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Over 100 years of environmental change recorded by foraminifers and sediments in Mobile Bay, Alabama, Gulf of Mexico, USA

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ABSTRACT

The marine microfauna of Mobile Bay has been profoundly influenced by the development and expansion of the primary shipping channel over the last ~100 years. Foraminifers and sediments from seven box cores with excess lead-210 chronology document that channel dredging and spoil disposal have altered circulation, reduced estuarine mixing, changed sedimentation patterns, and caused a faunal turnover within the bay. Beginning in the late 1800s, changes in estuarine mixing allowed for greater low-pH freshwater influence in the bay, and ultimately began environmental changes that resulted in the loss of calcareous foraminifers. By the early 1900s, box cores throughout Mobile Bay record a ~100-year trend of increasing calcareous test dissolution that continues to the present. Since the completion of the current shipping channel in the 1950s, restricted tidal flushing and increased terrestrial organic matter, documented by carbon-to-nitrogen ratios, stimulated an increase in agglutinated foraminiferal densities. However, in deeper areas of the bay, hypoxic water has negatively impacted the marine microfauna. Comparisons of the present-day foraminiferal assemblage with foraminifers collected in the early 1970s indicate that the continued biologic loss of calcareous foraminifers in the bay has allowed the introduction of a new agglutinated foraminiferal species into the bay.

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1. Introduction

Mobile Bay (Fig. 1), a triangular shaped, semi-enclosed bay located on the northern coastline of the Gulf of Mexico, is a complex estuary where salinity fluctuations, anthropogenic activities, and riverine inflow greatly affect the present environmental conditions. The bay is approximately 50 km long and 17–38 km wide with an average water depth of 3 m. The center of the bay is dissected by a dredged shipping channel of ~14 m water depth (mwd) and ~122 m width, with spoil banks deposited along the western side. Mobile Bay is a drowned Pleistocene river valley with two Pleistocene ridges, Dauphin Island and Morgan Peninsula, along the southern margin that limit marine water flow into the estuary. Additionally, these ridge barriers have expanded during the Holocene by sand accretion, further restricting the connection between Mobile Bay and the Gulf of Mexico.

Mobile Bay has a small volume (3.2×10^9 m^3) relative to the freshwater inflow from the Mobile–Tensaw river system (average flow ~1800 m^3 s^{-1}) that drains most of Alabama and parts of Georgia and Mississippi. The Mobile River system is the eighth largest average discharge in North America and the second largest emptying into the Gulf of Mexico (Kammerer, 1990). River flows entering the bay are highly seasonal, with highest discharge occurring during the late winter and early spring and lowest discharge occurring during the summer months. Salinities in the bay range from fully fresh (0) to fully marine (35) and are influenced heavily by riverine flow (Cowan et al., 1996; Noble et al., 1996).

The freshwater inflow is great enough to keep Mobile Bay highly stratified throughout most of the year (Schroeder and Wiseman, 1988; Noble et al., 1996). Stratification can be temporarily broken up by storms but quickly reforms (Park et al., 2007a). Storms occur year round, but are more frequent in the winter, and also act to redistribute sediment within the bay. The bay is primarily mixed via storm process and less by tidal exchange. Most of the exchange occurs through the narrow tidal pass between Dauphin Island and Morgan Peninsula (85%), with the remainder through the Pass aux Herons into Mississippi Sound. In the last 150 years, extreme storms and hurricanes have also impacted the bay, most recently hurricanes Katrina (2005), Ivan (2004), Fredrick (1979), and the earlier Miami (1926) hurricane (Park et al., 2007b).

Channel dredging near the port of Mobile began around 1827; however, the first ship channel to ~8 mwd was completed in the 1880–1890s (Table 1). Since then, continued deepening and expansion have occurred and more are currently planned.

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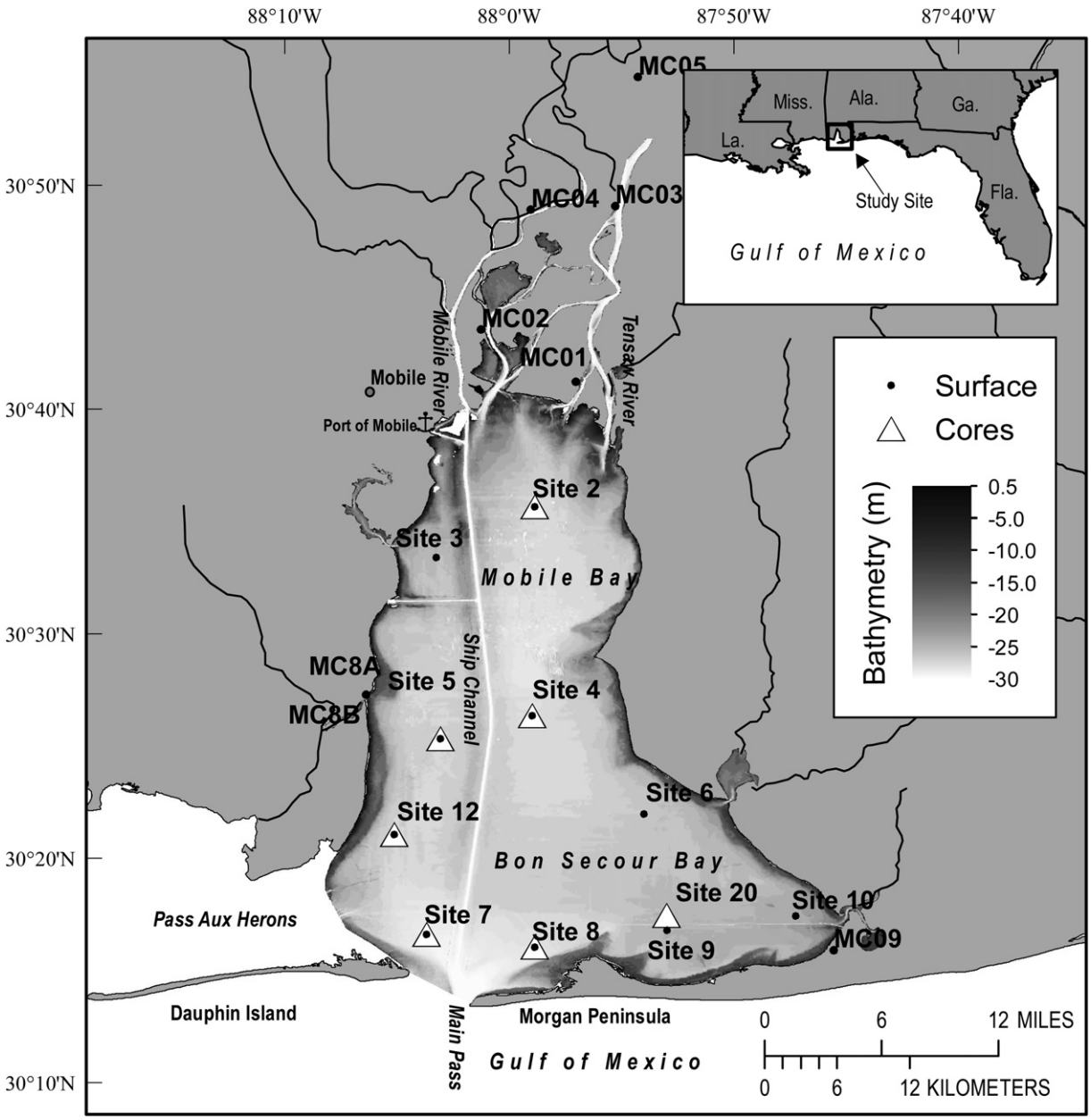

Fig. 1. Map of Mobile Bay showing place names, bathymetry, marsh core locations (black circles), and the locations of identically numbered bay surface samples sites (black circles) and box cores (white triangles) discussed in this paper. Site 9 approximates the location of box core MB0810-20BC. See Table 2 for more information.

Table 1
Chronology of the Mobile Bay ship channel construction.

Ship channel dimensions-depth (mwd) × width (m)	Approximate date completed	Entrance bar channel dimensions-depth (mwd) × width (m)	Approximate date completed
3–4 m	1870–1876		
5 m	1880–1889		
8 m × 15–30 m	1889–1896	9 m × 91 m	1902
9 m × 61 m	1910–1914		
9.7 m × 91 m	1918–1926	10 m × 137 m	1917
12 m × 122 m	1930–1934	11 m × 137 m	1930
12 m × 122 m	1955–1957	13 m × 183 m	1957
14 m × 122 m	Current	14 m × 183 m	Current
16.7 m × 167 m	Planned	17 m × 213 m	Planned

The dredged sediments are deposited as spoil banks primarily along the western side of the channel and at times the spoil banks have projected above the water surface, further reducing tidal exchange within the bay. Ryan (1969) reports that the construction of the 1890s ship channel altered the Mobile Bay circulation pattern. Prior to channel dredging, tidal flow was generally counter clockwise and stronger to the west as evidenced by a large, scoured channel recorded by mid 1800 bathymetric soundings (Ryan, 1969). Beginning in the 1880s the dredge spoil banks blocked the incoming westward tidal flow and deflected it eastward. Today, the majority of the tidal wedge enters the deep ship channel and travels up the center of the bay. Usually saltwater remains trapped in the channel, and mixes with the bay only during storms, or extremely low-river flow, through a process of wind-directed upwelling in the upper bay (Schroeder et al., 1996). In the mid 1940s the saltwater wedge was measured 23 miles up the Mobile

River during a time of low-river flow (Ryan, 1969). It is reported that the transport of more saline water in, and overflow from, the ship channel is the mechanism that supplies the denser water necessary for the development of seasonal stratification and hypoxia in upper Mobile Bay (Noble et al., 1996; Schroeder et al., 1996).

Channel dredging and associated spoil banks have also created isolated sub-basins within the bay where spring and summer hypoxic bottom water conditions have been recorded (May, 1973; Schroeder and Wiseman, 1988; Park et al., 2007a). During times of low-river flow, denser, more saline water escapes the channel and settles into deeper basins in the upper bay where oxygen is consumed and bottom water oxygen becomes limited. During these periodic hypoxic episodes, the mobile bottom-dwelling organisms escape from the deeper, low-oxygen areas into the shallow waters of the bay, and result in "jubilees" of seafood for coastal residents. Hypoxia in Mobile Bay is reported to be more a manifestation of the pronounced freshwater stratification and less a response to increased nutrients (Turner et al., 1987; Park et al., 2007a). However, there is some evidence that the hypoxia is becoming more frequent or of longer duration (Park et al., 2007a). The cause of Mobile Bay hypoxia is quite uncertain, but unlike the Gulf of Mexico Dead Zone excessive or increased riverine nutrient inputs is not believed to be a factor (Cowan et al., 1996). Hypoxic events (jubilees) have been documented since 1867 and averaged about five per year during 1970s (May, 1973). In August 2011, there were five reported events in one week (Raines, 2011).

There has been a considerable amount of research on Gulf of Mexico foraminifers. Poag (1981) notes, in a summary, that salinity is the main environmental factor controlling the distribution of foraminifers in the marginal marine environments of Florida, Alabama, Mississippi, Louisiana and Texas. Agglutinated species such as *Ammotium salsum* live close to river mouths and the low-salinity-tolerant, calcareous genus *Ammonia* is dominant in the main parts of these coastal bays and estuaries. As salinity increases in the offshore bays, calcareous *Elphidium* spp. and *Quinqueloculina* spp. become more common. In the only previous study of Mobile Bay, Lamb (1972) reports that four benthic foraminiferal species are present. In the upper half of the bay, due to the lower salinity, the agglutinated species *A. salsum* and *Miliammina fusca* are dominant, while in the lower bay calcareous species *Elphidium excavatum* and *Ammonia tepida* are dominant.

Mobile Bay, like other estuaries, serves as a repository of sediment and other detritus transported from the terrestrial realm prior to discharging to the coastal ocean. Additionally, marine inputs and internal sedimentation complicate this inherently simplistic view. The sedimentary record preserved by estuarine systems provides insight to certain aspects of land-use and climate forces that contributed to their long-term evolution. Deciphering these changes requires integrated data sets and applicable environmental proxies (e.g., foraminiferal assemblages, organic-matter tracers, etc.). The goal of this research is to understand and document environmental changes in Mobile Bay over the last ~100 years and to determine how natural and anthropogenic forces may have influenced recent geological evolution of Mobile Bay. To accomplish this, we: (1) quantify modern foraminiferal assemblages and examine organic-matter sources throughout the bay; and (2) use the downcore foraminiferal and geochemical record to interpret recent environmental changes.

2. Methods

2.1. Foraminifers

Benthic foraminifers are ideally suited to environmental studies because they have a calcareous or agglutinated test that is preserved in marine sediments and that can be used to monitor both the present conditions, through the analysis of modern surface sediments, and the past environment on variable time scales using core samples. However, on short, seasonal time scales their distributions are highly variable (Murray and Alve, 2000). In order to best remove the seasonal, spatial, and temporal variability from the determination of living foraminifers in Mobile Bay, multiple samples were used. Replicate surface sediments were collected at each site with a van Veen grab sampler in September 2009, May 2010 and August 2010 (Fig. 1; Table 2). Approximately 20 cc (11–35 cc) of the topmost oxidized, brown, surface sediments were collected using a spatula and immediately placed in two 50-cc centrifuge tubes to which a Rose Bengal stained solution of ethanol (2 g L^{-1}) buffered with sea water. Reduction detected by the smell of H_2S was not observed in any surface sample. The sample tubes were repeatedly shaken over the next two weeks to ensure equal staining of protoplasm and processed within one month.

During the August 2010 cruise aboard the U.S. Geological Survey R/V *Gilbert* box cores were also collected from the sites. Six-inch diameter plastic core liners were inserted into the box core. The sediment cores were refrigerated after collection and the sediment cores were extruded and the entire core was sampled at 1-cm intervals within 48 h after collection. No staining was done on the core sediments. All samples were refrigerated until lab processing, and all box-core samples were processed within six months.

Sediment volume measurements for the replicate stained surface samples (11–35 cc) were determined after settling in their centrifuge tubes. Standardized 20 cc volumes for box-core samples were measured by initially packing each sediment sample into a 0.5 ml graduated syringe. Both the stained surface and the core sediment samples for faunal analyses were not dried prior to the wet-sieving process. Each sample was soaked in water with a small amount of added 5% sodium hexametaphosphate solution, slowly agitated for up to 1 h to aid disaggregation, then washed over a stainless-steel, 63-μm sieve. The >63-μm fractions were oven dried at ≤60 °C, then dry sieved at 125 μm. Percent mud was determined by the mass of the >63-μm and the initial sample mass minus water content (see description later on how water content was determined). The foraminifers were picked, identified and counted from the >125-μm fraction. In most cases the entire sample was picked for stained tests. The six replicate surface samples (two for each season) were counted individually but eventually the data were combined to provide a complete representation of the living foraminiferal community over the year.

Table 2
Mobile Bay sample locations.

Marsh sample identification, site and box-core name	Latitude (°N)	Longitude (°W)	Water depth (m)
MB09-MC01	30.6875	87.9505	0.8
MB09-MC02	30.7262	88.0208	0.8
MB09-MC03	30.8180	87.9212	0.8
MB09-MC04	30.8152	87.9839	0.3
MB09-MC05	30.9139	87.9043	0.8
MB09-MC08A	30.4551	88.1061	0.4
MB09-MC08B	30.4551	88.1061	0.0
MB09-MC09	30.2652	87.7591	0.0
Site 2; MB0810-2BC	30.5944	87.9811	3.5
Site 3	30.5570	88.0543	3.1
Site 4; MB0810-4BC	30.4394	87.9830	3.9
Site 5; MB0810-5BC	30.4226	88.0511	4.0
Site 7; MB0810-7BC	30.2772	88.0612	5.1
Site 8; MB0810-8BC	30.2677	87.9811	4.4
Site 9	30.2801	87.8827	3.0
Site 12; MB0810-12BC	30.3514	88.0853	3.9
Site 20; MB0810-20BC	30.2910	87.8830	3.0

Processed box-core samples contained abundant benthic foraminifers. In all but a few cases, a representative subsample of approximately 300 specimens was obtained for faunal analysis using a microsplitter.

The purpose of this research was to document environmental change in Mobile Bay. Therefore, by using a >125-μm size fraction, this research deliberately does not strive to answer questions about the complete foraminiferal distribution. Nevertheless, comparison studies analyzing both the >63-μm and >125-μm faunal assemblages have identified similar faunal distributions and identical ecological controls on the different foraminiferal species found in each of the two different-size assemblages (Mojtahid et al., 2009; Bouchet et al., 2012). Useful ecological information is gained regardless of the size fraction examined. Additionally, a recent attempt to standardize foraminiferal-sampling protocols has made a mandatory recommendation for basing the faunal inventory on the >125-μm fraction (Schönfeld et al., 2012). The choice of the larger sieve size allows valid ecological information to be obtained more quickly.

The amount of calcareous specimens dissolution was graded on a three-point scale: no dissolution, mild dissolution, and strong dissolution. Mild dissolution is characterized by the first few chambers of the test affected by decalcification. Strong dissolution is characterized by a usually unidentifiable, star-shape test that occurs when the outer chambers of *Ammonia* spp. and *Elphidium* spp. are completely dissolved, leaving behind the inner organic lining (LeCadre et al., 2003). Complete surface sample and downcore foraminiferal counts can be found in Richwine and Osterman (2012).

2.2. Sediment characterization and geochronology of box-core samples

Subsamples from each 1-cm core interval were processed for basic sediment characteristics (i.e., dry bulk density, mass-water content, porosity, and organic-matter content) and geochronology via excess lead-210 ($^{210}Pb_{xs}$), difference between total and supported lead-210 (^{210}Pb) and cesium-137 (^{137}Cs). Mass-water content and bulk densities were determined by drying 30 ml of wet sediment at 60 °C. Differences in the pre- and post-drying masses were used to compute water content; dry bulk density was computed as the ratio of dry sediment to an initial wet volume of 30 ml. Volumes were determined using a 0.5 ml graduated syringe once samples were returned to the lab. Porosity was extracted from these data assuming pore water salinity was equivalent to bottom water salinity at the time of core collection (Smith et al., in press). Organic-matter content was determined using loss-on-ignition (LOI) at 450 °C from 5 to 10 g of dried sediment (Smith et al., in press).

Stable isotopes of carbon (^{12}C and ^{13}C) and nitrogen (^{14}N and ^{15}N) as well as bulk content of organic carbon and total nitrogen (presented as the ratio of carbon-to-nitrogen or C:N) were also measured on dried, box-core subsamples and surface samples. Each dried, homogenized sediment sample was pre-weighed in 5 × 9 mm silver capsules (Costech, Inc.) and fumigated with concentrated hydrochloric (HCl) acid for 4–6 h to remove inorganic carbon (Harris et al., 2001). Silver capsules were sealed and analyzed for ^{13}C and ^{15}N using a PDZ Europa ANCA-GSL elemental analyzer interfaced to a PDZ Europa 20-20 isotope ratio mass spectrometer (Sercon Ltd., Cheshire, United Kingdom) by the Stable Isotope Facility at University of California, Davis. Every third sample was replicated to evaluate sample heterogeneity. Relative standard deviations of $\delta^{13}C$ and $\delta^{15}N$ on all 85 replicates were 0.4% and 4.8%, respectively. The $\delta^{13}C$ and $\delta^{15}N$ are presented as ‰ differences relative to the reference standards Vienna PeeDee Belemnite and Air, respectively.

Additional subsamples were processed for the determination of $^{210}Pb_{xs}$ and ^{137}Cs by standard gamma-ray spectrometry (Cutshall et al., 1983). Each subsample (approximately 80 g of wet sediment) was freeze-dried and subsequently homogenized using a mortar and pestle. Between 15 and 20 g of dried sediments were sealed in airtight plastic containers for approximately 4 weeks prior to measurement to achieve secular equilibrium between radon-222 (^{222}Rn) and its parent radium-226 (^{226}Ra). Each sample was counted for a 2–3 day period on a planar-style, low energy, high purity germanium, gamma-ray spectrometer (Canberra, Inc.). Total ^{210}Pb and ^{137}Cs were determined using photopeaks 46.5 and 661.6 keV, respectively. Supported ^{210}Pb (sediment-bound ^{226}Ra) was estimated using the 295.1 and 352.1 keV photopeaks for ^{214}Pb and the 609.3 keV photopeak for ^{214}Bi (bismuth-214). Sample count rates were corrected for detector efficiency (referenced to the IAEA RGU-1 standard), standard photopeak intensity, and self-absorption corrections using a uranium-238 (^{238}U) sealed source (Cutshall et al., 1983).

Sediment geochronology for each box core was based on the vertical distribution of $^{210}Pb_{xs}$ and ^{137}Cs in the sediments. The $^{210}Pb_{xs}$ was determined by the difference between total ^{210}Pb and supported ^{210}Pb (i.e., sediment-bound ^{226}Ra). Typically, the excess ^{210}Pb tends to zero at depth or is within error of zero (i.e., typically within 0.3–0.5 dpm g^{-1}). Age–depth relationships were determined using the constant rate of supply (CRS) model (Appleby and Oldfield, 1978; Binford, 1990; Appleby, 2001). Briefly, the CRS model assumes a constant supply of $^{210}Pb_{xs}$ to the sediment surface. This allows the age of a sediment layer (years before collection) to be approximated by the decay correcting the disequilibrium between total depth-integrated ($^{210}Pb_{xs}$) inventory of the core and the inventory below the sediment layer of interest (Binford, 1990). With accurate knowledge of the date of core collection, the age of the sediment layer can be converted to calendar years. Age–depth relationships were then refined using either peaks in ^{137}Cs, which correlate with the peak in atmospheric nuclear testing in 1963, and/or the first occurrence of ^{137}Cs in the sediment record, which we assume is bracketed to the time period between 1954 and 1965.

3. Results

3.1. Foraminifers and geochemistry of surface sediments

Table 3 summarizes the living and dead foraminifers found in the surface sediment (Sites) of Mobile Bay and is in general agreement with previous work (Lamb, 1972). In the upper bay where salinity measurements ranged from 2 to 16 (Table 3), there are more agglutinated foraminifers (primarily *A. salsum*) than calcareous foraminifers (primarily *Ammonia* spp.), and there are ~four living species (S) in a sample. In the middle bay there are roughly equal numbers of calcareous and agglutinated living foraminifers, and the number of species increases slightly. In the southern bay where recorded salinity varies between 9 and 31, calcareous foraminifers are more abundant than agglutinated foraminifers, and there are greater than seven living species in each sample.

Differences between living and dead assemblages in the northern bay favor a relative decrease in calcareous foraminifers and/or a relative increase in the number of agglutinated foraminifers (Table 3). This is a clear indication that the dead calcareous foraminifers are being subjected to strong dissolution in the upper part of Mobile Bay. Dissolution causes calcareous foraminiferal tests to become opaque and often decalcified, leaving a characteristic star-shaped test of *Ammonia* spp. and *Elphidium* spp. (LeCadre et al., 2003). The percent of moderately to strongly decalcified foraminiferal tests observed in both the living and dead surface samples from Mobile Bay is shown in Fig. 2 (Richwine and Osterman, 2012). Living foraminifers show a small percentage of etched tests at Sites

Table 3
Foraminifers and geochemistry of Mobile Bay surface sample sites.

Sample station*	Linear distance from river mouth	Depth (m)	Salinity range	Mean number of species (standard error)	Density range number foraminifers/cc	Mean density – number foraminifers/cc (standard error)	Mean % agglutinated specimens (standard error)	Mean % calcareous specimens (standard error)	Molar C:N (standard error)	$\delta^{13}C$ (standard error)	$\delta^{15}N$ (standard error)
Marshes											
Freshwater Bay-head Marsh Stations (MC01, MC02, MC03, MC04, MC05)		–	–	–	–	–	–	–	12.28 (1.60)	−27.85 (0.83)	4.45 (0.36)
Juncus Brackish Marsh Stations (MC08A, MC08B, MC09)		–	–	–	–	–	–	–	16.59 (1.79)	−25.14 (0.40)	3.03 (1.35)
Upper Bay											
Site 2 (east)	12.6	3.5	2–8						10.13 (2.66)	−27.44 (0.15)	4.23 (0.94)
Living foraminifers				*3.6 (0.54)*	*0.47–11.9*	*6.38 (2.04)*	*51.4 (18.84)*	*46.8 (18.84)*			
Dead foraminifers				5.25 (0.41)	49–352	150.26 (65.3)	88.18 (2.48)	11.82 (2.48)			
Site 3 (west)	14.2	3.1	3–12						9.24 (0.59)	−26.72 (0.05)	5.46 (0.95)
Living foraminifers				*4.60 (0.67)*	*3.13–47.49*	*18.54 (7.31)*	*58.61 (21.04)*	*41.39 (21.04)*			
Dead foraminifers				3.75 (0.41)	264–1428	1048.11 (231.7)	91.56 (1.84)	8.44 (1.84)			
Site 4 (east)	26.3	3.9	3–16						10.12 (1.11)	−26.51 (0.04)	5.13 (0.33)
Living foraminifers				*3.00 (0.78)*	*0.82–4.9*	*1.52 (0.65)*	*18.12 (9.83)*	*81.88 (9.83)*			
Dead foraminifers				8.2 (0.52)	30–392	212.42 (68.06)	47.55 (7.65)	52.45 (7.65)			
Site 5 (west)	28.2	4	5–16						7.49 (2.00)	−26.02 (0.49)	4.36 (0.95)
Living foraminifers				*4.67 (1.07)*	*1.35–26.5*	*9.71 (3.55)*	*57.88 (16.76)*	*42.12 (16.76)*			
Dead foraminifers				7.8 (0.66)	301–1287	749.6 (188.01)	74.05 (1.97)	25.95 (1.97)			
Middle Bay											
Site 9 (east)	42.5	3	6–15						7.37 (2.31)	−25.13 (0.12)	4.74 (1.23)
Living foraminifers				*4.33 (0.69)*	*0.95–8.59*	*3.86 (1.18)*	43.64 (13.74)	*56.36 (13.74)*			
Dead foraminifers				12.75 (0.65)	114–498	310.42 (91.77)	40.09 (4.89)	59.91 (4.89)			
Site 12 (west)	36.2	3.9	7–17						8.24 (0.75)	−25.06 (0.13)	5.51 (0.86)
Living foraminifers				*7.50 (0.35)*	*8.27–30.58*	*19.43 (7.89)*	*53.96 (6.91)*	*46.04 (6.91)*			
Dead foraminifers				9.5 (0.35)	415–582	498.99 (59.29)	44.32 (12.73)	55.68 (12.73)			
Lower Bay											
Site 7 (west)	44.3	5.1	9–31						9.63 (0.33)	−24.23 (0.05)	5.51 (0.08)
Living foraminifers				*6.33 (1.43)*	*0.0.94–34.89*	*12.11 (16.35)*	*33.55 (10.96)*	*66.45 (10.96)*			
Dead foraminifers				15.2 (1.25)	77–231	142.02 (26.33)	33.86 (3.97)	66.14 (3.97)			
Site 8 (east)	45.3	4.4	15–30						8.47 (1.41)	−24.51 (0.46)	5.40 (0.9)
Living foraminifers				*9.00 (1.24)*	*4.55–124*	*47.3 (22.96)*	*19.78 (9.09)*	*80.22 (9.09)*			
Dead foraminifers				20.25 (0.35)	209–2502	1405.81 (483.72)	17.66 (2.35)	74.93 (2.35)			

* All data represents three seasons of measurements (fall 2009, spring 2010 and summer 2010) except for the Marsh cores (fall, 2009 and spring 2010) and Site 12 (spring and summer 2010).

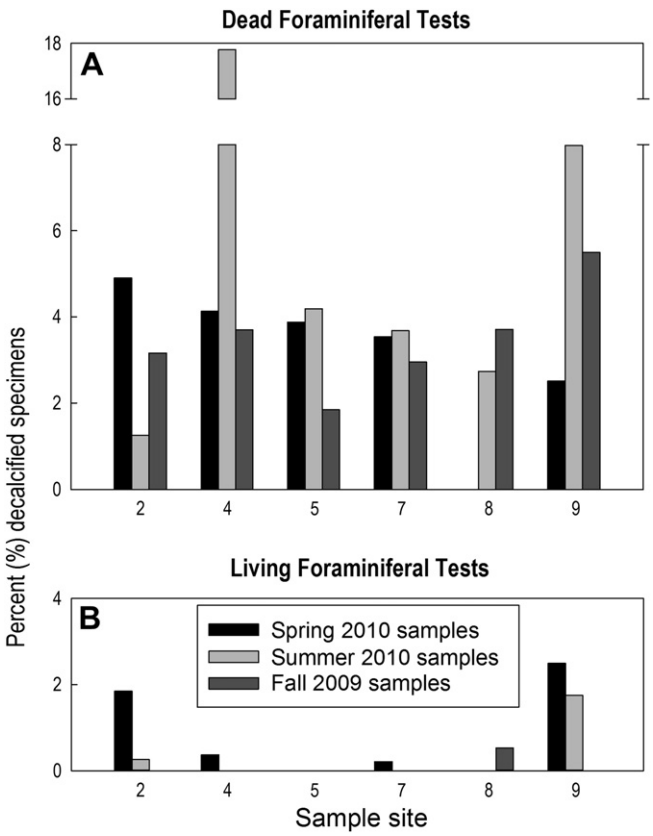

Fig. 2. Percent of moderately and strongly decalcified dead (A) and living (B) calcareous foraminiferal specimens (determined on a three-point scale) found in the surface sediment samples of Mobile Bay. Values computed at each surface sample site for the three-seasonal samples (fall 2009, spring 2010 and summer 2010) are shown on the *x*-axis. All surface sample sites correspond to same numbered box-core locations except for Site 9, which approximates the location of box-core MB0810-20BC. See Fig. 1 and Table 2 for site locations.

2 and 9, which are most influenced by freshwater inflow. Decalcification of the dead assemblage is fairly uniform throughout the bay, with about 4% of the dead specimens showing evidence of dissolution during all seasons. Site 4, collected in a deeper basin in the upper bay, records the highest numbers of decalcified dead specimens, >17% in the summer collection.

Organic-matter content in the surface sediments varies between 6 and 10% by weight (Table 3). Organic-matter content is higher in the eastern bay relative to the western bay sites ($p = 0.03$, $n = 6$). The δ^{13}C of the bulk organic matter increases along the longitudinal (north to south) axis of the bay. Molar C:N are highest at the most northern sites of the upper bay (2–4), decrease drastically along the middle bay, and increase slightly in the lower bay (7 and 8). Nitrogen isotopes do not have any systematic trend with either salinity or with distance from riverine source; however, the values are slightly enriched relative to marsh sediments (Table 3). The organic matter preserved at the upper bay sites with average C:N and δ^{13}C of 9.84 and −26.89‰, respectively, are similar to organic matter accumulating in the freshwater marshes (12.28 and −27.85‰, respectively; Table 3).

3.2. Radionuclides and geochronology of box-core sediments

Overall, the sediments obtained in the box cores from Mobile Bay were predominantly muds (relative one standard deviation < 10%)(Table 4). Except for MB0810-7BC, all cores had a vertically averaged mud content of greater than 90%. While

Table 4
Lead and cesium inventories for Mobile Bay box cores.

Core ID	Core length (cm)	Mixing zone (cm)	Maximum depth ^{210}Pb observed (cm)	Surface activity excess ^{210}Pb (dpm g^{-1})	Excess ^{210}Pb inventory (dpm cm^{-2})	Excess ^{210}Pb inventory error (±dpm cm^{-2})	Maximum depth ^{137}Cs observed (cm)	^{137}Cs inventory (dpm cm^{-2})	^{137}Cs inventory error (±dpm cm^{-2})	Average mud (<63 μm) content (%)	Mud (<63 μm) content 1 sigma (±%)
MB0810-2BC	45	15	44	2.71	41.72	4.01	30	3.06	0.182		
MB0810-4BC	35	10	28	4.26	32.76	2.66	19	1.764	0.105	96.05	1.97
MB0810-5BC	37.5	5	18	3.55	15.78	1.64	8.5	0.599	0.073	97.93	1.34
MB0810-7BC	27	3	26	4.5	49.57	3.42	26	2.82	0.299	95.20	3.82
MB0810-8BC	49	5	32	7.06	69.9	3.52	22	2.41	0.226	71.34	7.63
MB0810-12BC	38	2	22	4.62	23.10	2.19	15	1.3	0.131	90.15	8.52
MB0810-20BC	31	8	28	5.78	48.67	2.54	29	3.792	0.163	95.08	4.58
Average				**4.64**	**40.2**			**2.25**	**0.17**	**95.84**	**2.36**
Stdev				**1.32**	**16.9**			**1.02**	**0.07**		

MB0810-7BC had a higher sand content, the depth-averaged mud content still exceeded 70%. These muddy sediments are ideal for using particle reactive tracers such as ^{210}Pb and ^{137}Cs. The vertical distributions $^{210}Pb_{xs}$ and ^{137}Cs are presented in Fig. 3. For all cores, $^{210}Pb_{xs}$ profiles show near-vertical gradients in the upper few centimeters, indicating slight reworking (Table 4). Below the mixed zone of near constant activity, profiles have a more monotonic decrease of $^{210}Pb_{xs}$ with depth. For most cores, a complete $^{210}Pb_{xs}$ record was obtained as indicated by near zero values at the base of each core. The exceptions are MB0810-7BC and -2BC, which have appreciable $^{210}Pb_{xs}$ at depth. Surface $^{210}Pb_{xs}$ activities and cumulative inventories for the seven box cores average 4.64 ± 1.32 dpm g^{-1} and 40.2 ± 16.9 dpm cm^{-2} (Table 4), respectively, and are comparable to the activities (3.5 ± 1.1 dpm g^{-1}, $n = 4$) and inventories (31.5 ± 4.9 dpm cm^{-2}, $n = 4$) observed at marsh sites around the bay by Smith et al. (in press). This suggests that on average, the flux of ^{210}Pb and the total accumulation of ^{210}Pb in the bay sediments, where sediment redistribution is common, are comparable to less dynamic systems like marshes. However, individual variations in cumulative inventories are observed in the box cores. For example, MB0810-5BC has approximately half the total inventory as the average marsh and bay cores, indicating limited deposition or removal of sediment and associated ^{210}Pb at this site. In contrast, MB0810-8BC at the bay-gulf inlet has twice the inventory and surface activity, suggesting potential focusing from the marine and/or estuarine system.

The CRS model was used with measured $^{210}Pb_{xs}$ activities and inventories to determine age–depth relationships (Fig. 4) and corroborated with ^{137}Cs specific activity data (Binford, 1990; Appleby, 2001). The modeled temporal framework of each individual core varied; however, basal ages were 100–130 years (five–six half lives, $t_{½}$ of ^{210}Pb) with a ±15 years uncertainty. Any ages presented beyond 130 years are presented with caution and the inherent understanding that errors may exceed 50+ years. Profiles of ^{137}Cs did not show a well-preserved peak in any of the cores, which is assumed to be approximately 1963. However, based on the age models, the first detectable ^{137}Cs in all the cores occurred between 1945 (±8 years) and 1968 (±8 years). This first occurrence matches fairly well with atmospheric nuclear testing period between 1952 and 1968. The expected timeline of the study is then approximately the turn of the 20th century with corroboration of both ^{210}Pb-based models and the first occurrence of ^{137}Cs.

Overall, the age–depth relationship for all core sites is fairly linear from the 1950s to present day. A curvilinear transition during this period indicates a slight increase in the overall linear sedimentation rate. This acceleration is most pronounced on the east side of the bay. The highest sedimentation rates occur proximal to the subaerial bay-head delta (MB0810-2BC and -4BC) and near the

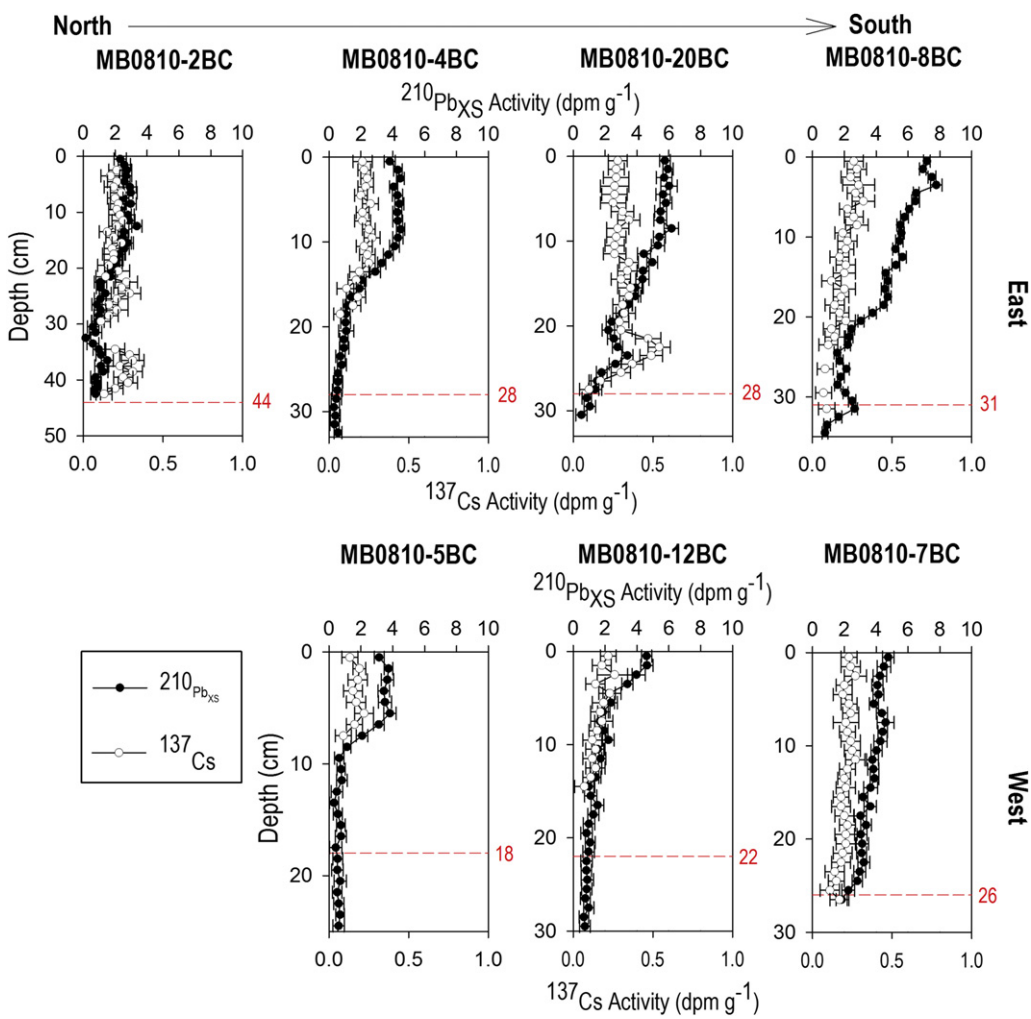

Fig. 3. Specific activity of $^{210}Pb_{xs}$ and ^{137}Cs presented as a function of depth for each of the box cores collected in Mobile Bay. Dashed line denotes inferred base of $^{210}Pb_{xs}$. Analytical uncertainty is presented at one standard deviation. Core records are arranged geographically from North (left) to South (right) and East (top) and West (bottom) of the ship channel. Error bars for specific activity are illustrated at 1 sigma.

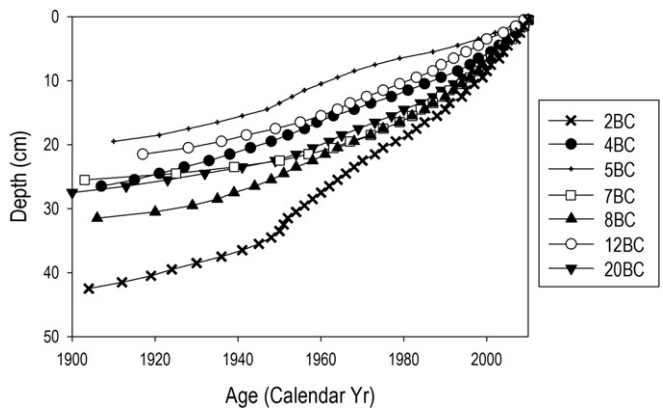

Fig. 4. Modeled age–depth relationship for each box core was established using measured $^{210}Pb_{xs}$ and assuming a constant rate of supply of ^{210}Pb over the last 100 years. A gradual transition from slower sedimentation rates to fairly linear sedimentation rates occurs around 1950.

inlet connecting Mobile Bay to the Gulf of Mexico (MB0810-7BC and -8BC). Cores collected along the western section of the bay (MB0810-12BC and -5BC) generally have a more compressed record relative to eastern sites indicating potential winnowing of fine-grained sediments and/or slower average sedimentation.

3.3. Foraminifers and geochemistry of box-core sediments

Benthic foraminifers were identified in seven box cores from Mobile Bay (Table 2). The relative percentage of agglutinated foraminifers in each sample is plotted in Fig. 5. The most striking result observed in the cores is the relative decrease of calcareous foraminifers, resulting in a faunal turnover in the bay during the last ~100 years. This is seen most clearly in cores MB0810-5BC, -7BC, -12BC and -20BC, and less in the marine dominated (MB0810-8BC) and freshwater dominated (MB0810-2BC, -4BC) cores. In most cores, sediments deposited prior to 1900 contain relatively more calcareous and fewer agglutinated foraminifers than are found in the modern bay sediments (Fig. 5). A second trend observed in the bay cores is increasing foraminiferal density (number of foraminifers [N] cc^{-1}) over the last century, with the youngest agglutinated assemblages having a greater density than the older mixed assemblages (Fig. 6). Exceptions to this trend are the marine-influenced MB0810-8BC and MB0810-4BC located in a deeper basin in the upper bay.

Organic-matter (OM) content in the box-core sediments varies considerably across both space and time (Fig. 6). The overall average percent OM is 9.6% and the total range is 5.8–15.0%. Downcore $\delta^{13}C$ and C:N data of the organic matter are presented in Fig. 7. All cores show a general depletion of ^{13}C upcore, which results in an approximate 2‰ decrease in the $\delta^{13}C$. The lightening-upward trend in bulk organic matter is statistically significant ($r^2 > 0.65$) in all cores except MB0810-8BC. Like the surface sediments, the upper 5 cm of all cores have down-estuary enrichment of ^{13}C moving away from the bay-head delta toward the Gulf of Mexico. The molar C:N downcore are extremely noisy, and no quantitative trend exists. However, qualitatively, C:N in the upper section (top 10-cm) of several cores, especially those on the western side of the bay (MB0810-2BC, -5BC, -7BC, and -12BC), do decrease relative to the mean.

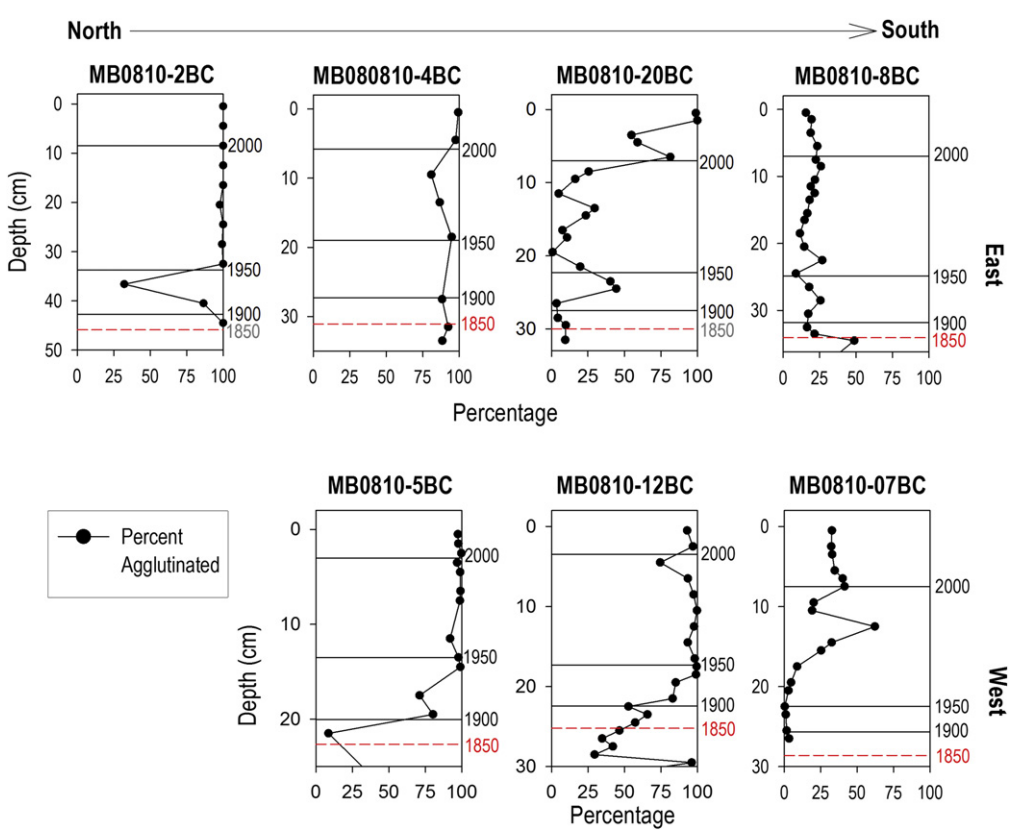

Fig. 5. Percentage of agglutinated foraminifers, relative to calcareous foraminifers, in the Mobile Bay box cores shown by sample depth. Excess ^{210}Pb chronology shown on right *y*-axis, with extrapolated ages shown with dashed line. Northern bay cores contain more agglutinated specimens than the southern bay cores. Younger sediments, in all but MB0810-8BC, contain relatively higher percentages of agglutinated specimens. Core records are arranged geographically from North (left) to South (right) and East (top) and West (bottom) of the ship channel.

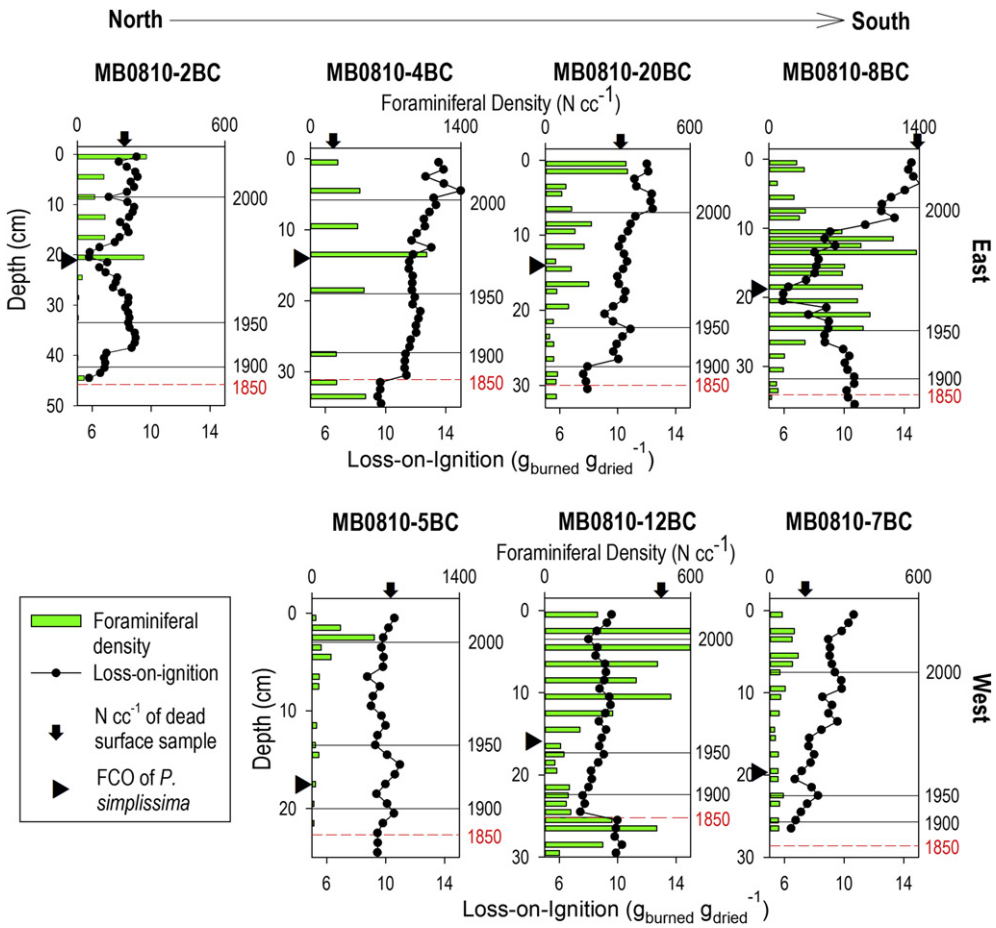

Fig. 6. Foraminiferal density (N cc^{-1}) and the organic carbon loss-on-ignition (LOI) in each box core shown by depth. Excess ^{210}Pb chronology shown on right *y*-axis, with extrapolated ages shown with dashed line. Note differing (N cc^{-1}values) on upper *x*-axis for cores MB0810-2BC, -7BC, and -20BC. Most box cores document higher foraminiferal densities after 1950 and lower foraminiferal densities prior to 1950. Also shown is the depth in each core (triangle on left *y*-axis) of the first common occurrence (FCO) of the agglutinated species *Paratrochammina simplissima*. A black arrow on the upper *x*-axis denotes the three-season average density (N cc^{-1}) of dead specimens in the surface sediment site at each box-core location (Table 3). In all box cores, except for the hypoxic site -4BC, surface densities are close to the maximum observed downcore density. Core records are arranged geographically from North (left) to South (right) and East (top) and West (bottom) of the ship channel.

4. Discussion

4.1. Taphonomy of surface foraminiferal assemblages

In over half of Mobile Bay, agglutinated foraminifers dominate the modern surface assemblage. This differs greatly from other worldwide industrial (Albani et al., 2007; McGann, 2008; Tsujimoto et al., 2008) and other Gulf of Mexico estuaries (Poag, 1981; Buzas-Stephens et al., 2011), which are dominated by the calcareous genera *Ammonia* and *Elphidium*. One contributing factor is that Mobile Bay is a drowned-river valley estuary where riverine freshwater from the Mobile–Tensaw River system and limited exchange with the gulf waters result in year-round low and spatially variable salinity. Similar dominantly agglutinated assemblages have been observed in low salinity, brackish sections of cores from other drowned-river valley estuaries like Albemarle (Vance et al., 2006) and Pamlico Sounds (Abbene et al., 2006) in North Carolina. Low to variable salinity conditions contribute to the observed living foraminiferal distribution; however, in Mobile Bay dissolution of calcareous tests also appears to affect the thanatocoenosis. Decreased pH in the bottom and/or pore water is common in systems with high freshwater discharge from the watershed to the bay and can result in a loss of calcareous foraminifers (Murray and Alve, 1999; LeCadre et al., 2003; Debenay et al., 2004; Scott et al., 2005). High freshwater discharge into Mobile Bay is one explanation for the loss of calcareous foraminifers.

Additionally, studies have also shown that diagenetic alteration of organic matter can alter bottom and/or pore water chemistry and enhance the dissolution of carbonate tests. The oxidation of organic matter by oxygen (O_2) can lead to a build up of carbon dioxide (CO_2) in pore waters that in turn drives down the carbonate saturation state and allows calcareous tests to dissolve. Under hypoxic to suboxic conditions, CO_2 can build up in the bottom water as oxidation of organic matter consumes O_2. These conditions are quite common in certain portions of Mobile Bay and can promote further taphonomic loss of calcareous specimens. For example, Site 4 is located in a deeper area of the bay with high organic matter ($\sim$10%) and documented summer hypoxia (Park et al., 2007a). At Site 4 surface foraminifers have >17% decalcified tests from the summer surface sample sediment (Fig. 2). For that reason the higher percentage of calcareous foraminiferal etching at Site 4 may be best explained by increased CO_2 concentration in the hypoxic water. The faunal assemblage of 74% living calcareous species at Site 4 (Table 3) also indicates a hypoxic environment as experimental data show that living calcareous foraminifers are more resistant to anoxia than living agglutinated species (Moodley et al., 1997).

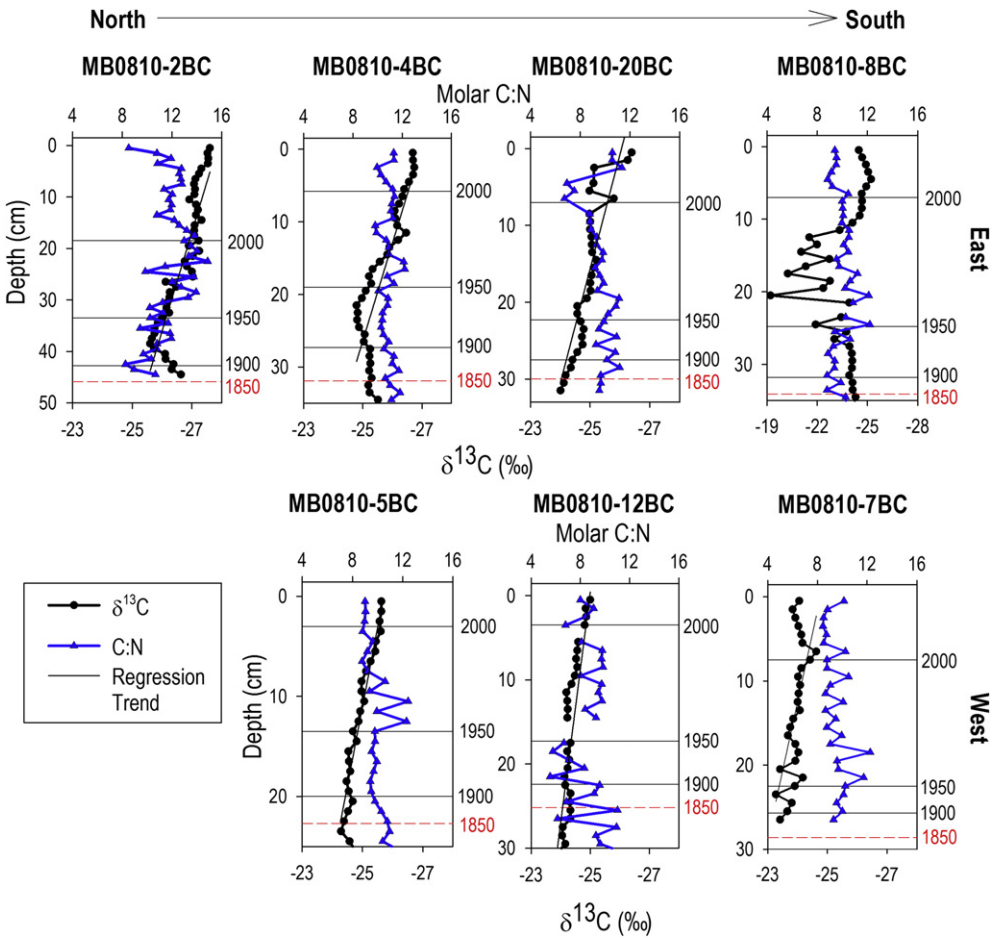

Fig. 7. Ratio of total organic carbon to total nitrogen (C:N) and $\delta^{13}C$ for bulk organic matter presented as a function of depth for each Mobile Bay box core. Excess ^{210}Pb chronology shown on right *y*-axis, with extrapolated ages shown with dashed line. Regression analysis on the $\delta^{13}C$ data (except MB0810-8BC) ($r^2 > 0.65$) show organic matter has increased by approximately 2–3‰ over the last 150 years. The magnitude of fractionation is comparable to that observed in modern surface samples (Table 3). Core records are arranged geographically from North (left) to South (right) and East (top) and West (bottom) of the ship channel.

Consequently, at Site 4 the living agglutinated percentage is the lowest in the upper bay (26%)(Table 3). Additional evidence for hypoxia at this site is also seen in the decreased foraminiferal density and will be discussed later.

Dissolution of calcareous tests can also occur in the absence of oxygen (Berkeley et al., 2007). Oxidation of organic matter by sulfate (SO_4^{2-}) reduction produces hydrogen sulfide (H_2S) and decreases pore water pH (Berkeley et al., 2007; Diz and Frances, 2009) that in turn dissolves carbonate. However, since the surface samples were restricted to the "oxic" sediments (interpreted based on brown coloration), this latter process is not considered significant in Mobile Bay. Also, the abundance of iron (Fe) in Mobile Bay sediments (Isphording, 1983) suggests that much of the H_2S that forms is probably sequestered as amorphous iron sulfides and buffers the pH (Berkeley et al., 2007).

Faunal comparisons with Lamb (1972) can be used to understand the recent history and severity of the calcareous dissolution problem in Mobile Bay. Table 5 shows the tabulation of the five major divisions reported by Lamb (1972) (two agglutinated, two calcareous and other calcareous) from Lamb's (1972) locations close to our sites throughout the bay. Comparisons between two studies document recent environmental changes in Mobile Bay. Unmistakably, there has been a loss of calcareous foraminifers from the bay during the last 40 years. In addition there has been the introduction of a new agglutinated species *Paratrochammina simplissima* into the bay since the early 1970s (Table 5).

The density (N cc^{-1}) of the living foraminifers in Mobile Bay is low (Table 3), but in line with the living densities of other comparable estuaries especially when considering the dissolution problem (Table 3) (Buzas-Stephens et al., 2011). As is usually the case, at each site the density of the dead assemblage is higher than the living. In upper and middle Mobile Bay, both the living and dead faunal densities are higher in the western bay closest to the city and port of Mobile (Table 3). The main source of food for foraminifers is organic matter and the associated bacteria. Increased availability of this nutrition will result in a larger standing crop or density of foraminifers (Yanko et al., 1994; Diz et al., 2004). The increased foraminiferal density observed in the northwestern bay indicates that foraminifers are responding positively to the increased trophic resources of the most polluted region of the bay (Peachey, 2003). Alternately, an additional source of nutrients and food could be through the introduction of the ship channel's saline gulf water into the upper bay during certain overflow events (Schroeder et al., 1996). Also as expected, Site 8 in the lower bay with the largest marine-carbon signature and highest salinity has the highest surface density in the bay (Table 3).

In summary, the dominance of agglutinated foraminifers in the surface sediments of Mobile Bay is not only related to lowered

Table 5
Comparison of surface foraminiferal data from Mobile Bay reported in Lamb (1972) and 2012 (this paper).

Sample locations within mobile bay from north to south	% *Miliammina fusca* (agglutinated)	% *Ammotium salsum* (agglutinated)	% *Paratrochammina simplissima* (agglutinated)	% Elphidium spp. (calcareous)	% Ammonia spp. (calcareous)	% other calcareous species
Site 2						
1972	19	26	No data	22	33	0
This paper	23	64	0.05	0.1	12	0
Site 3						
1972	22	28	No data	19	30	0
This paper	14	77	0	0.1	8	0
Site 4						
1972	0	0	No data	41	51	7
This paper	3	44	3	12	32	6
Site 5						
1972	0	0	No data	41	51	7
This paper	4	70	0.4	5	16	3
Site 9						
1972	0	0	No data	38	53	8
This paper	0.1	37	3	23	26	11
Site 12						
1972	0	0	No data	41	53	6
This paper	7	34	2	18	36	1
Site 7						
1972	0	0	No data	36	49	14
This paper	0.1	23	10	17	40	8
Site 8						
1972	0	0	No data	33	45	22
This paper	0	12	5	15	41	26

salinity, but also to the removal of calcareous species through dissolution. There are two possible causes of the dissolution. The first is lowered pH due to the large volume of freshwater delivered to the bay and the second appears to be a build up of CO_2 in deeper areas of the bay. Because ultimately hypoxia is directly linked to freshwater influx and organic matter supply, it may be difficult to determine what exactly is the cause of the increased calcareous dissolution at any particular site, but dissolution appears to have accelerated in recent times.

4.2. Discussion of downcore faunal changes

The foraminiferal data show a shift from lower density, mixed agglutinated-calcareous assemblages to higher density agglutinated assemblages over the last century (Figs. 5 and 6). Several likely explanations can explain the foraminiferal trends observed; however, three possible scenarios are most likely these: (1) taphonomic loss of agglutinated tests deposited prior to ~1900; (2) decrease in preservation of calcareous foraminifers after ~1900; or (3) a true ecological response (succession) due to changes in benthic water quality. Scenario 1 provides the only criteria where the bay conditions do not have to change while the other two scenarios require a change in the environmental and geochemical framework of the bay.

The near-to-total absence of agglutinated foraminifers at depth within sediment cores is often attributed to diagenetic sedimentary processes (Berkeley et al., 2007). In this scenario, the organic linings of agglutinated species (commonly *Ammobaculites* spp. and *Ammotium* spp.) are remineralized and sediment particles are reworked, with up to 90% reported loss of agglutinated species within the uppermost 10 cm in a marsh core (Goldstein and Watkins, 1999). The data from Mobile Bay do not support such an explanation for several reasons. First, agglutinated foraminifers occur in the cores to depths of 48 cm, while calcareous foraminifers are absent or in decline during the same approximately 150 year record (Fig. 5). Simple organic cement degradation cannot explain both trends. Second, downcore degradation has been shown to affect weakly cemented agglutinated species first (Goldstein and Watkins, 1999) and thus result in a downcore assemblage change. The downcore agglutinated assemblage composition does not change in the Mobile Bay cores. Downcore faunal assemblages are comparable to the younger sediments in that there is no change in assemblage composition (Richwine and Osterman, 2012). Third, downcore destruction of agglutinated tests cannot explain the documented differences between Lamb (1972) and current surface sample results (Table 5). Downcore diagenetic processes cannot affect surface assemblages. However, the relative abundance of agglutinated (to calcareous) foraminifers in Mobile Bay has increased when comparing the two surface sets collected 40 years apart. Thus, changes in environmental conditions (e.g., decrease in pH, decrease in salinity, prolonged stratification, and hypoxia, etc.) over the last century best explain the agglutinated data.

Scenarios 2 and 3 above cannot be readily differentiated with the current data set, but as previously mentioned both reflect a similar environmental driver. Undoubtedly, the same processes that result in the dissolution of calcareous foraminifers in present-day Mobile Bay surface samples can best explain the loss of calcareous foraminifers recorded in downcore sediments beginning just over 100 years ago.

The shift to more agglutinated taxa (MB0810-7BC and -8BC), domination by agglutinated taxa (MB0810-5BC, -12BC and, -20BC), and increased decalcification began during the late 1800s (Fig. 5). The increased dissolution would have to be in response to changes in environmental conditions around the bay in the late 1800s coincidental with early dredging of the main shipping channel. Ryan (1969) showed that modifications to the bathymetry of the bay (channelization and dredge disposal) had an impact on estuarine circulation, with the most notable effect being the focusing of saltwater intrusion up the axis of the main channel and dredge disposal deflecting tidal currents. Following closely in time were changes to the upper reaches of the watershed (e.g., small dam construction, changes in farming practices, etc.) that may have modified freshwater inputs and sediment and organic matter delivery to the bay. However, given the development of upstream dams and causeways on the Mobile–Tensaw–Tombigbee waterways it is more likely that water and sediment discharge have been

reduced (Smith et al., in press). On the other hand, the major upstream waterway modifications occurred in the mid 1900s, a time later than the faunal change observed in the core sediments. Thus, we conclude that the most likely explanation is that the original ship channel construction and spoil piles limited the mixing of Gulf of Mexico marine water into the bay, especially in the western bay, which allowed the low-pH freshwater to have more influence in the bay. A similar pattern, increasing agglutinated and decreasing calcareous foraminiferal specimens, is recorded in a core from New Bedford Harbor, Massachusetts, in response to the 1964 construction of a hurricane barrier built across the mouth of the bay that also limited marine exchange (Scott et al., 2005).

Bulk organic matter in the bay also supports the idea of a transition from a more marine-influenced system to a more riverine-influenced system (Fig. 6). The general upcore depletion of $\delta^{13}C$ in the organic matter in the box cores is consistent with an increased influx of organic matter derived from the bay-head delta marsh (Table 3). What is surprising is that the organic trend appears gradual through time and throughout the estuary, whereas the foraminiferal shift appears more abrupt. This may reflect a threshold for the foraminifers in the bay with respect to freshwater and terrestrial organic-matter influx.

While we infer that the observed changes of the bay are most easily attributed to the construction of the shipping channel, other factors including climatic change and increasing regional precipitation may also play a role in the decreasing pH. Longer-scale impacts of climatic change on the benthic foraminifers are currently being investigated in a late Holocene gravity core from Bon Secour Bay.

4.3. Eutrophication of Mobile Bay

A second upcore trend observed in the Mobile Bay box cores is increasing foraminiferal density (N cc^{-1})(Fig. 6) along with the decreased diversity (primarily due to the loss of calcareous species). All Mobile Bay box cores, except MB0810-8BC, have low foraminiferal densities with a more diverse mixed assemblage from ~1850 to ~1950 and a high density, low diversity agglutinated assemblages during the last ~50 years (Richwine and Osterman, 2012). Using the modern analog, it is believed that the increasing foraminiferal densities observed in the Mobile Bay cores since ~1950 are a result of increased food availability. High density and low diversity faunal assemblages are characteristic of waters that contain increased nutrients and organic matter usually from anthropogenic sources (Watkins, 1961; Bandy et al., 1965; Nagy and Alve, 1987; Alve, 1995; Tsujimoto et al., 2008). The post-channelization circulation scenario of increased freshwater influence in Mobile Bay would allow more terrestrial organic matter to remain in the bay due to decreased flushing/estuarine mixing and could result in increasing eutrophication. Alternately, increased foraminiferal density could also be the result of increased test preservation due to rapid burial (Tsujimoto et al., 2008). However, higher foraminiferal densities also occur in the west bay (MB0810-12BC and -5BC) where over the last 50 years sediment accumulation is lower (Smith et al., 2011, in press). So the most probable explanation for the increasing foraminiferal densities is increasing nutrients and food.

Eutrophication in Mobile Bay appears to be a recent phenomenon (since 1950) and might also be linked to increased nonpoint source loading in the upper watershed and more direct anthropogenic inputs from the city and port of Mobile (ADEM, 2006; Estes et al., 2009). Increase in organic matter and decrease in C:N in the upper 5–10 cm of the box-core sediments corroborate the foraminiferal trends. Higher $\delta^{15}N$ in surface sediments also supports a slight enrichment in the nitrogen source, which may be associated with anthropogenic sources. Evidence for the continued effect of eutrophication can be seen when comparing the densities of the dead surface assemblage to the uppermost box-core sediments (Fig. 6). All dead surface samples record foraminiferal densities near or in excess of the maximum downcore measurement, except for the periodically hypoxic Site 4. Foraminiferal densities in MB0810-4BC have decreased since the ~1980 (Fig. 6). This drop in density suggests a negative response to increasing nutrients and possibly expanded hypoxia occurring at this location (Park et al., 2007a).

Another possible response to eutrophication in both the surface and core sediments is the introduction of the agglutinated species *P. simplissima* (Cushman and McCulloch, 1939) into Mobile Bay. Lamb (1972) does not report the occurrence of this species, and given its unique shape, it is unlikely that this species was misidentified as either of the other two agglutinated species *A. salsum* or *M. fusca* (Table 5). In the surface samples, *P. simplissima* comprises up to 10% of the assemblage and occurs at depth within every box core (Fig. 6) (Richwine and Osterman, 2012). Shallow water rochamminids, including *Trochammina hadai* (Uchio, 1962) and *Trochammina* cf. *japonica* (Ishiwada, 1950) are common agglutinated species found in organic-rich, oxygen-poor estuaries throughout Japan and the west coast of the USA (Nomura and Seto, 1992; Takata et al., 2006; Tsujimoto et al., 2008; Nomura and Kawano, 2011; McGann et al., 2012). *Paratrochammina simplissima* is believed to be closely related environmentally to these other shallow water trochamminids. There have been several reports of agglutinated species migrating into polluted waters (Watkins, 1961; Bandy et al., 1965). This is especially true of agglutinated species migrating into a region where they can be more competitive when the calcareous species are removed by environmental conditions such as lowered pH (McGann et al., 2003). The first common occurrence (FCO) of *P. simplissima* (Fig. 6) occurs earlier in the more industrialized western bay (MB0810-5BC, ~1920–1930) and most recently in the least impacted site of MB0810-20BC (~1980). The introduction of *P. simplissima* into Mobile Bay is most likely a combination of several factors including increasing eutrophication, decreased salinity, the loss of calcareous species, and reduced competition. It is an indication of significant recent environmental change.

One contrary observation is that the density (N cc^{-1}) maximums (Fig. 6) in the upper part of the eastern cores (MB0810-4BC = 1000 N cc^{-1}, and -8BC = 1400 N cc^{-1}) are slightly greater than the maximum densities recorded in the western cores (MB0810-5BC = 600 N cc^{-1}, and −12BC = 800 N cc^{-1}), which is the opposite trend observed in the surface sediment values (Table 3). This observation is best explained by the greater taphonomic loss of calcareous foraminifers in the western bay. However, in both surface and subsurface core sediment, the lowest maximum density is found in MB0810-7BC (100 N cc^{-1}). The very low density observed throughout MB0810-7BC can be explained by several factors associated with the channelization. The chronology of MB0810-7BC indicates that the sediments deposited prior to the dredging of the ship channel were not penetrated at this site. Ryan (1969) calculated that one area of high sediment accumulation in the last 100 years was in the former tidal channel where the coarsest-grained (Table 3) MB0810-7BC is located. The rapid post-channel, sediment accumulation and short core length may explain why older sediment, with higher foraminiferal diversity, was not obtained at this site. The pronounced difference in foraminiferal density between MB0810-7BC and -8BC is seen as evidence that the channel and spoil bank effectively cut off the supply of marine salinity and negatively impact the fauna of the western bay. There is an order of magnitude difference in densities across the mouth of Mobile Bay; MB0810-7BC has the lowest

density and MB0810-8BC has the highest. Prior to channelization the foraminiferal densities were likely more similar.

5. Conclusions

Foraminifers and sediments from seven box cores with $^{210}Pb_{xs}$ chronology document that channel dredging and spoil disposal have altered circulation, reduced estuarine mixing, and changed sedimentation patterns in Mobile Bay. However, the most surprising result of this study is the documentation of the loss of Mobile Bay calcareous foraminifers linked to a reduction in water exchange with the Gulf of Mexico. Beginning in the mid 1800s restricted marine exchange allowed for greater low-pH freshwater influence in the bay, and ultimately began environmental changes that resulted in increasing calcareous test dissolution. By the late 1800s all sites throughout Mobile Bay record the initial loss of calcareous foraminifers, a trend that continues to the present. Since the completion of the current dredged channel (1950), agglutinated foraminiferal densities have climbed in a response to the increased organic matter and reduced estuarine mixing. Yet in deeper regions of the bay, hypoxic water formation has caused decreased foraminiferal densities and accelerated calcareous dissolution. Comparisons with foraminifers collected in the early 1970s indicate that the dissolution of calcareous foraminifers has allowed the introduction of a new agglutinated species into the bay.

The marine microfauna of Mobile Bay has been profoundly influenced by the development and expansion of the primary shipping channel over the last ~100 years. It is not difficult to predict that some impact would be observed in the biota, but the continued and accelerated degradation during the last 100 years was unexpected. This biologic degradation agrees with the observed loss of oysters and submerged aquatic vegetation from Mobile Bay during the last century (Estes et al., 2009).

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