Chapter 6

GEOLOGY OF MUD ISLANDS IN FLORIDA BAY

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INTRODUCTION

Florida Bay is a large triangular body of water located to the south of peninsular Florida and to the north of the limestone keys discussed in Chapter 5. Some 237 Holocene carbonate mud islands occupy about 1.73% (1,500 km²) of the total area of the bay (Enos, 1989). Despite having “Key” attached to their names, the mud islands of Florida Bay are composed of fine-grained, unconsolidated carbonate muds that have accumulated over the last 4,000 years as Florida Bay was flooded by rising sea level (Scholl, 1966; Davies, 1980). These islands, therefore, are distinct from the better-known Florida Keys to the south. Similar mud islands exist in other parts of the world, although they are not common and are limited to low-energy coastal areas lacking siliciclastic input. Examples occur along the west coast of Andros Island (Bahamas), Belize, and on both the northern and southern coasts of Cuba (Bathurst, 1971; Ginsburg and James, 1974).

The mud islands were used by the Calusa Indians while fishing and hunting within Florida Bay (Tabeau, 1968). None of the islands appear to have been used as a permanent camp; there are no shell mounds that typically mark Indian establishments on islands further to the north and along the Florida Keys. Settlement of the area began during the early 1900s by plume hunters, fishermen, and loggers. On some of the higher islands, charcoal burners set up camps and cut down the larger hardwood trees (Craighead, 1964). Attempts to cultivate tomatoes and other vegetables were periodically made on islands in the northwest portion of the bay, but most of these tiny establishments were short-lived and swept away by hurricanes. Today, the islands serve as important nesting and foraging habitats for a variety of birds, reptiles, and aquatic invertebrates.

Several features of the Florida Bay mud islands stand out hydrochemically and separate them from other carbonate islands. Their extremely low elevation (less than 50 cm above MSL) and small size (generally about 0.5 km²) allow seawater to flood over them during lunar tides and storms. As a result of this periodic flooding, shallow ponds collect on the interior of the islands and are subsequently evaporated. Consequently, both pond waters and groundwaters have salinities of 45–140 g kg⁻¹ in spite of the large annual rainfall. In addition, some of the older islands have sediment layers composed of up to 30% authigenic dolomite. Diagenetic alteration, therefore, has occurred, and it appears to be related to an island-specific process (Videlock, 1983; Swart et al., 1989a).

In this chapter, we review the geology and hydrology of Florida Bay and the mud islands. We will focus on the geochemistry and diagenesis using data collected from
earlier (Swart et al., 1989a,b; Burns and Swart, 1992) and ongoing studies (Kramer et al., 1993, 1994; Juster, 1995; Juster et al., in press). These studies, which make use of groundwater chemistry and stable isotope data, indicate that significant recrystallization is taking place in these sediments, but the data do not conclusively prove that dolomite is forming at the present time. Either rates of dolomitization are too slow to be properly documented by groundwater chemistry, or the dolomite is a relict from an earlier time in the island’s history.

REGIONAL SETTING

Geography of Florida Bay

Florida Bay is a shallow lagoon (depth < 4 m) located immediately below the southern tip of peninsular Florida (Fig. 6-1), approximately 80 km southwest of Miami. Bordered on the north by the Everglades and to the south and east by the Florida Keys, the majority of Florida Bay lies within the boundaries of Everglades National Park. The mud islands of Florida Bay are generally located atop or along the maze-like “mudbanks” which dissect Florida Bay into a series of “basins” (Fig. 6-1). “Mudbank” indicates a mound of unconsolidated carbonate mud that is exposed during low tides; “basin” refers to the large, relatively deeper water areas (>2 meters) amongst the mudbanks and islands. “Groundwater” as used in this chapter refers to the water in the saturated zone of these exposed carbonate sediments; this water is generally called “porewater” in the literature of diagenesis of these islands, which reflects the sedimentological context and marine-geochemical approach of the studies.

The Holocene sediments of Florida Bay lie unconformably on limestone bedrock, the late Pleistocene Miami Limestone (Hoffmeister et al., 1967) (Fig. 6-2A, B). The gentle slope of the Miami Limestone from east to west has allowed for thinner, narrow mudbanks in the east and thicker, wider mudbanks in the west (Fig. 6-2B). Most of the mudbanks and mud islands in the bay are positioned over karstic depressions and irregularities (30–200 cm in relief) within the Miami Limestone (Wanless and Tagett, 1989).

Climate

The Florida Bay region has a tropical to subtropical climate. Air temperatures are 16–33°C, with an average of 28°C. Prevailing winds are southeasterly and easterly, but swing around to the north to northeast during the winter and spring. Winds are on average stronger during the winter (10–20 kn) than during the summer (5–10 kn). Precipitation in Florida Bay can exceed 150 cm y⁻¹, but it is often localized and associated with afternoon thunderstorms that form over the Everglades and move out over the bay. The pronounced wet season extends from June through October, accounting for over 70% of the yearly rainfall, and is characterized by slightly higher
Fig. 6-1. Location map of Florida Bay showing position of major Holocene mudbanks (shaded regions), islands, and basins. The coastal Everglades swamp forms the northern boundary, and the Pleistocene-limestone Florida Keys define the southern boundary of the Bay. Four boxed areas are shown in detail in Fig. 6-4. Four zones of mudbank development have been added (dashed lines) from Wanless and Tagett (1989). Abbreviations of Keys in Florida Bay: Sa, Sandy; Cl, Chuet; Si, Sid; Ro, Roscoe; Br, Barnes; Co, Corinne; Tw, Twin; Ji, Jimmy; Cu, Club; BA, Bob Allen; Ch, Crab; Ru, Russel; Cr, Crane; St, Stake; Pr, Park; Ps, Pass; De, Deer; Cn, Cotton; Ra, Ramshorn shoun; CB, Cross Bank.
afternoon relative humidity (77%). The dry season, which generally starts in January and extends through May, has lower afternoon relative humidity (64%) and strong afternoon winds. In general, evaporation exceeds rainfall except during the late summer and fall months.

Florida Bay experiences approximately 15 major storms a year; mainly they coincide with the passage of cold fronts during the winter (Roberts et al., 1982). In
addition, hurricanes (Ball et al., 1967) strike south Florida approximately every 5–7 years. Hurricanes occur between June and October and can have a tremendous effect on Florida Bay by altering circulation, redistributing sediment, and removing vegetation on the islands.

**Hydrography of Florida Bay**

*Water levels.* Water levels within Florida Bay are controlled by tides, winds, and seasonal changes in sea level. Tides in Florida Bay are mixed diurnal-semidiurnal along the Gulf of Mexico boundary, and semidiurnal along the Atlantic. Tidal range is greatest at the open, western and southern portions of the bay (up to 80 cm). In the interior and northeastern areas of the bay, the tide is essentially damped (< 3 cm) by the numerous shallow mudbanks. Winds are significant controls of water level in these interior regions. For example, the water level in northeastern Florida Bay can be increased by up to 40 cm above normal tide levels when the wind blows strongly for several days from the southwest, and lowered by as much as 40 cm when it blows strongly from the northeast. A seasonal steric effect in the Gulf of Mexico causes water levels within Florida Bay to change annually by as much as 20 cm (Kramer et al., 1994). Because of this effect, water levels reach their yearly maximum levels during the fall (September–November) and their lowest levels during the spring (March–May).

*Salinity.* The variation in salinity of Florida Bay waters reflects intra- and interannual patterns. In general, there is less variation along the more-open western and southern portions of the bay and increased variation in the interior portion. Salinities as high as 80 g kg⁻¹ and as low as 15 g kg⁻¹ have been reported in the central portion of the bay. These variations are related to (1) freshwater input and (2) seawater penetration from the Gulf of Mexico and through the Florida Keys. The freshwater input into the bay is derived principally from three sources: Shark River, Taylor Slough, and local rainfall. Approximately 90 km³ y⁻¹ of water is discharged through Shark River to the west of peninsular Florida (Fig. 6-1). A portion of this runoff is believed to find its way into the western portion of the bay although the precise amount is not known. The smaller discharge of Taylor Slough (9 km³ y⁻¹) is perhaps volumetrically more important to Florida Bay as it enters directly into northeastern Florida Bay. Historically, the magnitude of the Taylor Slough runoff was probably larger, as it is now highly controlled by agricultural and urban interests in the south Miami area.

*Stable isotope composition.* Although there are slight differences in behavior between δD and δ¹⁸O in Florida Bay, the behavior of the two isotopes can be considered identical for the purposes of this account, and so discussion here will be limited to δ¹⁸O (Swart et al., 1989b). The δ¹⁸O composition of Florida Bay waters is governed by a combination of four distinct influences (Fig. 6-3). First is input of isotopically heavy freshwater (δ¹⁸O = + 3‰, SMOW) from the Everglades: these
Fig. 6.3. Hypothetical model showing possible causes of the oxygen isotopic composition and salinity in Florida Bay. Sources of water are Everglades, ocean, and local precipitation. The mixing of waters from these sources combined with the evaporation effect leads to the large range in Florida Bay water (shaded).

waters are enriched as a result of extensive evaporation that occurs during their slow flow through the Everglades. Second, there is the isotopically normal marine water from the Florida Keys ($\delta^{18}$O = 0% to +1% SMOW), and third, an input of isotopically depleted rainwater ($\delta^{18}$O = -3.0 SMOW; Swart et al., 1989b). Finally, and perhaps the most important influence, is the evaporation of water in the bay itself. The maximum $\delta^{18}$O isotopic composition that can be attained by the water within the bay is dictated by isotopic exchange between the atmosphere and the bay and, therefore, is related to the relative humidity and temperature. For conditions prevalent in south Florida, this maximum $\delta^{18}$O value is approximately +3% SMOW. Therefore, inundation of Florida Bay by marine water, which can act to either lower or raise the salinity, will usually act to decrease $\delta^{18}$O. Increased discharge from the Everglades, on the other hand, will decrease the salinity but will not affect the oxygen isotopic composition of the water (Swart et al., 1989b).

Sediments

Unconsolidated carbonate sediments comprise nearly 95% of the sediments within Florida Bay; the remainder consists of silica and detrital clays. The majority
of these sediments is believed to be the result of biogenic precipitation of skeletal material, principally as organisms which encrust the *Thalassia* communities (Nelson and Ginsburg, 1986; Bosence, 1989). As the organic portion of the grass dies and decomposes, the small carbonate encrustations (red algae *Melobesia membranacea* and *Fossiella farinosa*; serpulid worm *Spirophis* spp.) are released to form part of the sediment. The production from these encrustations in eastern Florida Bay has been estimated to be 118 g m⁻² y⁻¹, six times more than that derived from *Penicillus* in a similar area (Nelson and Ginsburg, 1986). Minor amounts of sediments are supplied by calcareous green algae such as *Halimeda* spp. and *Penicillus* spp., the small finger coral *Porites* spp., various species of mollusks, and foraminifera. Opaline silica (radiolaria, diatoms and sponge spicules) and organic matter are also found in the sediment.

The gentle east-to-west slope of the underlying Miami Limestone has led to marked differences between eastern and western Florida Bay. The eastern portions of Florida Bay, for example, generally have a sparse bottom fauna and lower carbonate production. Basins are characterized by smaller amounts of sediment; most of the finer material has been winnowed by wave action, leaving only a coarse lag deposit of molluscan shell fragments (Ginsburg, 1956; Enos and Perkins, 1979). In contrast, the western portions of the bay have luxuriant carpets of marine grasses (*Thalassia* spp., *Halodule* spp.), very high carbonate production rates, and thicker sediment cover over the basins.

The mineralogy of Florida Bay sediments reflects the relative contributions of the various biogenic components. On average, the sediments are 60% aragonite, 20% high-Mg calcite (HMC), and 15% low-Mg calcite (LMC) with minor quantities of detrital quartz and opaline silica. Detrital dolomite comprises up to 5% of the sediments found in the northwestern corner of the bay and is thought to originate from exposed portions of the Hawthorn Formation (Miocene) to the north (Taft and Harbaugh, 1964; Scholl, 1966). Samples rich in LMC occur principally in the northern portion of Florida Bay and are derived from freshwater marls which form in the Everglades.

**Mudbanks**

Mudbanks typically consist of bioturbated peloidal wackestone, grey molluscan wackestone, and minor amounts of molluscan packstone and pelleted mudstone (Enos and Perkins, 1979; Tagett, 1988; Wanless and Tagett, 1989). The mudbanks record a history of migration, with windward erosion and leeward sedimentation. Based on the fact that the northern and eastern margins of the banks are erosional, Wanless and Tagett (1989) concluded that winter storms rather than hurricanes are responsible for the deposition and movement of the banks. In some instances, mudbanks have migrated substantially across the bay bottom and in the process obliterated the record of earlier phases of the bank's history.

Wanless and Tagett (1989) also recognized four zones of mudbank development within the bay (Fig. 6-1): (1) an inner destructive zone (where mudbanks are
shrinking); (2) a central migrational zone (where mudbanks are migrating); (3) a western constructional zone (where mudbanks are growing); and (4) an outer destructional zone. As the names imply, different processes are taking place in different portions of the bay. The controls on these processes relate mainly to sediment supply and wave energy. In eastern Florida Bay, for example, sediment supply is limited; as a result, the mudbanks are discontinuous, and there is only a thin veneer of grainstone covering the basin floor. In contrast, there is an ample supply of sediment in central Florida Bay, and so a continuous network of mudbanks has been formed (Fig. 6-4D). In the western portion of Florida Bay, there appears to be a large increase in sediment supply, for banks have coalesced and are actively expanding on all flanks (Wanless and Tagett, 1989).

MUD ISLANDS

Physiography

Islands within Florida Bay have been divided into three groups or "stages" based on their vegetation and topography (Craighead, 1964): (1) low or early stage, (2) middle stage, and (3) high islands or late stage. In their early stage, the islands are covered by mangrove swamps, algal mats, and halophytic marshes; middle-stage islands support brackish-water vegetation, mainly black mangroves (Avicennia nitidae) and halophytic marshes; late-stage islands show growth of grass, palms and hardwoods. It is clear that the types and distribution of vegetation on these islands depends strongly on topography; elevation differences of mere centimeters often produce striking changes in vegetation (Davis, 1940). Extensive examination of diverse islands by Enos and Perkins (1979) led them to conclude that the "stages of development" are not related so much to island age as to the amount of storm deposition and sediment trapping. Ginsburg and Lowenstam (1958) recognized that nearly all of the supratidal sediment accumulating on the interior portions of islands is in fact brought in during storms. Hurricane Donna, which struck Florida Bay in 1960, is known to have deposited as much as 10 cm of well-sorted mud on the interior of some bay islands (Ball et al., 1967; Craighead, 1964).

Topographic features on the islands are small. Relief is generally measured in centimeters. Most islands have three principal topographic features: a high leeward side, a central depression, and a fringing levee. The high leeward side is 20–50 cm above MSL and often contains a small brackish-water lens, which supports a variety of hardwood trees and grasses. The lowest portion of a typical island includes a central area of saline mud flats and mangrove swamps, which are within 10 cm of MSL. The central mud-flat areas often contain small ridges (10–20 cm high), which are commonly colonized by black mangroves. The fringing levee is generally composed of skeletal beach sand 5–40 cm above MSL and borders much of the island shoreline. This levee is especially pronounced on low-lying islands and strongly influences the surface-water and salt balances on the islands.
Fig. 6-4 (locations on Fig. 6-1). (A) Cross section through Cluett Key, a late-stage island, western Florida Bay. This island, which is composed primarily of supratidal sediments, nucleated during the initial transgression. (From Videlock, 1983.) (B) Cross section through Crane Key, south-central Florida Bay. In this cross section, the island is shown to have nucleated on mudbank sediments some time after initial transgression, although subsequent work has shown that, on other parts of the island, supratidal sediments sit directly atop peat layers. (From Enos and Perkins, 1979). (C) Cross section of Jimmy Key, central Florida Bay, showing that island has only recently formed on underlying mudbank. Age of island is thought to be < 400 years. (Modified from Burns and Swart, 1992.) (D) Cross section of Russel Bank, central Florida Bay, showing leeward-dipping, layered mudstone sequences capped by seagrass-influenced sediments composed primarily of bioturbated wackestones and some packstone layers. (From Wanless and Tagett, 1989.)

Geologic history

The mud islands contain the most complete sedimentological history of Florida Bay's development over the past 4,000 years. Of the hundreds of small mud islands that dot the bay, only a handful of the more sizable islands (area > 0.5 km²) have been studied. Cores pushed through the soft mud-island sediments to the underlying
bedrock reveal a lower transgressive sequence beneath a regressive sequence (Fig. 6-4). Enos (1989) identified five distinct lithologies (Types I to V). Type I lithology, which is a supratidal mud found as the uppermost unit at most islands, is a highly oxidized, white to grey, laminated mud characterized by a crumbly texture thought to be the result of frequent drying and wetting (Enos and Perkins, 1979). Type II sediments are dark grey-brown wackestones that make up a marine mudbank succession underlying many younger islands. These sediments are characterized by numerous Thalassia rhizome sheaths containing successive accumulations of fining-upwards sequences broken up by occasional packstone shell layers (Wanless and Tagett, 1989). Type III sediments, which commonly overlie peat layers, consist of dark to medium grey, molluscan packstones and wackestones, often interlayered within peat layers and extensively rooted, these sediments are similar to those found forming today in the open basins of the bay. The two other lithologies are peats (IV) and freshwater calcitic mud (V), both of which are found at the base of many islands and mudbanks. Davies (1980) showed that the basal peats can be of either freshwater (Mariscus spp.) or brackish-water origin (Rhizophora-Avicennia), but all peats higher in the column are of marine origin.

Enos (1989) identified two types of island development: early colonization and late colonization, where “early” and “late” refer to the timing of colonization relative to the initial submergence of Florida Bay. Early-colonization mud islands (Fig. 6-4A, B) are characterized by a transgressive sequence consisting of calcite marls (Type V lithology) and basal peat (IV) intermixed with basin sediments (III, interpreted as representing an initial marine flooding of the bay) and an overlying regressive or progradational sequence consisting of a continuous succession of supratidal sediments (I). Enos (1989) classified the following islands as early-colonization mud islands: West Bob Allen, Calusa, Crane, Eagle, Lake, Man of War, Murray, Palm, Pigeon, Cluett, and Sid Keys.

In contrast, late-colonization mud islands, which by definition are thought to have nucleated on mudbanks some time after the initial flooding of the bay, are characterized by the occurrence of subtidal sediments (II) through a portion of the sequence, and, in some cases, a lack of basal peat. Such islands are not believed to be typical of Florida Bay; in fact, Enos (1989) suggested that the succession of sediment types on these islands may be simply the result of migration of a precursor island over an adjacent mudbank as a result of winds and currents. On the other hand, there is now good evidence that Jimmy Key (Fig. 6-4C) formed recently on a mudbank (Burns and Swart, 1992): distinctive shell layers can be traced from the islands into adjacent mudbanks, and δ^{13}C values of the organic material in the sediment change upward from an isotopically heavy marine signal (i.e., mudbank) to more depleted values characteristic of mangroves (i.e., island). In the case of Jimmy Key, the age of the veneer of island sediments is estimated to be only 200–400 years, and so this island has been emergent for only this period of time (Burns and Swart, 1992). In other regions, the mangrove colonization is known to be even more rapid; for example, Cowpens Cut through Cross Bank shows that entire area essentially has been colonized since 1949. Other islands which are suggested to have formed on mudbanks (hence late-colonization islands) include Bald Eagle, Bob
Allen (east), Bottle, Cotton, Johnson, Rabbit (north), Shell, and Stake (Enos, 1989).

**Hydraulic properties**

The unconsolidated carbonate muds which make up the island sediments are characterized by low hydraulic conductivity \(10^{-1} \text{ to } 10^{-3} \text{ m day}^{-1}\) and high porosity (45–85%). The majority of the sediment is micron-size aragonitic needles mixed with calcitic mollusks and HMC foraminifera. The sediments are dominantly mudstones interrupted by discontinuous wackestone and, less commonly, packstone units. In the upper 10 cm of Crane Key, the hydraulic conductivity is \(10^{-3}–10^{0} \text{ m day}^{-1}\) (Enos and Sadowsky, 1981). This large range is caused by root voids, gas bubbles, and large desiccation cracks, all of which can extend to depths of 30 cm and produce an extensive network of macroporosity (Enos and Sadowsky, 1981). In contrast, Enos and Sadowsky (1981) measured a hydraulic conductivity of \(10^{-4} \text{ m day}^{-1}\) at Ramshorn shoal, a predominantly fine-grained mudbank lacking shells. This low value may be attributable to the lack of macropores and is near the intrinsic value of pure carbonate mud, based on consolidation experiments (Juster, 1995). Hydraulic conductivity decreases towards this intrinsic value with depth on both islands and mudbanks due to compression of the pore matrix and clogging of the macropore network during burial (Juster, 1995).

Porosity in all the island sediments is very high. Enos and Sadowsky (1981) found values of 61–68% in samples from Crane Key. Videlock (1983), who used a gamma-ray attenuation method (GRAPE) for sediments from Cluett Key, found values of 53–68%, with porosity near 80% at the base due to the presence of peats. Juster (1995) measured a decrease in porosity from 69% at the surface to 65% at 2 m depth on Jimmy Key and the adjacent mudbank, but a reverse relationship was measured in Cluett Key sediments indicating that burial compression may not always be significant in reducing porosity.

**Surface waters**

A strong seasonality characterizes the water level and salinity of surface water bodies that form on the island interiors. During the late summer and fall months, when bay water levels are at their maximum, islands can be flooded daily with each high tide (Kramer, 1996). These months are also the wet season in south Florida, and precipitation on the islands can often exceed 20 cm during a single rain event; at such times, the pond water can be diluted to near freshwater salinities. In contrast, during the winter and spring months when rainfall is minimal and bay water levels are, on average, lower, it is not uncommon for the interior ponds of islands to dry out completely for several weeks. Evapotranspiration in the interior areas of the islands is from evaporation from free surfaces, evaporation from exposed soils, and transpiration from the low vegetation. Evapotranspiration on Florida Bay islands is not quantified. From pan-evaporation measurements, it is known (NOAA, 1989) that
evaporation in south Florida varies seasonally from 5 cm mo\(^{-1}\) to 22 cm mo\(^{-1}\). The annual average total exceeds 200 cm y\(^{-1}\) (NOAA, 1989).

The frequency that bay waters flood onto a particular island is controlled by the location of the island within Florida Bay's tidal regime and by the elevation of the levee surrounding the island. Islands with a low levee (e.g., Jimmy Key) are wet islands that are easily overwashed by tides. Higher islands (e.g., Cluett Key) are dry islands that are flooded mainly by the spring tides and when bay water levels are at their steric maximum (summer and fall). The frequency of tidal flooding, in turn, controls the salinity of surface waters that collect on the islands; wet islands have the lower-salinity ponds. Undoubtedly, major storms and hurricanes reshape levees from time to time, thus altering the water and salt balance on each island. This periodic restructuring of the levee may explain, in part, why massive mangrove die-off can occur on the interior of some islands following major storms (Ball et al., 1967).

*Groundwater*

Groundwater on these islands is derived from the ponds that occupy the island interiors. Salinity of groundwater collected from the upper 100 cm of island sediment ranges from as low as 20 g kg\(^{-1}\) after a large precipitation event to more than 200 g kg\(^{-1}\) during the last stages of pond evaporation (Kramer, 1996). The presence of wind-blown salts mixed in with the upper surface sediments can mask the true origin of the waters recharging the groundwater. For example, using the \(\delta^{18}O\) and \(\delta D\) composition of groundwater and surface waters from Cluett Key, Swart et al. (1989b) showed that isotopically light groundwater of meteoric origin can have salinities of 35 g kg\(^{-1}\) (essentially equivalent to seawater) as a result of dissolving these surface salts as the water seeps into the ground (Fig. 6-5). Although major changes occur in the upper 100 cm of groundwater associated with seasonal changes in the island's water balance, groundwater below 100 cm shows negligible salinity changes with time and probably represents a time-averaged value of the seasonal changes occurring in the overlying column. Similar processes have been documented in groundwaters of *Spartina* salt-marsh settings (Lord and Church, 1983, Casey and Lasaga, 1987).

Although it is well known that many mud islands in Florida Bay contain hypersaline waters (Davis, 1940; Halley and Steinen, 1979; Swart et al., 1989b), there has not been any extensive study of the islands to determine whether there are systematic patterns in the surface and groundwater between the various islands of different types. Our recent studies (Kramer, 1996) show that the deeper groundwater in most islands has a salinity in excess of 65 g kg\(^{-1}\). There is a tendency for the groundwater salinity in islands in the western portion of Florida Bay to be higher than that in the more eastern islands (Fig. 6-6). This geographic zonation can be attributed to three causes. First, the levees, and the islands themselves, are higher in the western and central portions of Florida Bay and, therefore, less frequently inundated. Second, the northeastern portions of the bay receive more rainfall, which dilutes the surface water and groundwater. Finally, the water of Florida Bay itself is
not as saline in the northeast part of the bay because of the freshwater runoff from the Everglades.

The movement of saline groundwater in Florida Bay mud islands is becoming known. We are completing a study of the hydrology of two islands, Cluett Key and Jimmy Key as part of a long-term project aimed at quantifying the water flux through island sediments (Kramer et al., 1993; Juster, 1995; Kramer, 1996; Juster et al., 1997). Mechanisms that can drive water through the sediments are topographic head, evaporative pumping (Hsu and Siegenthaler, 1969), and density-driven reflux (Adams and Rhodes, 1960), although the importance of each remains to be fully documented and understood. Hydrological observations on Cluett Key (Juster, 1995; Juster et al., in press) indicate that the pond floor is “perched” or elevated about 10 cm above mean sea level. Thus, a large but variable hydraulic gradient (~0.1) is produced between surface waters and the underlying limestone when the
Fig. 6-6. Porewater salinity profiles from 15 islands found throughout Florida Bay showing hypersaline character of the groundwaters. Islands have been divided into western, central, and eastern regions of the Bay. Islands in the east have lower salinities probably due to lower elevation and larger amounts of rainfall.

pond is present. This gradient has the ability to move brines vertically downward and is probably the dominant hydraulic drive on the higher islands. Estimated rates of downward velocities are on the order of 10–25 cm y⁻¹ (Kramer et al., 1993; Juster et al., 1997).

CASE STUDY: HYDROGEOCHEMICAL EVIDENCE OF DIAGENESIS

Diagenesis

As a result of the young age of the sediments (< 4,000 y B.P.) and the relatively slow rates of sedimentation (< 1 mm y⁻¹), the diagenetic stabilization of Florida Bay muds is in its early stages. The first study of the porewater geochemistry of the mudbanks in Florida Bay revealed little change in the concentrations of Cl⁻, Ca²⁺, Mg²⁺, and Sr²⁺ (Berner, 1966). Subsequent studies have shown small, but nevertheless significant, changes in the concentration of SO₄²⁻ (Rosenfeld, 1979) and Ca²⁺ and alkalinity (Walter and Burton, 1990) in the upper portions of cores through the mudbanks. Walter and Burton (1990) suggested that such changes in Ca²⁺, SO₄²⁻, and alkalinity in porewaters from mudbanks were probably affected by some type of advection process mediated by bioturbation. These workers proposed that the rate of carbonate dissolution in the mudbanks may in fact be much larger than that indi-
cated by the porewater profiles. According to Walter and Burton (1990, p. 602), "volumetrically significant dissolution may occur" in these sediments.

In contrast to porewaters of mudbanks, groundwater squeezed from cores taken on exposed islands reveal large changes in concentrations of Ca$^{2+}$, Mg$^{2+}$, and Sr$^{2+}$, and SO$_4^{2-}$ throughout the entire section (Swart et al., 1989a; Burns and Swart, 1992) (Fig. 6-7). The direction and the magnitude of these changes, however, are not always the same, and there are large differences in the nature of profiles between various islands. For example, Cluett Key shows a deficit of normalized Ca$^{2+}$ throughout the core, whereas, in Jimmy Key, the normalized Ca$^{2+}$ concentrations are close to that predicted from a simple evaporation model (Fig. 6-7). Differences between these and other islands relate to processes of evaporation, precipitation of minerals such as halite and gypsum, and carbonate dissolution and precipitation.

Fig. 6-7. Porewater concentration of Ca$^{2+}$, Mg$^{2+}$, SO$_4^{2-}$, and alkalinity taken from Jimmy and Cluett Keys. Ion concentration is given in relative mM values above or below the concentration that would be expected if Cl$^-$ were behaving conservatively; the value is calculated by:

$$
\text{ion}_{\text{rel}} = \text{ion}_{\text{measured}} - \text{ion}_{\text{seawater}} \times \frac{\text{Cl}_{\text{measured}}}{\text{Cl}_{\text{seawater}}}.
$$

(Data from Burns and Swart, 1992; R. Steinen, unpubl.)
As a result of seasonal cycles discussed previously, islands can dry out completely, which causes gypsum and halite to precipitate on the surface sediments. Although these evaporite minerals are not long-lived in that they are redissolved during subsequent flooding, they alter the chemistry of the surface water and, consequently, the underlying groundwater. For example, precipitation of gypsum preferentially removes Ca$^{2+}$ and SO$_4^{2-}$ and, therefore, the ratio of these species relative to Cl$^-$ decreases in the residual fluids (Fig. 6-8). In contrast, the ratio increases in the fluid that subsequently dissolves the minerals. The extent to which these processes alter the groundwater chemistry of an island is related to its hydrological balance; thus, a topographically lower, more frequently flooded island, such as Jimmy Key, will tend to have less precipitation of evaporate minerals than the slightly higher and relatively drier Cluett Key. Cross-plots of ion concentrations in the groundwaters reveal some of these processes. For example, the precipitation of calcite removes Ca$^{2+}$ and causes the groundwater to plot below the line one would expect from the simple evaporation of the fluid (evaporation line) (Fig. 6-9B); groundwater from the saline portion of Cluett Key is an example. In contrast, groundwater from Jimmy Key plots near the evaporation line and thus shows relatively little evidence of precipitation.

![Diagram](image_url)

Fig. 6-8. Cross-plot of total alkalinity vs. SO$_4^{2-}$ for four islands from Florida Bay. Evaporation of surface waters raises SO$_4^{2-}$. Precipitation of calcite lowers alkalinity (typical of "dry islands"). Sulfate reduction lowers sulfate and raises alkalinity (typical of "wet islands").
Fig. 6.9. (A) Cross-plot of sulfate vs. chloride. At low concentrations of sulfate, the process of sulfate reduction causes data to fall below the evaporation line. At high concentrations, the data fall below the line as a result of the formation of gypsum. Excess sulfate concentrations such as occur in Jimmy and Crane Keys are postulated to result from the oxidation of H₂S. (B) Cross-plot of calcium vs. chloride. Data falling below the line at high chloride concentrations are thought to reflect the precipitation of LMC. Data above the line arise from dissolution of aragonite and HMC.
The presence of sulfate reduction on most of the islands is evident from the pungent odor of H$_2$S emanating from the sediment. Jimmy and Crane Keys, however, are notable exceptions that exhibit a slight excess of sulfate in their groundwaters. Although this apparent excess may result from the presence of groundwater deficient in Cl$^-$, another explanation is that the excess results from oxidation of HS$^-$ to SO$_4^{2-}$—i.e., the HS$^-$ is produced lower in the sedimentary section by bacterial sulfate reduction and moves upwards through the pore space where it is eventually oxidized producing sulfate. Such a process may be more in evidence in low islands such as Jimmy Key which contain greater concentrations of organic material in the sediments. It should be noted that the sediments of Florida Bay islands differ fundamentally from iron-rich sediments in which the HS$^-$ would react with iron and form iron sulfide minerals. Pyrite is virtually absent in these sediments, except for very low quantities measured in the underlying peats (Davies, 1980).

Sulfate reduction also generates alkalinity and leads to the dissolution of carbonates by the generation of additional carbonic acid. A plot of sulfate vs. alkalinity (Fig. 6-8) shows three main trends. First is the trend which reflects the evaporation of the fluids. Second, precipitation of LMC or aragonite causes a drop in alkalinity with little change in sulfate. Third, sulfate reduction causes a decrease in sulfate and an increase in alkalinity. Based on this type of plot, it appears that islands such as Sid and Cluett Keys experience evaporation followed by precipitation of carbonate and some gypsum, whereas Jimmy and Crane Keys have sulfate reduction coupled with carbonate dissolution.

*Carbonate reactions*

Carbonate reactions can both decrease or increase the concentration of Ca$^{2+}$ and alkalinity and alter the ratios of Sr$^{2+}$/Ca$^{2+}$ and Mg$^{2+}$/Ca$^{2+}$ of the pore fluids. Dissolution of aragonite tends to increase the Ca$^{2+}$/Cl$^-$ ratio but does not alter the Sr$^{2+}$/Ca$^{2+}$ ratio appreciably, because aragonite has approximately the same Sr$^{2+}$/Ca$^{2+}$ ratio as seawater. In contrast, HMC has a lower concentration of Sr$^{2+}$, and so its dissolution lowers the Sr$^{2+}$/Ca$^{2+}$ ratio of the groundwater. Precipitation of LMC lowers the groundwater Ca$^{2+}$ content but increases the Sr$^{2+}$/Ca$^{2+}$ ratio, because the distribution coefficient for Sr$^{2+}$ into calcite is significantly less than unity. Finally, precipitation of dolomite generally lowers the Mg$^{2+}$/Cl$^-$ ratio and perhaps Mg$^{2+}$/Ca$^{2+}$. Depending upon the stoichiometry of the dolomitizing reaction, precipitation of dolomite may also lower the Mg$^{2+}$/Ca$^{2+}$ ratio. Dolomitization of aragonite generally causes an increase in the Sr$^{2+}$/Ca$^{2+}$ ratio of the fluid.

*Dolomite*

The occurrence of dolomite in these sediments is of particular interest, as hypersaline environments have traditionally been places in which dolomite has been thought to form (Hardie, 1987; Land, 1985). Examples of such settings are coastal sabkhas (McKenzie et al., 1980; Patterson and Kinsman, 1981; 1982), tidal flats
(Gebelein et al., 1980; Behrens and Land, 1972; Shinn, 1968), and islands (Murray, 1969). The occurrence of dolomite in the Florida Bay sediments and mud islands has been reported by numerous workers (Taft, 1961; Deffeyes and Martin, 1962; Degens and Epstein, 1964; Friedman and Sanders, 1967; Steinen et al., 1977; Videlock, 1983; Swart et al., 1989a; Andrews, 1991). The dolomite of Florida Bay islands typically consists of small (1–5 μm), euhedral rhombs, intimately intergrown with surrounding micrite and aragonite needles (Fig. 6-10). Dolomites recovered from the bay sediments are larger (> 10 μm) and generally abraded.

Although it is reasonably well established that the dolomite reported in the marine sediments of Florida Bay is of detrital origin, the dolomites within the islands are considered authigenic (Deffeyes and Martin, 1962; Degens and Epstein, 1964). This conclusion is based on the measured 14C activity, petrographic observations, and the δ13C and δ18O composition of the dolomite (Swart et al., 1989a). For example, Fig. 6-11 shows a comparison of the δ13C and δ18O values of the dolomite retrieved from the marine sediments compared to the dolomite from Crane Key and other sedimentary components. The dolomite studied by Degens and Epstein (1964) has a range of δ18O values (−1 to +1‰) that is clearly isotopically too light to have formed from the hypersaline fluids presently typical of the Holocene islands. In contrast, the δ18O value for Crane Key dolomites (+2‰) is in approximate equilibrium with porewaters throughout the core. δ13C values (−2 to −3‰) suggest an organic influence.

![SEM images of isolated dolomite taken from Crane Key showing small size of rhombs (scale bar = 1 μm). (From Swart et al., 1989a.)](image)
Although there is little doubt regarding the authigenic nature of the dolomite, it is still unknown whether dolomitization is taking place at the present time. In this regard, the most sensitive information we have is from groundwater concentrations of Ca\(^{2+}\), Mg\(^{2+}\), and Sr\(^{2+}\). The behavior of these minor elements during dolomite formation was noted above and leads to revealing cross-plots of Sr\(^{2+}\)/Ca\(^{2+}\) vs. Ca\(^{2+}\)/Cl\(^{-}\) and Sr\(^{2+}\)/Ca\(^{2+}\) vs. Mg\(^{2+}\)/Cl\(^{-}\) (Figs. 6-12, 6-13). The path of dolomitization is shown on these figures, and, as can be seen, none of the data from the islands that we have studied falls into these fields. Based on these analyses of groundwater from three islands in which dolomite is present, we must conclude that there is little evidence of present-day dolomite precipitation; that is, the dolomite that is present must have formed during an earlier time.

We cannot, however, conclude that no dolomite is forming at the present time for there is a finite lower limit determined by the sensitivity of the geochemical analyses. For example, consider dolomitization according to the following stoichiometry:

\[
2\text{CaCO}_3 + \text{Mg}^{2+} = \text{CaMg(CO}_3)_2 + \text{Ca}^{2+}
\]
If 1 g of aragonite (density, 2.86 g cm\(^{-3}\); porosity, 50\%) is filled completely with seawater, the groundwater would contain only \(1.7 \times 10^{-5}\) M of Mg\(^{2+}\). If we were able to use all the Mg\(^{2+}\) for dolomitization, we could dolomitize only 0.03\% of the sediment. Assuming, therefore, that we can detect a change of 2 mM Mg\(^{2+}\) in the groundwater, then our geochemical methods should be able to detect the formation of as little as 0.002\% dolomite in a closed system. This is more sensitive than X-ray diffraction methods by at least three orders of magnitude. Our results, therefore, suggest that if dolomitization is taking place at the present time, the rate must be so low that it does not produce measurable changes in the groundwater.

It is interesting to note that Jimmy Key, a young island, has very low concentrations of dolomite (\(<3\%) compared to Crane and Cluett Keys which have been islands for much of the history of Florida Bay (Fig. 6-14). Such a correlation between age and amount of dolomite—although preliminary and needing further substantiation—would tend to suggest that the dolomitization is an ongoing
Fig. 6-13. The ratio of Mg$^{2+}$ normalized to Cl$^{-}$ vs. the Sr$^{2+}$/Ca$^{2+}$ ratio from three islands, Jimmy, Crane, and Cluett Keys. Increases in the Mg$^{2+}$/Cl$^{-}$ ratio can be expected from the dissolution of HMC and the formation of NaCl. No change can be expected in the Mg$^{2+}$/Cl$^{-}$ ratio as a result of aragonite dissolution or calcite precipitation. Dolomitization would be manifested by a decrease in the Mg$^{2+}$/Cl$^{-}$ and an increase in the Sr$^{2+}$/Ca$^{2+}$ ratio. As in the case of the Ca$^{2+}$/Cl$^{-}$ plot, there is little evidence of dolomitization in the samples studied.

process, perhaps related to the slow passage of the hypersaline groundwater through the sediments. On the other hand, the distribution of significant amounts of dolomite (> 3%) in these older islands is limited to deep layers; the dolomite is not disseminated throughout the column (Swart et al., 1989a; Videlock, 1983). Recent work by Andrews (1991) suggests that dolomite formation is associated with the production of HMC-rich cyanobacterial mats near the surface. Concentrations of dolomite observed at deeper levels in our cores, therefore, may simply be a result of earlier episodes of dolomite formation of the surface sediments of these islands; the dolomitization may have been similar to that occurring on the tidal flats of Andros Island (Shinn et al., 1965).

CONCLUDING REMARKS

The Holocene sediments and islands in Florida Bay have long served as a modern laboratory for the study of ancient carbonate sedimentary sequences. They represent
a record of a broad array of depositional environments and are undergoing very early stages of diagenesis.

The occurrence of dolomite on the islands has prompted speculation that these unique hydrogeochemical environments may be sites of present-day dolomitization (Friedman and Sanders, 1969; Swart et al., 1989a; Steinen et al., 1977). Others have noted that the islands may be active sites of recrystallization, and this has led to speculation that sediments in this type of island environment may to be converted to calcite and dolomite relatively rapidly. Our geochemical data though not exhaustive indicate that recrystallization is taking place in the subsurface of these islands but that the rate is relatively slow, and, when the relatively long residence time of the groundwater is factored in, it appears that the recrystallization may be significant only if the present hydrological conditions persist over long periods of time.

There is no evidence of measurable changes in the Mg$^{2+}$/Cl$^{-}$ or Sr$^{2+}$/Ca$^{2+}$ ratios that would suggest that significant dolomitization is presently taking place in the subsurface. We do not rule out the presence of specialized environments of dolomite
formation such as described by Andrews (1991), but these are not widespread. Furthermore, we do not discount the formation of very minor amounts of dolomite throughout the sediment, but in concentrations that do not appreciably alter the hydrogeochemistry. Dolomite formation by this mechanism — like the recrystallization — would be significant only if the present hydrological conditions were stable over extended periods of time. We suggest that the major concentrations of dolomite found within the islands relate to formation during a previous time, perhaps when the dolomite was close to the surface and under conditions similar to those described by Andrews (1991).

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