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Late Pleistocene Climate, Vegetation, and Fire History from a Southern Appalachian Bog, Whiteoak Bottoms, Nantahala National Forest, North Carolina, U.S.A.

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Late Pleistocene Climate, Vegetation, and Fire History from a Southern Appalachian Bog, Whiteoak Bottoms, Nantahala National Forest, North Carolina, U.S.A.

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ABSTRACT

I examined loss-on-ignition, pollen, and charcoal evidence of climate, vegetation, and fire history at Whiteoak Bottoms (35º04'44"N, 83º31'50"W; 1032 m elevation), a peat-forming wetland located along the Nantahala River in western North Carolina. Previous research by J. McDonald and D. Leigh revealed that this wetland formed in a paleochannel of the Nantahala River between 15,000 and 14,000 cal yr BP. I obtained additional AMS radiocarbon dates, carried out high-resolution loss-on-ignition analysis, and examined pollen and microscopic charcoal assemblages in a 157-cm sediment core from the previous study. Radiocarbon dates and stratigraphic analyses indicate that much of the Holocene is missing from the Whiteoak Bottoms record; however, the site does contain an intact record of environmental conditions during the latest Pleistocene. Variations in organic matter content suggest variations in peat accumulation through the history of this wetland, likely driven in part by variations in effective moisture. Based on a comparison with GISP δ [delta] ^{18}O and Cariaco Basin sea surface temperature data, organic matter accumulation appears to trend with climate, generally increasing as temperatures decreased. Vegetation changes evident in the pollen record agree with the timing and trajectory of other pollen records for the region. These shifts in vegetation appear to correspond with changes in climate, potentially representing a response to shifts in the position of the Bermuda High. Fire indices reveal low incidence of fire during the late Pleistocene. In the early portion of the record, fire activity appears related to increasing temperature, and changes in fire activity seem to correspond with changes in vegetation.

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CHAPTER I
INTRODUCTION

Numerous proxies exist for reconstructing past environments. Tree rings, pollen grains, isotopes, conifer stomata, and charcoal particles, among others, have been used as proxies in the exploration of past climates, environments, and fire regimes. Although the southeastern United States has a long history of paleoecological investigation, few proxies other than pollen and tree rings have been widely applied within the region. Despite the fact that investigations into the environmental history of the southeastern United States have been ongoing since at least the 1950s, many questions remain unanswered.

One gap in knowledge concerns the environmental histories of southern Appalachian bogs and their surroundings. These highly diverse habitats are becoming increasingly rare due to development (Weakley and Schafale 1994). Because they often occupy the flattest ground within a rugged environment, they were often cleared and drained for agricultural and other purposes (Weakley and Schafale 1994). In the Southern Blue Ridge Province of North Carolina, non-alluvial wetlands support some 600 vascular plant species, including many endemic, disjunct, and relict species (Weakley and Schafale 1994; Pittillo 1994). Today, less than an estimated 400 hectares of these communities remain in the Southern Blue Ridge, and we currently know very little about the histories of these endangered ecosystems (Weakley and Schafale 1994). My thesis research concerns one of these wetlands, Whiteoak Bottoms in the Nantahala National Forest of North Carolina.

1.1 Research Questions

I examine evidence of past vegetation, fire, and climate at Whiteoak Bottoms by analyzing fossil pollen, sedimentary charcoal, and organic content in a sediment core from the

wetland. I then compare my results with local and regional paleoecological studies to answer the following questions:

1. Does the Whiteoak Bottoms core provide a continuous sedimentary record and what does core stratigraphy reveal about the environmental history of the site?

2. What does the late Pleistocene loss-on-ignition record at Whiteoak Bottoms reveal about late Pleistocene climate at the site, and how does this record compare with climate records for the Southeast?

3. What does the late Pleistocene pollen record at Whiteoak Bottoms reveal about vegetation and climate at the site, and how do the data compare with other late Pleistocene records of vegetation and climate for the Southeast?

4. What does the late Pleistocene charcoal record at Whiteoak Bottoms reveal about fire at the site, and how does this record compare to records of fire and climate for the Southeast?

1.2 Significance of Research

While the southeastern United States has been the subject of paleoecological study since at least the 1950s, several gaps in knowledge persist. First, few long-term fire reconstructions based on sedimentary charcoal exist. As a result, trends in fire activity over the Late Quaternary remain poorly understood. Second, few studies have combined charcoal and pollen analyses. The combination of these two proxies offers the opportunity to elucidate vegetation-fire interactions under varying climate conditions, providing knowledge that can help foster better understanding of our current and future environments. Third, Southern Appalachian bogs are a rapidly disappearing ecosystem (Pittillo 1994). Few studies have focused on these habitats or their vegetation histories (Weakley and Schafale 1994). These wetlands are hotspots of

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biodiversity and havens for rare endemic and disjunct flora, but the conditions responsible for the creation and maintenance of these communities are not well understood (Weakley and Schafale 1994). This lack of knowledge somewhat hinders the effective management of these unique wetlands. Finally, projected increases in evapotranspiration due to future climate change could lead to the reduction of moisture levels, resulting in drying of these ecosystems (Schultheis et al. 2010). This could result in increased decomposition, transforming these wetlands into carbon sources instead of sinks, and leading to the extirpation or extinction of a variety of wetland species (Schultheis et al. 2010). An understanding of the long-term history of both vegetation and fire at these sites could lead to more informed conservation and management decisions.

This thesis consists of seven chapters. In chapter two, I describe the site and environmental setting of my study area. I review literature pertaining to my methods and to the paleovegetation and paleoclimate of the southeastern United States in chapter three. The paleovegetation and paleoclimate subsections are presented geographically. In chapter four, I describe the laboratory methods used in my research. I present the results of my multi-proxy analysis in chapter five and discuss them within the framework of the southeastern U. S. in chapter six. In chapter seven, I summarize the key findings and conclusions of my research and make suggestions for further research.

CHAPTER II
STUDY AREA

2.1 Site Description

Whiteoak Bottoms (35º04'44"N, 83º31'50"W; 1032 m elevation) is a wetland of the type termed a "southern Appalachian bog" (Weakley and Schafale 1994), located in the Nantahala River Valley of the Nantahala National Forest, Macon County, North Carolina (Figure 1; Figure 2). This wetland is approximately 300 x 100 m in size and lies between the Nantahala River (on its western edge) and the western-facing slope of a neighboring ridge (McDonald and Leigh 2010).

McDonald (2010) reported a seasonally-fluctuating water table at the site, with water levels closest to the surface during winter; however, springs and seeps from the hill slope to the east appear to maintain water levels to within 15 cm of the surface even during the summer. Potential anthropogenic influences on the site stem from a nearby Forest Service campground and road. Also, according to McDonald (2010), an attempt was made to convert the bog into agricultural lands in the past; however, the ditches used for this purpose are now largely filled with sediments, and as such do not affect the current hydrology at the site. In addition to this attempted agricultural conversion, the site was also affected by historic period timber harvesting. According to informational billboards at the site developed by the USDA Forest Service (pers. obs.), as much as 90% of what is today the Nantahala National Forest was cleared during the late 1800s and early 1900s. During this time, a logging camp was located near Whiteoak Bottoms and a rail line ran parallel to the Nantahala River on the opposite bank. In addition to these human disturbances, beaver ponds and canals currently exist on the northern end of the wetland (McDonald 2010).
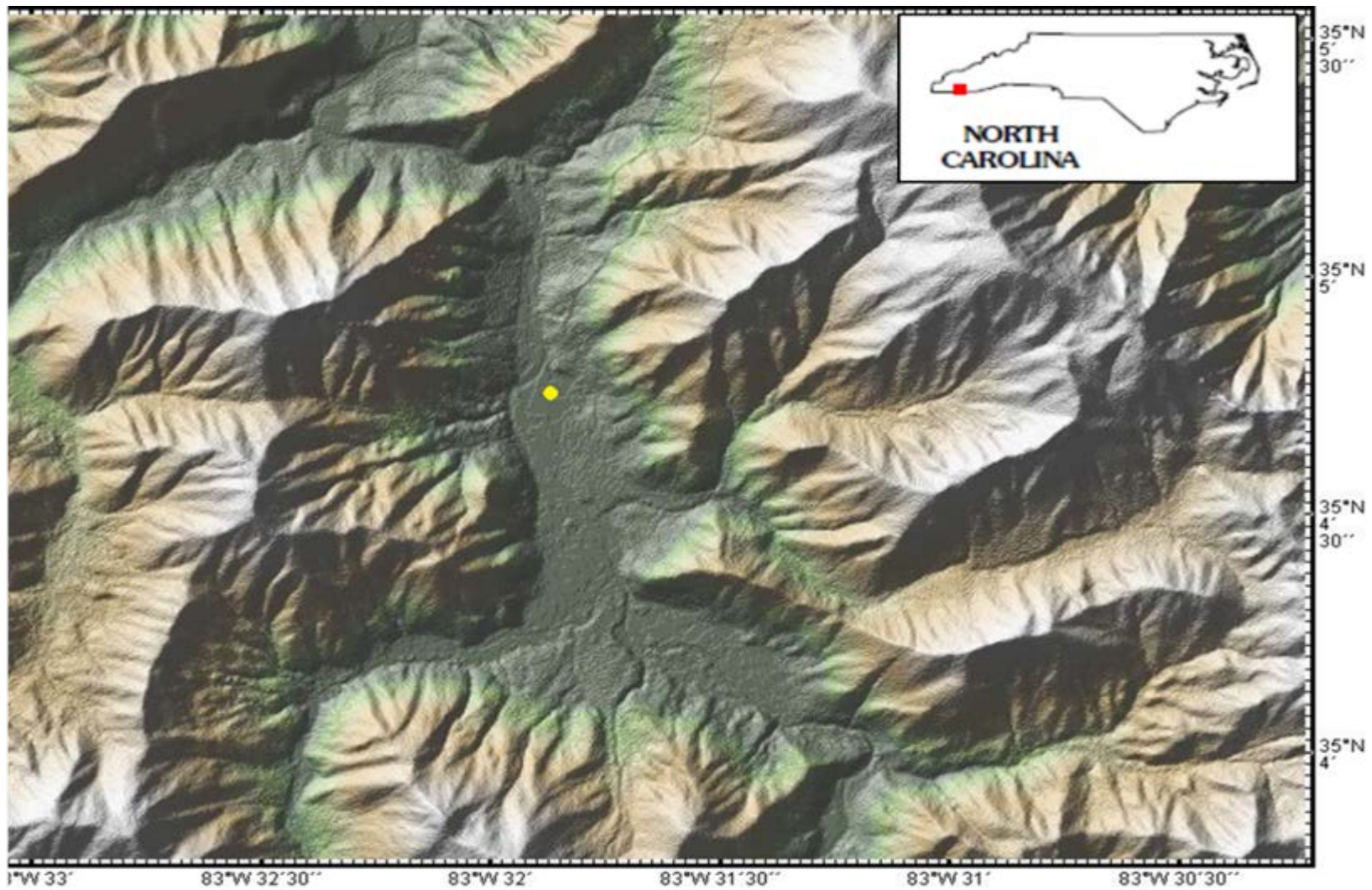
Figure 1. Digital Elevation Model for Whiteoak Bottoms, North Carolina. Figure created by author using mapping program and data available at http:/www.geomapapp.org and Ryan et al. (2009).

Figure 2. View looking south across the wetland at Whiteoak Bottoms, North Carolina. Photo courtesy of Sally P. Horn.

2.2 Climate

The annual temperature reported at the Coweeta Hydrologic Laboratory, located 9.5 km east of Whiteoak Bottoms, is 12.7 °C, with an average January temperature of 3.2 °C and average July temperature of 21.9 °C (Southeast Regional Climate Center 2011). Annual precipitation at the Coweeta Hydrologic Laboratory averages 189.6 cm, the majority of which occurs between December and March (Southeast Regional Climate Center 2011). In contrast, July and October are the driest months in the area (Southeast Regional Climate Center 2011).

2.3 Vegetation

Located within the oak-chestnut forest region of Braun (1950), Whiteoak Bottoms is classified by Weakley and Schafale (1994) as a Southern Appalachian Bog. Generally, these wetlands support a mosaic of shrub- and herb-dominated vegetation (Figure 2) (Weakley and Schafale 1994). Trees such as *Acer rubrum* L., *Betula alleghaniensis* Britton, *Nyssa sylvatica* Marsh., *Pinus strobus* L., *Tsuga canadensis* (L.) Carrière, and *Picea rubens* Sarg. may occur as scattered individuals or patches within the wetland or along its edge (Weakley and Schafale 1994; Spira 2011). Common shrubs can include *Alnus serrulata* (Aiton) Willd., *Ilex verticillata* (L.) A. Gray , *Rosa palustris* Marshall, and occasionally *Salix sp*. (Weakley and Schafale 1994; Spira 2011). The herb layer may include a variety of Cyperaceae sp., *Sphagnum* sp., *Osmunda* ferns, *Saxifraga pensylvanica* L., *Sagittaria latifolia* Willd., and rare species such as *Caltha palustris* L., *Helonias bullata* L., and *Lilium grayi* S. Watson (Weakley and Schafale 1994; Spira 2011).

The pollen record at Whiteoak Bottoms also receives input from neighboring cove and oak-chestnut forest communities on the surrounding slopes and ridgetops. Spira (2011) recognized two types of cove forest, rich cove and acidic cove. Rich cove forest has a greater

variety of mesophytic tree species and a diverse herbaceous layer. Acidic cove forests have fewer tree species, a dense ericaceous shrub layer, and a relatively sparse herbaceous layer. According to Spira (2011), common arboreal species for rich coves include *Liriodendron tulipifera* L., *Fagus grandifolia* Ehrh., *Tilia americana* L., and *Acer saccharum* Marsh. Less common species include *Fraxinus americana* L., *Betula lenta* L., *Ostrya virginiana* (Mill.) K. Koch, *Carpinus caroliniana* Walter, and *Acer pensylvanicum* L. Spira (2011) listed *Acer rubrum*, *Betula lenta*, *Liriodendron tulipifera*, and *Tsuga canadensis* as the most abundant arboreal taxa in acidic cove forests, along with shrubs such as *Kalmia latifolia* L. and *Rhododendron maximum* L. among others. Less common tree taxa include *Betula alleghaniensis*, *Fagus grandifolia* Ehrh., *Ilex opaca* Aiton, *Magnolia* sp., *Pinus strobus* L., and *Quercus rubra* L.

CHAPTER III
LITERATURE REVIEW

3.1 Peatlands and Loss-on-ignition as Paleoclimate Indicators

Peatlands are ecosystems in which net primary productivity exceeds decomposition, resulting in the long-term deposition of organic matter (Vitt 2006). Temperatures and precipitation are key factors in peatland initiation and maintenance due to their effect on moisture balance (Charman 2002; Rydin and Jeglum 2006; Vitt 2006). In peatlands, moisture balance acts as a catalyst for destructive or generative forces within the wetland (Charman 2002). While peat in cooler climates forms where soils are permanently waterlogged and temperatures are low, peat in the tropics may develop where high precipitation and poor drainage conditions exist (Rydin and Jeglum 2006). A positive moisture balance is the key factor in the maintenance of peatlands. Given these strong links with climate, peatlands could act as sensitive indicators of climatic change.

Loss-on-ignition (LOI) analyses measure the organic content of sediments. This content derives from both autochthonous and allochthonous sources (Dean and Gorham 1998), and varies due to changes in sediment composition and sediment accumulation (Shuman 2003). Factors such as environmental productivity, inorganic input, and decomposition affect the composition of sediments, while basin morphology and size, along with water level, affect accumulation (Shuman 2003). Both composition and accumulation are affected by climate, creating the potential for loss-on-ignition analyses to act as a paleoclimate proxy.

Sullivan et al. (1999) examined bulk density and organic matter content in sediments collected from two sub-alpine lakes in western Colorado and concluded that changes reflected in the organic matter content related to changes in temperature through time. Nesje and Dahl

(2001) performed high-resolution LOI testing on sediments from five lakes in southern Norway. Based on a comparison of their record with Greenland ice core data, they concluded that their LOI record provided a sensitive proxy for surface air temperature. Batterbee et al. (2001) interpreted variations in LOI, $\delta^{13}C$, and chironomid data from a small lake in the Cairngorm Mountains of Scotland as reflecting lake productivity. They hypothesized that these changes resulted from changes in climate. Taylor (2003), in undergraduate research directed by Sullivan, visually correlated LOI data from Church Camp Fen, Colorado, with Greenland ice core records, finding evidence for the Oldest Dryas, the Bølling, the Older Dryas, the Allerød, the Inter-Allerød Cold Period, and the Younger Dryas. Gilmore and Sullivan (2010) examined a series of small fens in eastern Colorado. Based on the analysis of peat from these and other sites, they proposed that the organic matter content in wetlands acts as a temperature proxy. Combining their LOI and peat humification data, they linked patterns in their data to known climatic anomalies, providing temperature and moisture proxy data for the last 2500 cal yr BP (Gilmore and Sullivan 2010).

3.2 Long-term Fire History

The lack of fire can have as dramatic an effect on the environment as a roaring blaze (see Dumas et al. 2007). Every ecosystem has an environmentally and biologically determined fire return interval. Fire suppression efforts, by derailing the natural fire cycle, have changed the composition of numerous forests in North America (Bratton and Meier 1998; Nowacki and Abrams 2008; Lafon 2010). Another byproduct of the suppression era is that we no longer know the natural periodicity of fire in particular environments (Weakley and Schafale 1994; Lynch and Clark 2002). As a result, a need exists for long-term fire history studies.

Several proxies are available for reconstructing the fire regime of an area. Each comes with its own strengths, weaknesses, and requirements. These include the tabulation of pollen types by degree of fire tolerance (Whitehead and Sheehan 1985; Delcourt et al. 1986; Whitlock and Larsen 2001), the dendrochronological dating of fire scars and stand age characteristics (Lafon 2010), and the quantification of charcoal in sediments and soils (Clark 1982; Whitlock and Larsen 2001; Lynch and Clark 2002; Gavin et al. 2007; Fesenmeyer and Christensen 2010). While the dendrochronological analysis of fire scars provides the most precise method for determining past fire occurrence, rampant logging in the Southeast over the last few centuries combined with high rates of wood decomposition have removed from most areas the trees and remnant wood necessary for long-term reconstructions. This leaves the analysis of charcoal in sediments and soils as a key proxy for the reconstruction of long-term fire histories.

3.2.1 Charcoal as a Proxy for Past Fire

Whitlock and Larsen (2001) listed five methods for charcoal analysis. Of these five, only three (the pollen slide, sieving, and thin section methods) appear repeatedly in the literature. For the pollen slide method, microscopic charcoal is quantified from slides prepared for pollen analysis (Clark and Royal 1995; Asselin and Payette 2005). The macrofossil, or sieving method, involves washing sediments through a series of sieves and then quantifying the macroscopic charcoal particles retained on each sieve (Whitlock and Larsen 2001). The thin section method requires mounting sediments in epoxy and thin-sectioning the sediments before counting the charcoal fragments (Whitlock and Larsen 2001). Each method has its pros and cons. Ultimately, however, the choice of method depends on the research objectives, the resources at hand, and the conditions at the site. Thin sections provide annual (if sediments are laminated) to millennial scale resolution; macroscopic charcoal can provide decadal to millennial resolution; and

microscopic charcoal reveals fire activity at the centennial or millennial scale (Whitlock and Larsen 2001).

Various indices for quantifying microscopic charcoal appear in the literature. These include charcoal area concentration (mm^2/cm^3), charcoal area influx ($mm^2/cm^2/yr$), and charcoal to pollen ratios (μm^2/pollen grain) (Whitlock and Larsen 2001; Patterson 2005; Tinner et al. 2006; Albritton 2009). For the interpretation of these indices, values are compared within the individual records and across similar records to identify periods of greater and lesser fire activity. Since interpretations are largely a factor of comparison, there are no fixed values defining what constitutes high, moderate, or low fire activity categories. Albritton (2009) reported charcoal values for all three indices based on an analysis of pond sediments from a pine rockland community in South Florida. Pine rocklands are adapted to frequent, low-intensity fires (Albritton 2009). Charcoal area concentration ranged between 80 and 862 mm^2/cm^3 and charcoal area influx reached over 200 $mm^2/cm^2/yr$ (Albritton 2009). Mean charcoal area concentration measured 308 mm^2/cm^3. In 20 of 21 samples analyzed charcoal area concentrations exceeded 100 mm^2/cm^3, and in 15 it exceeded 200 mm^2/cm^3 (Albritton 2009). Mean charcoal influx was 81 $mm^2/cm^2/yr$. Eighteen of 22 samples had influx values greater than 25 $mm^2/cm^2/yr$ (Albritton 2009). Albritton (2009) reported charcoal to pollen ratios of up to ca. 20,000 μm^2/pollen grain, with a mean of 7413 μm^2/pollen grain. Only two samples had charcoal to pollen ratios less than approximately 2500 μm^2/pollen grain; 13 exceeded 5000 μm^2/pollen grain (Albritton 2009).

In contrast, Tinner et al. (2006) estimated mean charcoal influx values of only 4.0 $\pm$7.2 $mm^2/cm^2/yr$ and 0.8 $\pm$0.6 $mm^2/cm^2/yr$ for two lakes in boreal Alaska. These area influx values were estimated from charcoal particle counts using equations in Tinner and Hu (2003). Charcoal

concentration values ranged from 0 to 87.5 mm^2/cm^3 and 0 to 157.1 mm^2/cm^3, respectively (Tinner et al. 2006). Fuller et al. (1998) reported charcoal to pollen ratios ranging from less than 100 to 600 μm^2/pollen grain for the past 1000 years at a series of lakes in north-central Massachusetts. The surrounding vegetation at these sites was classified as hardwood-hemlock-white pine forest. Values less than 200 μm^2/pollen grain were the most common (Fuller et al. 1998). MacDonald et al. (1991) reported ratios generally below 500 for a site in the boreal forest of northern Alberta Canada. They found a few higher ratios, but only two were above 1000 for their approximately 200 year record. Patterson (2005) reported charcoal to pollen ratios ranging between 0 and 5000 μm^2/pollen grain for a series of mesic deciduous and mixed conifer-deciduous forested ponds in Virginia, New York, Massachusetts, and Maine. Four of the six sites had charcoal to pollen ratios of less than 600 μm^2/pollen grain, with minimum values between 0 and 100 μm^2/pollen grain (Patterson 2005). Based on prior work from coastal Maine, Patterson (2005) suggested as a general guideline that charcoal to pollen ratios greater than 150 to 200 μm^2/pollen grain indicate fires within the watershed, while lower values probably represent fires farther afield. The above findings provide metrics by which sites in the Southeast may be interpreted.

3.2.2 Fire – Vegetation – Climate Relationships

Research on vegetation and fire history has improved our understanding of the sensitivity of plants to fire. Kirwan and Shugart (2000) used soil charcoal and fire models to test the hypothesis that *Fagus grandifolia* is an indicator of fire. They found that *Fagus grandifolia* and *Acer rubrum* are fire-intolerant species (Kirwan and Shugart 2000). Rogers (1978) found that *Tsuga canadensis* is extremely fire-intolerant. As all three species generally occupy more mesic sites, their occurrence in the pollen record indicates infrequent fires.

The combination of pollen and charcoal analysis has allowed researchers to analyze the effects of climate on fire regimes. At Browns Pond in Virginia, Kneller and Peteet (1999) associated warming after the Younger Dryas with a vegetation change from *Abies* to *Tsuga* that was accompanied by a change in fire activity. Millspaugh et al. (2000) combined macroscopic charcoal, pollen, and dendrochronological data to infer climate-fire relationships over the past 17,000 years in Yellowstone National Park. They found correspondences between their charcoal record and July insolation values, suggesting long-term control of fire activity by climate, even when vegetation remained constant (Millspaugh et al. 2000).

Gavin et al. (2006) found that climate is a major factor in the occurrence of forest fires in southeastern British Columbia and that this influence plays out over multiple spatial and temporal scales. In an examination of fire in the northwestern U.S., Whitlock et al. (2003) found similar results. Carcaillet et al. (2001) provided further support for this view at their study sites in eastern Canada. Grissino-Mayer and Swetnam (2000) linked changes in the seasonality of precipitation in the western U. S. to changes in the seasonality of fire. Le Goff et al. (2007) linked shifts in the fire regime near Quebec to climate teleconnections. They were able to connect decadal fire activity to the Pacific Decadal Oscillation (PDO). These connections serve to reinforce the bond between fire and climate.

Umbanhowar (2004) found that local differences in topography and fuels can lead to variations in climate and fire regime even between nearby sites in Minnesota. These differences point to the potential for variability between neighboring sites whether they are in the same general area, region, or climate zone. Gavin et al. (2007) found that this variability can thwart direct connections between fire and climate. Additionally, the type of vegetation at a site can affect the flammability of the site (Grissino-Mayer and Swetnam 2000; Gavin et al. 2007;

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Higuera et al. 2009). Tinner et al. (2006) tracked changing fire return intervals across changes in vegetation in Alaska. Their analysis revealed that changes in fuel species act to lengthen or shorten fire return intervals, depending on the fuel involved. Higuera et al. (2009) showed that shifts in vegetation over the last 15,000 years modified the fire regime of a site in Alaska by changing fuel loads. Marlon et al. (2009) discovered a strong connection between fire activity and rapid climate change. They found that evidence of fire increased in North America from the late glacial through the Younger Dryas period. Substantial increases in activity were apparent at 13,900, 13,200, and 11,700 cal yr BP, periods of rapid climate change (Marlon et al. 2009). This tendency for increased fire during these episodes requires further exploration, given the multitude of climatic oscillations shown in the pollen-climate record.

Lynch et al. (2004) showed that fires in the boreal forest of Alaska were more frequent under wetter conditions than dry ones. While this may seem counterintuitive, they argued that lightning strikes and seasonal moisture variability result in more frequent fires (Lynch et al. 2004). Wetter conditions can bring an increase in fuels; this increase can lead to a more flammable landscape (Grissino-Mayer and Swetnam 2000; Brown et al. 2005). On the Great Plains, Brown et al. (2005) found that fire was most prevalent following wet periods when grass cover was most extensive. Grissino-Mayer and Swetnam (2000) found that moisture levels in preceding years, by affecting the production and distribution of fine fuels, affected the flammability of their study site in New Mexico. These findings may not be restricted to the boreal forest, the Great Plains, or the American Southwest. Further investigation could reveal that wet season fuel build up followed by fire is the norm for many ecosystems.

3.3 Previous Research in the Southeastern United States

3.3.1 Vegetation History

Pollen diagrams spanning the Pleistocene to Holocene transitions in the southeastern United States generally reveal a tripartite pattern of vegetation changes that proceeded in a time transgressive manner from south to north. In the southern Appalachian mountains, coniferous forests dominated by spruce and pine gave way to more mesic deciduous forests consisting of oak, birch, hemlock, beech, *Ostrya/Carpinus*, willow, and elm, among others. As they transitioned to mesic, cove-like forests, these forests were probably similar to northern hardwood forests. They later gave way to oak-chestnut forests, and finally to the modern oak-hickory forest.

Today, in the Southern Appalachians, spruce-fir forest consisting of *Picea rubens* and *Abies fraseri* are restricted to elevations above 1372m (Spria 2011). *Abies fraseri* grows above 1524m, but dominates above 1829 m; *Picea rubens* is typically found between 1372 and 1890 m (Spira 2011).

At Cranberry Glades, West Virginia, Watts (1979) documented the transition from a forest dominated by spruce, pine, and fir to a forest rich in more mesic species shortly after 14,000 cal yr BP[1]. A change to a more xeric deciduous forest (oak, chestnut, and hickory) took place after 11,600 cal yr BP (Watts 1979). At Buckle's Bog, Maryland, a spruce-fir-pine forest existed from at least 14,250 to 11,600 cal yr BP (Maxwell and Davis 1972). After 11,600 cal yr BP, the forest changed to mesic deciduous before making a final transition to the modern oak-chestnut-hickory forest. At Browns Pond, Virginia, the coniferous to mesic deciduous transition

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[1] Calibrated ages mentioned in the text were interpolated from calibrated ages in original publications, or estimated from ^{14}C ages reported in the literature. For some sites, ^{14}C ages were calibrated using CALIB 6.0.2 and the dataset of Reimer (2009). For other sites, calibrated ages were approximated from ^{14}C ages based on conversions found at http://www.esd.ornl.gov/projects/qen/nerc14C.html.

occurred at approximately 11,700 cal yr BP (Kneller and Peteet 1993). The same pattern occurred at Potts Mountain Pond, Virginia, though less well dated, where coniferous forest of spruce-pine-fir transitioned to mesic deciduous forest prior to 10,000 cal yr BP, followed by a shift to oak-chestnut-hickory forest with some pine (Watts 1979). At Hack Pond in Virginia, the transition from coniferous to deciduous forest occurred at approximately 10,700 cal yr BP (Craig 1969).

West of the Appalachians, the sequence of vegetation change progressed in a similar manner. At Jackson Pond, Kentucky, Wilkins et al. (1991) inferred spruce-pine forest from about 24,500 to 12,900 cal yr BP. From about 11,600 to 8000 cal yr BP, mesic deciduous taxa dominated the landscape. Mesic taxa were replaced by oak-hickory-chestnut forest from approximately 8000 to 4500 cal yr BP. Based on my analysis of the Jackson Pond pollen data, I estimate that the transition from coniferous to deciduous forest occurred about 12,700 cal yr BP, based on the weighted means of the probability distributions of the calibrated radiocarbon dates. At Anderson Pond, Tennessee, coniferous forest gave way to a more mesic deciduous forest by 16,879 cal yr BP (Delcourt 1979; Ballard et al. 2012). Coniferous forest at Shady Valley Bog, in northeastern Tennessee, transitioned to mesic deciduous forest between 14,000 and 13,000 cal yr BP (Barclay 1957). At Bob Black and Quicksand Ponds, Georgia, the coniferous to deciduous vegetation change happened prior to 14,500 cal yr BP (Watts 1970). Until approximately 12,900 cal yr BP at Pigeon Marsh, Georgia, Watts (1975) reported pollen spectra dominated by pine and herbs, though pollen of deciduous forest taxa was also well represented. After 12,900 cal yr BP, mesic deciduous forest was established (Watts 1975). This forest was later replaced by a more xeric oak-hickory-chestnut-black gum forest, which persisted into modern times (Watts 1975).

On the coastal plain, a more hydric swamp or floodplain forest followed the mesic woodland, rather than the more xeric oak-hickory-chestnut forests seen in the piedmont and mountains. At Rockyhock Bay, North Carolina, Whitehead (1981) reported coniferous forest from around 24,500 to 11,600 cal yr BP. This forest was principally a pine-spruce-fir woodland, but a minute amount of *Larix* pollen was present (Whitehead 1981). From 11,600 to 8000 cal yr BP, mesic deciduous forest grew near the site, before becoming increasingly more hydric (Whitehead 1981). By 5900 cal yr BP, swamp forest taxa were dominant (Whitehead 1981). At a paleomeander along the Little River near Fort Bragg, North Carolina pine-oak forest with some hickory was the dominant vegetation between 10,400 and 10,100 cal yr BP (Goman and Leigh 2004). The site was part of a swamp forest between 10,100 and 6100 cal yr BP, before transitioning back to a pine-oak woodland (Goman and Leigh 2004). Coniferous forest occupied the area around White Pond, South Carolina, from approximately 23,500 to 14,500 cal yr BP (Watts 1980). Mesic deciduous forest vegetation dominated the pollen spectra between approximately 14,500 and 10,600 cal yr BP (Watts 1980). The site transitioned to a swamp forest after 10,600 cal yr BP (Watts 1980).

3.3.2 Climate History

While broad patterns of late Quaternary vegetation change have been documented in the southeastern U.S. (Delcourt and Delcourt 1985, 1987), high-resolution records of climate and fire history are few. Since the last synthesis of the literature (Jackson et al. 2000), the use of improved techniques and new proxies has generated more detailed data, leading to a more detailed understanding of the past environment. Environmental reconstructions for the southeastern United States and surrounding regions reveal significant variability in climate, as

well as in vegetation and fire over the late Pleistocene and Holocene. Change rather than stability has been the most consistent theme.

Proxy climate records for the majority of sites in the Southeast show a general warming trend from the Last Glacial Maximum (LGM) to the mid-Holocene, followed by a general cooling trend to the present; however, regional variability is evident. Lake level data assembled by Harrison (1989) indicate significant variability in Holocene lake levels for eastern North America. While late glacial conditions were generally wetter than present, the early Holocene brought a warming and drying trend with a peak in aridity at 6000 ^{14}C yr BP (approximately 6800 cal yr BP) (Harrison 1989). By 2000 ^{14}C yr BP (~ 2000 cal yr BP), lake levels were generally similar to present conditions (Harrison 1989). Harrison (1989) reported an approximately 4000 year lag between the timing of maximum summer insolation (given as 10000 ^{14}C yr BP; approximately 11,600 cal yr BP) and maximum aridity (at 6000 ^{14}C yr BP; ~ 6800 cal yr BP), and noted that early Holocene aridity first became apparent in eastern North America by 9000 ^{14}C yr BP (approximately 10,000 cal yr BP) (Harrison 1989). According to Harrison (1989), the Midwest was relatively wetter at this time; however, a drying trend, marked by the expansion of prairie, was evident between 9000 and 6000 ^{14}C yr BP (~10,000 and 6800 cal yr BP). After 6000 ^{14}C yr BP (6800 cal yr BP), the Midwest gradually returned to wetter conditions (Harrison 1989).

Webb et al. (2003) examined lake level proxies since 10,000 cal yr BP to map out trends in moisture-balance. At 10,000 cal yr BP, they showed drier conditions for the majority of eastern North America, with wetter conditions immediately south and west of the Great Lakes and in the northern Midwest. By 9000 cal yr BP, a band of wetter conditions extended diagonally from the upper Midwest to the coast of South Carolina, with dry conditions extending

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from coastal North Carolina to New England and west to the eastern Great Lakes (Webb et al. 2003). Webb et al. (2003) showed drier conditions in the upper Midwest by 8000 cal yr BP. Along the east coast, by this time, only southern New England was dry (Webb et al. 2003). By 7000 cal yr BP, wet conditions prevailed through the Southeast (Webb et al. 2003).

Terrestrial and aquatic remains in pond sediments from the Southeast also reveal shifts in moisture balance, providing support for the findings of Harrison (1989) and Webb et al. (2003). Based on pollen and sponge spicule analyses from Jackson Pond, Kentucky, Wilkins et al. (1991) inferred moist conditions from 10,000 to 7300 ^{14}C yr BP (~11,600 to 8200 cal yr BP). This was followed by a general drying trend from 7300 to 3900 ^{14}C yr BP (8200 to 4500 cal yr BP), followed by a gradual transition to the modern temperature and moisture regime found today. Delcourt (1979) reported progressive warming at Anderson Pond, Tennessee, beginning between 16,300 and 12,500 ^{14}C yr BP (~19,500 and 15,000 cal yr BP). Sediment influx data suggested that the Pleistocene-Holocene transition was a time of severe landscape instability (Delcourt 1979). From 12,000 to 10,000 ^{14}C yr BP (14,000 to 11,600 cal yr BP), cool mesic conditions prevailed, followed by warmer and drier conditions from 8000 to 5000 ^{14}C yr BP, approximately 8900 to 5900 cal yr BP. An essentially modern climate regime was in place by approximately 2000 ^{14}C yr BP, approximately 2000 cal yr BP. However, Tolliver (1998) studied diatoms in the Anderson Pond sediment profile, and found that assemblages revealed cool, wet conditions from 19,000 to 12,750 ^{14}C yr BP (23,000 to 15,000 cal yr BP). Water levels were lower from 12,500 to 10,000 ^{14}C yr BP (14,700 to 11,600 cal yr BP), followed by a period of relatively stable water levels until 2250 ^{14}C yr BP (~2250 cal yr BP), when seasonal precipitation increased. These finding appear to contradict interpretations from the pollen and macrofossil

record. Decreased water levels and potential seasonal drying have characterized the past two centuries at Anderson Pond, according to Tolliver (1998).

Sites in Missouri show similar moisture balance trends. Pollen analysis of Old Field Swamp in southeastern Missouri revealed vegetation changes interpreted as reflecting decreased groundwater availability from about 8700 to 5000 ^{14}C yr BP (9700 to 5900 cal yr BP), with lowest levels lasting until 6500 ^{14}C yr BP (~ 7400 cal yr BP) (King and Allen 1977). After 5000 ^{14}C yr BP (5900 cal yr BP), pollen frequencies showed an increase in arboreal taxa (King and Allen 1977). Royall et al. (1991) analyzed pollen and sediment data from Powers Fort Swale, also in southeastern Missouri. He noted plant assemblages indicative of warming and drying from 9500 to 4500 ^{14}C yr BP (10,800 to 5200 cal yr BP), followed by a switch to modern species composition.

Climate trends inferred at Jackson Pond and Anderson Pond compare favorably, allowing for a time lag due to the difference in latitude between sites. Anderson Pond showed cool, more mesic conditions from 12,000 to 10,000 ^{14}C yr BP (~ 14,000 to 11,600 cal yr BP). Jackson Pond showed these same conditions from 10,000 to 7300 ^{14}C yr BP (~ 11600 to 8200 cal yr BP). The warming and drying trends seen at Old Field Swamp and Powers Fort Swale are also similar to those at Anderson Pond. All three sites show a similar warm and dry interval from the early to middle Holocene, preceding a change to modern conditions.

Pollen data from Goshen Springs (Delcourt 1980) and Cahaba Pond in Alabama (Delcourt et al. 1983) indicated increasing moisture availability by 8500 ^{14}C yr BP and 8400 ^{14}C yr BP, respectively (approximately 9500 and 9400 cal yr BP). A close to modern climate regime had established at Goshen Springs by around 4700 ^{14}C yr BP (5400 cal yr BP), and at Cahaba Pond by around 8400 ^{14}C yr BP (9300 cal yr BP) (Delcourt 1980; Delcourt et al. 1983). In

contrast, an oxbow lake near the Tombigbee River, eastern Mississippi, showed evidence of aridity, in the form of lowered water levels, by 7300 ^{14}C yr BP (~8000 cal yr BP) (Whitehead and Sheehan 1985). According to Whitehead and Sheehan (1985), another drop in water levels occurred around 3500 ^{14}C yr BP (3700 cal yr BP); however, from 2400 to 500 ^{14}C yr BP (approximately 2400 to 500 cal yr BP) , more mesic taxa indicated a moister environment at this site.

Shifting moisture levels indicated by the initiation of drier conditions in Missouri, Kentucky, Tennessee, and Mississippi between approximately 9000 and 4500 cal yr BP concurrent with the initiation of moister conditions in Alabama by at least 9400 cal yr BP, could suggest a sharp west to east moisture gradient. The investigation of additional sites in Tennessee and Kentucky could help determine whether a longitudinal moisture gradient exists. This shifting of moisture balance agrees with the findings of Webb et al. (2003).

Watts (1971), however, examined a series of lakes in south Georgia and central Florida and found a dominance of sclerophyllous taxa from 8500 to 5000 ^{14}C yr BP (9500 to 5900 cal yr BP). This dry period is in line with drying found at the sites in Missouri, Kentucky, Tennessee and Mississippi. After 5000 BP (5900 cal yr BP), a more mesic association appeared, with pine as the dominant species. Whitehead (1981) produced a 30,000 year pollen record from Rockyhock Bay, a Carolina Bay in northeastern North Carolina that appears consistent with findings of Watts (1971). From 21,000 to 10,000 ^{14}C yr BP (25,500 to 11,600 cal yr BP), a colder, drier, more continental climate prevailed (Whitehead 1981). At 10,000 ^{14}C yr BP (11,600 cal yr BP), the local climate began to ameliorate, allowing for the influx of more thermophilous taxa (Whitehead 1981). From 7200 to 5000 ^{14}C yr BP (8100 to 5900 cal yr BP), the water table dropped, transforming the area into a swamp and marking the change to a more

modern climate (Whitehead 1981). This drop in water table could represent the drying trend seen at the other sites, indicating that dry conditions obtained throughout the southeastern United States during the early to mid-Holocene.

Sedimentary evidence from a paleochannel of the Little River on the North Carolina coastal plain appears to contradict the idea of a drier Southeast. Goman and Leigh (2004) found evidence for an extremely wet early to middle Holocene. From ca. 9000 to 6100 cal yr BP, at least 15 large flood events were recorded, in contrast to a total of six found since 6100 cal yr BP. Though this inordinate wetness seems anomalous, other research on the coastal plain of Georgia supports these results. LaMoreaux et al. (2009) found wet early to mid-Holocene conditions at a site south of Macon, with evidence for increased flood incidence. Frazier and Brown (2010) documented a wetter mid-Holocene in their study of sediments and pollen from a meander scar in west central Georgia.

The records of vegetation and climate change produced by Watts (1971) from lakes in south Georgia and central Florida seem consistent with changes later revealed at Cahaba Pond and Goshen Springs. The sequence developed by Whitehead (1981) matches records from Georgia and Florida summarized by Watts and those from Alabama and Missouri. These similarities reveal that at least this portion of the coastal plain has shared a common climate regime since the early Holocene. Work at additional coastal plain sites could help refine understanding of this and other spatial patterns.

The interpretation of a wet early to middle Holocene by Goman and Leigh (2004) and LaMoreaux et al. (2009) seem anomalous, contrasting with the findings of Watts (1971) and Whitehead (1981). Diffenbaugh et al. (2006), however, provide some support for a wetter Atlantic seaboard. Using a nested climate modeling system, they found that middle Holocene

solar forcing resulted in drier conditions in the central U.S. and wetter than present conditions on the Atlantic seaboard (Diffenbaugh et al. 2006). This apparent inconsistency could indicate that the Atlantic seaboard saw an increase in storm activity during this period. Based on an analysis of flood deposits from Brevard, North Carolina, Seramur and Cowan (2010) postulated increased summer tropical storm activity during the middle Holocene, lending some support to this possibility. Alternatively, increased solar radiation could have caused convective storms along the eastern seaboard, as found today in Florida and other parts of the coastal plain. Testing these ideas will require further investigation of coastal plain sites.

The climatic shifts that took place between the late Pleistocene and late Holocene were not always gradual or unidirectional. Depending on the proxy examined, numerous cyclical climatic events are evident (Li et al. 2007). Both Anderson et al. (2007) and Alley et al. (1997) noted numerous, often rapid, climatic shifts over the past 10,000 years. These climatic perturbations caused significant environmental changes in many areas, contradicting the notion of a stable Holocene.

In ice core data from Greenland, Alley et al. (1997) found evidence of a major cooling event at approximately 8200 cal yr BP. This excursion during the general warming trend of the early Holocene was interpreted as the result of a meltwater pulse into the North Atlantic, causing a decrease in deep water formation (Alley and Agustsdottir 2005). This change in the thermohaline circulation, though of short duration, caused modifications to storm tracks and changes in precipitation patterns, resulting in colder, drier, and windier conditions in eastern North America (Alley and Agustsdottir 2005). Reconstructions of sea surface temperatures and salinity for the North Atlantic based on Mg/Ca and $\delta^{18}O$ data from planktonic foraminifera reveal temperature decreases at 9300, 8200, 3700, 1800, and 900 cal yr BP (Came et al. 2007).

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Fleitmann et al. (2008) reported a short-lived climatic anomaly at 9200 cal yr BP, which they attributed to a meltwater pulse. While this anomaly shows up in multiple high-resolution proxy records across the northern hemisphere, the authors pointed out that the short duration of this episode (as well as that of the 8200 cal yr BP event) may leave few traces in lower resolution records (Fleitmann et al. 2008). Slow sediment accumulation rates and limited radiocarbon dates in many previous paleoenvironmental investigations in the Southeast make identifying short-term oscillations in regional records difficult (Kneller and Peteet 1999). Pollen evidence for the cold reversal at 8200 cal yr BP does appear, however, at Brown's Pond, Virginia (Kneller and Peteet 1999). A resurgence in *Picea*, *Abies*, and *Tsuga* pollen at this time provides evidence for a return to cool yet moist conditions at the site (Kneller and Peteet 1999).

Further evidence for the 8200 cal yr BP event comes from a Newfoundland peat core. Daley et al. (2009) noted declines in δ^{18}O of precipitation at 8450 cal yr BP and between 8250 and 8170 cal yr BP based on isotopic analysis of cellulose from *Sphagnum* moss. Viau et al. (2002) analyzed radiocarbon dates and pollen diagrams in the North American Pollen Database and identified nine vegetation shifts that they attributed to climate change. Climatic oscillations at 600, 1650, 2850, 4030, 6700, 8100, 10,190, 12,900, and 13,800 cal yr BP caused relatively rapid and synchronous changes in vegetation throughout North America (Viau et al. 2002). Carbon isotope reconstructions from speleothems in the Ozark Highlands of Missouri and Arkansas showed a decrease in δ^{13}C values between 9500 and 8200 cal yr BP, and again between 4500 and 3000 cal yr BP, indicating cooler and moister conditions separated by a warmer and drier period starting at 7500 cal yr BP (Denniston et al. 2000).

As the above examples show, a millennial scale cycle of alternating warm and cold periods has existed since at least late Pleistocene times. The Medieval Warm Period (ca. A.D. 750–

1300) and the Little Ice Age (ca. A.D. 1350–1850) could be seen as continuations of this pattern of change. Currently, few southeastern sites yield unequivocal evidence of these millennial-scale oscillations; however, this may be a product of the age and coarse temporal scale of most southeastern paleoenvironmental records. The careful selection of sites, combined with improved techniques, should allow for finer-grained analyses similar to that seen in studies by Watts et al. (1992), Watts and Hansen (1994), and Grimm et al. (2006). To this end, previously analyzed sites should be revisited to refine their records.

While data from the North Atlantic and northern North America show several cold excursions since the last Glacial, evidence from more southerly locations seems contradictory. Pollen and plant macrofossil data from Lake Tulane, Florida, revealed an inverse relationship between conditions in northern North America and the southeastern United States. Grimm et al. (2006) found that warm and wet conditions existed in Florida during cold episodes in the North Atlantic. They postulated that a reduction in thermohaline circulation during Heinrich events led to a reduction in the northward movement of heat, leading to heat retention in the subtropical Atlantic and Gulf of Mexico, and to a shift from scrub-oak and prairie to pine forest (Grimm et al. 2006). Donders et al. (2009) used a pollen-climate interface model to show that mean summer precipitation and mean November temperature increased during periods of pine dominance. They posited that a positive heat anomaly in the Gulf of Mexico and western Atlantic explained the increase in moisture during this time (Donders et al. 2009). Further evidence for regional warming comes from magnetic susceptibility readings of sediments from Hall's Cave, Texas, which revealed warmer and wetter conditions at 17,000, 8200, and 4300 cal. yr BP (Ellwood and Gose 2006). These apparent contradictions with more northerly data reveal the variability in the North American climate record. An increase in sea surface temperatures in

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the Gulf of Mexico and western Atlantic would serve to buffer the area against cold incursions from further north. Additionally, warmer ocean temperatures should lead to increased evaporation, and, consequently, increased precipitation.

Davis (1983), referencing CLIMAP data, stated that the Gulf Stream was displaced from its normal route along the Atlantic coast during the last glacial, and instead ran eastward across the mid-Atlantic from approximately the North Carolina coast. This southward shift in the flow, relative to post-glacial times, shifted wind directions, leading to changes in storm tracks and precipitation patterns. Carver and Brook (1989) analyzed the orientation of late Pleistocene parabolic dunes as a proxy for paleowind direction. They found evidence of predominantly westerly winds in Georgia, with wind direction shifting to a more northwestern orientation in North Carolina. During this same time, eolian deposits in Maryland and the Chesapeake Bay revealed strong northwesterlies and northerlies (Markewich et al. 2009).

The path of the winds marks the location of the Gulf Stream off the North Carolina coast. The Gulf Stream would mark the boundary between the cold North Atlantic and the warm South Atlantic. Russell et al. (2009), citing work on noble gas concentrations in groundwater by Clark et al. (1997), showed a difference in mean annual temperatures of approximately 5° C between Georgia and Maryland during the Last Glacial Maximum. LaMoreaux et al. (2009) noted that the enhanced summer insolation values of the early to middle Holocene, when combined with the location of the Laurentide Ice Sheet, resulted in a large temperature gradient over a relatively short distance. This gradient would suggest a more northerly position for the subtropical jet than seen at the present, resulting in increased storm activity over parts of the Southeast, and therefore increased precipitation (LaMoreaux et al. 2009). Whether the climatic oscillations dramatically

altered the location of the Gulf Stream and subtropical jets requires further investigation at coastal plain sites.

3.3.3 Southeastern Refugia

Delcourt and Delcourt (1996) provided a brief overview of the interpretations of Deevey (1949), whose reconstruction of ice age vegetation in the eastern U. S. included a wide expanse of boreal vegetation south of the ice sheet. Warm temperate species existed between this boreal forest and the Gulf of Mexico (Deevey 1949). Locations within this expanse served as refugia, providing thermophilous species a place to wait out the glacial cold (Deevey 1949). This view came to be regarded as fact, influencing a variety of later climate and vegetation models (Loehle and Iltis 1997). Deevey's interpretation conflicted with that of Braun (1950), who envisioned a relatively compressed zone of tundra and boreal forest south of the ice. Loehle and Iltis (1997) examined geologic, paleobotanical, and biogeographic evidence of ice age refugia. They concluded that the difficulties in differentiating species of *Pinus* from pollen and needle evidence, and the existence of pollen from species such as *Quercus* and *Carya* in association with *Abies* and *Picea,* argued against boreal forest conditions in the southeastern United States. They also found no support for the idea of refugia for deciduous forests, in the available paleoecological data.

3.3.4 Carbon Dioxide and Vegetation

Harrison and Prentice (2003) found that glacial to interglacial changes in vegetation depended on carbon dioxide levels as well as climate. Furthermore, Bennett and Willis (2000) showed that while changes in temperature were the predominant factor in glacial to interglacial changes in vegetation at middle to high latitudes (according to circulation models), a discrepancy existed at lower latitudes that is easily explained by carbon dioxide concentrations (Bennett and

Willis 2000). Additionally, Loehle (2007) found that reduced CO_2 levels also accounted for the temperature disparity seen in many vegetation models. Reduced carbon dioxide levels induce greater stomatal opening, privileging C_4 plants over C_3 plants, because of the increased potential for water loss. Plants that require less water to survive would have an advantage. Lowered CO_2 levels alter the dominance of taxa (Loehle 2007), as well as the freezing point of the leaves of some woody shrubs (Beerling et al. 2002), and are consequently one potential explanation for plant assemblages that have no modern analog. The coupling of temperature and carbon dioxide levels may help explain the strange associations of boreal taxa with oaks, hickories, and more mesic taxa that are apparent in late Pleistocene and early Holocene pollen diagrams from the Southeast, but infrequently discussed.

3.3.5 Charcoal Based Fire History Research in the Southeast

The vast majority of fire-related research in North America comes from boreal and western settings. However, a few analyses from the Southeastern U.S. exist. Kneller and Peteet (1999) estimated the volume concentration (mm^3/cm^3) of charcoal particles >500 µm in size in the sediments of Browns Pond, Virginia. Between 14,180 and 12,730 ^{14}C BP (17,000–15,000 cal yr BP) charcoal concentrations reveal relatively moderate fire activity. For this interval, which predates the Whiteoak Bottoms record, they inferred a montane spruce-fir forest with more mesophytic vegetation immediately surrounding the pond.

From 12,730 to 12,260 ^{14}C BP (15,000–14,320 cal yr BP) charcoal concentrations at Browns Pond decreased, indicating less fire activity, as vegetation transitioned to a more mesophytic forest similar to a northern hardwoods forest (Kneller and Peteet 1999). They attributed this change to warming temperatures, but noted that conditions were still cooler and

moister than present. Given the timeframe, the warming trend that they inferred could be the initial warming of the early Bølling interstadial.

After a peak in temperature at approximately 14,490 cal yr BP, conditions cooled. A decrease in deciduous tree taxa and increases in *Abies* and *Picea* pollen marks a reversion to cooler temperatures from 12,260 to 12,200[14]C (14,320–14,240 cal yr BP) (Kneller and Peteet 1999). During this period, charcoal concentrations remained similar to, but slightly higher than the previous period (Kneller and Peteet 1999).

Between 14,240 and 13,260 cal yr BP, temperatures increased at Browns Pond and the pollen diagram records an increase in *Tsuga* pollen (Kneller and Peteet 1999). Charcoal concentrations are similar to values for the period 15,000 to 14,320 cal yr BP (Kneller and Peteet 1999). This period spans the Bølling/Allerød; evidence of the Older Dryas appears not to be recorded at Browns Pond.

At Browns Pond from 13,260 to approximately 10,940 cal yr BP, *Tsuga* dominated the pollen assemblage, birch pollen increased, and *Sphagnum* reached its maximum values (Kneller and Peteet 1999). Spruce, oak, sedge, and Asteraceae pollen decreased, and charcoal concentrations increased, ranging from 0 to 3.6 mm^3/cm^3 (Kneller and Peteet 1999). Kneller and Peteet (1999) interpreted these changes as indicating a temperature increase, and found concomitant reductions in *Nuphar* and Nymphaeaceae pollen that they interpreted as evidence of a drop in water depth at the site. This period includes the Allerød, Inter-Allerød Cold Period, and the Younger Dryas. Peteet (2000), however, interpreted the increase in hemlock as indicating warmer and wetter conditions.

From 8410 to approximately 4870 ^{14}C BP (9400 to 5600 cal yr BP), Kneller and Peteet (1999) found increasing hickory and oak pollen percentages and decreasing percentages of

hemlock. Charcoal concentrations ranged from less than 1 to 4.5 mm^3/cm^3 (Kneller and Peteet 1999). They also found pollen of swamp and other wet ground taxa such as *Ilex* and *Cephalanthus* (Kneller and Peteet 1999). They inferred a fluctuating water table at this time, with water levels rising toward the end of the period. Additionally, they found evidence, in the form of brief increases in hemlock, spruce, and fir pollen, of a brief cool, but still moist, interval centered around 7500 ^{14}C BP (about 8400 cal yr BP).

Watts (1980), reported charcoal in nearly every level of the upper four meters of his core from White Pond, South Carolina. This interval spanned from 9550 ^{14}C BP (approximately 10,500 cal yr BP) to the present, and the dominant vegetation at the time was pine-oak forest (Watts 1980).

Lynch and Clark (2002) found that past fire regimes in the southern Appalachian region were best described as variable. They reported peaks in macroscopic charcoal accumulation at Browns Pond, Virginia at 17,600 to 16,800 cal yr BP, 16000 cal yr BP, 12,800 to 11,200 cal yr BP, and 10,400 to 8800 cal yr BP (Lynch and Clark 2002). The period from 10,400 to 8800 cal yr BP showed the highest charcoal accumulation values. Low accumulation characterized the period from 8000 cal yr BP to the present (Lynch and Clark 2002). At Pine Swamp, Maryland, Lynch and Clark (2002) found peaks in charcoal accumulation at 12,000 to 11,200 cal yr BP, 10,400 cal yr BP, and from 4800 cal yr BP to the present. The period from 10,400 to 4800 cal yr BP marked a time of low charcoal accumulation (Lynch and Clark 2002). Peaks in accumulation occurred at Pink Beds Bog, NC at 10,400 cal yr BP, 8800 to 8000 cal yr BP, 8000 cal yr BP, 7200 cal yr BP, 7200 to 6400 cal yr BP, 6400 to 5600 cal yr BP, 5600 to 3200 cal yr BP, 800 cal yr BP, and 0 cal yr BP (Lynch and Clark 2002). Low charcoal accumulation occurred between 3200 and 1600 cal yr BP.

Fesenmyer and Christensen (2010) examined soil charcoal to investigate the Holocene fire history of a 10 hectare section of the Wine Spring Creek Ecosystem Management Area, in the Nantahala National Forest, Macon County, North Carolina. They found that fires occurred on a regular basis across their 4000 year dataset. These fires were not confined to dry oak-pine stands, but extended downslope into more mesic areas. They noticed a distinct increase in fires approximately 1000 years ago that they attributed to Native Americans.

3.4 Previous Research at Whiteoak Bottoms

In an investigation of the geomorphic and stratigraphic history of Whiteoak Bottoms, McDonald (2010) and McDonald and Leigh (2011) found that the wetland began forming in a paleochannel of the Nantahala River between 15,000 and 14,000 cal yr BP. Peat accumulated between 15,000 and 9000 cal yr BP, forming the lower of two peat facies. Between 14,500 and 14,000 cal yr BP, woody peat transitioned to mossy peat, which the researchers interpreted to result from an increase in the height of the water table. Whether this increase was caused by an increase in precipitation or some other cause was unclear. Within the lower peat layer, they noted several sandy layers which they interpreted as evidence of overbank flooding. Between 13,400 and 9000 cal yr BP, the wetland sediments became increasingly more inorganic as the lower peat layer transitioned into a grey silty layer. McDonald and Leigh (2011) attributed this change in sedimentation to an increase in high magnitude flooding or an increase in hillslope erosion. A second, upper, peat layer overlies the grey silt. This peat layer represents the modern wetland surface.

Additional prior research at Whiteoak Bottoms included rudimentary pollen analysis by M. C. Sheehan in the early 1980s (Figure 3). Found within a set of photographic slides from the work of Drs. Paul and Hazel Delcourt that are held in the University of Tennessee's Frank H.
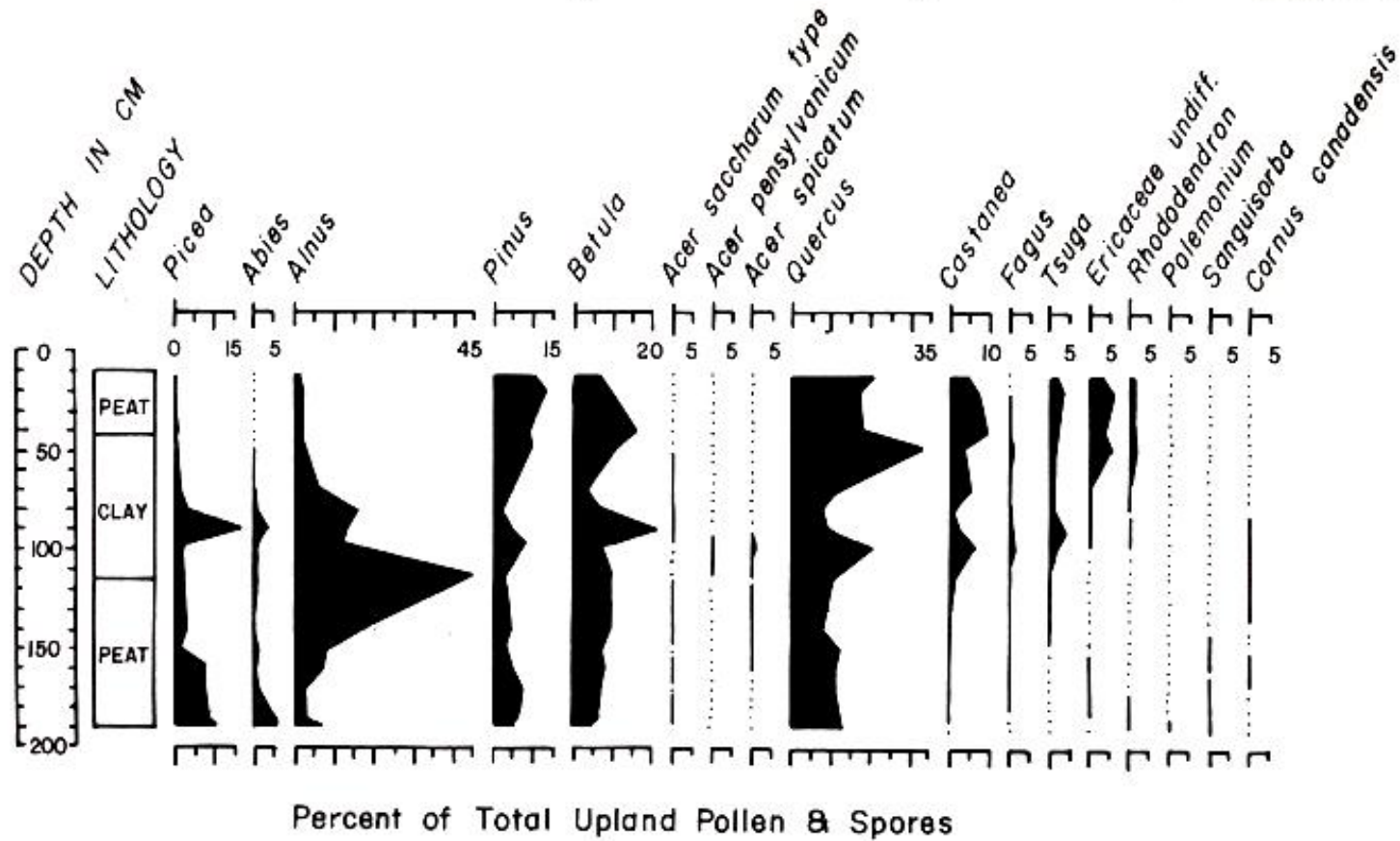
Figure 3. Unpublished pollen diagram from early work in the vicinity of Whiteoak Bottoms

McClung museum, Sheehan's pollen diagram includes at least 12 levels, but there is no chronological control for this sequence. Other slides found in the set are labeled Standing Indian Bog, or Whiteoak Bottoms (Standing Indian Bog), providing evidence that the wetland at Whiteoak Bottoms and Standing Indian Bog could be the same site. Though the name Whiteoak Bottoms is used here to designate the wetland next to Standing Indian Campground, the place name actually refers to the valley bottom in the vicinity of the wetland, as used on topographic maps of the area. Sally Horn (pers. communication) contacted Dr. Sheehan in September, 2011, and later his former advisor, Dr. Donald Whitehead. Both confirmed that they had never published the study and that no additional data exist. Dr. Whitehead, under whose employ Dr. Sheehan had undertaken the study, granted permission for us to use the diagram in whatever way might help the current research (S. Horn, pers. communication).

CHAPTER IV
METHODS

As part of his M.S. thesis on the geomorphological history of Whiteoak Bottoms, J. McDonald recovered a series of sediment cores in 2009 using a vibracorer and 3" (7.6 cm) diameter aluminum core tubes (McDonald 2010, McDonald and Leigh 2011). Two of these cores (WOB1A and WOB2A) were transferred from the University of Georgia to the University of Tennessee's Laboratory for Paleoenvironmental Research to perform pollen and charcoal analysis. Prior to transport to the University of Tennessee, the aluminum core tubes were cut along their longitudinal axis using a circular saw with a metal-cutting blade. The sediments were then sliced along the same axis, yielding two sets of core halves. One set, consisting of a half core from WOB1A and a half core from WOB2a, was archived at the University of Georgia; the second set was wrapped in plastic wrap and aluminum foil and transported to the University of Tennessee.

Upon arrival at the University of Tennessee, the 157-cm long WOB2A core was chosen for study. This core was photographed and described for Munsell color, and logged for stratigraphy and texture. X-radiographs were taken at the University of Tennessee Veterinary Medical Center. The WOB2A core was then sampled for loss-on-ignition, pollen, and charcoal. All samples were taken at the same time to reduce the exposure of the core to contamination and drying. Seeds, charcoal, and other materials were removed from the core for AMS ^{14}C dating prior to and during sampling.

4.1 Radiocarbon Dating

Organic materials in the Whiteoak Bottoms sediment core were submitted to Beta Analytic Laboratory, Inc., the Center for Applied Isotope Studies at the University of Georgia,

and the NSF-Arizona AMS Laboratory at the University of Arizona for AMS radiocarbon analyses. Samples sent to the NSF-Arizona lab underwent acid-base-acid pretreatment per Arizona protocol at the University of Tennessee prior to shipment. Radiocarbon determinations were calibrated using the CALIB 6.1 computer program (Stuiver and Reimer 1993), and the dataset of Reimer et al. (2009). Calibrated ages for levels between dated horizons were estimated using linear interpolation and the weighted means of the 2σ probability distribution of the calibrated ages.

4.2 Loss on Ignition

I performed high resolution loss-on-ignition (LOI) analysis on 156 1-cm^3 cubes of sediment removed from the WOB2A core at contiguous 1 cm intervals. This resulted in complete coverage of the entire core. LOI analysis provides an estimate of the organic and carbonate content of the sediments. LOI analysis also makes it possible to express the pollen and charcoal data on a dry mass or dry organic mass basis, and allows for the calculation of organic and inorganic sediment influx. Following a modified version of the protocol of Dean (1974), samples were weighed, dried overnight at 100 °C, and weighed again. Samples were then ignited at 550 °C for 2 hours (D. Sullivan, pers. communication), cooled and reweighed. Finally, the samples were ignited at 1000 °C for 1 hour, cooled, and weighed once more.

4.3 Pollen and Microscopic Charcoal

For pollen and microscopic charcoal analysis, I took 0.5 cm^3 samples at 4 cm intervals for a total of 39 samples. Sample processing followed standard palynological protocols including the use of HF, HCl, KOH, and acetolysis (Berglund 1986; Appendix A). Processing included the addition of control spores, allowing for the calculation of pollen concentration and influx values (Stockmarr 1971), and sieving with a 180 µm mesh sieve to remove large debris

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from the sample. The resulting residue was mounted on microscope slides in silicone oil for analysis.

For pollen analysis, a minimum of 300 pollen grains, excluding indeterminates, fern spores, and aquatics, were tallied per level. I classified Cyperaceae as an aquatic pollen type for this analysis as a large percentage of the native *Cyperus* and *Carex* species are wetland plants. Pollen was identified to the lowest taxonomic level possible using modern pollen reference samples, pollen atlases, and keys in the Laboratory of Paleoenvironmental Research (Kapp 1969; McAndrews 1973).

Unknown pollen grains were sketch and described, and their locations on the slides were recorded for potential reexamination. Fern spores were classified by morphology (monolete versus trilete). The spores of *Huperzia* and *Sphagnum* were tallied separately. Concurrent with pollen analysis, I noted the presence of conifer stomata on the pollen slides for possible future analysis.

I quantified microscopic charcoal on the pollen slides using a modified version of the point counting method described by Clark (1982) (Appendix B), with ca. 4000–5300 points applied to each slide. For 24 of 38 samples, the relative error of the charcoal estimate, calculated as in Clark (1982), was 11–50%. Fourteen samples with low charcoal abundance had higher relative errors (overall mean relative error 50%). Using the *Lycopodium* spike on the slides, the areal estimate of charcoal on the slide (mm^2) was converted to an estimate of charcoal in the entire sample, and used to calculate charcoal area concentrations (mm²/cm³), charcoal area influx (mm²/cm²/yr), and charcoal to pollen ratios (μm^2/pollen grain).

4.4 Data Processing

LOI, pollen, and charcoal data were entered into Microsoft Excel spreadsheets to calculate percentages, concentrations, and influx values. Stratigraphic diagrams were produced using the C2 software (Juggins 2010).

CHAPTER V
RESULTS

5.1 Core Stratigraphy

The Whiteoak Bottoms core shows distinct stratigraphic changes with depth (Figure 4). From its base at 157 cm to approximately 121 cm, the sediment core grades from a dark grey sand (10YR4/1) to a dark grey (10YR 4/1) loamy sand. Mica is abundant in this 36-cm section. At 121 cm, an abrupt change occurs from inorganic to organic sediment. Additionally, thin bands of dark grey sand (10YR4/1) are centered at 150, 147, 144, and 138.5 cm, all of which show up as light colored bands on the X-rays. Between 121 and 74 cm, the core consists of very dark grey (7.5YR3/1), silty organic material with mica. Several layers with higher organic content occur between 120 and 88 cm. Pieces of *Pinus* sp. wood occur from 91–81.5 and 78–77 cm. Additionally, a bent piece of wood of *Pinus* sp. extends across the core between 87 and 81 cm. Between 75 and 74 cm, a small silty clay inclusion measuring approximately 1 cm x 1.5 cm occurs. A layer of very dark brown (10YR2/2) silty sediment extends from 74 to 62 cm depth. Overlying this layer is a very dark brown (10YR 2/2) to dark greyish brown (10YR3/2) silty peat layer. At 37 cm, the sediments change abruptly to dark grey (10YR4/1) to grey (10YR6/1) silty loam, which extends to 28 cm. Overlying this layer, from 28 to 10 cm, is a layer of brown (10YR3/1) silty clay. From 10 cm to the top of the core, the sediments consist of a mix of very dark brown (10YR2/2) muck and peat.

5.2 Core Chronology

Seven AMS radiocarbon determinations were obtained on the Whiteoak Bottoms core (Table 1). Six of seven determinations were made on macrofossils (charcoal particles, wood fragments, and seeds) and one on bulk sediments. Results indicate that the basal sediments were deposited prior to 12,000 radiocarbon years (ca. 14,400 calibrated years), in the late Pleistocene.
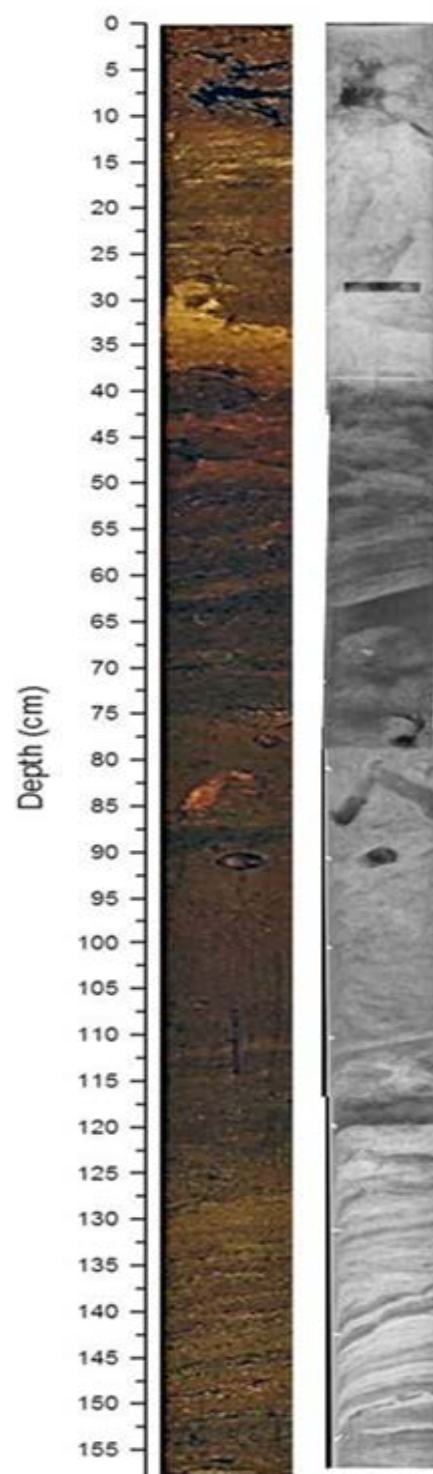
Figure 4. Photograph (left) and X-ray (right) of the Whiteoak Bottoms sediment core. Rectangular cut-out on the X-ray is the position of date β-282464.

Table 1. AMS radiocarbon determinations and calibrations for Whiteoak Bottoms.

Lab number[a]	Depth in profile (cm)	Material dated	$\delta^{13}C$ (‰)	Uncalibrated ^{14}C age (^{14}C yr BP)	Calibrated age ranges (2 σ)[b] (cal yr BP)	Area under probability curve	Weighted mean[c] (cal yr BP)
β-295805	14	charred material	−25.4	2200 ± 30	2324–2141	1.0	2229
β-282464	27	charred material	−25.7	1640 ± 40	1620–1413 1687–1673	0.975 0.025	1531
AA94829	34	charred material	−26.9	7926 ± 39	8616–8610 8814–8628 8874–8824 8979–8876	0.007 0.627 0.130 0.236	8782
UGAMS-9222	40	organic sediment	−29.1	10640± 30	12,655–12,544	1.0	12,597
β-295806	54	wood	−27.4	11,830 ± 50	13,820–13,475	1.0	13,670
UGAMS-07181	90	wood	−27.8	12,170 ± 40	14,173–13,860	1.0	14,020
AA94830	136	*Sparganium* sp. seeds	−28.6	12,331 ± 62	14,890–14,015	1.0	14,377

[a]Analyses were performed by Beta Analytic Laboratory in Miami, Florida; the Center for Applied Isotope Studies at the University of Georgia in Athens, Georgia; and the NSF-Arizona AMS Laboratory at the University of Arizona in Tucson, Arizona.
[b]Calibrations were calculated using CALIB 6.0.2 (Stuiver and Reimer, 1993) and the dataset of Reimer et al. (2009).
[c]Weighted mean of the 2 σ calibrated radiocarbon probability distribution.

Overall, the dates are in chronological order, except for the date of 2200 ± 30 ^{14}C yr BP, which overlies a date of 1640 ± 40 ^{14}C yr BP. This date was excluded from the age model (Figure 5) because changes in sediment type, sediment accumulation, organic matter content, and charcoal indicate that the upper 27 cm of the core may be the result of disturbance, and therefore the date is from an insecure context.

The age-depth model (Figure 5) revealed that the sedimentation rate for the Whiteoak Bottoms core has changed markedly over time. Between 136 and 91 cm, the sedimentation rate was 0.129 cm/yr (each centimeter representing 7.6 yr). From 90 to 55 cm, it was 0.103 cm/yr (9.7 yr/cm). Between 54 and 41 cm, the rate slows to 0.013 cm/yr (76.9 yr/cm). From 40 to 35 cm, the sedimentation rate was 0.002 cm/yr (500 yr/cm) and between 34 and 28 cm, the sedimentation rate was 0.001 cm/yr. This is the slowest sedimentation rate of the entire core and would suggest that each cm in this interval represents 1000 years of sediment accumulation. From 27 to 0 cm, the sedimentation rate was 0.017 cm/yr (58.8 yr/cm). Overall, sediment accumulation has slowed over the history of the site. Above 54 cm, the apparent sedimentation rate slows so dramatically that breaks or hiatuses in sedimentation are likely. The anomalous 2200 ± 30 ^{14}C yr BP date comes from this interval of slow sediment accumulation.

5.3 Loss on Ignition

Organic matter content as estimated by loss-on-ignition has varied over the history of the Whiteoak Bottoms wetland (Figures 6–8). From its base at 156 cm (approximately 14,532 cal yr BP), the organic content of the core shows a general rising trend to 37 cm (10,689 cal yr BP). Organic content between 156 and 120 cm (approximately 14,532 to 14,253 cal yr BP) ranges from 2 to 15%. Between 120 and 88 cm (14,253 to 14,001 cal yr BP), LOI values range from 14 to almost 36%. Organic content values range from 27 to 41% for the interval 88 to 77 cm
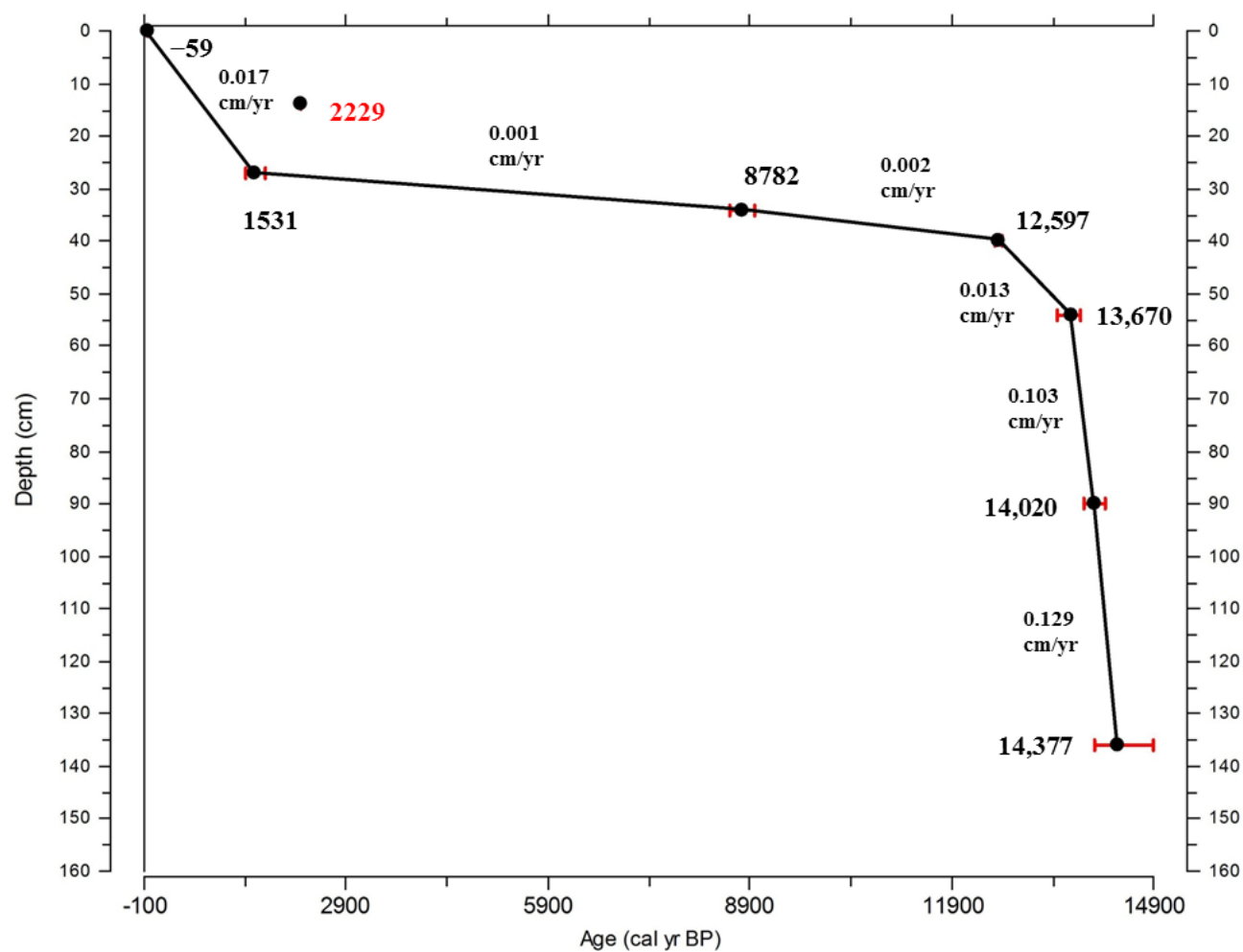
Figure 5. Age-Depth Model for Whiteoak Bottoms, North Carolina, core WOB2A. Bars mark the 2 σ calibrated age range.
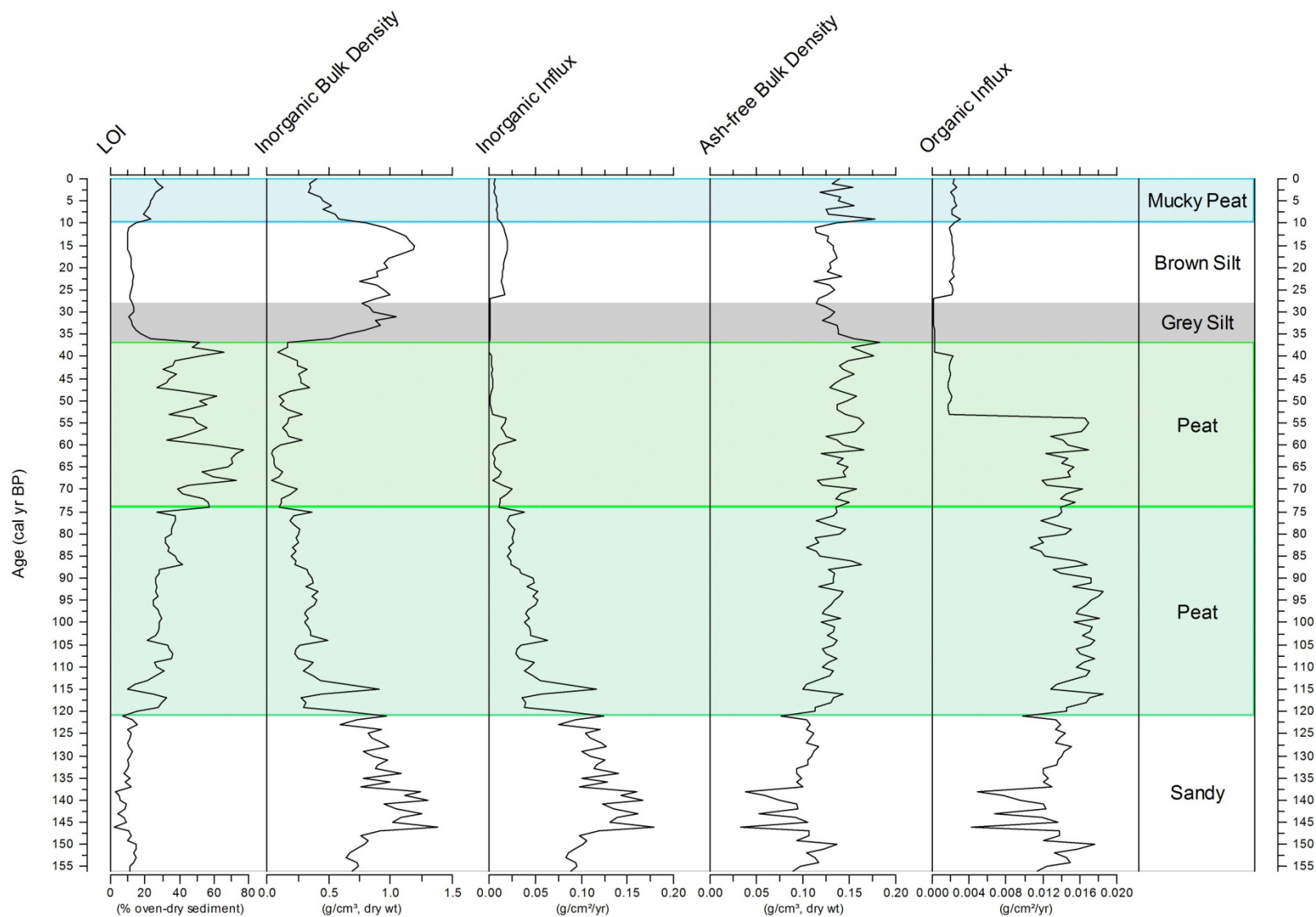
Figure 6. Loss-on-ignition, inorganic bulk density, inorganic sediment influx, ash-free bulk density, and organic sediment influx for Whiteoak Bottoms core WOB2A, graphed by depth. Right hand column shows the corresponding lithology for the core. The designation peat was assigned to sections of the WOB2A core with an organic content of 30% or more.
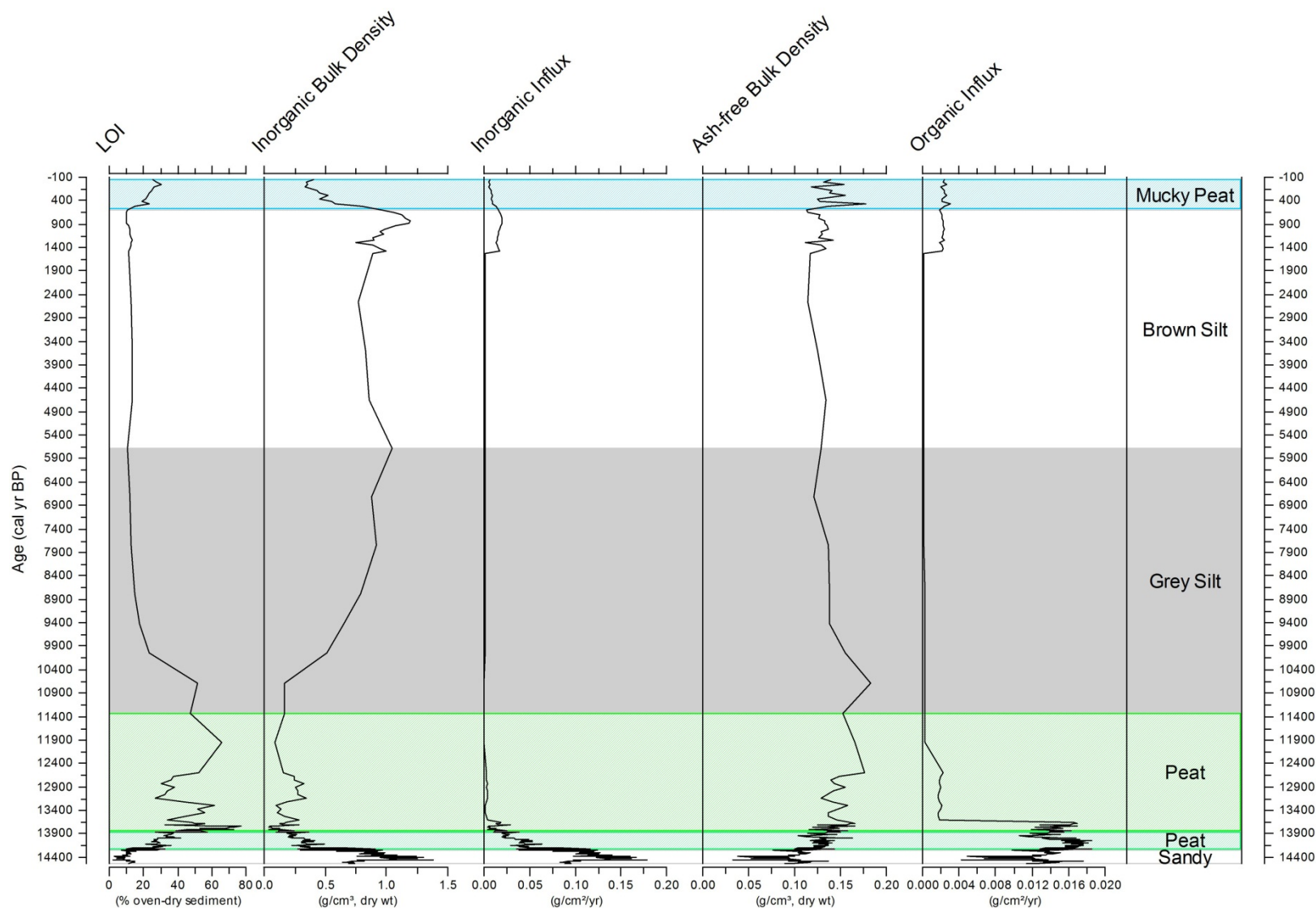
Figure 7. Loss-on-ignition, inorganic bulk density, inorganic sediment influx, ash-free bulk density, and organic sediment influx for Whiteoak Bottoms core WOB2A, graphed by interpolated age. Right hand column lists the corresponding lithology for the core. The designation peat was assigned to sections of the core with an organic content of 30% or more.
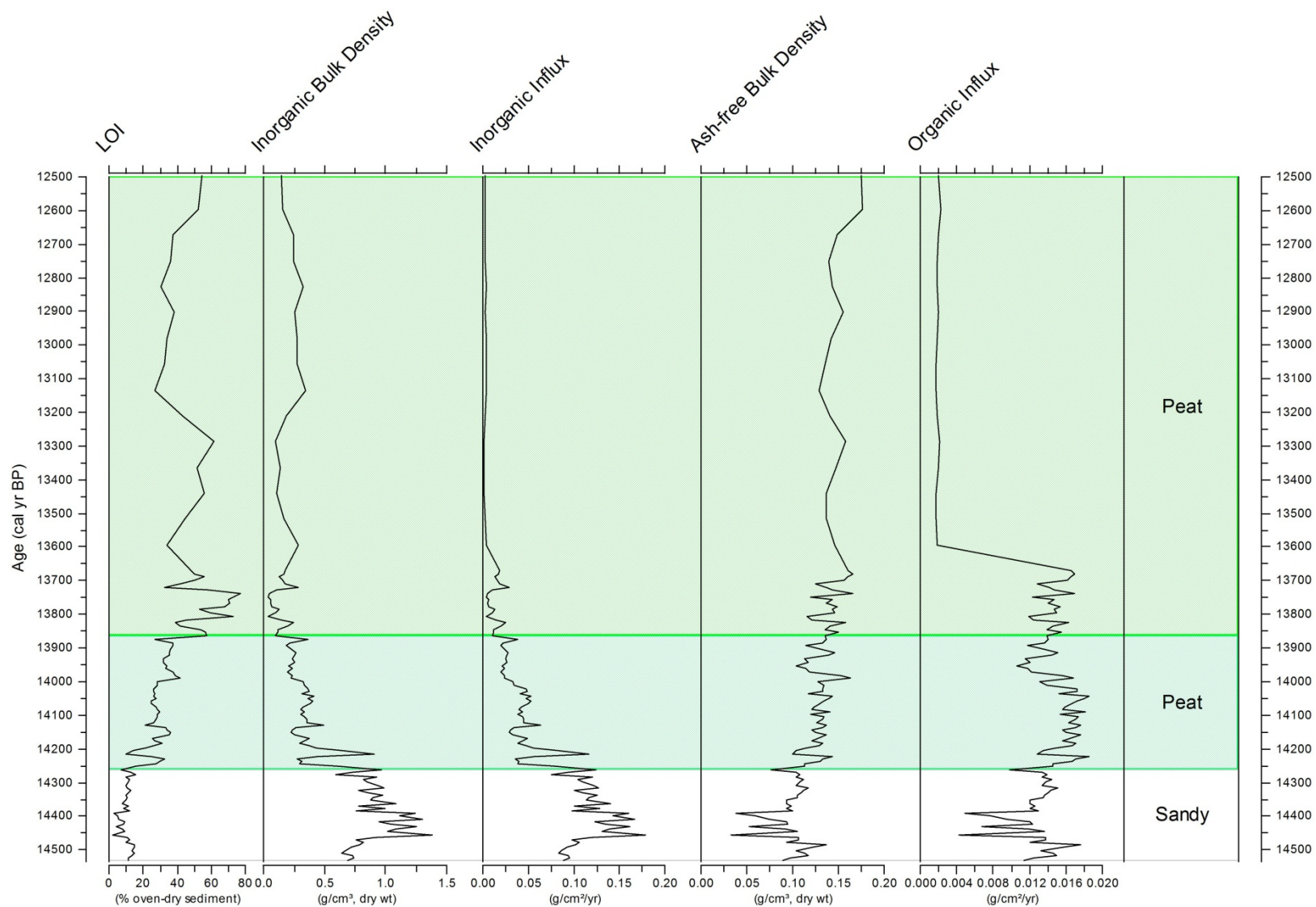
Figure 8. Loss-on-ignition, inorganic bulk density, inorganic sediment influx, ash-free bulk density, and organic sediment influx for the Pleistocene section of Whiteoak Bottoms core WOB2A, graphed by interpolated age. Data are for the period 14,532–12,500 cal yr BP. Right hand column lists the corresponding lithology for the core. The designation peat was assigned to sections of the core with an organic content of 30% or more.

(14,001 to 13, 894 cal yr BP). The highest organic values occur between approximately 13,894

and 10,690 cal yr BP, within the 77 to 37 cm interval. For this interval, values range from 27 to

over 77%. Between approximately 37 and 35 cm (10,690 to 9418 cal yr BP), the organic content

of the core drops dramatically, reaching values similar to those seen at the base of the core.

From 36 to 10 cm (10,054 to 530 cal yr BP) organic content drops to between 23 and 9%. From

10 cm to the surface (−59 cal yr BP), organic matter content rises again, reaching values

consistently above 18%.

5.4 Pollen

Pollen is moderately well preserved through the length of the core (Figure 9). Because of

the possibility of a disturbed signal in the upper part of the core, I focused the pollen analysis on

the portion of the core between 156 and 40 cm (14,532 and 12,597 cal yr BP). Based on the

radiocarbon dates, this portion of the core represents the late Pleistocene through early Younger

Dryas. Between 56 and 27 cm (13,690 and 1531 cal yr BP), I examined 8 samples for pollen

analysis. This resulted in a 4 cm sampling interval for this portion of the core; however, pollen

data above 36 cm (10,054 cal yr BP) are not shown in graphs. Between 120 and 56 cm (14,253

and 13,690 cal yr BP), I analyzed 4 additional levels, resulting in 16 cm intervals between

samples. Additionally, I analyzed pollen from 156 cm (14,532 cal yr BP).

I identified 34 different pollen types. This total does not include types classified as

unknown. Pollen grains classified as indeterminate were tallied but are not included in the pollen

sums. The percentage of indeterminates ranged from 15 to 45 percent, with the highest values

occurring in the two levels between 8782 and 1531 cal yr BP (34–27 cm).
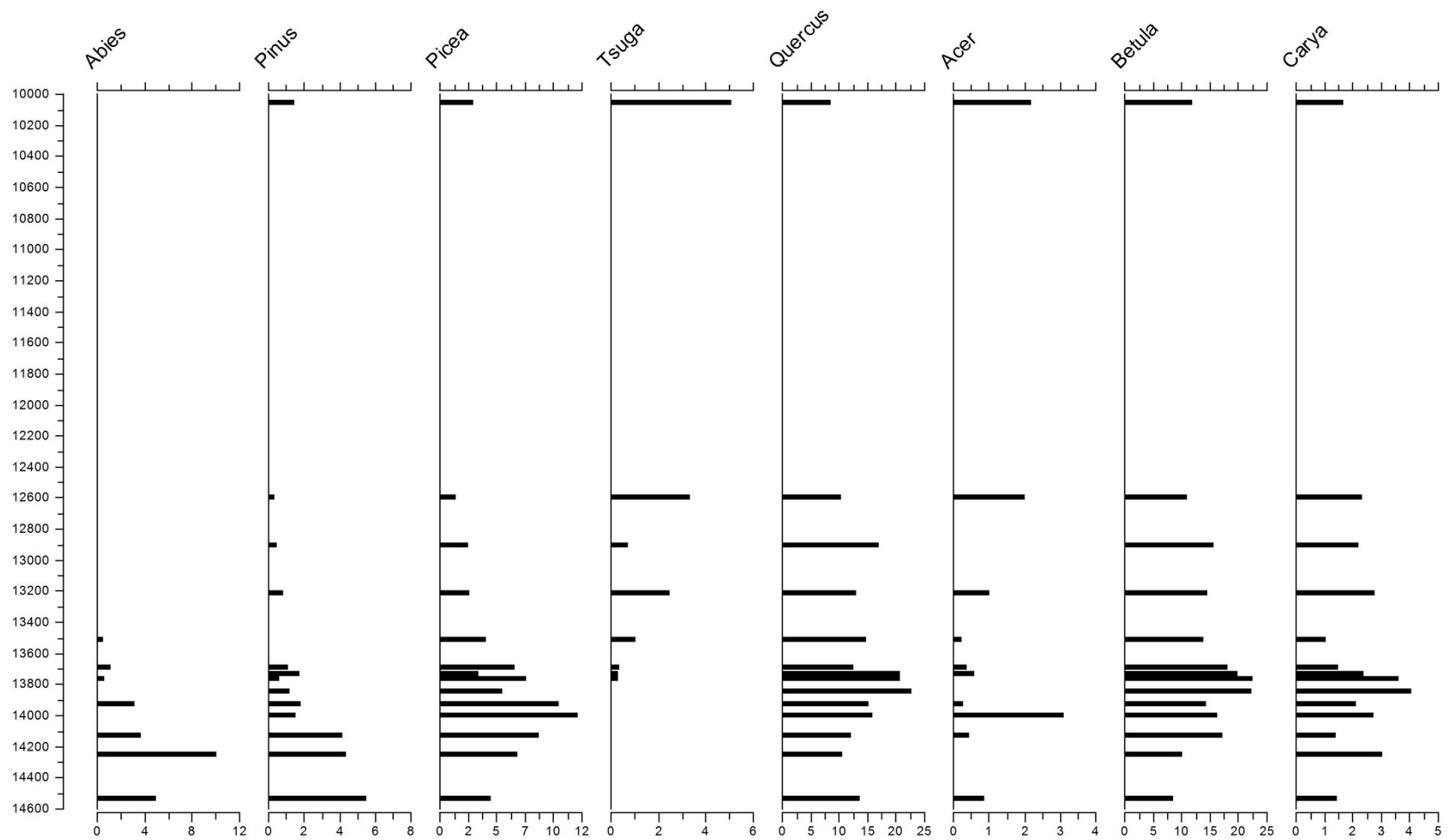
Figure 9. Pollen diagram for Whiteoak Bottoms for the period 14,532 to −59 cal yr BP. Pollen types other than Cyperaceae are plotted as a percentage of total pollen excluding Cyperaceae. Cyperaceae percentages are based on total pollen including this taxon.
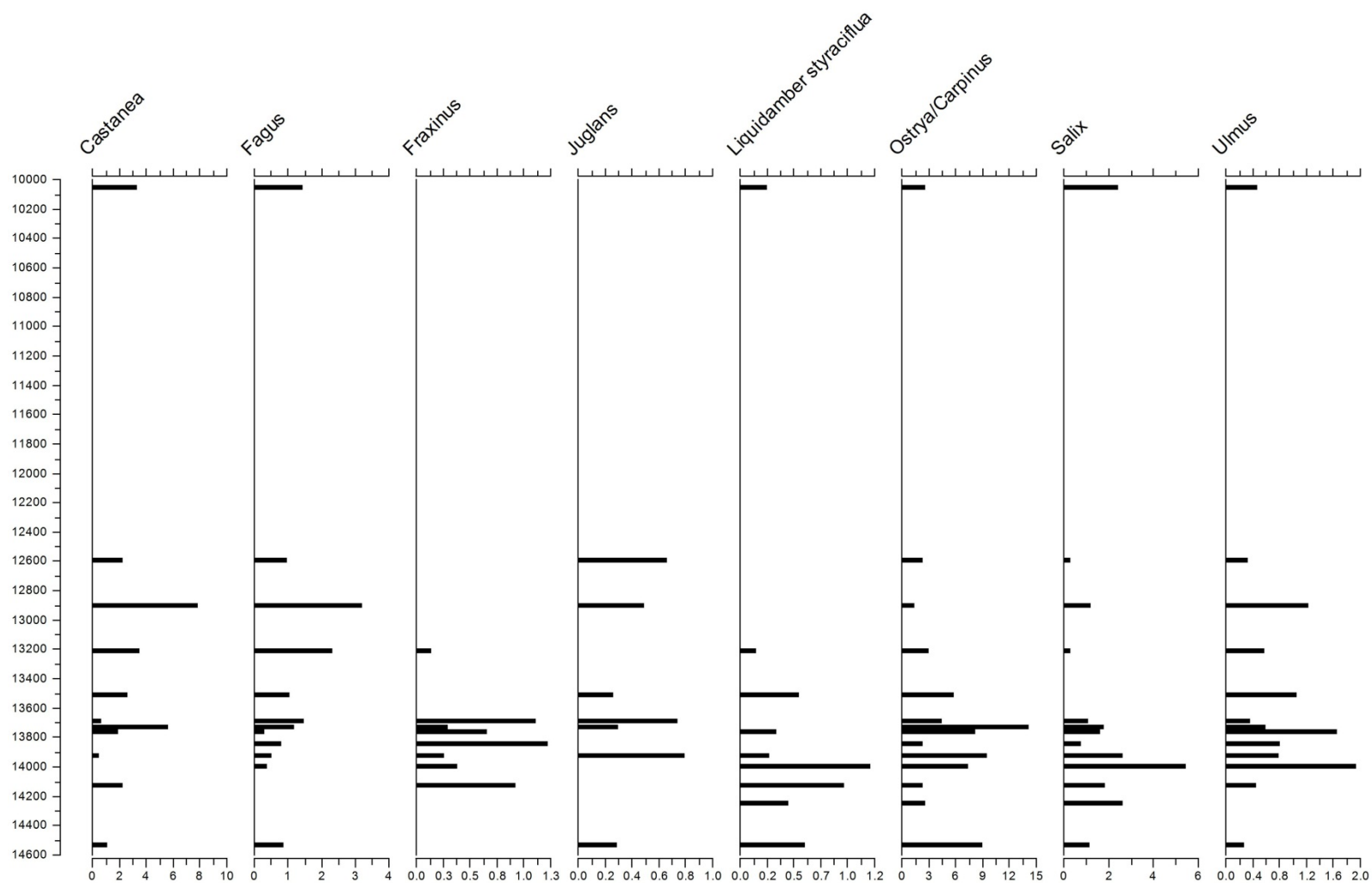
Figure 9. Pollen diagram for Whiteoak Bottoms for the period 14,532 to −59 cal yr BP (continued).

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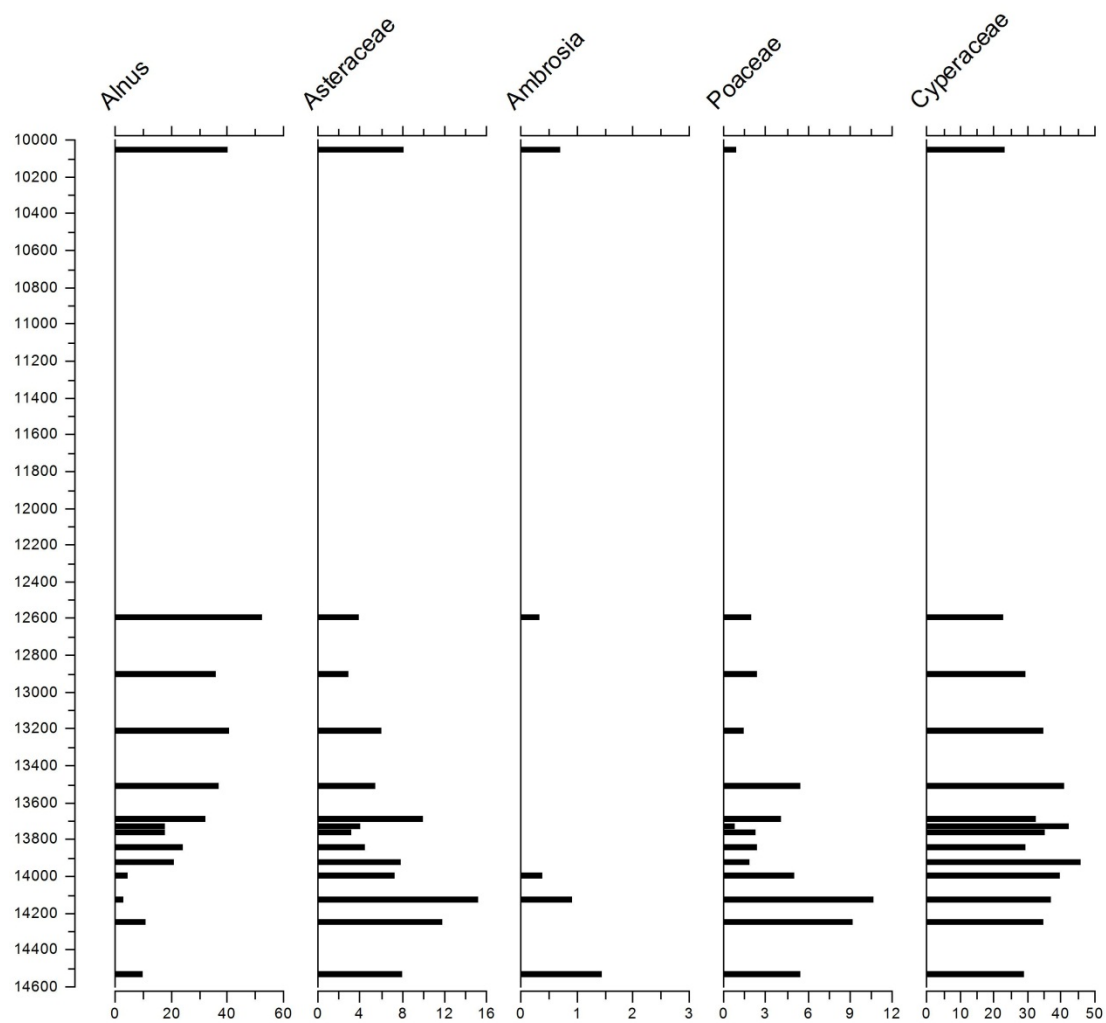
Figure 9. Pollen diagram for Whiteoak Bottoms for the period 14,532 to −59 cal yr BP (continued).

Cyperaceae pollen dominated the record with grain counts often twice those of the next most abundant taxon. *Alnus*, *Quercus*, and *Betula* were the next most common taxa, in that order. *Alnus* rises from 9.9% at 14,532 cal yr BP to 52% at 12,597 cal yr BP. From the peak at 12,597 cal yr BP, the percentage declines into the early Holocene. *Abies*, *Picea*, *Pinus*, and Pinaceae pollen percentages peak in the lower section of the core, showing a decreasing trend between about 14,500 cal yr BP to 14,000 cal yr BP and the early Younger Dryas at 12,597 cal yr BP. *Abies* percentages dip below 1% by 13,767 cal yr BP, disappearing entirely by 13,210 cal yr BP. *Picea* falls below 4% by 13,517 cal yr BP, but the taxon persists, in low quantities, into modern times. *Pinus* and Pinaceae also peak at 6710 and 177 cal yr BP. *Betula* percentages rise from 8.7% at the base of the core to 22% at 13,845 cal yr BP. *Betula* then slowly decreases into the Holocene, but rises again by 177 cal yr BP. *Ostrya*/*Carpinus* percentages peak at 14% by 13,728 cal yr BP before decreasing into the Younger Dryas. Percentages increase slightly at 6710 cal yr BP (35%) and again at 177 cal yr BP (6%). Pollen percentages for all other arboreal taxa are generally below 10% in the Whiteoak Bottoms record.

The most abundant nonarboreal taxa (excluding Cyperaceae) are Asteraceae and Poaceae. Percentages for both peak at 14,129 cal yr BP. Moving forward in time, they slowly decline into the Younger Dryas. Values increase for both families in the Holocene, peaking again at 2567 cal yr BP, before slowly declining into modern times. All other nonarboreal taxa contributed less than 10% to the pollen sum.

5.5 Microscopic Charcoal

Microscopic charcoal was quantified at 4 cm intervals through the core. Charcoal was present on all slides, but in varying concentrations (Figure 10). The interval between samples is ca. 30 years between 14,532 and 14,020 cal yr BP, ca. 39 years between 14,020 and 13,670 cal

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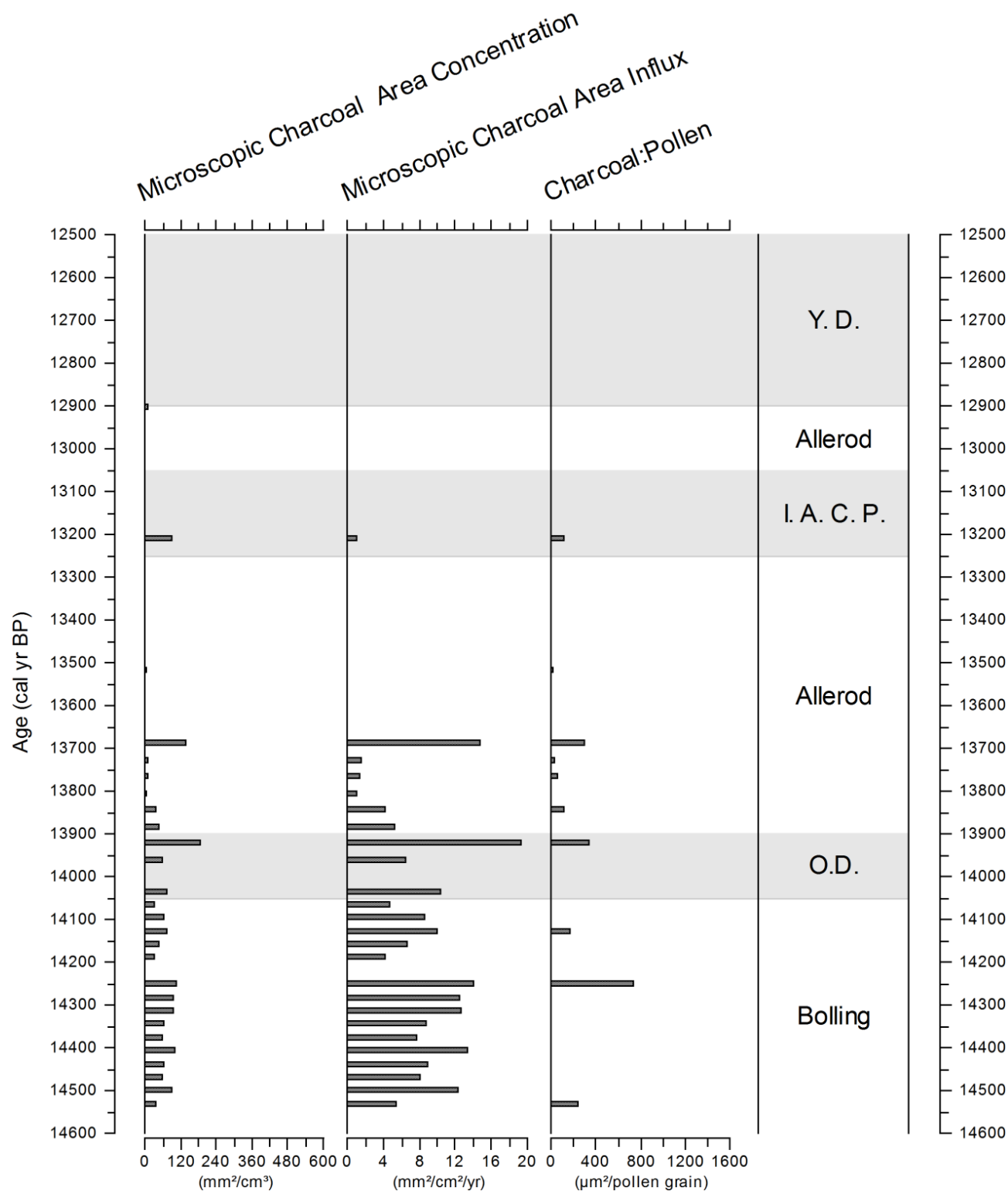
Figure 10. Charcoal indices for Whiteoak Bottoms for the period 14,532–12,500 cal yr BP. Shading shows intervals of cooler climate in the northern hemisphere based on Yu and Wright (2001) and Shuman et al. (2002). The Younger Dryas, Inter-Allerod Cold Period, and Older Dryas are abbreviated as Y.D., I.A.C.P., and O.D. respectively.

yr BP, and ca. 306 years between 13,670 and 12,597 years. After 12,597 cal yr BP (not shown in Figure 10), the record is of much lower resolution.

Charcoal area concentrations are relatively low through most of the core. Below 32 cm (6710 cal yr BP), 4 levels have concentrations over 100 mm^2/cm^3. These peaks are located at 140, 120, 80, and 56 cm (14,408, 14,253, 13,923, and 13,689 cal yr BP, respectively). Above 32 cm (6710 cal yr BP), there are 7 intervals with concentration values above 200 mm^2/cm^3. These occur at 32, 27, 24, 20, 16, 12, and 0 cm (6710, 1531, 1354, 1119, 883, 648, and −59 cal yr BP, respectively). Charcoal concentrations at 8 cm (412 cal yr BP) and 4 cm (177 cal yr BP) are above 150 mm^2/cm^3.

Over the length of the core, charcoal area influx ranges from 0.08 to 19.5 mm²/cm²/yr. Between 52 and 28 cm (~13,517 to 2567 cal yr BP), influx never exceeds 1.25 mm²/cm²/yr. The curves for charcoal concentration and charcoal influx show similar patterns. Corresponding stair-step patterns exist in both the Pleistocene and Holocene portions of the diagram. .

Mean charcoal area concentration for the entire core was 115 mm^2/cm^3; mean charcoal area influx was 6 mm^2/cm^2/yr. The mean charcoal to pollen ratio for the entire core was 296 μm^2/pollen grain. Prior to 12,500 cal yr BP, area concentration averaged 61.5 mm^2/cm^3, mean area influx was 7 mm^2/cm^2/yr, and the mean charcoal to pollen ratio was 174 μm^2/pollen grain.

CHAPTER VI
DISCUSSION

6.1 Does the Whiteoak Bottoms core provide a continuous sedimentary record and what does core stratigraphy reveal about the environmental history of the site?

The Whiteoak Bottoms core is primarily an archive of sedimentary evidence of late Pleistocene environments. The relatively fast rate of sediment deposition and the continuity in sediment type between the near-basal date of 14377 cal yr BP and the date of 12597 cal yr BP at 40 cm argue for continuous sedimentation. Above 40 cm, however, the sedimentary record is likely disturbed. At 37 cm, peat abruptly gives way to a grey silty layer. As the X-rays and the LOI data reveal (Figure 11), this grey layer is highly inorganic. The calculated sedimentation rate for this portion of the core decreases to 0.002 cm/yr, which indicates each cm represents ca. 636 years of accumulation. These shifts suggest that a break or hiatus in sedimentation exists between 12,597 and 8782 cal yr BP. Kneller and Peteet (1993) recorded a potential hiatus at Browns Pond, Virginia between 10,000 and 8000 ^{14}C years BP (approximately 11600 to 8900 cal yr BP). They interpreted a near absence of macrofossils and extremely inorganic sediments as indicative of lower water levels in the pond (Kneller and Peteet 1993; 1999). This portion of their core also showed the slowest sediment accumulation rate (Kneller and Peteet 1999). Webb et al. (2003) showed drier conditions in much of the Southeast at 10,000 cal yr BP (approximately 9000 ^{14}C yr BP), with moister conditions arriving in western Virginia between 9000 and 8000 cal yr BP (8000 and 7000 ^{14}C yr BP).

Another potential hiatus or break in the sedimentary record of Whiteoak Bottoms occurs at 27 cm, where sediments abruptly change from the grey silty layer to a brown silty layer. This shift is bracketed by radiocarbon dates of 8782 (34 cm) and 1531 (27 cm) cal yr BP. The

calculated sedimentation rate for this period in the record is 0.001 cm/year (ca. 1036 years of accumulation per cm), the slowest rate recorded in the Whiteoak bottoms core. The anomalous 2229 cal yr BP date comes from this section of the core. McDonald (2010) mentioned a failed attempt to drain the wetland at Whiteoak Bottoms and convert it to agricultural use during historic times, and significant logging also took place at the site. Both of these activities likely affected the wetland. The brown silty layer that includes the out of sequence date could be the result of this human disturbance. Clear cutting the forests on adjacent slopes during the late 1800s and early 1900s could have increased erosion and washed mineral sediment containing older soil charcoal into the wetland. Increased sedimentation into the river from clear-cutting upstream could have increased flooding in the valley, leading to deposition of sediments with old charcoal at the site. Human activity also could have resulted in loss of sediments from the wetland. The attempted draining of the wetland could have increased decomposition. Logging and other disturbances could also have resulted in erosion of sediments in the wetland.

The X-rays reveal numerous inorganic deposits within the peaty, late Pleistocene section of the Whiteoak Bottoms core (between 120 and 40 cm) (Figure 11). In their examination of the geomorphic history of the site, McDonald and Leigh (2011) noted several potential overbank deposits over the length of their cores. The silty grey layer between 37 and 28 cm in the WOB2A core is one of these deposits, corresponding to their T1f deposit. McDonald and Leigh (2011) interpreted this change in sediment as evidence of increased flooding or erosion from nearby hillslopes. At a site near Macon, Georgia, LaMoreaux et al. (2009) recorded several sand layers that they interpreted as evidence of increased fluvial activity between 11,000 and 4600 cal yr BP. Goman and Leigh (2004) found evidence for an increase in large flooding events (approximately 15 episodes) between 9000 and 6100 cal yr BP at their site on the coastal plain of
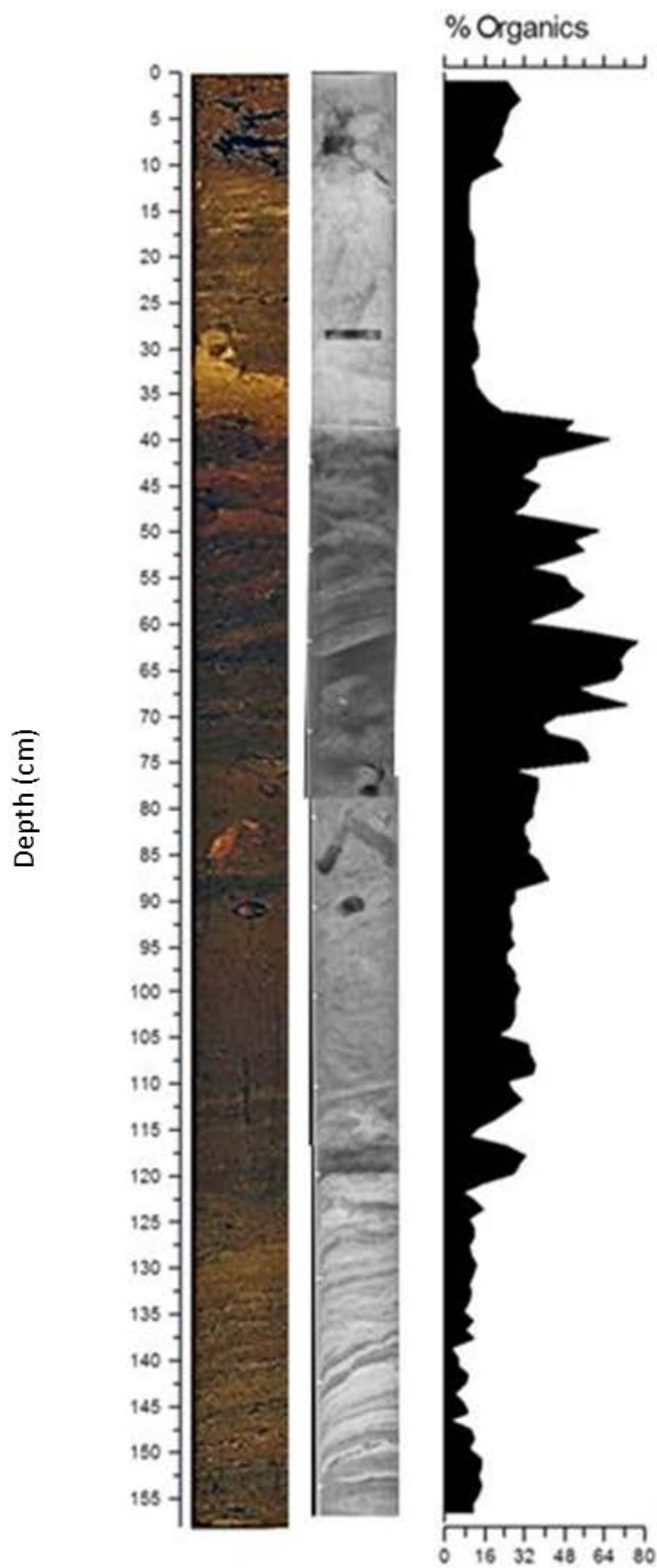
Figure 11. Whiteoak Bottoms photographic and X-ray images compared to LOI results.

North Carolina, positing a more northeasterly position for the Bermuda High as the cause. A more northeastern position for the Bermuda High would redirect moisture toward the Atlantic seaboard instead of the Gulf Coast, leading to more frequent flooding (Goman and Leigh 2004). Webb et al. (2003) offered some support for the northeastward movement of the Bermuda High. Mapping trends in moisture balance in the Southeast, they showed a west to east shift in positive moisture balance (increased moisture) from 10,000 to 7000 cal yr BP. Given its location, Whiteoak Bottoms would have experienced an increase in precipitation, potentially leading to increased flooding. Increased flooding could have led to the scouring and redeposition of sediments, resulting in an incomplete sedimentary record (and the deposition of the grey silty layer).

As a result of discontinuities in the upper section of the sedimentary record, I focused my attention on the most secure portion of the sedimentary record, the portion dating between ~14,532 and 12,597 cal yr BP.

6.2 What does the late Pleistocene LOI record at Whiteoak Bottoms reveal about late Pleistocene climate at the site, and how does this record compare with late Pleistocene climate records for the Southeast?

While organic matter has been accumulating, in low amounts, at Whiteoak Bottoms since its formation, LOI values first reach close to 30% (indicating peat; Rydin and Jeglum 2006) at 14,245 cal yr BP (119 cm) where the LOI is 27.8% (Figure 12). This is also the point at which sediments change from the basal sandy section of the core to more organic sediments. The LOI and organic influx curves from Whiteoak Bottoms show an interesting relationship with GISP2 $\delta^{18}O$ readings and reconstructed sea surface temperatures for the Cariaco Basin (Figure 13). As temperatures decrease, organic matter content increases and inorganic content decreases; the reverse occurs as temperatures increase. The increase in organic matter seen with decreases in
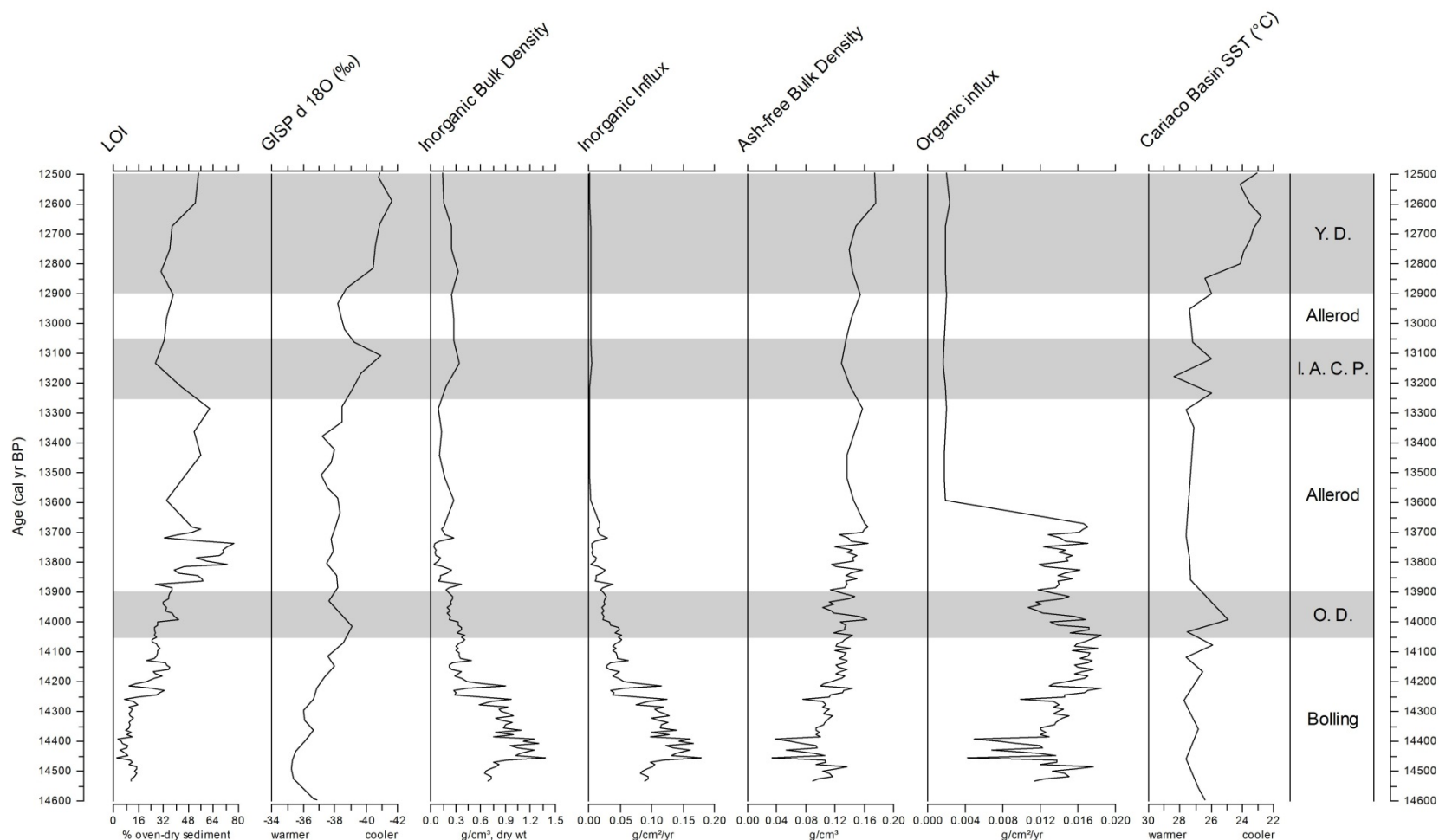
Figure 12. Comparison of LOI, inorganic and organic concentrations and sediment influx with GISP2 oxygen isotope data and reconstructed sea surface temperatures from the Cariaco Basin for the period 12,500 to 14,532 cal yr BP. Shading shows intervals of cooler climate in the northern hemisphere based on Yu and Wright (2001) and Shuman et al. (2002). The Younger Dryas, Inter-Allerod Cold Period, and Older Dryas are abbreviated as Y.D., I.A.C.P., and O.D., respectively. GISP2 δ [18]O data from Alley (2000, 2004). Cariaco Basin sea surface temperature data from Lea et al. (2003). Note differences in scale and direction of x axes.

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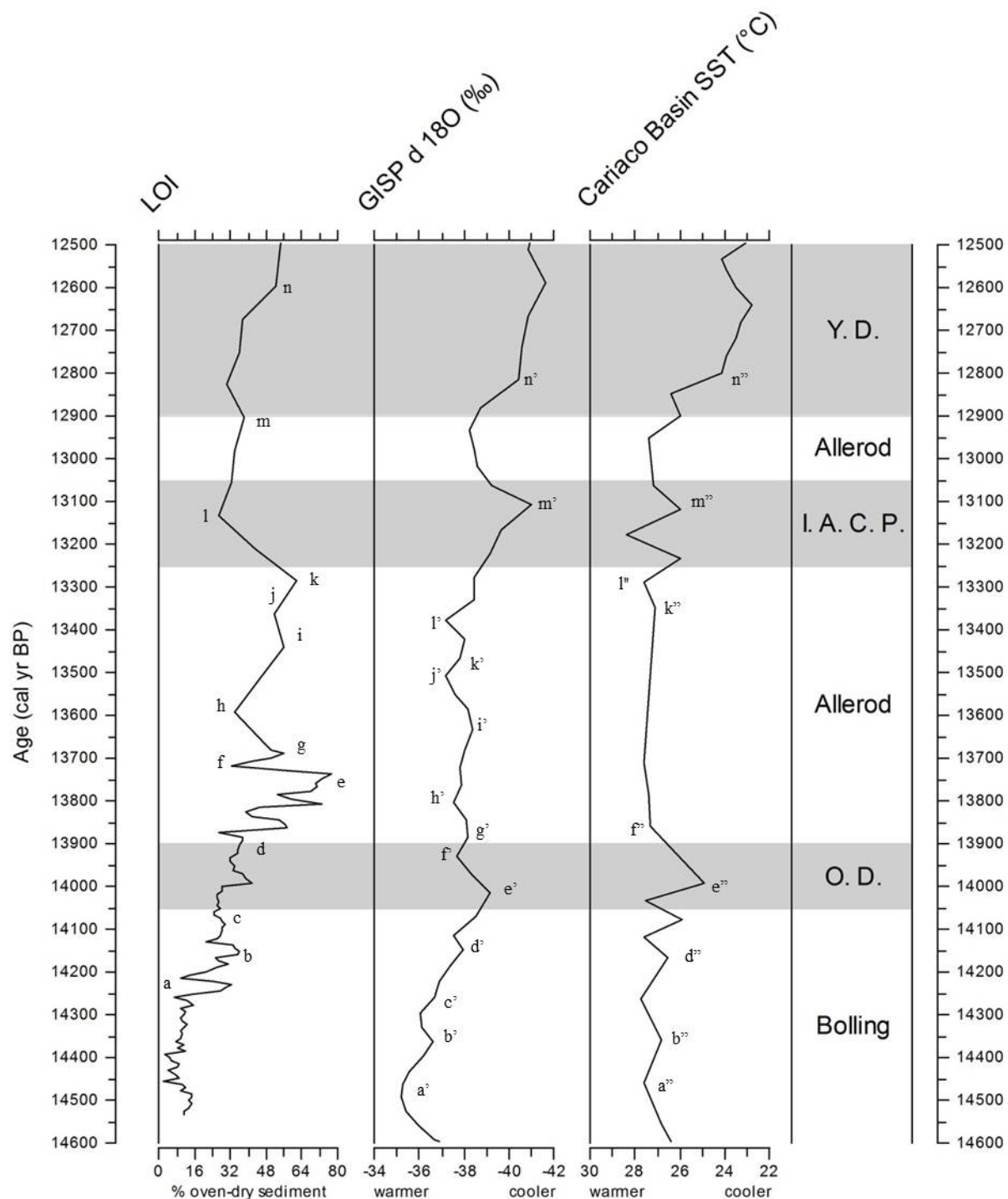
Figure 13. . Comparison of LOI with GISP2 δ ¹⁸O data (Alley 2000,2004) and reconstructed sea surface temperatures from the Cariaco Basin (Lea et al. 2003) for the period 14,532 to 12,500 cal yr BP. Shading shows intervals of cooler climate in the northern hemisphere based on Yu and Wright (2001) and Shuman et al. (2002). The Younger Dryas, Inter-Allerod Cold Period, and Older Dryas are abbreviated as Y.D., I.A.C.P., and O.D., respectively. Letters mark possible matches between sequences. (*i.e.*, a = a' = a'').

temperature may be the result of a more positive moisture balance. Cooler temperatures result in lower evaporation, resulting in a moister environment more conducive to organic matter preservation. Alternatively, cooler temperatures could be associated with greater rain or fog. Increased moisture could result in increased plant productivity, resulting in more deposition of organic matter, which is then preserved by the cooler, moister conditions. The increase in inorganic matter with increasing temperature may be the result of increased mineral sediment input (due to increased precipitation or erosion) or of increased decomposition of organic matter. At Whiteoak Bottoms, peat began to accumulate during the Bølling chronozone, as temperatures began to cool after reaching a peak in warmth at approximately 14,500 cal yr BP. Organic influx slowed during the Older Dryas chronozone, before rising slightly in the early Allerød. In the middle Allerød, sediment accumulation decreased sharply, but organic percentages remained high.

Based on the designation of peat as organic matter content of 30% or higher (Rydin and Jeglum 2006) episodes of peat formation occurred at Whiteoak Bottoms at 14,230 and 14,183, and from 14,160–14,136, 13,991–13,884, 13,864–13,210, and 13,057–10,690 cal yr BP (Figure 12). Interspersed between these organic zones were episodes where organic matter content fell below 30%. Peat accumulation occurred consistently, but not continuously, through the Whiteoak Bottoms core. These less organic intervals could represent periods during which decomposition outpaced deposition of organic matter. Another possible explanation, however, is that these intervals represent flooding events, as proposed by McDonald and Leigh (2011). As such they may be evidence of increased precipitation.

Overall, the LOI record shows a pattern of change resembling that of GISP2 oxygen isotope data and Cariaco Basin sea surface temperatures (Figure 13). The potential flooding

events seen in the core photographs and x-rays could represent single or multiple events. As these events would represent a change from the "normal" accumulation of sediments at the site, their presence potentially skews the calculated sediment accumulation rates, interpolated radiocarbon ages, and influx values. This could be the reason for an apparent 200 year lag between the GISP2 and Whiteoak Bottoms LOI records (Figure 13). Alternatively, this lag could indicate a delay between changes in the North Atlantic affecting the southeastern United States. Applying this 200-year lag would suggest that the Younger Dryas signal in the Whiteoak Bottoms record would last from approximately 12,700 –11,300 cal yr BP.

6.3 What does the late Pleistocene vegetation record at Whiteoak Bottoms reveal about vegetation and climate at the site, and how do the data compare with other late Pleistocene records of vegetation and climate for the Southeast?

The late Pleistocene to early Holocene transition in pollen records from the Southeast often shows a tripartite successional sequence of vegetation change. In the southern Appalachians, this pattern consists of coniferous forests dominated by spruce and pine transitioning to mesic deciduous forest consisting of oak, birch, hemlock, beech, *Ostrya*/*Carpinus*, willow, and elm, among others. This association, in turn, later gives way to oak-chestnut and, following the chestnut blight, to oak-hickory forests.

The establishment of mesic forest vegetation at Whiteoak Bottoms was underway by 14,532 cal yr BP. At 14,532 cal yr BP, the base of the core, total conifer pollen (excluding hemlock) made up 31% of the pollen sum. Total deciduous tree taxa (including hemlock) made up 38% of the sum. Hemlock was included in the deciduous taxa, instead of with the conifers, based on its common association with more mesic forest taxa in the pollen record. Additionally, *Tsuga* does not enter the pollen record until 13,767 cal yr BP, after charcoal values decline and *Abies* diminishes. *Tsuga* is extremely fire sensitive (Rogers 1978; Carey 1993). That *Tsuga* does

not enter the record until after evidence for fire diminishes could indicate that fires were frequent enough in the vicinity of Whiteoak Bottoms to exclude hemlock prior to 13,767 cal yr BP. Alternatively, temperatures could have been too cold for hemlock; however, *Tsuga* shows up in the pollen records of Jackson Pond, Kentucky and Cranberry Glades, West Virginia, before it appears in the Whiteoak Bottoms pollen record. This leaves fire as a more probable exclusionary force than temperature in this case.

The lower portion of the Whiteoak Bottoms core appears to represent a mixedwood forest, perhaps similar to the mixed coniferous-deciduous forests of eastern Canada. A brief return to conifer dominance (38% of the pollen sum as opposed to 29% for deciduous tree taxa) occurred at 14,253 cal yr BP, but by 14,129 cal yr BP, deciduous forest again dominated. At Jackson Pond, coniferous forest briefly returned to dominance at 12,483 and 11,843 cal yr BP (Wilkins et al. 1991). The return of coniferous forest to the area surrounding Jackson Pond marks the Younger Dryas chronozone. Cooler temperatures would have facilitated the return of more cold-adapted vegetation. The resurgence of coniferous taxa at Whiteoak Bottoms at 14,253 cal yr BP is somewhat anomalous as it appears to occur during the Bølling interstadial. Kneller and Peteet (1999) found a similar return to coniferous forest in their zone BR3b (14,320 to 14,240 cal yr BP). They interpreted increases in fir and decreases in many deciduous tree taxa to indicate a return to colder conditions.

The timing of the transition from coniferous boreal or montane-like forest to a northern hardwoods forest at Whiteoak Bottoms is in line with the timing at White Pond, Cranberry Glade Bog, Shady Valley Bog, Bob Black Pond, and Quicksand Pond. The date of mesic forest establishment at Anderson Pond seems anomalously early and will require further investigation. A rough west to east pattern emerges in the dates of deciduous forest dominance, with the

earliest dates southwest of the latest dates. This pattern is reminiscent of the moisture balance

patterns seen in Webb et al. (2003) and Shuman et al. (2002). Goman and Leigh (2004) posited

a more southerly to southwesterly position for the Bermuda High prior to 12,000 cal yr BP. This

would have changed precipitation patterns in the area. By 6000 cal yr BP, the Bermuda High

reached a more northerly or northeasterly position (Goman and Leigh 2004). Between

approximately 10,000 and 6000 cal yr BP (the period during which the Bermuda High was

migrating northeast), they noted evidence of increased flooding, which they attributed to

increased moisture availability. The transition from coniferous to northern hardwood forest may

have been the result of a shifting Bermuda High.

The Standing Indian Bog (SIB) pollen diagram produced by Sheehan (Figure 3) reveals

stratigraphy similar to that recorded by McDonald (2010) and McDonald and Leigh (2011), as

well as this work, for Whiteoak Bottoms (WOB). The lower peat facies from the SIB diagram

could equate to the lower peat facies in the WOB cores. The clay layer at SIB could correspond

with the silt layers at WOB, and the upper peat layer would equate to the upper/surface peat at

WOB. If Whiteoak Bottoms and Standing Indian Bog represent the same site, then the peat to

clay transition should mark the beginning of the Younger Dryas, based on dates obtained for

Whiteoak Bottoms.

The SIB diagram shows a large peak in *Alnus* across the transition from peat to clay

facies. Mayle et al. (1993) proposed the use of *Alnus* as a bio-indicator of the Younger Dryas,

based on work in the northeastern United States and Atlantic Canada. They found that alder

pollen percentages rose significantly at the beginning of the Younger Dryas and dropped sharply

at the end of the Younger Dryas. *Alnus* peaks at 12,597 cal yr BP in the WOB pollen record.

Tsuga offers an additional biostratigraphic marker for the SIB pollen diagram. *Tsuga* enters the

SIB record at approximately 140 cm. *Tsuga* enters the Whiteoak Bottoms record by 13,767 cal yr BP, and percentages peak at approximately 10,053 cal yr BP.

Standing Indian Bog, like the Whiteoak Bottoms record, lacks an obviously boreal vegetation assemblage. The transition from conifer-dominated forests to forests dominated by more mesic deciduous taxa seen in other southeastern pollen records also appears absent from the Standing Indian Bog diagram. The basal layers appear to represent a mixedwood forest, as at Whiteoak Bottoms. In the SIB pollen record, percentages for *Abies* and *Picea* between 90 and 80 cm return to or surpass basal levels for these two taxa. This change could represent a return to cooler or possibly wetter conditions, or both, and is not seen in the post-Younger Dryas portion of the Whiteoak Bottoms pollen record. While the Whiteoak Bottoms record shows a possible return to cooler conditions, this transition occurs at 14,252 cal yr BP, which is too early to correspond to increases in *Abies* and *Picea* seen in the Standing Indian Bog record. The correspondences in lithology and the *Alnus* peaks seen in the SIB and WOB records point to the possibility that the two sites are the same place, but without more data and dates on the SIB record, this remains speculative.

6.4 What does the late Pleistocene charcoal record at Whiteoak Bottoms reveal about fire at the site, and how does the record compare to other fire and climate records for the Southeast?

Fire activity, as reflected in the three charcoal indices, has varied over the history of the site. Pre-Younger Dryas, periods of greater than average charcoal concentration occurred between 14,501 and 14,253, 14,129 and 14,098, 13962 and 13,923, and at 14,036, 13,689, and 13,210 cal yr BP. Area concentrations measuring over twice the mean occurred at 13,923 and 13,689 cal yr BP; area influx exceeding twice the mean occurred at 14,252, 13,923, and 13,689 cal yr BP; and charcoal to pollen ratios greater than twice the mean occurred at 14,253 and

13,923 cal yr BP. Maxium values for the three charcoal indices, however, did not exceed 189

mm^2/cm^3, 19.5 $mm^2/cm^2/yr$, and 744 $\mu m^2/pollen$ grain. In comparison, mean values of these

metrics for pine rocklands, an ecosystem adapted to frequent, low intensity fires, were 308

mm^2/cm^3, 80.7 $mm^2/cm^2/yr$, and 7413 $\mu m^2/pollen$ grain. Charcoal area concentration values for

Whiteoak Bottoms are more similar to those found by Tinner et al. (2006) for two boreal Alaska

sites spanning the mid- to late Holocene than those found in pine rocklands by Albritton (2009)

during the late Holocene. These comparisons appear to indicate that fire activity at Whiteoak

Bottoms was similar to, though perhaps slightly greater than, that seen in boreal environments.

Charcoal area influx rates generally under 0.5 $mm^2/cm^2/yr$ prevailed through the

Holocene at southern Appalachian sites studied by Lynch and Clark (2002). Influx values were

generally under 1 $mm^2/cm^2/yr$ for a site in the coniferous boreal forest of western Quebec

spanning approximately the last 7000 cal yr BP (Ali et al. 2008). Charcoal area influx at

Whiteoak Bottoms appears more similar to influx values seen in boreal Canada and the southern

Appalachians than values seen in pine rocklands. The disparity between influx values reported

for Whiteoak Bottoms and those reported by Lynch and Clark (2002) could be a matter of

elevation. Of the sites studied by Lynch and Clark (2002), only one, Pink Beds Bog, has an

elevation similar to Whiteoak Bottoms. This record, however, only extends back to

approximately 11,200 cal yr BP. Influx values similar to those at Pink Beds Bog, between 0.5

and 1 $mm^2/cm^2/yr$, appear in the Whiteoak Bottoms record between 13,000 and 10,000 cal yr

BP. The charcoal to pollen ratios for Whiteoak Bottoms are more similar to those reported from

boreal and mesic deciduous to mixed conifer-deciduous forests, from at least the mid-Holocene

into historic times, (MacDonald et al. 1991; Fuller et al. 1998; Patterson 2005) than to those

reported by Albritton (2009) for pine rockland communities. Values for all three charcoal

indices indicate that the pre-Younger Dryas fire regime at Whiteoak Bottoms was most similar to that seen in more northern environments during the late Holocene.

Few sedimentary charcoal records exist for the southeastern United States; charcoal records that extend into the Pleistocene are even fewer. Between 14,532 and 13,670 cal yr BP, microscopic charcoal area concentration and influx values at Whiteoak Bottoms appear to correspond with warming in the GISP2 δ^{18}O and Cariaco Basin sea surface temperature records (Figure 14), as well as with the inferred warming at Browns Pond. This period of fire activity spans the Bølling/Allerød, with a brief interval of no recorded fire during the middle Older Dryas. Peak charcoal concentration and influx during the Bølling portion of the record appears to lag peak GISP2 temperature by approximately 200 years, the same potential offset seen in the LOI record. Within the Bølling portion of the record, a brief gap appears between 14,253 and 14,191 cal yr BP, during which one sampled level contained no charcoal. This charcoal gap corresponds with a brief peak in organic matter content, centered at approximately 14,222 cal yr BP.

There are three charcoal to pollen ratio data points during the Bølling portion of the WOB2A core. These points occur at 14,532 (the basal sand of the core), 14,252, and 14,128 cal yr BP. The highest charcoal to pollen ratio in the 14532–12,500 cal yr BP portion of the Whiteoak Bottoms record is at 14,253 cal yr BP. This was also the time of a brief return to conifer domination at the site, indicating cooler conditions. Kneller and Peteet (1999) found a similar short cold reversal marked by an increase in *Abies* at Browns Pond between 14,320 and 14,240 cal yr BP. Additionally, they reported low to moderate charcoal influx values for this period. The peak in organic matter between 14,253 and 14,222 cal yr BP (120–116 cm) in the

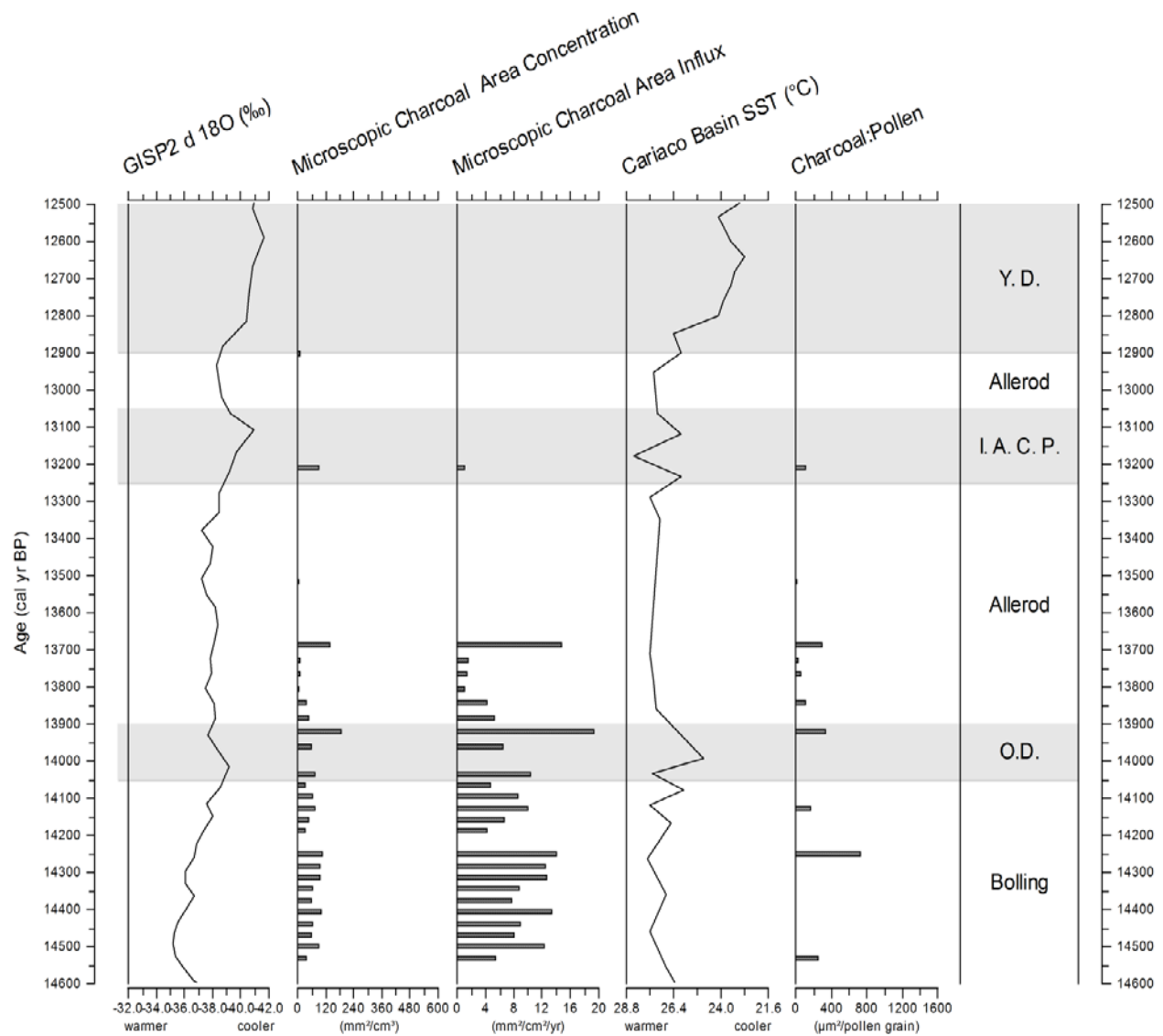**Figure 14. Comparison of charcoal indices with the GISP2 δ 18O record (Alley 2000, 2004) and Cariaco Basin sea surface temperature records (Lea et al. 2003). Shading shows intervals of cooler climate in the northern hemisphere based on Yu and Wright (2001) and Shuman et al. (2002).**

Whiteoak Bottoms core could be related to the peak in charcoal to pollen ratio at 14,253 cal yr BP. Given that microscopic charcoal point counting yielded zero hits at 14,222 cal yr BP, the increase in organic matter is unlikely to represent post-fire deposition of sediment eroded from nearby uplands. The interpolated age of 14,253 cal yr BP corresponds with 120 cm. This depth marks the transition from the sandy basal sediments to peat (Figure 13). I interpret the increase in organic content to indicate a brief return to cooler conditions.

A second charcoal gap exists at 14,001 cal yr BP at Whiteoak Bottoms. This gap falls within the Older Dryas chronozone and is concurrent with a brief increase in organic matter (Figure 14). Kneller and Peteet (1999) interpreted their pollen assemblage for the period 14,240 to 13,260 as representing a return to warmer, yet still cool and moist, conditions. At Whiteoak Bottoms, *Picea* reaches its highest value at 11%, and peaks occur in *Liquidambar styraciflua*, *Salix*, and *Ulmus*, at 14,001 cal yr BP. I interpret the presence of pollen of these taxa, constituents of the northern hardwood forest, and the increase in organic matter as indicating another brief cold reversal. Kneller and Peteet (1999) inferred warming at their site from the presence of hemlock needles. That Kneller and Peteet (1999) do not find evidence of the Older Dryas at Browns Pond may be related to elevation. The elevation at Whiteoak Bottoms is 1032 m; the elevation at Browns Pond is only 620 m. As a result, Whiteoak Bottoms may have been more sensitive to this brief climate excursion.

All three charcoal indices show values twice the mean at 13,923 cal yr BP. This period corresponds to the terminal Older Dryas, and the charcoal peaks could represent fire brought on by changing climate conditions. Warming temperatures could have led to the desiccation of upland areas, leading to an increase in fire activity.

In the Whiteoak Bottoms core, the period between approximately 13,000 and 10,900 cal yr BP includes one level that shows a charcoal peak in all three indices. At 13,210 cal yr BP, during the Inter-Allerød Cold Period, a charcoal peak appears just after a small peak in organic matter. Peaks in organic matter prior to or concurrent with fire could reflect an increase in plant productivity that led to the buildup of fuel on the landscape.

In the Whiteoak Bottoms core, charcoal to pollen ratios show relationships with percentages of some pollen types (Figure 15). The charcoal to pollen ratio appears to directly correspond with percentages of *Abies*, Asteraceae, Poaceae, and *Pinus*. Asteraceae and Poaceae are disturbance taxa, and *Pinus* is fire tolerant. Of the five taxa, *Abies* shows the best correspondence, with the pattern of charcoal to pollen ratios closely matching the pattern of fir pollen percentages. *Abies* disappears from the Whiteoak Bottoms pollen record after 13,210 cal yr BP, concurrent with a reduction in charcoal area concentration, area influx, and charcoal to pollen ratio. The highest charcoal to pollen ratios (over four times the mean value), at 14,253 cal yr BP, matches a peak in fir pollen. This appears to indicate a relationship between fire and *Abies* at Whiteoak Bottoms. Two species of *Abies* are extant in the Southern Appalachians today, *Abies balsamea* and *Abies fraseri*. Both are fire-sensitive species (Uchytil 1991; Sullivan 1993). *Abies fraseri* occurs in habitats rarely subjected to fire, as fuel moisture and humidity levels in the southern Appalachians are typically high enough to mute fire intensity and extent (Sullivan 1993). Uchytil (1991) reported that *Abies balsamea* is especially slow to reestablish following fire. Given the apparent relationship between *Abies* and charcoal at Whiteoak Bottoms, high charcoal to pollen ratios could be the result of relatively nearby fires. Given their sensitivity to fire, the charcoal to pollen ratio could reflect burning on nearby ridges, with the *Abies* pollen reflecting unburned vegetation growing adjacent to the wetland.
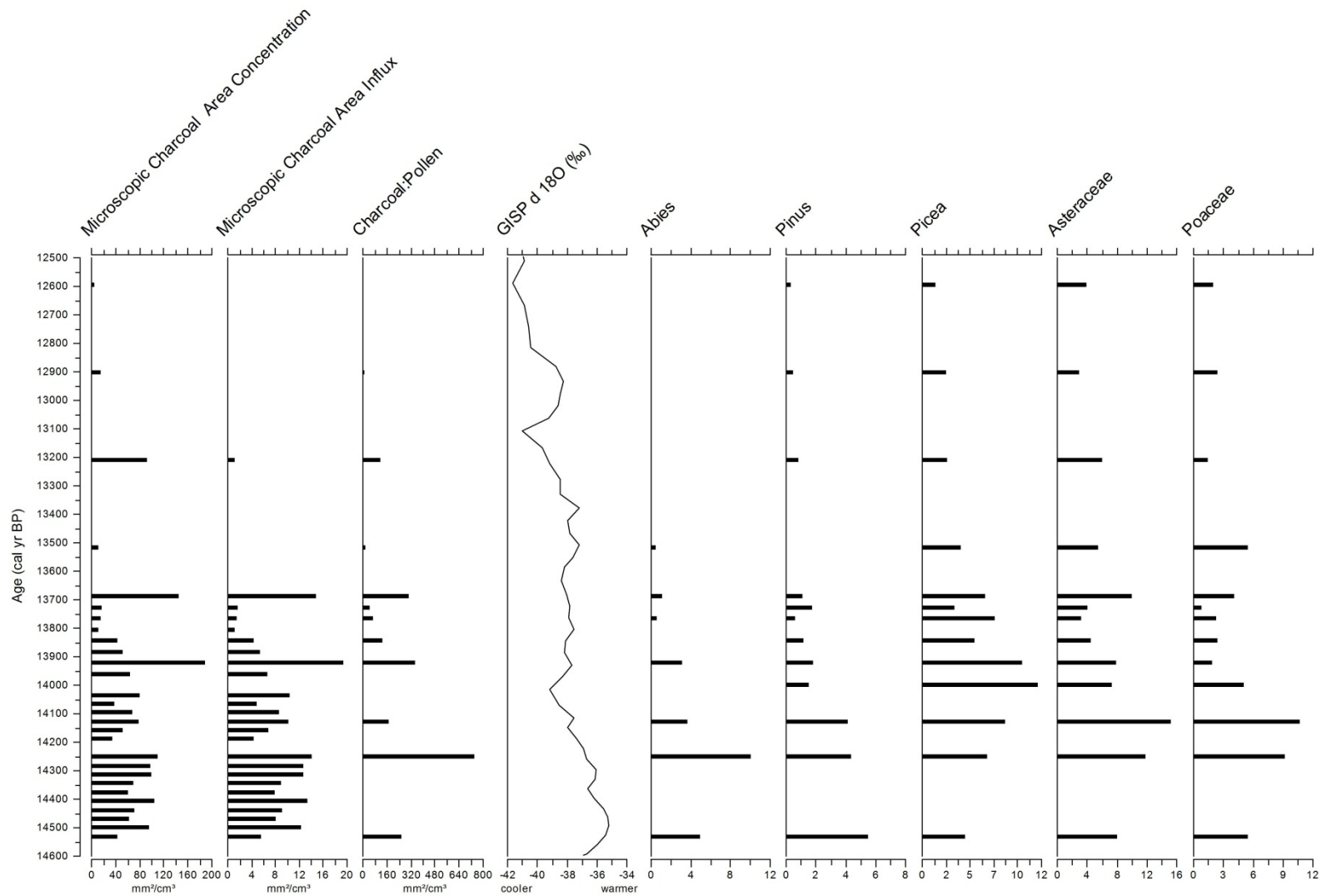
Figure 15. Comparison of charcoal indices and selected pollen taxa for the period 14,532 to 12,500 cal yr BP. Note similarities between the charcoal:pollen ratio and *Abies* pollen percentages. GISP2 δ ^{18}O data from Alley (2000, 2004). Note scale reversal for GISP data.

Whether the fires seen in the Whiteoak Bottoms record indicate periodic low intensity fires within the watershed or reflect fires farther away is unclear. That organic accumulation is fairly constant and charcoal values are relatively low in the lower 117 cm of the core argues against the wetland itself having burned.

As the Whiteoak Bottoms fire record seems to correspond with warming temperatures, fire events may represent dry periods. Based on moisture balance records reported in Webb et al. (2003), vast areas should have been susceptible to burning at around 10,000 cal yr BP. Results from Pink Beds, a bog in North Carolina, appear to argue against this (Lynch and Clark 2002). The mid-Holocene (especially between ca. 8000 and 6000 cal yr BP), not the earliest Holocene, seems to have had more fire activity (Lynch and Clark 2002). One peak between 10,000 and 9200 cal yr BP rises above 0.125 mm^2/cm^2/yr (Lynch and Clark 2002), but this peak is associated with high values of chestnut pollen and is half the value of the peaks between ca. 8000 and 6000 cal yr BP. Lynch and Clark (2002) proposed that fire may have played an important role in the regeneration and maintenance of chestnut forests. The peak between 10,000 and 9200 cal yr BP could reflect a local fire event instead of a more regional signal. At Browns Pond, charcoal concentrations reach their highest levels between 10,050 and 8410 ^{14}C BP (10,940 to 9400 cal yr BP), ranging from less than 1 to 8 mm^3/cm^3 (Kneller and Peteet 1999). Kneller and Peteet (1999) interpreted these values as indicating a local fire that led to the opening of the forest canopy. Oak replaced hemlock in dominance and *Nyssa* increased. They reported sediments to be highly inorganic during this period, with macrofossils virtually nonexistent. Kneller and Peteet (1999) interpreted these changes in pollen, macrofossils, and sediments as indicating the pond dried out, but that the climate remained at least as moist as present conditions. This interpretation agrees with the findings of Webb et al. (2003) of drier conditions

throughout much of southeastern North America during this time. A small peak in charcoal at 10,053 cal yr BP could be related to a slight increase in pine and spruce at Whiteoak Bottoms during this period.

For the period 6710 cal yr BP to the present, the Whiteoak Bottoms core exhibits the most consistently high charcoal concentration, influx, and charcoal: pollen ratios (not shown). This, however, is also the least stratigraphically and temporally secure portion of the Whiteoak Bottoms core. The possibility of one or more hiatuses in the record make influx values unreliable above 40 cm (12,597 cal yr BP). Additionally, high charcoal levels above 34 cm (8782 cal yr BP) may be the result of historic disturbance. Analysis of the Holocene assemblages, along with supplementary analyses of the Pleistocene record at Whiteoak Bottoms, awaits further study.

CHAPTER VII
CONCLUSIONS

The analysis of a sediment core from Whiteoak Bottoms, a small peat-forming wetland in western North Carolina, provides evidence of environmental change spanning the late Pleistocene and early Holocene. Previous research at the site revealed that the wetland formed between 15,000 and 14,000 cal yr BP in a paleochannel of the Nantahala River. Peat, defined as sediment with organic matter content equal to or greater than 30%, began accumulating by approximately 14,230 cal yr BP. Episodes of peat formation and preservation occurred at 14,183 cal yr BP, and from 14,160–14,136 cal yr BP, 13,991–13,884 cal yr BP, 13,864–13,210 cal yr BP, and 13,057–10,690 cal yr BP. Organic matter accumulation trended with climate. As temperatures cooled coming out of the early Bølling interstadial, organic matter built up in the wetland sediments. Periods of warming corresponded with decreases in organic matter content. Whether these more inorganic sediments were the result of the deposition of mineral sediment from floods or slopewash, or derived from the decomposition of organic matter, is uncertain.

The vegetation growing in and around Whiteoak Bottoms has changed over time. While most of the plant taxa recorded by pollen in the sediments currently inhabit the area, their importance has changed over the history of the peatland. Vegetation of the wetland itself has likely changed little, but surrounding forests changed from coniferous, probably fir-pine-spruce forest, to a more mesic, possibly northern hardwoods-type forest, and then to the modern pattern with a more xeric oak-chestnut-hickory forest on ridgetops and more mesic cove forest surrounding the wetland and on lower slopes. The timing of the transition from a coniferous to mesic forest is in line with previous research in the southeastern United States. The Whiteoak Bottoms record supports a potential west to east gradient in the timing of this transition,

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following a pattern similar to that of the moisture balance data reported by Shuman et al. (2002) and Webb et al (2003). The shifting of the Bermuda High to a more easterly or northeasterly position is one possible explanation for a west to east change in forest composition.

Concomitant with changes in vegetation, fire activity at Whiteoak Bottoms also changed. Though never a major ecological factor, fire activity was highest when coniferous forest dominated the pollen spectra. Additionally, fire activity, as reflected by charcoal concentration, charcoal influx, and charcoal to pollen ratios, increased as temperatures increased. Increased temperatures may have led to drying episodes which could have led to increased fire activity. Intervals of cooler and wetter climate, and the presence of more mesic taxa, muted fire activity in the surrounding area, but did not totally remove it.

Future particle size analysis of the Whiteoak Bottoms core may improve interpretation of the grey and brown silty layers in the Younger Dryas and Holocene sections of the core by revealing trends in particle size that may indicate the nature of depositional events. Thin-section analyses of sediments will also help reveal depositional dynamics. Documenting peat humification, as a proxy for past wetness, could help identify drying episodes in the peat record. These could be tied to changes in vegetation and fire to provide a more detailed climate history.

Obtaining additional pollen data would also be useful. Additional analyses could confirm trends in both the pollen and microscopic charcoal data. Additional pollen sampling across the Younger Dryas transition and across the grey silt layer could provide evidence, in concert with the results of the particle size analysis, that could help resolve the chronology and establish the extent of disturbance in the upper sediments of Whiteoak Bottoms.

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APPENDICES

Appendix A

Pollen Processing Procedure

The following procedure was used to process sediment samples from Whiteoak Bottoms for pollen and microscopic charcoal analysis. Samples were processed in 15 ml Nalgene® polypropylene centrifuge tubes. The centrifuge used was an IEC model CL benchtop centrifuge with a 6 x 15 ml swinging bucket rotor. All centrifugations were carried out at the highest speed.

1. Add 1 *Lycopodium* tablet to each centrifuge tube.

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 180 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.

Charcoal Point Counting Calculations

Calculations below are based on Clark (1982), with modifications in some cases by S. Horn. This appendix is based on an unpublished laboratory handout by S. Horn prepared February 8, 2005 and modified May 3, 2006.

Areal density of charcoal on the slide (P) (e.g., the estimated probability that a random point will fall on charcoal):

$$P = C/N$$

The accuracy, or relative error, of the estimate of P (Sp/P):

$$(Sp/P) = \sqrt{(1-P)/C}$$

Area of charcoal (total) in mm^2 in all the fields of view you counted (Af):

$$Af = P * Fat$$

Estimated area of charcoal in mm^2 in the entire pollen sample (Aps)

$$Aps = (M * Af)/Mc$$

Charcoal area in mm^2 expressed on the basis of volume of wet sediment (Acc):

$$Acc = Aps/V$$

Charcoal area in mm^2 expressed on the basis of wet sediment mass (Awm):

$$Awm = Aps/W$$

Charcoal area in mm^2 expressed on the basis of dry sediment mass (Adm):

$$Adm = Aps/W(1-Pw)$$

Charcoal area in mm^2 expressed on the basis of annual influx (Acy):

$$Acy = Acc * \text{sedimentation rate expressed in cm/yr}$$

Charcoal: Pollen ratio (C:P) expressed as mm^2 charcoal per pollen grain:

$$C:P = (Mpo * Af)/(Mc * Popc)$$

Charcoal: Pollen ratio (C:P) expressed as μm^2 charcoal per pollen grain:

$$C{:}P = (Mpo * Af * 10^6)/(Mc * Popc)$$

Definitions:

Po = points applied in each field of view

F = the number of fields of view you looked at

N = total number of points applied (equal to Po * F)

Fa = area in mm^2 of each field of view

Fat = Total area in mm^2 on slide in which you looked at charcoal (equal to Fa * F)

C = the number of applied points that "touched" charcoal

V = volume in cm^3 of original wet sediment sample processed for pollen

W = mass in g of original wet sediment processed for pollen (from LOI sheet)

Pw = percent water in the wet LOI sample from same level as pollen sample

M = number of *Lycopodium* marker spores added to original sample processed for pollen

Mc = the number of *Lycopodium* marker spores you counted in the fields of view in which you did the point counting

Mpo = the number of *Lycopodium* marker spores you counted in the pollen count from the same level

Popc = the total number of pollen grains (excluding spores and indeterminates) counted in the pollen count from the same level.

VITA

      Mathew S. Boehm was born in Philadelphia, Pennsylvania. He received his Bachelors of Arts in Anthropology from the University of West Georgia, in Carrollton, Georgia. Mathew entered the Master's program at the University of Tennessee in 2009. He served as a teaching assistant for introductory physical geography and dendrochronology classes from 2009 to 2010, and served as a research assistant from 2010 to 2012. Upon completion of his M.S. degree, he will enter the Ph.D. program in Geography at the University of Tennessee, where he will continue to study the paleoenvironments of the Southeast and, perhaps, places farther afield.