Coles Cave on Barbados (Machel et al. 2011; Chap. 10), and large chambers should form in the adjacent lens as a result of water mixing. High organic loading would be expected in a vadose stream cave gathering allogenic recharge as in composite and complex island settings.

For tectonically uplifted islands, even at a glacioeustatic sea-level highstand, as is the case today, uplift should have brought the cave chambers associated with the vadose stream caves and fresh-water lens boundary up into the vadose zone. This situation is observed in Awesome Cave on Guam, Mariana Islands (Fig. 4.10), where a series of stacked, globular phreatic chambers extend away from the stream cave at past lens positions (Jenson et al. 2006; Chap. 13). On Saipan, Mariana Islands, the juxtaposition of normal faults and low permeability volcaniclastics has created hydrologic compartments where confined conditions occur (Chap. 13). The fresh water drained the compartment by flowing upwards and over the confining barriers. Uplift later drained the compartment from below, and an entirely phreatic cave can be observed today, with a 30 m high phreatic lift tube that drained a converging series of passages within the compartment below.

4.3.6 Closed Depressions

On continents, one of the most distinctive karst landforms is the closed depression, drained internally by karst hydrologic flow. These depressions range from small sinkholes or dolines up to depressions kilometers across known as poljes. On eogenetic carbonate islands, closed depressions also exist in a variety of sizes and shapes, however their origin is commonly different. Because of sea level change, the existing island topography may be in part constructional, meaning that closed depression may be the result of uneven deposition of carbonates, and not by subsequent removal through dissolution by internal karst flow. In eogenetic carbonates, the limestone may be porous and permeable enough to drain karst depressions, or such drainage may be facilitated by pit caves, sinking streams, and other traditional karst pathways. Constructional closed depressions are known from continental settings, most commonly in glaciated areas where rock basins are scoured out in limestones, or where valleys are dammed by glacial sediments and water flow escapes through limestones in the valley wall (e.g. as in New York, Mylroie and Mylroie 2004).

Two common constructional basins in eogenetic carbonate islands are raised lagoons, and swales between eolianite ridges. In these cases, the closed depression is maintained by karst processes, but was generated by differential limestone deposition. Large closed depressions, especially in very young limestone islands such as the Bahamas, are constructional. There hasn't been enough time between the deposition of the limestone, its exposure to the subaerial environment, and the present to allow the large amounts of carbonate dissolution necessary to create a large depression. The same time constraints also limit the development of true dissolution sinkholes or dolines. Combined with the porous and permeable nature of eogenetic limestones, surface flow does not result in the classic bowl or conical sinkhole shape. The development of a hard calcrete crust promotes the development of steep-sided pit caves instead. On continents, cover-collapse sinkholes, produce a sinkhole by subsidence of insoluble overburden downward through a small dissolution pipe or dissolutionally widened crack. Such action rarely occurs in young eogenetic limestones as there is little insoluble cover to be piped downward. As islands mature over time, and capture insolubles as dust fall, ash fall, or from adjacent non-carbonate terrain, the ability to form cover-collapse sinkholes improves.

The least common sinkhole in continental settings is the cave collapse sinkhole, produced by failure of the dissolutional void down inside the bedrock. On young carbonate islands, however, cave collapse sinkholes are the most common sinkhole type. In the Bahamas, where banana holes have formed in great abundance with very thin roofs, their expression as collapse features is extremely common. Flank margin caves, because they form on the flank of the landmass enclosing the fresh-water lens, are also prone to collapse, and many flank margin caves have numerous collapse entrances. Because the position of banana holes and flank margin caves are very predictable, the risk posed by their abundance can be minimized by knowing where unexpressed collapse-prone voids are likely to exist (See Chap. 6 for additional discussion on modeling void development).

4.3.7 Blue Holes

Blue holes are intriguing explorational and scientific sites (Fig. 4.11). In island karst, the term "blue hole" has a fairly specific meaning, but for karst in general, the meaning is more general. Essentially, the term is descriptive, referring to a water-filled hole that has a deep bluish cast or color, commonly associated with the Bahama Islands, but also used elsewhere, such as Belize (Dill 1977). The term has also been used to describe artesian springs in karst settings unrelated to islands (Zans 1951; Sweeting 1973), such

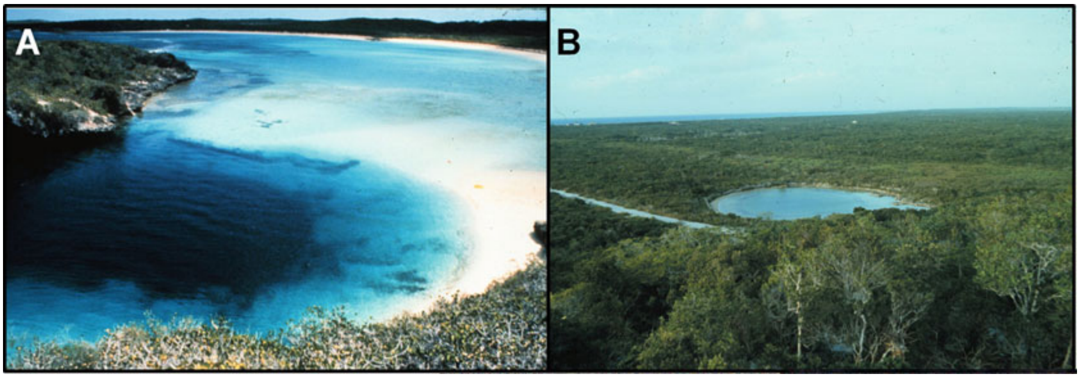

Fig. 4.11 Blue hole images. (**a**) Dean's Blue Hole, Long Island, Bahamas, an ocean hole with a connection to the sea. The blue hole is 200 m deep. (**b**) Watling's Blue Hole, San Salvador Island, Bahamas, an inland blue hole isolated from the open ocean

as Turner's Blue Hole in Kentucky, a typical continental interior karst spring (Mylroie and Mylroie 1991). In addition, features of similar morphology, that contain water, such as the cenotes of the Yucatan Peninsula, Mexico, are interpreted by some authors to be a form of blue hole (Ford and Williams 1989). The term blue hole proliferated in the last decades of the twentieth century because of more scientific research on karst environments in the Bahamas.

The Bahamas have been the focus of blue hole studies. Shaw (1993) reports the first blue hole reference in the literature to be by Catesby in 1725, who stated that local inhabitants in the Bahamas used the word "pit" to describe deep holes filled with water that fluctuated with the tide. The term blue hole appeared on the British Admiralty charts of the Bahamas in 1843 and 1844, to describe water-filled depressions and holes differentiated from the surroundings by a deep blue color (Shaw 1993). Nelson (1853) reported on fluctuating water levels in deep holes. Shaw (1993) also reports that Northrup in 1890 indicated the native Bahamians used the term "ocean hole", and that those holes demonstrating inflow and outflow were "boiling holes". Agassiz (1893) used the term blue hole in conjunction with the term ocean hole, to describe deep holes in the floor of Bahamian lagoons and banks. He gave no other definition or description. Shattuck and Miller (1905) used the terms ocean hole and blue hole to describe deep holes found on the banks whose sides flare-out beneath the opening and which usually have a constant circulation of water. The water in some of the blue holes they described responded in harmony with tidal fluctuations. Doran (1913) referred to blue hole as a local name given to ocean holes. Stoddart (1962) describes a blue hole as a deep depression in the floor of the lagoon which is perfectly round and reputedly bottomless. Stoddart preferred to use the term blue hole rather than the term ocean hole because of the deep blue color associated with the features. While obvious features of the Bahamas, blue holes did not receive much early scientific study (Sealey 1991). In-depth discussions of blue holes eventually became routine in textbooks about Bahamian geology and geomorphology (e.g. Sealey 1994).

Blue holes were popularized by a series of articles by cave diver George Benjamin (1970) in National Geographic magazine. The discovery of large underwater cave systems inside blue holes resulted in a wave of cave diver exploration, followed by scientific inquiry, much of it detailed in popular books on the subject (e.g. R Palmer 1985, 1989). A review of this period can be found in Mylroie et al. (1995b), who proposed a series of definitions to stabilize the varied use of the term "blue hole" in the literature. Their blue hole definition is: "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

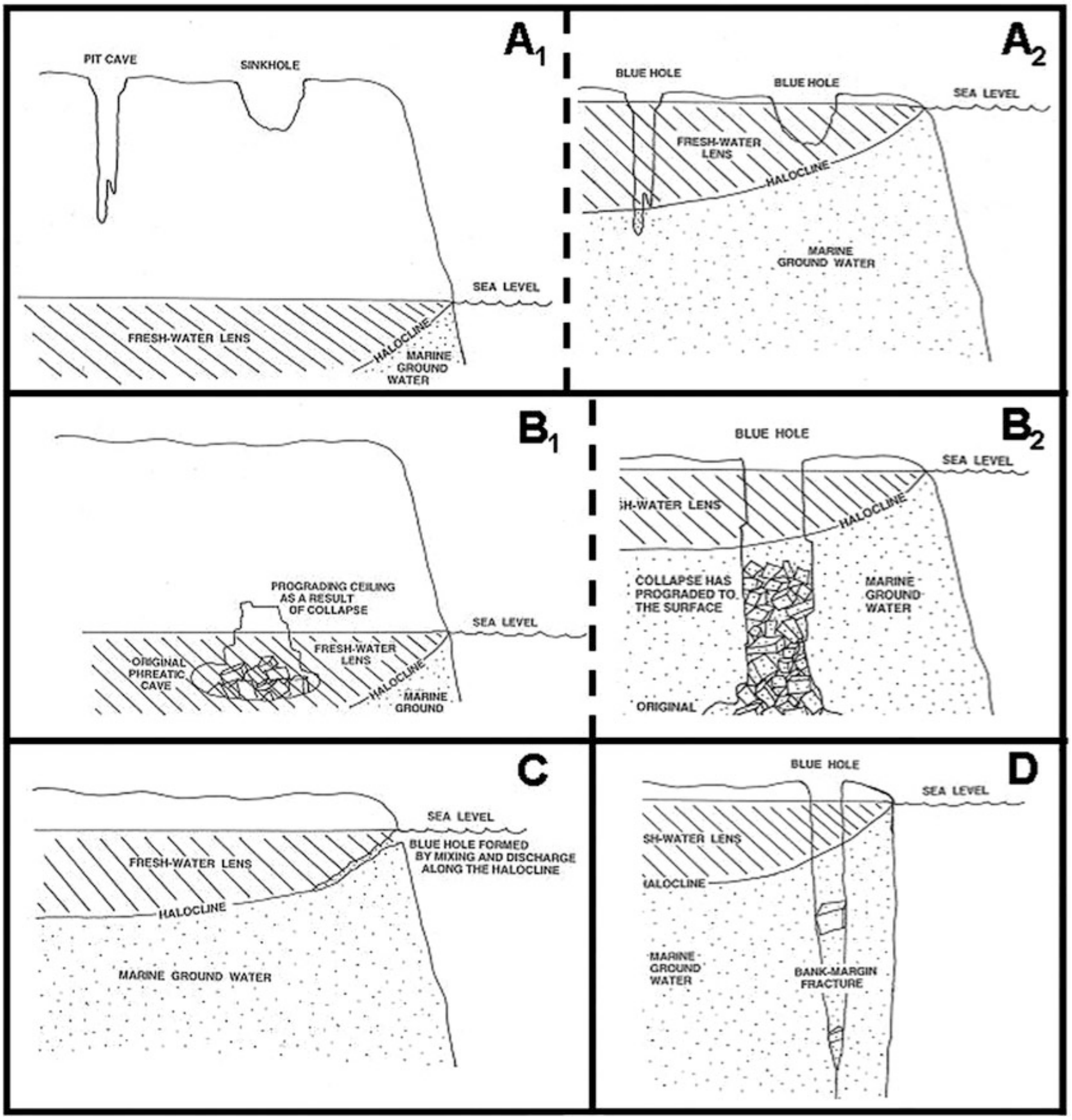

Fig. 4.12 Blue hole origin. (**a**) Vadose features such a pit caves and sinkholes can become small blue holes if sea level floods them. (**b**) Most blue holes result from progradational collapse of deep voids, perhaps originating as large conduits. (**c**) Downward trending passage following the halocline; the trend is misleading, a result of vertical exaggeration in making the figure. (**d**) Blue hole development by bank margin failure

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 were further characterized by their surface connectivity to the open ocean: "An ocean hole opens directly into a lagoon or the ocean, is usually tidally influenced, and contains marine water. An inland blue hole opens directly onto the land surface, or into an isolated pond or lake, may be tidally influenced, and may contain a variety of water chemistries from fresh to marine." (Mylroie et al. 1995b, p. 230–231). See Fig. 4.11 for examples.

The origin of blue holes is believed to be polygenetic, as presented in Fig. 4.12 (Mylroie et al. 1995b). Because pit caves can form independently of sea level, requiring only the presence of a vadose zone, it is likely that pit caves have formed at elevations below modern sea level. With sea-level rise to modern position, these pit caves would be flooded and would appear as blue holes (Fig. 4.12a). However, their size, both in diameter and in depth, would be somewhat limited based on the size of pit caves available for observation above modern sea level today, and this mechanism cannot be used to explain the vast number of blue holes that are larger in diameter and depth than known pit caves.

It was also proposed (R Palmer and Williams 1984) that blue holes could develop by following the mixing zone or halocline downward as the fresh-water lens thickened inland (Fig. 4.12c). This interpretation was based on the observation that in descending underwater caves in the Bahamas, divers could follow the halocline to significant depth. Therefore, it was reasoned, mixing dissolution along the descending halocline had created the passage that curved upward to become a blue hole. This proposed explanation has two flaws. First, the steep incline of the fresh-water lens, as shown in Fig. 4.12c, is not real. In textbooks and in scientific papers, the fresh-water lens is commonly presented with significant vertical exaggeration. An island 10 km across might have a lens only 10 m thick, an aspect ratio of 1,000. Dissolution along the halocline would not result in a steeply descending cave. Second, the observation of the halocline within the descending caves is real, but cause and effect are reversed. The cave represents an easy flow path for the lens, and so captures the lens flow. The halocline is steeply descending because of the pre-existing cave; the halocline did not cause the cave in this case. Some other mechanism must have been at work to make the cave.

One important mechanism of blue hole formation is bank margin failure (Fig. 4.12d). The Bahama Banks have very steep margins, which are gravitationally unstable and prone to vertical failure (e.g. Mullins and Hine 1989). Quaternary glacioeustatic sea-level change, which subaerially exposes the upper 100 m+ of the bank wall, might promote gravitational failure as a result of the loss of buoyant support. Bank-parallel vertical fissures are produced which can lead to water both at low and high sea-level positions. Small graben-like structures form (Carew et al. 1998) as material subsides parallel to the facture walls. In the Bahamas, blue holes form as a series of openings along the fracture trace; see R Palmer (1989) for a full review. Because of the length of these fractures, they can form a large number of blue holes, but these blue holes are restricted to the bank margin area. The fractures can trace from lagoons to dry land, producing both ocean holes and inland blue holes. Many of these blue holes can be linked up underwater along the fracture trace. These fractures also act as interceptors of the flow within the fresh-water lens towards the coast, creating unique hydrological conditions (Smart and Whitaker 1997).

The most common blue hole is the progradational collapse structure, sometimes referred to as an Aston collapse (R Palmer and Williams 1984). As presented in Fig. 4.12b, these blue holes originate as deep-seated voids large enough to accommodate the collapse material as ceiling failure progrades its way upward. If enough collapse occurs, and the collapse debris can be accommodated at depth, then the collapse will reach the surface. If, as shown in Fig. 4.12b, the accommodating void has active geochemical processes at work, the collapse material can be dissolved and carried away, maintaining accommodation space as the collapse progrades upward. As noted earlier, there is abundant evidence of a well-developed fossil conduit flow system at depth in the Bahamas. The conduits associated with the maximum glacioeustatic sea-level depression of the Quaternary would be at the 100–125 m depth, where the time to reverse glacioeustatic sea-level fall to a sea-level rise would provide the time window for large conduit systems to develop. As with any large conduit system, collapses would occur at irregular sites along the conduit where site-specific conditions created roof instability (for a review of cavern breakdown processes, see White 1988). These collapse blue holes can also intersect former conduits at a variety of elevations. Current research (e.g. Larson and Mylroie 2012) is now modeling blue hole abundances on the Bahama Banks to determine if patterns will appear that reflect conduit traces in the subsurface.

In the Bahamas, voids have been located by exploratory drilling at a variety of depths. Meyerhoff and Hatten (1974) report voids at depths of 21–4,082 m. The deepest of these voids was able to accept 2,430 m of broken drill pipe, indicating an extremely large chamber. Such a large chamber, if at a shallower depth, could easily accept sufficient collapse material to allow collapse to prograde all the way to the surface. Deans Blue Hole, on Long Island, Bahamas, is over 200 m deep; arguably the deepest blue hole in the world (Fig. 4.11a). The deepest 15 m of that blue hole is an extremely large chamber (Wilson 1994). The vertical migration of these voids by collapse allows blue holes to integrate conduit cave systems at a variety of depths. As a consequence, blue holes are considered one of the most diverse habitats for anchialine ecosystems (Mylroie and Mylroie 2011).

As noted for flank margin caves, biogeochemistry has been implicated in the development of the dissolutional environment for void development. Blue holes can receive significant organic loading, especially inland blue holes that can accept woody vegetative material. The speleogenetic impact of organic decay in blue holes was first reported by Bottrell et al. (1991). Subsequent arguments have been made that make biogeochemistry the primary cause of blue hole formation (e.g. Schwabe and Carew 2006). It is important to remember that major organic loading of a blue hole cannot occur until it expresses itself at the earth's surface. While DOC and particulate organics may play a role in dissolutional production and modification of voids at depth, large scale organic activity must await opening to the surface.

Blue hole diving, as with any cave diving, can be extremely hazardous. The inflow and outflow of water from ocean holes in response to tidal cycles is particularly dangerous. Many blue holes contain elaborate vadose speleothems (stalactites, stalagmites, flowstone, etc.) deposited during glacioeustatic sea-level lowstands when the caves were in the vadose zone and air-filled. In addition to minerals, abundant fossil and archeological remains are also present (R Palmer 1989). In Sawmill Sink blue hole on Abaco, fossil material includes a bat fossilized in flowstone at 30 m depth, a former owl roost with thousands of small vertebrate bones at 25 m depth, and incredible preservation of crocodiles and tortoises on a talus cone within the anoxic mixing zone at shallower depth (Steadman et al. 2007).

4.4 Modeling the Chronology of Coastal Cave Development

The development of banana holes as syndepositional caves implies a very rapid rate of cave development. Supporting evidence is found in flank margin caves reported in the Holocene Merizo Limestone of Guam, where the Pacific mid-Holocene sea-level highstand first emplaced the rocks above modern sea level. Tectonic uplift made the outcrop subaerial such that a fresh-water lens could form and create the caves (Miklavič et al. 2012). As sea level dropped to its current position 2 m below the Mid-Holocene highstand, the caves were drained and became accessible. Submarine work on San Salvador Island, Bahamas (Mylroie and Mylroie 2007) located flank margin caves at specific depths of 105–125 m. The survey began at a depth of 60 m, because at shallower depths Holocene coral growth masked the original platform wall. In the depth from 60 m down to 105 m, however, no caves were found. The interpretation is that during the major glacioeustatic sea-level excursions of the Quaternary, sea level, and the fresh-water lens associated with it, was not at a vertical position long enough to leave a macroscopic dissolution signature in the platform wall. When the sea-level excursion bottomed out, the time necessary to stop and begin the sea-level return upward was long enough that dissolution could occur. The depths of 105–125 m match well with the four major troughs located on the Mid to Late Quaternary sea-level curve, Chap. 1, Fig. 1.5. The flank margin caves found subaerially today in the Bahamas are also the result of a sea-level turn around, in this case the peak turn around as opposed to the trough turn around found by submarine.

As demonstrated by the banana holes, the time window to form a flank margin cave is indeed small, but how long do such caves persist in the rock record? Isla de Mona, Puerto Rico has some of the largest known flank margin caves in the world. The island has been tectonically uplifted, and many of the caves exist in the coastal cliffs at elevations up to 70 m above the sea (Fig. 4.5). This position has saved them from over-printing by later glacioeustatic sea-level events. The key to their great size is related to their age. Paleomagenetic analysis of a sequence of sediment, flowstone and stalagmites in Cueva de Aleman demonstrated the caves to be at least two million years old (Panuska et al. 1998). These data indicate that the caves can persist for millions of years. The age of the caves also places them far enough back in time that glacioeustasy had not yet become the high amplitude, short wavelength observed from the Mid to Late Quaternary. Sea level, and the fresh-water lens, were in a fixed, stable position for a long period of time, allowing the flank margin cave dissolutional environment to act continuously and produce the large, extensive rows of chambers seen there (e.g. Fig. 4.5b, c).

In northern Spain, work by Baceta et al. (2007) has described flank margin caves preserved as infilled voids in Danian (lowest Paleocene, ~66–63 million years ago) limestones of the Urbasa-Andia plateau. The caves are in carbonate facies that correlate with the edge of the platform as regression occurred and the platform became subaerial, allowing a fresh-water lens to form. These observations indicate that flank margin cave signatures can survive deep burial and well into the geologic past. Flank margin caves form rapidly, yet persist through time. That makes the caves a high resolution yet durable sea-level indicator; forming in a few thousand years but surviving for millions of years and further illustrate the interrelated processes of cave development within the context of evolving coastal landforms.

4.5 Coastal Landform Configuration and Cave Development

In the Bahamas are many flank margin caves of appreciable size (Fig. 4.6). As noted earlier, this dissolution has been quite rapid, given the volume of the cave chambers and the relatively short glacioeustatic sea-level highstands necessary to elevate the fresh-water lens and produce them. The Bahamian flank margin caves are developed mostly in eolian calcarenite ridges (fossilized carbonate sand dunes). These eolianite ridges come in all sizes from a few meters high to over 60 m high. They can be kilometers long. During MIS 5e, they were long, linear islands with a fresh-water lens; a "strip island" in the sense of Vacher (1988). Somewhat surprising is the development of flank margin caves during MIS 5e in what would have been very small islands or cays. In some cases, the eolianite paleo-island would have been only a 100 m long and 30 m wide, or less, and only 10–15 m high, yet the ridges contain flank margin caves with linear dimensions in the meters to tens of meters (Walker et al. 2008). Even very small fresh-water lenses are able to exploit the flank margin condition to produce caves.

The development of flank margin caves is in part controlled by the configuration of the shoreline. Shoreline shape controls how the distal portion of the fresh-water lens discharges to the sea. The control of cave development by the lens margin position causes the flank margin caves to faithfully follow the lens margin as it curves in conjunction with the shape of the island coastline (Fig. 4.5c). In the eolianite ridges of the Bahamas, each ridge terminates at the two linear ends of the ridge. These ridges have the appearance of a doubly-plunging anticline, and field workers casually call the ends of the ridges "noses" in a similar manner. Flank margin caves found in the vicinity of the ridge terminations have a unique appearance, and have been informally called "nose caves" (Fig. 4.13). At the ridge nose the active portions of the lens from each side of the island begin to overlap. If the ridge is

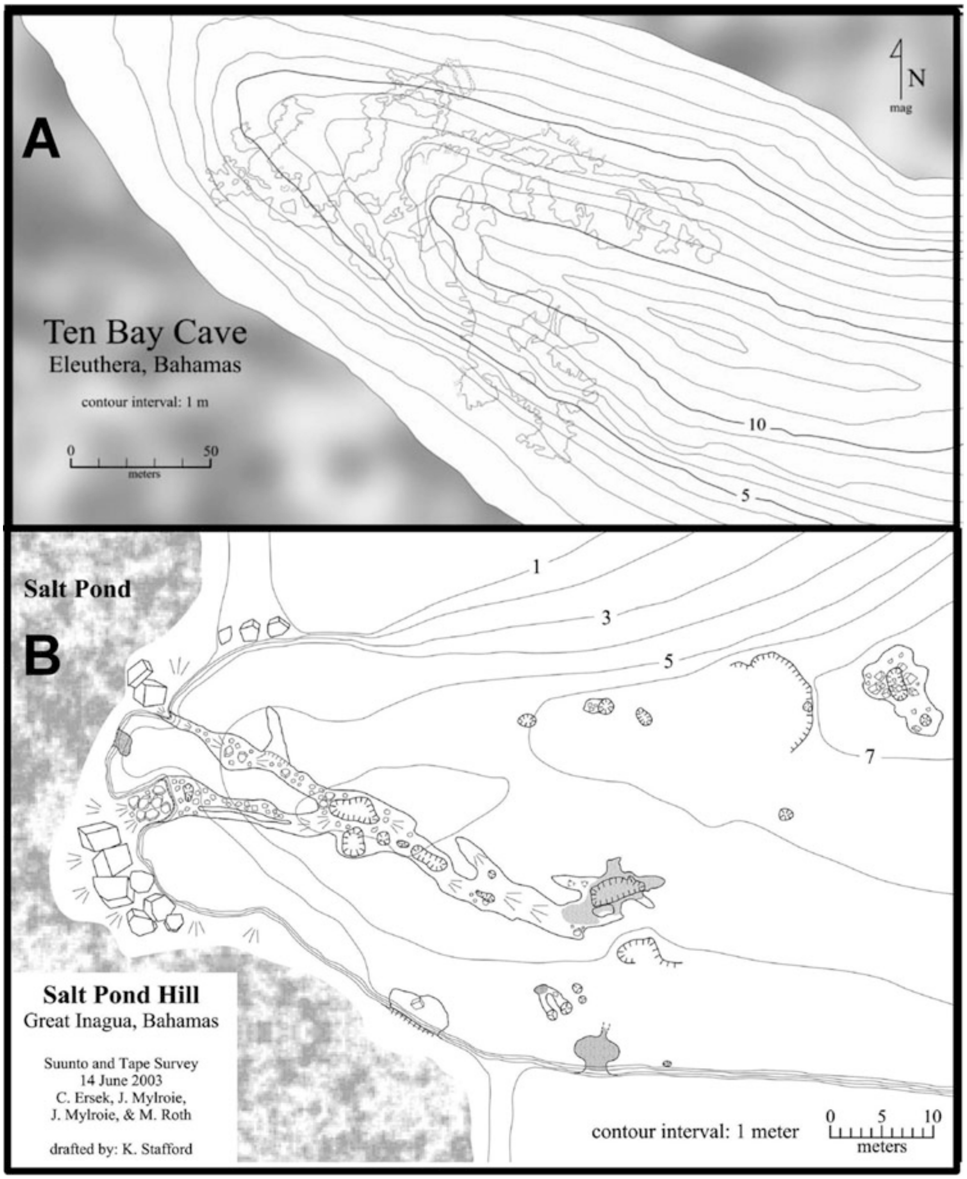

Fig. 4.13 Flank margin cave development at the end of Bahamian eolianite ridges, so-called "nose caves". (**a**) Ten Bay Cave, Eleuthera Island, wraps around the end of the dune and because the dune is wide enough the lens margin and resulting cave passages are separated. (**b**) Salt Pond Cave, Great Inagua Island, in a small dune ridge which overlapped the lens margin, creating a single passage down the dune axis

small, the overlap results in a single linear cave along the terminal axis of the ridge (Fig. 4.13b). If the ridge is large, the caves wrap around the nose of the ridge, cross-linking where the ridge is narrow, but as the ridge widens away from the nose, the cave separates along each flank of the ridge (Fig. 4.13a). Nose caves are another demonstration of the control the lens margin has on cave development, and the relative lack of dissolutional potential within the body of the fresh-water lens, where flow is slow and mixing does not occur.

Stream caves and linear conduits are not known in the Bahamas in the subaerial position today. However, cave divers report long, linear phreatic tubes, thousands of meters in length and meters wide and high at depths of 20–30 m in the Great Bahamas Bank (e.g. R. Palmer 1984). The openings to these caves were not visible on the platform edge during the submarine study as Holocene corals had obscured them. It has been argued (Mylroie and Vacher 1999) that the development of these cave passages is the result of sea-level position. Under current sea-level conditions, the islands on the Bahama Banks are relatively small, and highly linear (Chap. 7). This condition would be even more exaggerated during MIS 5e when more of the Bahamas Banks were flooded and most of the dry flank margin caves observed today developed. If sea level were only 20 m lower than today, however, almost the entire Bahama Banks would be subaerially exposed. The many islands of various sizes in existence today would combine to a few very large islands; small islands would exist only on the isolated small platforms (such as San Salvador Island). As the banks became exposed, their area would greatly increase with respect to their perimeter, as the former increases by the square but the latter does not. To discharge the collected meteoric water from the fresh-water lens by diffuse flow would become progressively less efficient, and at some area to perimeter ratio, conduit flow would be supportable. Cave divers have not located such conduit systems on the small platforms, as sea level fall does not increase their area to perimeter ratio by a significant degree. On large banks, flank margin caves could still develop, as water budget constraints would keep conduits widely spaced, and in the coastal locations between conduits, diffuse flow to create a lens margin would still exist. In the Akumal area of Quintana Roo in Mexico, flank margin caves from MIS 5e are found in the coastal eolianites, even as 100 km long conduit flow systems discharge below them from the interior of the Yucatan Peninsula (Kelley et al. 2006). The Yucatan, although it is part of a continental setting, is a possible analogue for behavior of the larger Bahamian banks when sea level was low enough to expose the bank tops (see Chap. 16).

4.6 Modeling the Interplay of Geomorphic Controls and Karst Processes in Littoral Settings

On Barbados (Machel et al. 2011), Kangaroo Island, Australia (Mylroie and Mylroie 2009a), Mallorca (Mylroie et al. 2012) and Curaçao, coastlines and paleocoastlines with deep inlets have flank margin caves developed within those inlet walls. Two models have been proposed to explain this cave distribution (Mylroie et al. 2012). One considers that expressions of flank margin cave development may exist several hundred meters inland from the active coast. Such a distance of cave development can be seen in Sistema Faro, Isla de Mona (Fig. 4.5c). If a field of cave voids was present, then as an inlet or incision was made into the coastline, by marine or fluvial processes, some of these caves would be exposed. The second model requires only a minor inland range of flank margin cave development. It suggests that after a valley or inlet has been incised into a coastline, a rise in sea level will result in marine invasion of the inlet, creating the flank margin speleogenetic environment in the walls of that coastal incision. After sea level falls, the caves are then exposed by slope retreat as happens on the open coast. The two models express a "chicken versus egg" situation; which came first, the caves or the inlet?

On Kangaroo Island, the inlets created a preservational bias situation in the eolian calcarenites (Mylroie and Mylroie 2009a; Chap. 18). The open coast fronts on the stormy, high wave amplitude Southern Ocean. As a result, the eolianites that could host flank margin caves have been stripped back in Holocene time, and many caves likely removed. Within the inlets, wave energy is low or non-existent as the inlets shoal to dry land, and numerous flank margin caves are preserved in the inlet walls. On Barbados (Chap. 10), work is currently ongoing to assess the inlet models in the development of flank margin caves found in the walls of steep-sided canyons called *gullies*. In the Mariana Islands, flank margin caves are commonly found ringing embayments into the coastline. Again, a chicken versus egg situation exists; are the flank margin caves there because the embayment places the lens margin along the embayment coast, or did flank margin cave development weaken the coast such that an embayment easily formed (Stafford et al. 2004)? Coastal configuration control by discharging ground water has been described for the Yucatan coast (see Chap. 16), creating inlets called *caletas* (Back et al. 1984).

Flank margin caves contain a record of past coastline configurations and conditions. For banana holes, they demonstrate a prograding strand plain. For nose caves, they show that marine water surrounded the ridge. In areas where the paleo-coastline is not well known, flank margin caves can clearly indicate where the coastline had to be for the caves to develop the configuration they present.

4.7 Flank Margin Cave Morphometrics

Flank margin caves, as noted earlier, develop without entrances. After abandonment by the fresh-water lens, usually due to sea level fall, the caves are breached by surface erosion. This breaching is most commonly from the side, but collapses or intersection by a pit cave (see below) can occur. Initially, slope retreat creates a tangential intersection of the cave. As time passes, and slope retreat continues, the cave will progressively disintegrate. Flank margin caves in all stages of erosional removal can be found (Fig. 4.14). The degree of removal is important, as it is a measure of denudation rate. On Tinian Island, Stafford et al. (2005) measured cave entrance widths and maximum cave interior

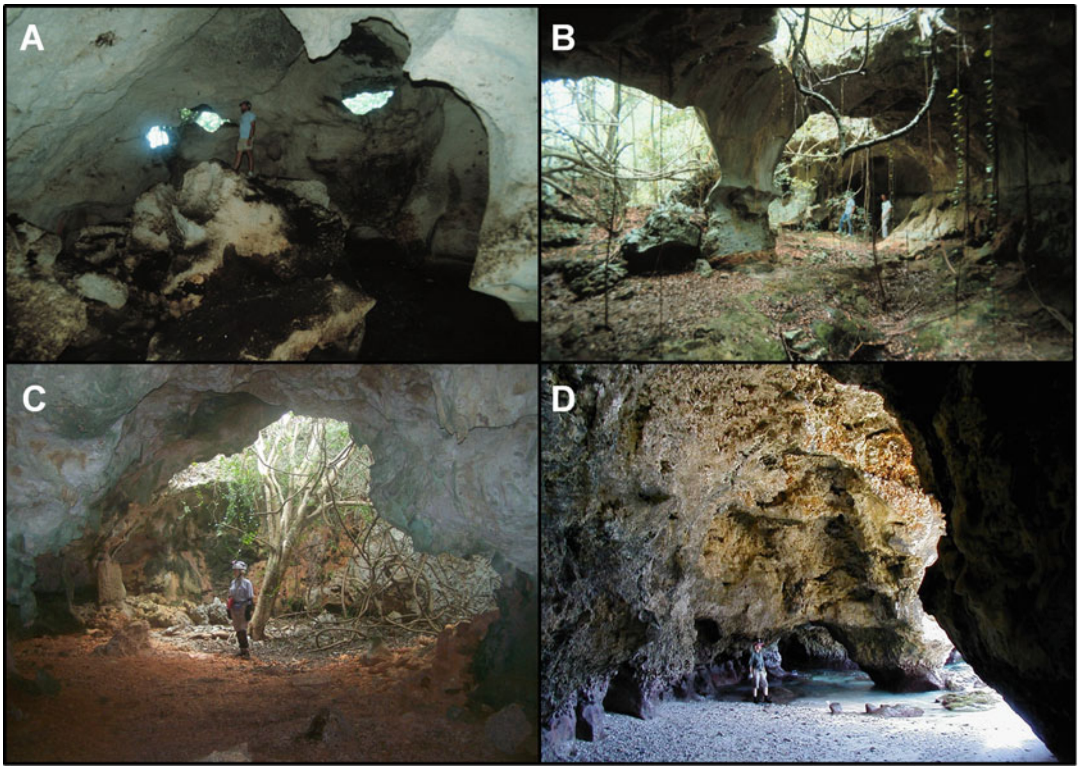

Fig. 4.14 Surface denudation intersects flank margin caves. (**a**) Caves Point West Cave, New Providence Island, Bahamas, displaying tangential surface intersection barely opening the cave chamber. (**b**) Harry Oakes Cave, New Providence Island, Bahamas, where slope retreat has removed almost all of the cave's outer wall. (**c**) Cave on Curaçao, where roof collapse has intersected the cave from above. (**d**) Faaligosey Cave, Fais Island, Federated States of Micronesia, where wave action has removed the outer wall of a flank margin cave

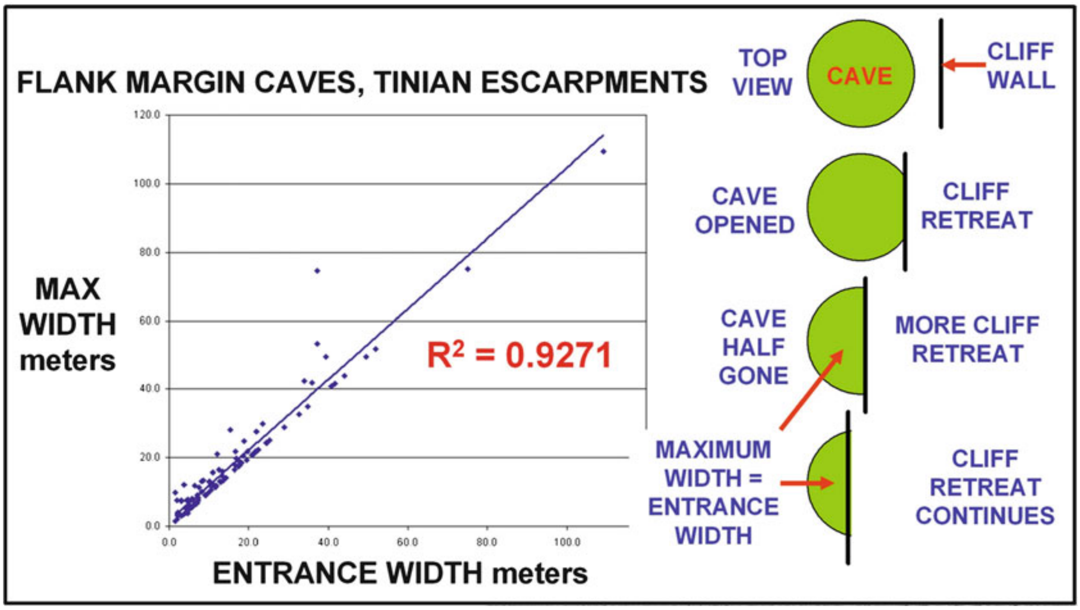

Fig. 4.15 Plot of maximum cave width versus entrance width for flank margin caves exposed in the cliffs on Tinian, Marianas Archipelago. Most caves fall on the 45° line, indicating the caves are 50 % or more eroded away, as shown diagrammatically at the far right of the figure

widths of flank margin caves found on the uplifted cliffs of the island. When these data were plotted (Fig. 4.15), they showed that the vast majority of the caves had an entrance width that equaled their maximum width. In other words, simplifying the cave chambers as simple flattened spheres, this equality between maximum width and entrance width required the cave to be half, or more than half, eroded away. Note that a few caves fall in the upper left of the graph, these are caves that are only partially breached, and they are few in number. It is not possible to have an entrance width greater than the maximum width, so no caves fall in the lower right part of the graph.

A series of other morphometric approaches have been used with flank margin caves. The simplest approach is a rank order plot of cave sizes. Cave size is itself somewhat argumentative, and for flank margin caves the convention is to use the areal footprint of the cave as seen on a high-quality cave map (see Mylroie 2007 for a full discussion of the issue). When first done for a suite of Bahamian flank margin caves (Roth et al. 2006), the resultant plot produced three straight line segments, each of which corresponded to a specific size range of caves (Fig. 4.16). When growth of flank margin caves in the distal margin of the lens was modeled on a computer (Fig. 4.17), the resulting rank order plot matched the empirical data set almost exactly (Labourdette et al. 2007). The computer model also displayed a fourth line segment at the very small end, representing caves so small they were not mapped. This lack of small cave data was an explorational bias of the survey teams, which did not bother to map very small caves. These size rank data support the growth model for flank margin caves as a series of small initial dissolution voids that enlarge and through time eventually intersect one another. As time proceeds, collections of chambers intersect. Each of these intersection events results in a jump in cave size, producing the individual straight line segments seen on the plot.

Another morphometric is the area to perimeter ratio. This approach takes advantage of the maze-like complexity of flank margin caves (Roth 2004), which produces a plot quite different from that created by linear stream caves (Fig. 4.18). Stream caves plot similar to an

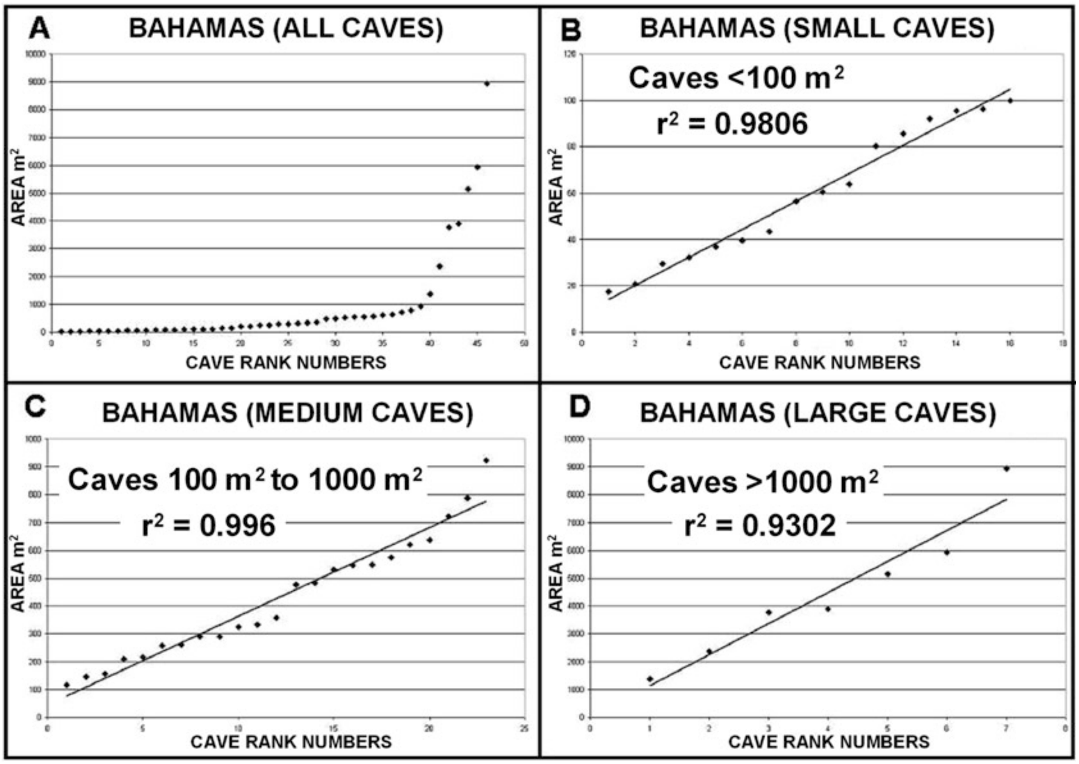

Fig. 4.16 Rank order plot of Bahamian flank margin caves, using areal footprint as a measure of cave size (From Roth et al. 2006). (**a**) All caves ranked, showing three segments of different slope. (**b**) Plot of caves under 100 m^2 area. (**c**) Plot of caves between 100 m^2 and 1,000 m^2 area. (**d**) Plot of caves over 1,000 m^2

extended rectangle with an aspect ratio of 1–100. Flank margin caves plot more as a 1–10 aspect ratio rectangle, a reflection of their globular, ramiform nature. However, as flank margin caves grow larger, they can only do so by connecting laterally along the lens margin; inward growth is limited. As a result, as Fig. 4.18 shows, the largest flank margin caves develop a linearity as a result of a switch from growth in all planar directions to a linear growth parallel to the lens margin. As with the area rank order plot, this pattern reinforces the model of flank margin cave growth as a function of chamber intersections.

The area versus perimeter (A vs P) approach was used by Lace (2008) to differentiate sea caves from flank margin caves in Puerto Rico. Waterstrat et al. (2010) did the same for San Salvador Island, Bahamas. The Bahamian approach utilized sea caves developed in Holocene eolianites as a control for sea cave pattern, as these eolianites could not host a flank margin cave from a Pleistocene sea-level event. Flank margin caves in Pleistocene eolianites, entered solely by roof collapse or pit caves were used as a control for the flank margin cave pattern, as wave energy could not carve a sea cave from above. The paper presented plots that indicated the A vs P ratio could separate sea caves from flank margin caves, but the statistical approach later proved incorrect (Curl and Mixon 2011), and when the correct approach was used, the statistical differentiation did not work (Waterstrat et al. 2011). The original argument that tafoni could be distinguished from sea caves and flank margin caves by their A vs P ratio did prove to be correct (see Chap. 8 for a discussion of tafoni).

The differentiation of sea caves from flank margin caves is not especially useful in sea-level studies, as the caves both form at sea level. However, differentiating a sea cave from a flank

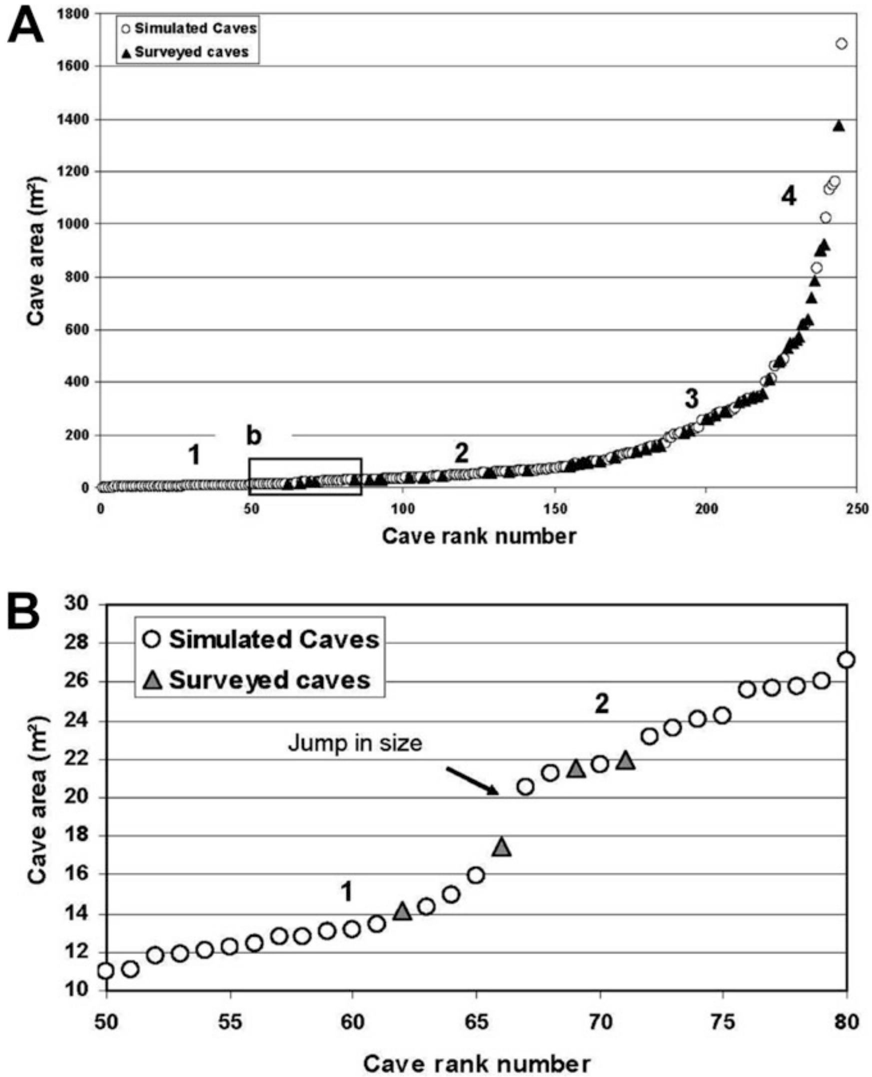

Fig. 4.17 A computer model was used to generate flank margin caves by the chamber amalgamation method. (**a**) The plot fits the empirical data of Roth et al. (2006) almost exactly, and (**b**) revealed a smaller subset of cave size, under 20 m^2, which cave mappers had ignored as too small (From Labourdette et al. 2007)

margin cave is critical for calculation of denudation rates based on degree of cave destruction (Waterstrat et al. 2010). Sea caves form from the outside inward, and are always open to the surface. Flank margin caves, on the other hand, form on the inside and are only enterable once they have been breached. The degree of denudation that could entirely remove a sea cave would only be enough to begin the opening of a flank margin cave (Fig. 4.19). A fossil sea cave incorrectly identified as a dissolutional cave would overestimate the amount of denudation; a dissolutional cave wrongly identified as a sea cave would underestimate the denudation.

Denudation has also been measured by measuring limestone pedestal heights where the limestone is protected by a cap rock. A common feature in glaciated areas, where a non-carbonate glacial erratic can protect the limestone beneath (Ford and Williams 2007), such features, called *karrentisch* (Miklavič et al. 2012), have been found on Guam in the Mariana Islands. In this setting, the cap rock is the Plio-Pleistocene Mariana Limestone, which has fallen from a cliff over

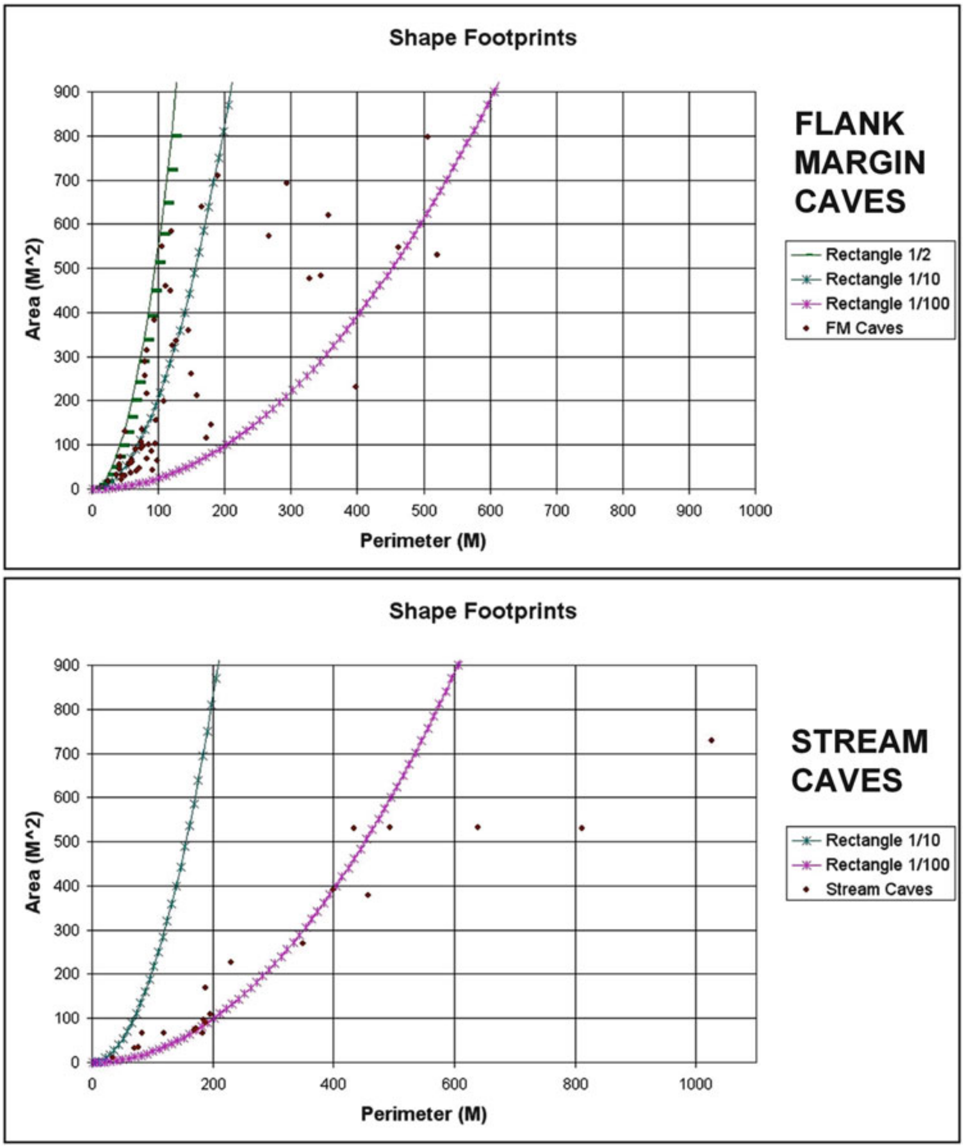

Fig. 4.18 Area vs Perimeter plots for flank margin caves and stream caves show that each cave type plots into different fields. The fields converge at large cave sizes, as flank margin caves can become large only by coast-parallel growth (see Fig. 4.5), which imposes a linearity on the globular cave pattern (From Roth et al. 2006)

100 m high on to a plain of Tarague Limestone, formed during the last interglacial (MIS 5e) sea-level highstand. Pedestals as much as 5 m high exist. Because the Tarague Limestone is so young, a minimum value for the denudation rate (assuming cap rock emplacement just after sub-aerial exposure about 114 ka) can be calculated.

4.8 Summary

Carbonate islands and coasts contain a variety of cave and karst features not found in the vastly more studied continental interiors of the world. The differences result from three major

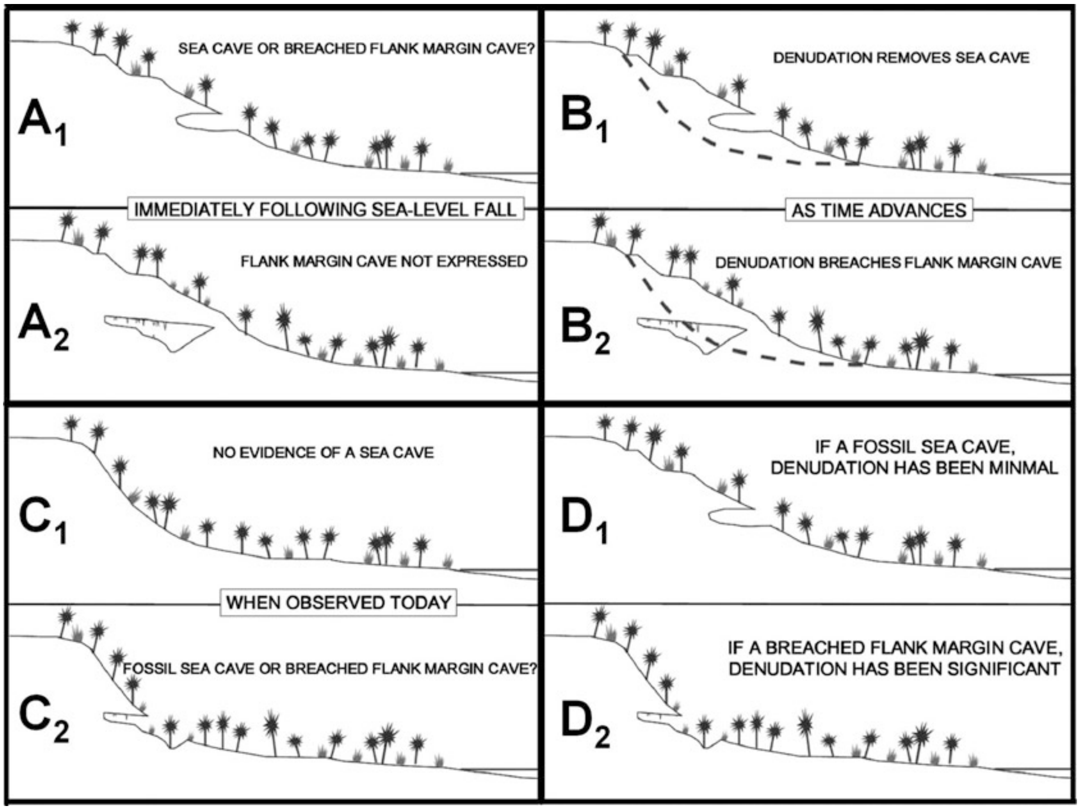

Fig. 4.19 Cartoon demonstrating the importance of differentiating relict flank margin caves from relict sea caves when assessing landscape denudation (From Waterstrat et al. 2010). (**a**) Initial condition. (**b**) Following denudation. (**c**) Denudation consequences. (**d**) Subsequent interpretations

influences: fresh water mixing with salt water, sea level and hence base level change, and commonly the young, eogenetic nature of the carbonate rocks and the continual reshaping of coastal landforms by littoral processes. The Carbonate Island Karst Model attempts to take into account the relationship of sea level, carbonate rocks, and non-carbonate rocks to explain the variety of caves and karst features found on islands, by creating four categories of island types, which can be discrete or form a continuum. The most common dissolutional cave in the world may well be the flank margin cave, as it has been found in great abundance on carbonate islands and continental coasts worldwide. Because sea-level change results in abandonment of old flank margin caves, and the production of new ones, the numbers of such caves are extremely high.

The study of island karst has many useful and practical applications. As shown by Labourdette et al. (2007), preservation of coastal caves in the rock record is of extreme interest to the hydrocarbon community. Carbonate island water resources are better managed when the CIKM can be applied to aquifer studies (e.g. Jocson et al. 2002; Mylroie et al. 2008c). The recreational caving community plays an important role in mapping of these caves (Mylroie 2007). The historical and archeological aspects of these caves are important (Chap. 5), as is the paleontology (Steadman et al. 2007). The complex relationship between carbonate deposition, sea level, and dissolution is an important contributor to our understanding of past Quaternary climate change, and by extrapolation, possible future consequences for society (Mylroie 2008).

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