A REEVALUATION OF FACIES ON GREAT BAHAMA BANK II: VARIATIONS IN THE $\delta^{13}$C, $\delta^{18}$O AND MINERALOGY OF SURFACE SEDIMENTS

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Abstract

In order to investigate the spatial distribution of $\delta^{13}$C and $\delta^{18}$O of modern carbonate sediments on Great Bahama Bank, ~290 surface samples were collected from a grid of stations approximately 10 km apart between 2001-2004. These samples were classified using a modified Dunham scheme, physically separated into six size fractions and subsequently analyzed for their mineralogy (aragonite, low-Mg calcite and high-Mg calcite) and $\delta^{13}$C and $\delta^{18}$O values. A striking feature of these data is the relatively positive $\delta^{13}$C values of all the samples. Based on measurements of $\delta^{13}$C and $\delta^{18}$O of the
dissolved inorganic carbon and the water, most of the sediments can be considered to be in C and O isotopic equilibrium with the ambient waters. The high $\delta^{13}C$ values are suggested to arise from isotopic enrichment of the dissolved inorganic carbon pool by photosynthesis of sea grasses, benthic algae and cyanobacteria on the platform and through the fractionation of $\text{HCO}_3^-$ during the precipitation of calcium carbonate. Sediments which are not in C and O isotopic equilibrium are dominated more by skeletal material. The data showed an absence of significant spatial variation in $\delta^{13}C$ of the sediments on the Great Bahama Bank and no clear spatial patterns relative to the margin of the platform. The $\delta^{18}O$ of the sediment showed more variation with the interior sediments being isotopically enriched relative to the platform margin. The absence of significant variations in the $\delta^{13}C$ in the modern surface sediments of Great Bahama Bank irrespective of facies type, suggests that in the case of Great Bahama Bank downcore variations in $\delta^{13}C$ cannot be related to changes in facies.

INTRODUCTION

The descriptions of Great Bahama Bank (GBB) and its surface sediments (Enos, 1974; Illing, 1954; Newell et al., 1959; Purdy, 1963a; Purdy, 1963b; Traverse & Ginsburg, 1966) made between 1950 and 1970 are widely used for modern biological and geological studies on Great Bahama Bank (GBB). In addition, GBB is often applied as an analogue for the study of ancient carbonate platforms. This paper together with a companion study (Reijmer et al., this volume) examines spatial variations in the facies and geochemistry of surface sediments on Great Bahama Bank. Although there have been several papers which provided some general geochemical characteristics on the sediments on GBB (Lowenstam & Epstein, 1957; Shinn et al., 1989) there have been no studies which have described spatial variation in the inorganic and organic mineralogy of the sediments on the same scale as the original sedimentary facies were investigated. Such data are important as changes in the $\delta^{13}C$ of platform and periplatform carbonates are increasingly being used for stratigraphic purposes (Saltzman et al., 2004; Vahrenkamp, 1996) and for providing information on the global $\text{CO}_2$ cycle in periods prior to the existence of pelagic records (Berner, 1987; Kump & Arthur, 1999). It is well
documented that global changes in the $\delta^{13}C$ of platform carbonates can be correlated between various strata during certain periods of the geological record (Föllmi et al., 1994; Immenhauser et al., 2002; Kump & Arthur, 1999; Saltzman, 2002a; Saltzman, 2002b; Saltzman et al., 2004; Saltzman & Young, 2005; Vahrenkamp, 1996; Valladares et al., 1996; Veizer et al., 1999; Weissert et al., 1998) as well as related to the pelagic record of $\delta^{13}C$ change (Gale, 1993; Weissert et al., 1998). Such variations are normally interpreted as reflecting changes in the partitioning of the global carbon cycle, with elevated values indicating the increased burial of organic carbon (Berner et al., 1983; Kump & Arthur, 1999). However, it is well established that changes in $\delta^{13}C$ can also reflect diagenesis (Allan & Matthews, 1982; Immenhauser et al., 2003) as well as variations in the relative proportion of periplatform and pelagic material (Reuning et al., 2006; Swart & Eberli, 2005). In this paper the origin of the $\delta^{13}C$ and $\delta^{18}O$ and mineralogy of ~290 samples collected from Great Bahama Bank on a regular grid spacing of ~10 km between 2001 and 2004 are reported.

SAMPLES

Samples were collected using a Shipek sampler during four cruises aboard the RV Bellows between 2001 and 2004 (Figure 1). The Shipek sampler differs from the Van-Veen sampler used during earlier studies (Purdy, 1963a; Purdy, 1963b) in that it is better at retaining the mud material. In total 291 bulk samples were analyzed. Of these approximately 120 were separated into size fractions (> 1000, 500-1000, 250-500, 125-250, 125-63 and < 63 μm) and the relative percentages of aragonite, high-Mg calcite (HMC) and low-Mg calcite (LMC) were determined together with their carbon and oxygen- isotopic compositions. Each sample was also separated into a modified Dunham facies classification. At the same time water samples were taken for salinity and stable oxygen and hydrogen measurements. Samples for carbon isotopic analysis of the dissolved inorganic carbon (DIC) were filtered using a 1 μm filter and preserved in sealed crimped vials with excess mercuric chloride.
METHODS

X-Ray Mineralogy
The percentages of aragonite, HMC and LMC were determined using the method of Swart & Melim (2000). In this method the sample is assumed to be completely composed of aragonite, HMC and LMC. Areas of the appropriate peaks for each mineral are determined using a scan between 23 and 32° 2θ (CuKα radiation). These are then compared to standard relationships between peak area and percentage mineral and the percentage of the mineral in the sample determined.

Stable Isotopes
Carbonates
The δ¹³C and δ¹⁸O of the carbonate materials was determined using dissolution in phosphoric acid using the common acid bath method (Swart et al., 1991) at 90°C. The gas produced was analyzed using a Finnigan-MAT 251. Later some samples were also analyzed using a Kiel III device attached to a Finnigan Delta plus. Data produced in both methods were corrected for isobaric interferences using the procedures (Craig, 1957) modified for a triple collector mass spectrometer. Data are reported relative to Vienna Pee Dee Belemnite (V-PDB) using the conventional notation. Average standard deviation based on replicate analyses of internal standards is < 0.1‰.

Waters
The oxygen and hydrogen-isotopic composition (δ¹⁸O and δD) were determined using the methods of Epstein & Mayeda (1953) and Coplen et al (1991) respectively. All δ¹⁸O and δD data for waters are reported relative to Vienna Standard Mean Ocean Water (V-SMOW). Average standard deviation, based on replicate analyses of internal standards is < 0.1‰ for δ¹⁸O and 2.0‰ for δD. The δ¹³C of the dissolved inorganic carbon (DIC) was determined by acidification of the sample followed by extraction using a flowing stream of He and analysis in a stable isotope ratio mass spectrometer (Europa 20-20). Data are reported relative to V-PDB using the conventional notation. Average standard deviation based on replicate analyses of internal standards is < 0.1‰.
Facies Classification
Immediately after collection, samples were assigned to a modified Dunham scheme (Dunham, 1962). In order to supplement the classification divisions were added between wackestone and mudstone (mud-rich wackestone), between a packstone and wackestone (mud-rich packstone) and finally between packstone and grainstone (mud-lean packstone). For display purposes the mudstone, wackestone, packstone, grainstone and rudstone classification were assigned numbers from 1 to 5, with the intermediary categories being assigned 1.5, 2.5 and 3.5 numbers. Samples from all cruises were preserved in cold storage and after the 2004 cruise all samples were re-examined to ensure consistency between cruises (Reijmer et al, this volume).

RESULTS
Stable Isotopes
Carbonates
The mean $\delta^{13}$C and $\delta^{18}$O values for the carbonate sediments and the various size fractions are shown in Table 1 and 2 and Figure 2. The $\delta^{18}$O of all fractions are statistically the same, but the $\delta^{13}$C of the coarser material (> 1000 μm) is statistically significantly more negative than the finer fraction and the bulk. Relative to the modified Dunham classification there is no statistically significant difference between the mean $\delta^{13}$C and $\delta^{18}$O of the coarser and finer sediment types (Table 3; Figure 3 and 4). However, the range of values of the modified Dunham categories increases with grain size (Table 3).

<table>
<thead>
<tr>
<th>Table 1: Carbon Isotopic composition of size fractions</th>
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</thead>
<tbody>
<tr>
<td>Bulk</td>
</tr>
<tr>
<td>Mean</td>
</tr>
<tr>
<td>s.d.</td>
</tr>
<tr>
<td>Min</td>
</tr>
<tr>
<td>Max</td>
</tr>
</tbody>
</table>
Table 2: Oxygen Isotopic composition of size fractions

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<th>Size Fraction</th>
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<th>&gt;1000 μm</th>
<th>1000-500 μm</th>
<th>500-250 μm</th>
<th>250-125 μm</th>
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<tr>
<td>Mean</td>
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<td>+0.3</td>
<td>+0.4</td>
<td>+0.5</td>
<td>+0.4</td>
<td>+0.5</td>
<td>+0.4</td>
</tr>
<tr>
<td>s.d.</td>
<td>0.4</td>
<td>0.4</td>
<td>0.3</td>
<td>0.4</td>
<td>0.44</td>
<td>0.4</td>
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</tr>
<tr>
<td>Min</td>
<td>-0.8</td>
<td>-0.5</td>
<td>-0.5</td>
<td>-0.40</td>
<td>-0.3</td>
<td>-0.3</td>
<td>-0.5</td>
</tr>
<tr>
<td>Max</td>
<td>+1.8</td>
<td>+1.5</td>
<td>+2.0</td>
<td>+2.0</td>
<td>+1.7</td>
<td>+1.3</td>
<td>+2.0</td>
</tr>
</tbody>
</table>

Table 3: The $\delta^{13}$C and $\delta^{18}$O in different modified Dunham classification

<table>
<thead>
<tr>
<th>Facies Description</th>
<th>1.5</th>
<th>2</th>
<th>2.5</th>
<th>3</th>
<th>3.5</th>
<th>4</th>
<th>5</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Mud-rich wackestone</td>
<td>Wackestone</td>
<td>Mud-rich packstone</td>
<td>Packstone</td>
<td>Mud-lean packstone</td>
<td>Grainstone</td>
<td>Rudstone</td>
</tr>
<tr>
<td>$\delta^{13}$C</td>
<td>+4.8</td>
<td>+4.8</td>
<td>+4.8</td>
<td>+4.8</td>
<td>+5.0</td>
<td>+4.9</td>
<td>+3.7</td>
</tr>
<tr>
<td>s.d.</td>
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<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.3</td>
<td>0.3</td>
<td>0.1</td>
</tr>
<tr>
<td>$\delta^{18}$O</td>
<td>+0.5</td>
<td>+0.6</td>
<td>+0.6</td>
<td>+0.6</td>
<td>+0.4</td>
<td>+0.2</td>
<td>-0.1</td>
</tr>
<tr>
<td>s.d.</td>
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<td>0.4</td>
<td>0.4</td>
<td>0.3</td>
<td>0.5</td>
<td>0.4</td>
<td>0.2</td>
</tr>
</tbody>
</table>

Stable Isotopes

Water

The $\delta^{18}$O of the waters are shown in Figure 5 and range between $+0.75$ and $+2.93$‰. There is an approximate positive correlation with salinity, similar to that shown previously (Lowenstam & Epstein, 1957; Shinn et al., 1989). The $\delta^{13}$C values of the DIC range from $+0.4$ to $+2.4$‰ and show no specific relationship with salinity (Figure 6).

Sediment Classification

The descriptions and distribution of these facies are fully described in a companion paper (Reijmer et al., 2007). The majority of the sediments are dominated by non-skeletal material (mud, pellets, peloids, ooids and grapestones), although all samples contained small amounts of skeletal debris. The percentage of mud in the samples increases rapidly moving away from the bank margin towards the interior of the platform with samples
completely composed of skeletal material only present near the bank margin. The facies
distribution patterns found are fairly similar to that shown in the facies distribution map
of Purdy (1963a; 1963b) and roughly agree with the map of Traverse and Ginsburg
(1966).

DISCUSSION

Previous Work on the Variation of $\delta^{13}$C on Modern Carbonate Platforms

There have been relatively few previous studies on the distribution of $\delta^{13}$C of sediments
on modern carbonate platforms (Dix et al., 2005; Lowenstam & Epstein, 1957; Shinn et
such as the one carried out at Heron Island (Weber & Woodhead, 1969) examined
variations over a relatively small area, a coral atoll approximately 5 x 20 km. The Heron
Island study showed substantial variation in $\delta^{13}$C progressing across the reef, with the
margins being significantly more negative (up to 2‰) in $\delta^{13}$C compared to portions of the
interior (Figure 10). These changes are strongly correlated with the percentage of
aragonite and arise because carbonate materials produced from the decomposition of
algae, such as the various calcareous green algae commonly found on coral reefs
(Halimeda sp., Penicillus sp., Acetabularia sp. and others), have very positive $\delta^{13}$C
values (typically > +4‰). In contrast, the $\delta^{13}$C of most LMC and HMC components is
close to or below 0 ‰ (Keith & Weber, 1965; Land, 1989; Milliman, 1974; Weber,
1965). Hence a mixing line between aragonite and LMC or HMC will produce a co-
varying trend. However, a close inspection of these data show two apparent trends, one
at concentrations of aragonite higher than 60% aragonite where there is a strong positive
association between percentage aragonite and $\delta^{13}$C and one at concentrations below 60%
where there is apparently no correlation between the percentage of aragonite and $\delta^{13}$C
(Figure 10). This apparent difference probably arises because there are at least two
sources of aragonite with very different $\delta^{13}$C values on GBB. One source is the
calcareous green algae mentioned above. The other source is the aragonite produced by
scleractinian corals and red calcareous algae which possess much lower $\delta^{13}$C values (-1
to +1‰) (Keith & Weber, 1965; Land, 1989; Milliman, 1974). Hence, increasing the concentration of aragonite by adding material derived from corals and red calcareous algae will not produce a correlation between the concentration of aragonite and δ^{13}C. A similar result was found at Enewetok, with relationships between aragonite and δ^{13}C being either positive or negative depending upon the location sampled (Weber & Schmalz, 1968). Another significant study which examined a much larger spatial area was that by Dix et al. (2005). Here workers examined approximately 100 samples collected over an approximate 600 km² area on the North West Australian Shelf and also identified a positive relationship between the percentage of aragonite and δ^{13}C, again produced by a mixing between isotopically positive aragonite and more depleted LMC. A similar explanation has been invoked for different correlations between C and O found in modern sediments collected from Belize and the Persian Gulf (Gischler et al., 2007).

RELATIONSHIPS ON GREAT BAHAMA BANK.

Salinity and Stable Isotopes

There have been no systematic temporal or spatial investigations of the water chemistry of Great Bahama Bank, but the analyses which have been made, show waters increasing in salinity as one progresses across Great Bahama Bank from west to east (Black, 1933; Cloud Jr, 1962; Lowenstam & Epstein, 1957; Shinn et al., 1989; Smith, 1940b). Very close to Andros Island, salinities have been shown to decrease as a result of runoff (Cloud Jr, 1962).

Oxygen-isotopic analyses of the waters generally show a positive relationship with salinity as do the data measured in this study (Lowenstam & Epstein, 1957; Shinn et al., 1989) (Figure 5). However, as a result of exchange between atmospheric water vapor and the water on GBB, the maximum δ^{18}O which can be attained is limited (Craig & Gordon, 1965; Gonfiantini, 1986; Lloyd, 1964) and therefore there is no clear relationship between salinity and δ^{18}O. In fact the maximum δ^{18}O value measured in this study was ~+2.8 ‰ at a salinity of 35.75. Waters with higher salinity values had actually lower δ^{18}O values. The δD data also show clear evidence of a strong evaporative trend
(Figure 7) with original source water close to 0‰. As the samples were collected over a four year time period during which the salinity and \( \delta^{18}O \) of the water varied as a result of weather conditions, it is not possible to combine the data and make a generalization about the distribution of \( \delta^{18}O \) on GBB. However, it is probable that waters further from the margin might have elevated \( \delta^{18}O \) values on a consistent basis. Hence carbonates formed further from the margin might also be expected to have higher \( \delta^{18}O \) values than carbonates found on or near the margin.

Previous studies on the \( \delta^{13}C \) of waters on GBB are limited to one which showed large ranges in \( \delta^{13}C \) ranging from ~ -5 to +1 ‰ (Patterson & Walter, 1994). It was suggested that these variations were related to oxidation of organic material, photosynthesis and precipitation of aragonite and LMC. The \( \delta^{13}C \) of the DIC measured in this study were significantly higher and varied from ~+0.5 to +2.5‰ (Figure 4). In no instance over five years of analyses were waters with \( \delta^{13}C \) values less than +0.5 ‰ analyzed. The lower values measured in this study are typical of open marine values (Kroopnick, 1974; Swart et al., 2005; Weber & Woodhead, 1971), while the higher ones suggest both fractionation of CO\(_2\) during photosynthesis and/or fractionation of the HCO\(_3^-\) pool as CO\(_2\) is produced during the precipitation of CaCO\(_3\). Photosynthesis by the abundant algae and sea grasses on GBB, preferentially remove the lighter isotope of carbon causing the residual bicarbonate pool to become enriched in the heavier isotope of carbon, while precipitation of calcium carbonate lowers the pH converting HCO\(_3^-\) to CO\(_2\). As there is an isotopic fractionation of approximately 8 ‰ in this process (Emrich et al., 1970; Romanek et al., 1992), the residual bicarbonate becomes isotopically enriched.

Sediments
Perhaps the most startling results arising from this study are (i) the absence of significant patterns in the \( \delta^{13}C \) across the GBB (Figure 3) and (ii) the lack of a relationship between facies type and \( \delta^{13}C \) (Figure 3). The only exception to this latter pattern is that very large grains (> 1000 μm) are slightly more negative (+4.2 vs. +4.9‰). The relatively depleted values are probably a result that these grains are more likely to be composed of skeletal
material which generally have lower $\delta^{13}C$ values (See previous discussion). The origin of relatively high $\delta^{13}C$ values of the sediments compared to pelagic LMC arises from two phenomena. First, as mentioned above the $\delta^{13}C$ of the DIC averages +1.5‰ ranging from +0.4 to +2.4‰. Second, enrichment of $\delta^{13}C$ in aragonite is about 2.7 +/- 0.6‰ relative to only about 1‰ for LMC (Romanek et al., 1992). Hence equilibrium values for aragonite precipitated directly from surface waters of GBB should range from ~ +2.6 to ~ +5.8‰ (this range is calculated using the error of 0.6‰ and the $\delta^{13}C$ values measured on the surface water DIC (Figure 8). This corresponds to a measured range of +3.6 to +5.5 ‰ (See Table 1 and 2). As regards the $\delta^{18}O$, the variation in water temperatures on GBB probably exceeds 10°C, but averages 25°C. Using the standard paleotemperature equation (Epstein et al., 1953) and the values reported in Table 2, the sediments if formed at equilibrium should lie between -1.2 and +0.8‰. This is a little more negative than the range of the sediments (-0.8 to +1.8‰) but still within the errors of the temperature estimate (Figure 7). So if the predominantly aragonite sediments on GBB are in equilibrium with the ambient waters, then why do all the sediments, from the mud dominated wackestones to the grainstones have similar $\delta^{13}C$ and $\delta^{18}O$ compositions? The simple answer is that the majority of the sediments are genetically related; the exception being the small amount of skeletal material mentioned previously.

The origin of the finest carbonate sediment on GBB has been the subject of speculation for over 50 years (Black, 1933; Cloud Jr, 1962; Smith, 1940a; Smith, 1940b). Authors have proposed that the material is either a result of direct or algal induced precipitation from seawater (Cloud Jr, 1962; Morse et al., 1984; Robbins & Blackwelder, 1992; Shinn et al., 1989), or that they are stirred up bottom sediments, ultimately originating from the decomposition of calcareous algae (Broecker & Takahashi, 1966; Morse et al., 1984; Stockman et al., 1967). This debate has been heated over the years, but regardless of the ultimate origin of the mud component, this fine grained sediment seems to be the building material for most of the other sedimentary particles on GBB. Mud becomes ingested by benthic organisms forming fecal pellets which harden to form peloids, some of which become coated to form ooids or clumped together to form grapestones. This process is supported by the fact that the finest sediments have the narrowest range of C
and O isotopic compositions (Figure 2a), gradually increasing in range as diagenetic and cementation processes add carbonate material of differing generations and isotopic compositions to the grains.

**RELATIONSHIPS BETWEEN $\delta^{13}C$ and $\delta^{18}O$**

The relationships between the $\delta^{13}C$ and $\delta^{18}O$ in the bulk samples and the separated size fractions are shown in Figures 2a and 2b. In $\delta^{13}C$ and $\delta^{18}O$ space, the data defines a triangle with the samples possessing the most positive $\delta^{18}O$ values also containing the narrowest range and the most elevated $\delta^{13}C$ values. These samples tend to be mud-rich wackestones. This pattern probably reflects the relationships between those areas on GBB where the water consistently possesses the highest $\delta^{13}C$ (DIC) and $\delta^{18}O$ values (Figure 11). These areas are probably fairly restricted in geographical extent. In contrast, the areas where there are wider ranges in $\delta^{13}C$ and $\delta^{18}O$ are more prevalent and account for the majority of the samples. The largest sized samples (500-1000 $\mu m$ and $> 1000 \mu m$) have a much shallower relationship between $\delta^{13}C$ and $\delta^{18}O$ reflecting a larger proportion of skeletal material in the samples (Figure 11).

**IMPLICATIONS FOR CARBON ISOTOPE STRATIGRAPHY**

This study on the $\delta^{13}C$ and $\delta^{18}O$ of surface sediments of GBB show that in spite of a large range in sediment sizes, there is a relatively small spread in the $\delta^{13}C$ and $\delta^{18}O$ values. The elevated $\delta^{13}C$ values are a result of a combination of isotopic fractionation of the DIC during photosynthesis and the natural enrichment of aragonite relative to LMC (see above). These sediments are exported from the platform and form periplatform sediments (Eberli, 2000; Eberli et al., 1997; Schlager & Ginsburg, 1981) which drape the flanks of the platform and extend into the pelagic realm. It has been shown that the $\delta^{13}C$ of such sediments can be correlated between units of equivalent age, but that these $\delta^{13}C$ values are not related to pelagic $\delta^{13}C$ values (Swart & Eberli, 2005). The reason for the correlation between units of similar age arises as varying amounts of surface sediments
enriched in δ^{13}C, mix with more depleted carbonate derived from pelagic sources. As all sediment size fractions have similar δ^{13}C values, the mixing does not appear to change as a function of distance from the platform. This pattern may not be present surrounding carbonate buildups with high percentages of aragonite derived from calcareous organisms such as corals. The question posed by this study is whether the results are at all applicable to carbonate platforms in previous geological time periods. For example, it has been proposed that the normal form of calcium carbonate precipitated from marine seawater has changed throughout geological time as a function of the Mg/Ca ratio of the oceans varying between ‘calcite’ and ‘aragonite’ seas (Hardie, 2003; Lowenstein et al., 2003; Sandberg, 1983). During periods of ‘calcite’ seas, clearly the δ^{13}C of carbonates formed on the platform would probably not be as isotopically positive as now since the only mechanisms which would enrich the platform carbonates would be the influences of photosynthesis and fractionation during the precipitation of calcium carbonate. The platform derived carbonate would nevertheless be more isotopically positive than pelagic material formed at the same time, although not as enriched as today. Hence it is proposed that there still might be a carbon isotopic effect as a result of precipitation on carbonate platforms during periods of ‘calcite’ seas, but that this might be reduced compared to periods of ‘aragonite’ seas.

**CONCLUSIONS**

1- The positive δ^{13}C values of surface sediments arises from a combination of the large fractionation between aragonite and bicarbonate (~2.7‰) and the elevated δ^{13}C of the DIC caused by the fractionation during photosynthesis of benthic algae, seagrasses and cyanobacteria and fractionation during the conversion of HCO₃⁻ to CO₂ during the precipitation of calcium carbonate.

2- All sediments formed upon GBB have similar δ^{13}C values suggesting a common origin. Mud, whether formed by direct precipitation or breakdown from algal carbonate, is ingested by benthic organisms forming pellets, which harden to produce peloids, which in turn are coated forming coated grains, grapestones
and/or ooids. As a result of this paragenetic sequence, the range of $\delta^{13}C$ and $\delta^{18}O$ of the sediments increases with increasing grain size.

3- Skeletal material only forms a small proportion of sediment on GBB, but has more negative $\delta^{13}C$ and $\delta^{18}O$ values.

4- The $\delta^{18}O$ of the carbonate sediment is elevated in the interior bank relative to the margin as a result of the persistent elevated salinity and $\delta^{18}O$ of the water.

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FIGURES

Figure 1
Location map of the Bahamas and position of sample collection points.

Figure 2
Stable δ^{13}C and δ^{18}O of all samples and size fractions from 2001-2004 cruises; a) data from < 63, 63-125 and 125-250 μm size fractions, b) data from bulk, 250-500, 500-1000 and >1000 μm size fractions.

Figure 3
Relationship between δ^{13}C and facies type (1= mud, 1.5= mud-rich wackestone, 2= wackestone, 2.5= mud-rich-packstone, 3= packstone, 3.5 = mud-lean packstone, 4= grainstone, 5= rudstone).

Figure 4
Relationship between δ^{18}O and facies type (1= mud, 1.5= mud-rich wackestone, 2= wackestone, 2.5= mud-rich-packstone, 3= packstone, 3.5 = mud-lean packstone, 4= grainstone, 5= rudstone).

Figure 5
Spatial contour map of the δ^{13}C of bulk carbonate sediments from GBB. Scale is in kilometres.

Figure 6
Spatial contour map of δ^{18}O of bulk carbonate sediments from GBB. Scale is in kilometres.

Figure 7
Relationship between $\delta^{18}$O and salinity for water samples from GBB measured by Shinn et al. (1989), Lowenstam and Epstein (1957) and this study. The line shows a model result of the evaporation of a water body with an initial salinity of 32 and $\delta^{18}$O of 0‰ using the equations outlined by Gonfiantini (1986) and the model of Craig and Gordon (1965).

Figure 8
Relationship between $\delta^{13}$C of DIC and salinity for samples collected in this study. There is no statistically significant correlation in these data between $\delta^{13}$C of DIC and salinity.

Figure 9
Relationship between $\delta^{18}$O and $\delta$D in the data measured in this study relative to the meteoric water line (dashed line). The solid line shows the evaporative trend of a water body using the same conditions as used in Figure 7.

Figure 10
a) Data from Heron Island, Australia showing the $\delta^{13}$C in a south to north transect (low numbers to high numbers) across the reef. See Weber & Woodhead (1969) for sample location. Changes in $\delta^{13}$C correspond to changes in facies with elevated values corresponding to areas dominated by calcareous green algae and lower values in areas where there are abundant red algae and scleractinian corals (Weber & Woodhead, 1969), b) $\delta^{13}$C data shown in Figure 10a correlated with the percentage of aragonite.

Figure 11
Ranges of equilibrium estimates of $\delta^{13}$C and $\delta^{18}$O of sediments based on data shown in Figures 5 & 6, compared to the mean $\delta^{13}$C and $\delta^{18}$O values measured on carbonates in this study. Carbonate data are the same as shown in Figure 2. Estimates for skeletal data are taken from literature data (Land, 1989; Milliman, 1974; Swart, 1983; Weber, 1965).
REFERENCES


Figure 2
Figure 3
Figure 4
Figure 5
Figure 7

Salinity

$\delta^{18}O$‰

Shinn et al. (1989)
Lowenstam and Epstein (1957)
This Study
Figure 8
Figure 10
Figure 11