Helium and thorium isotope constraints on African dust transport to the Bahamas over recent millennia

Stable deposition of African dust to the Bahamas over recent millennia

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Abstract

Despite its potential linkages with North Atlantic climate, the variability in Saharan dust transport to the western North Atlantic over the past two millennia has not been well-characterized. A factor of 4 increase in dust production in sub-Saharan Africa has been attributed to the onset of Sahelian agriculture 200 years ago. The regional extent of this anthropogenic dust increase, however, remains uncertain. Additionally, while millennial-scale cold periods of the last deglaciation have been associated with strong increases in North African dust emissions, few adequate records exist to observe dustiness during the Little Ice Age, a century-scale cooling of the North Atlantic (AD 1400-1800). Here we use a set of tidal flat cores from the Bahamas to reconstruct trans-Atlantic Saharan dust transport over the past 2000 years at approximately centennial resolution. Helium and thorium isotope-based proxies of dust deposition vary by less than 20% (RSD) in these cores. After accounting for the smoothing effect of bioturbation of the sediments, we can rule out the factor of 4 change in far-field dust transport to the North Atlantic between the pre-industrial and modern era. Furthermore, variations in western North Atlantic dust ~~loads~~ deposition associated with the Little Ice Age climate anomalies appear quite modest when compared to prior climatic events of the early Holocene or the last deglaciation.

1. Introduction

Aerosols constitute one of the largest uncertainties in global climate models, and mineral dust makes up roughly 70% of the global aerosol burden (Tsigaridis et al., 2006). Predictions of future climate scenarios disagree in the magnitude and sign of changes in mineral dust deposition (Jacobson and Streets, 2009; Liao et al., 2009; Mahowald et al., 2006; Woodward et al., 2005). Thus, we are motivated to provide observational records of dust deposition during past climatic changes to test and refine models of future change. Because the Sahara is the world's largest source of mineral dust, the North Atlantic Ocean has been a focus of dust time-series investigations. Dust flux has been reconstructed at millennial resolution using deep sea sediments, in particular over the Holocene (Adkins et al., 2006; Albani et al., 2015; deMenocal et al., 2000; McGee et al., 2013). Methods that provide annual-to-sub-annual resolution cover at most the past 5 or 6 decades, such as direct aerosol collection (Prospero et al., 2014; Prospero and Lamb, 2003; Rodríguez et al., 2015), corals (Mukhopadhyay and Kreycik, 2008) or satellite optical depth (Evan and Mukhopadhyay, 2010). In this study, we sought to fill this temporal gap between the instrumental period and the late Holocene using sedimentary records from a Bahamian tidal flat. ~~With their relatively rapid sedimentation (mm/yr), these cores offer centennial resolution of key climate anomalies of the past two millennia, from modern to the pre-industrial Little Ice Age to Medieval times.~~

A sediment core~~An existing 3200 yr record of African dust deposition to the North Atlantic comes~~ from ~~sediments on~~ the Mauritanian shelf (Mulitza et al., 2010)~~. The authors distinguish eolian particles from riverine particles using grain size, and~~ records a factor of four increase in ~~fluxes of the eolian component~~eolian dust deposition beginning at the intensification of agriculture in the Sahel 200 years ago. A similar phenomenon was seen in Late Holocene lake

records from the central Sahel (Cockerton et al., 2014). The inference of anthropogenic influence is consistent with observations over the past 50 years of deforestation with increased land use concurrent with increases in dust production in the Senegal River basin (Niang et al., 2008). However, anthropogenic dust emissions from North Africa, coming largely from the Sahel, account for only 15% of total North African dust emissions, which are dominated by natural emissions from the Sahara (Ginoux et al., 2012). Additionally, much of the anthropogenic increase seen on the Mauritanian shelf was due to dust of large grain-size (>10 μm) (Mulitza et al., 2010), which is generally only found close to shore (Stuut et al., 2005). Only very fine dust (<5 μm) is be exported out into the western Atlantic (Muhs et al. 2007). Furthermore, a model reanalysis suggests that dust optical depth over the North Atlantic was close to its present value in AD 1850 (Evan et al., 2016) and a study in the Everglades of South Florida found no change in the concentration of quartz grains, presumably from Saharan dust, over the past 2800 years (Glaser et al., 2013). ~~Thus, we suspect the modern increase in Sahel dust production may not be representative of changes in the larger North Atlantic region.~~

~~A far-field site like the Bahamas may more closely represent the dust deposition throughout the (sub)tropical North Atlantic.~~ There is a seasonal aspect for expecting dust deposition very close to the source to differ from that in the far-field North Atlantic. The Bahamas area receives Saharan dust predominantly in summer (Prospero, 1999) due to the seasonal movement of the Intertropical Convergence Zone (ITCZ). In summer, the northward excursion of the trade winds at the northern boundary of the ITCZ acts to both increase Saharan dust emissions as well as to facilitate trans-Atlantic dust transport to more northerly (20-30°N) destinations (Engelstaedter and Washington, 2007; Rodríguez et al., 2015). In contrast, in winter with the ITCZ further south, total African dust emissions are reduced and eastward dust transport

occurs mostly at latitudes south of 15°N. For these reasons, we expect the modern increase in Sahel dust production was not representative of changes in the Bahamas and the western North Atlantic region. ~~As a measure of the representativeness of the Bahamas site to dust deposition over the North Atlantic, aerosol optical depth due to dust (dust-AOD) in the Bahamas was found to be strongly correlated with dust-AOD throughout the tropical North Atlantic in summer in GEOS-Chem model simulations of the past 30 years [*Ridley et al.*, 2014; *Williams et al.*, submitted].~~

On longer timescales, millennial variations in North Atlantic dust deposition have been documented back through the last ice age. ~~Whereas the Greenland ice cores provide highly resolved records of East Asian dust transport [*Ruth et al.*, 2007; *Serno et al.*, 2015],~~ Holocene and last glacial variability in Saharan dust has been limited mostly to deep ocean sedimentary records. Periods of pronounced (several degrees Celsius) cooling in the North Atlantic region, such as Heinrich stadial 1 (17-18 ka) and the Younger Dryas (12-13 ka), have been associated with increases in African dust emissions to roughly 2-3 times modern levels (Jullien et al., 2007; McGee et al., 2013; Mulitza et al., 2008; Tjallingii et al., 2008). Furthermore the Early Holocene African Humid Period (5-11.7 ka), has been characterized with dust fluxes on the West African margin a factor of 2-5 lower than modern (McGee et al., 2013). Since the end of the African Humid Period, there is evidence for a further, gradual aridification of tropical Africa over the past 3000 years from the Mauritanian shelf dust record (Mulitza et al., 2010), continental precipitation records (Shanahan et al., 2015), reduced Niger river outflow (Weldeab et al., 2007) and dropping levels of Lake Chad (Armitage et al., 2015).

A general mechanism has been proposed for these variations, applicable to modern inter-annual variability as well, involving the meridional sea-surface temperature and/or pressure

gradients in the Atlantic and the position of the ITCZ over the African continent (e.g., (Evan et al., 2011; McGee et al., 2014; Rodríguez et al., 2015)). The present Andros Island records also offer an opportunity to test for this coherence of African dust and North Atlantic climate over the Little Ice Age. This period (roughly 1400 to 1800 AD) has been characterized by a modest cooling (~0.5°C) found most strikingly in the extratropical Northern Hemisphere continents (Mann et al., 2009), with relatively minor changes in interhemispheric surface temperature differences (Neukom et al., 2014). ~~In sea surface temperature compilations, gradual, rather than abrupt, cooling is found all throughout the pre-industrial Common Era (1 to 1800 AD) and the cooling from 800 to 1800 AD has been hypothesized to be a response to increased volcanic eruptions [*McGregor et al.*, 2015]. With the Andros cores, in addition to determining anthropogenic influences, we are poised to determine any roles African dust may have played in the Common Era cooling.~~

2. Approach

Our approach to reconstructing dust input here is geochemical. Thorium and helium share a unique property in that both elements have a dominant isotope (^{232}Th and ^{4}He) associated with aerosol mineral dust (Marcantonio et al., 2001; Marcantonio et al., 1998; McGee et al., 2016; Serno et al., 2014) and a minor isotope (^{230}Th and ^{3}He) that has a relatively constant source to the ocean surface (François et al., 2004; Marcantonio et al., 1996; McGee et al., 2010). In the Triple Goose Creek area of Andros Island (Fig. 1), we have cored within a fairly well-studied system of beach ridges, tidal channels, levee crests and mangrove ponds abutting inland algal marsh (Hardie, 1977; Shinn et al., 1969). The carbonate particles accumulating here are primarily derived from aragonite-producing marine algae and subsequently are washed inshore by tides. The levees are built up by overbank flooding and the low ponds are normally flooded twice daily

by tides with a range of roughly 40 cm (Hardie, 1977). The dust concentrations recorded in these accumulating sediments, therefore reflects dust deposited from the local atmosphere, dilution by the dominant carbonate sediments, and dust swept in by the tides from some portion of the shallow (~3 m water depth) Great Bahama Bank (Harris et al., 2015).

Our goal is to reconstruct atmospheric deposition, while the carbonate dilution and lateral components could vary in time due to factors unrelated to dust input such as variations in storm surges or evolution of tidal channel topography. Fortunately, any ^{232}Th and ^{4}He supplied to a core site laterally from the bank top should also be accompanied by a proportional amount of ^{230}Th and ^{3}He. Additionally, dilution will affect all these isotopes equally. Thus, the two ratios ^{232}Th/^{230}Th and ^{4}He/^{3}He provide proxies of atmospheric dust input that account for possible changes in lateral addition with time. The lateral addition of sediments is termed sediment focusing and the magnitude of sediment focusing can be expressed as a focusing factor (F): the ratio of the total accumulation rate of the isotope in question to its local production, or deposition, rate.

In the case of ^{230}Th, its production in water over the Bahamas Bank consists of two sources. The first is the strictly known production from the decay of dissolved ^{234}U in seawater. The second source is release of ^{230}Th from sediment pore waters derived from ^{234}U decay within the high U (~3 ppm, Fig. 2) aragonite sediments, which likely is a much larger source than the first but also is less quantifiable. Robinson et al. (2004) found evidence for this benthic source in that Bank-top water had much higher ^{230}Th content than at similar depths in the surrounding deep ocean. Using a box-model approach with reasonable assumptions, these authors determined the ^{230}Th supply from the sediments to the overlying water could easily be 30 times the in-situ production in the water column. There have been no independent determinations of this flux.

Without precisely knowing the benthic supply of ^{230}Th, we cannot make quantitative estimates of sedimentary fluxes as is done using ^{230}Th-normalization in deep ocean studies (François et al., 2004). Nonetheless, we do expect the benthic ^{230}Th flux to be constant over the timescales considered in this study. In other words, we ~~may~~ are not ~~be~~ able to determine an absolute focusing factor with ^{230}Th, but using it as a normalizing factor will account for relative changes in sediment focusing.

Once in the dissolved phase, due to its insoluble nature, ^{230}Th in the Bank-top waters is expected to be to be rapidly adsorbed (or scavenged) onto the abundant carbonate particles with a residence time of less than six months (Robinson et al., 2004). The same is true for any ^{232}Th that dissolves from dust, as it is estimated that 5-20% of aerosol Th dissolves in the upper water column (Hayes et al., 2013; Hsieh et al., 2011). Thus, Th delivery to the Andros Island cores consists of both structurally-intact Saharan dust, as well as scavenged Th on the carbonate mud.

As a noble gas, He in the sediments is fully contained in crystalline particles, in contrast to sedimentary Th, which has a fraction that dissolves and spends some time (months) in the water before being adsorbed and buried. In the helium system, at least during the Quaternary, ^{3}He is largely supplied to the Earth surface in the form of a relatively constant rain of interplanetary dust particles (IDP's). Its rain to Earth is calculated from its accumulation in deep sea sediments and ice cores, averaged over timescales ranging from 1 million years to the Holocene. These estimates range from 4 to 13 x 10^{-9} cc (pcc) ^{3}He at STP per m^2 per year, and most are consistent with 8 ± 3 pcc/m^2/yr (McGee and Mukhopadhyay, 2013). Thus, without a proper regional calibration, we can only be confident in the absolute ^{3}He input to our sites to within ~50%. Nonetheless, the constancy of IDP rain into Holocene ice cores (Brook et al., 2000; Winckler and Fischer, 2006) supports the view that the ^{3}He flux over the Bahamas has

remained relatively constant throughout the past two millennia. Furthermore, if we assume the 8 ± 3 pcc/m^2/yr $^3He_{ET}$ deposition rate is correct, this affords us an estimate of the focusing factor (F) in the sediments. Focusing factors are calculated as bulk averages between dated depth horizons, since it cannot generally be assumed that mass accumulation rates are constant between age controls. F equal to 1 implies no focusing, greater than 1 implies focusing and less than 1 implies sediment winnowing or that sediments have been removed.

3. Material and Methods

3.1 Coring and core descriptions

Fieldwork took place during March 2014 in the Triple Goose Creek area on northwest Andros Island in the Bahamas (Fig. 1). Sediment cores were collected using a 5 cm diameter, stainless steel and aluminum Livingstone piston corer from the University of Minnesota Limnological Research Center. Extruded cores were split open on plastic sheets in the field and sampled at cm-scale resolution. The corer was 1 m in length. Cores of 1.5-2 m were taken by re-entering the hole created by a first core. We use a composite depth scale for the core as a whole, assigning a depth to the top of the second core equal to the full length of the first core.

We chose three coring sites (Fig. 1): a mangrove pond in roughly 1 m of water (Core 2.4), a beach at the western edge of the tidal flat (Core 2.5) and a levee crest at the edge of a tidal channel (Core 2.7). We refer to these as the Pond, Beach and Levee Crest cores throughout the manuscript. All cores consisted of calcium carbonate mud (>94% by weight), bioturbated by gastropods and polychaete worms. The remaining, percent-level, components of the sediments were organic matter, largely in the form of decaying mangrove roots, and Saharan mineral dust. Due to the strong transport of the Gulf Stream in the Florida Straits, Andros Island is well-

isolated from continental input from Florida, making transported Saharan dust the only non-local source of sediments (Muhs et al., 2007).

All cores contained a stark transition with depth from white mud to gray mud (Fig. 2) which marks the transition from oxic to sulfidic conditions (Maloof et al., 2007). This redox transition is well marked by the enrichment of authigenic uranium (Fig. 2) as uranium becomes insoluble and precipitates in the sulfidic conditions. Reducing conditions at depth draw dissolved U from above down a gradient in dissolved U concentrations in porewaters (Klinkhammer and Palmer, 1991), as has been observed elsewhere in the Bahamas (Henderson et al., 1999b; Romaniello et al., 2013). The depth of this redox transition in the cores reflects the position of mean-tide level throughout the area (Maloof and Grotzinger, 2012). In the Beach core, the upper 8 cm consisted of broken shell hash, a product of the much stronger tidal energy available in this location. Below 8 cm, however, the core material was bioturbated mud, indicating that this location was once a pond environment in times of lower sea level (Maloof and Grotzinger, 2012). In the Levee Crest core, the upper 20 cm of white mud contains faint laminations, suggesting reduced bioturbation in this section.

3.2 Water sampling

Water samples were taken by filling acid-cleaned 5-L cubitainers. For "Pond" water and "Banktop" water samples (Fig. 1), 2 x 5-L samples were taken and when returned to the lab at MIT (~3 days later), 1 of each 5-L replicate was filtered at 0.45 µm. Subsequently all water samples, filtered or unfiltered, were acidified to ~pH 1.8 with ultra clean 6N hydrochloric acid for preservation. Aliquots of the seawater were taken prior to acidification for salinity analysis by conductivity at the Woods Hole Oceanographic Institution. The Pond and Banktop waters had salinities of 36.95 and 35.02, respectively, indicating evaporation in the pond water. Water

samples were collected on March 19, 2014 and the most recent rain event was March 13, 2014, according to the National Weather Service Weather Prediction Center (http://www.wpc.ncep.noaa.gov/)

3.3 Radiocarbon and excess lead-210 analysis

The benthic foraminifera *Peneroplis proteus* are abundant in the study area (Shinn et al., 1969), and were picked from sediments sieved at 355 µm, washed and sonicated in Milli-Q water, and dried at 80° C. Radiocarbon measurements were performed on 5 mg foraminifera samples at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory. Radiocarbon ages were converted to calendar ages using Marine13 (Reimer et al., 2013). Surface ocean ^{14}C reservoir ages have been determined for the region of interest using banded corals from the Florida Keys and Belize to be 390 ± 60 years for the period AD 1600-1900 (Druffel, 1982; Druffel and Linick, 1978; Druffel et al., 2008). The CALIB database (Reimer and Reimer, 2001) contains three other relevant reservoir age determinations in the vicinity: 546 ± 66 yrs from a coral in Goulding Cay, Bahamas (Lighty et al., 1982), and 360 ± 42 yrs and 456 ± 59 yrs from Bahamian gastropods (Broecker and Olson, 1961). Given the spread of these estimates, we apply the standard 400 yr correction with additional uncertainty, using a ΔR of 0 ± 70 yr.

Gamma spectroscopy was performed at MIT to analyze ^{210}Pb (46.5 keV) and ^{214}Pb (351.9 keV) using a Canberra Broad Energy Germanium Detector. Detector efficiency was determined using a pitchblende ore/silica mixture (US-DOE CRM 101) of known ^{238}U activity. ~~Four to six gram samples of freeze-dried sediments were sealed in plastic containers at least 3 weeks prior to analysis to allow secular equilibrium to be reached between ^{214}Pb and its grandparent ^{226}Ra. Sealing prevents loss of the gaseous intermediary ^{222}Rn. Therefore~~ Using 4-6 g samples that

were sealed in plastic containers for 3 weeks, ^{214}Pb activity was used as a proxy for the ^{210}Pb supported by decay within the sediments ($^{210}Pb_{supp}$). The difference between measured ^{210}Pb ($^{210}Pb_{meas}$) and $^{210}Pb_{supp}$ gives the excess ^{210}Pb ($^{210}Pb_{xs}$), unsupported by production within the sediments. In Bahama Bank sediments, as elsewhere, $^{210}Pb_{xs}$ is predominantly sourced from the atmosphere (Henderson et al., 1999a), where it is ultimately a product of radon emanation from the continents. Analysis times ranged from 5 to 14 days, depending on sample activity.

3.4 Thorium and helium derived parameters

Uranium, thorium and helium isotope analyses were performed by mass spectrometry after sample preparation following modifications of published methods (Muhkopadhyay… McGee et al., 2013, Hayes et al. 2015). Full procedures are included in the Supplemental Material. We make small corrections to the ^{230}Th data to account for detrital input and ingrowth and decay within the sediments to calculate the initial, excess $^{230}Th_{xs,o,}$ which is derived from decay of U in seawater. Similary, ~~and~~the He isotopic data is corrected to isolate the fraction of total ^{3}He from IDP's ($^{3}He_{ET}$) and the fraction of total ^{4}He from dust ($^{4}He_{TERR}$). Full descriptions are presented in the Supplement Material. .

The ^{232}Th concentration of fine-grained (<5 μm) Saharan dust that makes it to Caribbean is 13.7 ± 1.2 μg/g (Muhs et al., 2007). In this study, we undertook new measurements of ^{4}He in Northwest Providence Channel sediments whose dust content has been determined using ^{232}Th (Williams et al., submitted), core OCE205-2 103GGC (26°04' N, 78°03' W; 965 m water depth) on the south flank of the Little Bahama Bank (see Fig.1). From measured $^{4}He_{Terr}/^{232}Th$ ratios of 105, 96 and 64 mcc STP/g in samples of age 2.7, 3.3 and 4.1 ka, respectively, we infer that the fine grained Saharan dust reaching Andros Island has a ^{4}He concentration of 1212 ± 294 ncc/g. These ^{4}He concentrations and $^{4}He/^{232}Th$ ratios are very similar to values found for the <5 μm

fraction of dust source area samples from Australia, East Asia and South America (800-2000 ncc/g ^{4}He and ~50-250 mcc/g ^{4}He/^{232}Th, respectively) (McGee et al., 2016). This ^{4}He concentration for dust deposited in the Bahamas is much lower than concentrations of 5626 ± 3205 ncc/g found on the African margin in Cape Blanc sediments (Mukhopadhyay and Kreycik, 2008), consistent with the grain-size effect on ^{4}He concentration found by McGee et al. (2016). Sediments closer to the source in Cape Blanc have larger grains (e.g.,(McGee et al., 2013; Tjallingii et al., 2008)) and more ^{4}He than in the Bahamas.

4. Results

4.1 Core chronologies: ^{14}C

The Pond and Beach cores demonstrate stratigraphic order in their foraminifera ages, based on ^{14}C ages, (Fig. 3), but their core tops appear to be significantly older than modern.

For the Pond core, the sedimentation rate implied by the dates at 7 cm and 32.5 cm is 0.235 ± 0.037 mm/yr. Applying this constant sedimentation rate between 7 cm and the core top (depth = 0 cm) indicates a surface age of AD 1787 ± 82. Therefore roughly 225 years of sedimentation (~5.5 cm at the assumed rate) may have been lost due to disturbance/compaction during core collection, or due to natural processes associated with water movements in the pond.

The sedimentation rate in the Beach core between 9 cm and 19 cm is 0.260 ± 0.098 mm/yr, similar to the Pond core. This rate implies a core top age of AD 1325 ± 157, indicating that a much larger amount of sediment has been eroded away here by wave and storm action (as evidenced by the abrupt upward transition from mud to coquina) associated with the development of the beach environment as local sea level rises.

The fact that most of the foraminifera ages in both the Pond and Beach cores are older than AD 700 is surprising in itself, in consideration of previous work. Maloof and Grotzinger

(2012) found that *Peneroplis* specimens within the basal unit of 5 cores throughout the Triple Goose Creek area dated to AD 700-1005, averaging AD 903. Since the Bahama Banks were flooded by rising Holocene sea-level around 4500 ± 1000 years ago (or ~2500 ± 1000 BC) (Rasmussen et al., 1990; Slowey and Henderson, 2011), there may have been a several thousand year energetic period between initial inundation of the region and the beginning of quiescent pond-like sedimentation. Based on the bottom date of the Pond core, however, pond sedimentation may have begun as early as 1702 BC in that location. In the Beach core, if the sedimentation rate between 19 and 40.5 cm depth (1.27 mm/yr) is extended to the deepest of retrieved sediment (170 cm), this location may have had pond sedimentation as early as 600 BC.

Virtually all of the cores described by *Maloof and Grotzinger* [2012] were collected further inland than the Pond and Beach locations. Speculatively, the older bottom dates reported here suggest that these locations were among the first to develop undisturbed sedimentation, perhaps due to details of the topography of the underlying cemented Pleistocene reef bedrock. In the age models for these cores, we linearly interpolate age, assuming constant sedimentation rates between dated depth horizons (Fig. 3), with the understanding that continuous sedimentation may be a tenuous assumption in sections much older than AD 700. Further stratigraphic and sedimentological work is needed to clearly define the nature of sedimentation prior to this time.

Foraminifera dates for the Levee Crest were all from the lower 60 cm of this site's 180 cm core. Only in this lower section was foraminifera abundance adequate. These three dates are not in stratigraphic order (Figs. 2, 3, Table 1), but they are consistent with each other within dating uncertainties. The apparent age-inversion may be due to bioturbational mixing. Taking the average depth and average age of the dated foraminifera in this core, and assuming a modern age

for the core top (AD 2014 at depth 0 cm) yields a sedimentation rate of 1.83 ± 0.35 mm/yr. This sedimentation rate is consistent with the radiocarbon dating of other cores from this area by *Maloof and Grotzinger* [2012]. Noting the old core-top ages found in the other two cores, however, this assumption of modern core top is highly uncertain. We therefore analyzed the sediment profiles for unsupported ^{210}Pb which, as described next, corroborates the radiocarbon-based sedimentation rate, supports a modern core top age and provides an estimate of the depth-scale of bioturbation for the Levee Crest core.

4.2 Core chronologies: ^{210}Pb

The shape of the $^{210}Pb_{xs}$ profile in the sediment is affected by biological mixing, age decay during burial and, possibly, changes in sedimentation rate with depth. Ignoring changes in sediment porosity and assuming a constant sedimentation rate (w), $^{210}Pb_{xs}$ activity (A) can be modeled as a function at steady-state of depth (z), diffusional bioturbation rate (D), and radioactive decay constant (λ) (Eq. 3).

$$D\frac{d^2A}{dz^2} - w\frac{dA}{dz} - \lambda A = 0 \qquad \text{Eq. 3}$$

The logarithmic decay from the coretop activity (A_o) to A at depth z can be solved for (Eq. 4). Using a half-life of 22.2 yrs (Browne, 2003), λ for ^{210}Pb is 0.031223 yr^{-1}.

$$ln\left({}^{A}/_{A_o}\right) = \frac{w-\sqrt{w^2+4D\lambda}}{2D}z \qquad \text{Eq. 4}$$

$$ln\left({}^{A}/_{A_o}\right) = \frac{-\lambda}{w}z \qquad \text{Eq. 5}$$

If there were no bioturbational mixing (D = 0), the ^{210}Pb profile of the Levee Crest core (Fig. 4) would imply a sedimentation rate of 2.21 mm/yr using Eq. 5. Assuming, however, that

the ^{14}C-based sedimentation rate (w = 1.83 mm/yr) is correct, by solving Eq. 4, this core has a diffusion rate of 27.3 mm^2/yr. This is a lower than but reasonable diffusion rate compared to that found in 5 m water depth on the Little Bahama Bank of 106 ± 16 mm^2/yr by Henderson et al. (1999a). Additionally, we expected reduced bioturbation in the upper section of this core from its faint laminations (Fig. 2). Thus the ^{210}Pb results corroborate the ^{14}C-derived sedimentation rate of 1.83 mm/yr.

An annual ^{210}Pb deposition of 115 Bq/m^2 has been recorded at Bermuda (Turekian et al., 1983) and this result is consistent with the predicted deposition of 100-150 $Bq/m^2/yr$ to the Bahamas, based on a model constrained by global measurements (Preiss and Genthon, 1997). Additionally, we can infer the local ^{210}Pb deposition rate by integrating its depth profile (Fig. 4), assuming a dry bulk sediment density (ρ) of 1.14 g/cm^3, as measured on the Little Bahama Bank (Henderson et al., 1999a). The total inventory of $^{210}Pb_{xs}$ in the Levee Crest core is 3260 ± 70 Bq/m^2, meaning that the sediment column (per m^2) is losing 3260 atoms of $^{210}Pb_{xs}$ per second due to decay. To support a steady-state sediment inventory, this loss to decay must be balanced by atmospheric deposition, which converted back into activity units is 102 ± 2 $Bq/m^2/yr$. If we assume a modern (AD 2014) age for the coretop, using the local atmospheric $^{210}Pb_{xs}$ deposition rate, the coretop $^{210}Pb_{xs}$ activity of 48.5 ± 8.1 mBq/g (Fig. 4) can be used to derive a sediment mass accumulation rate (MAR) (MAR = atmospheric $^{210}Pb_{xs}$ deposition ÷ coretop $^{210}Pb_{xs}$ activity). This calculation gives a sediment MAR of 2100 ± 350 $g/m^2/yr$.

The sediment MAR based on the ^{14}C-derived sedimentation rate (1.83 ± 0.35 mm/yr), multiplied by the assumed ρ = 1.14 g/cm^3 gives 2085 ± 399 $g/m^2/yr$. That these two independent MAR estimates (2100 and 2085 $g/m^2/yr$) agree well within uncertainties is a good indication for a modern core top age for the Levee Crest core.

While the $^{210}Pb_{xs}$ profile of the Levee Crest core came from a faintly laminated section, we use the depth of $^{210}Pb_{xs}$ penetration to constrain a lower limit on the sediment mixed layer to 10-12 cm. Using the 1.83 mm/yr sedimentation rate, this mixing depth translates to a bioturbational smoothing in temporal space of about 60 years for the Levee Crest core. Sediments from the Pond and Beach locations are old enough that all of their $^{210}Pb_{xs}$ has decayed away (Fig. 4). Nonetheless, assuming a similar depth of bioturbation, the lower sedimentation rates from those cores (~0.25 mm/yr) imply longer temporal smoothing (~400 yrs). These smoothing time-scales should be considered lower limits.

4.3 Dust proxies

The He and Th isotope concentration data for the three Andros cores are presented as a function of calendar year in Fig. 5. All four isotopes show a decline in concentrations of roughly a factor of 5 between 500 BC and AD 500, followed by relatively uniform values between AD 500 and 2014. The Beach core shows some spikes in $^{3}He_{ET}$ at AD 200 and 500 that may represent the sampling of rare high-^{3}He IDPs in the relatively small 1 gram samples (Farley et al., 1997; Mukhopadhyay et al., 2001; Patterson and Farley, 1998).

On the basis of the dust indicators ^{232}Th and ~~$^{4}He_{ET}$~~ $^{4}He_{TERR}$ alone, one might have interpreted these records to reflect much higher dust input to the Bahamas prior to AD 500. Because our constant-flux proxies $^{230}Th_{xs,o}$ and $^{3}He_{ET}$ change in concert with the dust indicators, however, it appears that these sites received a relatively stable input of dust throughout. The increased concentrations~~, around the beginning of~~ prior to the Common Era~~,~~ likely reflect less total carbonate accumulation (thus reducing the dilution of the dust). This change clearly coincided with a change in sediment focusing (Fig. 6). In the Pond and Beach cores, the post-AD 500 focusing factors are within 10% of 1.0, while prior to this F = 11 ± 9 for the beach core and

3.3 ± 1.3 in the pond core, perhaps indicating that these were higher energy sedimentation environments before pond-like sedimentation set in. On the Levee Crest, F = 7.5 ± 3.0, and the stability in He and Th concentrations in this core suggests that focusing did not significantly change at this site for the past millennium.

We present the time series of the dust flux proxy ratios $^{232}Th/^{230}Th_{xs,o}$ and $^{4}He_{TERR}/^{3}He_{ET}$ in Fig. ~~6~~7. These ratios are linearly related to $^{230}Th_{xs,o}$-normalized ^{232}Th rain rates and $^{3}He_{ET}$-normalized $^{4}He_{TERR}$ rain rates, respectively, on the assumption of a constant supply of ^{230}Th or ^{3}He. The Th ratios have quite a limited range with a relative standard deviation (RSD) among the data from all cores of 8%. The He ratios have a RSD of 28% in all the data, typical of this type of analysis (Mukhopadhyay et al., 2001), and this lowers to 19% when the five ratios furthest from the mean are excluded. A similar average (720,000) and RSD (30%) of $^{4}He_{TERR}/^{3}He_{ET}$ ratios was found in another core collected in the same area in 2003 that had a bottom age of 1000 AD (Bhattacharya, 2012).

~~Before attempting to convert these ratios into dust flux,~~Let us ~~we~~ consider what dust flux is implied simply from the MAR of ^{232}Th in the Levee Crest core (since this core overlaps with the instrumental period). We focus on ^{232}Th accumulation because its concentration in fine-grained (< 5 μm) Saharan dust (13.7 ± 1.2 μg/g, (Muhs et al., 2007)) is known more precisely than $^{4}He_{TERR}$ (1212 ± 294 ncc/g, reported here). Using the average values in the upper 7.5 cm of the core (nominally representing AD 2014-1973), and a dry bulk density of 1.14 g/cm^3, the ^{232}Th MAR is 456 ± 40 μg/m^2/yr, translating to a Saharan dust MAR of 33 ± 4 g/m^2/yr. Recall that this estimate includes both direct atmospheric deposition as well as sediment focusing. ~~To account for focusing, we derive the focusing factor by comparing the $^{3}He_{ET}$ MAR in the same section (68 ± 19 pcc/m^2/yr) to the Quaternary average deposition of 8 ± 3 pcc/m^2/yr.The~~

~~$^{3}He_{ET}$~~Accounting for the focusing factor in this core (7.5 ± 3.0), ~~This calculation yields an apparent focusing factor of 8.5 ± 4.0 and applying this to~~ the estimated Saharan dust deposition is~~3.9 ± 1.3~~ 4.4 ± 1.8 g/m^2/yr.

Available observations from rain collections in nearby South Florida (1982-1984 and 1994-1996) (Prospero et al., 2010; Prospero et al., 1987) suggest an average dust deposition in this region of 1.3 g/m^2/yr, or roughly 3 times smaller than what was calculated in our core above. The most uncertain figure in our dust deposition calculation is the assumed $^{3}He_{ET}$ deposition to calculate the focusing factor (8 ± 3 pcc/m2/yr). Turning the problem around, the $^{3}He_{ET}$ deposition to our site would have to be about 3 times lower than the global average ~~, or more precisely 2.1 ± 0.7 pcc/m^2/yr,~~ in order to make dust deposition in our core exactly equal the modern rate of 1.3 g/m^2/yr. Lower $^{3}He_{ET}$ inputs at this site relative to the late Quaternary global mean may reflect temporal or spatial variability in IDP fluxes. While this is speculative, ~~T~~there is a possible justification of lower than global average $^{3}He_{ET}$ fluxes here in the subtropics, as other aerosol-bound isotopes of atmospheric origin, such as ^{10}Be or ^{90}Sr, are known to be transported by atmospheric circulation to be preferentially rained out in the mid-latitude storm tracks (Field et al., 2006; Lal and Jull, 2005).

Another possible explanation for lower-than-expected $^{3}He_{ET}$ fluxes in our core is fractionation of the two helium isotopes during sediment transport. IDP~~'~~s, the carrier phase of $^{3}He_{ET}$, are on average slightly larger than the Saharan dust grains, carrying the $^{4}He_{TERR}$ (5-20 μm for IDP's (McGee and Mukhopadhyay, 2013) versus 2.5-5.0 μm for far-field Saharan dust (Li-Jones and Prospero, 1998)). However, the two helium isotopes are accumulating at ~~the~~ nearly the same ratio when being gently washed into the quiescent ponds as when more energetically being flooded above channel banks onto the Levee Crest (Fig. 7). The same is true during

periods of low or high sediment focusing. We consider this evidence against fractionation by sediment transport and therefore our preferred explanation is regionally lower $^{3}He_{ET}$ fluxes here in the subtropics due to atmospheric circulation.

Finally, ~~I~~in analogy to inferring the local input of $^{3}He_{ET}$, we can use the ^{232}Th and ^{230}Th accumulation rates, assuming 1.3 g/m^2/yr dust input, to make an estimate of the ^{230}Th production rate on this region of the Bank, which could not be quantified *a priori* (Sec. 2). This ^{230}Th production rate works out to be 1.4 ng/m^2/yr, roughly 10 times the banktop benthic flux of ^{230}Th that Robinson et al. (2004) estimated from an analysis near Exuma Sound in the eastern Bahamas (see Fig. 1). For reference, the in-situ production of ^{230}Th in 5 m of seawater is quite small, about 3 pg/m^2/yr. It could be that on the northwestern Great Bahama Bank, there is more effective lateral transport of the benthic-sourced ^{230}Th into the Triple Goose Creek area than in the Exuma Cays. The difference may relate to contrasting groundwater hydrology between the two locations, as North Andros has significantly higher hydraulic conductivity than the Exumas on the southeastern portion of the Bank (Whitaker and Smart, 1997).

While more work needs to be done to define the absolute values of ^{230}Th and ^{3}He supply to the Bahama Bank, there is an undeniable stability in the dust proxy ratios in our cores. This stability argues against large changes in either IDP flux, ^{230}Th supply, or Saharan dust flux to our site over the covered time periods. ~~We also expect the ratios to vary linearly with dust flux. Therefore in Fig. 6, we made a rough scaling of the ratios into dust flux (Fig. 6) by centering the mean Th and He ratios to 1.3 g/m^2/yr dust flux. With this scaling, the dust proxy data suggest dust deposition in the Bahamas has varied only between 1.1 and 1.4 g/m^2/yr over the length of the records.~~

4.4 Implications for historical changes in North Atlantic dust deposition

Of course, bioturbation has smoothed these dust flux records, similar to a running average filter, contributing in part to the apparent stability. This is particularly true for the Pond and Beach cores (at least 400 yr filtering, based on the ^{210}Pb results discussed Section 4.2) in which any century-scale changes will have been muted. However, the Pond core dust proxy ratios do agree where they overlap (AD 1100-1800) with those of the Levee Crest core, which is much less affected by bioturbation (60 yr filtering). Even given the age model uncertainty prior to AD 700, the three Andros records together indicate no trend in dust flux during the millennium between AD 800 and 1800, which overlaps ~~long-term Common Era sea-surface cooling [*McGregor et al.*, 2015],~~ the Medieval Climate Anomaly (MCA) and the Little Ice Age (LIA) (Mann et al., 2009). We emphasize that we cannot assign absolute dust fluxes in our results, but the relative trends are robust.

During deglacial cooling events (Heinrich Stadial 1 and the Younger Dryas), the Early Holocene African Humid Period, and recent decadal scale variability, African dust appears to have been part of a coupled system involving North African aridity, trade wind intensity and subtropical North Atlantic cooling (Adkins et al., 2006; Bradtmiller et al., 2016; Evan et al., 2011; Evan and Mukhopadhyay, 2010; McGee et al., 2013; Williams et al., submitted). We hypothesize that the lack of apparent changes in Bahamas dust fluxes in response to centennial-scale climate variability between 800 and 1800 AD reflects the fact ~~that the observed Common Era cooling was global [*McGregor et al.*, 2015] rather than focused in the Atlantic. Similarly, the~~ that the Little Ice Age was marked by relatively warm conditions in the high-latitude North Atlantic (as opposed to the continents) [*Mann et al.*, 2009], limiting the trade wind response. It may be that increased volcanic aerosol loads during the LIA drove a climate response that did not significantly amplify the expected internal variability of the North Atlantic. ~~In other words,~~

~~the modern seasonal movement of the ITCZ that influences dust emission and transport to Bahamas was likely relatively unchanged over the past millennium in contrast to events of the Early Holocene or deglaciation.~~

On the issue of the regional impact of anthropogenic dust from the Sahel, we rely solely on the Levee Crest core. As described above, the radiocarbon and lead-210 dating in this core give us high confidence that the past 200 years are recorded here, albeit filtered by a 60 year running average. The dust proxy data indicate no variability over this time period outside the ~10% variations seen in the records as a whole. The bioturbation of this core would have muted the magnitude of decadal events, but any pre-industrial to modern trend in dust input would not be masked by a running average~~, unless the change was highly exponential and occurred more recently than the past 60 years~~. For instance, applying a 60 yr running average to the West African dust flux record from Mulitza et al. (2010) only reduces the apparent increase in dust flux between 1800 to 2000 AD from a factor of 4 to a factor of 2.5, also not supported by our data.

Furthermore, the dissolved $^{232}Th/^{230}Th$ ratio in the Banktop and Pond waters (average shown as a star in Fig. 7) nearly ~~exactly~~ matches the average value in the cores ~~of about 14,000~~. Since the dissolved Th ratio represents an average of dust deposition in accordance with the residence time of Th in the water (less than a year), this is another strong indication that dust flux to the Bahamas today is quite similar (within 10%) to that of 200 years ago. We thus conclude that the onset of commercial agriculture in the Sahel did not greatly modify the magnitude or transport of North African dust ~~over~~ to the western North Atlantic. This conclusion is consistent with the understanding that Saharan sources are the dominant contributors to the North African dust plume, with relatively minor inputs from the Sahel (Engelstaedter et al., 2006; Ginoux et al.,

2012) and will likely be important in the modeling of aerosol fields from the pre-industrial to modern day.

5. Conclusions and Future Work

We have described Saharan dust proxies in a set of tidal flat cores from Andros Island, Bahamas ~~that provide a continuous record of Saharan dust deposition over at least the past millennium~~. The chronological and geochemical work undertaken has led to several interesting avenues for future work. These are (1) to investigate areas of the Triple Goose Creek region in Andros Island that have apparent ages at 1.3 m depth of 3700 years, despite the fact the Pleistocene cement basement here was only flooded hundreds to thousands of years before this, (2) to determine if recent extraterrestrial ^{3}He deposition to the Bahamas could be a factor of 3 lower than has been found in the Holocene sections of the Greenland and Antarctic ice cores, and (3) to determine the magnitude and spatial distribution of the benthic flux of ^{230}Th from banktop sediments into banktop waters. These puzzles notwithstanding, the dust proxy data derived from the cores can be used to confidently determine that transport of Saharan dust ~~across~~ to the western tropical North Atlantic did not undergo large, factor of 4 magnitude changes from the pre-industrial period (AD 1800) to today. Furthermore, no long term trends in dust input to the Bahamas are apparent over the past millennium, indicating that transport of North African dust to the western Atlantic ~~emissions did~~ may not have participated in climate feedbacks associated with ITCZ movement over this time period. ~~global sea-surface cooling of the Common Era~~.

Acknowledgements

We are grateful for the students of the 2013-2014 Princeton Sedimentology course for coring work. Sample preparation by Clio Macrakis and Ashling Neary, during their High School and undergraduate research experiences, respectively, is greatly appreciated. We are thankful for logistical support from the staff of Small Hope Bay Lodge. CTH was supported by the W.O.

Crosby Postdoctoral Fellowship at MIT. Data from this study can be accessed in the electronic supplement. We thank Franco Marcantonio for reviewing an earlier version of this manuscript.

Figure and Table Captions

Figure 1. Study site locations based on maps extracted from Google Earth Pro. The overview map on the left is based on a Landsat image, using data from SIO, NOAA, U.S. Navy, NGA and GEBCO. The detailed map on the right, which is located roughly in the black box indicated in the overview map, is from DigitalGlobe © 2014. Coordinates on the detailed map are in kilometers and refer to UTM grid 17R. The black stars indicate sediment coring locations and the small white stars indicate water sampling locations.

Figure 2. Visual description of the Andros Island sediment cores. Overlaid on each core is the depth profile of sedimentary uranium concentrations. To the right of each core is indicated the calibrated (Marine13) radiocarbon dates in calendar years positioned at the depth of foraminifera samples used for dating. A 400 ± 70 year reservoir correction has been applied to the measured ages.

Figure 3. Calibrated (Marine13) radiocarbon dates based on the benthic foraminifera *Peneroplis proteus* found in each of the three Andros Island cores. A 400 ± 70 year reservoir correction has been applied to the measured ages. For the Pond and Beach core age models, linear interpolation between dated depth horizons is used. For the Levee Crest core, the age model, shown in gray, is a line drawn between the modern core top (depth 0 cm = year AD 2014) and the average depth and average age of the three measured samples.

Figure 4. Excess lead-210 depth profiles in the Andros Island cores. Data from the Levee Crest core is shown in detail in (A) and fit with models of exponential decay following sedimentation rates (w) and varying degrees of sediment diffusivity due to bioturbation (D). In (B), the Pond and Beach cores showed negligible excess Pb-210 in the upper 30 cm, indicating core tops older than at least 200 years. The upper 8 cm of the Beach core consisted of shell hash and therefore samples for Pb-210 were taken only below this section in the bioturbated mud (Fig. 2).

Figure 5. Sedimentary concentrations (from the top panel to the bottom) of terrigenous helium-4, extraterrestrial helium-3, thorium-232, and initial, excess thorium-230. ~~On the top panel, the black symbols represent the locations (depth/age) of age control in the three cores with the symbols corresponding to the same key shown in the bottom panel.~~

Figure 6. Focusing factors for the Andros Island cores. These factors are calculated using the inventory of 3HeET within dated depth horizons (indicated by the x-axis errors bars) compared to the global 3HeET deposition of 8 ± 3 pcc/m^2/yr. The uncertainty in the focusing factor is propagated from the uncertainties in the radiocarbon dates, the helium inventories, and the assumed 3HeET deposition rate.

Figure ~~6~~7. Dust flux proxy data based on ~~T~~thorium and helium isotope ratio data (units are atom ratios divided by 10^4 and 10^6, respectively) for the Andros Island cores. In the top panel, ~~also~~

shown, as a star, is the average $^{232}Th/^{230}Th$ ratio of dissolved Th from two Pond and Banktop water samples. To the far right and far left, we have made a rough linear conversion of the thorium and helium isotope ratios, respectively, to dust flux by setting the average isotope ratio to the average measured modern dust deposition for the region of 1.3 g/m^2/yr [*Prospero et al.* 1987; 2010]. The approximate range of temporal smoothing due to bioturbation in the cores is indicated in the bottom panel, 60 years for the Levee Crest Core and 400 years for the Pond and Beach cores.

Table 1. Radiocarbon data and calibration using Marine13, including a reservoir age correction of 400 ± 70 yrs, for the Andros Island cores. Radiocarbon age in years before 1950 is quoted using the Libby half-life of 5568 years following the conventions of Stuiver and Polach (1977). $\delta^{13}C$ values were assumed to be -1 ± 2 ‰ (VPDB) and this error is accounted for in the reported ^{14}C age.

For Supplemental Material:

3.4 Thorium and uranium analysis

Bulk sediment ^{232}Th, ^{230}Th and ^{238}U concentrations were determined by isotope dilution inductively-coupled plasma mass spectrometry (ICP-MS) following total dissolution of the sediment and anion exchange chromatography (Bio-Rad AG1-X8 resin). Sediment dissolution was facilitated using 16 M nitric acid (trace metal grade), 12 M hydrochloric acid (trace metal grade), 29 M hydrofluoric acid (trace metal grade), and 9.8 M hydrogen peroxide (reagent grade). These concentrations and purity apply to all mentions of the reagents below unless otherwise noted.

Fifty mg of freeze-dried sediments were weighed and spiked with 250 pg ^{229}Th and 1.1 ng ^{236}U into 22 mL PFA beakers containing a few mL high-purity (Milli-Q) water. Nitric acid was added in 100 µL increments until the sample was completely decarbonated. Next, after adding an additional 1 mL of HNO_3, the samples were heated to near-dryness at 150° C. Samples were cooled and taken up in 1 mL of HNO_3 plus 1 mL HCl. The Aqua Regia mixture was allowed to react for 2 hours and then heated down to near-dryness at 100° C. Cooled samples were taken up in 0.5 mL HNO_3 and 30 drops of H_2O_2 were added. About 30 minutes were

allowed for the sample to react and samples were heated to near-dryness at 100° C. Samples were then taken up in 0.5 mL HNO_3, heated to 150° C, and 1 mL of HF was added. Samples were then allow to dry completely at 150° C. At this point, if the sample did not fully dissolve in 0.5 mL HNO_3, either the H_2O_2 or HF addition steps (or both) were repeated. The dissolved samples were transferred to 15 mL centrifuge tubes with Milli-Q water and about 5 mg of Fe added as dissolved $FeCl_3$. pH was raised to >8 with 15-20 drops of 13 M reagent grade ammonium hydroxide. The precipitated iron (oxy)hydroxides were centrifuged at 3000 RPM for 5 minutes, decanted, washed with Milli-Q water plus a drop of ammonium hydroxide, centrifuged and decanted again. The remaining iron pellet was dissolved with nitric acid and transferred to the Milli-Q-water-rinsed PFA beakers used for sample dissolution. Samples were dried and taken up in 8 M HNO_3 for purification and separation of U and Th on 2 mL AG1-X8 (100-200 mesh) resin columns.

Anion-exchange columns were rinsed with 5 mL HCl, 7 mL Milli-Q water, and then primed with 2 x 2 mL 8 M HNO_3. Samples were added as 2 mL 8 M HNO_3, collecting the eluent as waste. Two 1 mL 8 M HNO_3 rinses of the beakers were added to the column and then 2 x 1 mL 8 M HNO_3 was added directly to the column. Thorium was then eluted into PFA beakers using 0.2 + 3 x 4 mL HCl. Uranium was subsequently eluted into separate PFA beakers using 3 x 4 mL 0.12 M HCl. All fractions were then dried at 150° C and taken up in 1 mL 0.48 M HNO_3 for ICP-MS analysis.

ICP-MS measurements were made on two multi-collector instruments at MIT: a Micromass IsoProbe and a Nu Plasma II. Th-230 was analyzed on an ion counter, while all other isotopes were measured on Faraday cups. Methods were designed to derive $^{232}Th/^{230}Th/^{229}Th$ and $^{238}U/^{236}U$ ratios at better than 1% precision. Instrumental mass-bias, ion-counter yields and ^{230}Th

tailing were corrected for using an in-house Th standard (MITh-1) and a certified U standard (CRM-112a). Five sediment samples were analyzed in triplicate and external reproducibility for ^{232}Th, ^{230}Th and ^{238}U concentrations were 2.1%, 4.6% and 4.4%, respectively.

Water samples were analyzed for ^{232}Th and ^{230}Th by published methods (Hayes et al., 2015) at MIT. U-238 concentrations in the water were also analyzed on 5 mL aliquots, using methods equivalent to those described above for dissolved sediments.

3.5 Helium analysis

One gram of freeze-dried sediment was transferred to 50 mL centrifuge tubes, and up to 50 ml of 1.7 M reagent grade acetic acid was added incrementally. To aid the dissolution of calcium carbonate, the samples were sonicated for 15-20 minutes or agitated on a mixing table for 1-2 hours. Samples were then centrifuged for 30 minutes. After the acetic acid was poured off, the sample residues were washed with 50 mL Milli-Q water and centrifuged again for 30 minutes. The overlaying water was then pipetted off until the residue remained at the bottom of the tube under approximately 1-2 mL of water. To ensure quantitative transfer of the entire sample, the wet residue was frozen in the centrifuge tube overnight. The frozen sample could then be extracted with a spatula and transferred to pre-weighed tin foil cups. The foil cups were dried in an oven at 80^{o} C. The remaining dry residue was weighed to determine a non-carbonate fraction of the sediments, which ranged from 1 to 6% by weight. The tin foil cups were then carefully rolled into sealed, solid balls and loaded into a high-vacuum furnace for analysis by Noble Gas mass spectrometry at Harvard University and the University of California, Davis. Three samples were run in triplicate and two others in duplicate. External reproducibility, based on these replicate analyses, for ^{4}He and ^{3}He measurements was 14% and 22%, respectively.

Th and He isotope corrections:

In the case of ^{230}Th, we wish to isolate the fraction of ^{230}Th in the sediments produced by decay of uranium in the water (or released from porewater in the Bahamas Bank), $^{230}Th_{xs}$. The ^{230}Th added from dust is subtracted using measured ^{232}Th and a $^{230}Th/^{232}Th$ atom ratio in dust of 4 ppm (Roy-Barman et al., 2002). Ingrown ^{230}Th from the decay of authigenic U in the sediments is subtracted based on the measured U concentration (Fig. 2) corrected for the U in dust using a U/Th activity ratio of 0.7 and the age-model derived sediment age (see Henderson and Anderson (2003)). These two corrections account for 4-16% of the measured ^{230}Th in the Andros cores. Finally, $^{230}Th_{xs}$ concentrations are corrected for ^{230}Th decay (half-life 75,584 yrs, Cheng et al. (2013)) since the time of deposition to derive the initial, excess or $^{230}Th_{xs,0}$. For ^{232}Th, we assume all of it is derived from Saharan dust.

In the case of He, both ^{3}He and ^{4}He have terrigenous and extraterrestrial sources. We use an end-member $^{3}He/^{4}He$ atom ratio for IDPs of 2.4 x 10^{-4} (Nier and Schlutter, 1990) and a terrigenous $^{3}He/^{4}He$ ratio of 3 x 10^{-8} ((McGee and Mukhopadhyay, 2013) and references therein). The ^{4}He derived from dust ($^{4}He_{TERR}$) and the extraterrestrial ^{3}He ($^{3}He_{ET}$), derived from IDPs, can then be derived from the measured (meas) ratios (Eqs. 1, 2). In the Andros cores, >99% of the ^{4}He is terrigenous and >97% of the ^{3}He is extraterrestrial.

$$^{4}He_{TERR} = \left[\frac{(^{3}He/^{4}He)_{meas} - (^{3}He/^{4}He)_{IDP}}{(^{3}He/^{4}He)_{TERR} - (^{3}He/^{4}He)_{IDP}}\right] \times {}^{4}He_{meas} \quad \text{Eq. 1}$$

$$^{3}He_{ET} = \left[\frac{1 - \frac{(^{3}He/^{4}He)_{TERR}}{(^{3}He/^{4}He)_{meas}}}{1 - \frac{(^{3}He/^{4}He)_{TERR}}{(^{3}He/^{4}He)_{IDP}}}\right] \times {}^{3}He_{meas} \quad \text{Eq. 2}$$

Possibly important, but not including due to space:

Along these same lines, given the opposite grain-size effects for ^{232}Th and ^{4}He (McGee et al., 2016), one might expect there to be some decoupling between these two elements during the sediment transport processes on Andros Island. The average $^{232}Th/^{4}He_{TERR}$ ratios in the Andros cores (0.012 ± 0.002 g/mcc, 1 sigma), however, are identical to the average ratio in the offshore (965 m depth) samples we analyzed from the NW Providence channel (0.012 ± 0.004 g/mcc). This is encouraging in that the dust accumulating in the more complex sedimentary setting of Andros Island has the same composition of the dust settling in a more pelagic environment.

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