A Multi-proxy Record of Paleoenvironmental Change

from a Southern Belizean Pine Savanna

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ABSTRACT

While tropical savannas are naturally fire-prone ecosystems, anthropogenic burning can influence their distribution. Precisely when human activity in southern Belize began to significantly alter regional fire regimes is unknown, but savannas there are composed of fire-adapted species and are regarded as having evolved with both anthropogenic and natural fire. This thesis presents a multiproxy record from lake sediment cores that documents the complex interaction between fire, people, climate, and vegetation during the late Holocene. The dataset derives from 1) high-resolution analysis (1-cm interval sampling) of local fire history based on macroscopic (>125 μm) charcoal; 2) pollen analysis to determine vegetation history; and 3) bulk geochemistry analyses. These analyses were carried out on two cores recovered in March 2018 from Pine Pond, a roughly circular pond of about 0.5 ha on the boundary of the Deep River Forest Reserve on the southern coastal plain of Belize. The pond is presently surrounded by an area of tropical pine savanna that may be threatened by increased, near-annual anthropogenic burning. Core 1, collected closer to shore, is ~2 m long and covers the last ca. 1400 years. Core 2 from near the center of the lake is ~6 m long and extends to ca. 4800 cal yr BP. The analyses of these cores demonstrate the presence of a neotropical savanna ecosystem, with varying degrees of fire occurrence, for at least the last ca. 4800 years in the area around Pine Pond. Fire tends to increase during drier periods, with relatively less burning during wetter periods. Beginning in the Classic Maya era and continuing to the present, an increase in fire occurrence may have been a product of changing land use practices in addition to climate changes. The evidence for changes in fire, vegetation, and climate during the prehistoric and historic eras may be relevant for

conservation practices within the Deep River Forest Reserve and the adjacent Payne’s Creek National Park.

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CHAPTER 1

INTRODUCTION

Geographers have long been interested in the topic of environmental change, and how changes in the environment are caused by and influence human activity (Pacione 1999, Zimmerer and Bassett 2003, Cutter et al. 2004, Bell and Walker 2013). This thesis research uses techniques in paleolimnology to decipher environmental changes during the late Holocene in southern Belize, based on sediment cores from a small pond located in an area of pine savanna.

Paleolimnology is the study of lake sediments as environmental archives (Horn 2017). Various materials and signatures preserved in lake-sediment cores can be used to reconstruct paleoclimates and prehistoric human-environment interactions at various spatiotemporal scales. Two commonly examined environmental indicators or "proxies" in sediment cores are pollen grains (revealing vegetation changes) and charcoal (revealing fire history) (Brenner et al. 2002, Horn 2007, Power et al. 2010). Across Central America, paleolimnological studies complement and may even supplant traditional archaeological studies in documenting dynamic human-environment feedbacks that shaped prehistoric cultural ecology and evolution (Islebe et al. 1996, Leyden et al. 1998, Dull 2004a, Wahl et al. 2006, Horn 2006, Lane et al. 2008). Many such studies across the region, however, are limited to pollen analysis. Studying a greater array of proxies including charcoal and stable carbon and nitrogen isotope ratios can provide records that better reveal complex human-climate dynamics and ecosystem change across varying temporal scales. For my thesis research, I examined multiple proxies in the sediments of Pine Pond, Belize, and compared the results to archaeological evidence, independent climate records, and other paleoecological reconstructions in the region. This analysis brings to light spatiotemporal changes in environmental processes, both human-caused and natural, that have affected savanna development around Pine Pond.

1.1 Research Questions

While ponds in some savanna regions of Central America desiccate during the December–May dry season, Pine Pond in Belize holds water year-round, allowing for the accumulation and preservation of organic sediments and proxy indicators that can be used to document past changes in the environment. The goal of my research is to reconstruct past changes in vegetation and fire history around Pine Pond using charcoal, pollen, and bulk stable carbon and nitrogen isotope analyses. I then compare my results with other paleolimnological reconstructions in the region, local speleothem-based climate reconstructions, and local archaeological evidence to better understand the dynamics of climate, fire, and anthropogenic land use on varying spatiotemporal scales.

Key questions driving my research include:

1. Did ecosystems change over the interval covered by the Pine Pond sediment record? What do proxy analyses reveal about the environmental history of the site?
2. What do proxy analyses reveal about late Holocene climate change at the site? Do changes in charcoal, pollen, and stable isotope values reveal evidence of climate events such as the Little Ice Age and Terminal Classic Drought?
3. What do the pollen and geochemical analyses reveal about human disturbance at the site, and how does the record compare to other paleolimnological records for the Maya lowlands?

4. Were ecosystem changes over the late Holocene primarily driven by people or climate? How does the Pine Pond record compare to the fire and climate record of Walsh et al. (2014)? When, if at all, did humans begin to significantly alter fire regimes around Pine Pond?

1.2 Thesis Outline

This thesis consists of seven chapters. In the following chapter, I describe the study site of Pine Pond and its environmental context. Chapter 3 reviews the literature on Central American savannas, local archaeology, and other paleoenvironmental studies in the region. The literature review also includes a consideration of the proxies I examined. I describe my methods for charcoal, pollen, and bulk stable carbon and nitrogen isotope analyses in Chapter 4. I present the results of my multi-proxy analysis in Chapter 5 and use these results to answer my research questions in Chapter 6. I then summarize the important findings and conclusions in Chapter 7.

CHAPTER II

PINE POND AND ITS ENVIRONMENTAL CONTEXT

2.1 Site Description

Pine Pond (0.5 ha, 7 m deep) is a roughly circular pond in the Deep River Forest Reserve of southern Belize (16.3682°N, 88.6311°W) (Figure 1; Figure 2). Pine Pond occupies an area on the southern coastal plain where Quaternary alluvium uncomfortably overlies the Paleocene-Eocene Toledo Formation (Geomedia 2014). The pond likely formed by collapse following dissolution of underlying carbonate deposits. The pond is surrounded by tropical pine savanna, a vegetation type composed of species adapted to and requiring recurring fires, including the Caribbean pine, *Pinus caribaea* (Figure 3). While savannas are naturally fire-prone ecosystems, anthropogenic burning can influence their distribution (Colombaroli et al. 2014). This may be particularly true in tropical areas with distinct wet and dry seasons where the frequency of natural fires might be low because lightning strikes are rare at times when fuels are dry (Walsh et al. 2014, Colombaroli et al. 2014). Belizean pine savannas evolved with both anthropogenic and natural fire (Bridgewater et al. 2002, Laughlin 2002), but the spatiotemporal variability of these disturbances on multi-century scales remains unknown. Today, many of these savannas may be threatened due to increased, near-annual burning in some areas coupled with the context of a still nascent system for ecosystem conservation (Young 2008, Cherrington et al. 2010).

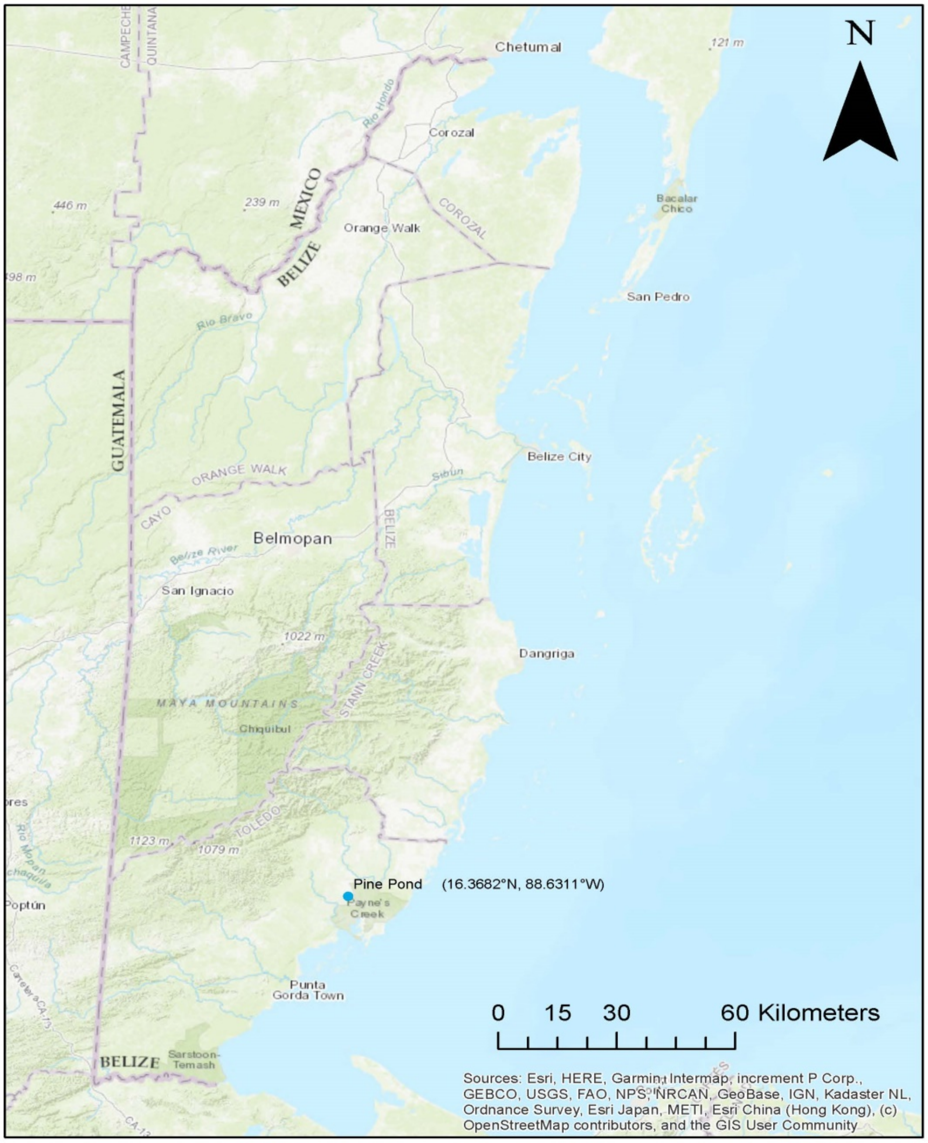

Figure 1. General reference map of Belize showing location of Pine Pond. This map was created by the author using the World Topographic Basemap in ArcMap.

Figure 2. Pine Pond in southern Belize. Photo courtesy of Rick Anderson.

Figure 3. Grass and sedge-dominated pine savanna vegetation on the edge of Pine Pond. Photo taken by Mathew Boehm, March 2018.

2.2 Climate

The area surrounding Pine Pond falls within the subtropical wet forest life zone in the Holdridge Classification (Holdridge 1967, Mcloughlin 2015). Temperature seasonality at the site is low, with mean monthly temperatures ranging from about 24 °C in January to 29 °C in June and July (Linares and Vinicio 2009, Mcloughlin 2013). Rainfall is highly seasonal, with a distinct wet and dry season as found throughout Belize. There is a pronounced south to north gradient in annual rainfall across the country, with annual means ranging from over 4000 mm in the south to as low as ca. 1500 mm in the north. The rainy season in the south often begins in early May, whereas the rainy season begins to affect the drier north around early June (Belize Meteorological Weather Service 2018). Mean monthly values at weather stations near Pine Pond are low during the dry season that lasts from January to early May (less than 100 mm); monthly values are higher during the wet season with peaks from June to August (greater than 300 mm) (Linares and Vinicio 2009). Average annual rainfall around Pine Pond is ca. 2500 mm (Belize Meteorological Weather Service 2018).

The annual migration of the ITCZ across the country largely accounts for the distinct wet and dry seasons in Belize (Dunning and Beach 2012, Akers et al. 2016). Rainfall totals in the south are higher than in the north, and the wet season lasts longer, because the area is located for a longer period within the zone of ITCZ migration. The position of the North Atlantic Subtropical High also influences precipitation variability. Westward movement of this high-pressure area across the tropical Atlantic produces stable southeasterly winds (Pollock et al. 2016) that correspond with the start of the warmer part of the dry season from March to May. Because of the unique geographical position of Belize, changes in atmospheric-driven ocean currents in both the Pacific and Atlantic produce climate variability (Giannini et al. 2000, Gamble and Curtis 2008, Wahl et al. 2014). Modern droughts and precipitation variability across Central America are linked to stronger El Niño phases of the ENSO (El Niño Southern Oscillation) cycle as well as southerly shifts in the mean annual position of the ITCZ (Poveda et al. 2006, Wang 2007). The effect of ENSO variability modulated by the Pacific Ocean is strongest on the western coast of Central America (Jury et al. 2007), while positive and negative phases of the Atlantic Decadal Oscillation and North Atlantic Oscillation are the predominant climate controls on the Atlantic side (Enfield et al. 2001, Elder et al. 2014). The two systems, however, are likely interconnected. ENSO cycles affect atmospheric conditions in the tropical Atlantic, which may add further interannual variation to the AMO (Atlantic Multi-decadal Oscillation)-driven precipitation variability (Giannini et al. 2000, Hu and Feng 2012).

2.3 Hydrology, Soils, and Fire

Pine Pond is located on the southern coastal plain of Belize within a lowland tropical pine savanna, an ecosystem type characterized by a seasonal drainage dichotomy from flooded to highly desiccated, along with nutrient-poor soil. In tandem, these factors are thought to be an important control on the regional distribution of the pine savanna ecosystem across the southern coastal plain (Baillie et al. 1993). On a local scale, minor gradients in topography and geological substrate affect vegetation distribution, with calcareous outcrops and alluvial deposits supporting patches of tropical broadleaf forest with high hardwood species diversity and drier ridges dominated by pines and oaks (Smith 2019). However, stands of oak and patches of lowland tropical broadleaf forest also occur around Pine Pond in areas with poor soil and drainage, indicating that another factor may be important for local pine dominance (Mcloughlin 2013, Figure 4). That factor appears to be recurrent fires, which favor pine and maintain the ecotone between savanna and the dense lowland tropical broadleaf forest that characterizes the broader region of the coastal plain (Baille et al. 1993, Meerman and Sabido 2001, Wyatt 2008, Walsh et al. 2014). Reports of lightning-ignited fires on the southern lowland plain in the vicinity of Pine Pond are rare and anecdotal (Linares and Vinicio 2009, Mcloughlin 2013). That lightning is rarely reported as a cause of fire suggests that human-set fires are an important influence on vegetation patterns.

In other areas of Belize, pine is known to spread into areas of human-induced soil degradation (Lentz and Hockaday 2009). The impact of Maya people across Belize on soils is well documented (Beach et al. 2006, 2017) and includes soil degradation, the creation of black soils by organic enrichment and extensive burning, and alterations to drainage and erosion patterns caused by different agricultural techniques. The presence of acidic topsoil above an impermeable clay layer likely represents a significant edaphic and lithographic control on savanna distribution around Pine Pond. In flat areas of the coastal plain such as the area around Pine Pond, the low topographic relief causes slow, sideways sheet erosion that thins the soil while stripping it of vital agricultural nutrients (Linares and Vinicio 2009, Mcloughlin 2013). Humans are unlikely to have engaged in large-scale burning for agriculture in the immediate vicinity of Pine Pond, but people currently set fires for hunting (Linares and Vinicio 2009, Mcloughlin 2013). Whether anthropogenic or natural in origin, the presence of fire-adapted plant-species (He et al. 2011, Goodwin et al. 2013) indicates a long history of fire.

Figure 4. View looking north east across the savanna to the north of the Deep River Forest Reserve. Photo courtesy of Cathy Smith.

2.4 Vegetation

In the first comprehensive floral survey of Belizean savannas, Goodwin et al. (2013) defined lowland savannas as "any natural or semi-natural, fire-influenced ecosystem under 500 m in altitude with a continuous herbaceous layer dominated by native grasses and sedges". While grasses, sedges, and other herbaceous plants dominate, some woody tree and shrub taxa may occur. This description characterizes the modern savanna ecosystem surrounding Pine Pond. Various species of Poaceae (grass), Cyperaceae (sedge), and Melastomataceae dominate the herbaceous layer among scattered to dense stands of *Pinus caribaea* (Caribbean Pine) and *Acoelorraphe wrightii* (Palmetto) (Figure 5). Other prominent woody taxa include: *Quercus oleoides* (Live Oak), *Byrsonima crassifolia* (Craboo), and *Myrica cerifera* (Tea Bark) (Linares and Vinicio 2009, Mcloughlin 2013, Goodwin et al. 2013). In areas of better soil quality and lower incidence of fire, *Scleria bracteata* (Cyperaceae) dominates the understory of lowland broadleaf forest (Mcloughlin 2013).

In Belize, tropical savannas constitute the second most extensive ecosystem type after broadleaf forest (Goodwin et al. 2013). While most lowland savannas across Central America and Belize are characterized by acidic soil and seasonal saturation of the soil, the uniformity of these edaphic and climate controls on savanna occurrence is still not well understood (Sarmieto 1983, Goodwin et al. 2013, Canché-Estrada et al. 2018). Canché-Estrada et al. (2018) demonstrated the importance and effect of local climate and physiographic factors on plant distribution in a comparative study of nine savanna sites across southern Mexico and Belize. Plants in the family Melastomataceae were absent in the savannas of the Yucatan and Oaxaca, but present in the Belizean savannas, a finding the authors attributed to differences in total annual rainfall. In their comprehensive vegetation survey of Belizean savannas, Goodwin et al.

(2013) also provided a summary of the development of classification systems that acknowledge the recent recognition of the heterogeneity between these diverse savanna ecosystems.

Beard (1953) provided some of the earliest descriptions of savanna vegetation across the northern Neotropics. He viewed Central American savannas as largely homogenous ecosystems that could be split into four broad phases: Open Savanna, Orchard Savanna, Palm Savanna, and Pine Savanna. The classification of Open savanna combined the Nanzal (dominated by *Byrsonima crassifolia*) and the Encinal (defined by stands of *Quercus oleoides*) savanna phases described earlier by Bartlett (1935). Bridgewater et al. (2002) expanded upon these classifications, identifying six distinct vegetation communities within the savanna landscape of the Rio Bravo Conservation and Management Area in northern Belize. In a survey of the Monkey Bay Wildlife Sanctuary in central Belize, Laughlin (2002) recognized three more distinct savanna types. These differences in nomenclature and floristics speak to the heterogeneity of Belizean lowland savannas.

Figure 5. Sparse stand of Caribbean Pine surrounded by Palmetto on the edge of Pine Pond. Photograph taken by Mathew Boehm, March 2018.

CHAPTER III

LITERATURE REVIEW

3.1 Central American Savannas

The origin and persistence of savannas across Central America has long been a topic of debate. Despite a lack of consensus, three general explanations for savanna occurrence dominate the literature: 1) they are Pleistocene relicts, from a time when the climate was cooler and drier; 2) they are the product of prehistoric deforestation, by the Maya in Belize and by other people elsewhere; or 3) they are controlled by edaphic or lithological factors (Leyden et al. 1998, Leyden 2002, Brenner et al. 2002, Beach et al. 2017). Several pollen and charcoal records from sites across the Maya lowlands document both sharp transitions between forests and grasslands and abrupt changes in fire regime during the Maya period that have been interpreted as resulting from changes in land use by humans rather than from abrupt changes in climate (Deevey 1979, Leyden et al. 1987, Islebe et al. 1996, Dull 2004a, Wahl et al. 2014).

At Pine Pond, the flat topography and the presence of acidic topsoil above an impermeable clay layer created hydrogeological conditions that favor savanna vegetation. However, fire may also have played a significant role in shaping the local ecosystem. Recent fires have been observed at the site (Figure 6) and some of the pine trees near the lake have charred lower trunks (Figure 5).

Several studies have been conducted on the complex interaction between fire, vegetation, and climate within the Mountain Pine Ridge ecosystem of Belize, an ecosystem found in the higher elevations of the Maya Mountains that is similarly characterized by poor, acidic soils (Kellman 1975, Kellman and Maeve 1997). These studies point to the importance and regular

Figure 6. Pine Pond following a fire in May 2017. Photo courtesy of Rick Anderson.

occurrence of lightning-ignited fires in the Mountain Pine Ridge ecosystem (Kellman 1984, Kellman et al. 1987, Kellman and Maeve 1997). However, no such studies have been conducted on any lowland savanna in Belize. There are currently no data on the frequency of lightning fires in the savannas of the southern coastal plain and no historic or prehistoric evidence that documents how commonly fires disturbed pine or broadleaf forests. In archival material not available to me, Cathy Smith (2019) documented only two recorded mentions of the sources of fire ignition on the southern coastal plain. When commenting upon fire occurrence in their annual report for 1963, the Forest Department noted that "19% were due to lightning and 81% were human in origin." Within a report attempting to assess the impact of a hurricane on fire occurrences three years later, Wolffsohn (1967) remarked that "in the coastal pine forests lightning fires are unknown." Modern conservation and management efforts in the savannas of the southern coastal plain are also characterized by purely anecdotal observations of fire and a belief in the primacy of human-driven fires in the long-term ecological development of the savannas the area (Smith 2019).

3.2 Savanna Conservation and Management

Despite their nutrient deficient soils, lowland Belizean savannas are home to nearly one-third of the country's flora and almost half of the country's endemic species (Mcloughlin 2015). These highly diverse, heterogeneous ecosystems are threatened by various anthropogenic activities: deforestation for agriculture and livestock, which alters fire regimes; increased, near annual burning in some areas for hunting; and impending threats due to accelerated climate changes (Mistry 2000, Young 2008, Cherrington et al. 2010). In the face of recent changes and threats, paleoecological reconstructions of past human and natural disturbances and subsequent ecosystem responses on different spatiotemporal scales are crucial for developing a more adaptive management framework. Such a framework uses a long-term range of variability (i.e. century to millennial), rather than recent reactions (i.e. decadal) (Hallet and Walker 2010, Dearing et al. 2012, Gillson and Merchant 2014, Dietl et al. 2015) to inform conservation management objectives. A major challenge of many neotropical environmental reconstructions from lake sediments is parsing out natural, climate-driven changes in vegetation and fire from those driven by humans, who may have altered the ecosystem around Pine Pond for centuries to millennia (Braswell and Prufer 2009, Robinson and McKillop 2013, Walsh et al. 2014). This thesis provides an empirical record of long-term fire and vegetation history in comparison with archaeological work, independent climate records, and collaborator Cathy Smith's recent political ecology work on fire management during the British colonial era (Smith 2019). Understanding these temporal changes of ecosystem development and fire disturbance is vital to inform conservation management plans that attempt to harmonize the competing forces of future climate change and cultural land-use activity.

Pine Pond is located on the western edge of the Deep River Forest Reserve, almost adjacent to the eastern boundary of Payne's Creek National Park (Figure 7). Given this location, the current management practices of both protected areas affect the ecosystem surrounding the pond. The fire return interval in areas of Payne's Creek is currently around one year or less due to illegal deer hunting, for which people set fires (Mcloughlin 2015). While *Pinus caribaea* is adapted to recurrent burning, with thick bark that protects against low-intensity fires and seedlings that require an open canopy to grow, seedlings of this savanna pine need at least one year of growth to avoid being killed by fire (Myers and Rodríguez-Trejo 2009). The park's current management plan recommends a prescribed burn schedule of every three to five years,

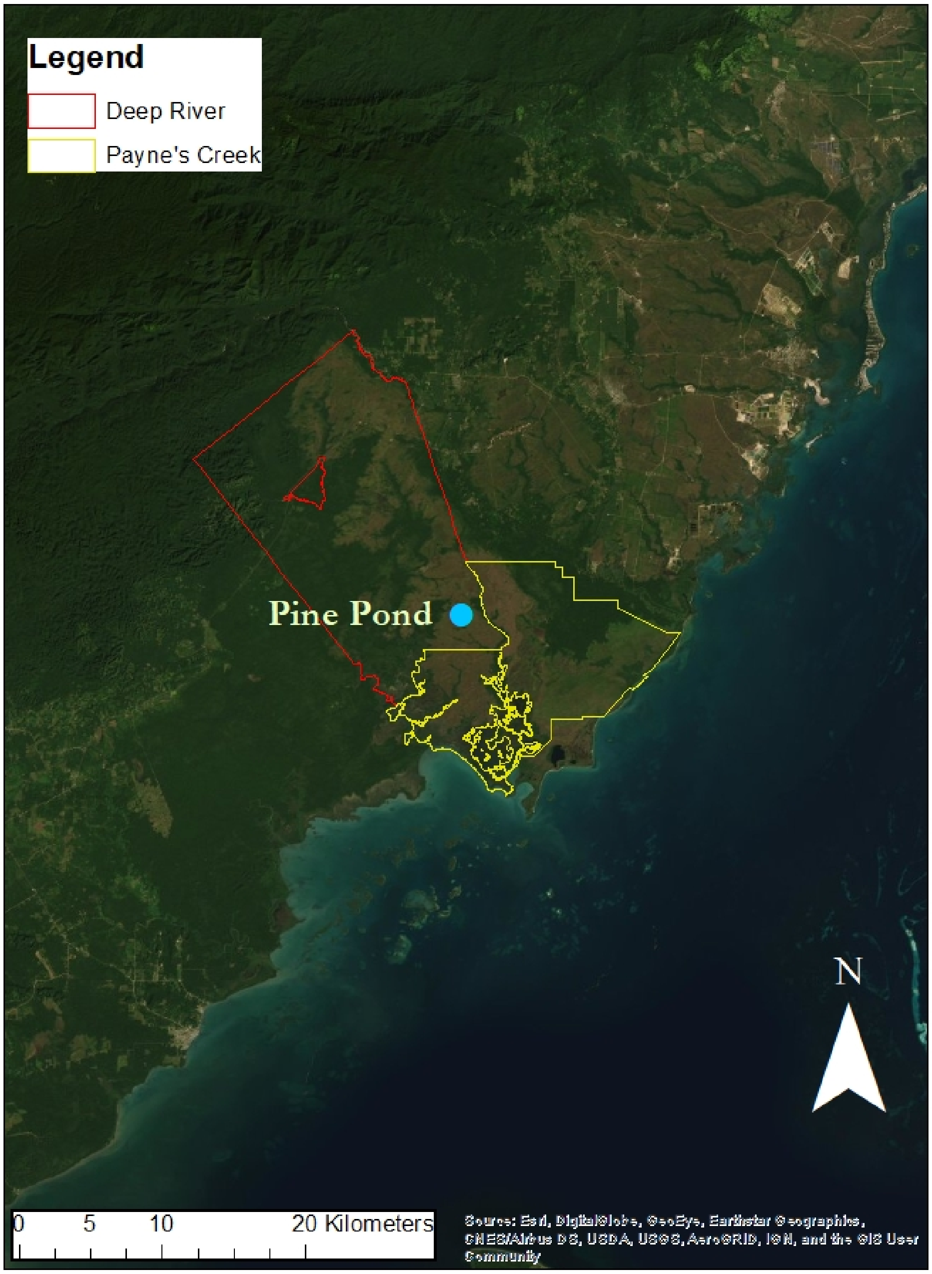

Figure 7. Map showing the location of Pine Pond in the Deep Forest River Reserve (outlined in red) and proximity to Payne's Creek National Park (outlined in Yellow). Created in ArcMap with the assistance of Bridgette Fritz, using shapefiles available at http://www.biodiversity.bz/.

with the idea that more frequent fires reduce pine density and threaten overall biodiversity (Mcloughlin 2015).

The management plan of the Deep Forest River Reserve also expresses concern that increasingly short fire return intervals from indigenous burning threaten the long-term health of the flora and fauna (Linares and Vinicio 2009). The Deep River Forest Reserve differs from Payne's Creek National Park in having specific commercial interests. Here the management plan supports a dual policy of suppression and controlled burning designed to create an ecosystem suitable for pine regeneration and harvesting (Linares and Vinicio 2009). Both plans are based on anecdotal and recent historical evidence and both castigate local indigenous burning without the long-term data on fire to support these interpretations and management objectives. Examining the effects of past changes in fire regimes may help inform more adaptive management policies.

3.3 Fires in Savannas

The flora of Central America included fire-adapted grasses, shrubs, and trees long before humans arrived near the end of the last glacial period (Horn and Kapelle 2009, Graham 2010, He et al. 2011). In the absence of human disturbance, fire regimes of tropical savannas—like ecosystems elsewhere—are governed by factors such as rainfall seasonality and drought duration, which affect vegetation growth and distribution (Gavin et al. 2007, Archibald et al. 2009, Power et al. 2010). A wetter climate may counterintuitively promote burning, as there is more available biomass to serve as fuel; conversely, a drought may actually limit burning due to decreased fuel (Nelson et al. 2012). Fire regime changes can induce positive feedback loops, where an open area of grasses encourages burning, while denser canopy tree stands create a wet microclimate that inhibits dry fuel build up (van der Werf et al. 2008, Archibald 2009, Colombaroli et al. 2014). Understanding changes in a complex climate-fire-vegetation dynamic is already difficult, and human land-use change further complicates these interpretations. Various paleoenvironmental studies from the prairies of North America and savannas in Africa indicate humans have for centuries to millennia controlled and maintained the distribution of savanna where forest might otherwise grow (Laris 2006, Nelson and Hu 2008, Archibald et al. 2009, Laris et al. 2011, Marlon et al. 2013, Colomboroli et al. 2014). Since their arrival, prehistoric peoples in Central America likely also engaged in burning practices that affected lowland tropical savanna ecosystems on timescales of centuries to millennia.

3.4 History of Ecological Thought on Development of Savannas in Belize

In her PhD dissertation on the political ecology of fire management during the British colonial era, Cathy Smith (2019) described the evolution of scholarly thought behind Belizean savanna development. Many early ecologists and foresters writing about Belizean savannas in the first half of the 20th century were influenced by the theories of Frederick Clements (1916, 1936) on ecosystem succession and climax. Savannas were viewed as degraded forests, with lowland broadleaf forest or pine forest being natural climaxes. By the 1950s, the importance of edaphic factors, either nutrient-poor soil or poor drainage or both, began to be emphasized, while the role of recurring fire was largely ignored. In the first comprehensive classification of land and soils for Belize, Wright (1959) noted the importance of both edaphic controls and recurring fire to explain savanna plant communities. Soon after, Belizean savannas were regarded as fire dis-climaxes, created by anthropogenic burning and deforestation. Over the last few decades, a greater appreciation for ecosystem heterogeneity has been incorporated into the literature. But lacking long-term empirical data, Clementsian views of equilibrium ecology still influence conservation efforts (Smith 2019).

In Payne's Creek National Park managers have focused on a three-to-five-year fire return interval as a management target (Mcloughlin 2015). This target is based on short-term anecdotal evidence, as well as recent work attempting to place current fire regimes of the Miskito Coast along the eastern coast of Honduras in the context of longer-term ecosystem development (Myers et al. 2006, Mcloughlin 2015). The long-term dynamics of fire, vegetation, climate, and human actions that influence the location of broadleaf forest, pine stands, and areas of more open savanna are still not well understood (Myers and Morrison 2006). While both the Deep River and Payne's Creek management plans acknowledge edaphic controls on vegetation, both plans express worry that a total absence of fire may lead to broadleaf forest succession. The Payne's Creek plan also discusses a long tradition of anthropogenically-driven changes in fire regimes, stating that fire frequency was highest during the Classic Maya era (250–900 CE) and the British colonial era. This thesis provides a long-term record of fire and vegetation history that may help inform these concerns and debates.

3.5 Local Archaeology and Relationships Between Prehistoric People and Pines

Humans have likely occupied the Maya region since the late Pleistocene, with archaeological and paleoenvironmental evidence revealing significant landscape alterations beginning around 3000 BCE that increased during the Archaic archaeological period beginning ca. 2000 BCE (Jones et al.1994, Pohl et al.1996, Coe et al.1999, Luzzadder-Beach et al. 2016a). Beach et al. (2015) dubbed the period between roughly 1000 BCE and 1000 CE the "Mayacene" due to the regional scale of landscape alterations affecting various ecological processes.

Population and agricultural activity peaked towards the end of the Classic Period (250–900 CE) before declining dramatically with a rapid abandonment of urban centers and agricultural structures during the Terminal Classic into the Postclassic, culminating by around 900 CE (Guderjan et al. 2009, Turner and Sabloff 2012, Valdez and Scarborough 2014, Beach et al. 2017). Current scholarly thought and paleoenvironmental evidence points to regional afforestation from this time onward because populations were smaller and less centralized (Graham et al.1989, Hodell et al. 2001, Luzzadder-Beach and Beach 2009, Mueller et al. 2010). There was, however, significant spatiotemporal variation in the impacts of continued Maya landscape alteration during the Postclassic (950-1539 CE) period (Anselmetti et al. 2007, Beach et al. 2008, Beach et al. 2017).

Archaeological evidence of prehistoric population growth and decline from the Belizean southern coastal plain generally follows the broad regional patterns that characterize the different Maya archaeological periods. Current evidence indicates the area was sparsely inhabited by people living in relatively small farming centers until the end of the Pre-Classic Period (250 CE) (Prufer et al. 2008, Braswell and Prufer 2009). The subsequent Classic period saw the growth of large, densely populated inland urban areas, with consequent environmental impact from agricultural activity and physical resource consumption that reached a zenith during the Late Classic (600–900 CE) (Hansen et al. 2002, Robinson and McKillop 2013, Walsh et al. 2014).

In archaeological work at an ancient saltworks in Payne's Creek National Park, Robinson and McKillop (2013) found salt production mirrored local population growth and decline, with both the salt works and many of the nearby large urban centers having been abandoned by the start of the Post-Classic period. Changes in the assemblages of taxonomically-identified wood charcoal at the site from predominantly coastal Black Mangrove (*Avicennia germinans*) in the

Early Classic (250–600 CE) to a more diverse array of broadleaf species in the Late Classic (600–900 CE) indicated local deforestation as production expanded. This taxonomic shift in the wood charcoal at the salt mines indicates deforestation of localized patches of lowland broadleaf forest or traveling across the savanna to reach the dense lowland forests further inlan. Recent archaeological evidence of Maya hunting sites suggests that increased activity within the savanna associated with the salt works may have coincided with increased burning for hunting (Smith 2019). While apparently avoided in saltworks production, pine was a culturally valued species in other contexts, based on paleobotanical studies at other sites in Belize. Pine remains have been found in the religious contexts of caves and tombs, and as the most popular source of construction material at numerous archaeological sites across a wide range of ecosystems in which pine does not now occur (Atran and Ucan Ek 1999, Morehart et al. 2005, Lentz et al. 2005, Wyatt 2008). These findings indicate either that pine was once present in ecosystems where it does not occur today, or that extensive trade took place as early as the Pre-Classic Period (Lentz et al. 2005).

3.6 Colonial Context (1800–1950 CE)

Coinciding with the abandonment of the salt works, major inland urban centers in the region experienced sudden depopulation at the end of the 9th century (Prufer et al. 2008). While it is probable some level of human occupation persisted in the region, little is known about human-land use patterns following the Classic Maya decline until the arrival of the British during the colonial era. When the Spanish arrived in the Caribbean and the Yucatan peninsula in the early 16th century, Maya populations in southern Belize likely became aware of their presence through extensive coastal trading among indigenous groups (Graham 2011). But apart from a few Franciscan churches in the north of the country, the Spanish did not establish major permanent settlements in Belize during their 16th and 17th century conquests in the Americas, despite claiming the territory as their own (Jones 1989, Graham 2011). Instead, the British were the first major European power to establish permanent settlement in Belize (Bulmer-Thomas and Bulmer-Thomas 2012). Although Belize did not become an official British colony until 1862, state-sanctioned resource extraction began long before. Extraction of the leguminous tree Logwood (*Haematoxylum campechianum*) was already taking place by the middle of the 17th century, and after a series of treaties through the 18th century, Spain gradually conceded British settlement rights along with rights to extract and export both logwood and mahogany (*Swietenia macrophylla*) (Bulmer-Thomas and Bulmer-Thomas 2012). However, resource exploitation during this time was largely confined to accessible areas in northern and central Belize, north of my study site of Pine Pond. At this time, Spanish concessions for legal settlement and logging only extended as far south as the Sibun River, which has its headwaters in the northern Maya mountains (Bridgewater 2012, Bulmer-Thomas and Bulmer-Thomas 2012; Figure 8).

By the middle of the 18th century, mahogany had surpassed logwood as the territory's main export, being highly valued in the furniture and ship-building industries (Pennington 2002). Despite the boundaries demarcated in the recent treaties with Spain, commercial-scale extraction of mahogany was reported as far south as the Sarstoon River, the present southern border of Belize, by the early 19th century (Smith 2019). The 19th century saw an increase in both capital investment and population across the southern coastal plain. Pre-nineteenth century, the European colonial population for entire territory never amounted to more than few hundred settlers (Bulmer-Thomas and Bulmer-Thomas 2012). An 1816 census recorded a total population of around four thousand (Smith 2019). By the middle of the 19th century, the British state began

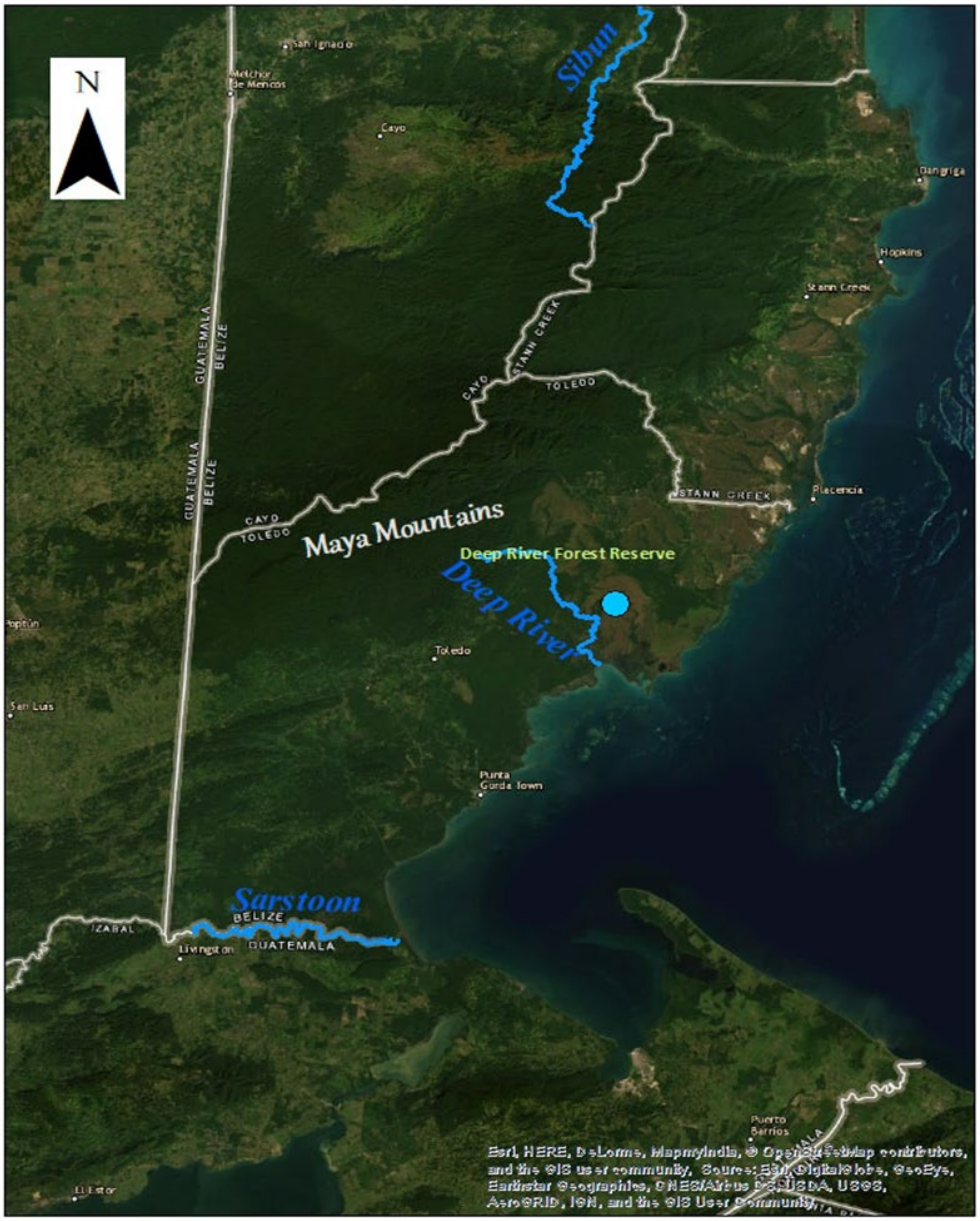

Figure 8. Reference map of southern Belize. Created in ArcMap with the assistance of Bridgette Fritz, using shapefiles available at http://www.biodiversity.bz/.

to actively support large-scale mahogany extraction, which required more land and labor than logwood cutting or commercial plantation-based agriculture, and led to labor importation from the Caribbean and China (Smith 2019). The indigenous population in Belize also dramatically increased during this time, as Mopan and Q'eqchi Maya sought to escape oppressive, exploitative conditions in Guatemala (Steinberg 2002, Wilk 1991). By the early 20th century, the British state established formal reservations based on settled agriculture (Wilk 1991). Indigenous Maya, European settlers, and their African slaves all employed some form of slash-and-burn agriculture in areas of broadleaf forest, but direct accounts are rare and anecdotal (Smith 2019). Descriptions of fire applications in savannas are also limited, but fire for hunting appears to have been common, and the burning of pine torches was seen as essential to warding of insects during the nocturnal harvesting of logwood and mahogany (Smith 2019). In the late eighteenth century, there was a burgeoning interest in pine extraction from savannas. In 1922 the state forestry department was established with the explicit goal of fire suppression to maximize pine stand density; this policy initially failed for a variety of reasons, but it laid the framework for substantial harvesting and exportation that peaked from 1948–1965 (Smith 2019).

3.7 Paleoenvironmental Context

Relatively few paleolimnological records that use both macroscopic charcoal and pollen to infer prehistoric land-use change are available for Central America as a whole (Nevle and Bird 2008), or for the greater Maya region, which extends from the Yucatan peninsula to northwest Honduras and encompasses the entire modern-day nation of Belize (Wahl et al. 2014, Walsh et al. 2014). Anderson and Wahl (2016) provided a comprehensive summary of charcoal work across the Maya lowlands. Most of the records quantified microscopic charcoal on pollen slides, which may document regional as well as local patterns of burning (Lafon et al. 2017). Only four of the studies they examined focused on macroscopic charcoal, which provides a local record of fire occurrence (Whitlock and Larsen 2001). Anderson and Wahl (2016) found that charcoal data in reconstructions from across the region do not always follow the same trends in the pollen data, which typically show alterations between intervals with high levels of forest taxa and intervals with high levels of herbaceous and disturbance taxa. While such discrepancies likely represent real variation in anthropogenic use of fire, comparisons across studies were hampered by differences in the methods used for quantifying charcoal. Difficulties in comparing records with limited chronological control also hampered the review by Anderson and Wahl (2016).

Only one paleolimnological study (Walsh et al. 2014) exists for southern Belize. Walsh and collaborators developed a 2800-year pollen and macroscopic charcoal record for the Agua Caliente wetland, located in an area of lowland broadleaf forest ca. 42 km southwest of Pine Pond. They found evidence of slash-and-burn agriculture beginning in the PreClassic ca. 2600 cal yr BP (650 BCE). Fire declined during the Terminal Classic and was lowest during the Post-Classic (ca. 950–450 cal yr BP or 1000–1500 CE). Fire activity then rose through the Colonial period, from 450 cal yr BP (1500 CE) to the present, coincident with increasing population in the region. Walsh et al. (2014) tied the results to the independent Yok Bolum Cave speleothem climate record and to inferred land use changes at the nearby archaeological site of Uxbenka to better understand spatiotemporal patterns of anthropogenic and natural changes in the environment.

3.8 Holocene Climate Shifts in Central America

Multiple paleoenvironmental reconstructions across Central America document long term (i.e. millennial scale) regional climate shifts during the Holocene driven by changes in ocean-atmospheric dynamics and shifts in solar insolation (Hodell and Curtis 1995, Hodell et al. 2001, Haug et al. 2001, Luzzadder-Beach et al. 2016b). Drivers across shorter time scales (i.e., centennial to decadal), are less well known (Wahl et al. 2013). Human impacts in the region over the past several thousand years complicate the identification and interpretation of climate dynamics (Horn 2007). Many paleolimnological studies across the Maya lowlands are based on pollen analysis alone. In such records, teasing apart climate and human impacts can be difficult or impossible. Where researchers do implicate climate changes, they may be looking at local changes in microclimate and/or hydrology, making regional climate extrapolations difficult (Leyden 2002, Wahl et al. 2013). There are also general issues of cross-study comparison inherent to the paleolimnological work due to limited radiocarbon dates and radiocarbon error ranges.

In 2001, Haug et al. described analyses of titanium and other materials in a marine sediment core from the anoxic Cariaco Basin off the coast of Venezuela. This study and other analyses of the Cariaco sediments have become standards for comparison, with Cariaco proxies interpreted as indicators of past precipitation across the circum-Caribbean region. Changing concentrations of titanium in the sediments have been used to estimate precipitation variability in northern Venezuela over the past 14,000 years, with high values indicating high rainfall. The Venezuelan rainfall has in turn been linked with long-term changes in the seasonal and annual positions of the ITCZ, with low Ti corresponding to a more southerly ITCZ position and lower levels of precipitation. Haug et al. (2003) documented an extended southerly annual displacement corresponding to a drying trend during the last 3000 years. Lowest total rainfall amounts were found between 1250 and 950 cal yr BP, a period known as the Terminal Classic Drought. The most consistent period of low rainfall totals corresponds to the Little Ice Age (ca. 500–100 cal yr BP). Evidence of both climate events is found in various paleolimnological reconstructions from other sites across the circum-Caribbean (Hodell et al. 2005, Lane et al. 2014).

Speleothem reconstructions provide more localized records of precipitation in Belize. These reconstructions are based on high resolution analyses of oxygen and carbon isotopes in speleothem calcite. Because of its karstic landscape, Belize is dotted with caves suitable for speleothem analysis. Kennett et al. (2012) provided an extremely high-resolution record of local precipitation change. Consistent with the general findings of Haug et al. (2001), the record showed a drying trend from 1340–930 cal yr BP with a period of extreme drought from 1130–1080 cal yr BP. The study directly correlated the timing to past climate conditions with different Maya prehistoric periods, noting that drought during the Late Classic period may have directly facilitated societal collapse (Kennett et al. 2012). Akers et al. (2016) reconstructed precipitation in the Maya Mountains of Central Belize and also found a major dry event between 1300 and 850 cal yr BP. While this study connected this broad climate event to the decline of Maya civilization, the authors posited that variability in rainfall during this time period rather than a broad drying trend had the greater societal effect, and noted that direct linkages between drought and societal disturbances may be an oversimplification and reductive view of Maya sociopolitical responses (Akers et al. 2016). That human societies are inextricably linked with their environment is indisputable, but pressures and responses due to climatic changes were not uniform across Maya societies. A palynological-based reconstruction by McNeil et al. (2010)

near the major Maya center of Copan in western Honduras pointed to high levels of resource management during a period of peak population, while a similar study by Lentz and Hockaday (2009) indicated the opposite around Tikal in northern Guatemala. This spatiotemporal discrepancy highlights the continued need for more localized paleoenvironmental reconstructions across the Maya region to better understand the timing and complexity of the effects of climate change on Maya sociopolitical developments.

3.9 Macroscopic Charcoal

For my macroscopic charcoal analysis, I used a 1-cm contiguous sampling scheme to provide a detailed, high-resolution record that would bring to light patterns undetectable with wide-interval sampling. I chose to focus only on macroscopic charcoal analysis, as microscopic charcoal may blow in to the lake from many kilometers away, and thus potentially include a regional as well as local fire signal (Pisaric 2002, Tinner et al. 2006, Lafon et al. 2017). While high-resolution macroscopic charcoal analysis is the best method for reconstructing a long-term record of local fire history, several constraints limit interpretation. First, there is a potential lag between a fire event and the subsequent deposition of charcoal in the lake sediment. Charcoal may sit on the land surface for decades or longer before washing into a lake, reducing the resolution of the record (Millspaugh and Whitlock 1995, Duffin et al. 2008, Conedera et al. 2009). Second, changes in charcoal abundance in sediments, expressed as either concentration (fragments/cm^3) or influx (fragments/cm^2/year), could reflect a combination of changes in fire frequency (how many fires burn in a given interval) and changes in fire intensity (temperatures reached during fires). For savannas specifically, various studies implicate fire intensity as the most important control on limiting tree invasion, rather than the just the frequency of burning or total burned area (Kuhlbusch and Crutzen 1995, Duffin et al. 2008, Colombaroli et al. 2014). But fire intensity cannot be separated from other drivers of charcoal abundance in sediments as determined in macroscopic charcoal analysis.

3.10 Pollen Analysis

One of the main strengths of paleolimnology is the ability to reconstruct vegetation history using palynology. Pollen and spores from various plants are blown or washed into a lake, where they are then preserved indefinitely with the accumulation of lake sediments because of anoxic conditions (Faegri and Iversen 1989). Pollen types can be identified by differing shapes, and by textures and patterns on the surface of the grain (Moore et al. 1991). By knowing the current environmental requirements of a given taxon, its presence or absence in sampled increments from the sediment profile can used to infer past environmental changes (Birks and Gordon 1985).

There are, however, certain limitations to using pollen analysis alone to infer the causes and timing of past environmental changes. Lake sediment cores are often taken from a centralized, deep area of a lake. This can create an averaging effect regarding the inflow and deposition of pollen grains and fern spores (Larsen and MacDonald 1994). Different plant taxa have different rates of pollen production and dispersal, and their proximity to the lakeshore may affect whether a particular taxon is represented in the record (Bennett and Willis 2001). Another significant limitation of pollen analysis is the ability to confidently discern shifts in vegetation caused by climate changes as opposed to changes resulting from human impact. The utility of pollen analysis for reconstructing climate shifts on long-term scales (i.e. over the entirety of the Holocene) across the Maya lowlands is well documented (Leyden 2002, Brenner et al. 2002, Walsh et al. 2014). On shorter scales of decades or centuries, human land use change may mask any climate signal. Shifts in pollen frequencies because of human activity are often abrupt. For example, a sharp increase in pollen of Asteraceae, a plant family with many weedy species associated with human disturbance and deforestation in tropical savannas and other ecosystems, may indicate anthropogenic land-use change (Faegri and Iversen 1989, Dull 2004b). Such a shift, however, might also be caused by the alteration of fire regimes due to a discrete disturbance such as a hurricane or a longer-term climatic shift such as drought (Marlon et al. 2013). Human impact hinders the ability to make simple, straightforward inferences about a predominantly dry or wet climate based only on pollen percentages or ratios of forest to non-forest taxa (Bush et al. 2002, Colombaroli et al. 2014). Despite the difficulty of inferring climate from pollen, palynological evidence still provides unparalleled information on changes in vegetation.

3.11 Stable Carbon and Nitrogen Isotopes

In the tropical ecosystems of Central America, stable carbon isotope analysis of lake and wetland sediments has been used to document shifts in relative proportions of C_3 and C_4 vegetation (Lane et al. 2004, Beach et al. 2009, Kerr et al. 2019). Globally, almost all plants use the C_3 photosynthetic pathway (Sage et al. 1999). Certain grasses, sedges, and other tropical plants have developed an evolutionary adaptation of extra carbon storage during photosynthesis known as the Hatch-Slack (C_4) pathway to cope with water shortage in warm, bright, dry environments (Johnson et al. 2007). These different photosynthetic pathways leave different carbon isotopic signatures in living plant matter, which are preserved in decayed plant matter that washes into lakes. C_3 plants preferentially use the lighter atmospheric ^{12}C isotope during photosynthesis. This produces $\delta^{13}C$ values ranging from –32 to –20 per mil, with tropical trees averaging around –30 per mil (O'Leary 1988, Luzzadder-Beach et al. 2016a). C_4 species do not discriminate as strongly between carbon isotopes and have a $\delta^{13}C$ range of –17 to –9 per mil (O'Leary 1988). Based on these isotopic differences, shifting organic $\delta^{13}C$ values in a sediment core may be used as a proxy to infer temporal shifts between C_3 (forest) and C_4 (grassland) vegetation in the watershed. Shifts between C_3 and C_4 vegetation may indicate a change in the ecosystem driven by climate or by human land use.

Multiple confounding variables might affect any inferred temporal shifts in C_3-C_4 vegetation within the ecosystem surrounding Pine Pond. Climatically, variability in temperature, aridity, and CO_2 levels are the main controls that govern C_4 plant distribution (Ehrlinger et al. 1997, Huang et al. 2001). Shifts between C_3 and C_4 signals in a sediment record may indicate a climate-driven change in ecotone boundary. Human land-use changes, however, may complicate this signal. Deforestation not only increases $\delta^{13}C$ values, but also alters the local microclimate, in turn creating a feedback loop of sunnier, drier conditions that promote the expansion of C_4 grasses (Taylor et al. 2013). The presence of autochthonous as well as allochthonous carbon in lake sediments further complicates the task of attributing shifts in $\delta^{13}C$ to climate or human influence.

Interpreting vegetation shifts from C_3 to C_4 values requires understanding the variability of productivity rates and sources of organic carbon. Changes in past lake productivity affect the $\delta^{13}C$ of sedimentary organic matter in profiles (Talbot 2001). Higher levels of organic matter may equate to higher levels of aquatic plant productivity (Meyers and Lallier-verges 1999). Conversely, autochthonous carbon from algae and aquatic plants may decrease $\delta^{13}C$ values (Lane et al. 2008). Analysis of C/N ratios in sedimentary organic matter can provide information on carbon sources that facilitates the interpretation of ^{13}C values for bulk sediment. Aquatic organic matter, due to high protein and a lack of cellulose, can be expected to exhibit lower carbon/nitrogen (C/N) ratios compared to plant-based organic matter derived from terrestrial inputs (Meyers 2001). In general, low C/N ratios likely reflect lower terrestrial input, while higher C/N ratios reflect a greater input of terrestrial organic matter. The nitrogen isotope composition of bulk organic matter in lake sediments is difficult to interpret (Talbot 2001). Changes in this composition may provide evidence of possible changes in lake chemistry and productivity triggered by changing climate or intervals of more frequent or intense fires, but other factors can also affect these values (Talbot 2001). Fires lead to nitrogen depletion in soil as ^{14}N is preferentially released to the atmosphere, resulting in more positive stable nitrogen isotope (^{15}N) ratios in remnant organic matter (Turekian et al. 1998, Torres et al. 2012). Given the current cultural practices of burning for hunting and for harvesting pine for construction, prehistoric people also may have engaged in large-scale burning and deforestation. Large influxes of charcoal should correlate with increased C/N ratios (Nave et al. 2011) and possibly with more positive nitrogen isotope ratios (Dunnette et al. 2014).

CHAPTER IV

METHODS

4.1 Sediment Coring

Sediment cores were collected from Pine Pond in March 2018 by Cathy Smith, then a doctoral student at the University of Edinburgh, and University of Tennessee graduate students Mathew Boehm, Bridgette Fritz, and Lizzie MacLennan. Sally Horn and Cathy Smith planned and obtained funding and permits for the coring, which was originally to be led by Sally Horn. Because of a last minute medical issue, Sally Horn could not travel and Mathew Boehm went to Belize in her place to lead the coring. Matt and his team recovered piston sediment cores from two positions in the lake, using a Colinvaux-Vohnout (C-V) locking piston corer (Colinvaux et al. 1999) operated from an anchored floating platform. Both cores were collected in one-meter increments from water depths of about 7 m. Core 1, about 2 m long, was collected closer to shore than Core 2, which was about 6 m long. Two mud-water interface (MWI) cores were collected at core site 1 using a plastic tube fitted with a rubber piston. MWI 1 was 84 cm long and MWI 2 was 30 cm long. The MWI cores were extruded in the field in 2 cm increments, and the sediments were placed in plastic bags for return to the lab. The C-V core sections were returned to the lab in their original aluminum coring tubes. Upon arrival, all sediment samples were stored in a cold room at 5–6 °C.

4.2 Core Cutting, Description, and Sampling

Back in the lab, Mathew Boehm, Cathy Smith, and I opened each aluminum core tube lengthwise using a modified wood shaper, and cut the sediments using a thin wire, dividing each core section into two halves. We then photographed each core section to record sediment color and stratigraphy. Cathy Smith and I examined the core sections, making notes on stratigraphy, texture, Munsell color, and the presence of plant macrofossils suitable for radiocarbon dating. One half of each core section was sampled for radiocarbon dating, loss-on-ignition, macroscopic charcoal, pollen, and bulk stable carbon and nitrogen isotope analyses.

4.3 Radiocarbon Dating

Sally Horn, Cathy Smith, and I removed seeds, charcoal, wood, and leaf fragments from the core sections, rinsed them in distilled water, and dried them at 50 °C in preparation for radiocarbon dating. Thirteen samples were submitted to university and commercial labs for radiocarbon dating, three from Core 1 and ten from Core 2. Dates were calibrated using the IntCal13 ^{14}C curve (Reimer et al. 2013) and the Clam modeling package in the statistical software program R (Blaauw 2019). Linear age-depth models for each core were produced in Clam by Mathew Boehm using depth models developed by Sally Horn. The depth model for Core 1 was based on the coring notes and the field calculation that the core began 31 cm below the mud-water interface. The depth model for Core 2 followed the field calculation that this core began 50 cm below the mud-water interface but included a one-meter upward adjustment in the depth of the second core section recovered. Following the first core drive covering the interval of 50–150 cm below the MWI, the piston jammed in the corer, and no sediments were recovered in drive 2. The raft was repositioned ca. 0.5 m to the side for drive 3, to avoid the disturbed sediment. While the intent of drive 3 was to sample the 150–250 cm interval in the new position, the watery nature of the sediments at the top of this section and the radiocarbon dates indicate that the drive began sooner, overlapping the depths for drive 1. The readjustment of drive 3 resulted in a gap of ca. 1 m in Core 2.

4.4 Charcoal Analysis

I carried out high-resolution (1-cm interval) sampling and analysis of macroscopic charcoal on all sections of Cores 1 and 2 and on the fifteen samples from the MWI 2 core, assisted by Cathy Smith. I used a brass sampler designed to take rectangular prismatic samples to remove contiguous samples of 2 cm^3 volume from the piston core sections. For the MWI samples, which were extruded in the field in 2 cm intervals, I sampled using a plastic spoon with a volume of 2.46 cm^3 (one half teaspoon). Following procedures outlined in Schlachter and Horn (2010), each sample was soaked overnight in 50 mL of 3% hydrogen peroxide. The peroxide helped to disaggregate the samples and to bleach some of the non-charred organic matter. Samples for Core 1 (n= 186), Core 2 (n= 581), and MWI 2 (n=15) were sieved using a 250 µm mesh sieve. Material retained on the sieves was transferred with distilled water to Petri dishes. All Petri dishes were then dried at ca. 50 ºC until free of water. Charcoal fragments from each sample were counted under a dissecting microscope. Charcoal was identified based on its black color, angular shape, and distinctive sheen.

Macroscopic charcoal counts from Cores 1, 2, and MWI 2 were combined to form a composite charcoal diagram. Counts from Core 1 were used to fill the gap in Core 2 between ca. 1300 and 600 cal yr BP, and to extend the record from the top of Core 2 at ca. 62 cal yr BP to 31 cal yr BP. Charcoal values from MWI core 2 were used to extend the record from 31 cal yr BP to the present. The charcoal concentration and influx values for Core 1 and MWI 2 were adjusted using a comparison of the values for Core 1 and Core 2 for the time span from 66 to 525 or 526 cal yr BP. The average values for charcoal concentration and influx in Core 2 during this period were divided by those in Core 1, and the resulting ratios were then multiplied by the Core 1 values to obtain adjusted values. This correction was also applied to the MWI 2 charcoal data, as this core was collected close to Core Site 1 and was assumed to also show inflated values resulting from the closer proximity to shore of Core Site 1.

4.5 Pollen Analysis

Pollen samples of 0.5 cm^3 were prepared from 28 levels of Core 2 and 2 levels of MWI Core 2, following a protocol used for other tropical lake-sediment samples in our lab (Appendix 1). Pollen slides were counted to a minimum of 250 pollen grains exclusive of spores and indeterminates at 400x magnification using a Leitz Laborlux microscope. To aid in the process of identifying unknown palynomorphs, I captured images using a digital camera.

Pollen grains were identified by comparison with reference pollen samples, pollen atlases, and keys, and by consulting with Sally Horn and Mathew Boehm. Grains were identified to the lowest taxonomic level possible, but with a focus on the most common types in the samples. One group that was difficult to separate was the order Urticales (including *Moraceae*, *Urticaceae*, and Ulmaceae). Many of the pollen grains produced by plants in the order are small (about 15 um), psilate to scabrate grains that vary in pore number (diporate to polyporate). I distinguished the genera *Trema* and *Cecropia,* and counted all other grains as “Other Urticales”. I found it difficult to distinguish *Byrsonima crassifolia* from *Rhizophora mangle* (red mangrove) and grouped them together. Fern spores were classified by morphology (monolete and trilete). Pollen and fern spore counts were entered in an Excel spreadsheet to calculate percentages. Pollen percentages were calculated as a percentage of the total pollen count (excluding indeterminate pollen grains and fern spores) for each level (i.e. percentage = (raw count/ total sum) x100). Fern spores percentages were calculated as a percentage of the total count of pollen grains and spores. Pollen and spore percentages and selected pollen ratios were plotted using the C2 software developed by Juggins (2010).

4.6 Loss on Ignition and Bulk Stable Isotope Analysis

For loss on ignition (LOI) analysis, samples of 1 cm^3 volume were taken at the same 30 levels initially sampled for pollen in Core 2 and MWI 2. Following Dean (1974), samples were dried overnight at 100 °C to determine water content, and then ignited at 550 °C and 1000 °C to estimate organic content and carbonate content, respectively. Samples of 2 cm^3 volume for bulk stable carbon ($\delta^{13}C$) and nitrogen ($\delta^{15}N$) isotope analyses were also taken at the same levels sampled for pollen. Preparation followed our standard laboratory protocol (Appendix 2), which includes acid fumigation for samples used for carbon isotope analysis. After final drying, samples for $\delta^{13}C$ and $\delta^{15}N$ analysis were placed into tin capsules and shipped to the Stable Isotope Laboratory at the University of North Carolina Wilmington. There they were analyzed with a Costech 4010 elemental analyzer interfaced with a Thermo Delta V Plus Mass Spectrometer. All samples were randomized in the trays relative to their linear position by depth in the core profile, and all samples were analyzed in duplicate to produce an average value for each sample point. Carbon and nitrogen isotope values are expressed in the standard δ-per mil notation with nitrogen values relative to atmospheric air and carbon values relative to the Vienna-Pee Dee belemnite (V-PDB) marine carbonate standard:

$$\delta^{13}C \text{ (per mil)} = 1000\,[(R\text{sample} / R\text{standard}) - 1]$$

$$\delta^{15}N \text{ (per mil)} = 1000\,[(R\text{sample} / R\text{standard}) - 1]$$

Carbon isotope values were derived from the acidified portion of each sample, while nitrogen isotope values correspond to the non-acidified fraction. The analyses of stable carbon and nitrogen isotopes also provides data on carbon and nitrogen values in the samples. These values were used to calculate C/N ratios. I used the mean percent organic carbon from the acidified fractions and the mean percent nitrogen from the non-acidified fractions to calculate these values.

CHAPTER V

RESULTS

5.1 Stratigraphy and Chronology

The two cores from Pine Pond exhibit little stratigraphic variation through the profiles, with each core section consisting of relatively homogenous, fine-grained organic sediment that varies in color from very dark grey (Munsell color 10 YR 3/1; 2.5Y 3/1) or greyish brown (10 YR 3/2; 2.5Y 3/2) to black (10 YR 2/1; 2.5Y 2/1). Scattered small plant fragments were noted during core description, but the cores are mainly fine-grained sediments without abundant visible macrofossils. The exception is an interval of very coarse organic material including wood fragments, leaves, and seeds present in Core 2 from 505–510 cm (Figure 9).

The LOI analysis of Core 2 showed relatively high-water content (62–92%), with highest values in the upper part of the profile (Figure 10). The sediment organic content ranges from 15 to 36% on a dry mass basis, and is also higher toward the top of the profile. Estimated carbonate percentages based on loss on ignition at 1000 °C are all below 3%. Values this low in lake sediments could be entirely the result of the loss of interstitial water in clays (Horn et al. 2018), as in Dean (1974).

5.2 Radiocarbon Dates and Age-Depth Models

The dates obtained on Cores 1 and 2 are in stratigraphic order (Table 1), with one exception. The exception is the date on a small piece of charcoal from near the top of Core 2

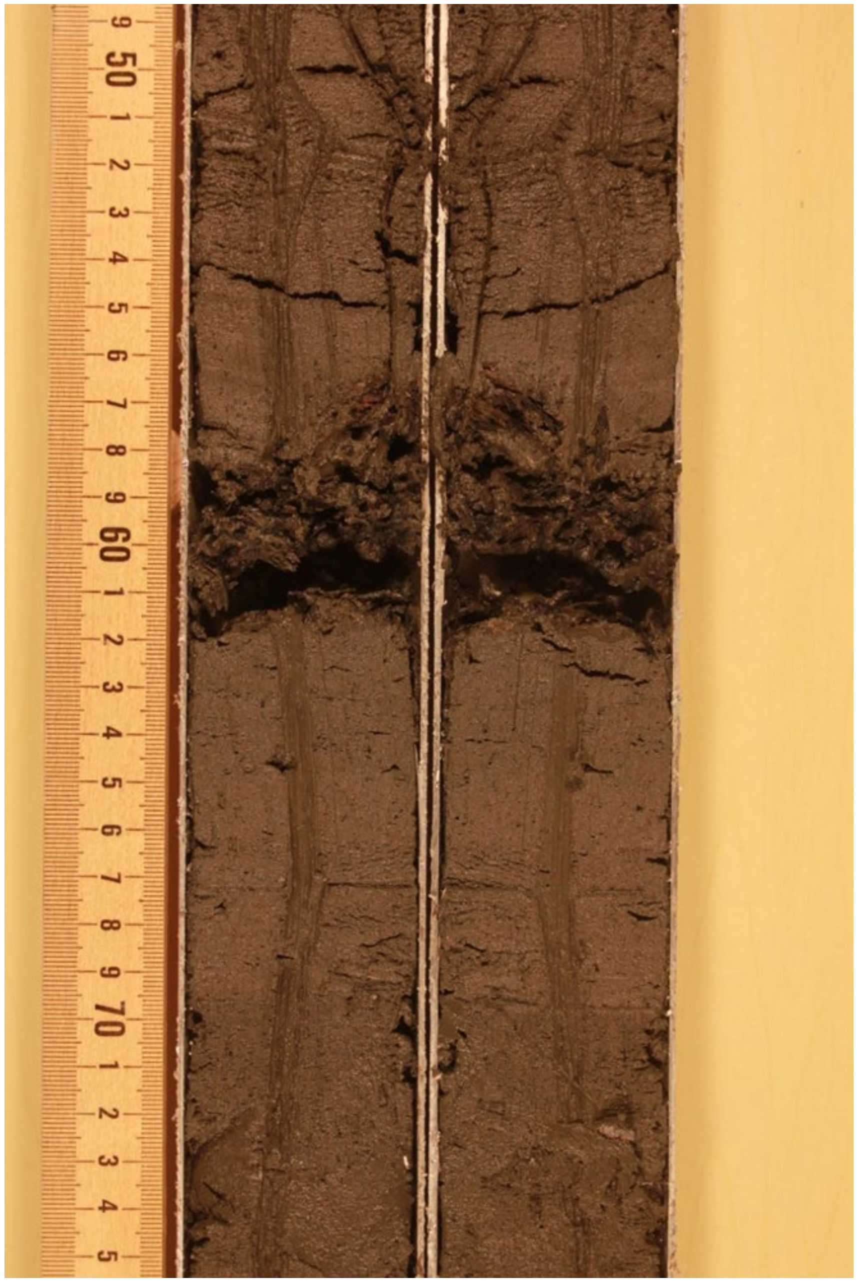

Figure 9. Interval of coarse organic matter from 505–510 cm in Pine Pond core 2.

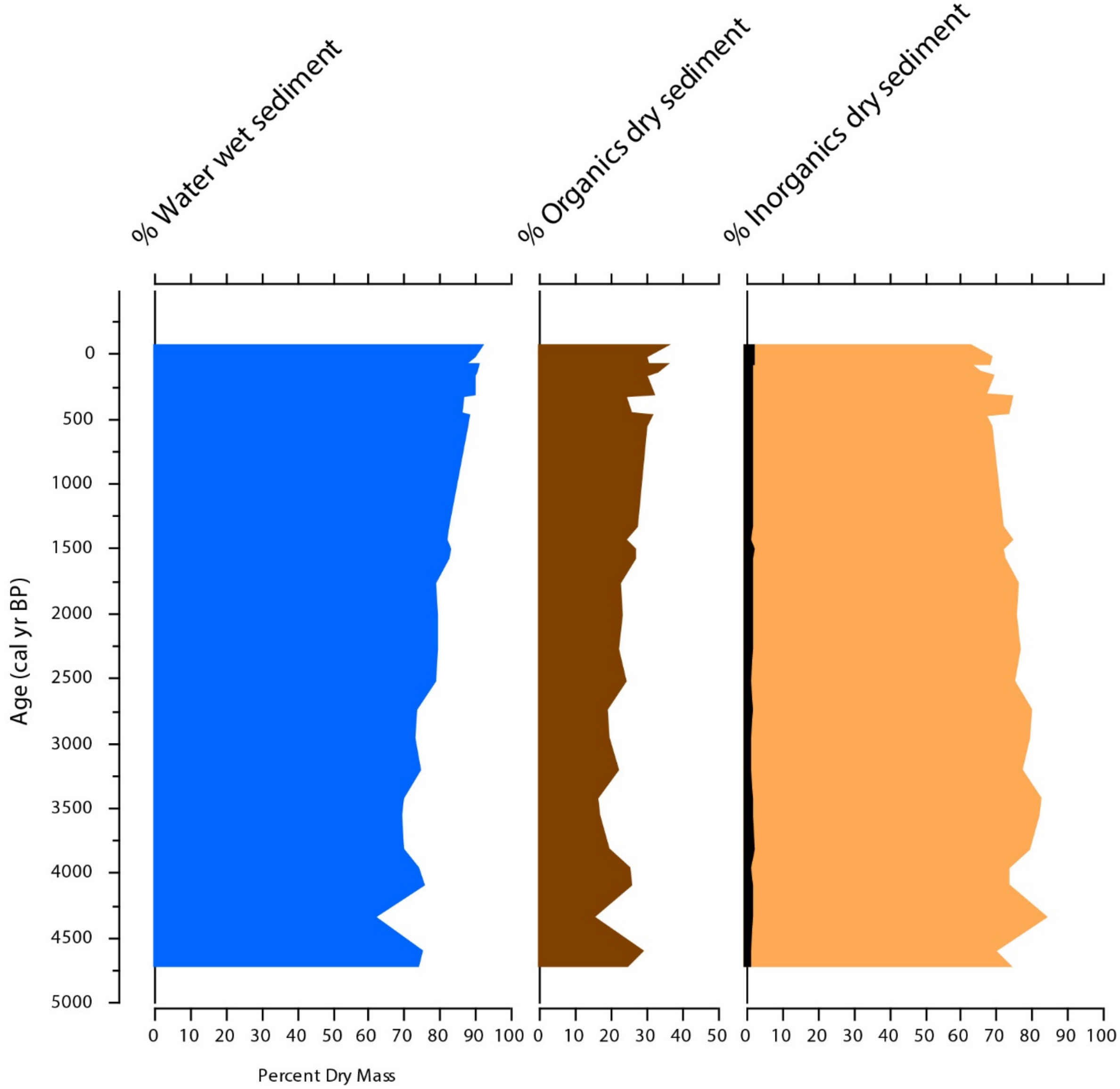

Figure 10. The water, organic, and inorganic content of Core 2, plotted by age (cal yr BP). The black curve within the inorganic curve is the estimated carbonate percentage.

Table 1. Radiocarbon determinations for Pine Pond Cores. Core 1 (top) and core 2 (bottom).[a]

Lab Number[b]	Core section/ interval	Depth (cm)	Material	δ^{13}C	Uncalibrated ^{14}C Age (^{14}C yr BP)	±2σ Range (cal yr BP)	Probability
D-AMS 033793	31–131/39.1	70.1	Leaf fragments		152 ± 25	36–0 118–68 125–125 154–131 231–167 249–244 283–251	17.7 14.7 0.1 10.7 35.4 0.8 15.51
OS-141176	31–131/74–76	106.0	Leaf	–29.0	370 ± 20	377–322 499–427	32.6 62.3
OS-141463	131–231/78.1	209.1	Plant fragments	–28.04	1250 ± 20	1108–1090 1135–1128 1158–1147 1271–1172	2.6 0.9 2.0 89.3
AA-113444	50–150/22–23	72.5	Charred plant fragment	-27.3	345 ± 75	167–154 519–283	1.4 93.5
OS-141171	150–250/23–24.5	73.8	Leaf fragments	–30.26	155 ± 20	33–2 79–73 104–81 114–106 154–136 224–168 283–254	18.6 1.2 5.2 2 11 41 15.9
OS-141177	50–150/42.0	92.0	Plant fragments	–29.26	215 ± 20	12 to –4 187–149 303–271	15.6 44.6 34.8
OS-144813	150–250/87.5	137.5	Leaf fragments & charcoal	–28.96	500 ± 15	537–512	95
UGAMS-42266	250–350/10	260	Wood fragment & charcoal	–26.5	1490 ± 20	1408–1331	95
OS-144814	250–350/21.5	271.5	Charred seed	–28.0	1600 ± 20	1467–1415 1544–1475	50.1 44.7
Beta-493701	250–350/51.75–53	301	Leaf fragment	–29.3	1710 ± 30	1639–1554 1698–1643	63.3 31.5

Table 1 Continued. Radiocarbon determinations for Pine Pond Cores. Core 1 (top) and core 2 (bottom).[a]

Lab Number[b]	Core section/ interval	Depth (cm)	Material	δ^{13}C	Uncalibrated ^{14}C Age (^{14}C yr BP)	±2σ Range (cal yr BP)	Probab ility
OS-141219	350–450/36.3	386.3	Leaf fragments	–29.0	2450 ± 20	2539–2363 2570–2568 2617-2587 2699–2632	51.3 0.3 11.4 31.9
OS-144815	450–550/60	510	Charred twig	–27.82	3390 ± 20	3649–3580 3691–3659	66.4 28.5
Beta-493700	550–650/88.25	617.3	Charcoal	–27.0	4200 ± 30	4639–4628 4763–4641 4790–4790 4842–4797	2.0 66.3 0.1 26.5

[a] Dates calibrated using the 'clam' package (v. 2.3.2; Blaauw 2019) for the R Statistical Environment (v. 3.5.2; R Core Team 2019) and the IntCal13 radiocarbon calibration curve (Reimer et al. 2013).

[b] Analyses performed by Direct AMS (D-AMS #), the National Ocean Sciences AMS Laboratory (OS #), The Accelerator Mass Spectrometry Lab at the University of Arizona (AA #), Beta Analytic, Inc (Beta #), and the Center for Applied Isotope Studies at the University of Georgia (UGAMS #).

that appears to represent redeposited, older material and that was excluded from the age model. This data has a large error range because it was run on a very small sample mass. The age models developed using the Clam modeling package indicate Core 1 extends to ca. 1350 cal yr BP (Figure 11) while Core 2 extends to 4800 cal yr BP (Figure 12). The models indicate that the two piston cores began at ca. 29 cal yr BP or 1921 CE (Core 1, 31 cm below MWI) and ca. 65 cal yr BP or 1885 CE (Core 2, 50 cm below MWI). Core 2 has a gap between ca. 1300 and 600 cal yr BP. Core 1 fills this time gap in Core 2, and the two MWI cores fill the time gap between the piston cores and the surface sediments. Based on the age model for Core 1, MWI 1 covers the time period from ca. 257 cal yr BP (1693 CE) to ca. -68 cal yr BP or 2018 CE, and MWI 2 covers the time period from ca. 26 cal yr BP (1924 CE) to -68 cal yr BP or 2018 CE.

The output from the Clam model indicates similar sedimentation rates for Cores 1 and 2, by depth where they overlap (Figure 13) and in general. From the bottom of Core 1 (Figure 11: 227 cm below the MWI, ca. 1347 cal yr BP) to 70 cm (ca. 152 cal yr BP), sediment accumulated at the rate of 0.13 cm/year. From 70 cm to the surface, sediment accumulated at the rate of 0.31 cm/year. From the base of Core 2 (Figure 12: 629 cm, ca. 4857 cal yr BP) to 300 cm (ca. 1616 cal yr BP), sediment accumulated at rates of 0.09, 0.10, and 0.11 cm per year. Between 300 and 271 cm (1470 cal yr BP) the age depth model indicates a short interval of more rapid sediment accumulation, followed by a return to a rate of 0.11 cm/year between 271 and 260 cm (1370 cal yr BP). The modeled sedimentation rates increase from 260 cm to the surface, from 0.11 cm below 260 cm to 0.14 cm/yr between 260 cm and 137 cm (ca. 522 cal yr BP), 0.16 cm/year from 137 cm to 74 cm (ca. 129 cal yr BP), and 0.38 cm/yr above 73 cm.

The similarity in radiocarbon dates (Table 1) and modeled sedimentation rates between Core 1 and the upper part of Core 2 (Figure 13) supports the depth model for Core 2, which

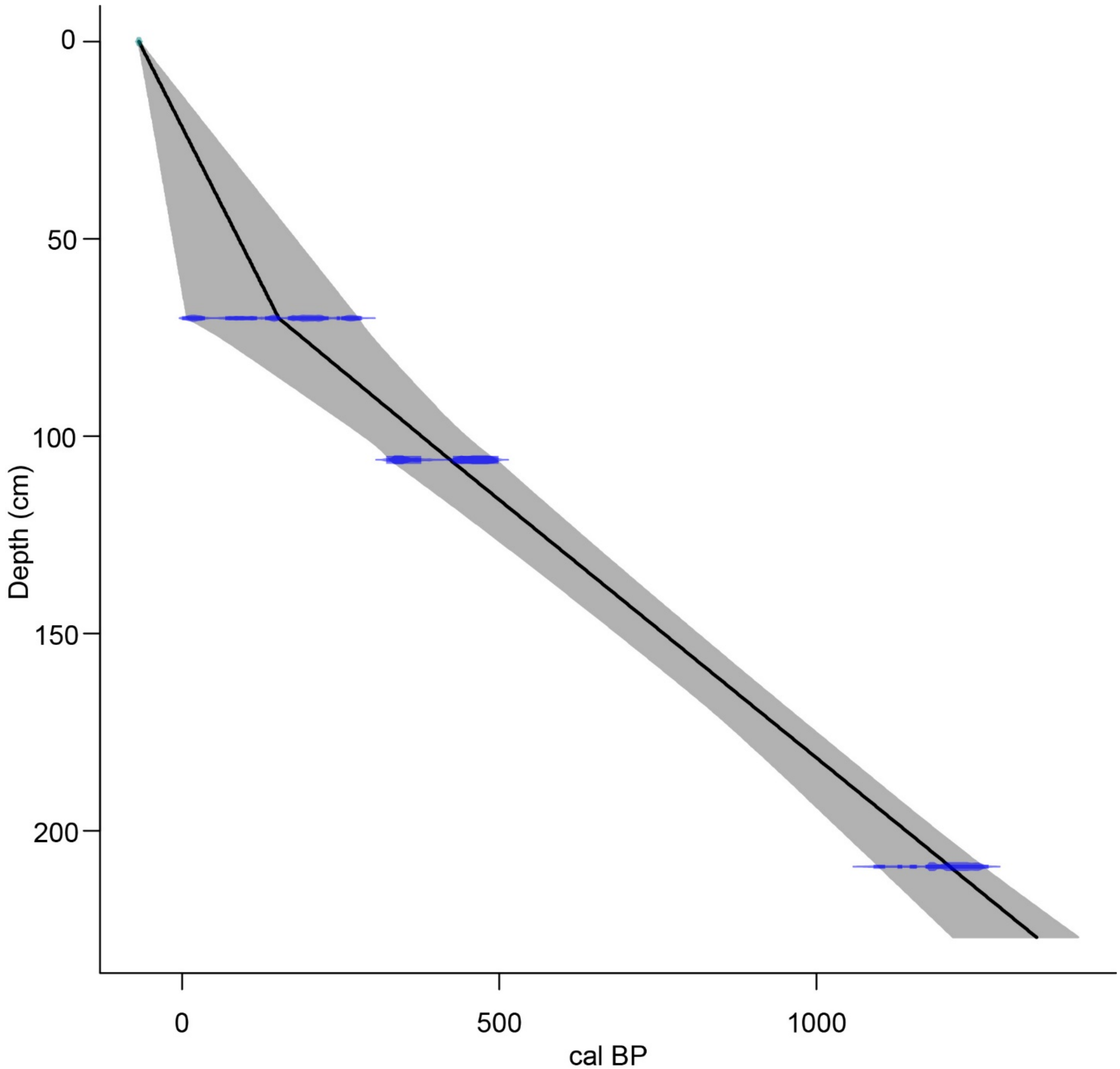

Figure 11. Age-Depth model for Core 1 made using the Clam modeling package (Blaauw and Christen 2011).

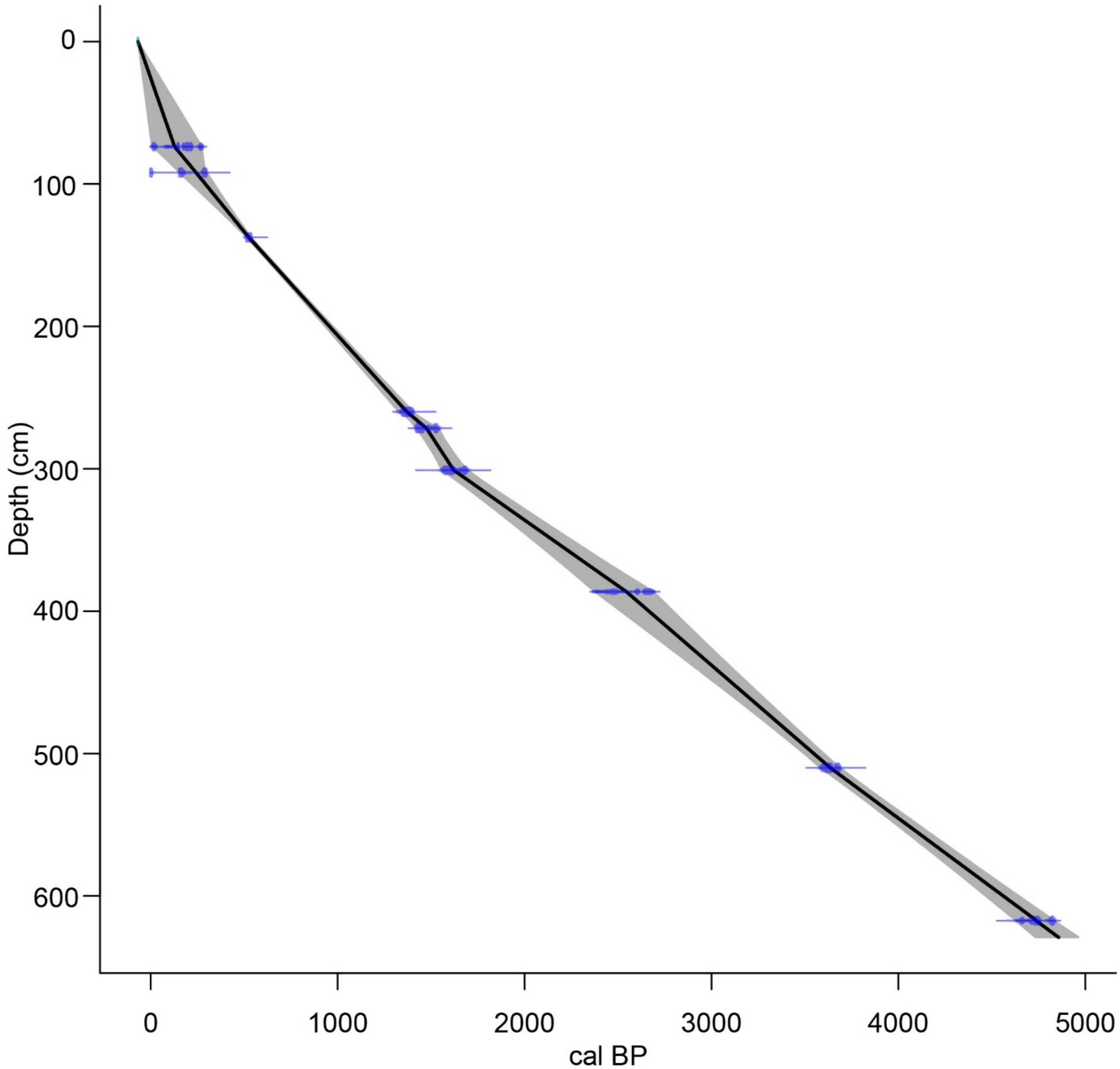

Figure 12. Age-Depth model for Core 2 made using the Clam modeling package (Blaauw and Christen 2011).

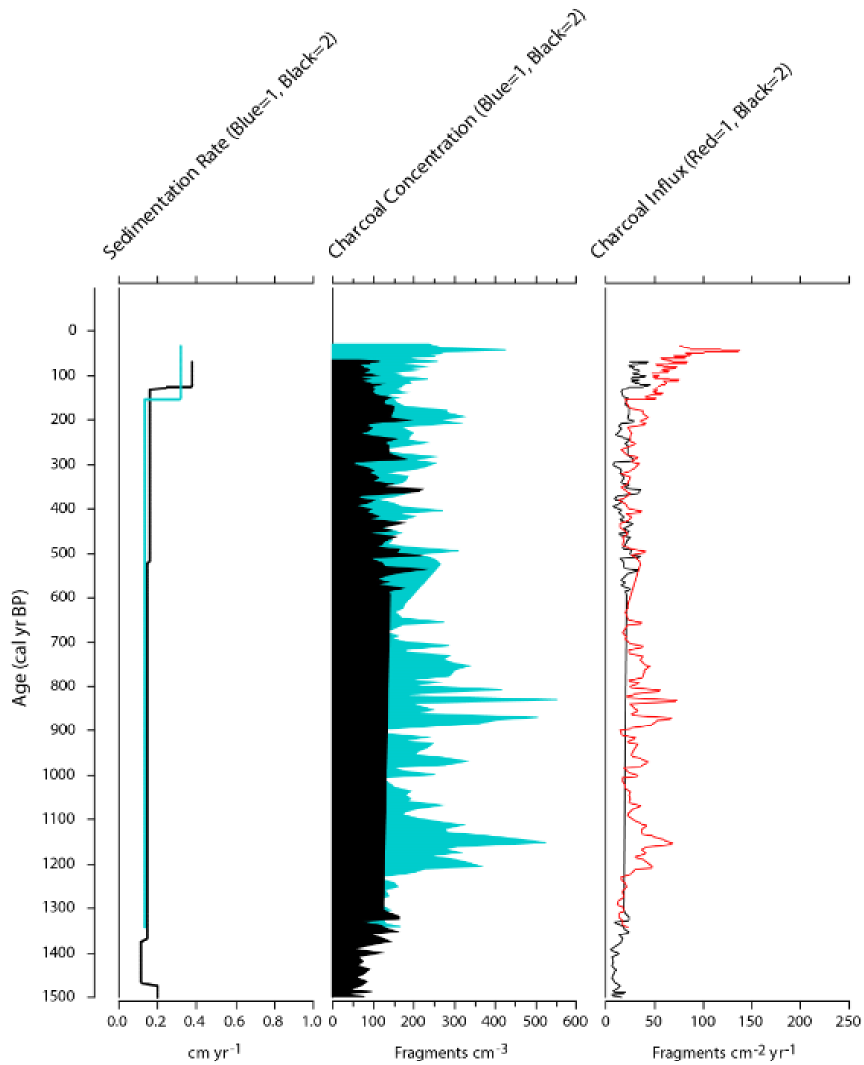

Figure 13. Comparison of sedimentation rates, charcoal concentration, and charcoal influx between Core 1 and Core 2 over the last 1500 cal yr BP. The blue lines and shading and the red lines in the graphs are for Core 1; the black lines and black shading are for Core 2. The straight lines in the charcoal graphs show the position of the gap in Core 2 between 1300 and 600 cal yr BP.

assumes that drive 3, the core section that the team intended to collect from 150–250 cm sub-bottom, was accidentally collected 1 meter higher, overlapping the depth of drive 1 and leaving a 1 m gap in the core. An age model created without this depth adjustment showed large differences in sedimentation rates between the two cores when compared by both depth and interpolated age and significant variation in sedimentation rates with depth in Core 2 that are difficult to explain given the proximity of the core sites (less than 40 m apart), and depths at the core sites (ca. 7 m in both cases). However, as a further check on our depth and age models we plan to obtain additional radiocarbon dates from the upper core sections.

Unfortunately, Sally Horn and I did not recognize the existence of the gap in Core 2 when we were designing the sampling strategy for my thesis. I carried out macroscopic charcoal analysis on both Cores 1 and 2 along with MWI 2, allowing the construction of a composite charcoal diagram that covers the last ca. 5000 years without large gaps in time. But I only carried out LOI, pollen, and isotope analyses on Core 2. In the future we will fill these gaps using material from Core 1. In the proxy diagrams, I include the LOI, pollen, and isotope data for both of the overlapping upper core sections of Core 2, providing slightly higher resolution for the last ca. 600 years at Pine Pond than for earlier intervals. Combined with the charcoal data in Core 1, these proxy analyses provide a fairly high-resolution picture of fire-climate-human dynamics during the Little Ice Age and period of Spanish and British colonization. However, because of the gap in Core 2, the end of the Classic Maya era and time of the Terminal Classic Drought are only represented in the charcoal data.

5.3 Charcoal Analysis

Macroscopic charcoal analysis reveals a long history of local fires at Pine Pond, with charcoal present in every 1-cm interval in Cores 1 and 2. During the period of overlap, Core 1 shows higher charcoal concentration and influx values (Figure 13), possibly because the core site is located closer to shore, where charcoal might be preferentially deposited following fires. The composite charcoal concentration and influx curves that were adjusted for these differences (Figure 14) show fairly steady oscillations until ca. 1600 cal yr BP, when values increase, especially for charcoal influx. A marked increase in both charcoal concentration and influx occurs at ca. 1200 cal yr BP, with high values persisting until ca. 50 cal yr BP.

5.4 Pollen Results

Eight pollen types account for 86% of all pollen grains I counted across the Pine Pond record. These pollen types are: Poaceae (26%), *Pinus caribaea* (20%), Cyperaceae (14%), *Quercus,* likely *Q. oleoides* (9%), Urticales exclusive of *Trema*, *Cecropia*, and *Ulmus* (5%), Myrica (5%), *Brysonima/Rhizophora* (4%), and *Trema* (3%). The percentage of unknown pollen grains per level averaged ca. 5%, only exceeding 10% in four levels. Indeterminate grains were rare, averaging between 1 and 2% per level.

Pollen percentages and total monolete and trilete fern spore percentages show some large changes in the Pine Pond record (Figure 15). Fern spores rise dramatically toward the top of the record. The most dramatic shift in pollen types is the large increase in pine pollen percentages in the last ca. 400 years (ca. 330 cal yr BP to present). Pine and grass are the two most common pollen types in the sediments, and their percentages are inversely correlated (Figure 16). The upper part of the profile with the highest charcoal concentration and influx values includes the

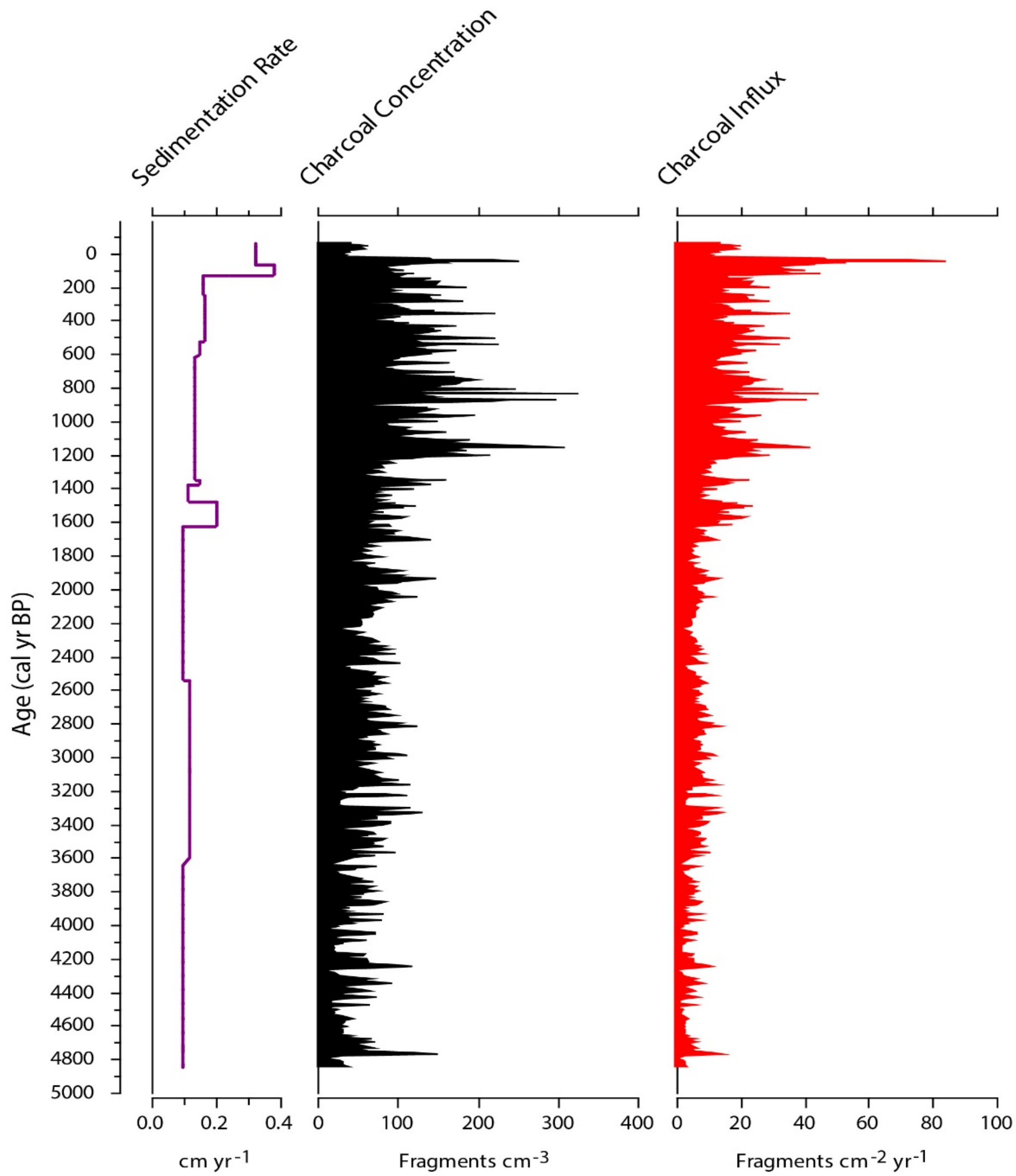

Figure 14. Composite charcoal concentration and influx diagram for Pine Pond. The sedimentation rate graph uses the rates calculated by the Clam models for Core 1 and Core 2; for the MWI core I used the Core 1 model.

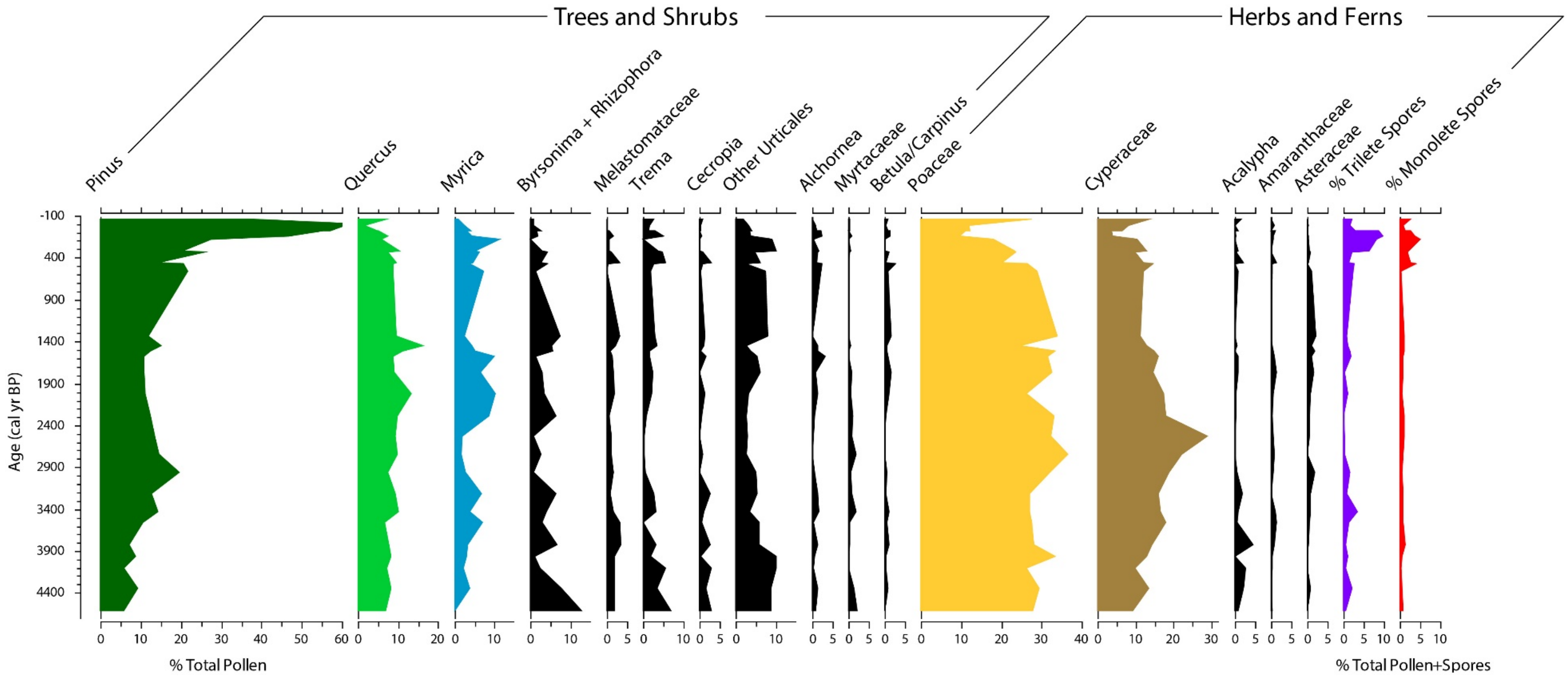

Figure 15. Pollen diagram for Pine Pond. Pollen types are plotted as a percentage of total pollen, and fern spores are plotted as a percentage of total pollen and spores. Types that were identified and counted but that were never higher than 2% in any sample level include: *Ilex*, *Ulmus*, *Alnus*, *Bursera*, *Myrsine*, Piperaceae, Palmae, Sapindaceae, ERA-type, *Hura*, *Utricularia*, *Dorstenia*, and *Mollugo*. The curve for *Pinus* pollen is slightly truncated; this pollen type reaches a peak value of 62.4% at 17 cal yr BP.

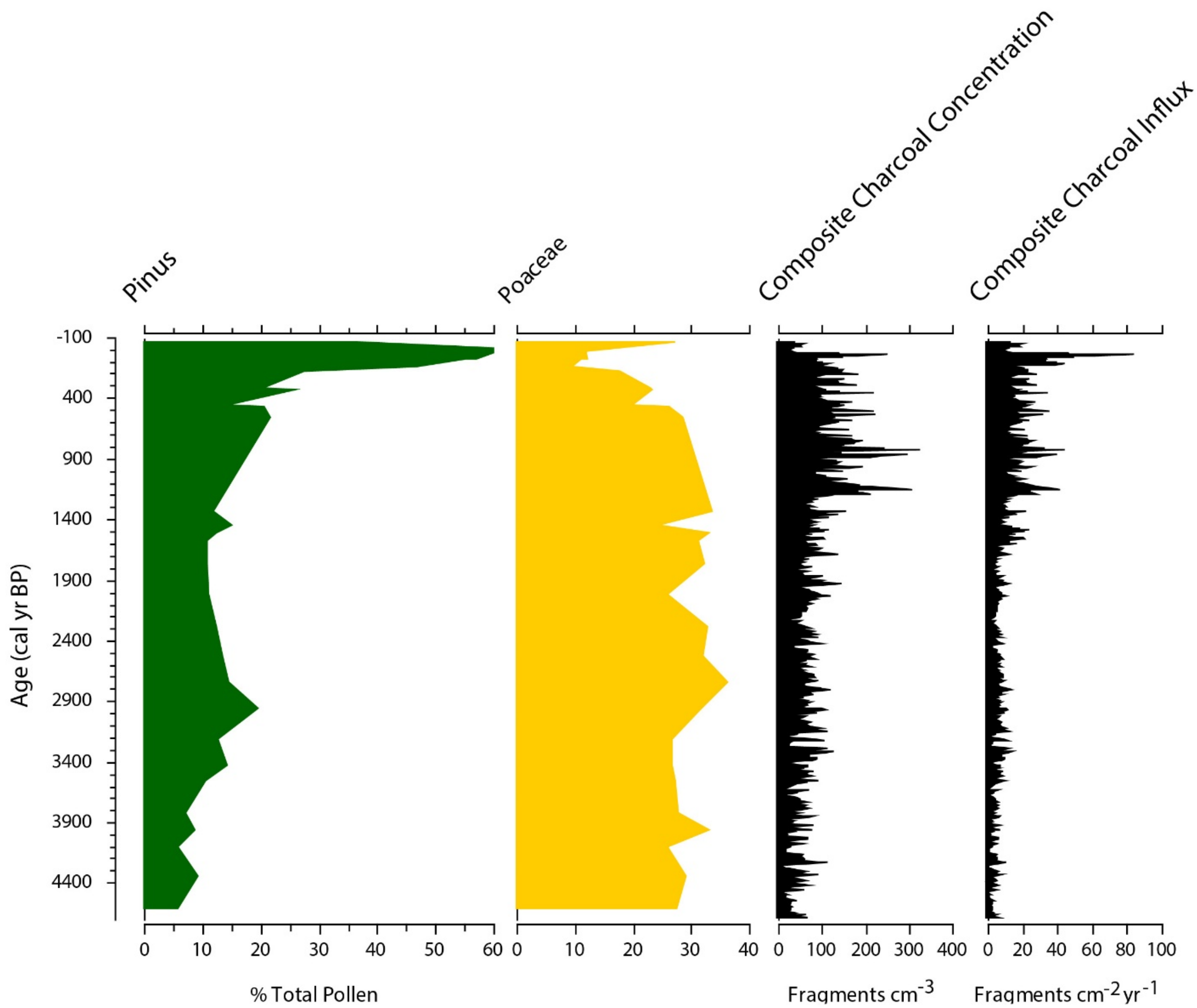

Figure 16. Relationships between pine and grass pollen percentages and charcoal influx vales plotted by age (cal yr BP). The curve for *Pinus* pollen is slightly truncated; this pollen type reaches a peak value of 62.4% at 17 cal yr BP.

period of highest pine pollen percentages, but pine is not high in all samples with high charcoal values. Charcoal influx in Pine Pond increases ca. 1600 cal yr BP and remains high through the Terminal Classic Drought and the Little Ice Age (Figure 17). No pollen or spore data are available for the Terminal Classic Drought, but percentages of trilete and monolete spores and ratios of Grass to Sedge pollen and Pine to Oak pollen show variations during the Little Ice Age that could relate to shifts in climate (Figure 17). The pollen ratios were calculated based on Mensing et al. (2008), with the Poaceae/Cyperaceae ratio computed as (total grass pollen – total sedge pollen) / (total Poaceae + Cyperaceae pollen) and the *Pinus*/*Quercus* ratio computed as (total *Pinus* – total *Quercus* pollen) / (total *Pinus* + *Quercus* pollen).

5.5 Bulk Geochemistry and Stable Carbon and Nitrogen Isotope Ratios

Organic carbon and nitrogen percentages and C/N ratios show relatively small changes over the Pine Pond record (Figure 18). Carbon and nitrogen percentages closely track one another and the organic matter (OM) percentages determined by loss on ignition. C/N ratios generally decrease towards the top of the core. Ratios range from 15:1 to 10:1, indicating a mixture of terrestrial and aquatic carbon inputs (Ishiwatari and Uzaki, 1987). Stable carbon isotope ratios remain within 2% of the mean (–21.79‰) through the profile, indicating that the terrestrial organic carbon reflects a fairly consistent contribution of C_3 and C_4 plant matter (Figure 19). Stable nitrogen isotopes ratios also exhibit little variation, remaining slightly negative through the profile, with a mean of –0.73‰ (Figure 20).

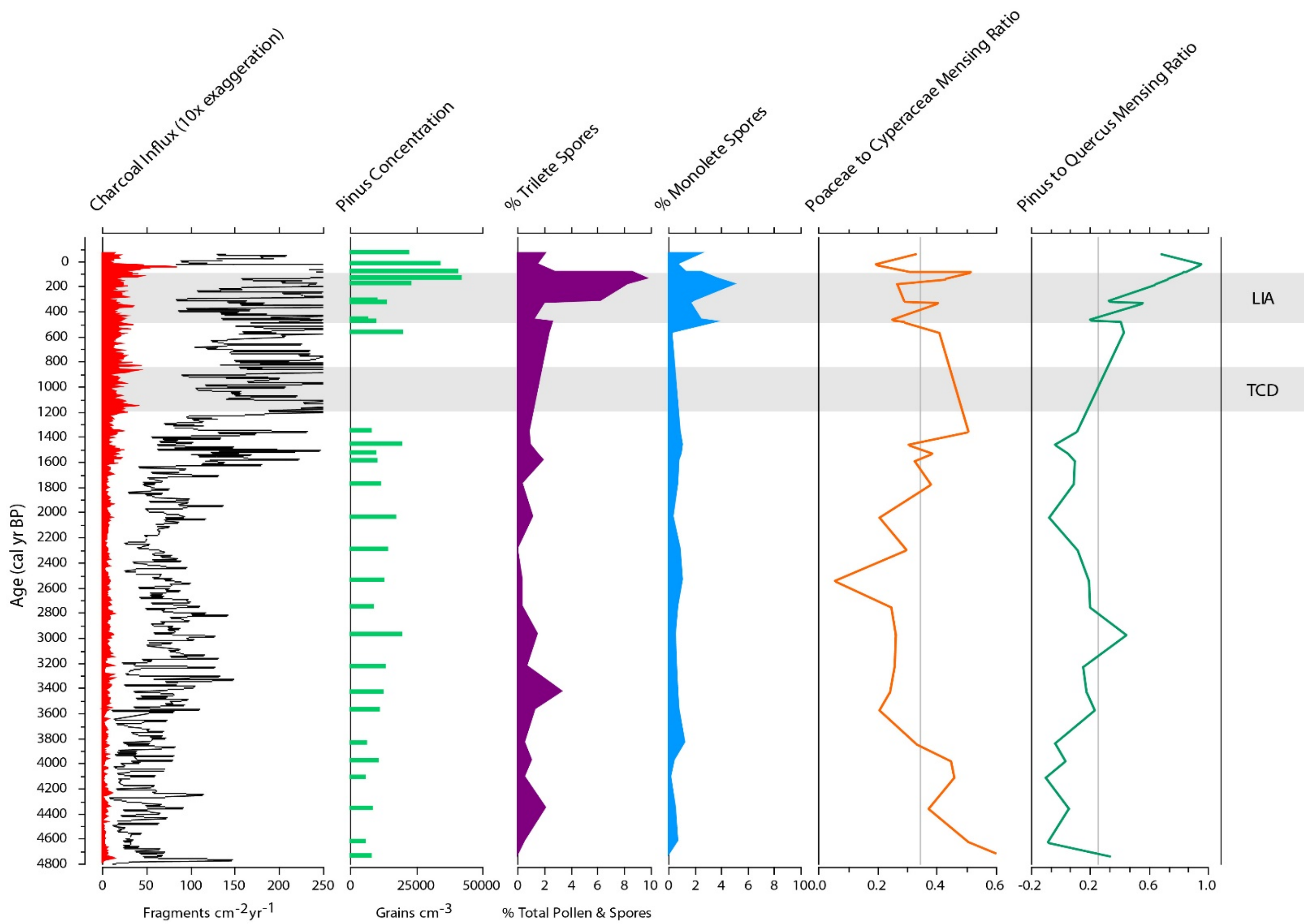

Figure 17. Charcoal influx, fern spore percentages, and selected pollen ratios calculated following Mensing et al. (2008). Shaded intervals (light gray) delineate the Terminal Classic Drought (ca. 1200–850 cal yr BP) and Little Ice Age (ca. 500–100 cal yr BP).

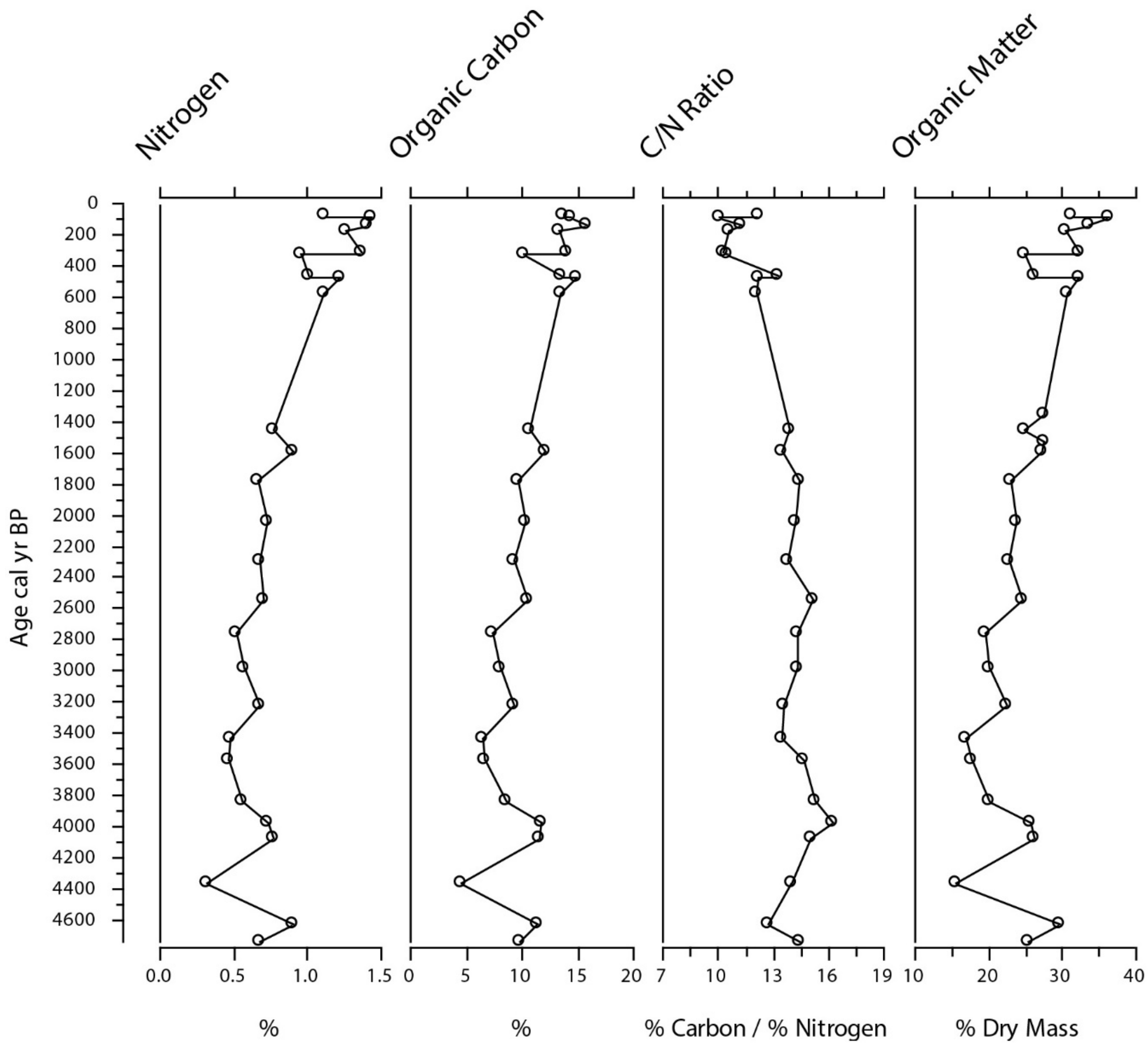

Figure 18. Values for %N, %C, C/N ratios, and % Organic Matter across the profile plotted by age (cal yr BP).

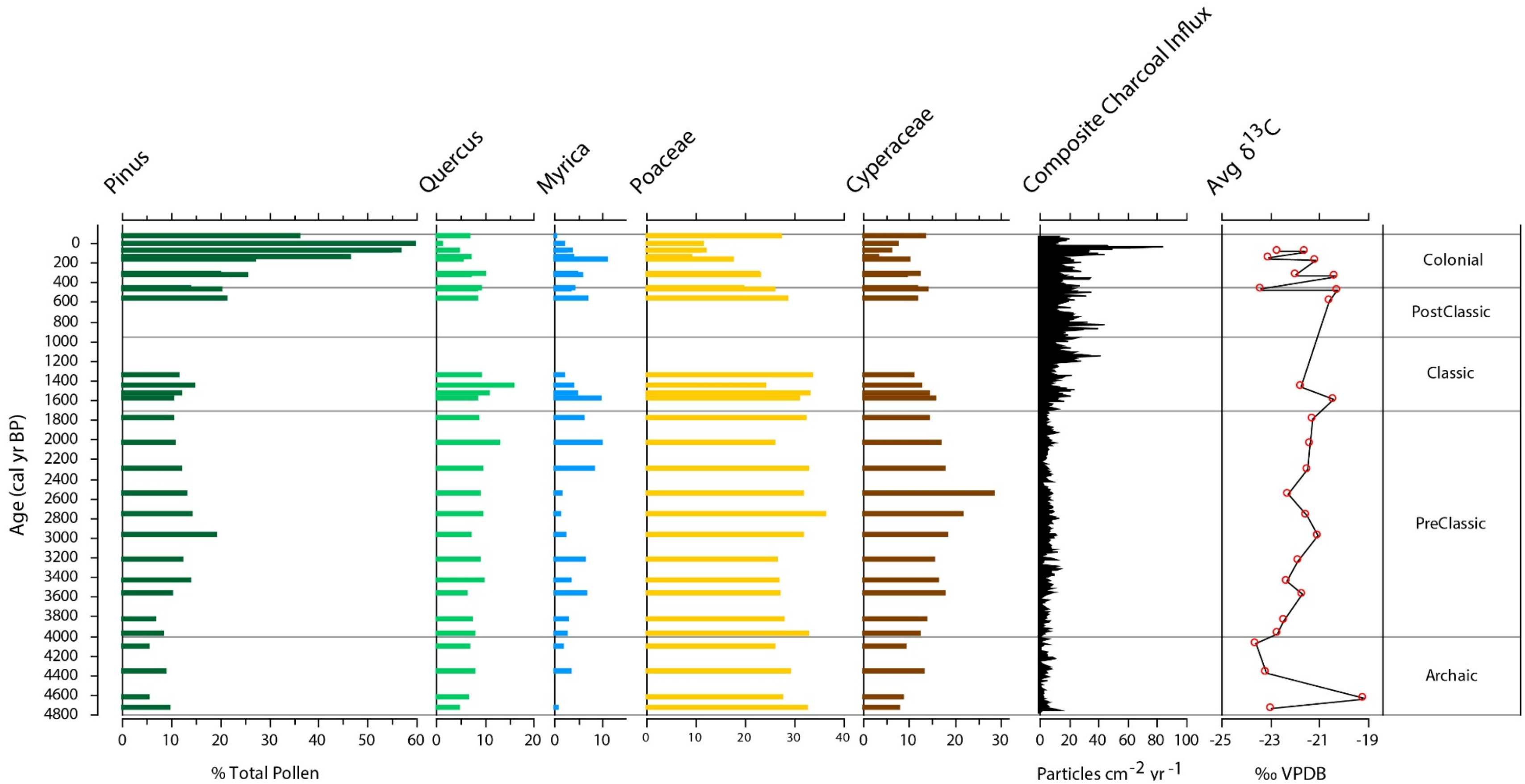

Figure 19. $\delta^{13}C$ ratios plotted by age (cal yr BP) in comparison with charcoal influx and the five main savanna plant taxa in the record. Sample values for pollen are displayed as bars to highlight the variations in resolution and to show the gap in the record. See text for sources used to delineate archaeological zones. The curve for *Pinus* pollen is slightly truncated; this pollen type reaches a peak value of 62.4% at 17 cal yr BP.

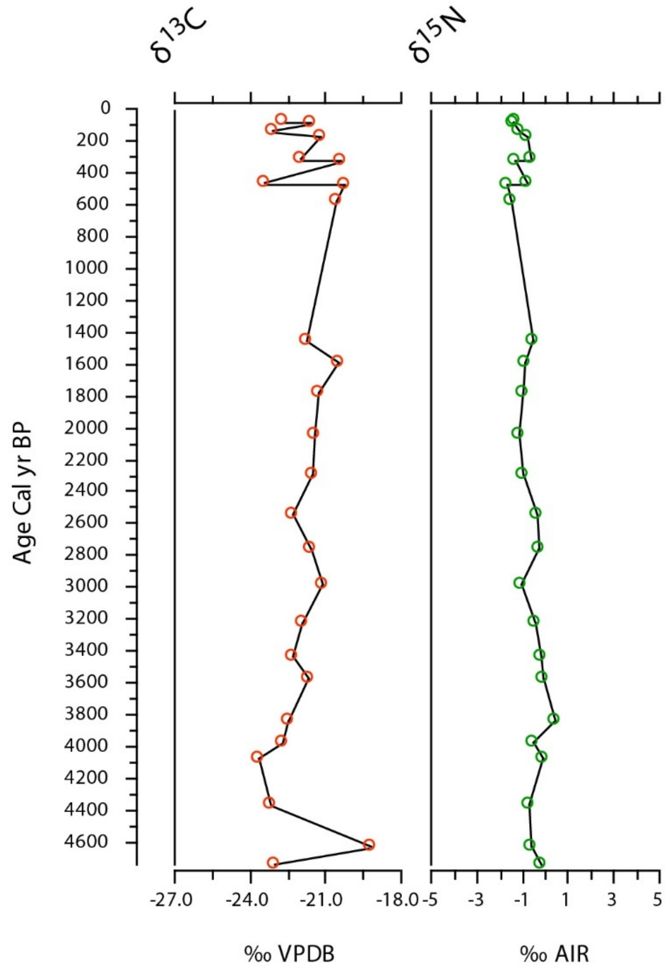

Figure 20. Variation in $\delta^{15}N$ and $\delta^{13}C$ of bulk organic matter plotted by age (cal yr BP).

5.6 Description of Results by Archaeological Zones

One of the main questions driving this research was when, if at all, did human land-use changes begin to impact fire regimes and affect ecosystem development. Here I divide and present proxy results by distinct archaeological periods (Figure 19). Specific time period delineations are based on studies by Rushton (2012), Walsh et al. (2014), and Anderson and Wahl (2016), and on personal communication with Cathy Smith. Rather than a lengthy description and comparison of all proxy changes per zone, only distinct shifts and/or trends are highlighted.

The Archaic Period (4800–4000 cal yr BP), represented by the four lowest levels sampled for pollen and bulk stable isotope analysis, includes the most positive $\delta^{13}C$ value (–19‰) in the entire profile at ca. 4600 cal yr BP. This sample corresponds with lowest percentage of pine in the record (5.8%), the highest level of *Byrsonima/Rhizophora* (12%), and a relatively high percentage of total Urticales (18.3%). Charcoal influx values were relatively low, rarely rising above 10 fragments/cm^2/yr.

Charcoal values during the PreClassic Period (4000–1700 cal yr BP) remain relatively low and stable, with no major peaks or notable absence. This zone contains the highest number of sample levels for pollen and isotope analyses (11), but the resolution is low with ca. 200-year intervals between each sample. Across this period the three most common pollen types are pine, grass, and sedge.

The gap in the Core 2 profile affects the pollen and isotope sample resolution during the Classic Period (1700–950 cal yr BP). There are two isotope samples and four pollen samples for the interval between ca. 1590 and 1350 cal yr BP, before a gap in the data until ca. 600 cal yr BP. The highest prevalence of oak in the record (16.3%) occurs ca. 1400 cal yr BP and corresponds to a relatively low percentage of grass pollen. In the composite core, charcoal influx rises significantly ca. 1600 cal yr BP, then briefly drops ca. 1400 cal yr BP before a dramatic rise until ca. 1300 cal yr BP. There is then another notable period of elevated burning between 1200 and 1100 cal yr BP. From here until the end of the Classic Period (950 cal yr BP) values decline and remain relatively low.

During the PostClassic period (950–450 cal yr BP), charcoal influx in the composite charcoal profile reaches a low point at ca. 900 cal yr BP. Values then rise sharply, and there is significant variation in peaks and troughs across the rest of this zone. The composite charcoal graph documents the consistently highest influx values of the record through the end of the PostClassic and across the Colonial Period (450 cal yr BP to present). Grass pollen begins to decline at the end of the PostClassic and reaches a nadir of 9.7% ca. 170 cal yr BP. The low values of grass pollen in this zone correspond to the highest percentages of pine pollen and charcoal influx in the entire profile. Zone 5 also contains the highest percentages of organic matter, the lowest C/N ratios, and the most negative $\delta^{15}N$ values.

CHAPTER VI

DISCUSSION

6.1 Did ecosystems change over the interval covered by the Pine Pond sediment record? What do proxy analyses reveal about the environmental history of the site?

6.1.1 Charcoal, C/N, and $d^{15}N$ Ratios

The main purpose of this study was to provide a long-term record of fire around Pine Pond as a complement to Cathy Smith's dissertation on the political ecology of British colonial fire management. Charcoal is the proxy I examined at highest resolution, and the contiguous sampling strategy reveals a continuous record of burning. Every 2 cm^3 of sediment sampled contained at least some macroscopic charcoal. Given the flat topography and lack of inflowing streams at Pine Pond, these macroscopic charcoal fragments should mainly be from fires that occurred adjacent to the pond or within a short distance of perhaps 1–2 kilometers (Whitlock and Larsen 2001). While there is no way to determine if a piece of charcoal was produced by a human hand or a bolt of lightning, the Pine Pond charcoal record indicates a long history of burning near the lake due to some combination of edaphic control, natural fires, and human actions.

Because of high rain and flood events during the wet season, peaks of burning in the record may contain charcoal deposited in the lake by erosion and overland flow for some years after a fire (Whitlock and Larsen 2001). However, the grasses and sedges that quickly regrow following fires may trap charcoal particles that remain on the soil following fires, reducing the lag in charcoal deposition if many particles never move into the lake following burning. In both Core 1 and Core 2, increases in charcoal influx values track increases in sedimentation rates as calculated by the clam age-depth models. This is apparent in the composite core record beginning ca. 1620 cal yr BP, where a rough doubling in sedimentation rate matches a rough doubling of charcoal influx values (Figure 13). From ca. 150–66 cal yr BP the highest rates of sedimentation also match the highest rates of charcoal influx values in both core profiles (Figure 12). In part this situation can reflect the mathematics behind the calculation of charcoal influx, which corresponds to charcoal concentration in fragments/cm^3 multiplied by the sedimentation rate in cm/yr. But both increases occur during the Classic Maya era and the British colonial periods, periods when we might expect increased burning due to anthropogenic landscape alterations in the region due to increased populations (Walsh et al. 2014, Smith 2019). Regardless of the source, the charcoal influx values throughout the core indicate a wetter or drier savanna environment rather than the tropical lowland broadleaf forest that covers most of the Southern Coastal Plain (Colombaroli et al. 2014, Walsh et al. 2014).

Geochemical and stable isotope data are available only at intervals through the core, so only tentative comparisons are possible between these proxies and the fire record. While C/N values do not show large shifts through the record, they are generally lowest during the time period of highest charcoal influx, indicating they may reflect changing levels of lake productivity. Elevated fire activity could reduce inputs of terrestrial organic matter (Dunnette et al. 2014), or otherwise alter nutrients or limnological conditions in the lake, affecting algal productivity and carbon content of the sediments.

Dunnette et al. (2014) examined the effects of fire on nitrogen delivery into soils and watersheds, and posited that large fire events may lead to an increase in $\delta^{15}N$ ratios before a subsequent decline in sedimentary nitrogen due to forest recovery. Stephan et al. (2015) also found a positive correlation between increased $\delta^{15}N$ and %N and higher intensity fire events in a comparison of nitrogen biochemistry across several Rocky Mountain watersheds. These studies were conducted in conifer forests of the Rocky Mountains, where relatively less frequent, high intensity crown canopy fires would likely produce more discernible shifts in nitrogen than more regular surface fires that we expect in the savanna environment at Pine Pond. During the period of the highest level of burning and the highest isotopic resolution near the top of the Pine Pond core, $\delta^{15}N$ values do not show any major positive shifts that might be attributed to fire events, consistently showing the most negative values in the profile. However, a true test of the possible link between burning and sedimentary $\delta^{15}N$ ratios would require high-resolution sampling of both stable isotopes and charcoal. Examining nitrogen isotopes and charcoal concentration at a resolution of 1 cm in the Pine Pond core could provide a test for whether savanna fires influence $\delta^{15}N$ ratios. However, many factors other than fire events can influence nitrogen dynamics and nitrogen content and isotope ratios in lake sediments (Talbot 2001).

6.1.2 Pollen and d^{13}C Ratio Interpretations

My examination of 30 pollen slides from the core resulted in a low-resolution record that is not as detailed as other work in the region. The intent behind the pollen analyses was to augment the fire record and potentially document ecosystem changes by capturing broad shifts related to either changes in climate or anthropogenic land use practices. I chose to focus on the identification and changing assemblages of several main vegetation types associated with lowland Belizean savannas, surrounding lowland broadleaf forest, and disturbance taxa associated with human land-use change. My interpretation of the ecological significance of different pollen types found in the record is based on descriptions of plant communities associated with lowland Belizean savannas by Bridgewater et al. (2002), Laughlin (2002), and Goodwin et al. (2013). I include Urticales as a forest taxon, and I also consider Melastomataceae, *Alchornea*, and *Acalypha* to be indicators of forest presence, based on other paleoecological studies in Belize (Bhattacharya et al. 2011, Rushton 2014, Walsh et al. 2014) and in other neotropical savannas (Brenner et al. 1990, Dull 2004a). Based on modern pollen spectra analyses of soil and surface sediment cores taken from eight distinct ecosystems in northern Belize, Bhattacharya et al. (2011) found Pine, Oak, and *Byrsonima* almost exclusively in pine savanna samples. The highest percentages of grass and sedge were found in samples from pine savannas, lake banks, and non-forested, seasonally flooded environments. The continual presence of *Pinus*, *Quercus*, and *Byrsonima*, and relatively high percentages of grass and sedge throughout the pollen profile, suggest the ecosystem around Pine Pond has been a pine savanna with variations in herbaceous and woody taxa dominance for at least the last ca. 4800 cal yr BP.

Interpretations of the pollen record in comparison to the charcoal data is limited by the much lower resolution, and that fact that not all taxa were identified to the lowest possible level. Pollen production and dispersal rates of different taxa also affect assemblages and my interpretations. *Byrsonima crassfolia* has been identified as an important indicator species of pine savanna, with its presence being more indicative than pine pollen alone because of its lower pollen productivity and dispersal rates (Ledru 2002, Gosling 2009, Bhattacharya et al. 2011), but I was unable to fully separate this pollen type from *Rhizophora mangle.* Low counts of *Acoelorrhaphe wrightii* (Arecaceae) were only found in the MWI samples and the top sample of Core 1 (ca. 79 cal yr BP). The absence of palm pollen in the rest of the record may be a consequence of large grain size (>100 μm) and poor pollen dispersal (Rushton 2014). Given the modern prevalence of palm stands in the savanna around Pine Pond, the low counts in near-surface samples indicate that its presence on the landscape in the past is likely underrepresented.

The pollen profile shows pine stand density in the savanna around the lake was much lower in the past compared to present day levels. This interpretation is supported by consistently high percentages of grass and sedge pollen, and relatively low levels of forest taxa. The narrow range of $\delta^{13}C$ values (–19.23‰ to –23.62‰) also supports this interpretation. This range compares with the upper limit of values found in a ca. 3000-year reconstruction of a more open, C_4 grass dominated savanna in the highlands of El Salvador by Dull (2004) (high value of –21.4‰).

6.1.3 Interpretations from Sediment Stratigraphy.

Each core section consists of fine-grained organic sediment with little variation in color or composition. A notable exception occurs in Core 2 from 505-510 (Figure 9). Given the depth of the pond (ca. 7 meters), desiccation is an unlikely explanation for appearance of this interval. I hypothesis this interval of coarse organic material that includes wood fragments and leaves may represent inwash associated with the high-winds and flooding of a hurricane. This interval is the only one of its kind in the core, but this may indicate the proximity of a past storm (i.e. passing directly over the lake) rather than a lack of hurricanes in the past. The unique geography and climatic influences of Belize point to a long history of intense tropical storms and hurricanes. Paleotempestology studies using coastal sediments and coral reefs as proxies document hurricane occurrence stretching back millennia (McCloskey and Keller 2008, Denommee et al. 2014). In 2001, Hurricane Iris passed near Pine Pond. The patches of broadleaf forest in the vicinity experienced extreme damage and defoliation, while the pine savanna exhibited fewer signs of disturbance (Linares and Vinicio 2009, Mcloughlin 2015). Such strong disturbance events in the past may have influenced the extent of savanna around Pine Pond by altering natural fire breaks and changing fuel availability, which in turn may have promoted fire in the disturbed areas of broadleaf forest (Myers and Rodríguez-Trejo 2009).

6.2 What do proxy analyses reveal about late Holocene climate change at the site? Do changes in charcoal, pollen, and stable isotope values reveal evidence of climate events such as the Little Ice Age and Terminal Classic Drought?

6.2.1 Pollen, Fern Spores, and Charcoal as Potential Indicators of Climatic Changes

Over the course of the Pine Pond record, increased charcoal influx broadly corresponds with drier intervals in the Macal Chasm and Yok Bolum speleothem records (Figure 21), while intervals with lower charcoal influx generally correspond to wetter periods. This is particularly apparent during the Terminal Classic Drought, with a dramatic rise in charcoal influx values coincident with the beginning of each inferred dry period. Influx values remain high across the rest of the record and peak towards the end of the Little Ice Age. Unfortunately, pollen and isotope data are lacking during the Terminal Classic Drought. But, by the end of the Little Ice Age there are higher Poaceae/Cyperaceae and *Pinus*/*Quercus* ratios, and higher percentages of fern spores —all potential indicators of drought.

Mensing et al. (2008) found the ratio of Poaceae to Cyperaceae pollen to be a reliable indicator of drought in the Great Basin of Nevada over the last ca. 2000 years when compared

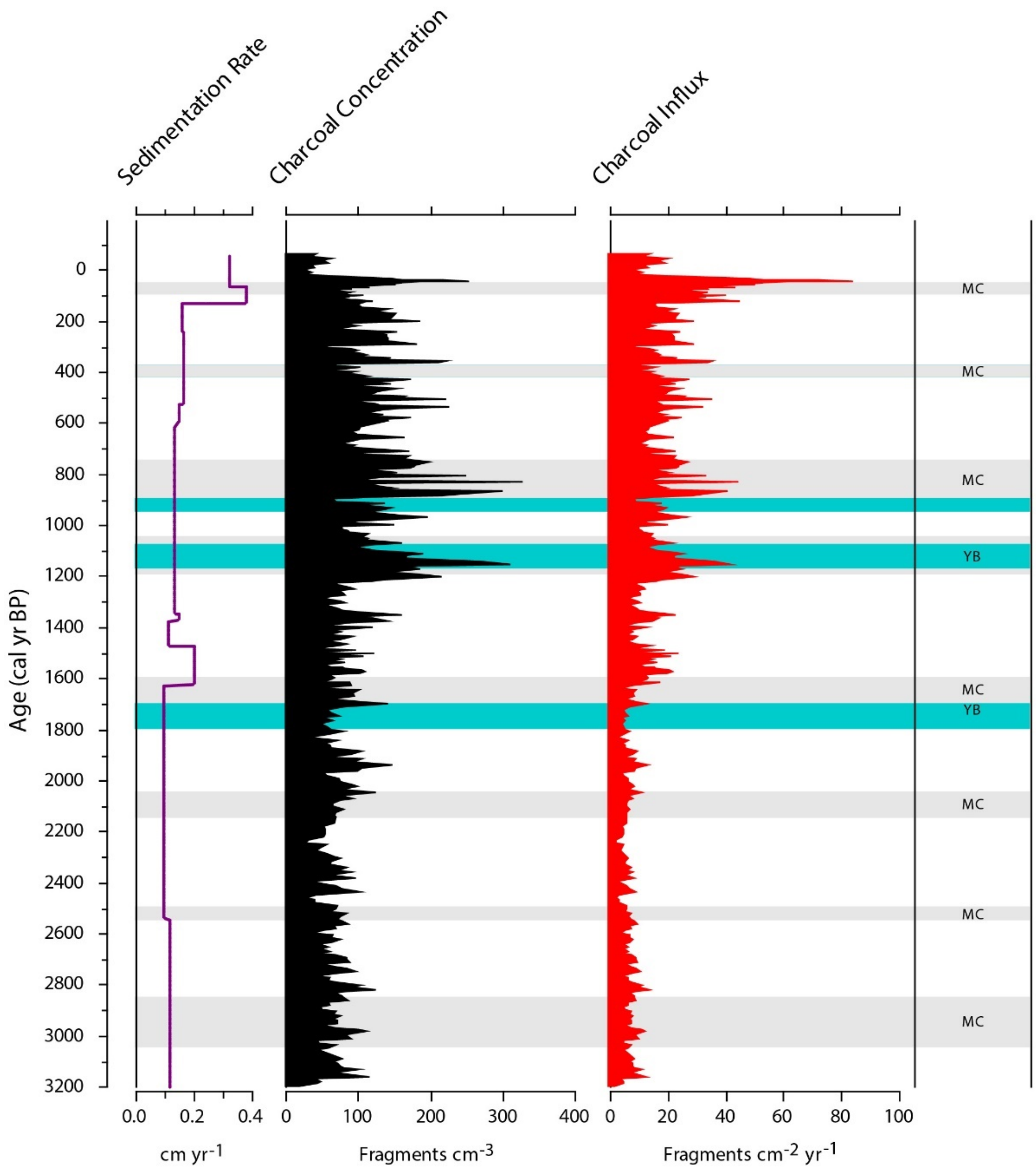

Figure 21. Dry periods inferred from stable oxygen isotope analysis of the Macal Chasm (MC) and Yok Bolum (YB) speleothem records compared to charcoal influx at Pine Pond. Diagram made using data from Table 4 in Akers et al. (2016).

with other independent climate records and paleoecological studies in the region. Higher Poaceae/Cyperaceae ratios appear to indicate drier conditions, which favor grasses, while lower Grass/Sedge ratios indicate relatively wetter conditions favoring sedges. I hypothesized a similar relationship might exist at Pine Pond, i.e. more grass = drier, more sedge = wetter. The Poaceae/Cyperaceae ratios throughout the Pine Pond profile do not show as clear a relationship with climate as found by Mensing et al. (2008). For example, the lowest ratio of grass to sedge pollen occurs ca. 2500 cal yr BP (Figure 17) during an extreme dry event recorded in the Macal Chasm record. High percentages of both grass and sedge pollen ca. 3000–2300 cal yr BP and low percentages of forest taxa and woody taxa associated with savannas indicate open, drier conditions at this time. The Poaceae/Cyperaceae ratio broadly increases across the rest of the record, and peaks at the end of the dry Little Ice Age. But percentages of both types reach their lowest in the core during this interval, in contrast to the highest levels of pine and charcoal. This large increase in burning may be tied to a drier climate, but it may also be a product of human altered fire regimes, a finding I discuss in a later section of this chapter.

I also hypothesized that a *Pinus*/*Quercus* ratio may act as a climate signal akin to the Poaceae/Cyperaceae ratio, with higher *Pinus*/*Quercus* ratios associated with drier conditions, and lower ratios with wetter conditions. Increases in *Pinus*/*Quercus* ratios do roughly follow increases in charcoal influx values. It is possible this ratio may be tied to broader climatic signals, i.e. a drier climate leading to more fires and/or lower moisture leading to relatively lower levels of oak compared to pine. Lower ratio values of *Pinus*/*Quercus* ca. 4600–3800 cal yr BP are concomitant with a decline in Poaceae/Cyperaceae ratios and a lower incidence of fire. This may represent a signal for wetter climatic conditions. *Pinus*/*Quercus* ratios then generally rise to a peak ca. 2900 cal yr BP, a rise coupled to an increase in fire. This increase in *Pinus*/*Quercus* ratios contrasts with a continued decline in Poaceae/Cyperaceae ratios. Between ca. 2500 and 1300 cal yr BP the climate signal comparison among both ratios is less clear, though the low sampling resolution also hinders interpretation. An increase in Poaceae/Cyperaceae ratio values across this interval contrasts with a general decline and stagnation of *Pinus*/*Quercus* values. Charcoal values increase dramatically at the start of the Terminal Classic Drought, and generally increase throughout the rest of the record. This increase in fire incidence matches an upward trend in *Pinus/Quercus* ratios through the Little Ice Age.

The Macal Chasm record (Figure 21) captures a severe drought ca. 100–50 cal yr BP. This inferred period of aridity coincides with the highest levels of burning in the record along with the highest percentages of pine and fern spores. Paleoecological reconstructions in the circum-Caribbean have linked increases in fern spores with increases in aridity, especially in conjunction with peak in grass pollen and charcoal, though definitive interpretations are complicated by high spore production and dispersal (Bush and Colinvaux 1994, Northrop and Horn 1996). The high levels of pine, fern spores, and charcoal contrasts with the lowest levels in the record of grass, sedge, and oak. The Poaceae/Cyperaceae and *Pinus*/Quercus ratios may signal a wetter climatic period at the base of record and a dry Little Ice Age. Both ratios also appear to be associated with changes in fire activity driven by some combination of changes in climate and human land-use practices.

6.2.2 d^{13}C Interpretation

I also hypothesized that shifts in δ^{13}C values in the Pine Pond record may represent vegetation responses to increases in aridity or moisture availability. Cross-checking native grasses and sedges listed by Goodwin et al. (2013) as occurring in Belizean savannas against lists of C_3 and C_4 graminoids (Bruhl and Wilson 2007, Osborne et al. 2014) shows that the dominant grasses are C_4 species, but some grass species may follow either the C_3 or C_4 pathway. The expected genera of sedges also include a mixture of C_4 and C_3 species. While the introduction of non-native grasses could cause changes in $\delta^{13}C$ values in sediment records, fire ecologist Rick Anderson commented that he has rarely observed invasive grasses in the modern savannas of the Deep Forest Reserve and Payne's Creek National Park, although invasive grasses do occur along roadsides and in fallow fields (R. Anderson, personal communication with S. Horn, October 1, 2019). Periods of aridity may have increased the abundance of C_4 species, causing a shift towards more positive $\delta^{13}C$ values. The most positive $\delta^{13}C$ value of the record (–19.2) at ca. 4600 cal yr BP does match higher Poaceae/Cyperaceae and *Pinus/Quercus* ratios along with a spike in charcoal influx. The most consistently negative values ca. 4350–3800 cal yr BP also match a decline in both pollen ratios and fire occurrence. This strengthens the inference of a wetter climate towards the start of the record. Values across the rest of the record, however, show little variation, including during the Little Ice Age, which was broadly dry based on speleothem records developed by Kennett et al. (2012) and Akers et al. (2016).

6.3 What do the pollen and geochemical analyses reveal about human disturbance at the site, and how does the record compare to other paleolimnological records for the Maya lowlands?

6.3.1 Pollen and Stable Isotope Evidence of Human Disturbance

Here, I discuss shifts in pollen taxa and stable isotope ratios that may indicate human disturbance. I consider pollen of Amaranthaceae and Asteraceae to represent weedy herbaceous taxa potentially associated with human disturbance (Rosenmeier et al. 2002, Dull 2004b, Bhattacharya et al. 2011, Walsh et al. 2014, Johanson et al. 2019). Among woody taxa, I include *Cecropia* and *Trema* as possible indicators of forest clearing or other large-scale disturbances to vegetation. Peaks in *Cecropia* and *Trema* (and to a lesser extent Melastomataceae) pollen have been interpreted to signal tropical forest regrowth after human and natural disturbances (Leyden 1987, Northrop and Horn 1996, Bhattacharya et al. 2011). But at Pine Pond, values of all these taxa remain relatively low through the record, failing to reveal evidence of increases in abundance that could signal intervals of forest clearance or major disturbances to vegetation.

The most relied upon palynological indicator of human presence and vegetation disturbance is pollen of *Zea mays* or corn (Clement and Horn 2001, Kennedy and Horn 2008, Lane et al. 2008, Taylor et al. 2013, Johanson et al. 2019). Due to large grain size and poor dispersal, the lack of maize pollen in a sample level does not necessarily indicate a total lack of cultivation in the surrounding area (Taylor et al. 2013). Sometimes maize grains are too rare to encounter in a standard pollen count but can be detected through low-power scanning of multiple microscope slides (Horn 2006). I did not find any maize pollen at Pine Pond, likely because the nutrient-poor, acidic soil and hyper-seasonal flooding limited the possibility for large scale prehistoric maize cultivation.

There is evidence of large agricultural terracing projects in the wet savannas of northern Belize during the Classic Maya period (300-500 CE) (Beach et al. 2009, Rushton et al. 2012). Iriarte et al. (2012) documented evidence for raised field agriculture to overcome seasonal flooding and the limiting of fire to improve soils in the coastal savannas of modern-day French Guyana. There is lack of archaeological evidence for raised terracing or indeed any major permanent settlements in the savanna near Pine Pond, and based on the continual presence of charcoal in the record there does not appear to have been a systematic limitation of fire to improve soils for agriculture. The pollen data and $\delta^{13}C$ values in tandem indicate a relatively open savanna environment dominated by a mixture of C_4 and C_3 grasses, sedges, and other woody savanna taxa long before the first known permanent settlements in the area (Dull 2004a, Prufer 2005). The narrow range of $\delta^{13}C$ values and forest taxa pollen percentages, while at a relatively low resolution, belies a lack of major ecotone shifts related to possible deforestation by humans in the vicinity of Pine Pond. The generally high percentages of organic matter and lack of major peaks in C/N ratios further supports the interpretation of low disturbance, in that large-scale human landscape alterations in the surrounding area could be expected to increase the input of inorganic sediment and terrestrial carbon to the lake. While I cannot say for certain, it is likely human disturbances associated with agricultural practices were restricted to riparian zones with more fertile soil and/or areas of broadleaf forest further inland.

6.3.2 Pine Pond within the Context of Other Paleolimnological Studies in the Region

This study adds to the limited number of paleolimnological studies in the region that investigate the relative effects of prehistoric and historic land-use changes, natural climate variability, and the importance of edaphic controls on the developments of Central American savanna ecosystems. Brenner et al. (1990) found evidence of significant forest regrowth around the time of European contact in the savannas of the Petén, Guatemala. This finding strongly implicates burning by humans as a driving factor in savanna maintenance, and depopulation as a driver of subsequent reforestation, though the authors do not discount the possible additional role of natural changes in climate and hydrology. Dull (2004a) developed a ca. 3000-year reconstruction of savanna history in the highlands of El Salvador and reported similar evidence of a major decline in burning and increase in forest taxa coincident with population decline around the time of first European contact in the region. Rushton et al. (2012) linked the relationship between concurrent increases in charcoal and decreases in pine pollen to selective pine harvesting in the nearby savanna for construction at the major Mayan population center of Lamanai. The nature and timing of my proxy results disagrees with the findings in all three of these paleolimnological reconstructions. The discrepancies speak to the heterogeneity of savanna development in the region across the Late Holocene. In the next section I discuss proxy changes across distinct archaeological time zones and independent climate records to explore the possible effects of localized human land-use and climate change on the savanna ecosystem around Pine Pond.

6.4 Were ecosystem changes over the late Holocene primarily driven by people or climate? How does the Pine Pond record compare to the fire and climate record of Walsh et al. (2014)? When, if at all, did humans begin to significantly alter fire regimes around Pine Pond?

6.4.1 Pine Pond Across Archaeological and Historical Time Periods

Archaic Period (4800–4000 cal yr BP)

Proxy results indicate the landscape surrounding Pine Pond may be classified as a type of savanna as far back as ca. 4800 cal yr BP, but fire does not appear to have been the singularly dominant mechanism of vegetation control during this period. The Archaic period shows both the consistently lowest charcoal influx and the lowest percentages of pine pollen. Here, changes in fire frequency and vegetation assemblages are likely predominantly linked to changing climate, as this period predates the earliest archaeological evidence of humans in the region (Prufer 2005). The lower incidence of fire and savanna taxa and higher incidence of forest taxa compared to later in the record broadly agrees with a period of wetter climate inferred from the Cariaco basin Titanium record and the speleothem record from the Macal Chasm in the Maya Mountains. A ca. 8700-year paleolimnological reconstruction from Lago Puerto Arturo in northern Petén, Guatemala along with a modeled climate record from El Junco Crater Lake in the Galápagos Islands also provide evidence for a relatively wet middle Holocene before a gradual drying trend across late Holocene (Conroy et al. 2008, Wahl et al. 2014) (Figure 22).

Given the relatively wetter climate and higher percentages of forest taxa during this time period, the continual presence of charcoal is notable. Because there is no evidence for major human activity in the area, edaphic factors were likely responsible for maintaining a savanna ecosystem conducive to burning. While there is no archaeological evidence for permanent human settlement in the region prior to the PreClassic period (Prufer 2005), I cannot definitively rule out the possibility of human-set fires. Paleoecological work in the north of Belize shows evidence of maize and manioc pollen dating back to 3400 BCE (5350 cal yr BP) and significant deforestation by people by at least 2500 BCE (4450 cal yr BP) (Jones 1994, Pohl et al. 2006). Though the dates are not absolute, fluted stone points discovered at sites in both the north and south of the country push human occupation back to the Pleistocene/Holocene transition ca. 13,500 to 10,000 years ago (Lohse et al. 2006). There are notable peaks in charcoal in the Pine Pond record at 4769 cal yr BP and 4248 cal yr BP, but these peaks roughly correspond with short-term peaks in aridity evident in the Macal Chasm speleothem and Lago Puerto Arturo sediment records.

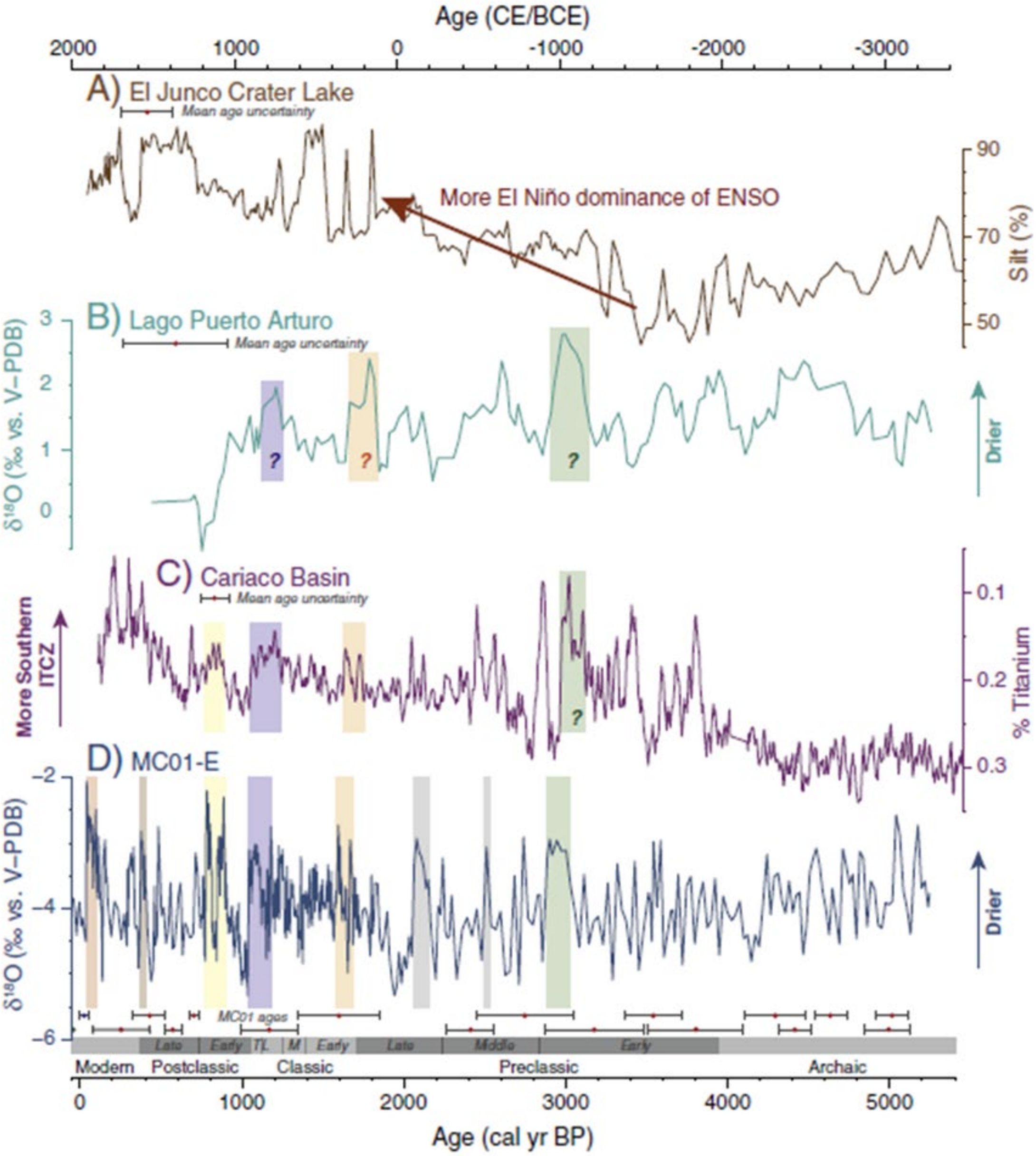

Figure 22. Comparison of climate inferences from the El Junco Crater Lake from the Galapagos (Conroy et al. 2008), Lago Puerto Arturo record (Wahl et al. 2014), Cariaco Basin record (Haug et al. 2001), and Akers et al. (2016) (MC01-E). Diagram taken from Akers et al. (2016).

PreClassic Period (4000–1700 cal yr BP)

The Macal Chasm speleothem record indicates a general trend of wetter climatic conditions from 5250 to 3100 cal yr BP characterized by low standard deviations in $\delta^{18}O$ and $\delta^{13}C$, a trend that Akers et al. (2016) interpreted as a proxy for increased moisture availability. This trend is also present in the Lago Puerto Arturo and Cariaco Basin records, though rapid shifts between wetter and drier conditions occurred during this time. Based on pollen evidence in a core from Lake Petén Itza in northern Guatemala, Mueller et al. (2009) inferred a drier climate ca. 4500–3000 cal yr BP, with a peak in aridity ca. 3500 cal yr BP.

The lower resolution pollen sampling and differences in microclimates limits comparisons between Pine Pond and the sites of these climate inferences, but generally forest pollen taxa decline with increases in fire and inferred periods of aridity. Pine pollen begins to rise ca. 3500 cal yr BP and reaches a peak at 2979 cal yr BP. This peak in pine matches relatively high charcoal influx values and an inferred abrupt shift to dry conditions in the Macal Chasm record ca. 3100 cal yr BP. The Lago Puerto Arturo and Cariaco Basin records also show a shift to drier climate at ca. 3200 cal yr BP. Wahl et al. (2014) attributed this shift to ENSO cycle changes in the tropical Pacific. This shift to drier conditions has also been linked to variations in solar insolation, the North Atlantic Oscillation, and the Atlantic Multidecadal Oscillation (Mueller et al. 2009). Irrespective of the dominant drivers behind this trend in increased aridity, the shift towards a drier climate is broadly reflected in the Pine Pond charcoal and pollen record.

A general increase in fire beginning ca. 2200 cal yr BP is consistent with a period of drier climate inferred from the Cariaco Basin record and major dry event ca. 2150–2050 recorded in the Macal Chasm record. Pine pollen values peak at 2979 cal yr BP during a period of elevated burning, then generally decline across the rest of this period, as do forest taxa. While the pollen record across the period lacks the resolution to make definitive linkage between changes in vegetation and climate, these broad declines may be tied to a general drying trend.

Archaeological work in southern Belize is lacking in comparison to the extensive excavations carried out in the lowlands of northern Belize, southern Honduras, and the Petén region of Guatemala. Currently there is no evidence for any large established prehistoric population in the immediate vicinity of Pine Pond, but ceramics from cave sites in the nearby Maya Mountains and evidence of settlement along the nearby Placencia coast to the northwest of the site indicate possible human activity in the area as early as the middle PreClassic period (Prufer 2005). While people were likely in the area, the archaeological site of Uxbenka (ca. 45 kilometers SW of Pine Pond) is the only major site permanently occupied prior to 500 CE (1450 cal yr BP) in the present-day Toledo district. Walsh et al. (2014) attributed low charcoal influx values and low percentages of forest taxa in the Agua Caliente profile between 2800–2600 cal yr BP to low human activity and a drier climate, conditions that also seem to have prevailed at Pine Pond during this interval. The authors linked a period of elevated burning between 2600–2350 cal yr BP to increased population and agriculture, while noting that human-ignited fires may have been exacerbated by drier climatic conditions as inferred from the Cariaco basin record. The Macal Chasm (2016) record, providing a more local signal, also shows a period of aridity centered around ca. 2500 cal yr BP. Around Pine Pond, fires appear to have been less prevalent across this interval, but again influx values are still relatively high compared to the Archaic Period. At Agua Caliente, fire activity begins to decrease ca. 2350 cal yr BP and remains low into the Classic Era. In the Pine Pond record, charcoal influx values remain relatively stable with a slight decrease around the end of the PreClassic Era.

Classic Era (1700–950 cal yr BP)

Charcoal values began to rise ca. 1700 cal yr BP, with a prominent spike that coincides with a dry period ca. 1850–1650 cal yr BP identified in the Yok Bolum speleothem record (Kennett et al. 2012) from a site 50 km SW of Pine Pond (Figure 21). The Macal Chasm and Cariaco records also indicate drier conditions ca. 1700–1600 cal yr BP. Charcoal influx values from ca. 1700–1500 cal yr BP indicate the period of the most consistently elevated burning in the record up to this point. Fire then declines and remains relatively low until ca. 1216 cal yr BP. The Yok Bolum record indicates the period 1510–1290 cal yr BP to be a period of sustained, anomalously high rainfall, with low variability indicated by the lowest standard deviations in $\delta^{18}O$ and $\delta^{13}C$ of the record. The period from ca. 1800–1250 cal yr BP in the Macal Chasm record also shows a period of wetter climate coupled with low deviations in $\delta^{18}O$ and $\delta^{13}C$. Though fire declines at Pine Pond during this inferred wetter period, charcoal values still remain relatively high compared to the Archaic Period. Forest taxa remain lower and pine pollen levels generally increase. There are also more woody savanna indicator taxa (Quercus, *Myrica*, *Byrsonima*) that may be associated with wetter conditions. Both periods of significantly increased fire activity correspond with periods of drought, but they also agree with the current archaeological understanding of the timing and scale of heightened human activity in the region. The Classic Era may be when humans first began to exert heavy and sustained influence on fire regimes around Pine Pond.

Though there are no large stone monuments in the immediate vicinity of Pine Pond that predate the Classic Era, excavations of the nearby saltworks reveal a significant coastal trade network dating at least as far back as 1450 cal yr BP (Robinson and McKillop 2013). Salt work production appears to have peaked ca. 250–500 CE (1700–1450 cal yr BP). Elevated levels of human deforestation in the area during this time is potentially reflected in the lower percentages of pine pollen than might be expected with elevated burning. While explicitly avoided for salt production, pine may have been harvested for other building purposes and/or was part of a larger commercial trade. Further afield, Walsh et al. (2014) attributed a high incidence of fire ca. 1500–1300 cal yr BP at Agua Caliente to higher rainfall that supported an increase in regional population, and thus higher levels of slash-and-burn agriculture. Fire then declines during the later Terminal Classic period (1150–950 cal yr BP). This period coincides with the Terminal Classic Drought, evident in both the Macal Chasm (1200–1050 cal yr BP) and Yok Bolum (1175–1020 cal yr BP) records as well as other paleoenvironmental reconstructions from the Maya lowlands (Wahl et al. 2013). In contrast to the findings of Walsh et al. (2014) at Agua Caliente, the Pine Pond record shows a notably elevated level of burning during the Terminal Classic Drought (Figure 21).

Archaeological evidence shows that the Classic Era was a time of high population density in southern Belize. Uxbenka is the largest excavated permanent settlement in the region, but ten other large settlements with stone monuments date to the Classic Era along with an estimated one-hundred smaller communities spread across the southern coastal plain (Prufer et al. 2008). Archaeological evidence suggests the regional population as a whole peaked ca. 550–900 CE (1400–1050 cal yr BP), while the population around the closest stone monument site to Pine Pond, Nim Li Punit (ca. 20 km SW of Pine Pond), peaked 734–810 CE (1216–1140 cal yr BP) (Braswell and Prufer 2009).

The ruins of Nim Li Punit currently lay underneath the canopy of a lowland broadleaf forest. Anderson and Wahl (2016) demonstrated that fires, either sparked by lightning or humans, are compatible with seasonally dry lowland broadleaf forests of the Petén in Guatemala dating back to at least the mid-Holocene. The climate of southern Belize is less seasonally dry than the Petén of northern Guatemala, but changes in microclimates influenced by shifting savanna ecotones may be similar. Deforestation of riparian zones within the savanna and possibly further afield to fuel nearby saltworks production ca. 600–900 CE (1350–1050 cal yr BP) may have enhanced the effects of the Terminal Classic Drought by creating conditions more conducive to the spread of fire by altering evapotranspiration and relative humidity in the local area. This drier climate may have facilitated a broader spread of anthropogenic fires ignited for hunting in the savanna or agricultural clearance in the nearby broadleaf forests. Walsh et al. (2014) linked a decline in charcoal influx during the later Terminal Classic to site abandonment and moist broadleaf forest regrowth that impeded the spread of either natural or human fires. This contradiction with the Pine Pond record indicates the influence of microclimates, edaphic factors, and the possible differences in the nature and timing of human-land use practices on ecosystem development and maintenance.

PostClassic Period (950–450 cal yr BP)

The Yok Bolum record documents an extended period of drought ca. 950–775 cal yr BP. This arid interval also represents the driest interval of the entire 5300-year Macal Chasm record. The highest charcoal influx up until this time occurs during the PostClassic Period, with peaks at 870 cal yr BP and 832 cal yr BP. A slight decline in burning until 618 cal yr BP coincides with a wetter climate while a subsequent increase in burning across the rest of the PostClassic era coincides with drier conditions as inferred from the Yok Bolum record. However, while the high charcoal influx values coincide with drier conditions, these findings conflict with paleoenvironmental records from across the Maya lowlands (Dull et al. 2007, Rushton et al.

2013, Wahl et al. 2013) as well as with Walsh et al. (2014), who found the PostClassic period to contain the consistently lowest levels of fire activity in the Agua Caliente record. The speleothem climate records indicate the most prolonged, sustained periods of aridity in the last ca. 5300 cal yr BP, which may have facilitated or driven elevated burning at Pine Pond. However, humans may have also driven fire occurrence. As major settlements such as Uxbenka were abandoned, people who continued to live in the region may have reverted to less efficient uses of fire for agriculture or hunting, leading to more fires spreading into the area around Pine Pond.

Colonial Period (450 cal yr BP to present)

During the Colonial Period, charcoal influx values remain consistently high. After the highest value in the record at 44 cal yr BP, values decline into the mud-water-interface core, but the level of burning still remains comparable to that seen during the Classic Era. These findings disagree with a general trend of dramatic declines in burning across the region attributed to population die-off during the time of European conquest (Nevle and Bird 2008, Dull et al. 2010). Based on a synthesis of charcoal records across the Americas, Power et al. (2013) concluded that climate was the predominant driver of reduced biomass burning in the Little Ice Age, rather than indigenous population collapse. Power et al. suggested that colder temperatures and reduced precipitation during the Little Ice Age dampened fire occurrence. The general co-occurrence of higher fire with drier conditions at Pine Pond contradicts the interpretation by Power et al. of fires being potentially less common during the Little Ice Age.

The increase in burning during the Colonial Period around Pine Pond also conflicts with charcoal data from other paleoenvironmental reconstructions in the surrounding Maya lowlands

(Rushton et al. 2013, Wahl et al. 2013), but it is consistent with findings of Walsh et al. (2014). The authors attributed an increase burning ca. 1500 CE to a proliferation of small farming communities in the area and a return to more fire-intensive, slash-and-burn agriculture. In accordance with Cathy's Smith's research, the major rise in burning and pine at Pine Pond during the colonial period likely corresponds to increased settlement by Maya immigrants and a heavier colonial presence and resource exploitation by the British. A drier climate during and after the Little Ice Age in tandem with increased regional deforestation and agriculture may have facilitated the spread of human-set fires, widening the expanse of savanna around Pine Pond. Given the current cultural practice of burning, it is highly likely an increase in population resulted in an increase of human-ignited fires whose spread was facilitated by the extremely dry conditions during and after the Little Ice Age.

CHAPTER VII

CONCLUSIONS

Taken collectively, the multi-proxy analyses presented here indicate the presence of a neotropical savanna ecosystem, with varying densities and compositions of hardwood and herbaceous taxa, for at least the last ca. 4800 years in the area around Pine Pond. The continual presence of charcoal points to the importance of fire in the savanna ecosystem well before the rise of slash-and-burn agricultural practices or indeed any significant known human impact in the area. The pollen and isotope evidence for the continued persistence of a savanna ecosystem across wet/dry climatic oscillations also implicates a strong edaphic control on vegetation. Burning alone does not appear to explain vegetation composition.

Generally, changes in fire track changes in climate as inferred from speleothem and other paleoenvironmental records in the region. Fire tends to increase during drier periods, with relatively less burning during wetter periods. Beginning in the Classic Maya era, changes in fire occurrence may have been a product of changing land use practices in addition to climate changes. Burning by local indigenous groups may have been exacerbated by an increased incidence of drought during the Terminal Classic Drought and into the Little Ice Age. The finding of human-influenced fire regimes since at least the early Classic Maya era holds important conservation implications.

Human pressure on savanna ecosystems across Belize has continued to increase across recent decades, whether it be through conversion to fallow pasture, slash-and-burn agricultural practices, or increased burning associated with poaching. The empirical data presented here coupled with Cathy Smith's historical documentation of fire management may aid current ecosystem conservation efforts. The charcoal record shows that the current level of burning around Pine Pond is not unprecedented. Modern day charcoal influx values are comparable to values during the Classic Maya Era. Burning by local indigenous groups has influenced and shaped ecosystem development for hundreds of years. This long-term paleoenvironmental reconstruction may inform local management and conservation groups on how best to manage and control fire occurrence in the face of current human disturbances.

This is one of the few paleolimnological studies in the Maya lowlands that combines pollen, charcoal, and bulk stable carbon and nitrogen isotope analyses to reconstruct the environmental history of a Central American savanna. The data add to the growing body of paleoenvironmental and archaeological work within the region. It also adds to the relatively understudied topic of Central American savanna development. Edaphic factors appear to have played a strong role in savanna maintenance around Pine Pond, as opposed to sustained levels of human-set fires as inferred in other studies in the region. More studies are needed to better understand the heterogeneity of edaphic factors, climate, and human land-use practices on the development and maintenance of savannas across the region.

Future charcoal and pollen analyses on the Pine Pond cores may be useful in providing a more detailed history of climate change and human impact. Pollen grains and fern spore types and morphologies could be identified in more detail. Identifying and quantifying pine stomata may generate a more accurate record of changing pine stand density in the vicinity of Pine Pond. Pine needles are heavier, less numerous, and less well dispersed than pollen, so their stomata in lake sediments provide evidence of local pine presence (MacDonald 2001). Analysis of microscopic charcoal particles on pollen slides could provide a record of fire history with a greater regional signal than the macroscopic charcoal analysis provides. These additional analyses combined with more pollen counts would further refine the records of Late Holocene environmental conditions at Pine Pond.

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APPENDICES

Appendix A

Pollen Processing Procedure

The following procedure way used to process sediment samples from Pine Pond for pollen analysis. Samples were processed in 15 ml polypropylene centrifuge tubes. The centrifuge used was a Thermo Scientific CL2 benchtop centrifuge with a 6 x 15 ml swinging bucket rotor. All centrifugations were carried out 3000 rpm.

1. Add 1 *Lycopodium* tablet to each centrifuge tube. Record the batch number for these control spores in the lab book.
2. Add a few ml 10% HCl, and let reaction proceed; slowly fill tubes until there is about 10 ml in each tube. Stir well, and place in hot water bath for 3 minutes. Remove from bath, centrifuge for 2 minutes, and decant.
3. Add hot distilled water, stir, centrifuge for 2 minutes, and decant. Repeat for a total of 2 washes.
4. Add about 10 ml 5% KOH, stir, remove stick, and place in boiling bath for 10 minutes, stirring after 5 minutes. Remove from bath and stir again. Centrifuge 2 minutes and decant.
5. Wash 4 times with hot distilled water. Centrifuge for 2 minutes each time.
6. Fill tubes about halfway with distilled water, stir, and pour through 125 µm mesh screen, collecting liquid in a labeled beaker underneath. Use a squirt bottle of distilled water to wash the screen, and to wash out any material remaining in the centrifuge tube.
7. Centrifuge down material in beaker by repeatedly pouring beaker contents into correct tube, centrifuging for 2 minutes, and decanting.
8. Add 8 ml of 49–52% HF and stir. Place tubes in boiling bath for 20 minutes, stirring after 10 minutes. Centrifuge 2 minutes and decant.
9. Add 10 ml 10% HCl. Stir well, and place in hot water bath for 3 minutes. Remove from bath, centrifuge for 2 minutes, and decant.
10. Add 10 ml hot Alconox® solution, made by dissolving 4.9 cm^3 dry commercial Alconox® powder in 1000 ml distilled water. Stir well and let sit for 5 minutes. Then centrifuge and decant.

11. Add more than 10 ml hot distilled water to each tube, so top of water comes close to top of tube. Stir, centrifuge for 2 minutes, and decant.

    Assuming that no samples need treatment with HF, continue washing with hot distilled water as above for a total of 3 water washes.

12. Add 10 ml of glacial acetic acid, stir, centrifuge for 2 minutes, and decant.

13. Make acetolysis mixture by mixing together 9 parts acetic anhydride and 1 part concentrated sulfuric acid. Add about 8 ml to each tube and stir. Remove stirring sticks and place in boiling bath for 5 minutes. Stir after 2.5 minutes. Centrifuge for 2 minutes and decant.

14. Add 10 ml glacial acetic acid, stir, centrifuge for 2 minutes and decant.

15. Wash with hot distilled water, centrifuge and decant.

16. Add 10 ml 5% KOH, stir, remove sticks, and heat in vigorously boiling bath for 5 minutes. Stir after 2.5 minutes. After 5 minutes, centrifuge for 2 minutes, and decant.

17. Add 10 ml hot distilled water, centrifuge for 2 minutes, and decant for a total of 3 washes.

18. After decanting last water wash, use the vortex genie for 20 seconds to mix sediment in tube.

19. Add 1 drop of safranin stain to each tube. Use vortex genie for 10 seconds. Add distilled water to make 10 ml. Stir, centrifuge for 2 minutes, and decant.

20. Add a few ml TBA, use the vortex genie for 20 seconds. Fill to 10 ml with TBA, stir, centrifuge for 2 minutes, and decant.

21. Add 10 ml TBA, stir, centrifuge for 2 minutes, and decant.

22. Vibrate samples using the vortex genie to mix the small amount of TBA left in the tubes with the microfossils. Centrifuge down vials.

23. Add several drops of 2000 cs silicone oil to each vial. Stir with a clean toothpick.

24. Place uncorked samples in the dust free cabinet to let the TBA evaporate. Stir again after one hour, adding more silicone oil if necessary.

25. Check samples the following day; if there is no alcohol smell, cap the samples. If the alcohol smell persists, give them more time to evaporate.

Appendix B

Isotope Sampling and Preparation Procedure

This is the standard protocol used for preparing bulk sediment samples for stable carbon and nitrogen isotope analysis in the Laboratory of Paleoenvironmental Research at the University of Tennessee. The first draft of this document was prepared by Joanne Ballard in July 2014, based on discussion with Chad Lane. The document was later edited by Erik Johanson with input from Matthew Kerr in July 2015. Matthew Kerr and Sally Horn made further edits in February and July 2019, and Sally Horn updated the formatting in October 2019.

Materials and Equipment Needed:

Nitrile gloves
Safety goggles
1 cc sampler and “pusher”
Metal spatulas
Agate mortar and pestle
Ethanol
Kim-Wipes
Distilled water
12M HCl
Scintillation vials
Glass Petri dish
Glass desiccator with ceramic plate
Sharpies
Meter stick for sampling core section
Core log sheets
Baking soda

Plastic waste bucket

Electric griddle

Fume hood

Oven

Note: If you are working with quarantined sediments or soils, you need to follow additional instructions and training to address this.

Instructions:

1. Gather the items listed above.

2. Pull the core section(s) from the cold room.

3. Get the core log sheets for the relevant core sections.

4. Determine the depths to sample based on a sampling design previously developed with Sally Horn and others. Make sure you can match the section to the core log and that you are following the sampling design and are confident that you are sampling in the right place. Cores are valuable and haphazard sampling will jeopardize your study and others.

5. Record sample locations on the core logs (using an ink pen) with your initials and date.

6. Label a pre-combusted scintillation vial with the site and depth info for each sample on the vial and cap. Use fine black Sharpie on the glass vial and ultrafine black Sharpie on the lid. Use a sampling code like this: Emerald Pond 2003 6.35-7.35m/14–15 cm.

7. Prepare a clean surface on the sediment using a spatula to scrape across the depth.

8. Take a sample of sediment from the core using a 1 cc sampler (unless another volume is required).

9. Place the sample in the first scintillation vial and set aside, with lid. Do not get sediment on the top of the vial or the threads.

10. Continue until all samples are ready in the vials. You will need to use a clean sampler and pusher each time so as not to contaminate the samples. Wash them as you go. Do not dry with a paper towel. Instead, put a glass dish in the oven and put clean, wet implements on the dish to dry while you work. An oven temperature of 80–100 °C is workable (don't burn your hand on implements).

11. When you are finished sampling, remove vial lids, wrap lids in clean Al foil, and place the vials on a pie plate or glass dish. Put the pie plate or dish in an oven and dry at 50 °C for 16–24 hours until dry. Put the foil packet with lids on a counter nearby. You can label it with a Sharpie with your name, for example, "Matt K's lids." The foil envelope is to keep dust off.

12. Remove vials from oven and let cool to room temperature in a box desiccator. When cool put the lids on them.

13. Grind each dried sample to a fine powder using the agate mortar and pestle.

14. Return the ground sample to its correct vial and cap it.

15. Clean the mortar and pestle between samples with ethanol and Kim-Wipes.

16. At this point, you may split the samples into different workflows: non-acidified ($\delta^{15}N$, XRD, etc.) and acidified ($\delta^{13}C$). Transfer roughly half of each sample to a new pre-combusted scintillation vial. The new vial should be labeled the same as the original, but with the addition of an "A" indicating acidified.

17. The original non-acidified vials should be recapped and ready for non-acidified analyses.

18. Once all samples have been ground and split, set up a glass desiccator in a fume hood for acid fumigation. Make sure you remove everything you are not using from the hood. The acid fumes will ruin everything metal, including centrifuges. **NEVER use the right fume hood in 422 that contains equipment for compound-specific isotope sample preparation!** If the hood contains hazardous waste, move it temporarily to an adjacent hood.

19. Wearing nitrile gloves and goggles, pour 12M HCl (37%, concentrated) into a glass Petri dish and place the Petri dish in the bottom of the desiccator. You want to use as much acid as you can have in the dish without slopping it all over the place when moving the dish into the desiccator.

20. Place the ceramic plate in the desiccator. It fits on a rim in the jar.

21. Moisten each sample with distilled water. You want them just moist, not swimming.

22. Carefully place the prepared sample vials on the ceramic plate.

23. Did you remember to moisten the samples? Place the lid on the desiccator and close the fume hood. No lubricant is used on the seal. Put a note on the hood sash so people are warned about the fumes.

24. Leave the samples in the desiccator to fumigate for 3–5 hours, depending on carbonate content (higher carbonate content requires longer fumigation time).

25. Vent the desiccator overnight by slightly opening the lid by a quarter turn.

26. Remove the vials from the desiccator and place them on an electric griddle in the fume hood on the lowest setting, "warm" (estimated to be ca. 50 °C) to dry. Leave them to dry for at least 48 hours. It's not possible to over-dry them at this low temperature and longer is better.

27. Put the used acid and Petri dish into a plastic waste container in fume hood and neutralize it with baking soda. When it stops bubbling you can dispose of the waste and wash the equipment.

28. When dry, recap the vials and store until you are ready to load capsules for analysis (which needs to be soon, see below). You may need to "regrind" the acidified samples, but this is more like mashing them up with a tool of some sort.

29. Load isotope samples into tin capsules for analysis. *Ideally, acidified samples would be left to dry on the griddle for a few days and then immediately loaded into capsules and sent for analysis.* If it is going to take longer you may need to use silver capsules (more expensive) for the samples or double wrap them in two tin capsules. But even then, the goal is to run them soon. You can store the acidified samples in vials and later load the capsules right before it's time to send them for analysis. The idea with drying on the griddle is to drive off residual HCl, but you can never get every bit of it. The consequences of this are (a) your acidified samples will now be a bit hygroscopic, so you have to keep them dry so water doesn't screw up your analyses, and (b) residual HCl will eat everything metal, including tin capsules, the liners of scintillation vial caps, desiccator cabinets, and so forth. So, you need to plan ahead and aim for loading and sending off samples in a short time frame.

Note on sampling: Mass is more important than volume for <u>isotope work</u>. None of your isotope data will be expressed on a volumetric basis. We sample sediments by volume for convenience. The goal is to have enough material for all the analyses you plan to do. For bulk isotope analysis, you need enough carbon and nitrogen by mass in your samples to be detectable on the mass spectrometer – the target is 0.5 mg pure carbon in each sample with little variation between duplicates and other samples. This target will also provide enough nitrogen in the samples, except maybe in rare cases where the samples have very low N content, but there's no practical way to know that in advance.

Calculate sample masses to load into capsules for isotope analysis:

Target mass (mg) = (0.0005 / ((%ORG / 100) * 0.4)) * 1000

where 0.0005 accounts for the desired 0.5 mg of C, %ORG is the organic fraction on a dry mass basis estimated by LOI for the sampled level, 0.4 is the assumed fraction of carbon in organic matter, and 1000 gets it in mg units.

VITA

Jacob A. Cecil was born in Knoxville, Tennessee. He received his Bachelors of Arts in History, Global Studies, and Geography from the University of Tennessee. Jacob entered the Master's program at the University of Tennessee in 2017. He served as a teaching assistant for introductory physical geography and paleolimnology classes from 2017 to 2019.