A MULTI-PROXY APPROACH TO DETERMINE PALEOECOLOGICAL CHANGE OF MANGROVES, DURING THE HOLOCENE, IN BELIZE, CENTRAL AMERICA

By

Natalie Marie Monacci

RECOMMENDED:

Advisory Committee Chair

Head, Program in Marine Science and Limnology

APPROVED:

Dean, School of Fisheries and Ocean Sciences

Dean of the Graduate School

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A MULTI-PROXY APPROACH TO DETERMINE PALEOECOLOGICAL CHANGE OF MANGROVES, DURING THE HOLOCENE, IN BELIZE, CENTRAL AMERICA

A

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Natalie Marie Monacci, B.S.

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Abstract

This thesis presents multiple analyses of mangrove peat cores from Spanish Lookout Cay (BT-79) and from along the banks of the Sibun River (SR-63), Belize to examine ecosystem responses to environmental change during the Holocene. Radiocarbon measurements showed these sites were colonized by mangroves ~8,000 cal. yrs BP and have decreased sedimentation rates from ~6,000 to ~1,000 cal. yrs BP, which is attributed to a decrease in sea-level inundation. Core SR-63 has a change in lithology from primarily mangrove peat to fluvial material at ~2,500 cal. yrs BP, which is attributed to erosion inputs of the drainage basin. Changes in the pollen assemblage, such as greater input from non-mangrove pollen, are coeval with changes in sedimentation rates at both sites. Subfossil mangrove leaves, from core BT-79, are used for stable isotope ($\delta^{15}N$, $\delta^{13}C$, and $\delta^{18}O$) analyses to illustrate past physiology and seawater inundation. The composition of organic material in core SR-63 changes from autochthonous to allochthonous sources, which is coeval with the change in lithology. A decrease in the rate of sea-level rise is assumed to be the cause of the significant changes seen in these mangroves, which counters existing sea-level curves.

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Chapter 1 General Introduction

Mangroves are forests located at the boundary between land and sea in the tropics and subtropics and are therefore sensitive to environmental changes such as fluctuations in sea-level (Alongi, 2002). This study examines mangroves in Belize, Central America on the Caribbean Sea. Belize is a small country that is part of the Yucatan peninsula with mangrove forests covering a large portion of its coastline and offshore islands (Murray, 2003). Peat deposits from these mangrove forests show that these habitats have been present throughout the Holocene (e.g. Woodroffe, 1988). I took two cores primarily composed of peat to study the paleoecology of these ecosystems: one from a site offshore on Spanish Lookout Cay (SLC) and one along the banks of the Sibun River on the mainland (Figure 1.1). Using a multi-proxy approach, I examined evidence preserved in the cores in order to reconstruct environmental change and compared these data to previous research findings. A similar study was completed on Twin Cays, Belize (Wooller et al., 2004 and 2007), which is ~100km south of SLC (Figure 1.1). I have examined if paleoecological change seen on one low-lying mangrove island is seen on another island and on the mainland to determine if our observations represent a local or regional signal. In addition, I compared my findings to published Holocene sea-level curves for the Caribbean region (e.g. Toscano and Macintyre, 2003).

Sea-level is not static; it has been constantly changing for millions of years and is highly correlated with times of glaciation and deglaciation. In the more recent past, sea-level has risen ~120m since the end of the last glacial maximum (LGM), 21,000 calibrated years before present (cal. yrs BP), but the rate of sea-level rise was not constant. In Figure 1.2, Blanchon and Shaw (1995) defined a Catastrophic Rise Event (CRE) which gives the sea-level curve a step-like feature and results from large discharges of ice or meltwater. The last abrupt change in the rate of sea-level rise defined by this study, CRE 3, was in the early Holocene at ~7,600 cal. yrs BP (Blanchon and Shaw, 1995). However, changes in the rate of sea-level rise through the mid and early Holocene are noted to have occurred without significant meltwater contributions (e.g. Lighty et al., 1982; Fairbanks, 1989; Toscano and Macintyre, 2003).

Two predominant sample types have been analyzed to study sea-level change in the Caribbean. The first are corals, mainly *Acropora palmata*, the Elkhorn Coral, which were used to constrain the sea-level curve in Figure 1.2. This particular coral lives approximately 5m below sea-level and has been used to track both the position and rate of sea-level rise through the last deglacial transgression (Toscano and Lundberg, 1998) using U-Th dating techniques. However, because the coral lives below sea-level, it is only helpful to estimate minimum sea-level curves. The second sample type used to analyze sea-level change is peat produced by mangroves, dated using radiocarbon (^{14}C) measurements. Basal peat from *Rhizophora mangle* L. (red mangrove) was used in conjunction with data from *A. palmata* to produce Figure 1.3. Since mangrove habitats exist at or above sea-level, these sample types provide maximum sea-level estimations. Both *A. palmata* and *R. mangle* are present in the Caribbean region, which is a popular area to study sea-level change due to the assumption that it has been relatively tectonically stable since the LGM (Toscano and Macintyre, 2003). The exception is the known uplift happening to Barbados and respective sea-level curves have been adjusted to assumed uplift rates (e.g. Fairbanks, 1989).

Although there are a limited number of studies that examine how mangrove ecosystems have responded to past environmental changes, there is a considerable amount of literature on modern mangrove ecology and physiology. It is this wealth of information that aids paleoecological reconstructions. Mangroves are halophytes, salt tolerant plants, which live in the intertidal zone of the tropics and subtropics. These are marginal ecosystems, and mangrove plants have adapted to tolerate extreme temperatures, salinity, winds, flooding, limited nutrients, and anoxic substrate (e.g. Feller, 1995; Ellison and Farnsworth, 1997; Sobrado, 2000). The evolutionary steps taken to allow such adaptability in mangroves makes them exclusively competitive to plants that do not have the mechanisms to tolerate such an extreme environment (Vannucci, 2001). Like other marginal ecosystems, biodiversity is low and there are only three main species in the Caribbean region: *Avicennia germinans* (L.) Stearn. (black mangrove), *Laguncularia racemosa* (L.) C.F. Gaertn. (white mangrove), and *R .mangle*. One mangrove associate, *Conocarpus erectus* L. (buttonwood) is also found in the region.

R. mangle is the dominant species in many Belizean mangrove ecosystems and is also the only species whose subfossil leaf fragments have previously been found in mangrove peat cores from Belize (e.g. Wooller et al., 2004 and 2007). *R. mangle* stands have an obvious height gradient exemplifying their nutrient status (e.g. Feller et al., 2003; McKee et al., 2002). Dwarf *R. mangle* (<1m) are phosphorous limited, have a more negative $\delta^{15}N$ (Wooller et al., 2003), and are generally located in the interior of mangrove islands. Tall mangroves (>3m) are generally nitrogen limited, have a more negative $\delta^{13}C$ (Wooller et al., 2003), and usually fringe the edge of mangrove islands. Through understanding the isotopic ranges of dwarf and tall *R. mangle* leaves, Wooller et al. (2004) reconstructed the paleoecology of the mangrove ecosystem at Twin Cays, Belize (Figure 1.1).

Stable isotopic analyses on mangrove leaves are limited to observations of one species, *R. mangle*, through the Holocene because the leaves of *A. germinans* and *L. racemosa* are not well preserved. Therefore, I applied a multi-proxy approach to achieve a comprehensive reconstruction of environmental change through the Holocene. Several stratigraphic analyses were performed on the two cores that I took from Belize to understand sedimentological changes. Magnetic susceptibility provides information about possible volcanic eruptions (Michaud et al., 2006), erosion or weathering of iron-rich rocks or clays (Curtis et al., 1998), or even the presence of bacterial magnetosomes found in seawater (Berquo et al., 2004). Bulk density measurements were performed to determine whether autocompaction of the mangrove sediments affected the chronology or sedimentation rates as suggested by Bird et al. (2004). Pollen analysis was used to understand changes in the mangrove species, as well as changes in the proportions of other plant species in the area. Changes in percentages of pollen types from species that are not halophytes (e.g. *Myrsine* sp.) can lead to observations of changes in the type of water (fresh or salt) available at a site.

The studies completed from Twin Cays, Belize (Wooller et al. 2004 and 2007) tracked changes in ecosystems that were present throughout most of the Holocene.

Researchers observing sea-level change have studied environments where a mangrove ecosystem has only been present during discrete periods of the Holocene (Gischler and Hudson, 1998; Gischler, 2003) or environments where mangrove peat was not continuous (Rull et al., 1999; Cohen et al., 2005; Vedel et al., 2006). I have applied the techniques described above to analyze the records at Spanish Lookout Cay and along the Sibun River. It is important to understand past sea-level changes to help predict the future of sea-level change and how it will affect our coasts. Changes in sea-level control the evolution of coastal environments. It affects the level of continental exposure, weathering processes (e.g. fluvial, aeolian), and the volume and nature of terrestrial derived sediments deposited on continental shelves (Mastronuzzi et al., 2005). By understanding how mangrove ecosystems respond to environmental changes, such as sea-level change, models will become more accurate at predicting the likely status of an ecosystem in light of future of sea-level change scenarios.

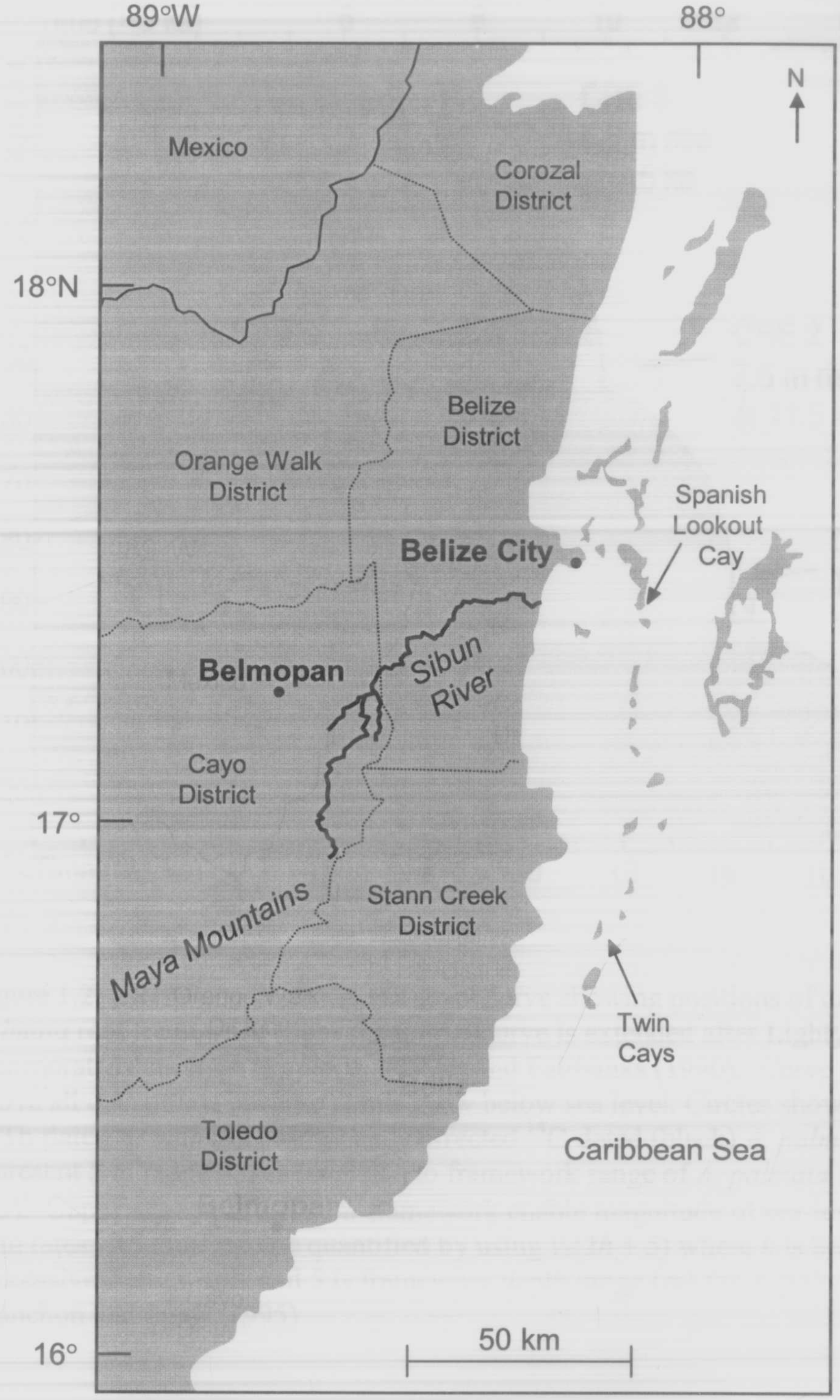

Figure 1.1: Map of Belize, Central America showing locations of the sites cored for this study.

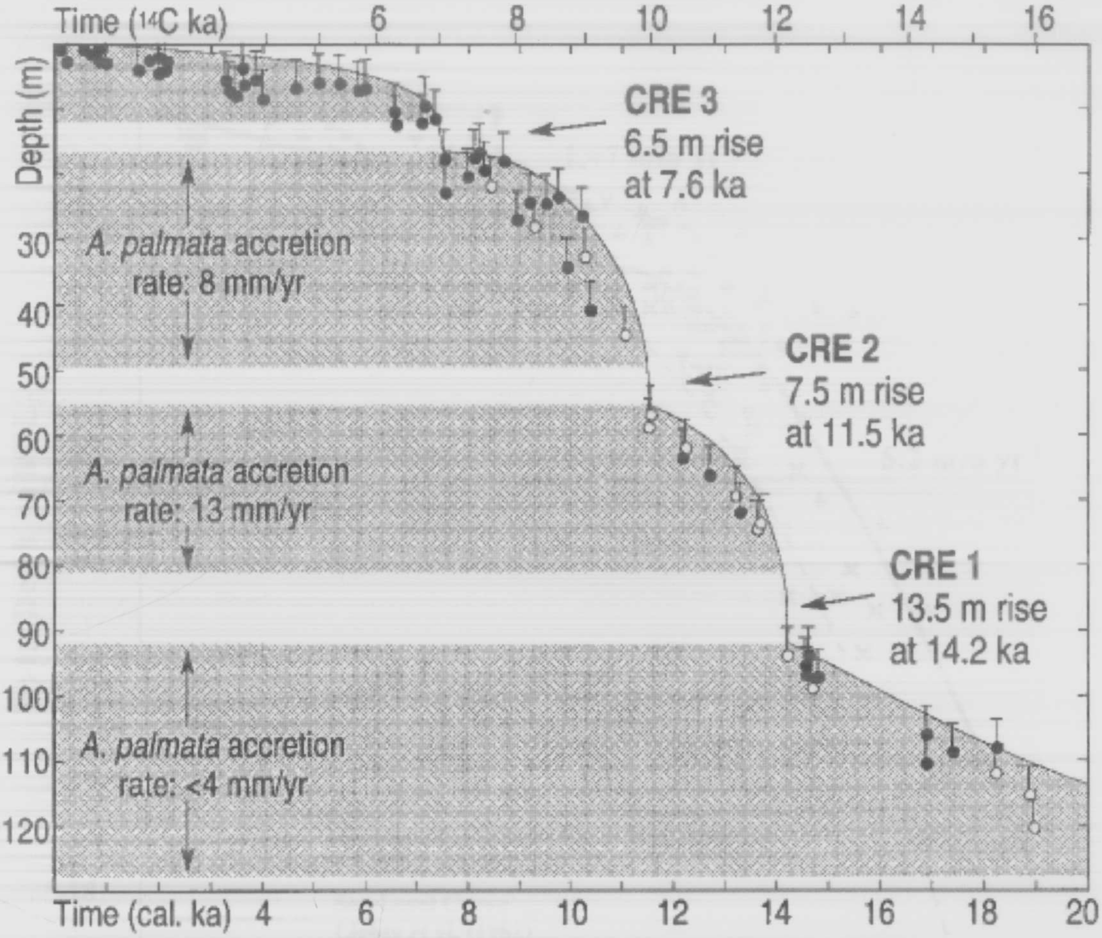

Figure 1.2: Caribbean deglacial sea-level curve showing positions of drowned *Acropora palmata* reef framework (light shading). Curve is extended after Lighty et al. (1982) and incorporates data from Bard et al. (1990) and Fairbanks (1990). Curve must lie on or above all data points because corals grow below sea level. Circles show positions of U/Th-dated *A. palmata* (white) and corrected ^{14}C-dated (black) *A. palmata*; error bars represent 5 m range of sea level due to framework range of *A. palmata* and age error (1σ). Gaps between *A. palmata* framework enable magnitude of sea-level rise events with rates >45 mm/yr to be quantified by using ½(2*h* + 5) where *h* is height (m) between successive frameworks and 5 is framework depth range (m) for *A. palmata*. (taken from Blanchon and Shaw, 1995)

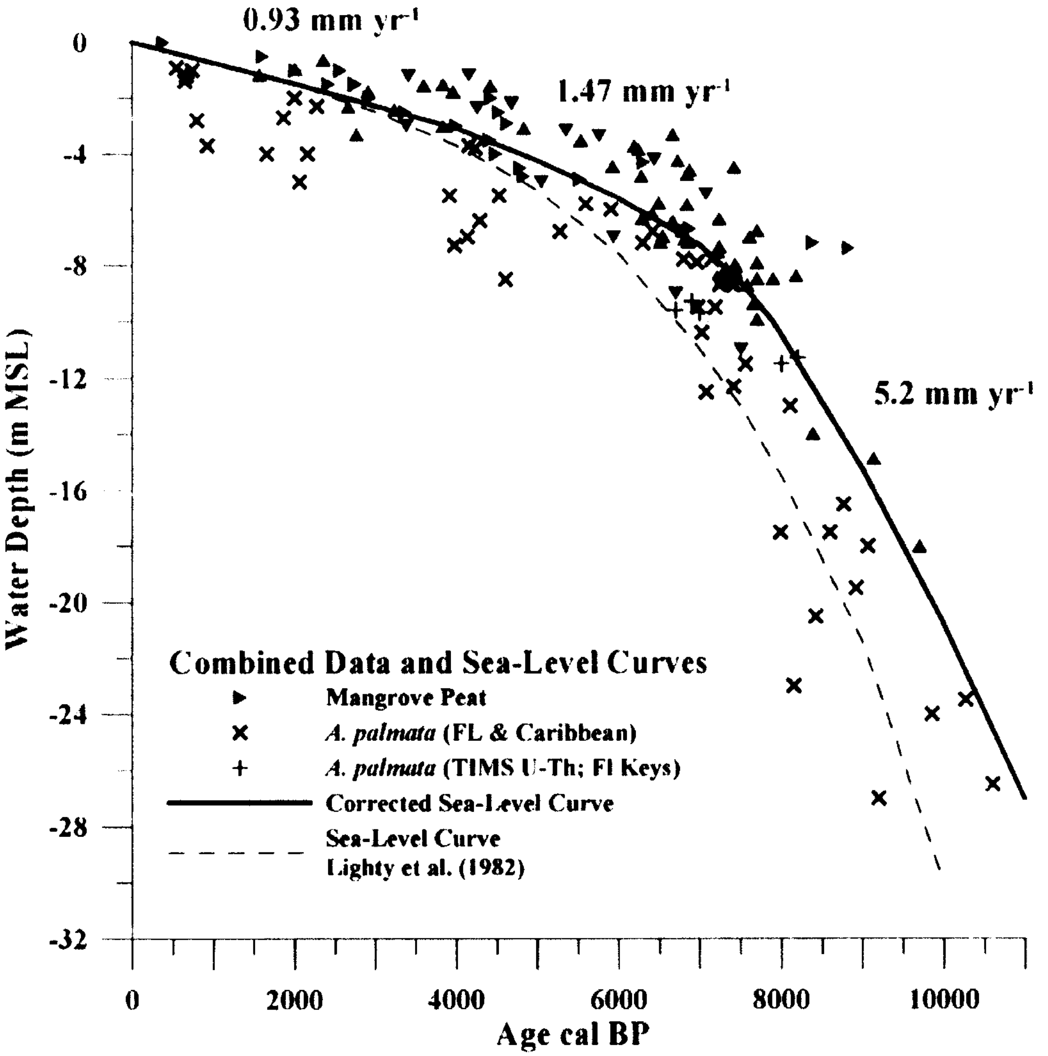

Figure 1.3: Combined peat and coral data. The sea-level curve describes a reasonable line of demarcation between intertidal peat data (maximum sea level) and shallow reef crest *A. palmata* data (minimum sea level required to submerge coral). Rates of SLR from the combined coral-peat record range from 5.2 mm yr^{-1} (10.6–7.7 ka), 1.47 mm yr^{-1} (7.7–2 ka), dropping to 0.93 mm yr^{-1} (2–0.4 ka). All rates are well below the maximum accretion rate of *A. palmata* and are not indicative of extreme pulses capable of effecting reef drowning and subsequent backstepping. (taken from Toscano and Macintyre, 2003)

References

Alongi, D.M., 2002. Present state and future of the world's mangrove forests. Environmental Conservation 29, 331-349.

Bard, E., Hamelin, B., Fairbanks, R.G. and Zindler, A., 1990. Calibration of the 14C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals. Nature 345(6274), 405-410.

Berquo, T.S., Thompson, R. and Partiti, C.S.M., 2004. Magnetic study of Brazilian peats from Sao Paulo state. Geoderma 118, 233-243.

Bird, M.I., Fifield, L.K., Chua, S. and Goh, B., 2004. Calculating sediment compaction for radiocarbon dating of intertidal sediments. Radiocarbon 46, 421-435.

Blanchon, P. and Shaw, J., 1995. Reef drowning during the last deglaciation: Evidence for catastrophic sea-level rise and ice-sheet collapse. Geology 23, 4-8.

Cohen, M.C.L., Behling, H. and Lara, R.J., 2005. Amazonian mangrove dynamics during the last millennium: The relative sea-level and the Little Ice Age. Review of Palaeobotany and Palynology 136, 93-108.

Curtis, J.H., Brenner, M., Hodell, D.A., Balser, R.A., Islebe, G.A. and Hooghiemstra, H., 1998. A multi-proxy study of Holocene environmental change in the Maya lowlands of Peten, Guatemala. Journal of Paleolimnology 19, 139-159.

Ellison, A.M. and Farnsworth, E. J., 1997. Simulated sea level change alters anatomy, physiology, growth, and reproduction of red mangrove (*Rhizophora mangle* L.). Oecologia 112, 435-446.

Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea-level record - Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature 342, 637-642.

Fairbanks, R.G., 1990. The age and origin of the ''Younger Dryas climate event'' in Greenland ice cores: Paleoceanography, v. 5, pp. 937–948.

Feller, I.C., 1995. Effects of nutrient enrichment on growth and herbivory of dwarf red mangrove (*Rhizophora mangle*). Ecological Monographs 65, 477-505.

Feller, I.C., McKee, K.L., Whigham, D.F. and O'Neill, J.P., 2003. Nitrogen vs. phosphorus limitation across an ecotonal gradient in a mangrove forest. Biogeochemistry 62, 145-175.

Gischler, E., 2003. Holocene lagoonal development in the isolated carbonate platforms off Belize. Sedimentary Geology 159, 113-132.

Gischler, E. and Hudson, J.H., 1998. Holocene development of three isolated carbonate platforms, Belize, Central America. Marine Geology 144, 333-347.

Lighty, R.G., Macintyre, I.G. and Stuckenrath, R., 1982. *Acropora palmata* reef framework: A reliable indicator of sea-level in the western Atlantic for the past 10,000 years. Coral Reefs 1, 125–130.

Mastronuzzi, G., Sansòb, P., Murray-Wallacec, C.V. and Shennand, I., 2005. Quaternary coastal morphology and sea-level changes—an introduction. Quaternary Science Reviews 24(18-19), 1963-1968.

McKee, K.L., Feller, I.C., Popp, M. and Wanek, W., 2002. Mangrove isotopic (delta N-15 and delta C-13) fractionation across a nitrogen vs. phosphorus limitation gradient. Ecology 83, 1065-1075.

Michaud, F., Ramirez-Sanchez, H.U., Parron, C., Zarate-del Valle, P.F., Fernex, F. and Barci-Funel, G., 2006. Strong magnetic levels in Lake Chapala sediments (western Mexico): Their mineralogy and stratigraphic significance. Journal of Paleolimnology 35, 819-836.

Murray, M.R., Zisman, S.A., Furley, P.A., Munro, D.M., Gibson, J., Ratter, J., Bridgewater, S., Minty, C.D. and Place, C.J., 2003. The mangroves of Belize Part 1. Distribution, composition and classification. Forest Ecology and Management 174, 265-279.

Rull, V., Vegas-Vilarrubia, T. and de Pernia, N.E., 1999. Palynological record of an early-mid Holocene mangrove in eastern Venezuela. Implications for sea-level rise and disturbance history. Journal of Coastal Research 15, 496-504.

Sobrado, M.A., 2000. Relation of water transport to leaf gas exchange properties in three mangrove species. Trees 14, 258-262.

Toscano, M.A. and Lundberg, J., 1998. Early Holocene sea-level record from submerged fossil reefs on the southeast Florida margin. Geology 26, 255-258.

Toscano, M.A. and Macintyre, I.G., 2003. Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated ^{14}C dates from *Acropora palmata* framework and intertidal mangrove peat. Coral Reefs 22, 257-270.

Vannucci, M., 2001. What is so special about mangroves? Brazilian Journal of Biology 61, 599-603.

Vedel, V., Behling, H., Cohen, M. and Lara, R., 2006. Holocene mangrove dynamics and sea-level changes in northern Brazil, inferences from the Taperebal core in northeastern Para State. Vegetation History and Archaeobotany 15, 115-123.

Woodroffe, C.D., 1988. Mangroves and sedimentation in reef environments: Indicators of past sea-level changes, and present sea-level trends? Proceedings of the Sixth International Coral Reef Congress 3, 535-539.

Wooller, M., Smallwood, B., Scharler, U., Jacobson, M. and Fogel, M., 2003. A taphonomic study of $\delta^{13}C$ and $\delta^{15}N$ values in *Rhizophora mangle* leaves for a multi-proxy approach to mangrove palaeoecology. Organic Geochemistry 34, 1259-1475.

Wooller, M.J., Behling, H., Smallwood, B.J. and Fogel, M., 2004. Mangrove ecosystem dynamics and elemental cycling at Twin Cays, Belize, during the Holocene. Journal of Quaternary Science 19, 703-711.

Wooller, M.J., Morgan, R., Fowell, S.J., Behling, H., Fogel, M., 2007. A multi-proxy peat record of Holocene mangrove paleoecology from Twin Cays, Belize. Holocene 17(8), 1129-1139.

Chapter 2

Mangrove ecosystem changes during the Holocene at Spanish Lookout Cay, Belize[1]

Abstract

Belize, located on the Caribbean Sea, has extensive mangrove ecosystems, which are a dynamic coastal habitat and are constantly adapting to environmental change. We observed multiple, independent proxies of a continuous mangrove peat core (BT-79) from Spanish Lookout Cay to examine how these ecosystems responded to environmental change during the Holocene. Radiocarbon measurements show this site was colonized by mangroves at ~8,000 cal. yrs BP and there was a significant decrease in peat accumulation rates from ~6,000 to 1,000 cal. yrs BP. Pollen analyses of the core illustrate high abundances of *Rhizophora* pollen and an increase in non-mangrove pollen at ~5,000 cal. yrs BP. Subfossil *Rhizophora mangle* leaves are used for stable nitrogen, carbon, and oxygen isotope ($\delta^{15}N$, $\delta^{13}C$, and $\delta^{18}O$) analyses, which provide a proxy of this plant's past physiology, stand structure, and seawater inundation. A decrease in inundation at the site is shown by decreased $\delta^{18}O$ values from ~7,000 to ~1,000 cal. yrs BP. Existing Caribbean sea-level data for the Holocene do not show evidence of a

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[1] Monacci, N.M., Meier-Grünhagen, U., Finney, B.P., Behling, H., Wooller, M.J., in prep. Mangrove ecosystem changes during the Holocene at Spanish Lookout Cay, Belize. Palaeogeography, Palaeoclimatology, and Palaeoecology.

decrease in the rate of sea-level rise or fluctuations that are assumed to be the major causes of change at site BT-79.

Introduction

Mangroves are part of a marginal ecosystem, able to tolerate extreme environmental factors such as high salinity and constant inundation (Vannucci, 2001). These halophytes have adapted physiological traits such as conservative water use, seen in *Avicennia germinans* (L.) Stearn. (black mangrove) and *Laguncularia racemosa* (L.) C.F. Gaertn. (white mangrove) (Sobrado, 2001 and 2005), or structural adaptations like the characteristic aerial roots of *Rhizophora mangle* L. (red mangrove) that allow them to exist in a marine environment. This dynamic habitat is associated with constantly changing environmental conditions, including sea-level. Predictions of future sea-level rise are highly dependent on temporal and spatial variability in rates of relative sea-level rise (IPCC, 2001). Understanding the influence sea-level rise will have on the physiology and composition of mangrove ecosystems will aid in estimating the impact of future sea-level scenarios. One way of examining how mangroves respond to sea-level change is to observe how these plants have responded during past sea-level fluctuations. Sea-level is known to have risen throughout the Holocene and rates of sea-level rise have been extensively studied throughout Central and South America (e.g. Lighty et al., 1982; Angulo and Lessa, 1997). Details associated with published sea-level curves, such as the timing associated with changes in the rate of rise of sea-level or the relative sea-level height, continue to be researched and debated (e.g. Fairbanks, 1989; Toscano and Macintyre, 2003; Blanchon, 2005; Gischler, 2006). In addition, possible fluctuations, or departures from a smoothly rising sea-level curve, have also been presented (Kearney, 2001; Angulo et al., 2006).

Belize, located in Central America on the Caribbean Sea, has extensive mangrove forests along its coast and offshore on low-lying mangrove islands. These islands comprise 3% of the country's 689km^2 land area (Murray et al., 2003), and many have large mangrove peat deposits. Paleoecological investigations have previously been conducted on peat deposits to determine the response of mangroves to past environmental change (e.g. Macintyre et al., 1995; McKee and Faulkner, 2000; Wooller et al., 2004 and 2007). Paleoecological records from Twin Cays indicate shifts between tall or transitional *R. mangle* stands throughout much of the Holocene, a switch to *Myrsine* vegetation and increased peat accumulation occurred at ~4,100 cal. yrs BP, and the proportions of available source water (seawater vs. precipitation) varied as the rate of sea-level rise changed (Wooller et al., 2004 and 2007). The majority of these previous studies have been conducted on mangrove islands (e.g. Tobacco Range, Carrie Bow Cay, Pelican Cays and Twin Cays) south of our current study site, Spanish Lookout Cay (SLC). The primary aim of this study is to determine whether environmental changes recorded in one relatively localized area (e.g. Twin Cays) can be identified in another mangrove habitat offshore Belize (SLC).

SLC is a mangrove island located off of Belize City, approximately 100km north of Twin Cays. We incorporate radiocarbon measurements, stratigraphic characteristics, pollen analysis, and stable isotopes to gain a picture of the past ecosystem throughout the last ~8,000 cal. yrs BP. Pollen data has previously provided evidence of vegetation changes in mainland (e.g. Rull et al., 1999; Vedel et al., 2006) and island habitats (e.g. Ellison, 1996; Higuera-Gundy et al., 1999; Wooller et al., 2004 and 2007). These data can be complemented with stable isotope analyses ($\delta^{13}C$, $\delta^{15}N$, and $\delta^{18}O$) of subfossil mangrove leaves to examine past physiological changes in a specific species (e.g. *R. mangle*). For example, nutrient limitation leads to changes in the height of *R. mangle* and is reflected in the stable nitrogen and carbon isotope signature of the leaves (McKee et al., 2002; Wooller et al., 2003a). Understanding the environmental changes at SLC will allow us to determine whether the mangroves in Belize responded to local or regional changes.

Study Site

This study took place on SLC, a 76ha, low-lying, mangrove island located approximately 15km east of mainland Belize (17°24′ N, 88°04′ W, Figure 2.1). A survey conducted on SLC shows an average of 8m of peat above Pleistocene limestone. The tidal range at SLC is assumed to be micro-tidal (~15-30cm) (Kjerfve et al., 1982) as it is at various other places throughout offshore Belize (e.g. Carrie Bow Cay). Belize is located in the transition zone from the tropics to subtropics and experiences a dry season from December to April and a wet season from May to November. The mainland is composed of several vegetation types including mangrove ecosystems (coastal), tropical evergreen needle and broadleaf forests (inland), and scattered agricultural areas (Meerman and Sabido, 2001). Several sand and coral cays are scattered throughout the region with either little vegetation or planted species of Arecaceae, (personal observation). Mangroves are the dominant vegetation on SLC and *R. mangle* is the principal species.

The stands of mangroves at SLC have a visible tree height gradient; tall *R. mangle* (>3m) are generally found on the fringe of the island, while dwarf *R. mangle* (<1m) are located in the interior of the island. On other islands along the coast of Belize (e.g. Twin Cays), this stand height gradient has been related to nutrient limitation (Feller et al., 2003; Lovelock et al., 2006), where dwarf trees are primarily phosphorus limited and tall trees are largely nitrogen limited. *R. mangle* in the transition zone (1-3m) is frequently mixed with the other two mangrove species found on SLC: *A. germinans* and *L. racemosa*. Relative to *R. mangle*, these species have physiological advantages in areas with high or fluctuating salinities (Sobrado, 2000). One mangrove associate, *Conocarpus erectus* L. (buttonwood), is also present on the island, but it is assumed to have been planted for aesthetic purposes and was not observed to be growing naturally (personal observation).

Methods

Sampling

Green *R. mangle* leaves were sampled from various locations around SLC (Figure 2.1) covering gradients in tree height and environmental characteristics. These samples were used for range comparisons of stable carbon and nitrogen isotope (δ^{13}C and δ^{15}N, respectively) variability as done in a similar study undertaken at Twin Cays, Belize (Wooller et al., 2003b). Stable oxygen isotope analyses (δ^{18}O) were also conducted on the living *R. mangle* leaves to determine modern variation on the island. All leaf samples were oven-dried (50°C) in the field before transport to University of Alaska Fairbanks (UAF) for analyses.

A variety of water samples were collected for δ^{18}O and δD analysis (Figure 2.2). Surface water samples came from shallow ponds or creeks in the interior of the island as well as ocean surface water on the edge of SLC, covering a large salinity range (0-56‰). Pore water samples were collected 20cm below ground using a pore water sipper. Precipitation samples were collected during rain events in a beaker at a constant location on the windward (northeast) side of the island. Several events were sampled per day, for two weeks in December 2005. All water samples were immediately transferred into 2mL glass vials and capped with no head space.

Upon visiting SLC, a survey of peat thickness was conducted using an Eijkelkamp Russian peat corer to determine peat accumulation patterns and the location with the thickest peat. The coring site selected (BT-79, Figure 2.1) was located at the north end of the island on the edge of a shallow pond. Dwarf *R. mangle* currently grow at the site. BT-79 was taken in 16 50cm x 5cm sections, individually wrapped in plastic, and kept in coolers at the Spanish Bay Conservation and Research Center on SLC. Layers with lesser or greater resistance and visual color transitions were noted in the field (Figure 2.3d). The core sections were later transferred to the Water and Environmental Research Center (WERC) at UAF where they are kept at ~4°C.

Core chronology

Radiocarbon measurements (^{14}C) were conducted on *R. mangle* leaf fragments, wood pieces, and macroscopic roots (Table 2.1). Samples were rinsed with deionized water, freeze dried, and measured using accelerator mass spectrometry (AMS) at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) at Woods Hole Oceanographic Institute (WHOI). All ^{14}C dates were calibrated using CALIB 5.0.2 according to Reimer et al. (2004) and rounded to the nearest decade.

Sediment characteristics

Magnetic susceptibility (MS), dry bulk density (BD), and loss on ignition (LOI) were measured on bulk peat samples. A $2cm^3$ sample was taken every 10cm throughout BT-79 and freeze dried. MS (SI) was measured on the dry sample using a Bartington

MS2 sensor. BD (g cm^{-3}) was determined by weighing the wet sample, freeze drying, and weighing the dry sample. Sequential LOI measurements were then taken following standard procedures (e.g. Dean, 1974; Heiri et al., 2001) to estimate organic carbon content (LOI, %), carbonate content ($CaCO_3$, %), and remaining material (Residue, %).

Pollen analysis

All pollen preparation and analytical procedures were performed in the Paleontology Laboratory in the Department of Geology and Geophysics at UAF. At least every 20cm down core, a $1cm^3$ sample was prepared using the standard method (Faegri and Iverson, 1989). One tablet of exotic *Lycopodium* spores (18,583 ± 762 spores $tablet^{-1}$) was added to each sample prior to processing in order to calculate pollen concentrations (grains cm^{-3}). Samples were mounted in a glycerin gelatin medium for counting. A minimum of 300 pollen grains were identified for each pollen sample; however, a few samples with extremely low concentrations were counted to a minimum of 100 grains. Pollen grains were identified using published pollen morphological descriptions (e.g. Palacios-Chavez et al., 1991). Pollen data are reported as percentages of the total pollen sum, which excludes fern and fungal spores and microforaminifera.

Stable isotope analysis: $\delta^{13}C$, $\delta^{15}N$

Subfossil *R. mangle* leaves were picked from 1cm thick sections at least every 10cm and analyzed for their stable carbon and nitrogen isotope composition. The leaves were rinsed with deionized water, measured to determine leaf area (mm^2) (Wooller et al., 2004), and freeze-dried. All stable isotope analyses were performed at the Alaska Stable Isotope Facility (ASIF) at UAF and reported in standard delta (δ, ‰) notation. The elemental compositions (%C and %N, wt. %) and stable isotope data ($\delta^{13}C$ and $\delta^{15}N$, ‰) were measured with an EA-IRMS utilizing a Costech Instruments Elemental Analyzer (EA) coupled with a Conflo III to a Thermo Electron Delta V Isotope Ratio Mass Spectrometer (IRMS). $\delta^{13}C$ is reported relative to Vienna Pee Dee Belemnite (VPDB) and $\delta^{15}N$ is reported relative to atmospheric nitrogen (atm). Peptone was used as a reference standard to determine analytical precisions: 1.4% (%C), 0.7% (%N), 0.1‰ ($\delta^{13}C$), and 0.4‰ ($\delta^{15}N$).

Stable isotope analysis: $\delta^{18}O$ and δD

The $\delta^{18}O$ of *R. mangle* leaves (modern and subfossil) was measured using pyrolysis-EA-IRMS with a Thermo Finnigan MAT high temperature TC/EA and Conflo III interface with a Delta V IRMS. All solid $\delta^{18}O$ samples were freeze-dried under vacuum for no less than 6 days prior to analyses and loaded into a Costech Zero-blank autosampler purged with Helium (see methods in O'Brien and Wooller, 2007). Benzoic acid was used as a reference standard for solid samples to determine analytical precision: 2.4% (%O) and 0.8‰ ($\delta^{18}O$). For $\delta^{18}O$ and δD analyses of water samples, a 0.2uL aliquot was injected into the TC/EA-IRMS with a CTC Analytics A200SE liquid autosampler with analytical precisions 0.7‰ ($\delta^{18}O$) and 2.0‰ (δD). All $\delta^{18}O$ and δD samples are reported relative to Vienna Standard Mean Ocean Water (VSMOW).

Results

The mean $\delta^{15}N$ of living, green *R. mangle* leaves from dwarf trees (-4.0±3.4‰, n = 3) was lower than that from tall trees (1.4±1.3‰, n = 6). The mean $\delta^{13}C$ of living, green *R. mangle* leaves from tall trees (-29.6±1.0‰) was lower than that from dwarf trees (-26.0±0.6‰). Wooller et al. (2003b) also found dwarf trees to have more negative $\delta^{15}N$ values (-10‰) and more positive $\delta^{13}C$ values (-25.3‰) than tall *R. mangle* ($\delta^{15}N$ = 0.0‰, $\delta^{13}C$ = -28.3‰). The total $\delta^{18}O$ range for both dwarf and tall trees was 22.4 to 28.3‰. The mean $\delta^{18}O$ values of dwarf (26.3±1.0‰) and tall (25.3±1.5‰) *R. mangle* were not significantly different (t-test: $P > 0.4$, n = 9).

The $\delta^{18}O$ of all water samples collected at SLC ranged from -4.6 to 2.3‰, and δD ranged from -10.2 to 17.0‰ (Figure 2.2). The mean $\delta^{18}O$ and δD of surface (0.7±0.9 and 9.2±3.4‰, respectively) and pore waters (0.6±0.8 and 8.6±3.1‰, respectively) were not significantly different (t-test: $P > 0.2$, n=26). However, the rain water (n = 9) had lower

$\delta^{18}O$ and δD values relative to the surface and pore waters: -2.6±0.9‰ and 0.6±5.4‰ respectively.

Four pollen zones are used to illustrate the data from core BT-79: zone BT-79 I (735-800cm), zone BT-79 II (325-735cm), zone BT-79 III (125-325cm), and zone BT-79 IV (0-125cm). These zones were established through a Constrained Incremental Sum of Squares (CONISS) analysis performed in TILIA (Grimm, 2004) on the pollen data.

Core chronology

The radiocarbon (^{14}C) measurements from core BT-79 and their calibrated ages are presented in Table 2.1. Most samples used for dating were *R. mangle* leaf fragments. Exceptions were samples from depths 324-325cm and 725-726cm, which were macroscopic roots, and 524-526cm, which was a wood fragment. Upon examination of the age vs. depth distribution (Figure 2.3a), it was concluded that three linear equations allowed determination of ages for all depths of BT-79 and calculation of sedimentation rates. The bottom section of the core (380-800cm) was fitted and has linear equation:

$$Depth_{cm} = 0.24Age_{cal.yrs\,BP} - 1154.14, R^2 = 0.90, n = 4 \qquad (1)$$

yielding a sedimentation rate of 0.24cm yr^{-1}. The calibrated dates were extrapolated to estimate a basal date of ~8,100 cal. yrs BP for BT-79. The middle section (260-370cm) was fitted and has linear equation:

$$Depth_{cm} = 0.02Age_{cal.yrs\,BP} + 246.87, R^2 = 0.89, n = 3 \qquad (2)$$

yielding a sedimentation rate of 0.02cm yr^{-1}. The top section of the core (0-250cm) was fitted assuming an age of -55 cal. yrs BP (Reimer et al., 2004) for depth 0-1cm, yielding a linear equation:

$$Depth_{cm} = 0.39 Age_{cal.\,yrs\,BP} + 5.16, R^2 = 0.97, n = 3 \qquad (3)$$

yielding a sedimentation rate of 0.39cm yr^{-1}. We chose to use multiple linear fits over a comprehensive logarithmic fit as it resulted in negative sedimentation rates and inverse modeled ages for each depth. We acknowledge that this method may give the false impression that the change in the sedimentation rates was abrupt rather than more gradual (Figure 2.3a).

Sediment characteristics

Magnetic susceptibility (MS) ranges from 0.6 to 2.2 throughout BT-79 (Figure 2.3b). Maximum and minimum values are both observed in zone BT-79 I. MS increases from 1.3 at the start of zone BT-79 II to a zone maximum of 2.1 at 530cm and then decreases to 1.0 at the end of the zone. In zone BT-79 III, MS fluctuates with peaks of 1.8 and 1.6 at 150 and 240cm, respectively. MS averages 0.9 during BT-79 IV with little variability (±0.2).

BD (dry) ranges from 0.08 to 0.60g cm^{-3} throughout BT-79 (Figure 2.3c). However, BT-79 I is the only zone with values greater than 0.15g cm^{-3}, minimal variation (0.12±0.02g cm^{-3}) is evident in the other zones.

LOI ranges from 20 to 75% throughout the core (Figure 2.4a). Low LOI values, <54%, are evident in zone BT-79 I. LOI values are relatively constant throughout zone BT-79 II (64±3%), but values generally decrease from 68% at the beginning of zone BT-79 III to 58% at the end of the zone. LOI values begin increasing at 70cm in zone BT-79 IV, with the maximum value observed at 10cm (75%).

$CaCO_3$ ranges from 6 to 38% with the maximum value observed in zone BT-79 I (Figure 2.4b). $CaCO_3$ generally decreases through zone BT-79 II to the minimum value of 6% at 350cm. $CaCO_3$ increases slightly during zone BT-79 III and decreases in zone BT-79 IV.

Residue after the 550 and 850°C burns ranges from 0 to 72%, with both the minimum and maximum values observed in BT-79 I (Figure 2.4c). Percent residue increases overall during zone BT-79 II, with ~0% at the beginning and ~25% at the end. A slight decrease in the residue occurred at the transition between zone BT-79 II and BT-79 III, and a noticeable peak was observed in BT-79 IV: 29% at 40cm and is the highest value seen outside zone BT-79 I.

Pollen analysis

All pollen data is presented in percentages: Figure 2.5 illustrates pollen summaries based on the type of plant and Figure 2.6 illustrates percentages of common genera. Pollen concentrations range from 2,392 to 28,791grains cm^{-3}, and pollen accumulation rates range from 6,133 to 812,195grains cm^{-2} yr^{-1}. Typically, greater than

300 grains were counted per sample, with a few exceptions (counted to between 105 and 262) due to particularly low pollen concentrations.

BT-79 Zone I: 781-735cm

Pollen zone BT-79 I is characterized by several unique features. *R. mangle* abundance is relatively low (<75%) due to high amounts of *A. germinans* pollen (≥5%), particularly at depth 760cm (18%) (Figure 2.6). Other shrubs and trees are evident, including *Pinus*, *Quercus*, and *Alnus*. The highest values of Moraceae/Urticaceae (8%) are observed at depths 740 and 781cm. A grain of Sapindaceae is found at 760cm. The maximum abundance of Poaceae (8%) is recorded at the base of the core (781cm). Depth 781cm was remarkably different from other samples in terms of low pollen concentration (5,623grains cm^{-3}) and the absence of microforaminifera. However, a relatively high number of fungal spores (292% of pollen sum) and fern spores (4% of pollen sum) are observed.

BT-79 Zone II: 325-735cm

BT-79 II contains the highest abundance of *R. mangle* (85%) in the entire core; therefore, there is a reduced percentage of pollen from other shrubs and trees (Figure 2.5). The abundance of *A. germinans* pollen is low (2%) with minor peaks at 340 and 500cm. *Ilex*, *Juniperus*-type, and *Ulmus* pollen are observed solely in this zone. The highest percentage of Cyperaceae (3%) is observed at 640cm, but it is not evident throughout this zone. Pollen concentrations are among the highest (>21,000grains cm^{-3}) in the middle of BT-79 II (520-600cm), and the pollen accumulation rate begins to increase towards the end of the zone. Microforaminifera are found in every sample and show a slight increase from the base of the zone to 620cm.

BT-79 Zone III: 125-325cm

The abundance of *R. mangle* decreases to <75% at depths 240 and 310cm, where the increase in the abundance of *Myrsine*-type (8 and 6% respectively) is observed. The abundance of pollen from other shrubs and trees also increase during this zone. At depth 310cm, herb and grass abundance peaks (20%) and the highest value (17%) of Amaranthaceae/Chenopodiaceae was recorded. Also at 310cm, large numbers of microforaminifera are seen (21% of pollen sum) and fern spores are also present (1% of pollen sum). Pollen concentrations were variable in this zone, ranging from 3,587 to 16,898grains cm^{-3}.

BT-79 Zone IV: 0-125cm

The amount of pollen from non-mangrove shrubs and trees increases in pollen zone BT-79 IV: 25% (90cm) is the highest value and 16% is the highest average of any zone. This translates into the lowest values of *R. mangle* pollen and mangroves (<70%).

Pinus and *Quercus* are the major contributors of pollen from other trees and show maximum values in this zone (11 and 8% respectively). Anacardiaceae makes a sole appearance at the top of the core (0cm). Minimum pollen concentrations are recorded during BT-79 IV, with 2,392grains cm^{-3} at 40cm.

Stable isotope analysis (R. mangle leaves): $\delta^{15}N$, $\delta^{13}C$, and $\delta^{18}O$

Figure 2.7a illustrates leaf area, which ranges from 0 (no leaves found) to 600mm^2. Nitrogen content ranges from 0.3 to 2.0%, Carbon content ranges from 19.5 to 49.8%, and Oxygen content ranges from 23.7 to 34.2%. The atomic carbon:nitrogen ratios ($[C:N]_{a}$) range from 75 to 139mol:mol. Throughout the core, δ^{15}N range from -5.5 to 13.5‰, δ^{13}C range from -28.8 to -26.1‰, and δ^{18}O range from 15.3 to 22.1‰ (Figure 2.7b-d).

The δ^{15}N and δ^{13}C of the organic material in the mud layer (796-798cm) average 3.0±1.0‰ and -27.1±0.4‰, respectively. Only one sample (740cm) provided leaves for δ^{15}N and δ^{13}C analysis from zone BT-79 I, these values are -1.0 and -26.9‰, respectively.

δ^{15}N showed little variability at the base of zone BT-79 II, with a mean of -0.4±1.9‰ from 725 to 460cm. Exceptions are a low δ^{15}N value of -5.1‰ at 615cm and two high δ^{15}N values of 5.2 and 4.8‰ at depths 580 and 485cm, respectively. Lower δ^{15}N values ranging from -3.5 to -1.0‰ are observed from 445 to 380cm, but return to 1.4‰ at 347cm. There is a general increase in δ^{13}C from -27.4‰ at 725cm to -26.1‰ at

580cm. The middle of zone BT-79 II (~550-400cm) shows the most variation, with core-wide maximum (-26.1‰) and minimum (-28.8‰) $\delta^{13}C$ values at depths 505 and 485cm, respectively. There is a slight decrease in $\delta^{13}C$ at the top of the zone, from -26.8‰ at 380cm to -27.5‰ at 325cm. $\delta^{18}O$ is generally more positive at the base of zone BT-79 II with a mean of 19.6±1.7‰ from depths 715 to 515cm. A shift to more negative values begins at 505cm and the end of the zone (505-347cm) average 17.4±1.5‰, including the minimum $\delta^{18}O$ value observed (15.3‰) at depths 401 and 501cm.

The base (250-325cm) of zone BT-79 III has few data points due to the lack of leaves found in this section. Above this (205-245cm), zone BT-79 III is characterized by high $\delta^{15}N$ values (>6‰) with a maximum value of 13.5‰. In contrast, the most negative $\delta^{15}N$ value (-5.5‰) is also seen in BT-79 III, at 180cm. $\delta^{13}C$ values generally increase throughout zone BT-79 III to values of -26.8‰ at the end of the zone. $\delta^{18}O$ values start at 15.7‰ at 301cm and increase to the maximum value observed, 22.1‰ at 205cm. $\delta^{18}O$ decreases again to 17.1‰ at 151cm.

BT-79 IV has the highest average leaf area with three samples exceeding 400mm^2. $\delta^{15}N$ values show less variability in BT-79 IV than in the other three zones, with a mean of -1.5±1.4‰. $\delta^{13}C$ values show a slight increase from -28.0‰ at the base of the zone to -27.0‰ at the top of the zone. Zone BT-79 IV shows the least amount of variation in $\delta^{18}O$, with a 2.6‰ range.

Interpretation and Discussion

BT-79 Zone I: 8,100 to 7,890 cal. yrs BP

The transition of site BT-79 from a completely marine to a more terrestrial ecosystem began at ~8,000 cal. yrs BP. Evidence of a marine ecosystem recorded in zone BT-79 I, which is very different from the rest of the core, includes the presence of mud and carbonate and the absence of mangrove peat and leaves at the base of the core (Figure 2.3d). Further to the east of SLC and Twin Cays, other offshore carbonate platforms were being flooded and colonized by mangroves at the same time. Gischler et al. (2003) observed mangrove peat on Glover's Reef between 8,500 and 7,500 cal. yrs BP, although a lagoon developed at this site and mangrove peat is not continuous.

After mangroves became established at SLC (~8,000 cal. yrs BP), the pollen data indicates that both *R. mangle* and *A. germinans* were dominant. Relative to *R. mangle*, *A. germinans* is a low pollen producer (Behling et al., 2001) and therefore, we conclude the peak abundance (9%) observed of *A. germinans* indicates that the species was growing at the site (Vedel et al., 2006). The high abundance of fern and fungal spores in this zone also demonstrate the transitional nature of the site, as ferns are often the first plants to colonize an area (Traverse, 1988).

BT-79 Zone II: 7,850 to 4,180 cal. yrs BP

A mangrove ecosystem was established at the site of BT-79 from 7,850 cal. yrs BP to the present. The relatively uniform BD values for this zone indicate no significant compaction of the mangrove peat and therefore, do not affect the chronology or sea-level interpretation as discussed by Gischler (2006). If compaction was significant at this site, the bulk density would increase with depth and time (Bird et al., 2004).

The pollen data from zone BT-79 II show that *R. mangle* consistently dominated the ecosystem and pollen concentrations were relatively high between ~7,300 and 6,900 cal. yrs BP. A denser zone in the lithology, and relatively high MS values were also observed between ~7,300 and 6,900 cal. yrs BP. Higher MS values could be indicating the presence of bacterial magnetosomes, which take up iron minerals, as seen in Brazilian deposits during this time (Berquo et al., 2004). If only two source waters, precipitation and seawater, are assumed to be important for site BT-79, it can be assumed that there was a significant ocean water source at ~7,000 cal. yrs BP to deliver the iron needed for the bacterial magnetosomes. Lacustrine studies of these bacteria have showed greater prevalence in highly organic-rich samples (Kim et al., 2005), which is characteristic of mangrove peat. In core BT-79, the peak MS values are still less than values caused by volcanic eruptions (e.g. Michaud et al., 2006). Therefore, these values are most likely not indicative of erosion of magnetic minerals, as they are at mainland sites (Curtis et al., 1998).

Another feature of zone BT-79 II is the 3‰ shift in the $\delta^{18}O$ of *R. mangle* leaf fragments at ~7,000 cal. yrs BP. A similar shift (~20 to 17.0‰, Wooller et al., 2007) has been documented in a mangrove peat core (TCC2) from Twin Cays, Belize. Wooller et al. (2007) ascribe decreasing $\delta^{18}O$ values to a greater proportion of rainfall than ocean water in mangrove source water. Our analysis of precipitation data and modern *R. mangle* leaves from SLC, in addition to other proxies, supports this hypothesis. For $\delta^{18}O$, we measured an average of 25.6±1.4‰ for modern *R. mangle* leaves and an average of 0.0±1.5‰ for water on SLC. Although we analyzed whole leaves instead of performing the recommended cellulose extraction (Ferrio and Voltas, 2005), we find it encouraging that the overall difference in $\delta^{18}O_{water}$ and $\delta^{18}O_{leaves}$ (~26‰) is consistent with the observed fractionation associated with the synthesis of cellulose (DeNiro and Epstein, 1981; Saurer et al., 1997; McCarroll and Loader, 2004). The $\delta^{18}O_{rain}$ (~-3‰) at SLC was expected to be less than the $\delta^{18}O_{ocean}$ (~0‰). Though our samples are limited to a two-week period during the transition from the wet to dry season, these values are similar to calculated precipitation data based on spatial distributions (Bowen et al., 2005). Our $\delta^{18}O_{water}$ data illustrates that precipitation has a more negative signature than ocean water. Therefore, *R. mangle* may have begun taking up a higher proportion of precipitation relative to ocean water, ~7,000 cal. yrs BP.

Greater precipitation vs. ocean water delivered to the site could be achieved by two different scenarios. First, there may have been a decrease in sea-level or a decrease in the rate of rise of sea-level at this time, with no change in precipitation. Alternatively, there may have been an increase of the amount of precipitation during this time and no change in sea-level inundation. Due to the change in sedimentation rates in core BT-79, we assume that a change in sea-level was more likely. The sedimentation rate is 0.24cm yr^{-1} in zone BT-79 II until ~6,000 cal. yrs BP. Assuming our sedimentation rates record sea-level rise, the 0.15cm yr^{-1} rate proposed by Toscano and Macintyre (2003) is relatively comparable. However, they witness a consistent sea-level rise from 7,700 until 2,000 cal. yrs BP, which does not correspond with our observation of a significant decrease in the sedimentation rate to 0.02cm yr^{-1} at ~6,000 cal. yr BP. Therefore, it is possible that there was a fluctuation in sea-level rise that is not shown by the data presented by Toscano and Macintyre (2003).

BT-79 Zone III: 3,680 to 320 cal. yrs BP

A marked decrease in leaf area of subfossil *R. mangle* leaves is observed at the base of BT-79 III, ~4,000 cal. yrs BP. Loss of leaf fragments and higher abundances of *Myrsine*, herbs and grasses, and foraminifera suggest a significant overall change in the ecosystem, suggesting significant stress on *R. mangle.* Below the lower boundary of zone BT-79 III there is a slight increase in *A. germinans* pollen and the maximum abundance of Amaranthaceae/Chenopodiaceae. This is potentially due to an increase in *Salicornia bigelovii* Torr., a chenopod, which is often found in association with *A. germinans*. Based on these observations, it appears that site BT-79 was experiencing a less *R. mangle* dominated system from ~3,600 to 600 cal. yrs BP. Research on Twin Cays, Belize shows a similar change in vegetation at ~3,500 cal. yrs BP. Cores TCC1

(Wooller et al., 2004) and TCC2 (Wooller et al., 2007) record significant abundances of *Myrsine*-type pollen at this time, suggesting a less saline environment that allowed non-halophytic plants to flourish. Presently, *Myrsine* sp. are neither found on Twin Cays nor Spanish Lookout Cay.

Greater inundation of seawater, and perhaps a more tolerable environment for mangrove species at site BT-79, resumes when the sedimentation rate increases to 0.39 cm yr^{-1} at approximately 600 cal. yrs BP in zone BT-79 III. Also at ~600 cal. yrs BP, leaf area of subfossil *R. mangle* leaves increases and some of the most marked changes in the stable isotopic values of these leaves are observed. Relative to the other zones, exceedingly high (>6‰) $\delta^{15}N$ values are seen from ~600 to 500 cal. yrs BP in zone BT-79 II. Such highly enriched nitrogen isotope values have previously been observed in Belize in locations receiving bird guano (Wooller et al., 2003a). These types of $\delta^{15}N$ values in plants have also been recorded in various other geographic locations receiving animal waste (e.g. Erskine et al., 1998). Therefore, we propose that a bird rookery was established near site BT-79, at ~600 cal. yrs BP, delivering enriched nitrogen to the plants in the form of bird guano.

At ~600 cal. yrs BP, there is also a 6‰ increase in the $\delta^{18}O$ values of subfossil *R. mangle* leaves. If mangrove source water is the driving force behind the $\delta^{18}O$ of the bulk leaf, then it is assumed that precipitation was no longer the dominant contribution and seawater inundation had increased. Seawater has greater concentrations of nutrients than precipitation that *R. mangle* requires, thus aiding their growth; presence of predominantly tall *R. mangle* at site BT-79, at ~600 cal. yrs BP is indicative by more depleted (<-27‰)

δ^{13}C values. Stimulated mangrove growth at this time may have made the site an attractive roost for marine birds. We conclude that the site experienced less inundation with seawater and the mangroves were under stress from the base of zone BT-79 III until ~600 cal. yrs BP. After this point, accumulation of peat increases, site BT-79 has tall *R. mangle*, and there is greater inundation of seawater.

BT-79 Zone IV: 300 cal. yrs BP to present

R. mangle is highly abundant throughout zone BT-79 IV, as demonstrated by very high leaf areas. Core TCC1 from Twin Cays (Wooller et al., 2004) demonstrate higher leaf areas near the base of the core; therefore, we do not believe that depth or age is correlated with leaf area. A significant amount of pollen accumulated from other shrubs and trees, particularly *Pinus*, in zone BT-79 IV. The increase in abundance of pollen from non-mangrove shrubs and trees observed at SLC is attributed to a greater influx of pollen from mainland Belize. It is not likely that these plants were present at the site in the last 300 yrs, because we did not see any evidence (e.g. pine cones, tree stumps) on the island (personal observation). It is possible that there were more favorable conditions for pollen transport, such as greater prevailing winds, during this time.

At the base of zone BT-79 IV, low δ^{13}C values denote the presence of a tall mangrove stand. However, δ^{13}C values generally increase to the top of the zone, and we infer a transition to a dwarf stand, which is seen at the site today, throughout the last 100 yrs. Therefore, it may be likely that the transition of the site from a tall to dwarf *R.*

mangle stand in the last 300 yrs produced less pollen, which would accommodate the greater abundance of non-mangrove shrubs and tree pollen seen throughout the zone. Low local pollen production can exist in environments of low productivity (e.g. a dwarf *R. mangle* stand) (Traverse, 1988).

Relatively high $\delta^{18}O$ values (>20%) are observed throughout zone BT-79 IV. We concluded earlier that this signal was associated with a higher proportion of ocean water source relative to precipitation. Greater inundation with ocean water may account for the *R. mangle* leaves with relatively low $\delta^{13}C$ values, which denotes the presence of a tall mangrove stand at the base of zone BT-79 IV. However, we believe that this site was in transition for the last 300 yrs. Currently, the site has less frequent inundation with seawater and a dwarf *R. mangle* stand is present, but we do not see the expected decrease in $\delta^{18}O$ values. Therefore, it is assumed that the BT-79 site has only recently (<100 yrs) become a dwarf *R. mangle* stand.

Summary

It appears that the development of a mangrove-dominated habitat began at approximately the same time on SLC and Twin Cays, ~8,000 cal. yrs BP, when sea-level was ~8m lower than present. Analyses of $\delta^{18}O$ support the use of this proxy to further understand changes in mangrove source water. A decrease in $\delta^{18}O$, pollen concentration, and MS values and an increase in $\delta^{13}C$ at ~7,000 cal. yrs BP all suggest the beginning of a decrease in seawater inundation at the BT-79 site. Interestingly, the Twin Cays sites record similar evidence (e.g. dwarf *R. mangle* stands and low $\delta^{18}O$ values) during the same time period (Wooller et al., 2007), suggesting that this environmental change was regional.

A high abundance (>20%) of *Myrsine*-type pollen was observed on Twin Cays ~4,000 cal. yrs BP and was correlated with decreased inundation of seawater to the site of core TCC1 (Wooller et al., 2004). We expected to see a similar *Myrsine* signal at site BT-79 during the interval of decreased sedimentation rates from ~6,000 to 1,000 cal. yrs BP, but this was not evident. SLC core BT-79 did not yield the high percentages of *Myrsine*-type pollen as seen in Twin Cays, but moderately elevated values (<20%) were seen from ~2,800 cal. yrs BP to the present. We therefore conclude that the prominent *Myrsine* signal seen on Twin Cays was a local ecosystem response to regional change in seawater inundation. At ~1,000 cal. yrs BP, we observed that there was an increase of inundation of seawater to the site based on increased leaf area and $\delta^{18}O$ values. As a result, peat accumulation rates increase and we assume that the mangrove stand at the site was more productive as a result of an increased rate of sea-level rise.

Conclusions

These data show the response of a mangrove ecosystem to environmental change during the past ~8,000 yrs. Sea-level studies throughout the Caribbean region demonstrate the significance of environmental change prior to 7,000 cal. yrs BP. Several curves (e.g. Lighty et al., 1982; Toscano and Macintyre, 2003; Gischler and Hudson,

2004) show an inflection point in sea-level rise at ~7,700 cal. yrs BP when the rate of sea-level rise is proposed to have decreased. The data used to create these sea-level curves are from Belize and other areas and suggest that the change was regional. Core BT-79 showed a major environmental change after 7,000 cal. yrs BP, causing sedimentation rates to decrease dramatically between ~6,000 and 1,000 cal. yrs BP. It is possible that SLC witnessed a "stepped" sea-level change as presented by Blanchon (2005). Only coarse temporal resolution is available in BT-79 during this time period due to greatly reduced sedimentation rates; however, fluctuating sea-level in this area is inferred rather than smoothly rising sea-level due to the significant environmental change seen through multiple, independent proxies of the mangrove ecosystem at site BT-79.

We have shown the local response of a mangrove habitat to regional environmental change can be different. For example, there was a high influx of *Myrsine*-type pollen witnessed on Twin Cays during decreased inundation with seawater at ~3,000 cal. yrs BP. At this time on SLC, we also observe decreased inundation based on low peat accumulation rates and a decrease in *R. mangle* productivity seen by low leaf area, but a dominant *Myrsine* habitat did not develop. Our study has used a multi-proxy approach to understand sedimentation, vegetation, and stand structure to identify events seen in other mangrove habitats in Belize. Further comparison with other islands and mainland Belize will aid in interpreting regional changes and provide more information for predictions of future habitat change.

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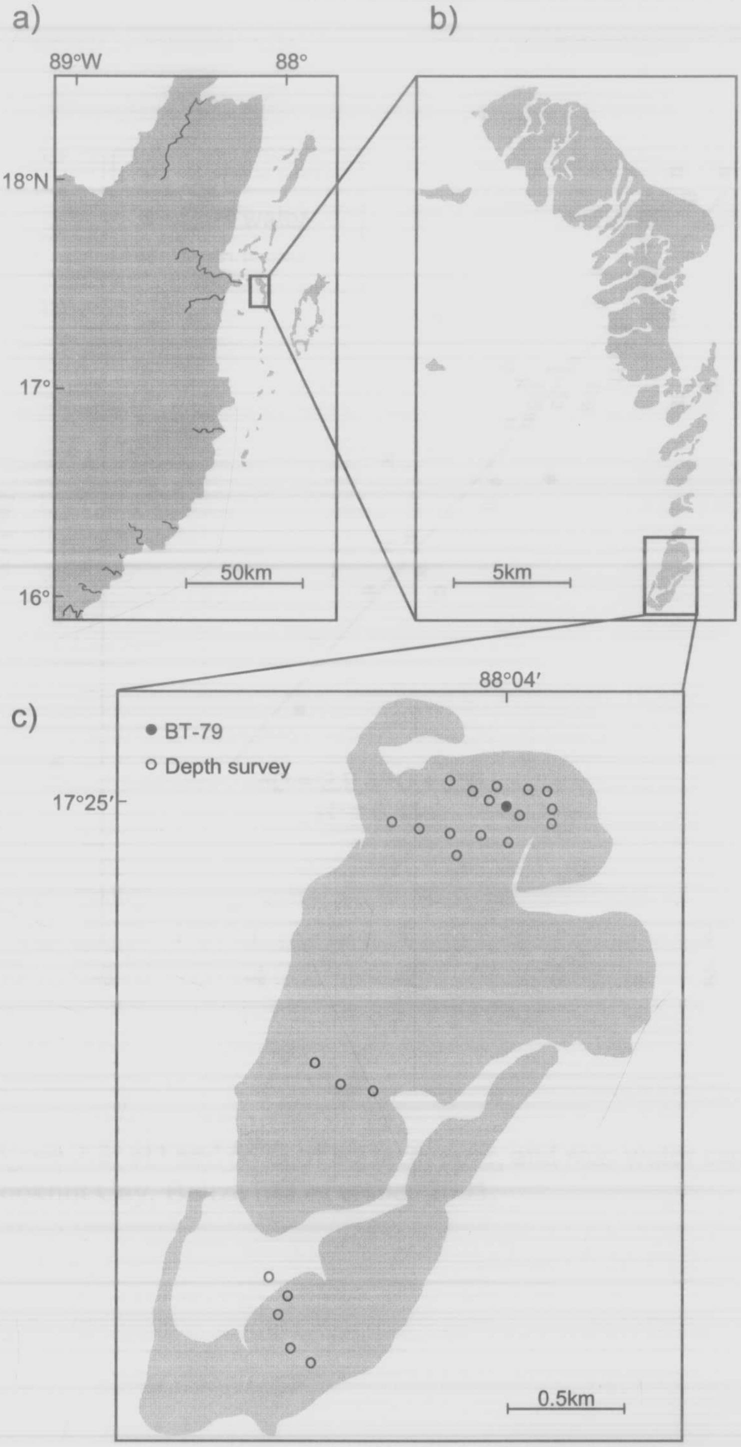

Figure 2.1: a) Belize, Central America. b) Drowned Cays archipelago located offshore mainland Belize. c) Coring sites on Spanish Lookout Cay, Belize.

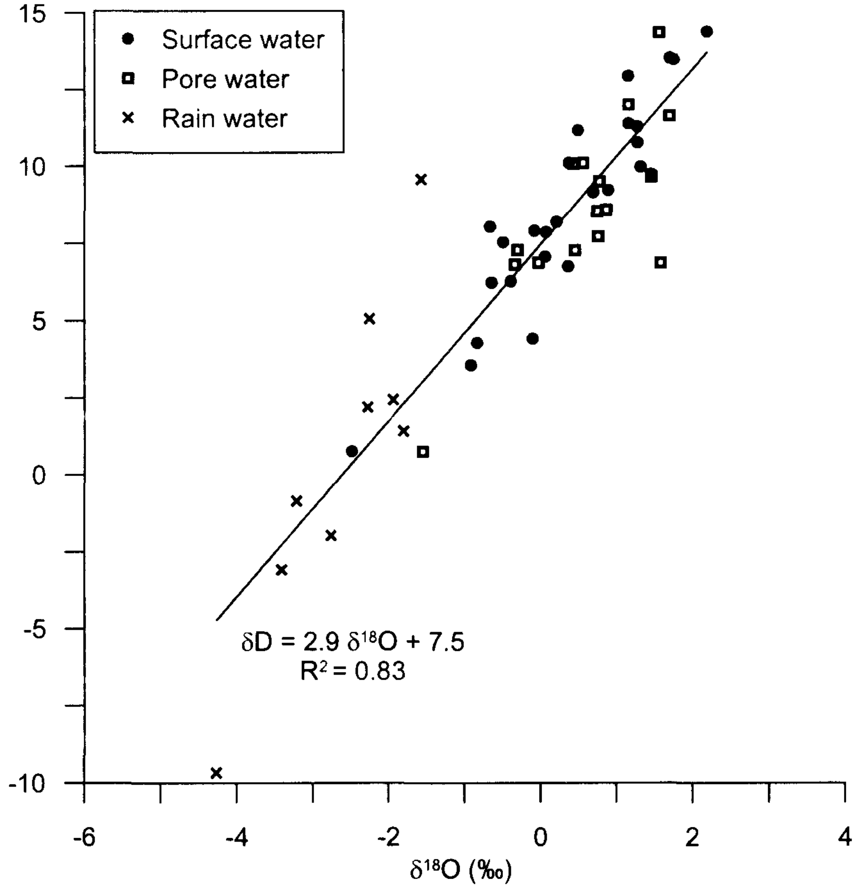

Figure 2.2: δD and $\delta^{18}O$ of surface, pore, and rain water samples taken at Spanish Lookout Cay, Belize in December 2005.

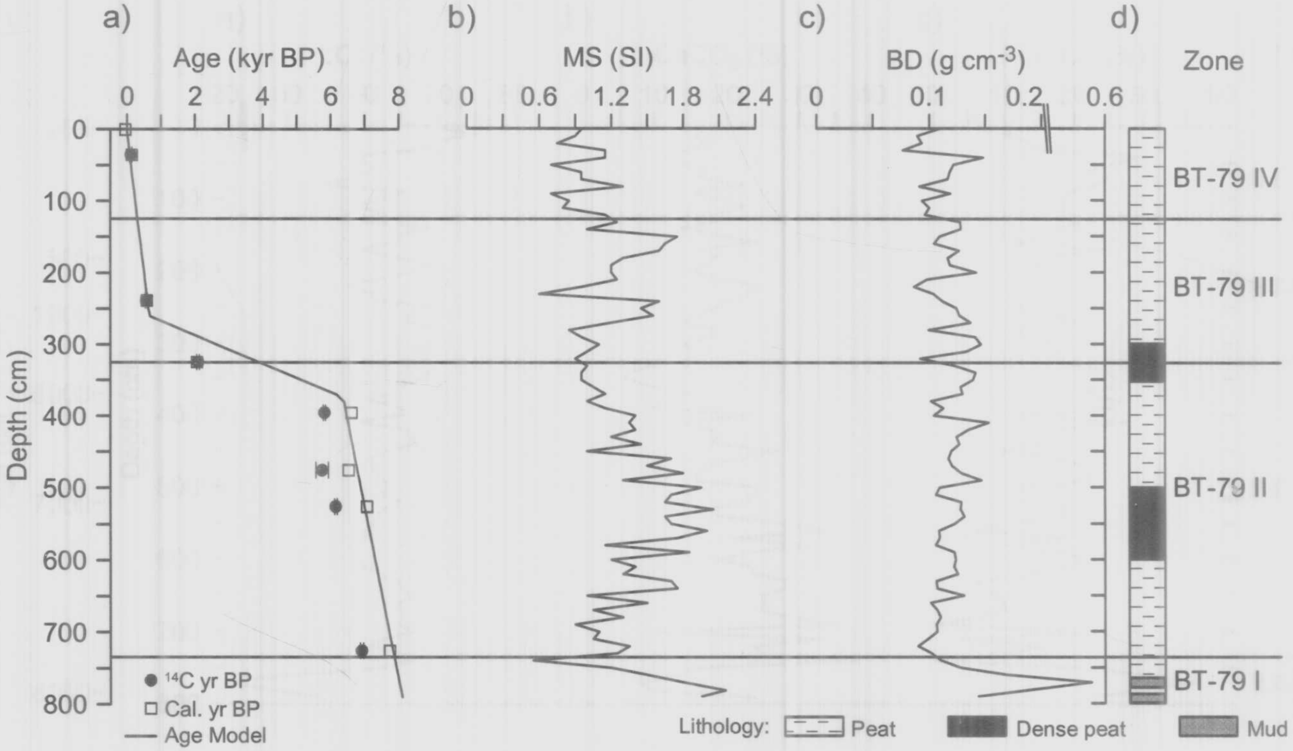

Figure 2.3: a) Radiocarbon ages (^{14}C yr BP), calibrated ages (cal. yrs BP), and age model (using equations 1-3). Error bars were calculated by NOSAMS at WHOI. b) Magnetic susceptibility (MS, SI units). c) Dry bulk density (BD, g cm^{-3}). d) Lithology based on notes taken in the field during coring.

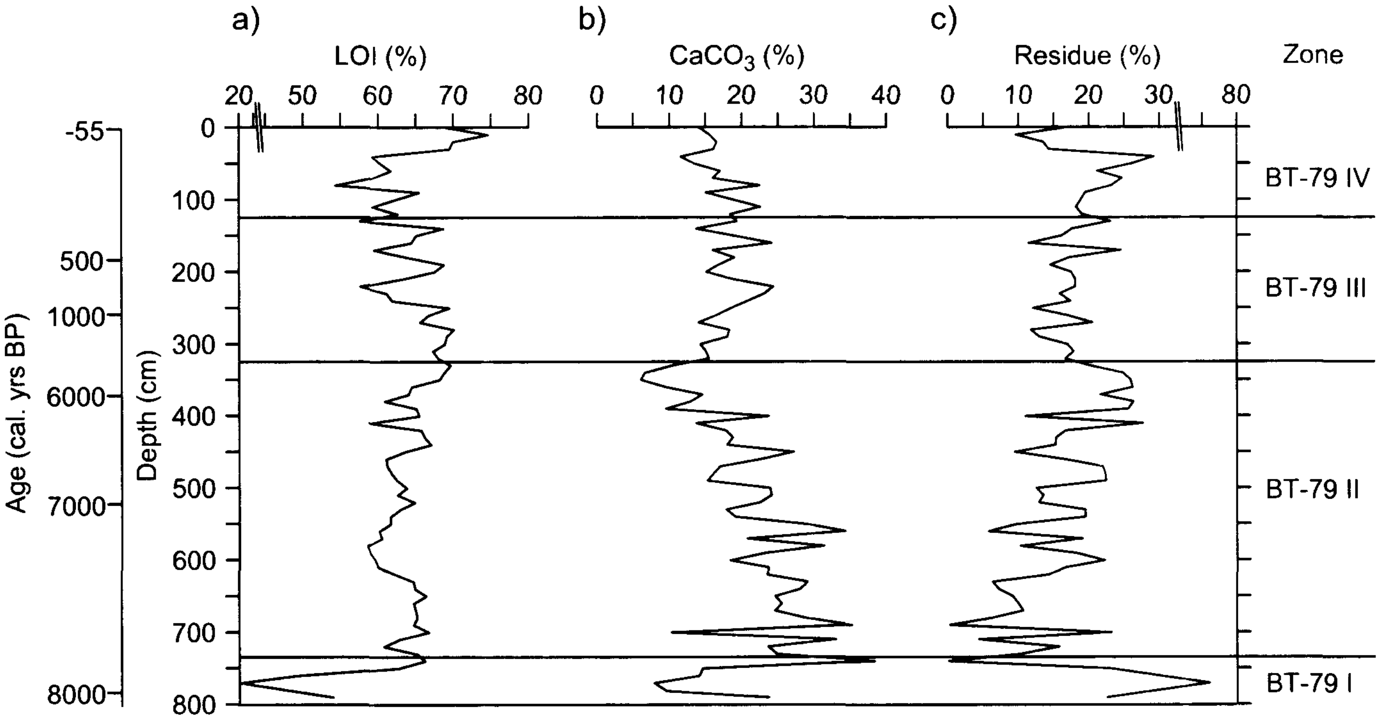

Figure 2.4: Loss on ignition sequence: a) Loss on ignition (LOI, %). b) Calcium carbonate ($CaCO_3$, %). c) Remaining material (Residue, %).

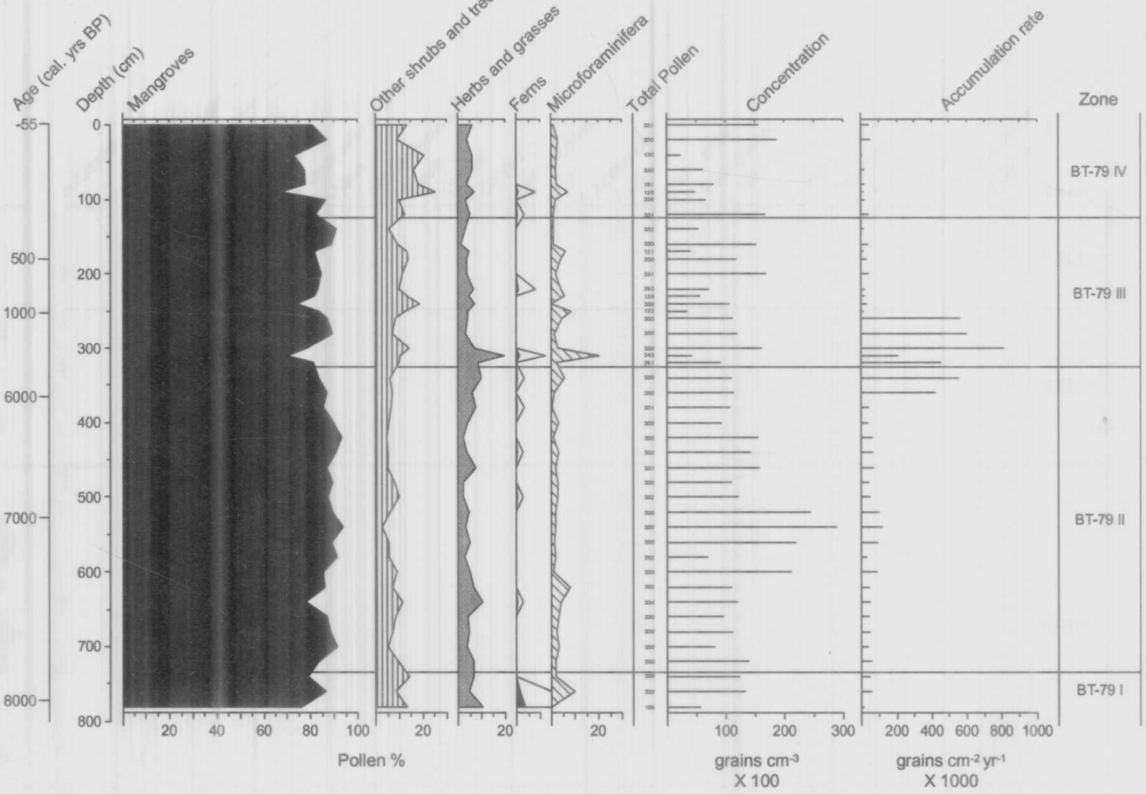

Figure 2.5: Summary pollen diagram showing vegetation groups, pollen concentration, pollen accumulation rate, and zones. Fern percentages are illustrated with a double line: the hollow line is exaggerated by a factor of 10.

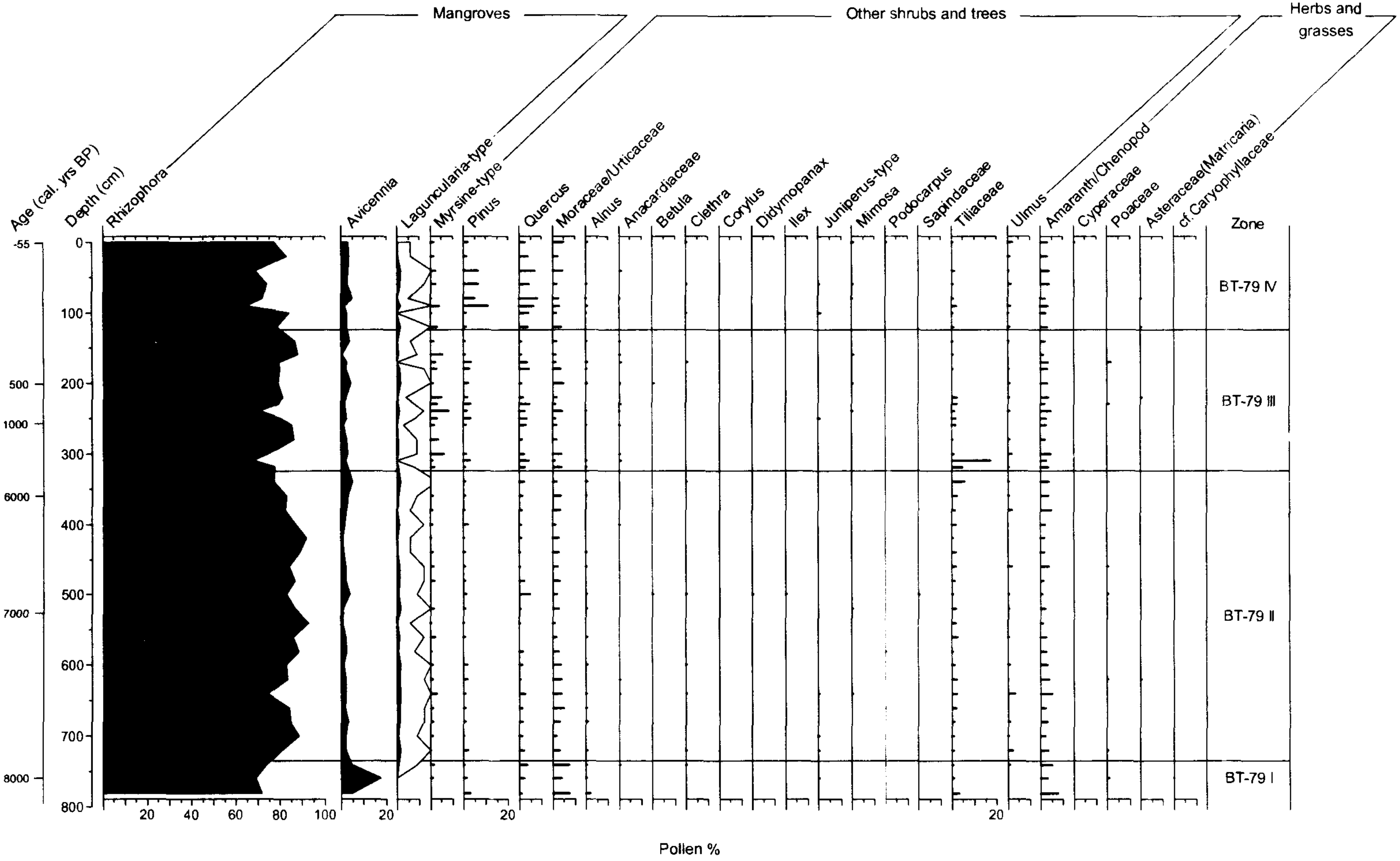

Figure 2.6: Percentage pollen diagram showing the most frequent pollen taxa. Laguncularia-type percentages are illustrated with a double line: the hollow line is exaggerated by a factor of 10.

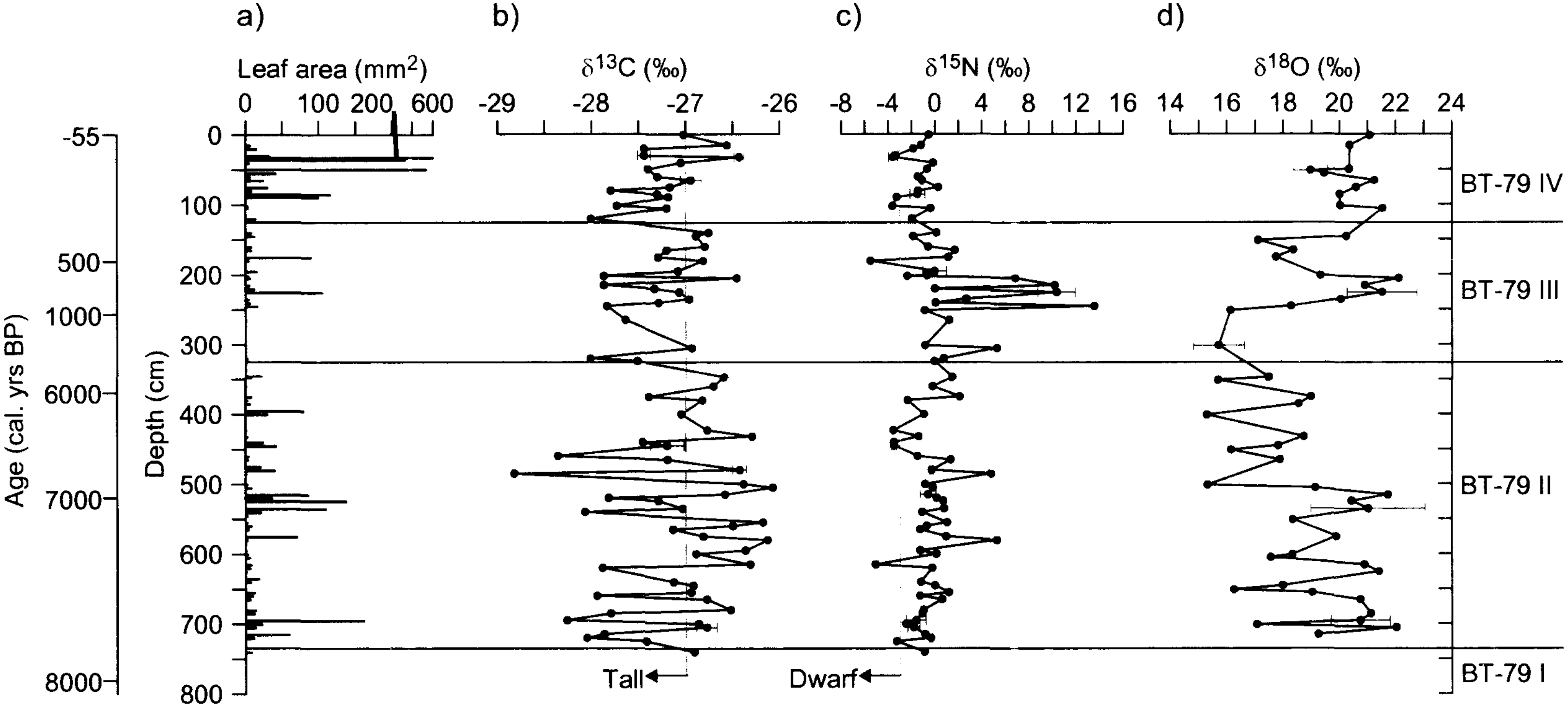

Figure 2.7: Subfossil *R. mangle* leaf data: a) Leaf area (mm^2). b) $\delta^{13}C$ (‰) vs. VPDB. c) $\delta^{15}N$ (‰) vs. air. d) $\delta^{18}O$ (‰) vs. VSMOW. Error bars show standard deviations of triplicate analyses. Tall and dwarf limits were taken from Wooller et al., 2004.

Table 2.1: List of AMS radiocarbon dates and calibrated dates using CALIB (Reimer et al., 2004). *Not included in age model.

Depth (cm)	Sample Type	^{14}C age (yrs BP)	Cal. age (cal. yrs BP)	Cal. range (cal. yrs BP)	NOSAMS number
35-36	leaves	105 ± 30	140	32-256	OS-58900
126-127*	roots	1,010 ± 30	940	918-959	OS-55440
238-239	leaves	560 ± 30	580	535-627	OS-58742
324-325	roots	2,090 ± 35	2060	2,003-2,114	OS-55648
395-396	leaves	5,820 ± 35	6620	6,565-6,672	OS-58890
475-476	leaves	5,750 ± 180	6530	6,322-6,742	OS-59240
525-526	wood	6,150 ± 35	7070	6,992-7,156	OS-55416
635-636*	leaves	4,270 ± 280	4870	4,446-5,286	OS-59247
725-726	roots	6,910 ± 50	7740	7,685-7,788	OS-55649

References

Angulo, R.J. and Lessa, G.C., 1997. The Brazilian sea-level curves: a critical review with emphasis on the curves from the Paranagua and Cananeia regions. Marine Geology 140(1-2), 141-166.

Angulo, R.J., Lessa, G.C. and de Souza, M.C., 2006. A critical review of mid- to late-Holocene sea-level fluctuations on the eastern Brazilian coastline. Quaternary Science Reviews 25(5-6), 486-506.

Behling, H, Costa, M.C.L., Lara, R.J., 2001. Studies on Holocene mangrove ecosystem dynamics of the Braganca Peninsula in north-eastern Para, Brazil. Palaeogeography, Palaeoclimatology, Palaeoecology 167, 225-242.

Berquo, T.S., Thompson, R. and Partiti, C.S.M., 2004. Magnetic study of Brazilian peats from Sao Paulo state. Geoderma 118(3-4), 233-243.

Blanchon, P., 2005. Comments on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. Coral Reefs 24, 183-186.

Bird, M.I., Fifield, L.K., Chua, S. and Goh, B., 2004. Calculating sediment compaction for radiocarbon dating of intertidal sediments. Radiocarbon 46(1), 421-435.

Bowen, G.J., Wassenaar, L.I. and Hobson, K.A., 2005. Global application of stable hydrogen and oxygen isotopes to wildlife forensics. Oecologia 143(3), 337-348.

Curtis, J.H., Brenner, M., Hodell, D.A., Balser, R.A., Islebe, G.A., and Hooghiemstra, H., 1998. A multi-proxy study of Holocene environmental change in the Maya lowlands of Peten, Guatemala. Journal of Paleolimnology 19(2), 139-159.

Dean, W.E., 1974. Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: Comparison with other methods. Journal of Sedimentary Petrology 44(1), 242-248.

DeNiro, M.J. and Epstein, S., 1981. Isotopic composition of cellulose from aquatic organisms. Geochimica et Cosmochimica Acta 45(10), 1885-1894.

Ellison, J.C., 1996. Pollen evidence of Late Holocene mangrove development in Bermuda. Global Ecology and Biogeography Letters 5(6), 315-326.

Erskine, P.D., Bergstrom, D.M., Schmidt, S., Stewart, G.R., Tweedie, C.E., and Shaw, J.D., 1998. Subantarctic Macquarie Island - a model ecosystem for studying animal-derived nitrogen sources using N-15 natural abundance. Oecologia 117(1-2), 187-193.

Faegri, K. and Iverson, J., 1989. Textbook of Pollen Analysis, 4th edn. Wiley, New York.

Fairbanks, R.G., 1989. A 17,000-Year Glacio-Eustatic Sea-Level Record - Influence of Glacial Melting Rates on the Younger Dryas Event and Deep-Ocean Circulation. Nature 342(6250), 637-642.

Feller, I.C., McKee, K.L., Whigham, D.F. and O'Neill, J.P., 2003. Nitrogen vs. phosphorus limitation across an ecotonal gradient in a mangrove forest. Biogeochemistry 62(2), 145-175.

Ferrio, J.P. and Voltas, J., 2005. Carbon and oxygen isotope ratios in wood constituents of *Pinus halepensis* as indicators of precipitation, temperature and vapour pressure deficit. Tellus Series B-Chemical and Physical Meteorology 57(2), 164-173.

Gischler, E., 2006. Comment on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. Coral Reefs 22, 257-270 (2003), and their response in Coral Reefs 24, 187-190 (2005). Coral Reefs 25, 273-279.

Gischler, E., Hauser, I., Heinrich, K. and Scheitel, U., 2003. Characterization of depositional environments in isolated carbonate platforms based on benthic foraminifera, Belize, Central America. Palaios 18, 236-255.

Gischler, E. and Hudson, J.H., 2004. Holocene development of the Belize Barrier Reef. Sedimentary Geology 164(3-4), 223-236.

Grimm E., 2004. TILIA and TGView, 2.0.2. Illinois State Museum, Springfield.

Heiri, O., Lotter, A.F. and Lemcke, G., 2001. Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. Journal of Paleolimnology 25(1), 101-110.

Higuera-Gundy, A., Brenner, M., Hodell, D.A., Curtis, J.H., Leyden, B.W. and Binford, M.W., 1999. A 10,300 C-14 yr record of climate and vegetation change from Haiti. Quaternary Research 52(2), 159-170.

Intergovernmental Panel on Climate Change (IPCC), 2001. Coastal zones and marine ecosystems. In: McCarthy, J.J, Canziani, O.F., Leary, N.A., Dokken, D.J and White, K.S. (Eds.), Climate Change 2001, Working group II: Impacts, adaptation, and vulnerability. Cambridge University Press, UK. pp 1000.

Kearney, M.S., 2001. Late Holocene sea level variations. In: Douglas, B.C., Kearney, M.S., Leatherman, S.P. (Eds.) Sea Level Rise: History and Consequences. Academic Press, San Diego, CA, pp. 13-36.

Kim, B.Y., Kodama, K.P. and Moeller, R.E., 2005. Bacterial magnetite produced in water column dominates lake sediment mineral magnetism: Lake Ely, USA. Geophysical Journal International 163(1), 26-37.

Kjerfve, B., Ruetzler, K. and Kierspe, G.H., 1982. Tides at Carrie Bow Cay, Belize. In: K. Ruetzler and I.G. Macintyre (Eds.), Smithsonian Contributions to the Marine Sciences. Smithsonian Institution Press, Washington, pp. 539.

Lighty, R.G., Macintyre, I.G. and Stuckenrath, R., 1982. *Acropora palmata* reef framework: a reliable indicator of sea-level in the western Atlantic for the past 10,000 years. Coral Reefs 1, 125–130.

Lovelock, C.E., Ball, M.C., Choat, B., Engelbrecht, B.M.J., Holbrook, N.M. and Feller, I.C., 2006. Linking physiological processes with mangrove forest structure: phosphorus deficiency limits canopy development, hydraulic conductivity and photosynthetic carbon gain in dwarf *Rhizophora mangle*. Plant Cell and Environment 29(5), 793-802.

Macintyre, I.G., Littler, M.M. and Littler, D.S., 1995. Holocene history of Tobacco Range, Belize, Central America. Atoll Research Bulletin 430.

McCarroll, D. and Loader, N.J., 2004. Stable isotopes in tree rings. Quaternary Science Reviews 23(7-8), 771-801.

McKee, K. and Faulkner, P., 2000. Mangrove peat analysis and reconstruction of vegetation history at the Pelican Cays, Belize. Atoll Research Bulletin of Marine Science 468, 47-60.

McKee, K.L., Feller, I.C., Popp, M. and Wanek, W., 2002. Mangrove isotopic (delta N-15 and delta C-13) fractionation across a nitrogen vs. phosphorus limitation gradient. Ecology 83(4), 1065-1075.

Meerman, J. and Sabido, W., 2001. Central American Ecosystems Map: Belize, Volume II: Ecosystem map and descriptions. Belize City: Programme for Belize, pp. 88.

Michaud, F., Ramirez-Sanchez, H. U., Parron, C., Zarate-del Valle, P. F., Fernex, F. and Barci-Funel, G., 2006. Strong magnetic levels in Lake Chapala sediments (western Mexico): Their mineralogy and stratigraphic significance. Journal of Paleolimnology 35, 819-836.

Murray, M.R., Zisman, S.A., Furley, P.A., Munro, D.M., Gibson, J., Ratter, J., Bridgewater, S., Minty, C.D., and Place, C.J., 2003. The mangroves of Belize Part 1. distribution, composition and classification. Forest Ecology and Management 174(1-3), 265-279.

O'Brien, D.M. and Wooller, M.J., 2007. Tracking human migration using stable oxygen and hydrogen isotope analyses of hair and urine. Rapid Communications in Mass Spectrometry 21, 2422-2430.

Palacios-Chavez, R., Ludlow-Wiechers, B. and Villanueva, R., 1991. Flora Palinologica de la Reserva de la Biosfera de Sian Ka'an, Quintana Roo, Mexico, 1st edn. CIQRO: Chetumal.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hogg, A.G., Hughen, K. A., Kromer, B., McCormac, G., Manning, S., Ramsey, C. B., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., van der Plicht, J., and Weyhenmeyer, C.E., 2004. IntCal04 terrestrial radiocarbon age calibration, 0-26 cal kyr BP. Radiocarbon 46(3), 1029-1058.

Rull, V., Vegas-Vilarrubia, T. and de Pernia, N.E., 1999. Palynological record of an early-mid Holocene mangrove in eastern Venezuela. Implications for sea-level rise and disturbance history. Journal of Coastal Research 15(2), 496-504.

Saurer, M., Aellen, K. and Siegwolf, R., 1997. Correlating delta C-13 and delta O-18 in cellulose of trees. Plant Cell and Environment 20(12), 1543-1550.

Sobrado, M.A., 2000. Relation of water transport to leaf gas exchange properties in three mangrove species. Trees 14, 258-262.

Sobrado, M.A., 2001. Hydraulic properties of a mangrove *Avicennia germinans* as affected by NaCl. Biologia Plantarum 44(3), 435-438.

Sobrado, M.A., 2005. Leaf characteristics and gas exchange of the mangrove *Laguncularia racemosa* as affected by salinity. Photosynthetica 43(2), 217-221.

Toscano, M.A. and Macintyre, I.G., 2003. Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat. Coral Reefs 22(3), 257-270.

Traverse, A., 1988. Paleopalynology. Unwin Hyman, Boston, pp. 600.

Vannucci, M., 2001. What is so special about mangroves? Brazilian Journal of Biology 61(4), 599-603.

Vedel, V., Behling, H., Cohen, M. and Lara, R., 2006. Holocene mangrove dynamics and sea-level changes in northern Brazil, inferences from the Taperebal core in northeastern Para State. Vegetation History and Archaeobotany 15(2), 115-123.

Wooller, M., Smallwood, B., Jacobson, M. and Fogel, M., 2003a. Carbon and nitrogen stable isotopic variation in *Laguncularia racemosa* (L.) (white mangrove) from Florida and Belize: implications for trophic level studies. Hydrobiologia 499(1-3), 13-23.

Wooller, M., Smallwood, B., Scharler, U., Jacobson, M. and Fogel, M., 2003b. A taphonomic study of δ^{13}C and δ^{15}N values in *Rhizophora mangle* leaves for a multi-proxy approach to mangrove palaeoecology. Organic Geochemistry 34, 1259-1475.

Wooller, M.J., Behling, H., Smallwood, B.J. and Fogel, M., 2004. Mangrove ecosystem dynamics and elemental cycling at Twin Cays, Belize, during the Holocene. Journal of Quaternary Science 19(7), 703-711.

Wooller, M.J., Morgan, R., Fowell, S.J., Behling, H. and Fogel, M., 2007. A multi-proxy peat record of Holocene mangrove paleoecology from Twin Cays, Belize. Holocene 17(8), 1129-1139.

Chapter 3

Paleoecological changes of a mangrove ecosystem during the Holocene along the Sibun River, Belize[1]

Abstract

The specialized ecology and physiology of mangrove ecosystems have allowed them to exist in a constantly changing environment throughout the Holocene. Previous paleoecological studies of mangrove ecosystems in Belize, Central America have largely been restricted to the cays offshore. This study examines a mangrove ecosystem along the Sibun River, Belize (core SR-63), which is subject to both changes in sea-level and in the characteristics of the drainage basin. Radiocarbon measurements show a decreased sedimentation rate from ~6,000 to 1,000 cal. yrs BP and a dramatic change in lithology from primarily mangrove peat to fluvial-derived material at ~2,500 cal. yrs BP. Pollen analyses show a decreased abundance of *Rhizophora* pollen and an increased abundance of *Avicennia* pollen and non-mangrove pollen coeval with the decreased sedimentation rates. Elemental ratios ($[N:C]_a$) and stable isotope analyses ($\delta^{13}C$ and $\delta^{15}N$) show changes in the composition of the organic material are coeval with the change in lithology. We attribute changes in the sedimentation rates and pollen assemblage at site SR-63 to changes in sea-level inundation, which is similarly observed in paleoecological

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[1] Monacci, N.M., Meier-Grünhagen, U., Finney, B.P., Behling, H., Wooller, M.J., in prep. Paleoecological changes of a mangrove ecosystem during the Holocene along the Sibun River, Belize. Prepared for submission to Quaternary Research.

studies offshore Belize. We compare the changes in lithology to paleolimnological and anthropological studies conducted in areas with previous Mayan occupation.

Introduction

Mangrove ecosystem research has shown the importance of this dynamic habitat as a food supply for biological production (e.g. Lalli and Parsons, 1993) and as a protective barrier against erosion during storm or flood events (e.g. Danielsen et al., 2005; Kathiresan and Rajendran, 2005). Mangroves are highly adaptive plants tolerant of extreme environmental conditions such as high salinity, anoxia, and constant inundation (Vannucci, 2001). This adaptability has allowed mangroves to withstand environmental change throughout the Holocene; evidence of this is recorded in the peat they produce (Monacci et al., in prep.).

Mangrove ecosystems present along coastlines are subject to changes in the marine environment (e.g. sea-level change and salinity) and climatic changes (e.g. changes in precipitation) (Alongi, 2002). Studies throughout Central and South America have shown that sea-level has been rising throughout the Holocene (e.g. Fairbanks, 1989; Angulo and Lessa, 1997; Gischler, 2003). Published sea-level curves incorporate data from mangrove peat and *Acropora palmata* (Elkhorn Coral) (e.g. Lighty et al., 1978; Toscano and Macintyre, 2003). Analyses of paleoecological changes with respect to Holocene sea-level changes have been conducted on the offshore, low-lying mangrove islands of Twin Cays (Wooller et. al, 2004 and 2007) and Spanish Lookout Cay (Monacci et al., in prep.) to determine past habitat changes. However, changes in sea-level could also have influenced mangrove ecosystems on the mainland. There are few stratigraphic records from coastal Belize; in the recent past, paleoenvironmental studies have focused on lacustrine environments (e.g. Jacob and Hallmark, 1996). Several limnological studies in Central America have shown that lake levels and productivity have been affected by changes in sea-level when the lake is connected to a phreatic aquifer (e.g. Curtis et al., 1998; Hodell et al., 2007). In fact, lacustrine deposition only began at Lake Peten-Itza in Peten, Guatemala after 8,000 calibrated years before present (cal. yrs BP), which Curtis et al. (1998) assumed was related to rising sea-level in the early Holocene. Likewise, Hodell et al. (2007) also observed that the groundwater at Punta Laguna in Quintana Roo, Mexico is controlled by both precipitation and sea-level. These paleoenvironmental studies also relate Holocene climatic changes to observed changes in local Mayan culture (e.g. Leyden, 1987; Hodell et al., 1995; Curtis et al., 1996; Wahl et al., 2006).

The primary goal of this research is to determine whether environmental change, such as the effects of changes in sea-level in offshore mangrove ecosystems, can be directly related to changes in mainland mangrove ecosystem along the banks of the Sibun River, Belize. We also aim to determine whether changes in the mangrove paleoecology and fluvial dynamics of the Sibun River correspond to broad scale climatic changes recorded in upland lake environments. To achieve this, we apply a multi-proxy approach to recover evidence of past environmental change preserved in a sediment core taken from a riverine mangrove habitat in Belize. Radiocarbon measurements are used to determine the timing of observed changes in the core and to correlate these changes with other research. Bulk density, magnetic susceptibility, and loss on ignition analyses of the sediments are used to examine changes in the stratigraphic characteristics of mangrove peat and sediments. Pollen analyses of the core are used to illustrate vegetation dynamics. Finally, atomic nitrogen:carbon ($[N:C]_a$) ratios of the sediments are combined with stable isotope analyses ($\delta^{15}N$ and $\delta^{13}C$) to determine the source of deposited organic matter.

Study Site

This study took place along the Sibun River located within the Cayo and Belize political districts in central Belize (Figure 3.1). Northern Belize is located on the tectonically stable Yucatán platform and is underlain by Tertiary limestone, while the mountainous, southern region is composed of Paleozoic sediments and Cretaceous limestone (Gischler and Hudson, 1998). The Maya Mountains (960m asl) serve as the headwaters for a river that flows northeast from the Sibun Gorge through a riparian forest, karst limestone belt, coastal marsh, and mangroves to terminate in the Caribbean Sea (Achee et al., 2006).

Belize is located in the transition zone from the tropics to subtropics and experiences a dry season from February to May (Meerman and Sabido, 2001). Mean annual rainfall in Belmopan (Figure 3.1) is 2020mm and the region experiences flooding in the rainy season between July and December (Piaskowski et al., 2006). Approximately

3,000 people currently live in villages along the Sibun River, but many settlements remain undocumented (Achee et al., 2006).

The location of our coring site along the Sibun River, SR-63 (17° 25′ 13.9″N, 88° 16′ 19.7″W, Figure 3.1), is approximately 4km inland of the mouth of the river and 20m from the river edge. The Sibun River watershed has several vegetation types: tropical evergreen seasonal mixed needle forest (e.g. *Pinus* sp.), broadleaf forest (e.g. *Quercus* sp.), mangroves (e.g. *Rhizophora mangle*), and agricultural (e.g. Poaceae, *Citrus* sp.) (Meerman and Sabido, 2001). The canopy height at the coring site is approximately 30m and is primarily composed of *Rhizophora mangle* L. (red mangrove) with interspersed *Avicennia germinans* (L.) Stearn. (black mangrove), *Laguncularia racemosa* (L.) C.F. Gaertn. (white mangrove), and Arecaceae sp. (palm). At the time core SR-63 was taken, Belize City (~10km north) had a flood tide and the Sibun River had a salinity of 0.5‰; however, it is possible that SR-63 is periodically exposed to more saline water, especially during storm events. Belize experiences a semidiurnal micro-tide (15-30cm) (Kjerfve et al., 1982). The Sibun River drains an area of 970km^2. Tributaries are fast and turbulent, giving the main channel a more sluggish, flood prone reach with discharge ranging from 35 to 65m^3 s^{-1} (BERDS, 2007).

Methods

Sampling

An Eijkelkamp Russian peat corer was used to obtain an 8.5m long, continuous core (SR-63) that is composed of mangrove peat and fluvial sediments. During coring, layers with lesser or greater resistance and visual color transitions were noted in the lithology (Figure 3.2). In the field, SR-63 was collected in 17 50cm x 5cm sections; wrapped in plastic, and kept in coolers at the Spanish Bay Conservation and Research Center on Spanish Lookout Cay. The core sections were then transferred to the Water and Environmental Research Center (WERC) at UAF, where they are stored at ~4°C.

Core chronology

Radiocarbon measurements (^{14}C) were made on wood pieces, macroscopic *R. mangle* roots, and bulk peat. Samples were measured using accelerator mass spectrometry (AMS) at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) at Woods Hole Oceanographic Institute (WHOI). All ^{14}C dates were calibrated using CALIB 5.0.2 according to Reimer et al. (2004) and rounded to the nearest decade.

Sediment characteristics

Several geological and stratigraphic analyses were performed on bulk peat samples. A $2cm^3$ sample was taken every 10cm throughout SR-63. Magnetic susceptibility (MS, SI) was measured on the dry sample using a Bartington MS2 sensor. Dry bulk density (BD, g cm^{-3}) was determined by weighing the wet sample, freeze drying, and weighing the dry sample. Sequential loss on ignition (LOI) measurements were then taken following standard procedures (e.g. Dean, 1974; Heiri et al., 2001) to estimate organic carbon content (LOI, %), carbonate content ($CaCO_3$, %), and remaining material (Residue, %).

Pollen analysis

All pollen preparation and analytical procedures were performed in the Paleontology Laboratory in the Department of Geology and Geophysics at UAF. At least every 20cm down core, a $1cm^3$ sample was prepared using the Faegri and Iverson (1989) standard method. One tablet of exotic *Lycopodium* spores (18,583 ± 762 spores $tablet^{-1}$) was added prior to processing in order to calculate pollen concentrations (grains cm^{-3}) and accumulation rate (grains cm^{-2} yr^{-1}). Samples were mounted in glycerin gelatin medium for counting; a minimum of 300 pollen grains were identified for each pollen sample, with a few exceptions that are counted to >100 due to particularly low pollen concentrations. Pollen grains were identified using published pollen morphological descriptions (e.g. Palacios-Chavez et al., 1991). Pollen data are reported as percentages of the total pollen sum, which excludes fern and fungal spores and microforaminifera. Pollen diagrams were created and zoned using TILIA (Grimm, 2004).

Elemental and stable isotope analysis: %N, %C, $\delta^{15}N$, $\delta^{13}C$

Bulk peat samples were freeze-dried, ground, and acidified with 2N HCl to remove carbonate in preparation for elemental and stable isotopic analyses of organic material (OM). Stable isotope analyses were performed at the Alaska Stable Isotope Facility (ASIF) at UAF and reported in standard delta (δ) notation in units of permil (‰). The elemental compositions (%C and %N, wt. %) and stable isotope ratios (δ^{13}C and δ^{15}N, ‰) were measured with an EA-IRMS utilizing a Costech Instruments Elemental Analyzer (EA) coupled with a Conflo III and a Thermo Electron Delta V Plus Isotope Ratio Mass Spectrometer (IRMS). δ^{13}C is reported relative to Vienna Pee Dee Belemnite (VPDB) and δ^{15}N is reported relative to atmospheric nitrogen (atm). Peptone was used as a reference standard to determine analytical precision: 0.9% (wt. %N), 2.8% (wt. %C), 0.5‰ (δ^{15}N) , and 0.3‰ (δ^{13}C).

Results

All data from core SR-63 are presented within the context of three pollen zones: zone SR-63 I (585-850cm), zone SR-63 II (195-585cm), and zone SR-63 III (0-195cm).

These zones were established through a Constrained Incremental Sum of Squares (CONISS) analysis performed in TILIA on the pollen data.

Chronology

Table 3.1 presents the radiocarbon data (^{14}C) and their corresponding calibrated dates and ranges. Ages for all depths of SR-63 and sedimentation rates were interpolated using three linear equations. The top of the core (0-1cm) is assumed to be -55 cal. yrs BP. The bottom section of the core (710-850cm) was fitted and has linear equation:

$$Depth_{cm} = 0.56Age_{cal.\,yrs\,BP} - 10.94, R^2 = 0.84, n = 3 \tag{1}$$

yielding a sedimentation rate of 0.19cm yr^{-1}. The calibrated dates shown in Figure 3.2a were extrapolated to estimate a basal date of ~6,700 cal. yrs BP for SR-63. The middle section (400-700cm) was fitted and has linear equation:

$$Depth_{cm} = 0.06Age_{cal.\,yrs\,BP} + 349.00, R^2 = 0.98, n = 3 \tag{2}$$

yielding a sedimentation rate of 0.06cm yr^{-1}. The top section of the core (0-390cm) was fitted assuming an age of -55 cal. yrs BP (Reimer et al., 2004) for depth 0-1cm, yielding a linear equation:

$$Depth_{cm} = 0.19Age_{cal.\,yrs\,BP} - 437.34, R^2 = 1, n = 2 \tag{3}$$

yielding a sedimentation rate of 0.56cm yr^{-1}. We chose to use multiple linear fits over a comprehensive logarithmic fit to avoid negative sedimentation rates and inverse modeled ages for each depth, which could not be avoided in a variety of curve fitting equations using all the dates.

Sediment characteristics

The lithology of SR-63 is divided into two main parts. The bottom of the core (850-500cm) is primarily composed of mangrove peat with some minor fluvially deposited sediment, while the top of the core (500-15cm) is primarily fluvial sediment with interspersed peat (Figure 3.2d). There is a gray, mud/clay deposit from 150-100cm with little visual evidence of peat. The uppermost 15cm is composed of peat material and leaf litter present from the modern mangrove stand.

Throughout SR-63, magnetic susceptibility (MS) ranges from 0.0 to 12.3 (Figure 3.2b). Zone SR-63 I has relatively constant MS values throughout, with an average of 1.6±0.3. MS values remain <2.5 in zone SR-63 II until reaching a peak value of 12.3 at 290cm. Above this peak, MS values vary between 2.0 and 4.0 until another significant peak of 7.2 is observed at 170cm in zone SR-63 III. MS values are less than 2.5 for the remainder of the core.

BD ranges from 0.20 to 1.31g cm^{-3} (Figure 3.2c). An average value of 0.31±0.10 in zone SR-63 I is greater than the BD observed in mangrove peat with little to no allochthonous material (<0.15g cm^{-3}; Monacci et al., in prep.), but zone SR-63 I is nevertheless the least dense. Throughout zone SR-63 II, BD values steadily increase from a basal value of 0.3g cm^{-3} to a maximum value of 1.31g cm^{-3}. BD generally decreases from 1.07g cm^{-3} at 190cm to 0.35g cm^{-3} at the top of zone SR-63 III.

The LOI sequence from core SR-63 (Figure 3.3) correlates with changes in the lithology. LOI ranges from 6 to 67% with values less than 20% in fluvial sediments (500-15cm). A similar trend is observed in $CaCO_3$ and residue. $CaCO_3$ ranges from 2 to 15%, with values less than 10% in the fluvial section. Percent residue, remaining material at the end of the LOI sequence, ranges from 19 to 91% with the fluvial sediment showing greater than 70% of the fraction as neither organic material nor calcium carbonate. There are differences at the pollen zone boundaries, in addition to the marked lithological difference between the top and bottom of SR-63. Peaks are observed in LOI, $CaCO_3$, and residue in zone SR-63 I at ~750cm until the transition into zone SR-63 II. Another comparable peak is observed from ~560cm, in SR-63 II, to immediately below the transition from peat-dominated to sediment-dominated layers at 500cm.

Pollen analysis

Figure 3.4 illustrates pollen summaries and Figure 3.5 illustrates percentages of common genera. Pollen concentrations range from ~2,000 to 30,000grains cm^{-3} and pollen accumulation rates range from ~9,000 to 190,000grains cm^{-2} yr^{-1}.

SR-63 I: 850-585cm

Pollen zone SR-63 I is dominated by mangroves (70%, Figure 3.4). *R. mangle* is the primary species, with a mean of 50%. *A. germinans* is the only other mangrove species present in the core; no pollen from *L. racemosa* or *Conocarpus erectus* L. (buttonwood mangrove) is observed despite their prominence in Belize (Murray et al., 2003). *A. germinans* is least abundant in zone SR-63 I, accounting for less than 20% of the pollen sum, until a peak of >30% is reached between 660 and 630cm. Pollen from other shrubs and trees (non-mangrove) are rare in SR-63 I, with a mean of 20% throughout the zone. *Myrsine* does not appear to be a significant genus at this site. It is seen throughout the core, but it never accounts for more than 2% of the pollen sum. Other than mangroves, *Pinus* and *Quercus* are the dominant plant types in SR-63 I, with means of 5% and 11%, respectively. Members of the Aracaceae family are present throughout the core in low levels never exceeding 3%. In general, herbs and grasses are consistently present throughout the core, averaging 8%. The highest presence of herbs and grasses (15%) is observed in SR-63 I. This peak is attributed to high amounts of Amaranthaceae/Chenopodiaceae (up to 10%) and Poaceae (up to 11%). Microforaminifera are only present in zone SR-63 I. They comprise 4% of the pollen sum at depth 840cm. Ferns averaged 3% throughout the core and are most abundant (7%) at depth 780cm. Fungal spores are present in most of the core, averaging 16%, with 20% and 385% of the pollen sum at 840 and 780cm, respectively. Due to low pollen concentrations (<4,000grains cm^{-3}) and pollen accumulation rates (<15,000grains cm^{-2} yr^{-1}) until 735cm, only 100grains slide^{-1} were counted from samples at the base of the core.

SR-63 II: 585-195cm

Pollen zone SR-63 II has the lowest sum of mangroves with an average of 53% including the overall minimum value for the core of 27% at 480cm. *R. mangle* percentages decrease significantly at the beginning of SR-63 II to 6% at 480cm, but an average 20% throughout the zone. Pollen from *A. germinans* increases and averages 33% in SR-63 II, with a significant peak from 390 to 330cm. As expected given the decrease in mangrove pollen, the percentage of pollen from other shrubs and trees increases to an average of 39%, with the maximum value of the core observed at 480cm. This peak is most likely due to the increase in *Pinus* pollen in this zone, which averaged 26%. *Quercus* percentages remain approximately constant and average 9% throughout SR-63 II. *Podocarpus* makes a small contribution (average 1%) to the pollen assemblage in this zone, but it is not found in intervals with high percentages of *A. germinans*. Herbs and grasses remain the same as for zone SR-63 1 (8%), although greater inputs of Asteraceae (average 2%, e.g. *Matricaria*) are recorded, with maximum values of 5% at 420 and 315cm throughout SR-63 II. Fungal spores are prominent at 510cm, 15% of the pollen sum. No microforaminifera are evident in this pollen zone. Pollen concentrations are low, 3,000grains cm^{-3}, at the base of SR-63 II and generally increase to the maximum value observed in the core, 30,000grains cm^{-3} at 210cm. Pollen accumulation rates also increase from the base of SR-63 II to a maximum value of 190,000grains cm^{-2} yr^{-1} at 450cm and then decrease for the remainder of the core.

SR-63 III: 195-0cm

The amount of pollen from mangroves increases into pollen zone SR-63 III, while they return to an average of 68%, similar to SR-63 I. Pollen from *R. mangle* increases to an average of 27%, with a significant decrease (6%) at 120cm and a peak zonal value of 55% at the top of the core. *A. germinans* pollen increases from 180 to 90cm, where the maximum percentage of 73% is observed, but then decreases to 4% at the top of the core. Other shrubs and trees decrease to 26%, with a low value at 90cm corresponding to peak abundance of *A. germinans*. *Pinus* and *Quercus* pollen both decrease to averages of 14% and 5%, respectively. *Podocarpus* averages 2%, with a maximum value of 4%. A maximum value of Arecaceae (3%) is recorded at 75cm. Herbs and grasses increase to 10% between 90 and 75cm, and then decrease to 7% to the top of the core. Pollen concentrations average 11,000grains cm^{-3} in SR-63 III and pollen accumulation rates are the lowest in the core with an average of 19,000grains cm^{-2} yr^{-1}.

Stable isotope analysis: $\delta^{15}N$ and $\delta^{13}C$

Stable isotope data and elemental ratios downcore are presented in Figure 3.6. Nitrogen content throughout the entire core is low, ranging from 0.1 to 0.6% with no observable trends (data not shown). Carbon content in the sediments is variable and correlates with the lithology (data not shown). Peat-dominated regions (850-500cm, core surface) contain >10%, whereas sediment-dominated portions of the core (500-15cm) contain <10%. δ^{15}N ranges from -2.7 to 3.0‰ and also generally correlates with the lithology with lower values in peat. Zone SR-63 I has the lowest δ^{15}N values with an increase of nearly 3‰ from the base of the core to the top of the zone at 590cm. The values continue to increase to 2.9‰ in zone SR-63 II and then slightly decrease from 500 to 200cm. In SR-63 III, the δ^{15}N values are relatively constant (1.8±0.8‰).

δ^{13}C ranges from -29.4 to -25.7‰ throughout the entire core, but averages -27.1±0.3‰ from 850 to 500cm. More variation in δ^{13}C values are seen after the lithological change to fluvial sediment at 500cm in zone SR-63 II. There is a slight decrease in δ^{13}C to -28.3‰ until 340cm, followed by an increase to -26.4‰ at 270cm. The transition from zone SR-63 II into zone SR-63 III has the lowest δ^{13}C values (-29.4‰) in the whole core, but then values increase nearly 4‰ to the highest δ^{13}C at 40cm before decreasing again to -28.7‰ at the top of the core.

Atomic elemental ratios of nitrogen to carbon are shown in Figures 3.6 and 3.7. Following the suggestion of Goñi et al. (2003), we chose to express $[N:C]_a$ rather than $[C:N]_a$ due to the low nitrogen composition of these samples. $[N:C]_a$ values range from 0.01 to 0.09mol:mol; however, values greater than 0.02mol:mol were not observed in SR-63 I. At 400cm in SR-63 II, $[N:C]_a$ values begin to increase until the top of the core. Two peaks are recorded, 0.07mol:mol at 230cm and 0.09mol:mol at 140cm. Figure 3.7 shows $[N:C]_a$ versus δ^{13}C values of core samples relative to possible sources of organic matter (OM). This approach, which displays the different signatures of OM, has been used in tropical, fluvial environments such as the Fly River in Papua New Guinea (Goñi et al., 2006) and the Amazon River in Brazil (Hedges et al., 1986). Values from horizons below 500cm depth are tightly clustered. The OM in these samples is most likely derived from C_3 vascular plants, primarily mangroves given the lithology. Intervals at depths above 500cm in SR-63 II are more variable, but appear to record a dominant C_3 source. SR-63 III shows the greatest variability in $[N:C]_a$ and $\delta^{13}C$ sources and may contain a greater portion of soil or algal OM.

Interpretation and Discussion

Zone SR-63 I: 6,780 to 3,930 cal. yrs BP

The site of core SR-63 had a well-established mangrove ecosystem during the middle Holocene; this is evident from the peat that comprised the basal section (850 to 500cm) of the core and the high percentages of mangrove pollen in this interval. This was expected, because sea-level inundated other mainland sites at ~7,000 cal. yrs BP (e.g. Jacob and Hallmark, 1996), providing the slightly marine or brackish habitat needed to maintain mangrove ecosystems. By 7,000 cal. yrs BP, mangrove stands had been established for ~1,000 yrs on mangrove islands offshore Belize, including Twin Cays (Wooller et al., 2004 and 2007) and Spanish Lookout Cay (Monacci et al., in prep.). Likewise, lakes in Guatemala connected to a phreatic aquifer began to fill after 8,000 cal.

yrs BP as a result of a higher sea-level (Curtis et al., 1998). Microforaminifera present at the beginning of zone SR-63 I, also record the influence of marine water during this time.

The first decrease in the rate of sea-level rise from published sea-level curves began at 7,700 cal. yrs BP (Toscano and Macintyre, 2003) and is too old to be responsible for the observed change in sedimentation rates. A dominant feature in zone SR-63 I is the decrease in sedimentation rates (Figure 3.2) from 0.19cm yr^{-1} to 0.06cm yr^{-1} at ~6,000 cal. yrs BP. Sedimentation rates also decreased on Spanish Lookout Cay (Monacci et al., in prep.), at this time or slightly earlier. Blanchon (2005) propose that a smooth sea-level curve for the Caribbean during the Holocene is inaccurate and that a "stepped curve" is more realistic. Based on this interpretation, it is assumed that a decrease in the rate of sea-level rise could be followed by an increase in the rate of sea-level rise. Therefore, we conclude that there was a step in sea-level rise at ~6,000 cal. yrs BP. In contrast, an interpreted flooding event at Cobweb Swamp, Belize, ~60km to the north of site SR-63, was observed between 6,400 and 5,500 cal. yrs. BP (Jacob and Hallmark, 1996) and was associated with a rapidly rising sea-level rather than a decreased rate of sea-level rise.

Discussions of the Caribbean sea-level curve reconstruction for the Holocene stress the need for caution when interpreting radiocarbon dates from mangrove peat. The problem of autocompaction of peat is raised by multiple authors (e.g. Woodroffe, 1988; Bird et al., 2004; Gischler, 2006). However, the BD data from zone SR-63 I do not suggest a problem with autocompaction, because there is no increase with depth and time (Bird et al., 2004). It is also noted that penetration of fine roots and old carbon contamination should be taken into consideration (Gischler, 2006), but by not using bulk samples, problems can be minimized.

The LOI sequence from zone SR-63 I shows a significant change just before the decrease in sedimentation at ~6,000 cal. yrs BP. The material at the bottom of zone SR-63 I (~850-750cm) has a greater biogenic content (organic material and calcium carbonate) than the lithogenic content (residual material, non-carbonate/inorganic). There are no corresponding features seen in the elemental or stable isotope ratios to explain a source change in the organic material at this time. BD and MS data during the change in accumulation rate are also relatively uniform. However, there is relatively high abundance of fern spores observed at site SR-63 prior to the change in accumulation rate (>6,000 cal. yrs BP), indicative of a transitional or disrupted site (Traverse, 1988).

The carbonate content at the bottom of zone SR-63 I is similar to the percentages seen at BT-79 on Spanish Lookout Cay (Monacci et al., in prep.), but values from the top of the zone (~750-600cm) are significantly lower. If it is assumed that no significant changes in precipitation or discharge rates of the river occurred at this time, then perhaps a decrease in the rate of rise of sea-level altered both the sedimentation rate and the composition of the deposited materials, according to the LOI sequence, with no obvious changes in the lithology. Microforaminifera are present in zone SR-63 I prior to ~6,000 cal. yrs BP, but are not observed after this time. If the carbonate primarily indicates the presence of these planktonic organisms, this line of evidence supports decreased intrusion of saline water. This could also account for a change in the productivity of the mangrove stand at that time and the subsequent decrease in peat accumulation rates. Decreased seawater inundation of this site was not sufficient to affect the composition of vegetation types, as mangrove species represent greater than 60% of the pollen with no marked changes in the percentages of pollen from other shrubs and trees (freshwater species).

Zone SR-63 II: 3,930 to 370 cal. yrs BP

The lower boundary of zone SR-63 II is marked by a significant change in pollen types. Overall, there is a decrease in mangrove pollen and an increase of other shrubs or trees, which are assumed to be less salt tolerant than the mangroves. There is a marked decrease in *R. mangle* percentages and an increase in *A. germinans*. Compared to offshore mangrove ecosystems (Wooller et al., 2004 and 2007; Monacci et al., in prep.) and those along the Amazon River (Behling and da Costa, 2001), the high percentages (>20%) of *A. germinans* pollen observed in core SR-63 are greater than expected. *A. germinans* is a low pollen producer (Behling et al., 2001), thus a prolific black mangrove stand must have been in the immediate vicinity of site SR-63 to produce such high percentages (Vedel et al., 2006). *Avicennia* are typically not dominant in the same mainland vegetation zone as *Rhizophora* and are usually found more inland (Murray et al., 2003), in more saline conditions. The general increase in *A. germinans* pollen from ~6,000 to 600 cal. yrs BP closely correlates with a decrease in sedimentation rate. If it is assumed that there was a decrease in the rate of sea-level rise during this period, the SR-63 site would have experienced decreased inundation and may have become more saline through increased evapotranspiration and evaporation. The hydraulic properties of *A.*

germinans include higher water use efficiency, providing an advantage in areas with higher salinities (Sobrado, 2000). This and other physiological differences between *A. germinans* and *R. mangle* can also prohibit *A. germinans* from being an effective competitor in areas with lower salinities (Sobrado, 2000). It is therefore assumed that the high abundance of *A. germinans* indicates high salinity and evaporation. However, it is possible that the increase in salinity could be attributed to a decrease in the rate of sea-level rise, and a change in the dynamics of the Sibun River or its drainage basin, or both.

The transition from basal peat-dominated lithology (850-500cm) to the overlying sediment-dominated lithology (500-0cm) occurs in zone SR-63 II. This change from predominantly peat material to predominantly fluvial material occurs at ~2,500 cal. yrs BP (Figure 3.2). Several proxies reflect this transition: the LOI sequence becomes virtually uniform for the remainder of the core, there is an overall increase in bulk density to the top of zone, indicating more compaction, and there are changes in the elemental and isotopic composition of the organic material. $\delta^{15}N$ becomes fairly uniform for the remainder of the core, there is a slight decrease in $\delta^{13}C$ towards the top of the zone, and there is a slight increase in $[N:C]_a$ at the lithological boundary. These characteristics contrast with data from peat cores taken from offshore environments, which have fairly uniform stratigraphic and geochemical characteristics. For example, $\delta^{13}C$ analyses on bulk peat from a 7.5m continuous mangrove peat core, have shown minimal variation, -27.1±0.5‰ (Monacci et al., unpublished data 2005). Therefore, the majority of the material being deposited at the top of core SR-63 is presumed to be allochthonous, indicative of a regional, rather than local, signal. These sediments can be compared with the "Mayan clay", which has been observed in several limnological studies (e.g. Leyden, 1987; Jacob and Hallmark, 1996).

Mayan clay has been recognized as a thick, clay-rich stratigraphic unit that has been found in the Peten Lake District of Guatemala (e.g. Binford et al., 1987), and can also be identified through analysis of magnetic susceptibility and pollen. Fine materials with high MS and disturbed pollen taxa were produced by human-induced erosion and deforestation at ~2,800 cal. yrs BP (Curtis et al., 1998). The change in the lithology at ~2,500 cal. yrs BP at the SR-63 site suggests that similar anthropogenic mechanisms could have been responsible. Swidden agriculture, a slash and burn technique used by Mayans (e.g. Hodell et al., 2007), may have increased erosion and deforestation. An increase in erosion related to anthropogenic influences coeval with an increase in agricultural pollen (e.g. maize or *Zea mays*) have been observed at sites in Venezuela (e.g. Leyden, 1987). Some agricultural plants used by the Mayans (e.g. maize) are C_4, which have distinct $\delta^{13}C$ signatures from mangroves or other C_3 plants (e.g. Polk et al., 2007). Despite the similarity in lithological characteristics, no significant changes in the percentages of herb and grass pollen types (Figure 3.4). There was also no shift in the $\delta^{13}C$ of the organic material (Figure 3.7) towards values more typical of C_4 plants, such as those used in Mayan agriculture (e.g. maize). Such a shift in $\delta^{13}C$ has been recognized at other Mayan sites (Polk et al., 2007).

Although human-induced erosion may have been responsible for the increased lithogenic material and allochthonous organic matter at ~2,500 cal. yrs BP in core SR-63, it is important to consider other possible environmental changes. For instance, past fluctuations in the position of the Intertropical Convergence Zone (ITCZ) may have resulted in an increase in precipitation and erosion at the site (Haug et al., 2001; Hodell et al., 2007). Haug et al. (2001) described how the ITCZ dynamically links changes in biogenic and terrigenous sediments with precipitation in the Cariaco Basin off Venezuela and effects of variation in the ITCZ can be seen throughout the Caribbean region. Increased precipitation occurs in the Maya Mountains when the ITCZ is in the northern position and decreased precipitation when the ITCZ is in the southern position (Haug et al., 2001; Polk et al., 2007). Research on cave sediments in the Vaca Plateau region, near the headwaters of the Sibun River, have shown evidence of changes in precipitation and more frequent intervals of aridity during the past ~2,600 cal. yrs BP (Polk et al., 2007).

In addition to significant changes in composition and source of material in the middle of zone SR-63 II, the sedimentation rate increases to 0.56cm yr^{-1} between ~1,000 cal. yrs BP and the present. Unfortunately, no wood pieces or subfossil leaves were found in the area of low sedimentation (~6,000 to 1,000 cal. yrs BP) in core SR-63. The two radiocarbon dates reported from this area were analyzed using macroscopic roots (Table 3.1), which are less than optimal for dating compared with leaf fragments (Wooller et al., 2007). However, any error in dates associated with roots would have resulted in relatively young ages, as roots from living trees grow down into peat. This scenario would produce a significant decrease in the sedimentation; however, such errors would bring the change in the lithology of the core at ~2,500 cal. yrs BP closer in age to the subsequent change in sedimentation rates.

Zone SR-63 III: 370 cal. yrs BP to present

Sedimentation rate during zone SR-63 III is constant, as implied by our ages, and the composition of the deposited material is fairly uniform as evident from the LOI sequence (Figure 3.3). The bulk density shows an overall decrease from the beginning of the zone to the top of the core, where values are similar to those in mangrove peat of zone SR-63 I. This indicates that site SR-63 transitioned back to a mangrove-dominated environment, or flood events were less frequent, which is consistent with the habitat present at the site today. Pollen analyses support this interpretation, as mangrove pollen percentages increase during zone SR-63 III. Even though *A. germinans* pollen is still abundant in this zone, *R. mangle* increases significantly. As a result, the pollen from other shrubs and trees, most notably *Pinus*, decreases. This could be caused by a greater inundation of seawater at this site during the last 370 yrs.

The elemental and stable isotope ratio data are variable during zone SR-63 III. This contrasts with the dry bulk density and LOI data, which are relatively constant. $\delta^{13}C$ increases until ~200 cal. yrs BP, followed by an overall decrease to the top of the core, resulting in a slight increase in $[N{:}C]_a$ of the organic matter. Given the relatively high sedimentation rates, data from SR-63 III represent a relatively high resolution record.

Summary

Previous studies have shown the response of aquatic ecosystems such as coral reefs, mangroves, and lacustrine habitats to Holocene climate change in the greater Caribbean region (e.g Woodroffe, 1988; Hodell et al., 2001; Gischler, 2003). Interpretations of factors driving these changes include orbital forcing (e.g. Leyden et al., 1994), sea-level change (e.g. Fairbanks, 1989), variations in precipitation (e.g. Higuera-Gundy et al., 1999), anthropogenic influences (Hodell et al., 2007), often in combination. In order to better understand the driving mechanisms and responses, a greater number of sites need to be examined.

Core SR-63, from a mangrove ecosystem on the banks of the Sibun River, has shown a decrease in sedimentation from ~6,000 to 1,000 cal. yrs BP. A similar decrease is observed in a core of continuous mangrove peat from Spanish Lookout Cay, 15km offshore (Monacci et al., in prep.). Changes in the sedimentation rates observed in mangrove ecosystems offshore have been attributed to changes in relative sea-level and the rate of sea-level rise (Wooller et al., 2007). A coeval decrease in sedimentation rate at the site of SR-63 on the mainland suggests that sea-level changes most likely influenced mainland mangrove ecosystems. Toscano and Macintyre (2003) showed decreases in the rate of sea-level rise at 7,700 and 2,200 cal. yrs BP. We propose that higher resolution fluctuations in the rate of sea-level rise than concluded by some previous studies can be identified through examination of continuous records of mangrove paleoecology.

Conclusions

The core collected at site SR-63, a mangrove ecosystem on mainland Belize, has undergone significant changes through the Holocene. The site was established in the early Holocene, >7,000 cal. yrs BP. A significant decrease in sediment accumulation from ~6,000 to 1,000 cal. yrs BP was most likely related to a decrease in the rate of sea-level rise. There was also a change in the source of material being deposited at the site ~2,500 cal. yrs BP, which was likely a result of changes in the drainage basin of the Sibun River.

By comparing environmental change at sites on the mainland to sites offshore on mangrove islands and paleolimnological studies in Central America, it is possible to examine regional paleoenvironmental changes. It is understood that sea-level has risen from the start of the Holocene to the present day; however, a more in-depth study that focuses on both geographically and physically different settings will provide the most valuable information for future predictions of climate and sea-level change.

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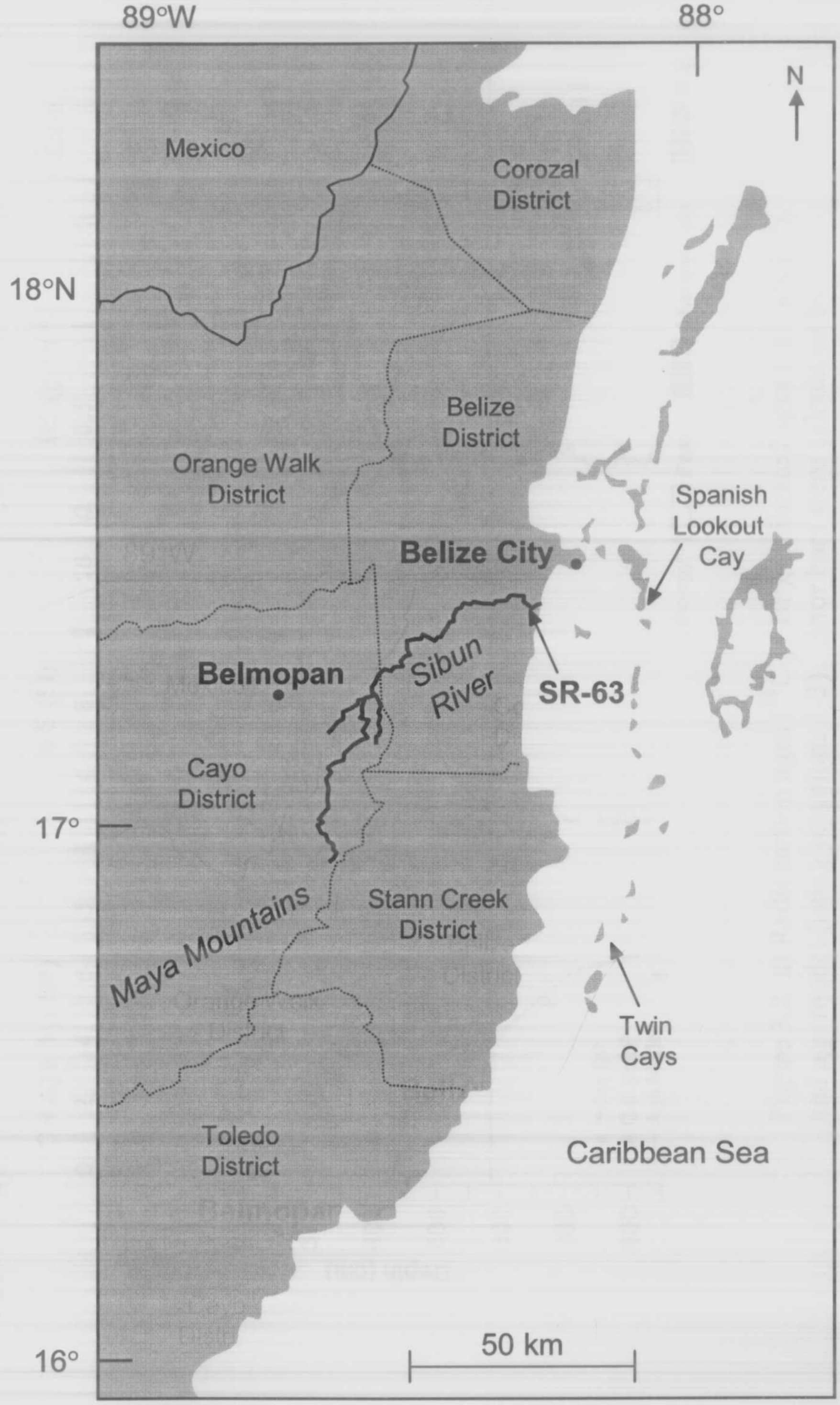

Figure 3.1: Map of Belize, Central America showing core site SR-63 along the Sibun River.

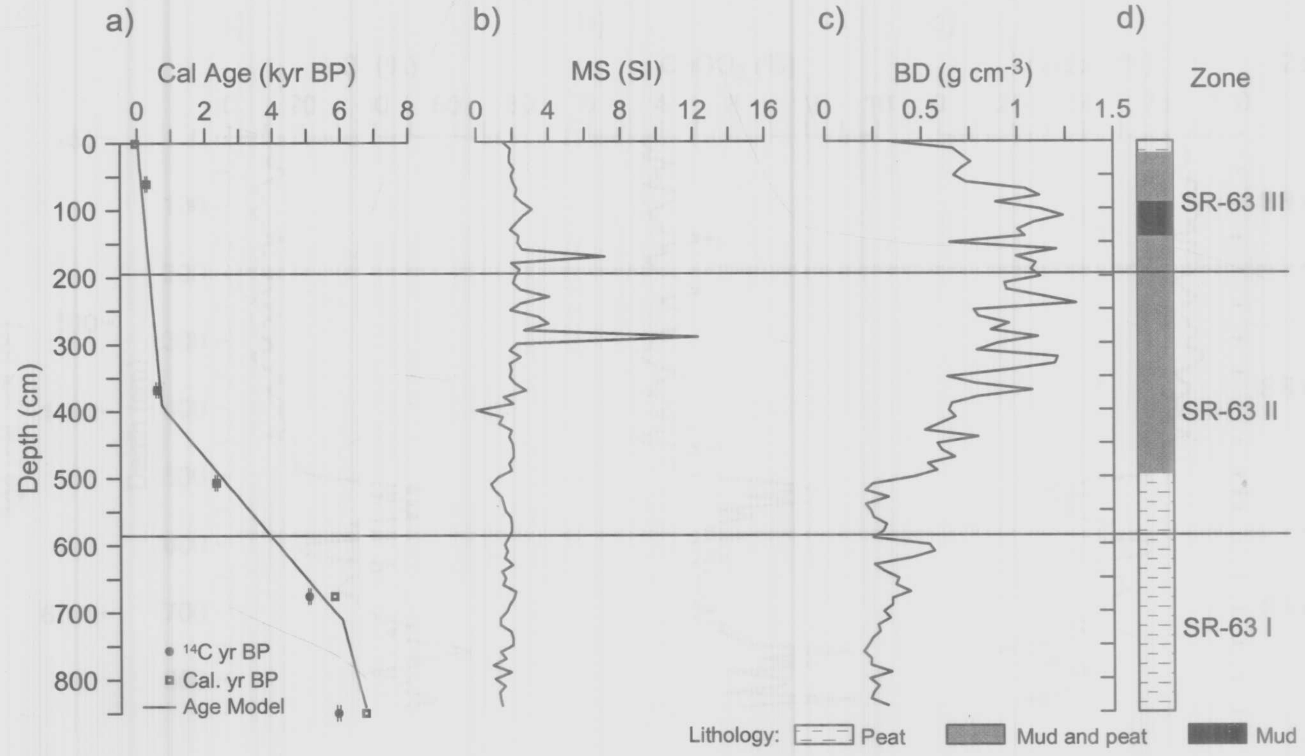

Figure 3.2: a) Radiocarbon ages (^{14}C yr BP), calibrated ages (cal. yrs BP), and age model (using equations 1-3). Error bars were calculated by NOSAMS at WHOI. b) Magnetic susceptibility (MS, SI units). c) Dry bulk density (BD, g cm^{-3}). d) Lithology based on notes taken in the field during coring.

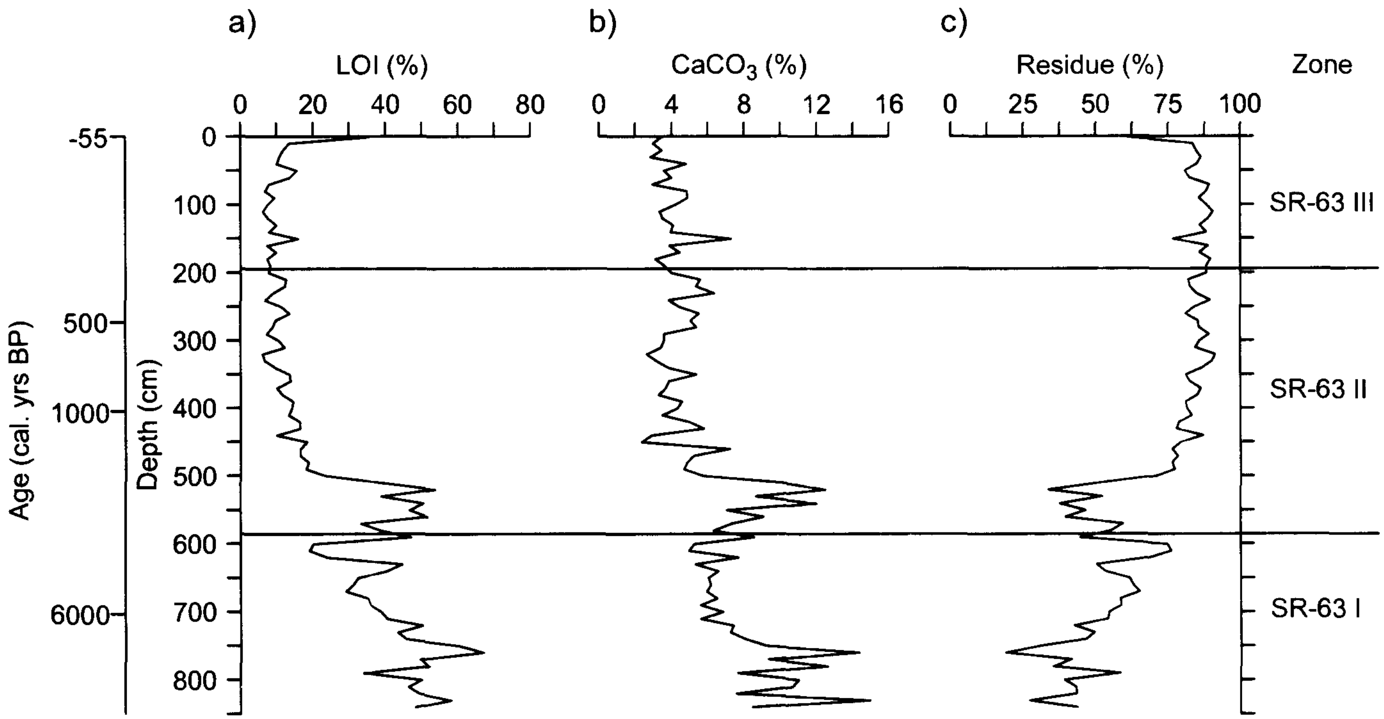

Figure 3.3: Plot of loss on ignition sequence: a) Loss on ignition (LOI, %). b) Calcium carbonate ($CaCO_3$, %). c) Remaining material (Residue, %).

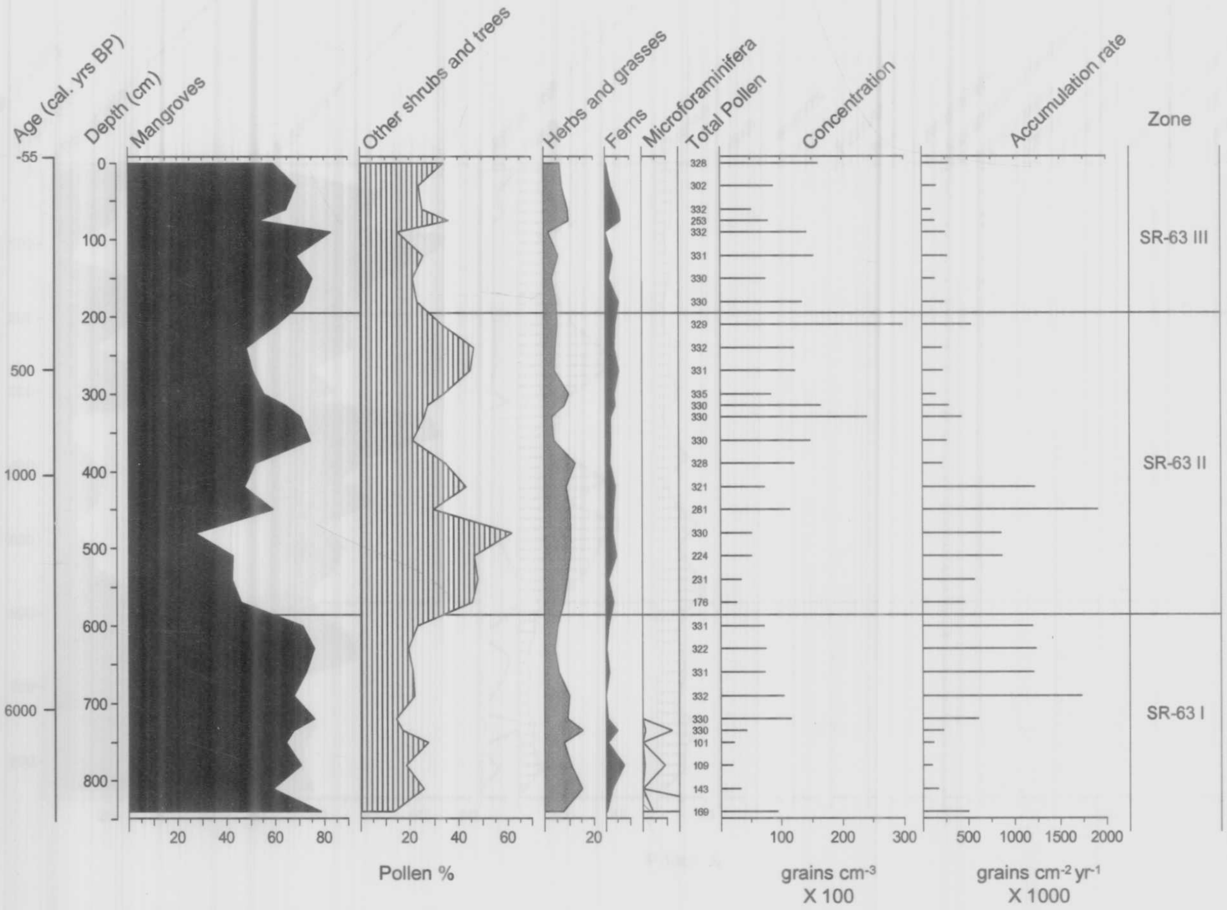

Figure 3.4: Summary pollen diagram showing vegetation groups, pollen concentration, pollen accumulation rate, and zones. Microforaminifera percentages are illustrated with a double line: the hollow line is exaggerated by a factor of 10.

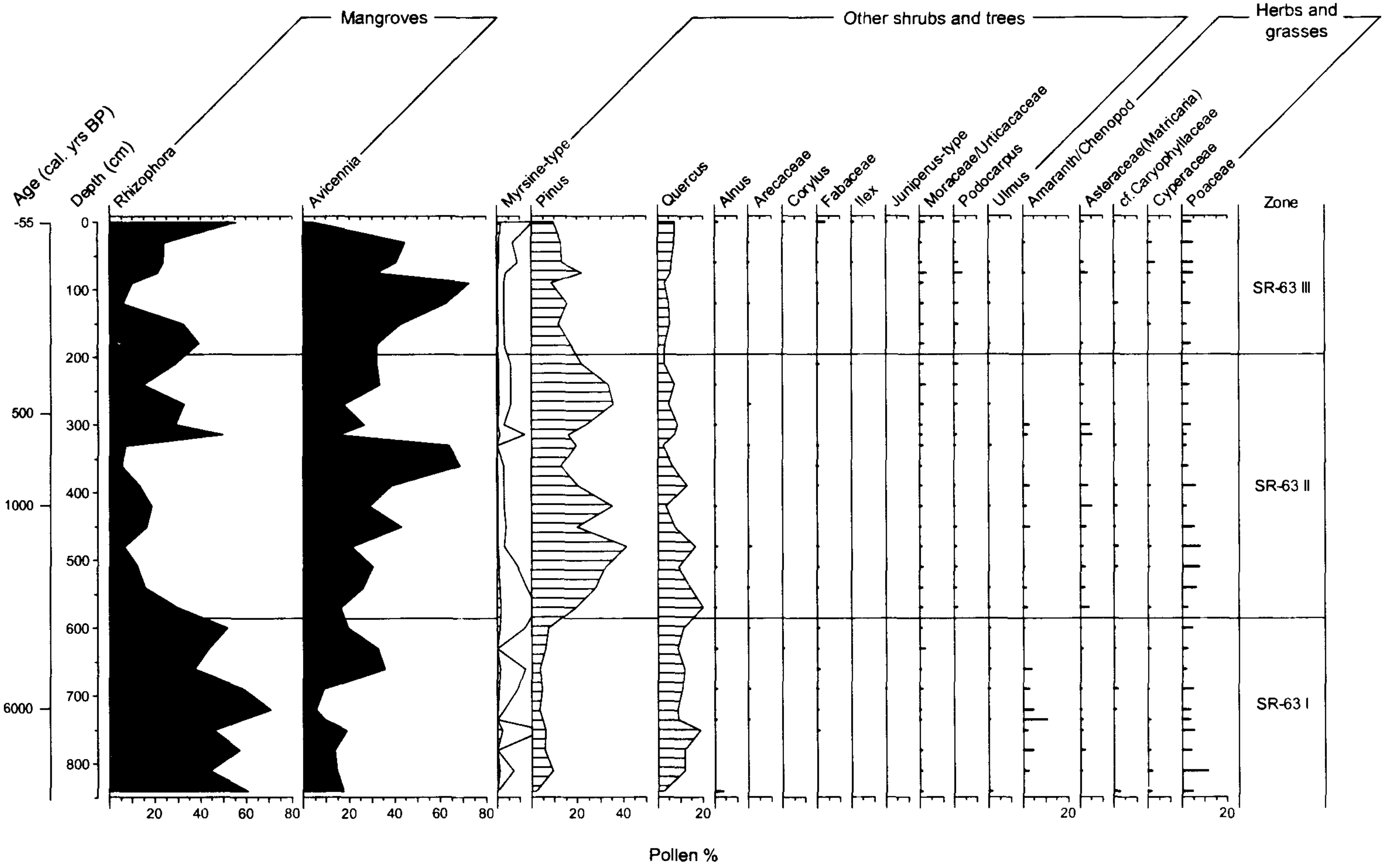

Figure 3.5: Percentage pollen diagram showing the most frequent pollen taxa. *Myrsine* percentages are illustrated with a double line: the hollow line is exaggerated by a factor of 10.

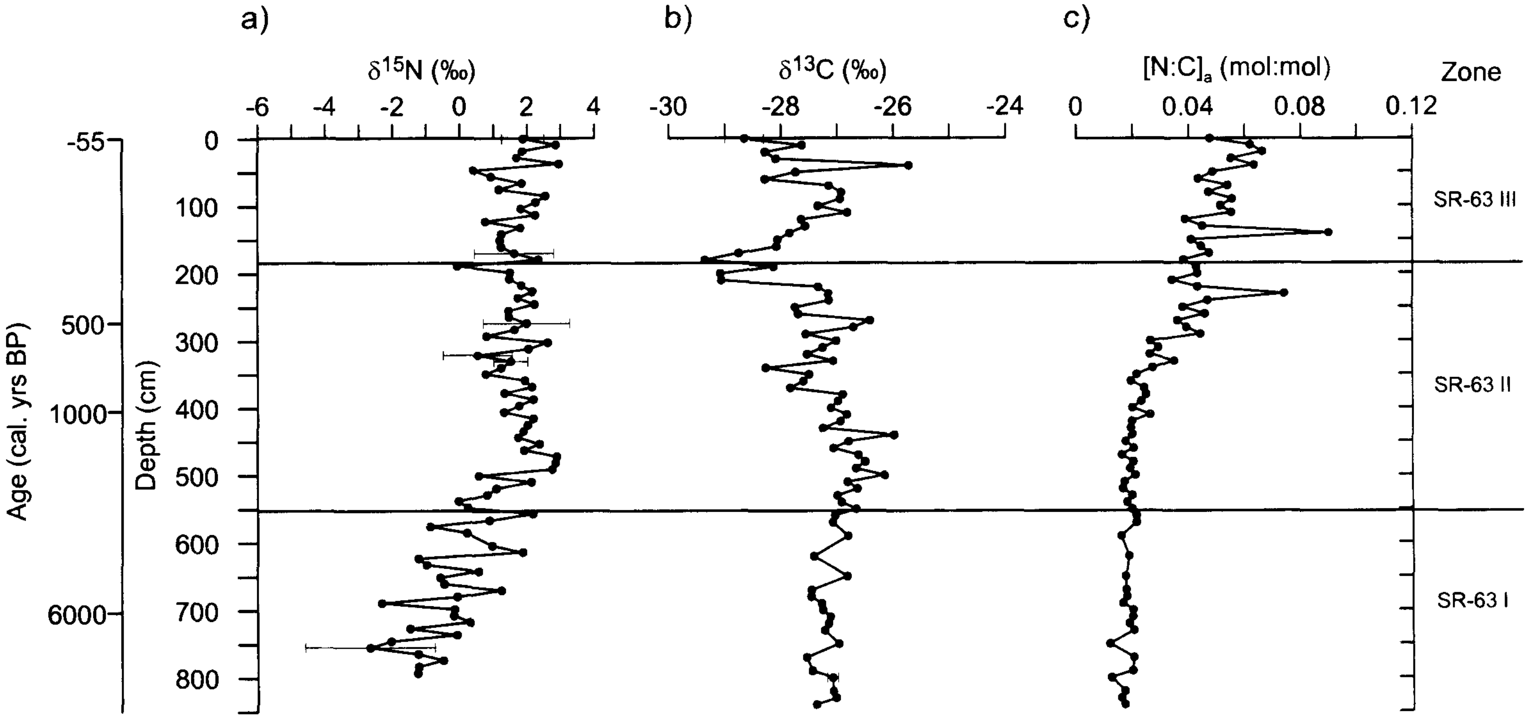

Figure 3.6. Plot of organic matter data: a) δ^{13}C (‰) vs. VPDB. b) δ^{15}N (‰) vs. air. c) [N:C]$_a$ (mol:mol) Error bars show standard deviations of triplicate analyses.

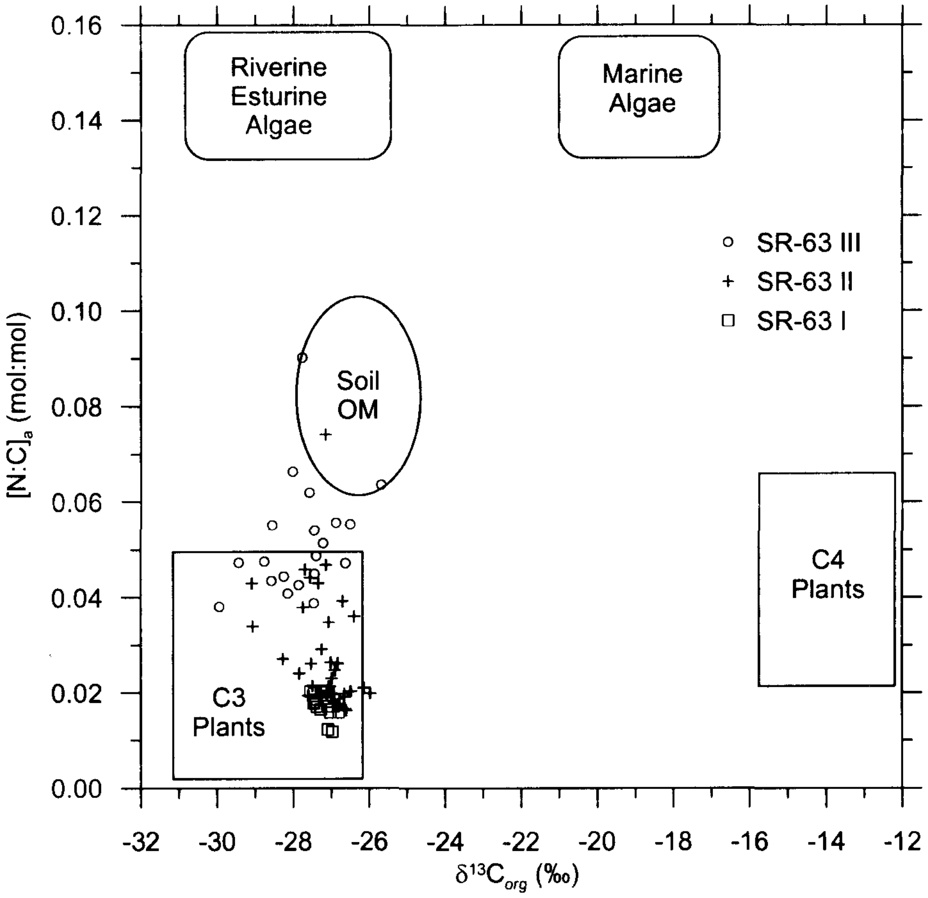

Figure 3.7: Plot of [N:C]a (mol:mol) vs. $\delta^{13}C$ (‰) where symbols indicate the zone. The ranges drawn have been previously published from comparable environments (e.g. Hedges et al., 1986; Goñi et al., 2006).

Table 3.1: List of AMS radiocarbon dates and calibrated dates using CALIB (Reimer et al., 2004).

Depth (cm)	Type	^{14}C age (yrs BP)	Cal. age (cal. yrs BP)	Cal. range (cal. yrs BP)	NOSAMS number
60-61	wood	265	290	159-422	OS-55507
366-369	wood	580	590	542-635	OS-55441
506-507	roots	2,320	2340	2,332-2,352	OS-55442
675-676	roots	5,070	5820	5,752-5,894	OS-55451
849-850	bulk peat	5,920	6730	6,676-6,786	OS-54384

References

Achee, N.L., Grieco, J.P., Masuoka, P., Andre, R.G., Roberts, D.R., Thomas, J., Briceno, I., King, R. and Rejmankova, E., 2006. Use of remote sensing and geographic information systems to predict locations of *Anopheles darlingi*-positive breeding sites within the Sibun River in Belize, central America. Journal of Medical Entomology 43(2), 382-392.

Alongi, D.M., 2002. Present state and future of the world's mangrove forests. Environmental Conservation 29, 331-349.

Angulo, R.J. and G.C. Lessa, 1997. The Brazilian sea-level curves: a critical review with emphasis on the curves from the Paranagua and Cananeia regions. Marine Geology 140(1-2), 141-166.

Behling, H. and da Costa, M.L., 2001. Holocene vegetational and coastal environmental changes from the Lago Crispim record in northeastern Para State, eastern Amazonia. Review of Palaeobotany and Palynology 114(3-4), 145-155.

Behling, H., Cohen, M.C.L. and Lara, R.J., 2001. Studies on Holocene mangrove ecosystem dynamics of the Braganca Peninsula in north-eastern Para, Brazil. Palaeogeography Palaeoclimatology Palaeoecology 167(3-4), 225-242.

BERDS, 2007. Biodiversity and Environmental Resource Data System of Belize. http://www.biodiversity.bz. Accessed October 2007.

Binford, M.W., Brenner, M., Whitmore, T.J., Higueragundy, A., Deevey, E.S. and Leyden, B., 1987. Ecosystems, Paleoecology and Human Disturbance in Subtropical and Tropical America. Quaternary Science Reviews 6, 115-128.

Bird, M.I., Fifield, L.K., Chua, S. and Goh, B., 2004. Calculating sediment compaction for radiocarbon dating of intertidal sediments. Radiocarbon 46(1), 421-435.

Blanchon, P., 2005. Comments on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. Coral Reefs 24(2), 183-186.

Curtis, J.H., Hodell, D.A., and Brenner, M., 1996. Climate variability on the Yucatan Peninsula (Mexico) during the past 3500 years, and implications for Maya Cultural Evolution. Quaternary Research 46(1), 37-47.

Curtis, J.H., Brenner, M., Hodell, D.A., Balser, R.A., Islebe, G.A., and Hooghiemstra, H., 1998. A multi-proxy study of Holocene environmental change in the Maya lowlands of Peten, Guatemala. Journal of Paleolimnology 19(2), 139-159.

Danielsen, F., Sorensen, M.K., Olwig, M.F., Selvam, V., Parish, F., Burgess, N.D., Hiraishi, T., Karunagaran, V.M., Rasmussen, M.S., Hansen, L.B., Quarto, A., and Suryadiputra, N., 2005. The Asian tsunami: A protective role for coastal vegetation. Science 310(5748), 643-643.

Dean, W.E., 1974. Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: Comparison with other methods. Journal of Sedimentary Petrology 44(1), 242-248.

Faegri, K. and Iverson, J., 1989. Textbook of Pollen Analysis, 4th edn. Wiley, New York.

Fairbanks, R.G., 1989. A 17,000-Year Glacio-Eustatic Sea-Level Record - Influence of Glacial Melting Rates on the Younger Dryas Event and Deep-Ocean Circulation. Nature 342, 637-642.

Gischler, E., 2003. Holocene lagoonal development in the isolated carbonate platforms off Belize. Sedimentary Geology 159, 113-132.

Gischler, E., 2006. Comment on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. Coral Reefs 22, 257-270 (2003), and their response in Coral Reefs 24, 187-190 (2005). Coral Reefs 25(2), 273-279.

Gischler, E. and Hudson, J.H., 1998. Holocene development of three isolated carbonate platforms, Belize, Central America. Marine Geology 144(4), 333-347.

Goñi, M.A., Teixeira, M.J. and Perkey, D.W., 2003. Sources and distribution of organic matter in a river-dominated estuary (Winyah Bay, SC, USA). Estuarine Coastal and Shelf Science 57(5-6), 1023-1048.

Goñi, M.A., Monacci, N., Gisewhite, R., Ogston, A., Crockett, J. and Nittrouer, C., 2006. Distribution and sources of particulate organic matter in the water column and sediments of the Fly River Delta, Gulf of Papua (Papua New Guinea). Estuarine Coastal and Shelf Science 69(1-2), 225-245.

Grimm E., 2004. TILIA and TGView, 2.0.2. Illinois State Museum, Springfield.

Haug, G.H., Hughen, K.A., Sigman, D.M., Peterson, L.C. and Rohl, U., 2001. Southward migration of the intertropical convergence zone through the Holocene. Science 293(5533), 1304-1308.

Hedges, J.I., Clark, W.A, Quay, P.D., Richey, J.E., Devol, A.H. and Santos, U.M., 1986. Compositions and Fluxes of Particulate Organic Material in the Amazon River. Limnology and Oceanography 31(4), 717-738.

Heiri, O., Lotter, A.F. and Lemcke, G., 2001. Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. Journal of Paleolimnology 25(1), 101-110.

Higuera-Gundy, A., Brenner, M., Hodell, D.A., Curtis, J.H., Leyden, B.W. and Binford, M.W., 1999. A 10,300 C-14 yr record of climate and vegetation change from Haiti. Quaternary Research 52, 159-170.

Hodell, D.A., Curtis, J.H., and Brenner, M., 1995. Possible Role of Climate in the Collapse of Classic Maya Civilization. Nature 375(6530), 391-394.

Hodell, D.A., Brenner, M., Curtis, J.H. and Guilderson, T., 2001. Solar forcing of drought frequency in the Maya lowlands. Science 292, 1367-1370.

Hodell, D.A., Brenner, M. and Curtis, J.H., 2007. Climate and cultural history of the northeastern Yucatan Peninsula, Quintana Roo, Mexico. Climatic Change 83(1-2), 215-240.

Jacob, J.S. and Hallmark, C.T., 1996. Holocene stratigraphy of Cobweb Swamp, a Maya wetland in northern Belize. Geological Society of America Bulletin 108(7), 883-891.

Kathiresan, K. and Rajendran, N., 2005. Coastal mangrove forests mitigated tsunami. Estuarine Coastal and Shelf Science 65(3), 601-606.

Kjerfve, B., Ruetzler, K., and Kierspe, G.H., 1982. Tides at Carrie Bow Cay, Belize. Smithsonian Contributions to the Marine Sciences. K. Ruetzler and I. G. Macintyre (Eds.). Washington, Smithsonian Institution Press. 12, 539.

Lalli, C.M. and Parsons, T.R., 1993. Biological Oceanography: An Introduction. 2nd ed. Elsevier: Oxford, pp. 314.

Leyden, B.W., 1987. Late Quaternary Aridity in the Lake Valencia Basin (Venezuela) - Reply. Ecology 68(5), 1553-1555.

Leyden, B.W., Brenner, M., Hodell, D.A. and Curtis, J.H., 1994. Orbital and Internal Forcing of Climate on the Yucatan Peninsula for the Past Ca.36-Ka. Palaeogeography Palaeoclimatology Palaeoecology 109, 193-210.

Lighty, R.G., Macintyre, I.G. and Stuckenrath, R., 1978. Submerged early Holocene barrier reef south-east Florida shelf. Nature 275, 59-60.

Meerman, J. and Sabido, W., 2001. Central American Ecosystems Map: Belize, Volume II: Ecosystem map and descriptions, Programme for Belize, Belize City.

Monacci, N.M., Meier-Grünhagen, U., Finney, B.P., Behling, H., Wooller, M.J., in prep. Mangrove ecosystem changes during the Holocene at Spanish Lookout Cay, Belize. Palaeogeography, Palaeoclimatology, and Palaeoecology.

Murray, M.R., Zisman, S.A., Furley, P.A., Munro, D.M., Gibson, J., Ratter, J., Bridgewater, S., Minty, C.D. and Place, C.J., 2003. The mangroves of Belize Part

1. distribution, composition and classification. Forest Ecology and Management 174(1-3), 265-279.

Palacios-Chavez, R., Ludlow-Wiechers, B., and Villanueva, R. 1991. Flora Palinologica de la Reserva de la Biosfera de Sian Ka'an, Quintana Roo, Mexico, 1st edn. CIQRO: Chetumal.

Piaskowski, V.D., Teul, M., Williams, K.M. and Cal, R.N., 2006. Birds of the Sibun riverine forest, Belize. Ornitologia Neotropical 17(3), 333-352.

Polk, J.S., van Beynen, P.E. and Reeder, P.P., 2007. Late Holocene environmental reconstruction using cave sediments from Belize. Quaternary Research 68(1), 53-63.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hogg, A.G., Hughen, K. A., Kromer, B., McCormac, G., Manning, S., Ramsey, C. B., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., van der Plicht, J., and Weyhenmeyer, C.E., 2004. IntCal04 terrestrial radiocarbon age calibration, 0-26 cal kyr BP. Radiocarbon 46(3), 1029-1058.

Sobrado, M.A., 2000. Relation of water transport to leaf gas exchange properties in three mangrove species. Trees 14, 258-262.

Toscano, M.A. and Macintyre, I.G., 2003. Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated C-14 dates from *Acropora palmata* framework and intertidal mangrove peat. Coral Reefs 22(3), 257-270.

Traverse, A., 1988. Paleopalynology. Unwin Hyman, Boston, pp. 600.

Vannucci, M., 2001. What is so special about mangroves? Brazilian Journal of Biology 61(4), 599-603.

Vedel, V., Behling, H., Cohen, M., and Lara, R., 2006. Holocene mangrove dynamics and sea-level changes in northern Brazil, inferences from the Taperebal core in northeastern Para State. Vegetation History and Archaeobotany 15, 115-123.

Wahl, D., Byrne, R., Schreiner, T., and Hansen, R., 2006. Holocene vegetation change in the northern Peten and its implications for Maya prehistory. Quaternary Research 65(3), 380-389.

Woodroffe, C.D., 1988. Mangroves and sedimentation in reef environments:indicators of past sea-level changes, and present sea-level trends? Proceedings of the Sixth International Coral Reef Congress 3, 535-539.

Wooller, M.J., Behling, H., Smallwood, B.J. and Fogel, M., 2004. Mangrove ecosystem dynamics and elemental cycling at Twin Cays, Belize, during the Holocene. Journal of Quaternary Science 19(7), 703-711.

Wooller, M.J., Morgan, R., Fowell, S.J., Behling, H. and Fogel, M., 2007. A multi-proxy peat record of Holocene mangrove paleoecology from Twin Cays, Belize. Holocene 17(8), 1129-1139.

Chapter 4 General Conclusion

This study used multiple, independent proxies of past environments to understand and compare the paleoecology of mangrove ecosystems both onshore and offshore Belize. Radiocarbon analyses, magnetic susceptibility, bulk density, loss on ignition, pollen analyses, and stable isotope ratios were compared between mangrove habitats. Previous work documents some of these techniques at coring sites TCC1 and TCC2 on Twin Cays (Wooller et al., 2004 and 2007).

Two core sites on Twin Cays were the first to be analyzed by our group and the mangrove habitats were established ~8,000 cal. yrs BP. Significant environmental changes were noted throughout the Holocene. Core TCC1 records a significant increase in *Myrsine*-type pollen after ~5,000 cal. yrs BP coeval with an increase in peat accumulation, presumably attributed to thriving *Myrsine* and decrease in inundation of seawater. After ~4,000 cal. yrs BP, productivity at TCC1 declined with a greater abundance of dwarf *R. mangle* and decreased peat accumulation rates (Wooller et al., 2004). Core TCC2 also records elevated *Myrsine*-type pollen in the mid Holocene, although not at such high percentages as core TCC1 (Wooller et al., 2007), suggesting local variability across different island settings.

The coring sites described in this thesis, BT-79 and SR-63 on Spanish Lookout Cay (SLC) and along the Sibun River respectively, provide evidence of coeval environmental change with Twin Cays, suggesting that these ecosystems are responding to regional environmental changes, albeit with different local results. Core BT-79 records colonization by mangroves at ~8,000 cal. yrs BP (Figure 4.1). It is assumed that the same holds true from site SR-63; however, this record only extends back to ~7,000 cal. yrs BP (Figure 4.2). In Figure 4.3 taken from Toscano and Macintyre (2003), there are only five ages of mangrove peat that are older than 8,000 cal. yrs BP and three samples are from Belize. Despite these few samples, it appears that mangrove colonization throughout most of the region did not occur until after 8,000 cal. yrs BP, but there is no reported evidence of a drastic change in sea-level at this time. Blanchon and Shaw (1995) proposed an abrupt sea-level rise in this region at ~7,600 cal. yrs BP and Toscano and Macintyre (2003) discuss a decrease in the rate of sea-level rise at ~7,700 cal. yrs BP, but these findings are after mangrove colonization had begun. It is possible that conditions immediately following the paleoclimatic instability in the early Holocene, reported as the "8.2kyr event", lead to a favorable framework for mangrove ecosystem development. The 8.2kyr event is described as a cold, dry, windy event in and around the North Atlantic (e.g. Alley et al., 1997), which appears to be a climatic signal that is observed globally (e.g. Ljung et al., 2007). Therefore, it can be assumed that the conditions that immediately followed this event were relatively warmer, wetter, and less windy, which are opportune characteristics for mangroves.

The most obvious similarity between sites BT-79 and SR-63 is the drastic decrease in peat accumulation rates at ~6,000 cal. yrs BP that lasted for ~5,000 yrs. There is no significant increase in *Myrsine*-type pollen, as seen at Twin Cays, which was attributed to a decrease in seawater inundation to allow a non-halophyte species to proliferate at the sites (Wooller et al., 2004 and 2007). Wooller et al. (2004 and 2007) also presented the *Myrsine* signal as a response to a change in the prevailing wind direction bringing *Myrsine*-type pollen over from the mainland. Unfortunately, no modern or paleopalynological analyses were performed by our group on the mainland to the west of Twin Cays. It is evident from pollen analyses at site SR-63, to the west of SLC, that *Myrsine* is not extremely prevalent and perhaps did not have the significant pollen concentrations to be transported offshore to SLC to support the type of habitat that was sustained at Twin Cays. The peak of *Myrsine*-type pollen as seen in core TCC1 was recorded to have been a brief event, ~240 yrs, and it is possible that peak was missed in core BT-79 due to the coarse resolution at ~4,000 cal. yrs BP. Nevertheless, the *Myrsine* signal seen at Twin Cays is most likely a local result to a decrease in seawater inundation to the site or a change in winds.

Changes in sedimentation rates, pollen assemblages, nutrient status, and source water of *R. mangle* are seen throughout the Holocene at all four sites studied in Belize. Published sea-level curves such as Toscano and Macintyre's (2003) synopsis of data from the greater Caribbean region do not suggest that there was a decrease in sea-level in the Holocene, nor do they suggest a decrease in the rate of sea-level rise followed by an increase in sea-level rise. They concluded that there was a decrease in the rate of rise of sea-level at 7,700 and 2,000 cal. yrs BP. Unfortunately, these curves are based on one date from many sites that clearly do not have the same environmental characteristics. From our analyses of four different sites in Belize that have been inhabited by mangroves throughout the Holocene, we observe evidence that there was significant, and perhaps disruptive, environmental change between ~6,000 and 1,000 cal. yrs BP. Due to dating uncertainty, the exact determination of the timing of such events is not well constrained.

The goal of this study was not to produce a sea-level curve, although a mangrove ecosystem is a natural tracker of sea-level position due to the placement of this habitat at the water's edge. Therefore, there is an assumed hiatus of seawater inundation to these sites during the Holocene. In Figure 4.4, the similarity in accumulation rates at Spanish Lookout Cay and along the Sibun River is obvious. Although age versus depth profiles look slightly different at Twin Cays, it is still evident that a smoothly rising sea-level does not correspond with the shown data. The advantage of observing one site through time allows more evidence of the position of the relative sea-level to that site and the response of the local habitat.

Sea-level is not static; it has been constantly changing for millions of years and it will continue to change. A changing sea-level has societal importance because of its potential to alter ecosystems and habitability in coastal regions where a large percentage of the globe's population lives (Douglas, 2001). Many predictions are currently being made about the response of coastal environments to a rise in sea-level. Figure 4.5 illustrates a possible sea-level rise scenario for the next century and estimates approximate 4mm yr^{-1}, according to the Intergovernmental Panel on Climate Change (IPCC, Bindoff et al., 2007). These values are comparable to the accumulation rates observed at site BT-79 from ~1,000 cal. yrs BP to the present (3.9mm yr^{-1}) and less than rates observed at site SR-63 (5.6mm yr^{-1}) during the same time period. Comparisons of a total ecosystem response to environmental change between past, present, and future scenarios provides an in-depth understanding of our dynamic environment.

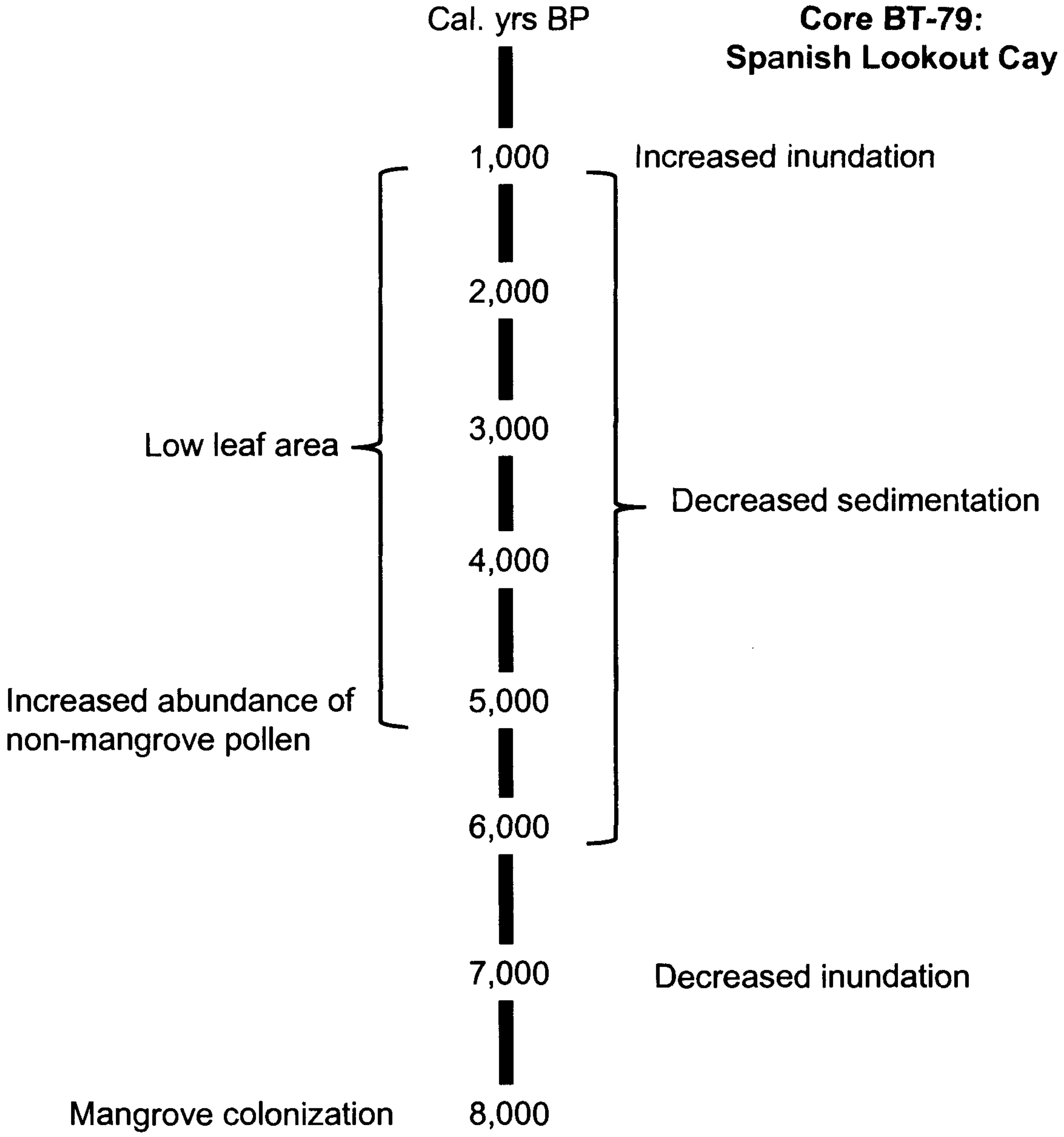

Figure 4.1: BT-79 timeline, approximate ages of noted observations from core BT-79, Spanish Lookout Cay.

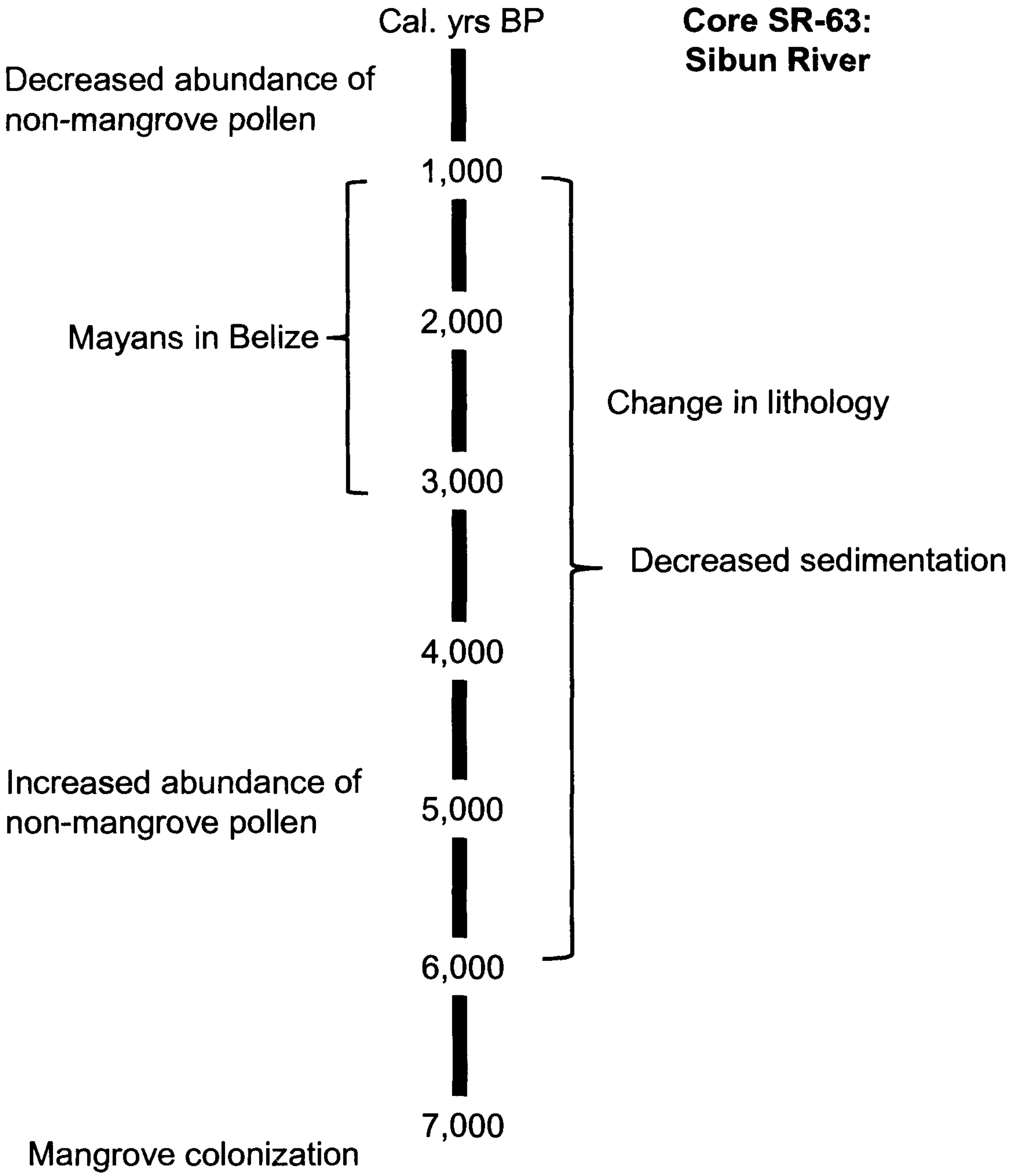

Figure 4.2: SR-63 timeline, approximate ages of noted observations from core SR-63, Sibun River.

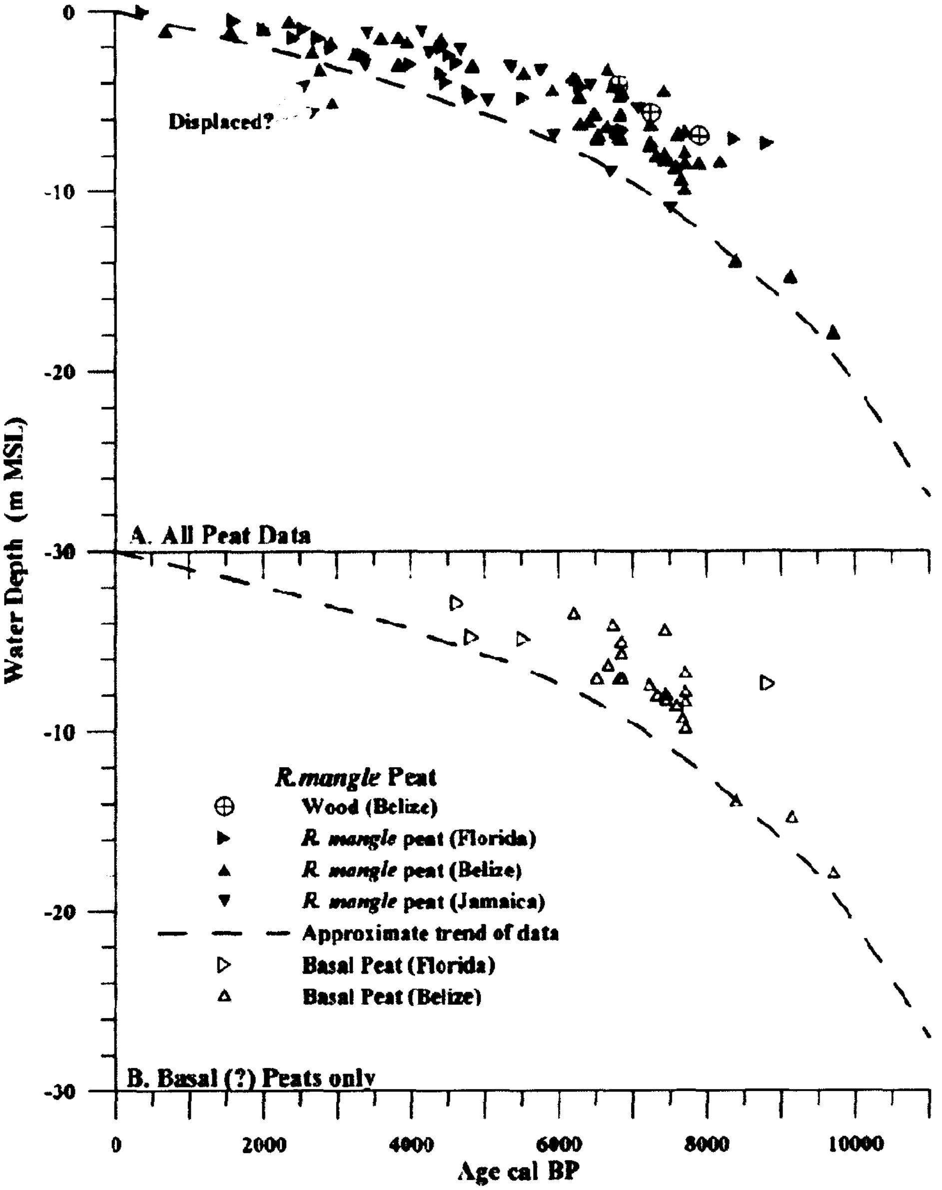

Figure 4.3: Caribbean mangrove peat data. A) Calibrated mangrove peat data from Florida, Belize (Tobacco Range, Twin Cays, and Carrie Bow Cay), and Jamaica in time–depth space. Dashed line drawn along base of data field indicates an upward trend over time (but is not a sea-level curve). Displaced refers to two samples which plot deeper than the data field or trend line, due either to compaction or to dating complications such as younger root contamination. B) Interpreted basal or near-basal peat data from Florida and Belize (as designated in original publications). (taken from Toscano and Macintyre, 2003)

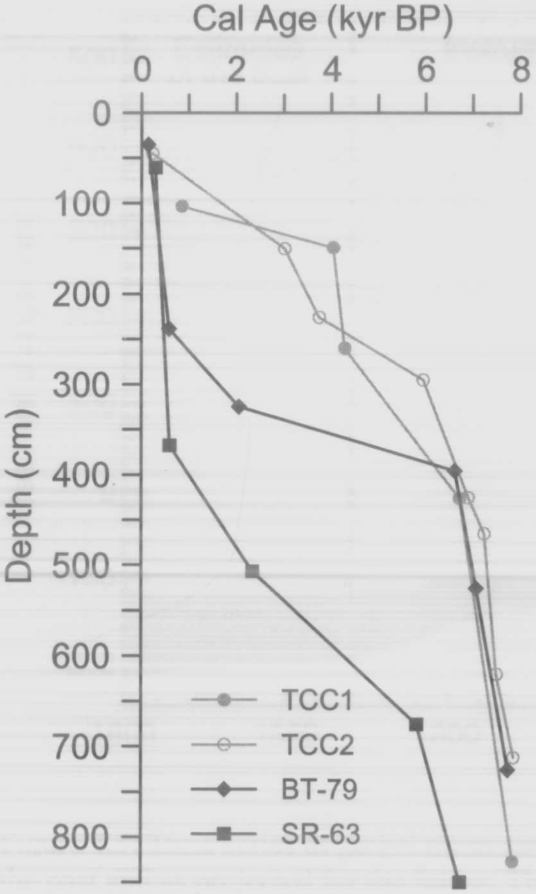

Figure 4.4: Plot of depth vs. age showing all calibrated age samples from cores TCC1 and TCC2 from Twin Cays (Wooller et al., 2004 and 2007, respectively) and cores BT-79 and SR-63 from this study.

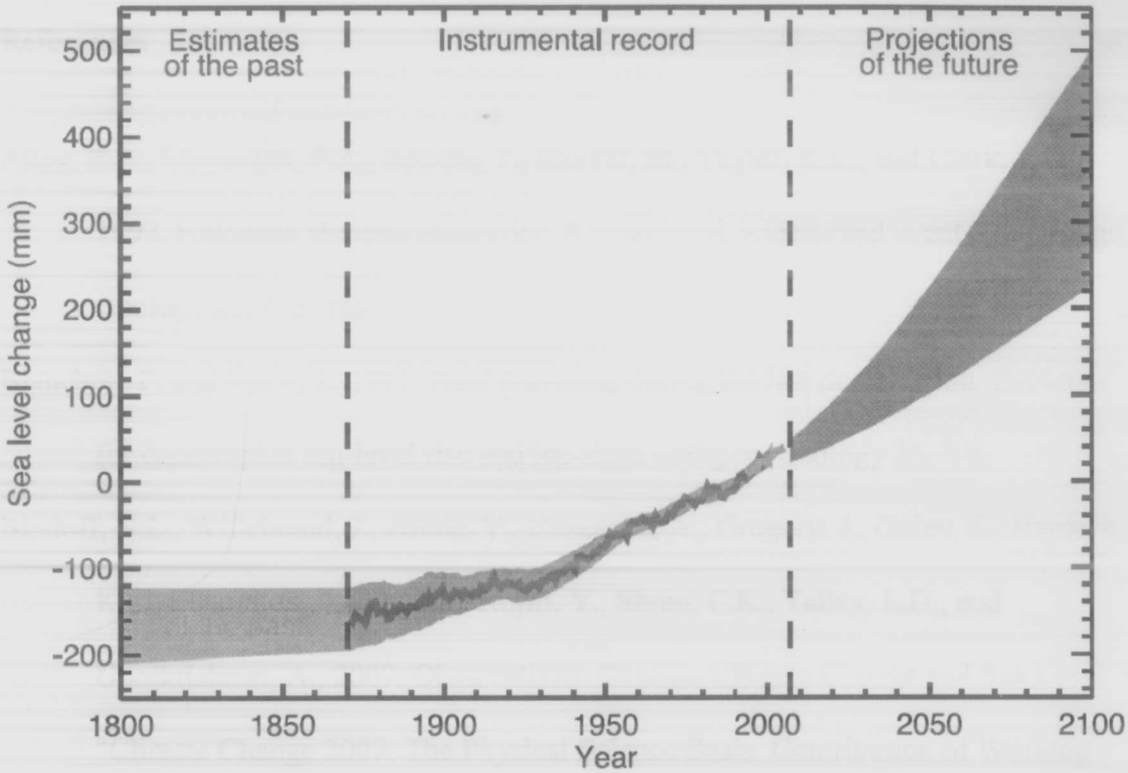

Figure 4.5: Time series of global mean sea-level (deviation from the 1980-1999 mean) in the past and as projected for the future. The grey shading shows the uncertainty in the estimated long-term rate of sea-level change. (taken from Bindoff et al., 2007)

References

Alley, R.B., Mayewski, P.A., Sowers, T., Stuiver, M., Taylor, K.C., and Clark, P.U. 1997. Holocene climatic instability: A prominent, widespread event 8200 yr ago. Geology 25, 483-486.

Blanchon, P. and Shaw, J., 1995. Reef drowning during the last deglaciation: Evidence for catastrophic sea-level rise and ice-sheet collapse. Geology 23, 4-8.

Bindoff, N.L., Willebrand, J., Artale, V., Cazenave, A., Gregory, J., Gulev, S., Hanawa, K., Le Quéré, C., Levitus, S., Nojiri, Y., Shum, C.K., Talley, L.D., and Unnikrishnan, A., 2007. Observations: Oceanic Climate Change and Sea Level. In "Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change." (S. Solomon, Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, M., and Miller, H.L., Eds.). Cambridge University Press, Cambridge and New York.

Douglas, B.C., 2001. Sea Level Rise: History and Consequences. International Geophysics Series, 75.

Ljung, K., Bjorck, S., Renssen, H., and Hammarlund, D., 2007. South Atlantic island record reveals a South Atlantic response to the 8.2kyr event. Climate of the Past Discussions 3, 729-753.

Toscano, M.A. and Macintyre, I.G., 2003. Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated ^{14}C dates from *Acropora palmata* framework and intertidal mangrove peat. Coral Reefs 22, 257-270.

Wooller, M.J., Behling, H., Smallwood, B.J. and Fogel, M., 2004. Mangrove ecosystem dynamics and elemental cycling at Twin Cays, Belize, during the Holocene. Journal of Quaternary Science 19, 703-711.

Wooller, M.J., Morgan, R., Fowell, S.J., Behling, H., Fogel, M., 2007. A multi-proxy peat record of Holocene mangrove paleoecology from Twin Cays, Belize. Holocene 17(8), 1129-1139.