**Biogenic sedimentation and geochemical properties of deep-sea sediments of
the Demerara Slope/Abyssal Plain: Influence of the Amazon River Plume**

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ABSTRACT

The Amazon River Plume has a major impact on sedimentation at abyssal depths throughout a broad area (450,000 km²) and far (500-1000 km) from the river mouth. As the plume mixes into the North Atlantic Ocean, it may support ecological niches favorable for N₂ fixation driven by the symbiosis between diatoms and diazotrophic cyanobacteria. This proposed diatom-diazotroph association (DDA) should result in increased carbon export from a limited region of plume influence. Here we used 32 multicores collected throughout the Demerara Abyssal Slope and Plain at depths ranging from 3000-5000 m to assess patterns and rates of export and accumulation of organic carbon (C_{org} or POC) and biogenic SiO₂ (bSi). Sediment accumulation rates were determined from ¹⁴C ages of foraminifera obtained at 2-5 depths from 9 cores and were further constrained by ²³⁰Th measurements. The sedimentation rate (1-5 cm/ky), representing the average accumulation over the past 8000 years, was combined with C_{org} and bSi sediment concentrations to establish biogenic matter burial fluxes. Burial of POC represents 4-16 % of the C_{org} raining to the deep sea floor, which is high for water depths >4000 m. Burial of bSi represents 1-9% of the total bSi rain. Clastic sedimentation dominates regions underlying the plume, while carbonate sedimentation dominated otherwise. Sediment mixing (bioturbation) was studied in 13 cores using ²¹⁰Pb as a tracer and showed no spatial patterns. Diffusional mixing coefficients ranged from 0.03 to 3.2 cm² yr⁻¹, values consistent with other deep sea mixing rates, however, sites with the highest mixing rates did not have the highest C_{org} content. Sediment focusing was studied in at 7 sites using ²³⁰Th as a tracer; surficial sediments had

focusing factors that ranged from 1-2, indicating little sediment redistribution. Biogenic sedimentation in general and bSi deposition in particular is not concentrated in the region of DDA abundance. The NW axis of primary plume trajectory is coincident with the maximum deposition and burial of POC and bSi but not CaCO_3 .

Keywords: carbon, biogenic silica, carbonate, sedimentation, deep-sea, bioturbation

1. INTRODUCTION

The outflow of the Amazon River forms a plume of low salinity (<36 ppt) water that extends to the northwest over the north Brazilian continental shelf and then passes over the Demerara Slope and Abyssal Plain. These waters can be traced northwest to the Caribbean (Froelich et al., 1978; Goes et al., 2013; Lentz and Limeburner, 1995; Neumann, 1969; Smith and DeMaster, 1996), as well as eastward into the eastern North Atlantic due to the seasonal retroflection of the North Brazil Current (Muller-Karger et al., 1988). The plume is responsible for the delivery of both nutrients and organic matter to the deep western tropical North Atlantic (WTNA) (Aller et al., 1996; DeMaster and Pope, 1996; DeMaster et al., 1996; Richey et al., 1990; Smith and DeMaster, 1996), as well as dispensing 1.2×10^9 tons yr^{-1} of suspended sediment (Meade et al., 1985), though the majority of the sediment and nutrient load is deposited or utilized in continental shelf waters.

CO_2 uptake has been observed in the WTNA, particularly in the vicinity of the Amazon River plume, in an area of the Atlantic generally considered to be a net

source of CO₂ to the atmosphere (Cooley et al., 2007; Cooley and Yager, 2006; Mikaloff Fletcher et al., 2007; Takahashi et al., 2002; Terner et al., 2000). The areal extent of this uptake and its impact on the sediment column has not been previously studied. The plume contains waters that are depleted in N but contain a relative excess of P, Fe and dissolved Si (Subramanian et al. 2008, Goes et al. 2013), creating ideal conditions for the proliferation of diatom-diazotroph associations (DDAs), which have been suggested to be a factor in carbon export (Carpenter et al., 1999; Foster et al., 2007; Shipe et al., 2006; Subramanian et al., 2008). Goes et al. (2013), following the work of Subramanian et al. (2008), identified a 'sweet spot' of enhanced DDA abundances in a location associated with salinities of 31-34 ppt. Using oxygen isotope tracers, Yeung et al., (2012) showed that DDA blooms do increase carbon export efficiency, but it is unknown how much POC or bSi reaches the deep ocean.

Previous work in the WTNA utilizing a moored sediment trap at 13.5°N 54.0°W (the northern limit of our study region; Fig. 1) showed that the mass flux at 3755 m and 5068 m was nearly identical (46.4 and 47.0 mg m⁻²d⁻¹, respectively) and there was little temporal variability between Nov-1977 and Feb-1978 (Honjo, 1980). A later study found that the mass flux of sinking material at 3200 m at 13.22°N, 57.68°W displayed a high amount of inter-annual variability (42 ± 24 mg m⁻² d⁻¹ Jun 1981-Apr 1982; 115 ± 47 mg m⁻² d⁻¹ Apr 1982-Jul 1983) and suggested that the variability was derived from variability in surface blooms in Amazon plume waters (Deuser et al., 1988). Floating sediment traps deployed by Subramanian et al. (2008) showed a correlation between the presence of DDAs and the overall mass

flux at 150 m in waters of salinity ~ 32 ppt (152 ± 26 mg m⁻² d⁻¹ at stations with salinity 30-34 versus 42 ± 8 mg m⁻² d⁻¹ at stations with salinity >35), but this is not entirely diagnostic for quantifying the amount of biogenic material that reaches the sea floor, nor were DDAs identified as the primary vector of export.

Here, we describe the compositional variability of surface sediments and patterns of biogenic sedimentation and accumulation throughout the Demerara Rise/Abyssal Plain. We construct a budget of C_{org} and bSi rain (remineralization + accumulation) using remineralization fluxes established from pore water modeling (Chong et al., 2014). We also examine the pattern of sediment mixing and sediment re-distribution (focusing) throughout the region.

2. STUDY AREA

This study presents results from the ANACONDAS (Amazon iNfluence on the Atlantic: CarbOn export from Nitrogen fixation by Diatom Symbioses) project. Three cruises took place throughout the WTNA aboard the *R/V Knorr* (May-June 2010), the *R/V Melville* (September-October 2011), and the *R/V Atlantis* (July 2012). Most of the coring sites are quite far from the river mouth but still clearly under river plume influence as inferred from satellite imagery. The study region covers an approximate area of 450,000 km² (Fig. 1).

The mean flow of the Amazon River is 1.8×10^5 m³ s⁻¹, discharging 1.2×10^9 tons y⁻¹ of suspended sediment, most of which moves northwest in the form of suspensions and coastal mudwaves (Kineke et al., 1996; Meade et al., 1985; Nittrouer et al., 1995; Wells and Coleman, 1981). Within the surface sediments of

these shelf deposits, both terrestrial and marine organic matter, dominated by the marine fraction (60-70%), contribute to early diagenetic reactions (Aller and Blair, 2006; Aller et al., 2004). Rapid sediment mixing and suboxic diagenesis depletes the inventory of terrestrial organic material (Aller and Blair, 2006). Although most Amazon River sediments appear to be trapped on the continental shelf (Damuth and Embley, 1981; Damuth and Kumar, 1975; Milliman et al., 1975), the plume may carry a significant load of fine, suspended material far into the N. Atlantic. The sediments of the continental rise and abyssal plain adjacent to the Amazon mobile mud belt consist predominantly of brown pelagic and grey hemipelagic sediments with sedimentation rates ranging from 4 to 7 cm kyr⁻¹ on the lower continental rise to 2 to 4 cm kyr⁻¹ on the abyssal plain (Damuth, 1977).

3. METHODS

3.1 Sediment Cores

We collected sediment cores at 32 sites (Fig. 1, Table 1) throughout the Demerara Slope/Rise and Abyssal Plain using a multi-corer with 9.8 cm diameter (I.D.) core tubes (Barnett et al., 1984). After recovery on deck, cores were immediately transferred to a temperature-controlled van (~2°C). One core from each station was sectioned in 1 cm intervals from the sediment-water interface (SWI) to 5 cm, and in 2 cm intervals thereafter. Two aliquots of mud were collected from each interval. The first aliquot was collected into a pre-weighed glass container, which was used to determine the sediment porosity by weight loss after drying. Dried sediment was ground to a fine powder using an agate mortar and

pestle. The second aliquot was sieved (200- μ m mesh) and the material remaining on the mesh was rinsed with de-ionized water and dried.

3.2 Total Carbon, Organic Carbon, Carbonate, and $\delta^{15}\text{N}$

Weight percent total carbon (TC) and wt% inorganic carbon (TIC) were measured on aliquots of ground sediment. Total carbon was analyzed using 15-50 mg of ground sediment in an Elantech 1110 Elemental Analyzer (EA) connected to a Picarro G2121-i isotopic CO_2 analyzer. The standard used for TC analysis was developed in our lab by powdering a Pre-Cambrian sedimentary rock (AR-15) collected from Death Valley, CA (F. Corsetti, pers. comm.) that records the Shuram isotope excursion. The chemical composition was standardized by analysis of 7 replicates (data not shown) at the UC Davis Stable Isotope Laboratory (<http://stable-isotopes.geology.ucdavis.edu/>) yielding an average C composition of 8.23 ± 0.01 ‰C. The relative uncertainty of the TC measurements, based on the uncertainty of the AR-15 standard from every run, was <1.5% of the value measured.

The wt% TIC was also measured on the Picarro analyzer using 15-50 mg of ground sediment weighed into quartz test-tubes capped with gas-tight septa (10 ml vol). Each tube was evacuated for 30 seconds, injected with 3 mL of 10% H_3PO_4 and vortexed for 1 minute to homogenize the mixture. Samples sat for a minimum of 30 minutes at room temperature prior to analysis. The standard used for TIC analysis was an optical calcite (OPT) standard (12.002 ‰C). The relative uncertainty in wt% of TIC was determined from the average uncertainty of the standards from every

run. This averaged 2.5% of the wt % measured. Weight percent C_{org} was determined as the difference between wt% TC and wt% TIC. This determination by difference leads to an uncertainty in wt % C_{org} of $\pm 10\%$ of the value measured. Some discussion of the precision and accuracy of the Picarro analysis can be found in Subhas et al. (2015).

$\delta^{15}N$ was measured on bulk sediments from the two uppermost depth intervals from the 2010 and 2011 stations. Samples were analyzed at the UC Davis stable isotope lab. Uncertainty in standard replicates was $\pm 0.1\text{‰}$.

3.3 Biogenic Silica

Solid biogenic SiO_2 (bSi) was measured using an alkaline leaching method (DeMaster, 1981; Strickland, 1968). Approximately 50 mg of ground sediment was placed in a 50 mL plastic centrifuge tube with 50 mL of 5% (wt/vol) of a sodium carbonate solution. The samples were incubated in an 80°C water bath for 5 hours, with periodic mixing. Aliquots (0.5 mL) were removed at hours 3, 4, and 5 into pre-weighed plastic test tubes. The samples were re-weighed, neutralized with dilute hydrochloric acid, and analyzed colorimetrically using the molybdate blue assay. Blanks were also analyzed to correct for colorimetric variations from using different batches of Na_2CO_3 . Dissolved Si concentrations were plotted versus time, and a linear fit to the data was extrapolated back to time zero, where the difference between the blank and the intercept was assumed to be the amount of Si released from biogenic SiO_2 . The product of molar concentration of Si times leach solution

volume was used to calculate wt% bSi, assuming a molecular weight of 60 g/mol.

The uncertainty in wt% was $\pm 5\%$ of the measured value.

3.4 ^{210}Pb

Lead-210 activity was measured using a high purity intrinsic germanium well detector (HPGe ORTEC, 100 cc active volume) in the radioisotope lab at USC. Standards used included IAEA-315 marine sediment and EPA diluted pitchblende SRM-DP2. Measurements were made at 7 depths between the SWI and 20 cm from 13 cores (Table 2). Approximately 1.5 g of dried, ground sediment was used for each analysis. Corrections for sample geometry were made based on the sample height within the capped 9-mm measurement test-tube. Samples were counted for 1-6 days and the gamma spectra (keV) were analyzed for the nuclides ^{210}Pb (46), ^{214}Pb (296, 352), ^{214}Bi (609), and ^{137}Cs (662). Supported ^{210}Pb activity was found by calculating a weighted average for ^{214}Pb and ^{214}Bi activity, and using a correction of 1.18 ± 0.10 to account for the loss of ^{222}Rn . This correction was based on measurements of ^{222}Rn loss from powders that were prepared similarly and assumes that each sample is of uniform texture and comparable grain size distribution. Integrated excess ^{210}Pb activity was calculated as the difference between total and supported ^{210}Pb summed over the depth containing excess. The analytical uncertainty in a measurement was typically 35 Bq kg^{-1} . There may be additional uncertainty in the excess ^{210}Pb profiles due to the uncertainty in the value of supported ^{210}Pb stemming from the assumed loss of ^{222}Rn , but we believe these to be small relative to other uncertainties.

3.5 ^{14}C

Sieved material from 2-5 depths from nine sites was examined under a microscope to determine the type and quality of material present. The visual examination of the sieved material used for ^{14}C dating showed that it was primarily comprised of planktonic foraminifera. Bulk sieved material was hand selected, excluding miscellaneous grains (black, detrital material), and aliquots of ~40 mg of material were sent to the UC Irvine Keck AMS facility for radiocarbon analysis (Table 3). The ^{14}C ages were converted to calendar age using the Calib 6.0 conversion program and the Marine09 calibration curve (Reimer et al., 2009; Stuiver and Reimer, 1993). The uncertainty after calibration was $\pm <150$ yr. The marine calibration assumes a global ocean reservoir age correction of 400 years, which was adjusted for local effects following Stuiver and Braziunas (1993) using a ΔR value of -1 ± 42 from the Marine Reservoir Correction Database (<http://www.calib.qub.ac.uk/marine>).

4. RESULTS

4.1 Organic Carbon, Total Carbonate, and $\delta^{15}\text{N}$ in Sediments

Cores obtained from twenty-eight stations were analyzed from 0-45 cm (at 1-2 cm intervals) for Total Carbon and Total Inorganic Carbon. At all stations, the difference between these two ($= C_{\text{org}}$) content is generally < 1 wt% (Fig. 2) and declines with depth. Stations closest to the river mouth (Fig. 2A) have slightly higher surface C_{org} content (0.75-0.97 wt%) compared to those in the north (0.46-0.89

wt%) and the east (0.23-0.76 wt%). Overall, the C_{org} profiles decrease with depth in a somewhat exponential fashion, reflecting remineralization with depth, yet it is not clear that the profiles reach an asymptotic value over the depths sampled.

Weight percent TIC was converted to wt% $CaCO_3$ and surface sediment values range considerably, from 4.7 to 55.4 wt% calcium carbonate (Fig. 2D-F). Stations closest to the river mouth and within the main axis of the NW-trending plume (Hu et al., 2004) were generally low in surface $CaCO_3$, ranging from 4.7 – 15.0 wt%. Downcore, these stations generally increase in wt% carbonate with depth. To the north, surface $CaCO_3$ range from 9.3 – 28.6 wt% and downcore sediment profiles remain fairly consistent or increase slightly to ~20-30 cm, below which $CaCO_3$ fraction decreases. The eastern stations show greater variability in surface $CaCO_3$, ranging from 8.8 – 55.4 wt%, and downcore, nearly all of the stations in this region show a precipitous decrease below 20 cm with the exception of station 2011-9 that increases with depth.

Surface sediment $\delta^{15}N$ isotopes (PON from bulk sediments) were generally uniform throughout the study region, ranging between 4.50-5.16‰ (Fig. 3). Although contours can be placed on this plot showing most sites in the middle of the study region with higher $\delta^{15}N$ values, the differences in either the N-S or E-W were very small.

4.2 Biogenic Silica

Biogenic SiO_2 in the sediment was measured at 15 stations throughout the study region (Fig. 4). Surface values range from 0.62 wt% to 2.55 wt%. Downcore,

most stations show a decrease with depth, reflecting remineralization. Station 2011-13 has constant values (~ 2.5 wt%) through the upper 15 cm before decreasing. All other sites show a decline with depth below the SWI. At 2011-20, the southern-most station, the bSi composition is low (< 0.7 wt%) throughout the entire sediment profile.

4.3 ^{210}Pb

Excess ^{210}Pb in surface sediments varies from $65 - 1146 \text{ Bq kg}^{-1}$, and most profiles show a generally exponential decrease with depth (Fig. 5). Most of the profiles approach the background value of ^{210}Pb by 12 cm, though some show excess ^{210}Pb persisting to ~ 16 cm. Station 2010-27 shows an excess at depths below 10 cm, however, the profile cannot be fit with an exponential model, and is not included in subsequent discussions.

4.4 ^{14}C Calendar Ages

The overall calendar ages of the sediments analyzed range from 614 ybp to 28,069 ybp (Table 3). The age of sediments from 4.5 cm are youngest at stations 2011-9 and 2010-19 (663 ybp and 1407 ybp, respectively). These stations lie close to the axis of the plume. The surface age increases to the northwest at stations 2011-1 and 2011-31 (2284 ybp and 3573 ybp), and the sediments are the oldest at station 2011-23 (4985 ybp) in the east.

All sedimentary data, that presented here, as well as additional unpublished data, is available in tabular form in Chong (2013) and on the ANACONDAS data repository web site (<http://www.bco-dmo.org/project/2097>).

5. DISCUSSION

5.1 Sediment Mixing

Bioturbation influences the physical structure of the sediments and can also impact the burial efficiency of biogenic constituents (Green et al., 2002; Sun et al., 1993). We assessed the amount of sediment reworking using measurements of the radionuclide ^{210}Pb . In the absence of sediment mixing, excess ^{210}Pb should only be found in the uppermost millimeters given the low sedimentation rates in this area and the lifetime of ^{210}Pb (~100 years). Any excess ^{210}Pb in sediments at depths > mm will have been introduced, presumably, by macrofaunal mixing (Aller, 1982; DeMaster et al., 1994; Rhoads, 1974).

Measurements of excess ^{210}Pb in a sediment profile also allow for an estimate of the bioturbation diffusional mixing coefficient (D_b) and the inventory of excess ^{210}Pb in the sediments. Another measure of mixing provided by ^{210}Pb is the depth of the sediment mixed layer (ML). To determine D_b , we assumed that excess ^{210}Pb input is constant and bioturbation is a diffusion-like process (Berner, 1980; Guinasso and Schink, 1975). The fit to the excess ^{210}Pb profile, using the steady-state diagenetic equation as described by Stephens et al., (1997):

$$A = A_0 \exp(-\alpha z) \quad (1)$$

where: z is depth in cm, A is the activity of excess ^{210}Pb in Bq kg^{-1} , A_0 is the activity concentration at $z=0$, and α is the attenuation coefficient (cm^{-1}). A_0 was allowed to be a free parameter in the model fit to the data. The attenuation coefficient is related to D_b by the following:

$$\alpha = (\lambda/D_B)^{1/2} \quad (2)$$

where λ is the radioactive decay constant for ^{210}Pb (yr^{-1}).

Values of D_b ranged from 0.07 to 3.2 $\text{cm}^2 \text{yr}^{-1}$ with most values around 1.6 $\pm$ 0.4 (Table 2). These values are consistent with the mixing rates observed in the Equatorial Pacific (<0.1-16.9 $\text{cm}^2 \text{yr}^{-1}$) (Stephens et al., 1997) and similar to other estimates from the deep ocean (Barsanti et al., 2011; Basso et al., 2004; DeMaster et al., 1985; Kuehl et al., 1993; Teal et al., 2008; Thomson et al., 1993). There was no significant spatial pattern of D_b throughout the study area. However, there does appear to be an inverse relationship between D_b and surface sediment wt% C_{org} (Supplemental Figure S1); in areas of high % C_{org} , the D_b was quite low (i.e. stations 2010-20, 2010-21), while where C_{org} was low, the D_b was high (i.e. stations 2010-26, 2011-23). This is curious, because it is generally believed that higher rates of mixing are correlated to higher macrofaunal abundance, and thus higher input and availability of C_{org} . Here, high macrofaunal activity apparently acts to transport the incoming organic carbon deeper in the sediment.

The relationship between higher wt % C_{org} and lower D_b was examined quantitatively by first assuming that the concentration of C_{org} in the sediments can be described by an exponential equation that approaches a constant value at depth:

$$C = C_0 \exp(-\beta z) + C_{\text{inf}} \quad (3)$$

where C_0 is the concentration of C_{org} at the SWI, C_{inf} is the baseline concentration of C_{org} at depth, $\beta = (k/D_b)^{0.5}$, and k is the rate constant (1/time) of C_{org} degradation.

Using Fick's first law to describe the C_{org} flux into the sediments, we obtain

$$C_{org} \text{ flux} = -D_b \frac{dC}{dz} \quad (4)$$

Inserting Eq. 3, differentiated and solved for $z = 0$ yields

$$C_{org} \text{ flux} = C_0 \sqrt{k D_b} \quad (5)$$

If the flux and k are constant, there should be a linear relationship between C_0 and

$[\sqrt{D_b}]^{-1}$. A quasi-linear relationship is observed (Supplemental Figure S1), which

indicates that bioturbation rate may have some influence on surface C_{org}

concentration. Some of the non-linearity in this plot likely stems from assumptions

about constant C_{org} flux and k , or to variable inputs of refractory C_{org} . We will

explore refractory C_{org} later in this discussion.

If scavenged ^{210}Pb is carried by sinking particles to the deep ocean, and there is no sediment focusing, then water column ^{210}Pb -deficiency should be equal to the inventory of excess ^{210}Pb found in the sediments. To calculate the sediment inventory, we converted the ^{210}Pb data to disintegrations per minute (dpm) and calculated the integrated sediment density based on the porosity profile over the upper 17 cm, interpolating values for depths not counted and extrapolating the trend linearly to the sediment surface. We graphically integrated the amount of excess ^{210}Pb in each layer, obtaining the sum of excess ^{210}Pb , which ranged from 39 to 133 dpm cm^{-2} . Strict error propagation from the counting data indicates inventory uncertainties less than 5%, but given uncertainties in the interpolation

and perhaps in the porosity measurements, we estimate the inventory might have an uncertainty of 15%. Stations with a higher ^{210}Pb inventory appeared to correlate to the stations with elevated wt% C_{org} , south of 10°N and west of 50°W (Table 2).

Cochran et al., (1990) estimated the total water column deficiency of ^{210}Pb at various stations throughout the North Atlantic. At stations in the tropical north Atlantic closest to our study area (15°N , 48.5°W), the water column deficiency was estimated to be $54 \pm 6 \text{ dpm cm}^{-2}$ at a 3800 m site, and $82 \pm 8 \text{ dpm cm}^{-2}$ at 4900 m, resulting in an average of $68 \pm 19 \text{ dpm cm}^{-2}$ across our study sites. Many of our stations fall within this inventory range, though 2010-19, 2010-21, and 2011-8 were higher than the average by $> 1\sigma$. These three stations with high inventories of excess ^{210}Pb may have experienced sediment focusing by deep ocean currents. We examined this possibility by using ^{230}Th -normalized fluxes to calculate sediment focusing factors (Ψ) following Francois et al. (2004). ^{230}Th activity was measured from surface sediments (1-3 cm), and focusing factors were evaluated as the accumulation rate of unsupported ^{230}Th in the sediment divided by the production rate of ^{230}Th in the overlying water column (Table 4). Generally, ^{230}Th showed relatively minor effects of sediment focusing, as most ratios are around 1. We compared Ψ with the inventories of excess ^{210}Pb normalized to the average water column inventory of 68 dpm cm^{-2} (Fig. 6). Values close to 1 indicate that all of the water column ^{210}Pb is present in the sediments below, while greater values imply the import of ^{210}Pb to the area. Stations in the region south of 10°N and just west of 50°W showed relatively high values for the inventories of excess ^{210}Pb , while the ^{230}Th -derived focusing factors were close to 1 (1.05-1.07). Overall, there does not

appear to be a relationship between the normalized inventory of ^{210}Pb and ^{230}Th -derived focusing factors.

A large inventory of excess ^{210}Pb does not necessarily mean that there is additional sediment being delivered to a particular site. In areas with a high sedimentation rate, the amount of ^{210}Pb scavenged can be inflated by the addition of ^{210}Pb advected into the area from nearby waters with a lower sedimentation rate. The additional ^{210}Pb could also be brought in from outside the study area, perhaps from the continental shelf and/or from particles carried by the river plume. Suspended riverine material has an excess ^{210}Pb activity ranging from $600\text{--}2100 \pm 360 \text{ dpm kg}^{-1}$ ($10\text{--}35 \pm 6 \text{ Bq kg}^{-1}$), and activities of excess ^{210}Pb from the seabed of the Amazon shelf range from $20,010\text{--}120,000 \text{ dpm kg}^{-1}$ ($333.5\text{--}2000 \text{ Bq kg}^{-1}$) (DeMaster et al., 1986; Smoak et al., 1996).

The decoupling between ^{230}Th focusing and ^{210}Pb inventory may also be caused by differences in their particle reactivity – ^{230}Th is highly particle reactive, but ^{210}Pb has the potential to be preferentially attached to sinking C_{org} . It has been previously documented that POC export flux can be correlated to ^{210}Pb flux (Cochran et al., 1990; Moore and Dymond, 1988), however, this correlation is not ubiquitous (Smoak et al., 1999). A plot of the ^{210}Pb inventory versus C_{org} and bSi rain rates (discussed later) shows a general, positive correlation indicating that ^{210}Pb could be a proxy for C_{org} and bSi rain in this region (Fig. 7). Variation in figure 7 may result from the fact that ^{210}Pb is not completely scavenged from the water column.

The depth of the mixed layer was determined by examining where excess ^{210}Pb reached the typical measurement detection limit of ± 75 Bq/kg (Fig. 5). Generally, this depth was ~ 12 cm below the SWI, though it was as shallow as 6 cm at station 2010-20. With a half-life of 22.3 years, mixing determined by this tracer is only applicable to sedimentary processes that occur on the order of 30-100 years (Cochran et al., 1990; Craig et al., 1973; Rama and Goldberg, 1961). Downcore profiles of C_{org} , CaCO_3 , and bSi indicate which sedimentary constituents are strongly affected on this time scale (Figs. 2, 4). We would expect constituents with much longer residence times than ^{210}Pb to have a near-constant concentration in the upper 10 cm of the profile. The profiles of C_{org} do not show this consistency, rather, many profiles decrease exponentially with depth within the mixed layer, indicating regeneration of C_{org} on less than centennial time scales. In contrast, the upper 5-15 cm of the CaCO_3 profiles appeared to be well mixed for nearly all of the stations. This suggests that the relative flux of CaCO_3 has not varied substantially over the century time scale, and that the impact of any diagenetic reactions that involve CaCO_3 must occur either very near the surface of the sediment, or deeper than 10-15 cm. Two of the bSi profiles, stations 2011-6 and 2011-13, appeared to be mostly homogenous to ~ 8 cm and 16 cm, respectively, however, the rest of the stations had a significant concentration gradient in the upper sediments.

5.2 Sedimentation Rates

A linear sedimentation rate below the ML was established with a regression of age vs. depth (Table 5, Fig. 8). At most stations, the age was measured within the

ML, using a sample obtained from 4.5 cm. Sedimentological studies on ANACONDAS sediment cores (Lund et al., in prep) show little evidence of physical mixing much below 6 cm, and a previous study on the Iberian Abyssal plain found that the ML from excess ^{210}Pb profiles and a high-resolution profile of ^{14}C ages was ~ 5 cm for both tracers (Carvalho et al., 2011). However, our ^{210}Pb data suggest that the ML may extend to 12-16 cm. As such, we applied a linear fit to the data assuming three different ML depths: 4.5 cm, 10 cm, and 15 cm. Assuming the depth of the ML extended to 4.5 cm, we included this data point in the linear fit. Alternately, we extrapolated the age of the ML at 4.5 cm to a depth of 10 cm and 15 cm to explore how this changed the calculation of sedimentation rate. For the purposes of this discussion, we have chosen to assume a ML at an average depth of 10cm as an estimate of the true sedimentation rate. Given this baseline, a ML of 4.5 cm would increase the sedimentation rate by 18%, and a ML of 15 cm would result in a decrease in the sedimentation rate of 16% (Fig. 8)

Three stations only had 2 depths from which to calculate the sedimentation rate and are therefore more sensitive to potential errors in the ^{14}C age estimates. At these stations, the trend from two data points was extrapolated to 8 cm. At stations 2010-21 (from which we extrapolate three age dates) and 2010-27, this extrapolation yielded an age of the ML that is within the range of the other stations, and so we believe these rates to be reasonable. In contrast, stations 2011-20 and 2011-25 extrapolated to ages that are not consistent with an average ML age of <2000 ybp, and these stations would require a significant decrease in the sedimentation rate in the upper 20 cm in order to agree with our estimations of the

ML depth and age. The age of the sediments from these cores correspond to the earliest Holocene, and so it is possible that the sedimentation rates were higher at that time due to lower sea levels. However, it is also possible that recent sediments are missing from this site, either due to loss during coring (unlikely as ^{210}Pb profiles appear typical) or erosion prior to 100 years ago. Since we do not have a good constraint on the Holocene sedimentation rate at stations 2011-20 and 2011-25 due to the limited sampling resolution, we have chosen to omit them from the subsequent discussion of biogenic burial fluxes below.

Since the age of the mixed layer is a function of the sedimentation rate and the ML depth, we employed a simple box model (Broecker et al., 1999) to assess sedimentation rate at the estimated age of the ML and compare it to the sedimentation rate from deeper sediment ages (Supplemental Figure S2). Assuming steady state, the ^{14}C activity of incoming material to the ML will be equal to the loss due to burial and decay. Assuming the depth of the ML is 10 cm, and a reservoir correction of 400 years for planktonic foraminifera (Broecker et al., 1999), we calculated the sedimentation rate within the ML and compare it to the sedimentation rate below. Results (Supplemental Table ST2) from six stations show that the sedimentation rates within the ML are generally higher than the rates below (4.7 ± 3.2 cm/kyr for ML, 3.4 ± 1.3 below ML). However, the ML sedimentation rate is highly dependent on the ML depth and mis-estimation of the ML depth will affect the calculation of the ML sedimentation rate. Given the uncertainties discussed, it is our interpretation that sedimentation rates have generally remained constant since the middle Holocene.

Our estimates of the sedimentation rate below the ML ranged from 1.2-4.7 cm kyr⁻¹, which are in good agreement with previously reported rates for the Demerara Abyssal Plain (2-4 cm kyr⁻¹) (Damuth, 1977). Higher sedimentation rates occur in the area south of 10 °N and west of 50°W (Fig. 9); within this area, the average sedimentation rate is 4.7 cm kyr⁻¹. North and east of this area, sedimentation rates are lower and generally similar, 2.9 ± 1.5 and 2.9 ± 1.1 cm kyr⁻¹, respectively. The zone of higher sediment accumulation rates includes the central study region and incorporates sediments that underlie the NW and retroflection plume positions.

5.3 Surface Sediment Distribution and Deposition of Biogenic Materials

The overall concentration of C_{org} in surficial sediments was <1 wt% throughout the study area. However, there is a distinct pattern in its distribution (Fig. 10A). The zone of the highest sediment C_{org} content is centered near 10°N 50°W, covering much of the central part of our study area. North of 10°N and east of 50°W, the C_{org} wt% was about 2x lower.

The carbonate content of the sediments displayed a pattern opposite to C_{org} (Fig. 10B). The central stations south of 10°N and west of 50°W had low carbonate concentrations (<10%), while the stations to the north and east were generally higher (~30 wt%), reaching ~55 wt% in the stations near 45 °W

The surficial sediments had a distribution pattern of bSi that is similar to C_{org} (Fig. 10C), with the highest surface sediment values appearing south of 11 °N and

west of 50°W. The bSi content of the sediments dropped off sharply further north and east decreasing from ~2.5% in the center to <1% in the north and east.

The $\delta^{15}\text{N}$ values of the surface sediments showed very little variation and there did not appear to be a distinct pattern throughout the region (Fig. 3). We also did not observe any zones with particularly light isotopic values that would have been indicative of material derived from N_2 -fixation. Amazon River NO_3^- collected at Manaus has an isotopic value of $4.1 \pm 0.03\text{‰}$ (Brandes and Devol, 2002), and the $\delta^{15}\text{N}$ of DON in the Amazon Plume region has been observed to range from 2.7-4.8‰ (Chong, 2013; Knapp et al., 2011), with the assumption that the lighter values indicate a contribution from N_2 -fixation (Carpenter and Capone, 1992; Carpenter and Romans, 1991; Montoya et al., 2002). Thus, our data indicates that either very little organic matter derived from diazotrophy reaches the deep ocean floor, or that C_{org} produced by N_2 -fixation is more efficiently regenerated than the rest of the available C_{org} during early sediment diagenesis, leaving no signal in the sediment record.

Seiter et al., (2004) regionalized the global ocean with respect to benthic biogeochemical provinces. Our data fall near the border of the GUBRACO and NOATL regions, defined as the continental margin of SE-America (Guyana) and the Northern Atlantic deep sea, respectively. For these regions, they found the mean Opal %, Calcite % and TOC % were 2.4%, 46.5%, 0.4% (GUBRACO) and 3.5%, 44.3%, 0.4% (NOATL) respectively. For our study, the mean surface sediment composition was 1.5% 21.0% and 0.67%, respectively, indicating that the deep-sea sediments

influenced by the Amazon River plume have a greater C_{org} content and less carbonate and biogenic opal than other deep North Atlantic sites.

The composition of the sediments is a function of the relative input and preservation of biogenic components and detrital material (Seiter, 2005). The pattern of detrital deposition in surface sediments provides a footprint for plume location integrated over several hundreds-1000's years. The contribution of detrital material was derived using the measured ^{232}Th concentration, assuming 13 ppm thorium in the detrital fraction, and shows surface sediments are comprised nearly entirely (~90%) of detrital material in the central zone of the plume - extending far to the east relative to the plume axis (Fig. 10D). Stations to the far north and particularly to the SE contain a larger biogenic component. The distribution of high wt% detrital sediment (Fig. 10D) in the central study area is consistent with the pattern of fraction C_{org} buried (Fig. 11A), which is highest where detrital sediment dominates. This coherence suggests that C_{org} burial efficiency may be linked to the deposition of riverine-derived particulate matter and possibly terrestrially-derived C_{org} . It has been suggested that organic matter can be protected by the micro-pores of aluminosilicate minerals or mesopore sized spaces contained within aluminosilicate aggregates (Mayer, 1994a; 1994b; Mayer et al., 2004), which may explain the patterns of increased C_{org} burial. Of note is the observation that although sedimentation rates are similar, C_{org} burial rates vary as a strong association with detrital content.

One way to extract information about the role of oceanic ecology and sediment dilution in the burial of organic matter is to calculate biogenic

accumulation fluxes. Burial fluxes were determined in two ways: by using ^{14}C -derived sedimentation rates combined with sediment composition to find mass accumulation rates (Supplemental Table ST2), and also by normalization with ^{230}Th fluxes (Table 5). Comparison of the two methods shows that they are generally in agreement (Supplemental Figure S3). However, ^{230}Th -normalized fluxes have been shown to be more precise and accurate than fluxes derived from ^{14}C -derived mass accumulation rates (Anderson et al., 2008; François et al., 2004). Thus, subsequent discussion of burial fluxes is based on ^{230}Th -normalized fluxes.

Burial fluxes were quite low, ranging from $0.01 - 0.10 \text{ mmol C m}^{-2} \text{ d}^{-1}$ and $0.004 - 0.05 \text{ mmol Si m}^{-2} \text{ d}^{-1}$ (Table 5). To assess deep sea rain rates, we added the burial fluxes to the remineralization fluxes reported by Chong et al., (2014). As remineralization dominates, the spatial patterns of C_{org} and bSi rain (Fig. 11) closely mirror the remineralization patterns observed by Chong et al., (2014). The C_{org} rain is highest nearest the river mouth, where a 'tongue' of elevated C_{org} rain trends to the northwest, following the path of the spring river plume (Fig. 11D). Somewhat elevated C_{org} rain rates were also observed to the east, extending approximately along 10°N . This latitude is at the northern edge of the fall retroflexion plume, when the majority of the Amazon plume waters are carried eastward by the North Equatorial Counter Current (Hu et al., 2004; Muller-Karger et al., 1988; Salisbury et al., 2011). The highest bSi rain also extends to the NW, but the attenuation gradient covers a larger longitudinal space than with C_{org} (Fig. 11E). Thus, while it appears that the retroflexion plume is important for C_{org} export, it may play a greater role in

enhancing the deposition of bSi in the deep ocean even though it only affects those waters for a few months out of the year.

Comparing the burial fluxes spatially, it is evident that not only are C_{org} and bSi fluxes higher in the vicinity of the plume, they are also well correlated with one another (Fig. 12). Chong et al., (2014) found that the highest remineralization fluxes are found in a zone roughly occurring south of 10 °N and west of 50 °W. The fluxes of bSi and C_{org} in this zone are higher than the surrounding area, and are also well correlated with one another (Fig. 12A). $CaCO_3$ flux is not correlated with opal flux (Fig. 12B) because preservation of $CaCO_3$ is water-depth dependent, as can be seen by the negative relationship between $CaCO_3$ flux and seafloor depth (Fig. 12C). However, we note that the burial flux of $CaCO_3$ is relatively lower south of 10 °N and west of 50 °W than to the north or to the east. While there may be individual data points that do not fit expectations, in general the spatial pattern of all of the fluxes is consistent with the Amazon plume playing a major role in controlling the fluxes of both lithogenic and biogenic phases to the sediments.

A large quantity of previous work has shown that nearly all of the C_{org} that reaches the sea floor in the deep ocean is remineralized (Bender and Heggie, 1984; Burdige, 2007; Canfield, 1994; Hedges and Keil, 1995; Premuzic et al., 1982), thus burial fluxes are generally expected to be small (<2%) relative to remineralization fluxes at our study sites where water depths exceed 4000 m. Aller (2014) demonstrated that in ‘normal marine sediments’, preservation can reach 40% when the accumulation rates are high (10-500 g cm⁻² kyr⁻¹), however, at accumulation rates <10 g cm⁻² kyr⁻¹, preservation is rarely above 10%. At our sites, we find C_{org}

burial fluxes comprise 5-17% of the POC rain (Table 5) while the mass accumulation rates range from 1.6 – 5.9 g cm⁻² kyr⁻¹ (Supplemental Table ST2). On average, in each of the three zones, burial constitutes 7-9% of the total C_{org} rain. However, in the center and, in particular, the northern section of the study area, burial can account for 11-17% of the POC rain. This is a surprisingly large fraction given the low accumulation rates. The large fraction of C_{org} buried may be inflated due to a number of uncertainties including: uncertainty in sedimentation rate (too large), uncertainty in the wt % C_{org} value (unlikely), or inappropriate extrapolation of sedimentation rate to adjoining stations. Another possibility is that there is a fraction of C_{org}, either produced or associated with the Amazon Plume that is more refractory and hence has a greater chance of burial. Chong et al., (2014) determined that approximately half of the remineralized C_{org} in the center of the study area is ‘slower’ reacting carbon, possibly of terrestrial origin, which may contribute to enhanced burial. Finally, as previously discussed, the enhanced C_{org} preservation may be the result of ‘mineral protection’ by the large amounts of aluminosilicate minerals derived from the Amazon River (Mayer, 1994b; 1994a; Mayer et al., 2004).

5.4 Downcore Patterns

The downcore profiles of the biogenic constituents reveal how patterns of deposition have changed throughout the Holocene. However, making a comparison across the region is complicated by the fact that sediment ages are not well constrained and diagenesis is not a uniform ‘filter’ in this region. Nonetheless, by

examining sediments of similar age, we were able to identify changes in biogenic material accumulation across the region.

Examining the history of biogenic material preservation on a transect following the NW plume, we observe that there are trends in deposition of C_{org} and bSi in this region. Both sediment accumulation rates are greater in the southeast (as evidenced by the depressed 5000 year isochron) and the fraction biogenic content is greater in these sediments (Fig.13). The amount of C_{org} and bSi in the sediments is fairly high to depths of >20 cm at stations that are in the south, and they remained this way to $\sim 10^{\circ}N$, $53^{\circ}W$ where there is a fairly abrupt change in biogenic content and sedimentation rate.

Downcore, there is little variation in detrital content for the majority of the sediment profiles (Fig. 14), however there was a substantial increase in the fraction of detrital material at depth at the eastern stations 2011-20 and 2011-23. This likely represents the boundary between glacial and Holocene sediments, as the increase occurs in sediments whose age is 10-14 kybp (Fig. 14C). In general, clay deposition in the Atlantic was higher during glacial times as opposed to interglacial times, resulting in lower $CaCO_3$ concentrations (Bacon, 1984; Balsam and McCoy, 1987; François and Bacon, 1991; Thomson et al., 1990). Alternatively, the wt% $CaCO_3$ may be low due to lower carbonate preservation during glacial times (François et al., 1990; Reimer et al., 2009; Stuiver and Reimer, 1993). Based on the ^{14}C calendar ages, we know that station 2011-23 reaches glacial ages at ~ 16 cm (Table 3), consistent with the marked increase in fraction detrital and decrease in carbonate.

6. CONCLUSIONS

The sediments underlying the Amazon River plume have a central zone near 10°N 50°W that contain elevated amounts of C_{org} and bSi compared to the surrounding area. The wt% of these biogenic constituents decreases by a factor of 2 to the north and east. Calcium carbonate concentration in the sediment has the opposite pattern, with the highest values appearing in the north and east. Mean surface sediment C_{org} , $CaCO_3$, and bSi wt % in the study area was 0.67% 21.0% and 1.5%.

The sediment mixed layer depth determined by profiles of excess ^{210}Pb was generally ~8-12 cm. Diffusive sediment mixing coefficients ranged from 0.03 – 3.2 $cm^2 yr^{-1}$, agreeing with other values from the deep sea, yet no spatial patterns were evident. Sites with higher bioturbation rates did not correlate to sites with higher wt% C_{org} , in fact, the inverse was found. Excess ^{210}Pb inventory in sediments was higher than the water column deficiency by ~2X at sites south of 10°N 50°W, but outside this zone, sediment inventories were within $\pm 40\%$ of the water column deficiency. ^{230}Th -derived focusing factors (from similar stations) revealed a pattern of minor sediment focusing and winnowing throughout the region that was not geographically systematic. Generally, the sediments in the study region are not subject to large re-distribution or enhanced scavenging.

Carbon-14 age dates on the >200 μm carbonate fraction (pelagic forams) define sediment accumulation rates in this region from 1.2-4.7 $cm kyr^{-1}$. Model calculations of mixed layer accumulation show that the sedimentation rate has generally remained the same (to $\pm 30\%$) since the early-mid Holocene. The average

sedimentation rate was highest in the central zone, while rates to the north and east were comparable to each other and about 50% lower than the accumulation rates in the central zone. The central region is where river plume waters are most consistently located, both during the NW axis of flow in the spring and the eastward retroflection period in the fall-winter.

Sediment accumulation rates were used to estimate burial fluxes of C_{org} and bSi, which ranged from 0.01-0.09 mmol C m⁻²d⁻¹ and 0.01-0.03 mmol SiO₂ m⁻²d⁻¹. The burial fluxes comprised 4-16% of the total C_{org} rain arriving on the sea floor. Burial fluxes of C_{org} are quite large and regionally coherent with sites of highest detrital sedimentation. This may reflect a locus of deposition of refractory terrigenous C_{org} .

Downcore profiles reveal that more C_{org} and bSi accumulates in the west than in the east, and more in the south than the north along the NW plume axis. Detrital and biogenic deposition has remained fairly constant throughout the Holocene. Stations that include the boundary between glacial and Holocene sediments show a rapid decrease in the fraction of detrital material concurrent with an increase in CaCO₃ across the transition from the last glacial period into the Holocene.

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Figure 1.

Figure 2

Figure 3.

Figure 4.

Figure 5.

Figure 6.

Figure 7.

Figure 8.

Figure 9.

Figure 10.

Figure 11.

Figure 12.

Figure 13.

Figure 14.

Figure Captions

Figure 1. Map of study sites.

Figure 2. Sediment C_{org} and $CaCO_3$. The left column represents stations from the central plume (south of $10^\circ N$ and west of $50^\circ W$), the center column represents the northern stations, the right column represents the eastern stations. Uncertainty in C_{org} was $<1.5\%$, and uncertainty in $wt\%$ $CaCO_3$ was 2.5% .

Figure 3. $\delta^{15}N$ of surface sediments

Figure 4. Downcore profiles of bSi. A) Stations from the zone south of $10^\circ N$ and west of $50^\circ W$. B) All other stations.

Figure 5. Downcore profiles of excess ^{210}Pb . A) Stations south of $10^\circ N$. B) Stations north of $10^\circ N$. Red: stations south of $10^\circ N$ and west of $50^\circ W$. Yellow: stations north of $10^\circ N$. Blue: stations east of $50^\circ W$.

Figure 6. A) ^{230}Th -normalized sediment focusing factors. B) Excess ^{210}Pb inventories normalized to an average water column inventory.

Figure 7. ^{210}Pb inventory versus biogenic rain. A) ^{210}Pb inventory vs C_{org} rain B) ^{210}Pb inventory vs bSi rain.

Figure 8. Linear fits to the calendar ages. Units of sedimentation are cm/kyr. The solid line reflects sedimentation rates calculated assuming a mixed layer (ML) depth of 4.5 cm. $\square$ and dashed lines show the sedimentation rate (italicized) calculated if the ML was extended to 10 cm. For stations without a ML age, fits have been extrapolated back to the surface. Bold black lines assume the depth of the mixed zone extends to 10 cm. Red lines reflect sedimentation rates that are inconsistent with a ML with an average age of 2 kyr. Dotted lines reflect an extrapolation to an assumed ML of 2kyr extending to 10 cm depth. Calendar age uncertainties are smaller than the size of the data points.

Figure 9. Extrapolation of sedimentation rates to surrounding stations. Colored circles indicate the stations where ^{14}C ages were measured and the color represents the sedimentation rate. The colored box around each group of stations shows where this sedimentation rate was applied to other stations. Starred stations are where bSi was measured. All stations shown had C_{org} measured.

Figure 10. Spatial patterns in surface sediment biogenic composition and calculated detrital composition.

Figure 11. Spatial map of the C_{org} , bSi, and $CaCO_3$ burial fluxes compared to rain fluxes. (Chong et al., 2014)

Figure 12. Burial fluxes. Colors represent the relative location of each station to the mouth of the river. Circles: South and west of 10°N 10°W, Squares: North of 10°N, Triangles: East of 10°W. A) Comparison of C_{org} and bSi burial fluxes. B) Comparison of $CaCO_3$ and bSi burial fluxes. C) Comparison of $CaCO_3$ burial flux and water column depth.

Figure 13. Downcore profiles along the NW axis of the river plume. Dashed line indicates sediments of similar age.

Figure 14. Downcore estimations of the fraction of sediments composed of detrital material. A) Stations with >5% decrease in wt% with depth. B) Stations with ≤5% decrease with depth and stations that increase with depth. C) Calendar age vs. % detrital material for stations with an available, constrained sedimentation rate.

Table 1. Site descriptions and data available from each station.

Station	Latitude (°N)	Longitude (°W)	Depth (m)	Traps	TC	TIC	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$	bSi	^{210}Pb	^{14}C
2010											
1	11.57	56.79	4373		✓	✓	✓	✓			
5	6.81	49.98	3975		✓	✓	✓	✓	✓	✓	
6	6.82	47.63	4088		✓	✓	✓	✓	✓		
7	7.00	45.02	4394		✓	✓	✓	✓			
9	5.92	50.73	3014	✓				✓			
13	6.50	51.11	3500*	✓							
19	8.29	50.75	4854		✓	✓	✓	✓		✓	✓
20	10.00	49.97	4854		✓	✓	✓	✓		✓	
21	9.77	50.75	4863		✓	✓	✓	✓	✓	✓	✓
23	10.62	54.39	4486	✓	✓	✓	✓	✓	✓	✓	
25	11.41	56.36	4465	✓	✓	✓	✓	✓	✓	✓	
26	11.80	54.21	4777		✓	✓	✓	✓	✓	✓	
27	12.16	52.13	5044	✓	✓	✓	✓	✓	✓	✓	✓
2011											
1	11.60	56.86	4362		✓	✓	✓	✓	✓		✓
3	10.27	54.53	4393		✓	✓	✓	✓			
6	6.85	51.35	3760		✓	✓	✓	✓	✓		
8	6.86	49.92	3997		✓	✓	✓	✓		✓	
9	6.69	48.73	4030		✓	✓	✓	✓	✓		✓
13	8.07	51.18	4450	✓	✓	✓	✓	✓	✓		
19	6.78	45.96	4205		✓	✓	✓	✓			
20	6.19	45.43	4197		✓	✓	✓	✓	✓		✓
22	7.00	44.90	4435		✓	✓	✓	✓			
23	8.50	45.00	4709		✓	✓	✓	✓	✓	✓	✓
24	9.45	46.99	4784		✓	✓	✓	✓	✓		
25	8.47	48.03	4526		✓	✓	✓	✓	✓	✓	✓
26	7.45	48.89	4200*	✓							
27	8.08	49.53	4415		✓	✓	✓	✓			
29	9.55	50.36	4784	✓							
31	10.55	52.70	4889		✓	✓	✓	✓	✓	✓	✓
33	11.32	55.00	4650		✓	✓	✓	✓			
2012											
2	9.42	52.66	4764		✓	✓	✓		✓		

* Depth estimated from nearby stations

Table 2. Excess ^{210}Pb Inventory, mixing coefficient, and sediment mixed layer depth.

(*) Data did not fit an exponential, and so D_b was not determined

Station	Depth (m)	Inventory ^c (dpm cm ⁻²)	D_b (cm ² yr ⁻¹)	D_b Error ^b ±	Mixing Depth (cm)
<i>2010</i>					
5	3975	75	1.1	0.3	12
6	4088	51	2.2	0.6	16
7	4394	54	0.3	0.2	16
19	4854	100	1.4	0.3	12
20	4854	25	0.1	0.3	6
21	4863	133	1.0	0.2	16
23	4486	61	1.8	0.4	16
25	4465	39	0.5	0.2	16
26	4777	56	1.8	0.3	16
27*	5044	55			
<i>2011</i>					
8	3997	95	3.2	0.4	12
23	4709	41	2.6	0.7	10
25	4526	53	1.9	0.4	10
31	4889	47	2.7	0.3	12
<i>Water Column^a</i>					
	3800	54			
	4950	82			

^a Total water column deficiency (in situ production plus atmospheric input) from the central tropical North Atlantic (Cochran et al., 1990)

^b Error was determined through propagation of the error of the fitting parameters.

^c Inventory error is estimated to be ±15%

Table 3. Calendar ages calculated from the $\Delta^{14}\text{C}$ measurements.

Station	UCIAMS #	Depth cm	$\Delta^{14}\text{C}$ ‰	$\pm$	Calendar Age Range ybp		Average
2010-19	117876	4.5	-207.9	1.4	1459	1355	1407
	107612	18	-300.7	1.1	2741	2540	2641
	107614	38	-562.1	0.8	7234	7063	7149
2010-21	117877	24	-455.4	1.0	3453	3314	3384
	107620	30	-353.7	1.2	5300	5085	5193
	107621	42	-572.0	0.8	7428	7289	7359
2011-1	117880	4.5	-277.0	1.2	2360	2208	2284
	107608	16	-423.6	0.9	4633	4482	4558
	117881	22	-533.6	0.9	6627	6467	6547
	117882	28	-590.1	0.8	7707	7589	7648
	107609*	34	-490.6	0.8	5872	5709	5791
2011-9	117883	4.5	-127.1	1.6	711	614	663
	107610	16	-311.8	1.0	2820	2750	2785
	107611	38	-627.5	0.7	8458	8359	8409
2011-20	117887	16	-686.2	0.8	10240	10058	10149
	117888	24	-711.1	0.7	11100	10753	10927
2011-23	117885	4.5	-447.2	1.0	5061	4908	4985
	107615	16	-699.1	0.8	10600	10441	10521
	107616	41.5	-883.0	0.8	20289	19827	20058
2011-25	117889	16	-608.0	0.8	8047	7923	7985
	117890	36	-687.7	0.8	10279	10133	10206
2010-27	117878	16	-650.6	0.8	9134	8980	9057
	117879	38	-944.8	0.5	28069	27003	27536
2011-31	117884	4.5	-183.5	1.1	3642	3503	3573
	107617	16	-480.2	0.9	5665	5547	5606
	107619	38	-696.5	0.9	10541	10334	10438

* Calendar age is significantly different from the other samples and was not used in the calculation of the sedimentation rate.

Table 4. Excess ^{210}Pb inventories normalized to the water column and sediment focusing factors derived from ^{230}Th

Station	^{210}Pb -Norm. Inventory ^a	Focusing Factor Ψ
<i>2010</i>		
5	1.1	
6	0.8	
7	0.8	
19	1.5	1.05
20	0.4	
21	2.0	1.07
23	0.9	
25	0.6	
26	0.8	
27	0.8	0.60
<i>2011</i>		
1		1.41
8	1.4	
9		0.50
23	0.6	1.60
25	0.8	
31	0.7	1.51

^a Excess ^{210}Pb inventories were normalized to a water column average of 68 dpm cm⁻²

^b Focusing factors derived from ^{230}Th -normalized fluxes

Table 5. ²³⁰Th-normalized burial fluxes.

Station	C _{org} Burial (mmol m ⁻² d ⁻¹)	bSi Burial (mmol m ⁻² d ⁻¹)	CaCO ₃ Burial (mmol m ⁻² d ⁻¹)	C _{org} Remin. ^a (mmol m ⁻² d ⁻¹)	bSi Remin. ^a (mmol m ⁻² d ⁻¹)	C _{org} buried (%)	bSi buried (%)
2010-1	0.03		0.16	0.18	0.14	15	
2010-5	0.06	0.02	0.17	0.53	0.67	11	3
2010-6	0.04	0.01	0.19	0.35	0.18	11	6
2010-7	0.02		0.22	0.40	0.19	5	
2010-9	0.07			1.13	0.89	6	
2010-19	0.06		0.07	0.45	1.35	11	
2010-20	0.03		0.07	0.41	0.28	6	
2010-21	0.06	0.02	0.04	0.47	0.34	11	5
2010-23	0.04	0.01	0.13	0.61	0.45	6	3
2010-25	0.03	0.01	0.15	0.23	0.30	12	4
2010-26	0.02		0.09	0.16	0.26	13	
2010-27	0.01	0.004	0.08	0.16	0.27	7	1
2011-1	0.03	0.01	0.16	0.29	0.48	9	2
2011-3	0.05		0.14	0.64	0.94	8	
2011-6	0.08	0.05	0.04	1.38	0.95	5	5
2011-8	0.09		0.15	0.59	0.49	13	
2011-9	0.10	0.04	0.18	0.47	0.56	17	6
2011-13	0.06	0.04	0.05	0.75	0.99	8	4
2011-19	0.03		0.24	0.21	0.34	13	
2011-20	0.01	0.01	0.34	0.23	0.40	5	1
2011-22	0.02		0.20	0.46	0.23	4	
2011-23		0.01	0.12	0.32	0.36		2
2011-24	0.07	0.02	0.12	0.65	0.25	10	6
2011-25	0.03	0.01	0.19	0.34	0.51	7	2
2011-27	0.05		0.16	0.53	0.55	8	
2011-28				0.33	0.35		
2011-30				0.54	0.26		
2011-31	0.03	0.02	0.05	0.39	0.36	8	4
2011-33	0.02		0.11	0.44	0.32	5	
2012-1				0.22	0.29		
2012-2	0.05	0.02	0.05	1.04	0.71	5	3
2012-5				1.92	0.76		

^a Remineralization fluxes from Chong et al., 2014Burial fluxes were determined by normalization to ²³⁰Th as described in Anderson et al., 2008