Quantifying carbonate sediment mixing across a formational boundary, an example from the Late Pleistocene Miami Limestone

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Abstract
Holocene marine sediment from proto-Florida Bay has been eroded, suspended, transported landward and has infiltrated downward (to 1 m depth) to fill secondary pores in subaerially exposed Late Pleistocene limestone. Sediment mineralogy (up to 95% aragonite), sediment texture (mostly needle-like, <15 µm), and bulk sediment stable-isotope values (upward trend in δ18O from −1 to −3‰ and δ13C from 3.5 to −2‰) of infill sediment is used to document inter-stratal movement of sediment ahead of marine transgression. Nearby Florida Bay probably supplied needle-like aragonite crystals by a storm surge mechanism that transported sediment landward and perhaps sorted the marine mud fraction. Currently, the core site is covered by freshwater microbial lime mud with a texture different than most infill sediment. That sediment crosses a formational and sequence boundary can have several implications: microfossil displacement; sedimentologic interpretation of mixed mud and sand-size grains; porosity and permeability reduction; and associated diagenesis with perched ground water. If sediment infiltration is recognized in core and thin section, these mud-filled pores can aid in recognition of subaerial exposure. Stratigraphically, the occurrence of marine sediments landward of the shoreline and below freshwater deposits may be interpreted as the result of a rapid rise and fall of sea level.

KEYWORDS
Carbonate mud, cross-formation mixing, Pleistocene Miami Limestone, selective diagenesis, storm sedimentation.

1 INTRODUCTION
Mixing of younger sediments downward into older sediments can occur in several ways, either physical, chemical or biological. Here, distinct and identifiable pockets of admixed sediment contained within the host sedimentary limestone is examined. This cross-stratal mixing can occur in numerous ways and at different scales. For example, earthquake-related seismic deformation can produce 5–20 cm wide cracks in the host sediment that are later filled (Velázquez-Bucio & Garduño-Monroy, 2018). Chemically, limestone dissolution can generate voids that range from millimetres to decimetres. Biological activity, mainly bioturbation, is well-known to create open voids in un lithified sediments.
A main focus of most downward mixing, however, has been through processes related to biological activities, mainly bioturbation while sediment is still unconsolidated (Bianco, 2017; Droser, Jensen, & Gehling, 2002; Frey, 1971; Kennedy, 1975; Seilacher, 1964; Teal, Bulling, Parker, & Solan, 2008; Zhang et al., 2017). Bioturbation structures can be held open by mud-packed burrow linings or partial cementation, and sediment then infills open burrows (Tedesco & Wanless, 1991; Wanless, Tedesco, & Tyrrell, 1988; Warme, 1975). In semi-consolidated sediment, bioturbation involves a firmground (stiff but uncremented sediment close to the sediment–water interface, i.e., Droser et al., 2002). The firmground contains openings from either biological-related boring or burrowing (Warme, 1975), and these cavities are then sediment filled. Bioturbation-related mixing and associated structures are usually recognizable based on destruction of layering or from a contrast in sediment between host sediment and infilled bioturbation sediment (Bromley, 1975; Frey, 1975).

Karstic dissolution in limestone, both surficial and subsurface also provides a mechanism for infiltration of sediments across stratal boundaries, or at least makes for a highly irregular surface with sediment-filled depressions. Surface dissolution features in limestones can range from millimetres to many metres in width and height (Choquette & James, 1988; James & Choquette, 1984). Mechanisms for downward sediment movement include conditions where host rock is dissolved away to form a surface cavity, a series of subsurface cavities or caves, or a collapse-related structure. Dissolution and collapse is especially prominent in limestone host rocks. For example, quartz sand-filled dissolution holes in oolitic limestone during subaerial exposure (Truss, Grasmueck, Vega, & Viggiano, 2007), collapse depressions with marine sediment infill around reefs (Bonem, 1988), large karst "blue holes" (Shinn, Reich, Lockers, & Hine, 1996), surface dissolution in “banana holes” (Harris, Mylroie, & Carew, 1995) and influx of sediments in cave systems (van Hengstum, Scott, Gröcke, & Charette, 2011; Wicks, Paylor, & Bentley, 2018). Of recent interest is surface to subsurface movement by surface–water drainage and flow within karst aquifers. For example, Herman, Tancredi, Toran, and White (2007), Herman, Toran, and White (2012) describe suspended load movement in karst springs, and Bosch and White (2007) document movement of silicilastic sediment in karst limestones. Pronk, Goldscheider, Zopfi, and Zwahlen (2008) proposed percolation and particle transport in the unsaturated zone above a karst aquifer system by groundwater turbidity during natural and artificial recharge events.

In the marine realm, the primary porosity created in coarse-grained deposits, for example a reef framework, can have sediment infill by fine-grained mud to produce geopetals (Ginsburg, Shinn, & Schroeder, 1967; James & Ginsburg, 1979). These geopetals, by definition (Sander, 1951), are cavities that have been partially infilled by fine-grained sediment. The partial infill suggests that infiltration of sediment was restricted by either insufficient time or that the pathways were blocked by sediment or cementation. Similar features, diagenetic vadose silts, were extensively discussed by Dunham (1969) and interpreted by Dunham to have formed in porous rock while in the vadose zone during early emergence. Dunham (1969) also evaluated and discussed the feasibility that internal sediments (detrital) may also occur in subaerially exposed limestones that were transgressed and subjected to currents strong enough to forcibly emplace sediment. A confirmation of infiltrated marine sediment would validate the (slightly modified) combined subaerial–submarine model of Dunham.

In some sedimentary strata, downward movement of sediment from an overlying bed is relatively easy to distinguish by the presence of distinct sediment types (Bromley, 1975; Frey, 1975; Zhang et al., 2017). In some cases, however, especially limestone where sediment colour or composition is similar, mixing of different sediments can be much more difficult to recognize. It may even produce a rock where distinct grain types and grain sizes commingle and result in an unusual texture difficult to interpret hydrodynamically. It is proposed here that fine-grained sediment can move across stratal boundaries, in this example a subaerial exposure surface, and penetrate into the underlying rock unit.

A short core (1.05 m) from the Miami Limestone in Everglades National Park (Figure 1) was examined to investigate this sediment infiltration hypothesis. This host rock was deposited approximately 120 kyr ago and was subaerially exposed until the mid-Holocene (Osmond, Carpenter, & Windom, 1965). Subaerial exposure has created secondary porosity in the Miami Limestone, which factors prominently for sediment infiltration into host rock. Subsequently, Holocene sea-level transgression on the southern Florida shelf has raised the water table and created a freshwater microbial habitat that today covers the core site with low-Mg calcitic mud. These freshwater sediments are distinct from marine sediments in Florida Bay (mostly aragonite and high-Mg calcite), and provide a means to determine provenance of infill sediment (marine vs. terrestrial). Several questions still remain, however, especially with regards to the aragonite mineralogy of infilled sediment and an absence of high-Mg calcite, which is currently abundant in Florida Bay. This study provides an opportunity to quantify sediment transfer across a subaerial exposure boundary and examine the mechanism and implications (porosity, permeability and diagenesis) for cross-stratal sediment mixing.

1.1 | Host rock: Miami Limestone

The bedrock beneath modern Everglades freshwater sediments consists of Miami Limestone (Cooke & Mossom, 1929). The name Miami Limestone was proposed by Hoffmeister, Stockman, and Multer (1967) to include a combination of “oolitic” and “bryozoan” lithofacies on the south-eastern Florida peninsula and lower Florida Keys. More recently, Evans (1983, 1987) divided the
Miami Limestone into three lithofacies types. These include subdivision of oolitic facies into "bedded and mottled" in the east where they form the Atlantic Coastal Ridge, a shoal and tidal channel complex, and "bryozoan" facies in more quiescent western interior settings (protected from waves by an inner shelf position or by the shoal complex to the east). The bryozoan lithofacies beneath the Everglades is a mixture of mainly peloids, skeletal fragments and a small percentage of ooids nearer the tidal shoals. The most dominant skeletal form is cheilostome bryozoa that can in places be so dominant as to form a framework (Hoffmeister et al., 1967). Bryozoan facies is thought to underlie most of the Everglades, although relatively few rock cores have been taken in the southernmost Everglades, especially in areas south (seaward) of the shoal complex to the east. The location of this study is purportedly within the bryozoan facies, and rock can frequently exhibit a mottled or vuggy fabric (Evans, 1987). These vuggy pores are of key interest for downward infiltration of marine sediment.

**METHODS**

**2.1 | Core ENP-2 location, core collection and processing**

An approximately 7 cm diameter core was drilled as part of a study on overlying freshwater microbial mud at Paurotis Pond, within Everglades National Park (25°18′43.59″N, 80°47′44.81″W, Figure 1). The veneer (<10 cm) of surface sediment was removed from the drill site. The 104 cm long core was drilled with a diamond bit on a gasoline-powered field drill, and cut lengthwise with a diamond tile saw. The cut core produced two flat slabs that were used to assess lithology and to provide samples (Figure 2).

**2.2 | Mineralogy (Miami Limestone, white sediment, surface sediment)**

Mineralogy of Miami Limestone and infills of unconsolidated white sediment was determined using X-ray diffraction. In cut core, unconsolidated sediment was gently scraped from 10 sites (5.6, 12.2, 21.6, 34.8, 42.0, 55.7, 64.5, 78.3, 88.6 and 101.0 cm depth). Three host rock (Miami Limestone) samples were collected near the core's base (83.6 cm), middle (53.7 cm) and top (13.2 cm). All samples were ground to a powder by hand and mounted as a paste on a glass slide. Slides were scanned from 20° to 50° 2θ on a Panalytical X'Pert Pro automated diffractometer. Determination of percent aragonite and calcite was determined by peak-height ratios (Griffin, 1971) and confirmed by an in-house peak area Excel macro for carbonate mineral mixes. Mole percent Mg in calcite was determined based on (112) peak position and compared to the calibration chart of Goldsmith, Graf, and Heard (1961).
The mineralogy of surface sediment was determined by Pederson (2017) using similar preparation methods and an X-ray diffractometer. These surface samples are plotted as depth “0 cm” on the mineralogy plots. Mineralogy is the main method for differentiating marine sediment (mainly aragonite, high-Mg calcite) from non-marine sediment (mainly low-Mg calcite).

### 2.4 Scanning electron microscopy petrography of white sediment

A fine-grained unconsolidated white sediment was collected for Scanning electron microscopy (SEM) analysis from three locations in the core. Raw sample (ungrounded) was sprinkled on an SEM stub and then sputter coated with palladium. Samples 1 (12 cm), 6 (35 cm) and 18 (101 cm) were examined with a Philips XL-30 environmental scanning electron microscope and SEM was used to identify the crystal forms within the unconsolidated sediment.

### 3 RESULTS

#### 3.1 ENP-2 core description

Core ENP-2 consisted of a well-indurated limestone and patches of white, unconsolidated fine-grained sediment. The limestone was tan to very pale brown (10YR 7/2 light grey to 10YR 8/2 very pale brown). Based on regional geology, this rock is Miami Limestone. This indurated limestone consists of peloidal, skeletal and oolitic limestone with abundant rounded to subrounded intraclasts. Macroscopically, remains of bryozoa are absent, but might have been present prior to meteoric alteration. Texturally, the rock consists of an intraclast-skeletal floatstone, with a peloidal and oolitic matrix.

On the cut surface were numerous irregular to oval shaped patches that contained a white (between a Munsell value of 9 and 10) unconsolidated sediment. Host rock immediately around these white sediment areas had a ~1 mm thick, slightly darkened surface lining the pore. This boundary between indurated host rock and sediment was confirmed by emptying several sediment-filled structures to reveal small cavities. White, unconsolidated fine-grained sediment occurs throughout the 104 cm core, but becomes less prominent in the bottom 10 cm, where some unfilled pores occur. The small cavities in the rock appear to have been oval, rounded and elongate voids, with some that crosscut matrix and intraclasts.

The white sediment was extremely fine-grained and did not contain sand or silt-size grains, as it had no grittiness to the touch. A cursory examination of several samples under a binocular microscope likewise showed no coarse (silt or sand) grains or skeletal fragments.

The maximum diameters of pores that contain white sediment were measured on cut core surfaces and are about a...
centimetre or greater (mean = 18.3 mm, median 16.5 mm, 
$SD = 11.2$ mm, $n = 74$) (Figure 3). Filled pores greater than 
4 cm are elongate in shape and may be either joined vugs or 
burrow-related structures. Open (unfilled) pores are gener-
ally smaller (mean = 5.7 mm, median 4.8 mm, $SD = 4.0$ mm, 
$n = 47$), except in the lower part of the core where white sed-
iment is absent (Figure 3). Below 95 cm depth, several >1 cm 
unfilled pores occur (Figure 3).

3.2 | Mineralogy

3.2.1 | Miami Limestone host rock

Three intervals (shown in Figure 4) in core ENP-2 were found 
to be a mix of aragonite and low-Mg calcite. The percentage 
of aragonite was ~60% in the lowest sample, 52% in the mid-
dle sample and 57% in the uppermost sample (Figure 4).

3.2.2 | Unconsolidated white sediment

The white sediment that infills limestone vugs is a mixture of 
just two minerals, aragonite and low-Mg calcite (Figure 4). 
The amount of aragonite decreases upward from over 90% at 
the base of the core (about 1 m depth) to about 75% at 20 cm, 
and about 35% at 12 cm. Conversely, the amount of low-Mg 
calcite increases upward, to 65% at 12 cm, and to 100% in a 
filled vug at about 6 cm, and in surface sediment that overlies 
Pleistocene rock (Pederson, 2017). The calcite was all of the 
low-Mg variety (<4 mole percent Mg) based on the position 
of the calcite (112) X-ray peak.

3.3 | Stable-isotope data and profiles

Samples of unconsolidated white sediment from 18 sites 
were analysed for their stable isotopic composition. Stable 
carbon and oxygen values show a similar upward trend ($R$-
value of .88) from more positive values at the core's base, to 
more negative values towards the core top (Figure 5).

3.3.1 | $\delta^{13}$C values and trend

The white sediment has $\delta^{13}$C values that range from ~3.5‰ 
at the base of the core to nearly ~2‰ near the top (Figure 5). 
An $R$-value of .78 confirms that there is a trend with depth. 
Adjacent data values can vary by nearly 2‰ (Figure 5) (e.g. 
data points at 67.4, 44.1 and 37.4 cm).

Two $\delta^{13}$C values from host rock at 83.6 and 13.2 cm, are 
1.91 and 2.39‰, respectively (solid red circles on Figure 5). 
Both values are generally consistent with the range of carbon 
isotope values from the white sediment (Figure 5). The lower 
sample (83.6 cm) is slightly more negative (~1‰) than that 
of the depth equivalent white sediment. The upper sample 
(13.2 cm) value is slightly more positive (~1 to 2‰) than 
the depth equivalent white sediment.

3.3.2 | $\delta^{18}$O values and trend

Infilled white sediment has $\delta^{18}$O values that range from 
~1‰ near the base, to about ~3‰ at the core top (Figure
5). Similar to the trend in δ¹³C, a linear correlation (R-value of .81) of δ¹⁸O values versus depth was observed.

Two δ¹⁸O values of host rock from 83.6 and 13.2 cm depth, were −3.13 and −3.36‰, respectively (open red circles on Figure 5). These δ¹⁸O values are both lighter than the oxygen-isotope values of white sediment by as much as ~1‰ (Figure 5).

3.4 | SEM crystal morphology

Three distinct crystal forms (equant, needle, bladed/trellis) dominantly occur in SEM images of the white sediment. A sample of sediment at 101 cm depth had mostly equant-type crystals 0.5–1.0 µm in diameter, with some short (2–4 µm) needle-like forms (Figure 6A and B). In places, the needles appeared to consist of segmented crystals or partitions and did not appear to be one single crystal. At ~35 cm, white sediment had mainly needle-like crystals mixed with a lesser amount of equant crystals (Figure 6C and D). The needles are mostly 6–10 µm, but can be up to ~15 µm. Most needles show a rough surface that appears to be partly altered or segmented. Some needle-like crystals appear to have sharp crystal faces and blunt end terminations, as defined by Macintyre and Reid (1992). The uppermost sample at 12.2 cm was also mostly made up of larger needles (5–10 µm), with some up to 15 µm (Figure 6E and F). The needles were mixed with equant crystals of about 1 µm diameter, or smaller. In a few places, the sample contained a cross-hatched pattern of elongate blades in a trellis-like form (~5–7 µm) that was distinct and recognizable relative to other less organized crystals in the matrix. The SEM images also showed a finer (≤1 µm) groundmass of debris that had irregular shape and size.

4 | DISCUSSION

4.1 | Origin of the pores

Pores infilled with white sediment are of a size and shape consistent with dissolution-related vugs. Pores were perhaps also influenced by callianassid (shrimp) burrows as these are common and consistent with 1–3 cm diameter pores in nearby Miami Limestone exposures (Halley & Evans, 1983). The mineralogy of the host rock (~50% aragonite–50% calcite, Figure 4) probably reflects this dissolution and on-going stabilization of primary aragonite to secondary low-Mg calcite by meteoric fluids (Evans, 1982; Evans & Ginsburg, 1987; Halley & Evans, 1983). Within rock of the coastal ridge to the east, aragonite percentages range from 0% to 45% and a small amount of high-Mg calcite is preserved in benthic foraminifera (Evans, 1982; Evans & Ginsburg, 1987).

4.2 | Aragonite mineralogy of white sediment

The high aragonite content of the white sediment within vuggy pores was unexpected given the inland location of the drill site, and the core’s position beneath a freshwater cyanobacterial calcitic mud layer. The only modern local source of aragonite is from freshwater gastropod shells. Two species of gastropod, Pomacea paludosa and Planorbella duryi (Pederson, 2017) produce aragonite in their shell, and the crystal form is typical of molluscan aragonite (dense, lamellar and crossed lamellar). The occurrence of aragonite needles, along with other forms in the white sediment infill, is not consistent with a molluscan source of aragonite. In addition, no shell fragments were found in the infill sediment, and no fragments were identified under SEM.

An excess of 90% aragonite, and a needle-like form found in infill sediment is also inconsistent with aragonite from the host rock. Although Miami Limestone contains aragonite (~50% at core site), aragonite is held in grains: peloids, skeletal grains and a few ooids. If the host rock was the source, one would expect a mineral mix closer to the 50% aragonite and 50% calcite found in the core and a textural form (fragments from ooid cortex) different than large, individual needle-like crystals. The authors are not aware of an example, or how to preferentially weather ooids to produce individual aragonite needles.
Aragonite has been, and is currently, the dominant mineral in Florida Bay marine mud (Scholl, 1966 has averages of 58% aragonite, 27% high-Mg calcite, 15% low-Mg calcite). The infilling white sediment has a trend of decreasing aragonite and increasing low-Mg calcite (from >90% upward to about 75%) from the core bottom to about 20 cm depth (Figure 4). From a depth of 20 cm to the core top the sediment becomes more enriched in low-Mg calcite (Figure 4) representing the onset of freshwater cyanobacterial formation (Enos & Perkins, 1979; Merz, 1992; Pederson, 2017; Scholl, 1964).

4.3 | Texture of infilled sediment

The texture of carbonate material at three depths in the core, and at the surface, show distinct needle, bladed, equant and rhomboid crystal shapes. The SEM analysis of infill sediment confirms an absence of skeletal fragments or other grains (siliciclastic, dolomite rhombs or calcispheres).

Needle forms, of various sizes, are thought to form mainly in a marine environment (Macintyre & Reid, 1995). The occurrence of needle-like crystals at depth (34.8 and 101 cm), is consistent with a predominantly aragonite mineralogy of the infill sediment. The occurrence of large crystals (up to 15 μm) at 34.8 cm depth, some of which have blunt ends, sharp crystal faces and possible prismatic shape, is typical of needles formed in marine algae (Macintyre & Reid, 1992, 1995). Many needle-like forms are less than 4 μm long, and these are crystals probably formed in the inter-utricle spaces of green algae such as *Halimeda* (Macintyre & Reid, 1995).

Equant nannograins (≤2 μm) found in the lowest infill sample at 101 cm have a shape and size consistent with equant crystals (of aragonite) produced from micritization of aragonite needles (Macintyre & Reid, 1995). Equant-shaped crystals occurred in all three SEM samples, but they were especially abundant in the 101 cm sample. Grains at 101 cm were >90% aragonite and are probably anhedral equant aragonite nannograins produced by marine micritization based on their size and shape. Segmented or partitioned surface features (Figure 6C and D) may be evidence of this micritization process, however, crystals were eroded and redeposited before final conversion to anhedral equant forms.

An uppermost infill sample at 12.2 cm core depth has a mixture of mostly large needles, some equant nannograins, and a few trellis-like forms that may be remnants of cyanobacterial freshwater crystals. Freshwater low-Mg calcite
crystals form single-crystal dendrites (Gleason & Spackman, 1974) that consist of a trellis, or cross-hatch array of elongate blades organized in a rhombohedral crystal form (Gleason & Spackman, 1974; Merz, 1992; Pederson, 2017).

### 4.4 Stable-isotope perspective

Stable-isotopic values of infill sediment are used to confirm a marine origin of the mud, especially δ¹³C data. Both δ¹⁸O and δ¹³C values trend upward toward lighter values (Figure 5) and co-vary (R = .88, Figure 5). The isotopic signature is probably a result of mixing between marine mud and a progressive influx of freshwater cyanobacterial (low-Mg calcite) mud. This mixing shows scatter above and below the best-fit line (Figure 5) and is likely due to either different proportions of marine sediment and freshwater sediment, a change in source area (isotopic composition) of infill mud, or perhaps some degree of diagenesis (Walter, Bischof, Patterson, & Lyons, 1993).

The δ¹³C values are mainly dominated by a marine lime mud signature (3–0.5‰), similar to bulk values found in Florida Bay mudbanks and islands (Andrews, 1991; Swart et al., 1989). Isotopic values are also generally consistent with living Halimeda values and Halimeda sediment values that range from 5 to −1‰ (Walter et al., 1993). Brackish and freshwater mud are typically more negative and range from about 0.5 to −3‰ (Andrews, 1991; Merz, 1992; Pederson, 2017) (Figure 7). The profile of infill sediment would represent mainly marine mud from the core base at 101 cm up to about 35 cm, a depth where several δ¹³C values are similar to freshwater values (more negative than 0.5‰). Fluctuation in δ¹³C values suggests that multiple infill events probably occurred, with sediment of different sources and isotopic composition. The contribution of freshwater calcite becomes dominant above 20 cm depth.

A δ¹⁸O profile of infill, although trending from less negative values to more negative values upward, has a relatively narrow range (−1.5‰) below 20 cm (depth where aragonite becomes dominant). An aragonite mineralogy and original, well-preserved marine texture (needles) would preclude significant in situ diagenesis by freshwater (see selective diagenesis comments above). Most infill values are more negative (−1 to −3‰) than bulk values of marine mud on Florida Bay mudbanks (Cross Bank, most values −0.2 to −0.5‰) and on islands (Crane Key, most values −0.5 to −0.9‰) (Andrews, 1991; Swart et al., 1989) (Figure 7). The δ¹⁸O values are, however, generally consistent with Halimeda values (−1 to −2.2‰) reported by Walter et al. (1993) for Florida Bay and the Florida shelf. Similarly, Milliman (1974) gives a range of δ¹⁸O values for green algae of between 1 and −3‰. Isotopic values of benthic calcareous algae are also complicated by metabolic processes in their precipitation that is tied to different growth stages and possible environmental effects (Wefer & Berger, 1981). A slight up-core trend in δ¹⁸O toward more negative values might partly represent a change from cooler to warmer water temperature (Epstein, Buchsbaum, 

**FIGURE 7** Stable-isotope profiles of infilled sediment (solid black circles), the same data as in Figure 5 but with isotopic ranges of nearby freshwater and marine sediments. (A) δ¹³C values of infilled sediment versus depth in core ENP-2. Range of δ¹³C values from nearby marine, brackish and freshwater carbonate sediment are shown as horizontal bars. (B) δ¹⁸O values of infilled sediment versus depth in core ENP-2. Range of δ¹⁸O values from nearby marine, brackish and freshwater carbonate sediment are shown as horizontal bars.
Lowenstam, & Urey, 1953). The observed ~1‰ δ^{18}O more negative shift would equate to a ~4°C temperature change. Perhaps such warming was possible as Florida Bay was flooded and the developing mudbanks restricted tidal circulation from contact with cooler open-water sources. It seems unlikely, however, that this signal is fully temperature related over such a relatively short time window and the authors believe it more probably represents variation in source material and source mixing as discussed above.

4.5 | Transgression and 3,200 BP shoreline position

The Holocene transgression on the southern Florida peninsula, near present-day Florida Bay, occurred in a series of distinct stages on seaward-inclined Pleistocene limestone (Wanless & Tagett, 1989). Basal freshwater peat and freshwater calcitic muds were deposited from about 5,000 to 3,500 BP as sea-level transgressed the limestone. Sea-level rose at a rate between 3 and 5 mm/year (Wanless, Parkinson, & Tedesco, 1994). Between 3,500 and 3,000 BP Wanless et al. (1994) document a rising sea level that overstepped shoreline deposits and existing coastal levees. Mudbanks were initiated on these drowned coastal levees between about 3,300 and 2,500 yr, and layered mudstone was deposited in flooded parts of western Florida Bay. By ~3,200 yr the shoreline position during this transgression was about 5 km landward of the modern shoreline and coastal levee and approximately 10 km south of the ENP-2 core site. A combination of southward sloping limestone and absence of Holocene sediments allowed sea-level transgression to this inland position. By 1,500 yr a slower relative sea-level rise (0.4 mm/year, Wanless et al., 1994) coupled with sufficient sediment production was able to maintain existing mudbanks and develop interbank areas (called "lakes") of Florida Bay. This slower sea-level rise probably allowed formation of today's coastal levee and >8 km-wide fringe of mangroves. The ~5 km-wide zone between the 3,200 yr inland shoreline and modern coastal levee was subsequently infilled with mangroves and brackish water lakes to form the modern shoreline configuration (Wanless & Tagett, 1989; Figure 1). Today's levee and mangroves effectively block inland transport of marine sediment, consistent with observations made during several recent hurricanes (Irene 1999; Irma 2017; Wilma 2005).

A 10 km distance, between the 3,200 yr shoreline and the location of core ENP-2 was the closest occurrence of in situ marine deposits. Thus, mud-size aragonitic marine infill sediment was transported at least 10 km inland, and probably further, from subtidal marine sediment producing areas of Florida Bay to the ENP-2 site.

Sea level at ~3,200 BP was estimated to be 1.5–2 m lower than today (Robbin, 1984; Scholl, 1964; Scholl, Craighead, & Stuiver, 1969). However, between 3,200 and 2,500 yr, the absence of a mangrove fringe and coastal levee could have allowed storm surges to more easily penetrate inland and transport suspended sediment landward.

4.6 | Model of sediment transport and infiltration

Storm surge is the probable mechanism for onshore transport of fine-grained sediment. Hurricanes and tropical storms would be strong enough to erode, suspend and transport substantial amounts of sediment (see below). Winter storms associated with passing cold fronts are not sufficiently strong enough to transport sediment-laden sea water that far inland (10 + km). Although cold front winds progress clockwise from their normal south-easterly direction, to a northern or north-eastern direction, the strongest winds are incapable of such inland surge and sediment transport (based on modern observations of cold front effects).

Hurricane force winds affect the southern Florida peninsula on an average of once every seven years and produce storm surges that can exceed 4 m (Ball, Shinn, & Stockman, 1967; Wanless & Tagett, 1989; Warzeski, 1976). For example, during Hurricane Donna in 1960, Ball et al. (1967) described erosion, suspension, transport and redeposition of Florida Bay mud. They noted that ebb storm surge left large amounts of layered lime mud stranded on supratidal areas within the Bay. It was reported by Ball et al. (1967) that Hurricane Donna deposited lime mud 8 km inland on the south end of the Florida mainland. However, the mud was unfortunately not confirmed to be marine in origin, especially because local freshwater cyanobacterial mud can be similarly eroded and redeposited on inland areas. For example, freshwater calcitic mud accumulations were observed after Hurricane Irene in several areas north of the mangroves near core ENP-2 and near U.S. Highway 1. Perkins and Enos (1968) compared the effects of Hurricane Betsy to that of Hurricane Donna and noted that supratidal sedimentation was considerably less than that reported for Donna (Ball et al., 1967). Although hurricanes are highly variable in strength and wind direction, they are usually sufficient to erode, suspend and mobilize marine mud.

Between ~4,000 BP and slowing of sea-level rise at ~1,500 BP (Wanless et al., 1994), and before development of coastal levee and mangroves, a 2,500 year period exists for tropical storms to mobilize sediment and for a surge to emplace it inland. If an average storm period of seven years is used (Ball et al., 1967), nearly 300 tropical storms and hurricanes could have potentially transported the marine mud to the ENP-2 site. Support for mud suspension and redeposition also comes from the Stockman, Ginsburg, and Shinn (1967) study of lime mud production, where they proposed that marine mud is widely mobilized, such that it might even have been imported into Florida Bay from the shelf west and south-west of Florida Bay during tropical storms. A viable model for deposition of aragonitic mud in large pores of subaerially exposed Pleistocene limestone is through
storm surge deposition (Figure 8). A windblown mechanism would not have been viable if we use today for comparison, with a lack of wind suspended aragonite (Prospero, Landing, & Schulz, 2010). Aragonite mud in Florida Bay is mostly subtidal, with relatively small amounts in the supratidal zone. The supratidal mud is either bound in microbial mats, too cohesive, or too indurated when dry, so as to be easily suspended. The infiltration model proposed here envisages formation of an open pore system through subaerial-related dissolution and meteoric stabilization of aragonite to low-Mg calcite. Prior to establishment of the modern cyanobacterial mud production, tropical storm surges were able to transport suspended aragonite sediment landward. This muddy storm layer then migrated downward into Pleistocene rock by gravity infiltration as storm water flowed across the limestone, or perhaps later after deposition, helped by percolating rainfall. An upward progression from nearly pure aragonite to more calcite probably records the establishment of the palustrine cyanobacterial mud deposits as the water-table rose and periods of standing water occupied areas around the ENP-2 site.

4.7 | Uncertainties and unanswered questions of the model

The aragonite and low-Mg calcite mineralogy of the sediment is somewhat puzzling given the mineralogy found today in Florida Bay; the two minerals mentioned above plus high-Mg calcite (Bosence, 1989; Bosence, Rowlands, & Quine, 1985; Stockman et al., 1967; Taft & Harbaugh, 1964). The mud production from codiacean green algae (Stockman et al., 1967) and from Thalassia epibionts (Bosence, 1989; Nelson & Ginsburg, 1986) produce most of the modern mud-size aragonite and high-Mg calcite. There is, however, some partitioning in the distribution of aragonite and high-Mg calcite sediment in Florida Bay. For example, epibions on mudbanks can have an aragonitic suite of encrusters, or a mixed aragonite and high-Mg calcite suite of encrusters (Bosence, 1989). During a tropical storm, these different mineralogy-type grains are likely mixed.

Sediment mineralogy (<62 µm fraction) of cores from a Holocene mudbank (Cross Bank) and island (Crane Key) show a generally consistent mix of high-Mg calcite and aragonite (Andrews, 1991). For Cross Bank, the mix is about 50% high-Mg calcite, 40% aragonite, with the remainder consisting of low-Mg calcite. For Crane Key, the mineralogy consists of about 60% high-Mg calcite, 30%–35% aragonite, with the remainder comprised of low-Mg calcite. This stability through time (up core), suggests that even the earliest sedimentation in Florida Bay had a source of mixed aragonite and high-Mg calcite.

The main question is then, how is it sediment that infills Pleistocene rock at ENP-2 is just aragonite and low-Mg calcite, and absent of high-Mg calcite? One can only speculate that a high-Mg calcite absence may be either related to source (one of mainly needle aragonite from green algae), selective sorting of storm suspended material, or selective diagenesis. A selective source mechanism may have to do with where storm suspended sediment originates. Florida Bay is 10% mud banks and 90% basin “lakes” by area (Enos & Perkins, 1979; Stockman et al., 1967). These “lakes” have a higher standing crop of green algae such as Penicillus sp., and combined with their extensive area, may favour an aragonitic mineralogy for the suspended sediment. Much unbroken epibiont material is larger (most >50 µm) than infill mud, plus it does not produce needle-shaped crystals. The <10–15 µm size crystals of infill sediment
would suggest suspension winnowing to a fine-grained mud, preferentially excluding some of the larger high-Mg calcite grains. This disparity in grain size might also be important in the downward movement of the marine sediment. Similar to geopetal fills, perhaps only small mud-size grains (aragonite needles) are able to infiltrate downward through small pores and pore throats, thus effectively filtering out large high-Mg calcite grains. Some combination of size and source, may perhaps, restrict infill sediment to predominantly marine aragonite.

A third mechanism is selective diagenesis. Solubility of high-Mg calcite is partly dependent on the mole percent MgCO$_3$. Magnesian calcites with about 12% or greater mole percentage of Mg$^{+2}$ are more soluble than aragonite (Morse & Mackenzie, 1990). High-Mg calcite recorded from Cross Bank and Crane Key cores (Andrews, 1991) show most values (29 of 33) at between 12 and 14 mole percent. These MgCO$_3$ values would make most high-Mg calcite more soluble than aragonite. This solubility difference makes selective dissolution of high-Mg calcite grains in the infill sediment feasible. Some amount of freshwater would be required to move through the mud-sized infill sediment. Any evidence of dissolved high-Mg calcite has obviously been lost, so testing this hypothesis is difficult. Perhaps a much more extensive infill sediment analysis (from additional cores) might provide evidence for a selective dissolution mechanism.

5 SUMMARY AND IMPLICATIONS

Marine aragonite, mostly needles (<15 µm) and smaller equant grains, have infilled vugs in Pleistocene limestone down to a depth of about 1 m. The infilled sediment has a composition, grain size and mineralogy distinct from host rock material and modern microbial low-Mg calcite that forms at the site today. In core, the highest amount of aragonite (>90%) infill is at the base, and mineralogy transitions upward to about 75% aragonite at 20 cm depth. The absence of high-Mg calcite, that today forms with aragonite in Florida Bay, is puzzling. A tropical storm mechanism is proposed for suspension and transport of fine-grained aragonite at least 10 km landward of the sediment source. Once deposited, aragonite mud infiltrates downward into open and connected pores previously formed by dissolution of host Miami Limestone. This emplacement of muddy aragonite during the transgression, differs from Dunham’s (1969) model of internal silt-size sediment formed in the vadose zone upon early emergence (from marine conditions).

Infilled sediment effectively moves across a subaerial exposure surface, a sequence boundary, to infiltrate porous rock below. Although in this case no microfossils were recognized in Holocene infill sediment, this process may have implications if suspended sediment is microfossil rich. Infilled sediment may also decrease effective porosity and permeability across these formational contacts. Such porosity decreases may allow groundwater to perch on these boundaries, eventually affecting the local diagenetic processes such as meteoric dissolution or cementation. Infiltration sediment that crosses depositional boundaries can produce a complex sedimentological mix of relatively finer grains within the coarser grained matrix of the host rock. For example, the cut surface of core ENP-2 was 17.3% (by area) infilled mud. Infiltrated sediment, similar to Dunham’s (1969) diagenetic silt, might aid in the recognition of subaerial exposure. The occurrence of marine sediments landward of the shoreline and stratigraphically below freshwater deposits may, however, be interpreted a result of a rapid rise and fall of sea level.

Preservation of such muddy infill sediment and recognition in ancient limestone is dependent on the diagenetic pedigree. Aragonite will be recrystallized, but the muddy texture may be preserved within a diagenetic setting with a low water-to-rock ratio. An example of such a scenario, perhaps, is well-preserved muddy sediment in geopetals of ancient limestones.

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CONFlict OF INTEREST

The authors have no conflict of interest to declare.

DATA AVAILABILITY STATEMENT

The data for this study are shown in the various figures. These data are available in tabular form upon request from the corresponding author.
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