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PALEOENVIRONMENTAL AND PALEOLANDSCAPE RECONSTRUCTIONS
OF GREENWICH BAY REGION, RI

BY
CAMERON E. MORISSETTE

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

IN

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UNIVERSITY OF RHODE ISLAND

2014

MASTER OF OCEANOGRAPHY THESIS

OF

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DEAN OF THE GRADUATE SCHOOL

UNIVERSITY OF RHODE ISLAND

2014

Abstract

A paleoenvironmental and paleolandscape reconstruction of two separate sites in the Greenwich Bay region of RI was completed for use in the Paleocultural Landscapes Project. Investigations of a lacustrine sediment record in Warwick, RI were completed in order to develop a regional paleoenvironmental reference record. Investigations of the acoustic sub-bottom record in Greenwich Bay, RI were completed in order to develop paleolandscape reconstructions of the observed depositional environments for future use in an archaeological site prediction model.

For Chapter one, we took a single sediment core from Gorton Pond, Warwick, RI in the summer of 2013 for the purpose of developing a regional paleoenvironmental reference record. The core was dated using both AMS-radiocarbon dating of four terrestrial macrofossil samples, and correlations with a radiometric/pollen age model from a surface core taken from Gorton Pond for a previous study. The oldest terrestrial macrofossil was a spruce cone dated to ~12,100 calendar years BP, but an estimated 18,000 years of sediment is in the Gorton Pond record, based on bracketing ages from dated recessional moraines in the region. However, we only sampled an estimated 16,000 years, based on a Younger Dryas transition of 13,000 BP.

Physical, magnetic, and elemental/isotopic (C, N, S, D/H) proxy measurements were made, and six distinct environmental zones were interpreted. The Gorton Pond record indicates significant changes in lake productivity and temperature. The Late Pleistocene Deglacial Period (16.0–13.0 ky BP) is indicated by low OC%, high concentrations of magnetic material, and terrestrially-sourced organic matter. The Younger Dryas chronozone was observed (13.0–11.8 kyr BP) as large decreases in

$\delta^{13}\text{C}$, OC%, and $\delta\text{D}_{\text{BA}}$, indicating cold conditions. A transition from the Late Pleistocene to Early Holocene (11.8–9.6 ky BP) was observed by rapid warming and increased aquatic productivity. The Middle Holocene (9.6–3.6 ky BP) was characterized by depleted $\delta\text{D}_{\text{BA}}$ ratios, which may suggest Gorton Pond is sensitive to changes in air mass/moisture sources, but other local effects are being investigated.

In chapter two, a geophysical survey was conducted using acoustic sub-bottom (CHIRP) seismic reflection techniques, and 24 lines covering 40 km were processed and interpreted. Four distinct seismic units were interpreted, spanning from the Pleistocene-aged deglacial sediments to modern day estuarine sediments. A basal till/bedrock surface marks the limit of seismic penetration, and overlying thick (up to 42 m) varved proglacial lake sediments suggest that the area was part of Glacial Lake Narragansett (GLN). Paleochannel stream cuts unconformably lie above the proglacial lake sediments, indicating a draining of GLN and a period of subaerial exposure in which an organized tributary system existed in Greenwich Bay. A GIS-based Local Polynomial Interpolation model was used to create a representative surface of the stream-dissected paleolandscape that existed prior to marine inundation. While useful, denser CHIRP coverage (100 m line spacing minimum) is recommended in order to take stress off of the interpolation. The proximity of channels to archaeological finds at Cedar Tree Beach would suggest that they might have been important resources for ancient inhabitants. Comparisons to the USGS East Greenwich Quadrangle Surficial Geology Map (Smith, 1955) shows significant kame terrace and other glacial meltwater deposits that may indicate a paleo-drainage of the modern day Pawtuxet river into the paleolandscape observed in the Greenwich Bay seismic record.

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about \$5 six-packs of Narragansett tall-boys, and they absolutely helped me throughout my time here!

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CHAPTER 1

A Paleoenvironmental Reference Record from Lacustrine Sediments at Gorton

Pond, Warwick, RI

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Abstract

A single sediment core from Gorton Pond, Warwick, RI was taken in the summer of 2013 for the purpose of developing a regional paleoenvironmental reference record. The core was dated using both AMS-radiocarbon dating of four terrestrial macrofossil samples, and correlations with a radiometric/pollen age model from a surface core taken from Gorton Pond for a previous study. The oldest terrestrial macrofossil was a spruce cone dated to ~12,100 calendar years BP, but an estimated 18,000 years of sediment is in the Gorton Pond record, based on bracketing ages from dated recessional moraines in the region. We sampled approximately 16,000 years of record, estimated based on a Younger dryas onset of 13,000 years.

Physical, magnetic, and elemental/isotopic (C, N, S, D/H) proxy measurements were made, and six distinct environmental zones were interpreted. The Gorton Pond record indicates significant changes in lake productivity and temperature. The Late Pleistocene Deglacial Period (16.0–13.0 kyr BP) is indicated by low OC%, high concentrations of magnetic material, and terrestrially-sourced organic matter. The Younger Dryas chronozone was observed (13.0–11.8 ky BP) as large decreases in $\delta^{13}\text{C}$, OC%, and $\delta\text{D}_{\text{BA}}$, indicating cold conditions. A transition from the Late Pleistocene to Early Holocene (11.8–9.6 ky BP) was observed by rapid warming and increased aquatic productivity. The Middle Holocene (9.6–3.6 ky BP) was characterized by depleted $\delta\text{D}_{\text{BA}}$ ratios, which may suggest Gorton Pond is sensitive to changes in air mass/moisture sources, but other local effects are being investigated. The Late Holocene is characterized by rising lake levels and a possible transition to cyanobacteria algal domination, based on nitrogen isotope ratio decreases.

Introduction

1.1.1. Lacustrine Sediments as Paleoenvironmental Records

The analysis of sediment core records has long been a fundamental basis for geologist's understanding of the Earth's dynamic history. The funding of large programs such as the Ocean Drilling Project has been an indicator of the importance of the retrieval and study of long-term (10^4 - 10^8 years) marine sediment records, and much of what we know today about large-scale paleoenvironmental change has come from these proceedings. However, due to low-sedimentation rates and geographic location, marine records can be limited in temporal and spatial resolution. Other depositional environments must be consulted, especially for the study of relatively recent Earth history (i.e. the Late Quaternary period).

Lacustrine (lake) sediments have proven to be excellent high-resolution archives of paleoenvironmental change when a multi-proxy approach is used (Cohen, 2003). Paleolimnology is a large, multi-disciplinary branch of geology that involves the use of physical, chemical, and biological proxy data in order to extract information from the sediment record. Lacustrine records are important tools for scientists for many reasons.

First, lakes are widespread and can provide archives of local and/or regional changes at all latitudes, from the sub-glacial polar regions (Seigert, 2005), to the tropics (Haberyan and Heckey, 1987) to the temperate regions in between (Menking et al., 2012). Because of this global distribution, long-term lacustrine records can provide a means to interpret continental-scale change that cannot be attained by focusing solely on marine records.

Secondly, lake genesis differs drastically, and therefore a range of sediment record lengths are available depending on the basin. Tectonically-derived lakes such as Russia's Lake Baikal (Williams et al., 1997; Kashiwaya et al., 2001), provide some of the longest (10^7 years), most robust records of continental environmental change on Earth, while at a finer scale, coastal lakes such as those found on the United State's east coast barrier beach systems have been used to develop paleo-storm frequencies at higher (10^0 - 10^2 yr) resolutions, albeit at shorter (10^3 yr) time intervals (Scillepi and Donnelly, 2007). Glacially-derived lakes such as the Finger Lakes in western New York have been extensively studied (Mullins and Hinchey, 1989, Ellis et al., 2004) in order to better understand paleoclimatic changes associated with the deglaciation of the last (Wisconsinan) ice age. These glacially-derived lacustrine records are of special interest and relevance to this thesis because of their usefulness in interpreting paleoenvironmental changes during the Late Pleistocene and Holocene epochs.

Lastly, the depositional environment of an ideal lake locality is one of low-energy and relatively rapid, continuous sedimentation. Moreover, because the erosional forces in deep, tectonically-inactive lakes are relatively weak as compared to other depositional environments (estuarine, fluvial, eolian), the preservation potential of lacustrine sediments tend to be much greater. Organic-rich lake sediments in particular provide useful paleoenvironmental proxy information for a wide array of geoscience research, especially in the context of deglacial-period environmental change in New England. This research includes major changes in the biosphere (Deevey 1939; David 1968), hydrosphere (Kirby et al., 2002; Menking et al., 2012), atmosphere (Hubeny et al., 2011), and cryosphere (Mullins and Hinchey, 1989).

Evidence for the reliability and reproducibility of lake-sediment derived paleoenvironmental change is robust. Here, we use a multi-proxy approach on lacustrine sediments from Gorton Pond, Warwick, Rhode Island in order to develop a regional environmental reference record for the time period spanning the Late Pleistocene to the present.

1.1.2 Study Site Characteristics

Gorton Pond is located in the highly urbanized region of Warwick, Kent County, Rhode Island (Figure 1.1.). It is a natural-basin pond covering 62 acres, has an average depth of approximately 4.9 meters, and a maximum depth of 13.7 meters (Guthrie and Stolgitis, 1987). The pond has suffered significant environmental degradation over the past several decades, and previous studies in 1955 (Guthrie and Stolgitis, 1987) and 1981-82 (Keyes Associates, 1982) have detected late summer anoxia in its bottom waters. Moreover, an unpublished study by King et al. (1993) showed that significant trace metal contamination and cultural eutrophication has occurred at Gorton Pond as a direct result of anthropogenic watershed changes and urban-development since 1850 A.D. A history of recorded land-use changes in the Gorton Pond watershed (from King et al. 1993, unpublished) can be found in Table 1.

Despite the negative effects of anthropogenic influences in the most recent sediments, Gorton Pond shows ideal characteristics for a paleolimnological study. Primarily, as a kettle-derived water body, Gorton Pond could potentially contain continuous paleoenvironmental information dating back to initial deglaciation of the region. Furthermore, given this type of basin genesis, it is unlikely that significant

erosion has occurred at Gorton Pond, and lengthy sedimentary hiatuses are not anticipated. Lastly, the pond is proximal to other important paleoenvironmental studies (Davis, 1969; Newby et al., 2000; Kirby et al., 2002; Ellis et al., 2004; Shuman et al., 2004; Hubeny, 2006; Oswald et al., 2010; Zhao et al., 2010; Menking et al., 2012) that can be valuable resources for regional correlation. It is for these reasons that I developed these three initial hypotheses concerning Gorton Pond:

- 1.) Because it is a kettle pond generated by the calving of glacial ice, the sediments of Gorton Pond will have begun to be deposited within centuries following the retreat of the Laurentide Ice Sheet, and therefore will span the entire time period following deglaciation;
- 2.) The sediments of Gorton Pond will be important recorders of local and regional paleoenvironmental change since deglaciation, and;
- 3.) By using a multi-proxy approach, we will be able to generate an environmental reference record based on interpretations of the physical, chemical, and magnetic character of the sediment archive at Gorton Pond.

1.1.3 Geological Context

The study sites in both chapters (Gorton Pond and Greenwich Bay, respectively) of this thesis lie within the United States Geological Survey (USGS) East Greenwich Quadrangle (Lat: 41.7500° to 41.6250° N; Long: -71.5000° to -71.3750° W). The East Greenwich Quadrangle covers about 119 km² area, much of which is occupied by Greenwich Bay and western Narragansett Bay. Maps of both the bedrock geology (A.W. Quinn, 1952) and surficial geology (J.H. Smith, 1955) of this 7.5-minute quadrangle provide important background information.

The bedrock geology of the East Greenwich Quadrangle is Esmond-Dedham (E-D) subterranean material (Hermes et al., 1994), and spans from the Pre-Cambrian-era Blackstone series to the Carboniferous East Greenwich Group (Mississippian) and Narragansett Bay Group (NBG) (Pennsylvanian). The Gorton Pond basin lies completely within the NBG Rhode Island formation (RIF), which consists of gray to dark gray, variable to irregular bedded sandstones, shales, and conglomerates, and has a predominant quartz, feldspar, biotite, and muscovite mineralogy (A.W. Quinn, 1952). The RIF is the thickest and most widespread formation in the East Greenwich Quadrangle, although the contact with the older East Greenwich Group Cowesett granite is less than a kilometer to the west of Gorton Pond. Bedrock elevations are greater in the East Greenwich Group than in the Rhode Island formation, and Late Quaternary glacial processes have resulted in the deposition of sand and till-rich ground moraine material in bedrock elevation highs at Drum Rock Hill and Bald Hill, and sand-rich outwash material in bedrock elevation lows (J.H. Smith, 1955).

The bedrock geology of the East Greenwich Quad underlies thick deposits of unconsolidated Quaternary-age sediment cover. These deposits consist largely of Late Pleistocene ground moraines and sequential outwash deposits associated with the retreat of the Laurentide Ice Sheet (LIS), although recent swamp, alluvium and shoreline deposits exist (J.H. Smith, 1955). Regionally, Gorton Pond lies within the well-sorted sand facies associated with Pleistocene glacial outwash deposits. However, significant lenses of undifferentiated sand and gravel surround Gorton Pond's western shoreline. The presence of these deposits as well as kame terrace and ice-channel deposits suggests that the area surrounding Gorton Pond was once part of a local or regional glacial drainage pattern.

Gorton Pond is a kettle pond. Kettles form from the melting of calved ice-blocks, or from the collapse of marginal ice masses buried beneath glacially derived sediments (Maizels, 1995), and are common within glacial outwash plains. Because kettles are depressions resulting from ice-block loading then melting, they tend to have characteristic "bulls-eye" bathymetries (Figure 1.2), with the deepest lake depths corresponding to the specific area of ice-block deposition. Kettle holes that accumulate some volume of water are classified as kettle lakes (or ponds), and their basal sediments are typically a glacial sand facies associated with outwash deposits. During sediment core analysis, it is generally assumed that a basal sand lithology demarcates the initial genesis of the kettle.

1.1.4 The Deglaciation Chronology of Southern New England

Much of the Quaternary period of Earth history has been punctuated by recurring, cyclical glacial “ice ages” thought to be caused by changes in incident solar radiation (insolation) and the resulting feedback mechanisms. These insolation changes are initiated by variations in planetary orbit, axis tilt, and precession of the equinoxes (Hays et al., 1976). The latest glaciation (Wisconsinan) spanned from ~75,000 years BP to ~11,000 years BP, and the term Last Glacial Maximum (LGM) is used to describe the southernmost extent of the Laurentide Ice Sheet (LIS) in North America.

Fortunately, the subject of the LIS history has been a focus for many branches of the geosciences over the past several decades. In the past decade specifically, there has been an increased interest in the post-LGM retreat of the LIS in New England (Boothroyd and Sirkin, 2002; Balco et al., 2002; Balco and Shafer, 2006; Boothroyd and August, 2008; Oakley, 2012; Ridge et al., 2012; Oakley and Boothroyd, 2013). A combination of two general approaches characterizes this type of research: the analyses and correlation of varved sediments, and/or the mapping and dating of recessional moraines. Varved sediments provide high-resolution (annual) archives of deglaciation that can provide paleoenvironmental proxy data and be correlated to the New England Varve Chronology (NEVC) (Antevs, 1922; Antevs, 1928; Ridge, 2011). Glacial moraines are large (2–10km wide; 30–200km long; 50–100m high) ice-marginal deposits that indicate periods of glacial advance (Balco et al., 2002). Traditionally, absolute ages of moraines have been determined by radiocarbon dating of organic material proximal to moraine localities (Stone and Borns, 1986; Boothroyd,

1998); however, the rarity of organic matter prior to ~15,000 calibrated years before present (cal. yrs BP) has made early deglaciation age control difficult through ^{14}C -dating alone. Recently, cosmogenic-nuclide exposure methodologies (Gosse and Phillips, 2001) have allowed for direct dating of morainal sediments (Balco and Shafer, 2006).

While the general spatial and temporal extent of the Late Wisconsinan LIS is well documented, areas of debate still exist, particularly in the time domain. Within the literature, the southern New England deglacial history can vary in absolute age precision by thousands of years, as interpretations are based on inherent assumptions, precision of methodologies, and robustness of datasets, among other factors. For example, in the context of our study region (mid to north-west Narragansett Bay), Boothroyd and August (2008) suggest that the LIS margin was north of Narragansett Bay by ~17,500 cal. yrs BP, but more recent work by Oakley and Boothroyd (2013) points to local deglaciation prior to 20,000 cal. yrs BP. Oakley's work was based around varve reconstructions in the Providence River, but the sediments there could not be correlated to the NEVC. For the purpose of this thesis, the deglaciation chronology of Balco and Shafer (2006) is preferred. By summarizing and adding to the existing regional cosmogenic-exposure data, and by reevaluating the precision of the two methodologies used in the interpretation (Be-10 and varve-year-calendar-year discrepancies) they were able to correlate absolute ages in morainal sediments to the NEVC. Their findings are generally well accepted within the scientific community, and their deglacial history will be referenced (Figure 1.3) to determine a bracketed basal age for Gorton Pond.

The Wisconsin ice age culminated approximately 25,000 cal. yrs BP, and net glacial retreat followed this last glacial maximum. By 24,000 cal. yrs BP, the LIS margin had retreated to coastal RI and deposited the Martha's Vineyard/Ronkonkoma moraine system before retreating to the Buzzards Bay/ Charlestown/ Fishers Island/ Harbor Hill recessional moraine by 19,200 cal. yrs BP. Continued recession brought the margin to the Old Saybrook moraine by 18,790 cal. yrs BP, and then the Ledyard moraine by 18,740 cal. yrs BP. Glacial varve deposition began in the Lower Quinnipiac Valley and in Glacial Lake Hitchcock (Rocky Hill Dam locality) at minimum dates of 18,400 cal. yrs BP and 17,800 cal. yrs BP, respectively. If one were to assume temporal synchronicity between the Old Saybrook and Wolf Rocks locations, and that the minimum date at Rocky Hill Dam is accurate, then deglaciation at Gorton Pond can be bracketed by 18,790 cal. yrs BP at the oldest and 17,800 cal. yrs BP at the youngest (Figure 1.3).

Methods

A note on reported ages: all radiocarbon dates referenced throughout this thesis are calibrated dates before present. In the following sections, ages will be reported in the format of years BP (i.e. 18,000 BP), and can be taken as true calendar years

1.2.1 Sediment Core Collection

The sediment core collection at Gorton Pond was completed in summer 2013, using a pontoon boat as a coring platform. Each core used in this study was taken from the deepest hole (41.704126°N, 071.45638°W) at Gorton Pond with the exception of GPRI90-1, which was taken roughly 100 meters from our coring location (Figure 1.2). Water depth at the deep hole was approximately 13 meters during each coring expedition. Composite core GP13 is composed of 2 different sediment cores (Table 2). Details of the basic sediment core collection and extrusion methods used for this study can be found in Glew et al. (2001).

Sediment core GP13BC1 was taken with a biological coring system composed of an 8-foot (2.43m) polycarbonate core liner, square threaded pushrods, and an internal piston. This system was used to retrieve the flocculent sediments that comprise Gorton Pond's sediment/water interface. This method worked well in the field, and the core was strapped upright on the boat and allowed to settle before being cut down and capped.

The longest record retrieved from Gorton Pond is GP13LC2, which was taken with a large-diameter Livingstone-type piston corer. This corer consists of a steel core barrel (in lieu of core liners) and an internal piston and cable, and is pushed via 10-foot fiberglass rods. Coring is completed through multiple 1-meter drives down the same core hole, and the collected sediments associated with each drive are extruded

from the corer on site. In order to insure that each drive was taken from the same location, 45-feet of aluminum casing was lowered through the water column and driven approximately 1 meter into the sediment. Coring was then accomplished by lowering the corer through the casing, pushing to the desired depth interval, retrieving the corer and extruding the sample into liners on deck, and then repeating the process for the subsequent interval. GP13LC2 consists of six drives (D1-D6); however, the first drive (D1) was lost in the field.

Our sediment cores were correlated with freeze core GPRI90-1. The freeze-coring system is capable of retrieving ~1.5m cores from the uppermost sediments in a lake. The freeze-coring apparatus is a hollow, aluminum rectangle that is filled with a dry ice and methanol solution, which cools the corer to approximately -80°C. At this point, the corer is pushed into the sediment surface vertically, and left stationary for 20 minutes. Upon retrieval, the corer is covered with frozen sediment. This system is especially useful for retrieving undisturbed, *in situ* surface sediments.

1.2.2 Core splitting, imaging, and logging

All sediment cores taken for this study were transported to the University of Rhode Island for physical analyses and storage. All cores were halved on a specially designed core splitter, which incorporates track-driven routers to cut the liners. Each core section was separated into a working half designated for imaging, logging, and sub-sampling, and an archive half designated for later analyses when material from the working half was exhausted. All core material is stored in walk-in refrigerators at URI's Rock and Core Repository.

After splitting, manual initial core descriptions (ICD) were documented, and cores were then cleaned and imaged using the line-scanner digital camera attachment on our Geotek® multi-sensor core-logger (MSCL) (Zolitschka et al., 2001). Gamma-ray attenuation and porosity evaluator (GRAPE), magnetic susceptibility, resistivity, and compressional wave (P-wave) sensors are housed on the track-driven MSCL, and provided valuable, non-destructive physical proxy measurements at one-cm resolution for each core.

1.2.3 Paleomagnetic proxy methods

Changes in the Earth's magnetic field can be characterized by three main types of variations in space and time: (1) geomagnetic secular variation, (2) excursions, and (3) reversals. These three types of variations are defined based on amplitude, direction, and duration of change (Jacobs 1994), and each are valuable tools for dating and correlation of sediments. Changes on scales of 10^2 or 10^3 years are found in the sediments of rapidly deposited environments (Thompson and Oldfield, 1986), and Mackareth (1971) showed that organic-rich lake muds have the potential to carry continuous paleomagnetic records for tens of thousands of years. Therefore, the sediments at Gorton Pond have the potential to contain valuable paleomagnetic information.

Following the non-destructive analyses of core logging and imaging, the working halves of each core was subsampled into U-channels for paleomagnetic analyses. A U-channel is a 1 cm^3 rectangular plastic tube that is cut to core section length and pressed into the sediment face. U-channels samples are taken from the

center of the core sections in order to minimize potential sediment fabric deformation that may occur along the walls of the polycarbonate liner during the coring process. Once in place, tools are then used to remove the U-channel with entrapped sediment. This method of subsampling works well with relatively consolidated sediments, but can prove difficult with flocculent sediments.

Paleomagnetic measurements were made using a 2-G Enterprises 755R pass-through superconducting cryogenic magnetometer and alternating field (AF) demagnetizer. This automated system measures high-resolution (1 cm) paleomagnetic and physical properties of U-channel samples (Weeks et al., 1993). Each core section taken at Gorton Pond was analyzed for natural remanent magnetization (NRM) in each of three main axes. Progressive demagnetization steps were performed with a diminishing alternating field (AF) in order to “clean” the sample from magnetic overprints that can occur during coring and subsampling. Demagnetization steps (Oe units) of 0, 25, 50, 100, 150, 200, 250, 300, 400, 500, and 600 were performed in between each subsequent pass through the magnetometer.

In addition to NRM, anhysteretic remanent magnetization (ARM) and isothermal remanent magnetization (IRM) were also imparted. ARM was imparted on each sample with a constant DC field of 1 Oe in the presence of a decreasing alternating field. IRM was imparted at room temperature under a 1.2T DC magnetic field. Both remanences were measured on a 2-G cryogenic magnetometer.

1.2.4 Elemental/Isotopic Proxy Methods

Lacustrine sedimentary organic matter can record changes in paleoenvironments, climate regimes, and anthropogenic influences (Meyers and Ishiwatari, 1993). Gorton Pond composite core GP13 was subsampled for loss-on-ignition analysis, elemental and stable isotopic concentrations, and a compound-specific paleotemperature proxy.

Loss-on-ignition is a quasi-quantitative method for determining dry bulk density, water content, and organic matter concentration of sediments (Dean, 1974; Hierl et al., 2001). Sediment subsamples (1 cm³) were taken at 2-cm resolution down composite core GP13, put into ceramic crucibles, and placed in a muffle-furnace overnight at 100°C to burn off pore water. The portion of mass lost was recorded, and sediment crucibles were then returned to the muffle-furnace and burned for one hour at 550°C. Organic matter begins to burn at 200°C and is fully combusted by 550°C (Dean 1974), and the portion of mass lost during this step (loss-on-ignition, or LOI) is equal to the concentration (%) organic matter by:

$$LOI_{550} = ((DW_{100} - DW_{550}) / DW_{100}) * 100\%$$

where LOI₅₅₀ is the bulk organic matter percentage and DW₅₅₀ the mass of the sediment after burning at 550°C for an hour. Organic carbon content was estimated as being 44% of the total LOI mass (Dean, 1974). Although the precision of the loss-on-ignition method has been called into question (Hierl et al., 2001), it is still an important quasi-quantitative measurement that can be performed with basic equipment available in most laboratories.

Composite core GP13 sediment subsamples (1cm^3) were taken at 8-centimeter resolution for elemental and stable isotopic analyses. Elemental compositions and stable isotope ratios were carried out on a coupled Elementar Vario Micro CHNOS Elemental Analyzer/Isoprime continuous-flow isotope-ratio mass spectrometer (EA/IRMS). Stable isotope ratios of carbon ($\delta^{13}\text{C}$), nitrogen ($\delta^{15}\text{N}$), and sulfur ($\delta^{34}\text{S}$) and their elemental compositions were utilized in this study. Prior to running samples, a sulfanilamide standard (Table 1.2) was run 3 times through the elemental analyzer, and a one-point calibration curve was generated for a daily correction factor. The daily Qa/Qc factor had to fall between 95% and 105% of the sulfanilamide standard in order for the run to be continued. Bracketing standards were also used for stable isotope determination. USGS-40 and USGS-41 were used as the respective low and high standards for both carbon and nitrogen isotopes, and the sulfur isotope bracketing standards were IAEA S-2 and IAEA S-3. Prior to running any samples, triplicate runs of these aforementioned IRMS standards were run, and the mean values are used to generate a two-point calibration curve used for later correction. Instrumental organic matter analyses were performed via two separate EA/IRMS “modes”, each involving a detailed preparation method.

Carbon isotope mode (C-mode) provides concentrations for sediment total organic carbon content (TOC) and the mass ratio of carbon isotopes $^{13}\text{C}/^{12}\text{C}$. Dried sediment samples were homogenized via mortar and pestle, and ~ 1 mg was placed in silver capsules. These samples were acidified via the fumigation technique of Harris et al. (2001) in order to remove any inorganic carbon present as carbonates. This technique is completed by placing the silver capsules in an assay tray, wetting each

sample with deionized water, and placing the open tray into a dessicator with 12M hydrochloric acid (HCL) solution for 6 hours. The carbonate is removed as carbon dioxide gas, and samples are then placed in a drying oven at 60°C overnight. After this fumigation process was completed and samples dry, the silver capsules were placed in tin capsules, and 3 times the sample mass of catalyst tungsten trioxide (WO₃) was added to assist in sample combustion. The isotopic composition of carbon samples are reported as $\delta^{13}\text{C}$, where:

$$\delta^{13}\text{C} = \left[\frac{(^{13}\text{C}/^{12}\text{C})_{\text{sample}}}{(^{13}\text{C}/^{12}\text{C})_{\text{standard}}} - 1 \right] \times 1000$$

$\delta^{13}\text{C}$ values are expressed as permille (‰) relative to the Vienna Pee Dee Belemnite (VPDB) standard.

Nitrogen and sulfur isotope mode (NS-mode) provides elemental concentrations for nitrogen and sulfur, as well as mass ratios for $^{15}\text{N}/^{14}\text{N}$ and $^{34}\text{S}/^{32}\text{S}$ isotopes. Dried samples were homogenized via mortar and pestle, and ~8 mg of sediment was placed in tin capsules, compressed to remove residual atmospheric nitrogen, and run through the EA/IRMS. The isotopic compositions of nitrogen and sulfur isotopes are reported as $\delta^{15}\text{N}$ and $\delta^{34}\text{S}$, respectively, where:

$$\delta^{15}\text{N} = \left[\frac{(^{15}\text{N}/^{14}\text{N})_{\text{sample}}}{(^{15}\text{N}/^{14}\text{N})_{\text{standard}}} - 1 \right] \times 1000$$

and

$$\delta^{34}S = \left[\frac{(^{34}S/^{32}S)_{\text{sample}}}{(^{34}S/^{32}S)_{\text{standard}}} - 1 \right] \times 1000$$

Both $\delta^{15}\text{N}$ and $\delta^{34}\text{S}$ values are expressed as *permille* (‰) with respect to AIR for nitrogen and the Vienna Canyon Diablo Triolite (VCDT) for sulfur.

In addition to isotope ratios, organic carbon to nitrogen (OC/N) and organic carbon to sulfur (OC/S) ratios were calculated from elemental compositions, and are dimensionless values.

Isotopic analysis of compound-specific constituents of sediment organic matter is a relatively new but promising method within the paleoclimate realm (Huang et al., 2002; Hou et al. 2006; Hou et al., 2007). Deuterium/hydrogen ratios of medium-chain C_{22} behenic acid (*n*-acid) were measured by Pamela Wagener at Brown University using the method of Hou et al. (2006).

Gorton Pond sediments were subsampled at the University of Rhode Island and transported to Brown University, where they were freeze-dried for three days and then ground to a fine powder with a mortar and pestle. Lipids were extracted using an Accelerated Solvent Extractor 200 (Dionex). Acids were isolated and methylated, and fatty acid methyl esters (FAMES) were extracted. FAMES were purified by elution through silica columns with dichloromethane. Procedural blanks were prepared for the *n*-acid samples in order to test for laboratory contamination with every batch of extracted samples. No significant contamination was determined.

Quantification and identification of compounds was carried out using GC and GC/MS, respectively. Hydrogen isotope analysis of behenic acid C_{22} was performed using an HP 6890 GC interfaced to a Finnigan MAT Delta+ XL mass spectrometer through a high-temperature pyrolysis reactor. Six pulses of hydrogen reference gas

with a known δD value (checked manually once every other day) were injected via interface to the IRMS, for the computation of δD leaf wax values of sample compounds relative to Vienna Standard Mean Ocean Water (VSMOW). A fatty acid methyl ester (FAME) standard with a known δD value was run frequently throughout the analyses, and a long-term average C_{22} δD standard value of (-180‰) was used to correct for drift in the standard. Each sample was analyzed in duplicate, and standard deviations were used as an indication of error.

1.2.5 Age Constraints

Age control for Gorton Pond was achieved through a multi-disciplinary approach (Table 1.2) utilizing accelerator mass spectrometer (AMS) radiocarbon dating and correlations to a separate basin core with a $^{210}\text{Pb}/^{137}\text{Cs}$ radiometric decay dataset and known pollen horizon (King et al., 1993, unpublished report).

Four terrestrial macrofossil samples (two leaves, one twig, and one cone) were sent to Beta Analytic, Inc. for AMS radiocarbon dating. An acid-alkali-acid sample preparation procedure was used in order to remove carbonates (initial acid), humic acids (alkali base), and any addition of atmospheric carbon (final acid) during initial acidification. The reference standard used by Beta is 95% of the radiocarbon activity of Oxalic Acid (National Institute of Standards and Technology standard SRM 4990C), and calculations were based on the Libby ^{14}C half-life of 5568 years. Quoted errors represent 1 standard deviation (68% probability) counting errors based on the combined measurements of the sample, background, and modern reference level. Results were provided as conventional radiocarbon ages (in ^{14}C years before present;

“present” being 1950 A.D.) and were calibrated using the Calib 6.0 freeware program and Intcal09 dataset (Reimer et al., 2009).

Dr. Peter Appleby of the University of Liverpool created the radiometric ($^{210}\text{Pb}/^{137}\text{Cs}$) chronology for a previous study in 1993. Gorton Pond freeze core GPRI-90 was measured non-destructively for ^{210}Pb , ^{226}Ra , ^{137}Cs , and ^{241}Am by direct gamma assay (Appleby et al., 1986). ^{210}Pb dates were calculated using the CRS model of Appleby and Oldfield (1978). Well defined ^{137}Cs and ^{241}Am peaks occurred at some sites downcore and were assumed to be a recorder of atmospheric fallout from testing of nuclear weapons, and were assigned a date of 1963 A.D..

In addition to the aforementioned radiometric dating of GPRI-90, pollen analyses were prepared using a standard chemical extraction procedure, and pollen slides were counted with standard light microscopy methods (Faegri and Iversen, 1975). The *Ambrosia* horizon, long used as a proxy for European settlement, was identified and both pollen and radiometric dating was used to create the final age model for sediment core GPRI-90.

In order to attain this radiometric decay/pollen age model and for the most recent sediments at Gorton Pond, the upper section of composite core GP13 was correlated to surface freeze core (GPRI-90) using physical and chemical proxy measurements (Figure 1.4). Comparisons of magnetic susceptibility, carbon %, and nitrogen % suggest that approximately 14 centimeters of sedimentation has occurred at Gorton Pond between 1990 and 2013. Freeze core GPRI-90 was shifted down 14cm and the age control points understood to be 23 years older than reported in the previous unpublished study by King et al. (1993).

1.2.6 Spectral Analysis

Spectral analysis was calculated using K-Spectra software. Analyses were done on the organic carbon time-series after resampling the data at a 75.5-year average spacing in the Analyseries 2.0.4.2 software program. Spectral analysis was performed using a multi-taper method with 3 tapers. Confidence intervals were calculated assuming a red-noise first-order autoregressive (AR(1)) background (Mann and Lees, 1996).

Results

1.3.1. Paleomagnetic Results

Mackareth (1971) showed that organic-rich lake sediments have the potential to carry continuous paleomagnetic records for tens of thousands of years. However, the sediments at Gorton Pond did not show a continuous, valuable natural remanent magnetization record for much of the record. Correlations to the Northeast Regional Inclination Curve of King and Peck (2001) could not be made. Sediment core deformation during the extrusion process could have disturbed the magnetic fabric of sediment core GP13, and rendered the paleo-inclination signal useless.

Environmental magnetic results can be found in Figure 1.5. Magnetic susceptibility, anhysteretic remanent magnetization, and isothermal remanent magnetization are useful proxies for concentration of magnetic material. Susceptibility in lake systems can be thought of as an indication of magnetite concentration (Cohen 2004), ARM is biased toward fine-grained concentration, and IRM is biased toward coarse-grained magnetic material. NRM and ARM median destructive fields (MDF), as well as ARM(30)/ARM(0) are indicators of magnetic grain size. Higher values indicate finer magnetic grains, and lower values indicate coarser. Hard isothermal remanent magnetization (HIRM) and the S-ratio are both indicators of changes in the magnetic mineral assemblages.

1.3.2 Lithology and Physical Properties

A Geotek® multi-sensor core logger (MSCL) was used to image and log the core, and plots of magnetic susceptibility (K) and gamma-ray attenuation and porosity

evaluator (GRAPE) density measurements are shown in Figure 1.6. Magnetic susceptibility is a quantification of the bulk volume of magnetic mineral content in a sediment sample, and it is measured directly by subjecting sediments to a low magnetic field to induce an observable magnetization that will disappear when the field is shut off (Nowaczyk, 2001). For the purpose of discussing susceptibility, values are expressed as whole-number SI units, but it should be noted that values should be multiplied by 10^{-5} in order to be actual SI units. Environmental changes are characterized by variations in weathering, erosion, transport, and deposition, and these mechanisms result in different sedimentation packages. The lithology, susceptibility, and density of these packages can tell us about changes in depositional environments.

The lowermost sediments at Gorton Pond (670 – 450 cm) consist of alternating beds of yellow-brown silts and blue-gray silty-clays. These sediments are depleted in organic material (although a single spruce seed was subsampled at 507 cm for AMS-dating), have a larger relative grain size (fine-medium silt) than the overlying material, and show characteristic laminations (1 – 2 mm in thickness) throughout the facies. Magnetic susceptibility values are elevated (~ 3 SI) at the base of the core (Figure 1.6) as a result of the poorly sorted and finely magnetic sediments associated with glacial deposits. Susceptibility shows a decreasing trend going up-section, dropping ~ 3 SI units until relative stability around 0 SI units. The GRAPE density largely reflects the shifts observed in the susceptibility, with the densest ($\sim 1.7 \text{ g/cm}^3$) sediments occurring at the base of the core and a decreasing trend moving up-section. The sediment density remains relatively stable between $1.3 - 1.4 \text{ g/cm}^3$ following the initial decreasing trend. These lowermost facies are interpreted as glacially derived terrestrial material

due to the color, grain size, lack of organic material, and elevated susceptibility values, and the presence of laminae suggests deposition may have been controlled by the freeze-thaw seasonal cycles characteristic of glaciated regions. Despite coring to failure with the Livingstone-type corer, we were not able to reach the sand material that typically marks the base of a glacially derived deposit. However, the presence of laminated clays at the base of our core suggest there is probably not much more material (<1 meter) in the post-glacial sediment record at Gorton Pond.

Overlying these terrestrially dominated silts and clays is a thick (449 – 115 cm), light brown to dark brown, organic-rich gyttja facies. Significant plant debris is observed throughout this facies, and the bulk of AMS-datable material was subsampled from hence. The transition from glacially derived material to that of a more modern lacustrine setting occurs gradually; a transitional gyttja exists from ~449 – 350 cm. These transitional sediments display variations in both proxies that may be a result of continued seasonal ice control over sediment availability. Susceptibility in particular increases by 1.5 SI from 440 – 420 cm, before a decreasing trend ~2.5 SI in magnitude occurs over the next meter. Above this transitional facies, susceptibility peaks (~0.5 SI) around 275 cm depth before reaching a stable range (~ -0.75 SI) before the core break at 192 cm. Sediment density decreases until ~240 cm depth, but increases to ~1.3 g/cm³ before the break.

The uppermost (192 – 0 cm) sediments at Gorton Pond are also organic gyttja, but a shift in color, susceptibility, and density occurs at ~40 cm depth. Above 40 cm the sediments are very similar to the underlying organic gyttja, although increased gas and water content result in a much more flocculent texture. Sediment susceptibility

values are comparable on either side of the void in the record, but density is lower and displays larger amplitude variations. Above 40 cm the susceptibility values of the sediment increase drastically to ~20 SI units. These uppermost sediments are indicative of those contaminated by anthropogenic sources, and do not reflect natural changes to the environment surrounding Gorton Pond.

1.3.3 Organic Matter Analysis and Stable Isotope Proxy Results

Organic matter constitutes a minor but important fraction of lake sediments, and provides proxy information that can be used to determine both natural and anthropogenic environmental changes (Meyers and Teranes, 2001). A variety of analyses were conducted on the sediments of Gorton Pond core GP13, including loss-on-ignition, elemental and stable isotopic concentrations for carbon, nitrogen, and sulfur, and also deuterium to hydrogen ratio measurements.

Organic carbon and $\delta^{13}\text{C}$

The results of the organic-matter based carbon analyses can be found in Figure 1.7. Organic carbon content is very low at the base of the core (~1%) but increases steadily to ~5% by 540 cm depth. This increase is mirrored by the carbon isotope ratio $\delta^{13}\text{C}$, which starts at very low values (-29.8 ‰) at the core base, but become less negative (~ -28 ‰) until 540 cm. There is a significant shift at this depth in both organic carbon content and $\delta^{13}\text{C}$, with OC% dropping by 2% and $\delta^{13}\text{C}$ ratios decreasing to less than 30‰. Values remain in these respective ranges until ~490 cm depth.

Above this depth, OC% shows an increasing trend until the core break at 192 cm depth. This is a ~9% increase, with pronounced OC peaks of 13.77%, 12.30%, and 14.50% at 412, 346, and 226 cm depth, respectively. Values above the core break are comparable to those below, but a decrease of ~8% is observable until ~35 cm depth, when OC% begins to increase to modern values. It should be noted that above ~350 cm depth the OC percentages calculated from LOI analyses differ from those calculated by elemental analyzer analysis. This difference may be due to the fact that organic carbon was estimated as 44% of OM (Dean, 1974), which may not be a precise estimate in all sediments in the core. However, the general pattern of OC deposition is recorded in both methods.

$\delta^{13}\text{C}$ values show a sharp increase to less negative values following the initial drop at 540 cm. Values reach a maximum of -24.44‰ at 390 cm depth, approximately 6‰ increase from the minimum values associated with the initial drop. The sediments from 390 – 258 cm show small amplitude variations, but generally stay in the same range of values, averaging -25.41‰. A sharp decrease to more negative values occurs from 258 cm until the core break at 192 cm. Values above the break are comparable to those just below, but show an increasing trend until 24 cm depth (-26.78‰) when ratios decrease to a modern value of -30.18‰.

Elemental nitrogen, $\delta^{15}\text{N}$, and OC/N ratios

Elemental nitrogen values were run on an elemental analyzer and can be found in Figure 1.8. Similarities of the elemental nitrogen trends with organic carbon trends would suggest that the %N signal is largely a function of organic matter preservation.

Nitrogen (N) percentages are very low ($\sim 0.1\%$) at the base core GP13, but increase to 0.49% by 553 cm depth. A rapid decrease to 0.37% occurs after this initial increase, and values remain near this value until 483 cm (0.41%) when a sharp increase to 1.15% occurs at 414 cm. A rapid decrease to 0.68% follows; however, a gradual, steady increase after this drop characterizes the curve until the core break at 192 cm. Significant peaks of 1.23% , and 1.57% occur along this increase at 348 cm and 208 cm, respectively. Values above the core break are comparable to those below, and a decreasing trend is observable until a minimum of 0.62% at 24 cm depth. N values increase during the anthropogenic era to modern values of 1.48% .

$\delta^{15}\text{N}$ ratios were measured using an isotope-ratio mass spectrometer and reported relative to AIR (0‰). These ratios show significant variability in the lowest core section, fluctuating between $\sim 0\text{‰}$ and $\sim 2\text{‰}$ until 569 cm depth, when an increasing trend starts from 0.28‰ and culminates at a maximum value of 2.74‰ at 505 cm depth. Ratios then decrease to 0.67‰ at 470 cm depth, before increasing again to 2.75‰ at 406 cm. The sediments between 406 cm and 242 cm have the most stable $\delta^{15}\text{N}$ values of the whole core, displaying a general increase from 2.75‰ to 3.09‰ over a 164 cm interval, with only small amplitude variations between. Values decrease sharply until the core break. Values above this break are comparable to those below, and appear to continue the decrease observed before the void in the record. The uppermost $\delta^{15}\text{N}$ values indicate a sharp increase of nitrogen to the sediment, suggesting significant anthropogenic influence, especially in the upper 60 cm.

Organic carbon to nitrogen ratios are useful for determining the source of organic matter to the sediment record, as the OC/N ratios of organic matter from algal

sources differ from those of vascular plant sources (Meyers and Teranes, 2001). By calculating the OC/N ratios of sedimentary organic matter from Gorton Pond, we could distinguish between aquatic sources versus terrestrial sources. In general, OC/N ratios of terrestrial plants are much higher (~20) than the ratios from aquatic algal sources (< 10).

Figure 1.9 shows the organic carbon content, OC/N mass ratios, and elemental nitrogen concentration for Gorton Pond core GP13. The base of the core shows elevated OC/N ratios (>20), but these values decrease to a minimum ratio of 9.5 at 594 cm depth. An increase to 12.6 culminates at 505 cm depth before a decrease back to 10.3 at 470 cm. Much like the OC and N% proxies, the most stable period in the history of the sediments occurs from ~440 – 220 cm depth, averaging an OC/N ratio just under 12, and experiencing only minor amplitude fluctuations. Values above and below the core break are slightly less than 12, but are comparable and seemingly reflect similar stability. Increases in the OC/N ratio start at ~60 cm depth, and are probably the result of anthropogenic alterations to the watershed. Clear-cutting of forested watershed areas and the development of impervious surfaces can lead to significant increases in surface run-off and therefore deliver more terrestrial sourced organic matter to the lake.

Elemental sulfur, $\delta^{34}\text{S}$, and OC/S ratios

Elemental sulfur (S%) was run simultaneously with elemental nitrogen on an elemental analyzer, and results can be found in Figure 1.10. Values for S% are very small (~0.2%) at the base of the core (Figure 1.10A.), and fluctuate around this range

until a minimum of 0.17% at 462 cm depth, averaging 0.31% for this depth interval. Following the minimum at 462 cm is a sharp increase to 0.66%, where %S begins to decline gradually again until a minimum of 0.30% at 340 cm. Higher amplitude variations are evident in the depth interval from 438 – 340 cm, but values average 0.43%, approximately 0.13% higher than the interval below the initial increase at 462 cm. The depth interval from 340 – 242 cm is characterized by elemental sulfur concentrations more similar to those in the lowermost meter of the sediment core. A slight increase is observed in the ~60 cm just below the core break, and values above the break are comparable to those just below, but exhibit a decreasing trend going up-core. The minimum S% concentration in the uppermost core section is 0.34% at 56 cm depth, and sediments overlying this increase to >2.50% in the most recent sediments.

The sulfur isotopic signature in core GP13 is very depleted in ^{34}S at the base, with values less than -4.5‰ (Figure 1.10B.). An enriching trend occurs until 561 centimeters below sediment surface, where a sharp decrease from 0.15 to -3.96‰ occurs. Values begin to trend to more positive $\delta^{34}\text{S}$ ratios over the next 80 cm, peaking at 4.50‰ at 470 cm. A smaller amplitude decrease from 4.50 – 2.37‰ is observable from 470 – 454 cm, concomitant with the shift from glacially-derived sediments to the transitional gyttja mentioned in previous sections.

$\delta^{34}\text{S}$ ratios in the middle of the core continue to show an enriching trend, with a core maximum peak of 12.50‰ at 242 cm. Depletion in the isotope record follows, and values below the core void fall within the 6 – 8‰ range. Ratios between the void and anthropogenically altered facies (112 – 64 cm) average 6.36‰.

The ratio of organic carbon to sulfur (OC/S) can also provide insight into lacustrine versus marine influences, water column stratification, and sulfate limitation (Cohen, 2003). OC/S ratios at the base of core GP13 are relatively low (~6.5) but show a steady increase to a ratio of 20.22 at 446 cm (Figure 1.11). An interval of high amplitude variability characterizes the region from 446 – 348 cm, but an increasing trend persists. Variations stabilize (excluding a sulfur outlier associated with a core break at 286 cm) until 274 cm, when fluctuations return until the sediment void. Excluding the outlier at 286 cm, OC/S ratios average 39.13 below the void in core GP13, and 39.58 in the “natural” signal above the void (112 – 48 cm). Elemental sulfur loading in the anthropogenic interval has caused a drop in OC/S ratios to core minimum values below 5.0.

Elemental sulfur can covary with organic carbon because it is found in organic matter. Therefore, changes in %S may be driven by preservation of organic matter. By looking at the ratio of OC/S, we can begin to understand changes in sulfur normalized to organic carbon. The increasing trend in the natural record at Gorton Pond indicates that the OC/S ratio at Gorton Pond is reflecting the increased OC values, as elemental sulfur percentages remain relatively constant.

1.3.4 Paleotemperature Proxy Results and Proposed Environmental Shifts

Deuterium-to-hydrogen ratios (δD) are sensitive to paleoenvironmental changes. Ambient temperatures, water availability and source, and plant growth rates all affect the fractionation of hydrogen isotopes by land plants (Meyers and Teranes, 2001), and therefore interpretations from these sources of organic matter are

complicated. However, because water is essentially unlimited to aquatic algae, and temperature variations in lakes are smaller than those on land, δD ratios of aquatic organic matter preserved in sediments primarily reflect to origin of lake water. D/H ratios can then be interpreted in terms of precipitation/evaporation balances and air mass trajectories, and surface air temperature can also be inferred (Hou et al., 2007).

Compound-specific D/H ratios of aquatic sources are of particular interest because of their ability to record hydrogen isotope ratios of lake water (Huang et al., 2002; Hou et al., 2006; Huang et al., 2007). Fatty-acid lipid compounds of organic matter are especially useful paleoclimate studies because they are relatively abundant in lakes and are relatively easy to extract and measure. Ficken et al. (2000) showed that medium-chain (C_{20} to C_{24}) fatty-acid lipids are especially abundant in submerged and emergent aquatic macrophytes, and therefore have high potential to record D/H ratios of lake water. Hou et al. (2006) showed a significant relationship between the δD ratios of medium-chain (C_{22}) behenic acid of lake surface sediments and their respective water δD values. This would suggest that δD_{BA} is an excellent recorder of changes in lake water D/H ratios, and therefore a useful proxy for paleoclimate changes.

Hou et al. (2007) showed a relationship between surface sediment δD_{BA} values and surface air temperature in 19 lakes, spanning a temperature gradient of 2–23°C. They were able to infer surface air temperatures (T) from δD_{BA} by using the following relationship:

$$\delta D_{BA} = 4.3T - 208.4$$

$$R^2 = 0.96, p < 0.001$$

Rearranging this function, temperature can be inferred by:

$$\frac{208.4 + \delta D_{BA}}{4.3} = T$$

Figure 1.12 shows the δD_{BA} variations, as well as the inferred temperature versus depth for core GP13. Error bars in the δD_{BA} plot were calculated as the standard deviation of duplicate mass spectrometer runs, and these errors were carried over in the temperature calculations. These results will be reported in the context of inferred temperature changes coincident with multiple proxy shift “zones” (Figure 1.13).

Lake water temperatures for the entire Gorton Pond average 10.9°C, but show significant variation throughout the history of the pond (Figure 1.12B.). Inferred temperature was moderately low (9.9°C) at the base of core GP13 but show an increase of 1.9° by 646 cm before falling to ~8.6° by 612 cm depth. Following this relatively cold interval is a rebound to a moderate average temperature of 11.0° from 595 – 528cmbss. Environment Zone A (Figure 1.13) is characterized by decreasing sediment susceptibility and density, a steady trend towards a less negative $\delta^{13}C$ signature, and a shift toward autochthonous organic matter sources as evidenced by the decreasing C/N ratio. Temperatures in Zone A are moderate, with evidence for a brief cooling early in the sediment record.

A drop of nearly 3° follows this moderately warm period, bringing lake surface temperatures to an Environmental Zone B minimum of 7.7°C. Unfortunately, the sediment sample taken from 511 cm could not be run due to saturation issues on the IRMS, so this temperature drop is characterized by a single data point. However, large

concomitant shifts in nearly every other proxy (Figure 1.13) provide evidence for our interpretation that this temperature decline is not an outlier.

A temperature peak of 13.1° follows this large drop, and marks the boundary between Zone B and Zone C. Warming is short-lived however, and values fall back to an average of 11° by 460 cm. Immediately after this slight cooling is the most rapid and highest magnitude warming observable in the entire record, with lake temperatures rising to an average of 15.1°C over the depth interval 460 – 409 cm. Coincident with this rapid warming are spikes in $\delta^{13}\text{C}$ and OC%, and increasing trends in both elemental nitrogen and sulfur percentages. Much like Zone B, the large changes observed in Zone C are characterized by rapid, large-scale changes that suggest significant alterations in the paleoenvironment.

Zone D can be associated with moderately cool temperatures, averaging 10.8°C, and a relatively stable proxy record. The coolest temperature in Zone D (8.8°C) occurs at 341 cm, and is associated with a small peak in organic carbon deposition. There is a slight leveling-off of magnetic susceptibility values at this depth, as well as a small amplitude decrease in $\delta^{13}\text{C}$ values. Slight increasing trends are observable in both $\delta^{15}\text{N}$ and $\delta^{34}\text{S}$ isotope ratios, but Zone D as a whole suggest a stable paleoenvironment absent of large amplitude proxy fluctuations.

Zone E is difficult to interpret due to the large amount of missing core material. This is unfortunate because large changes mark the boundary between Zone D and Zone E, particularly in the $\delta^{13}\text{C}$ and OC records. Inferred temperatures suggest that an overall cooling trend is occurring at the onset of Zone E, and δD values after the sediment void are similar to those before it. It appears as if this zone is probably

one of relative stability, with a much more negative $\delta^{13}\text{C}$ signature, decreasing $\delta^{15}\text{N}$ values, and lake record maximum values of $\delta^{34}\text{S}$.

The uppermost 60 cm of Gorton Pond composite core GP13 has been interpreted as Zone F. Temperatures appear to be very low in the first 25 centimeters of this Zone, averaging 8.0°C and recording a sediment core minimum temperature of 6.5° at 36 cm. Nevertheless, δD -inferred temperatures rise rapidly in the uppermost 35 centimeters, and average 12.6°C in the anthropogenic era. This value is comparable to a modern (2011 A.D.) mean annual air temperature of 11.4°C in Warwick, RI (National Climatic Data Center, 2013).

Paleoenvironmental changes coming from the isotopic analysis of organic matter can be difficult to interpret, largely because the sources of organic matter come from both terrestrial (allochthonous) and aquatic (autochthonous) sources. Differentiating the sources of organic matter is essential in paleoenvironmental studies, because organic matter provenance will help to constrain interpretations (Cohen, 2003). The amounts of sedimentary organic matter that originate from aquatic versus terrestrial sources can be distinguished by the characteristic OC/N ratio compositions of algae versus vascular plants (Meyers and Teranes, 2001). Figure 1.14 shows a plot of the $\delta^{13}\text{C}$ signatures of each environmental “zone” versus their respective atomic OC/N ratios. Atomic OC/N ratios were calculated by multiplying the mass ratios by the atomic weight of nitrogen and carbon (1.167). Domains and ranges for typical OM signatures of lacustrine algal, C_3 vascular land plants, and C_4 vascular land plants are demarcated in the plot.

Organic matter has been derived from a mixture of lacustrine algae and C₃ vascular land plants for much of the environmental history of Gorton Pond. One point early in Zone A (666 cm) shows a terrestrial-sourced signature, as well as one point in the anthropogenically-altered Zone F (24 cm). A subequal mixture of aquatic and terrestrial organic matter sources is expected for most lakes, but alterations to watersheds have been shown to increase OC/N ratios (Kaushal and Binford, 1999). It is possible that the increases in the OC/N ratio in both Zone A and Zone F are due to increased runoff from the watershed, although the mechanism causing this increase is inherently different. It is likely that runoff increases in the anthropogenic era are due to deforestation associated with agricultural techniques, while increases in runoff prior to European settlement may be due to varying vegetation regimes, increased precipitation events, an unstable watershed, or a mixture of any of these.

1.3.5 Age to Depth Model

The age model for Gorton Pond has been constructed using a variety of multi-disciplinary dating techniques (Table 1.3) from two different sediment cores taken from locations proximal to each other (Table 1.2). Although the sediment cores were taken over two decades apart, similarities in multiple proxies (Figure 1.4) provide evidence that these cores can be correlated, and that the previous radiometric age model from core GPRI-90 can be incorporated. Absolute ages from a variety of sources were plotted downcore in Figure 1.15 in order to generate an age-to-depth model. Due to complications involving the large void in the record from 115 – 192

cm, these data were broken up into 3 distinct domains for the purpose of estimating sedimentation rates.

The lower domain spans from the basal core depth of 670 centimeters to the lowermost age control point, a radiocarbon date, at 507 cm depth. A linear regression was used to develop the sedimentation rate within this depth interval, and is defined by the function:

$$y = 36.184x - 6243.3$$

The basal age of the core could not be constrained by absolute dating techniques, but an estimate of 18.79 – 17.80 ky BP was used based on the bracketing ages at the Rocky Hill Dam site (Ridge et al., 2012) and the Old Saybrook/Wolf Rocks moraine (Balco and Shafer, 2006) (Figure 1.3). Working on the assumption that Gorton Pond formed at the estimated time of deglaciation in the region, and adopting a conservative ice retreat rate of 30 meters per year (Oakley and Boothroyd, 2013), the time it would have taken for the ice margin to retreat the 23 kilometers from Wolf Rocks to Gorton pond is approximately 750 years. These calculations would suggest that a maximum possible basal age at Gorton Pond is 18,000 years BP. This calculation is also assuming that the entire sediment archive at Gorton Pond was cored, but this is not known to be valid. In any case, the ages associated with the lowermost domain should be studied with these assumptions in mind.

The depth interval from 507 – 212 cm consists of four calibrated ¹⁴C dates. Linear sedimentation rates were assumed between radiocarbon points, and the equations of linear regressions were used to convert depths to absolute ages. This interval in the core is arguably the best constrained, and spans the longest period of

Gorton Pond's history (~10,400 years). The bulk of the environmental change interpretations fall within this range of time, and these shifts can be converted to time with confidence.

The uppermost domain incorporates the top 211 centimeters, spanning from the radiocarbon date at 212 cm, through the void in the record, and terminating at the sediment-water interface. Age constraints include one radiocarbon dates, eight points from the GPRI-90 radiometric age model, the well-documented *Ambrosia* horizon, and the sediment/water interface. A 2nd-order best-fit polynomial regression was passed through the data in this domain in order to convert depth to age, and can be defined by the equation:

$$y = 0.285x^2 - 2.4531x$$

$$R^2 = 0.996$$

A separate trendline was used for this interval for the purpose of best characterizing the intervals in the core with strong age constraints. A significant change in the depth to age curve in the uppermost sediments is evident in core GP13, and this phenomenon is regularly seen in lake archives, as anthropogenic influences effect sedimentation rates, and compaction of material increases with depth. It is evident in the available record that a change in sedimentation rate occurred at some time in the past 2000 years. Unfortunately, the likely inflection point is within the large sediment void from 115 – 192 cm. Until a more complete record can be retrieved and used for correction, two separate regressions are used.

1.3.6 Spectral Analysis

The LOI organic carbon time series was used for spectral analysis due to the high-resolution sampling (2 centimeter) performed on the sediment core. However, due to core breaks associated with section changes and the large sediment void from 115 – 192 cm, significant gaps in the record exist. It is for this reason that we ran spectral analysis only in the interval with the most robust and complete age chronology: 507 – 212 cm. Based on the age-to-depth model, this would provide evidence of any significant organic matter periodicities occurring from approximately 12.1 – 1.8 kyr BP, or the bulk of the Holocene epoch.

Spectral analysis of organic matter content was performed using the K-Spectra software program. The time-series was resampled using the Analyseries 2.4.0.2 software program, and an average sampling interval of 75.5 years was used for spectral analysis. The multi-taper method was employed, using 3 tapers, and confidence intervals of 90%, 95%, and 99% were calculated based on a red-noise first order autoregressive (AR(1)) background (Mann and Lees, 1996). Significant (99% confidence) spectral peaks were observed in centennial (*ca.* 439-, 537-, 568-, 750-, 802-yr) periodicities (Figure 1.16).

Discussion

The environmental history as preserved in Gorton Pond has been categorized into six distinct chronozones based on varied proxy analysis and a multi-disciplinary age model. The record from the available Gorton Pond record is largely characterized by productivity, OM source, and temperature proxy changes. A chronozone-specific discussion of environmental characteristics, the timing of key changes, and comparisons with available literature will follow (Figure 1.17). Our Environmental “zone” interpretations will be investigated in the context of a number of other paleolimnological studies done in the region (Figure 1.18). It is our goal to investigate how Gorton Pond’s environmental history correlates with New England regional climatic changes (i.e. Shuman et al., 2004).

4.1 Environment A: Late Pleistocene Deglacial Period (16.0 kyr – 13.0 kyr BP)

The basal sediments at Gorton Pond indicate that the sediment core penetrated to a lithology characteristic of deglacial conditions. Laminated silts and clays point to a depositional environment controlled by seasonal glacial melt, but unfortunately a basal sand facies was not penetrated, which suggests that the entire deglacial package was not retrieved. Moreover, there is no age constraint from 507 – 670 cmbss, so a linear sedimentation rate was used in this depth interval, with an estimated basal age of 16.0 kyr BP. Bracketing ages established by previous work done in the region (Antevs, 1922, 1928; Balco and Shafer, 2006) suggest a maximum age of 18.0 kyr BP. However, it is very probably that the lowermost facies at this location is glacial sand, deposited as outwash deposits from proximal ice sources (Boothroyd, 1991; Hubeny,

2006). The fact that we did not retrieve this facies suggests that core GP13 is not complete. An estimate of 16,000 years was based on the shift from Environment A to Environment B. Environment B is interpreted as the Younger Dryas chronozone (YDC), and a basal age of 16.0 kyr BP would place the onset of the YDC at 13.0 kyr BP, consistent with other literature in the area (Hou et al., 2007).

Environment A spans from ~16.0 kyr BP until *ca.* 13.0 kyr BP. As deglaciation of the area occurred, shifts in regional vegetation and climate followed, and many studies have documented the erratic nature of the climate system during this period of significant change. The basic laminated silt/clay lithology of Environment A was probably deposited following retreat of glacial ice far enough out of the Gorton Pond region for limited deposition of sandy material. Magnetic susceptibility and OC/N values are elevated at the beginning of Zone A, indicating a terrestrially-dominated depositional environment. Magnetic ARM and IRM values suggest that the concentration of coarse, high coercivity magnetic mineral grains were abundant at the base of GP13 (Heil et al., 2009), representative of glacier derived material. Organic carbon content is very low at the base of the sediment record, suggesting low productivity, but shows an increasing trend throughout the Late Pleistocene record.

The terrestrially dominated OC/N is due to low values of both OC and N, and the overall deposition of organic matter is very small. Increasing $\delta^{34}\text{S}$ values suggest Gorton Pond was becoming more sulfate-limited, as terrestrial glacial material began to be exhausted, or stabilized via watershed vegetation growth. Increasing $\delta^{13}\text{C}$ values is interpreted as an indication of increased productivity in the lake, and a concomitant increase in the %N values during this time may be an indication of the forcing

mechanism behind the increase in productivity. Increases in %N may be due to increased runoff of nitrates from soils, and this increased delivery of nitrates to a lake system could result in increased algal productivity, which will then lead to a decrease in the supply of dissolved inorganic carbon (DIC) in the water column (Meyers and Teranes, 2001). A decrease in DIC will limit algal bias of ^{12}C and result in higher $\delta^{13}\text{C}$ values. (Meyers and Teranes, 2001, and references therein).

These proxies indicate a late Pleistocene deglacial environment. Palynological reconstructions at Rogers Lake in Southern CT (Davis 1969) suggest a dominance of tundra and herb pollen (N.E. pollen zone “T”) in the region from ~18.0 kyr BP until 14.6 kyr BP. Davis (1969) points to the overall lack of pollen grain abundance during this time interval, and compare this rarity in deposition of all pollen grains to records found in modern Arctic settings. This study is consistent with our interpretation of a late deglacial environment lacking terrestrial vegetation, as increased susceptibility and density values are probably due to increased runoff as a result of unrooted sediments in the watershed.

Following the herb-tundra zone, pollen dominance shifts to Zone A-1-3 approximately 14.6 kyr BP, indicating a change in vegetation to spruce and oak dominated taxa (Davis, 1969). Concomitant with this change in pollen zone is a stabilization of susceptibility, and generally higher δD_{BA} -inferred temperatures are observed (Figure 1.19). It may be possible that the cold-warm signal observed at *ca.* 14.7 kyr BP could be indicative of the Oldest Dryas – Bølling-Allerød warming transition, as this signal is consistent with the Greenland ice transition of Stuiver et al. (1995). However, this observation is based on only two data points within the δD_{BA} ,

and the event cannot be confidently identified because it is not observed in other proxies.

Shuman et al. (2004) categorizes the climate of the pre-Younger Dryas Late Pleistocene as cold and dry based on a compilation of proxy data from lakes in central New England (Figure 1.17), but surface air temperatures inferred from δD_{BA} at Gorton Pond reflect cool temperatures like those found in southern N.E. localities in CT and NJ (Peteet et al., 1993), and southeastern NY (Menking et al., 2012). It is unclear as to why there would be a discrepancy between these proximal locations, though it is likely that the record at Gorton Pond during this time is reflecting local conditions.

4.2 Environment B: Younger Dryas cold interval (13.0 – 11.8 kyr BP)

The Younger Dryas (YD) cold interval is arguably one of the most studied climate anomalies in Earth's Quaternary history. It is especially well defined in European records, as well as southern New England, but is less constrained in other parts of the northeastern part of North America. The most accepted interpretation for the cause of the YD is a breakdown of thermohaline circulation in the world's ocean (Broecker et al., 1989), due to the draining of Glacial Lake Agassiz. The YD is constrained in the oxygen-isotope record in Greenland (Stuiver et al., 1995) and reflects a cold, dry climate. Palynological studies (Deevey, 1939; David, 1969) indicate that warming in the pre-YD late deglacial period had led to boreal forestation in the region, dominated by warm-tolerant taxa. The spruce-dominated pollen zone A began during this time, but increased percentages of fir, oak, ash, beech, and hornbeam were common. However, the onset of the YD resulted in a shift to more

cold-tolerant taxa, and spruce pollen percentages reached maximum values during this chronozone (Shuman et al., 2004). The YD chronozone is often referred to as the last return to glacial conditions before warming and relative climate stability began in the Holocene.

The Younger Dryas stadial is reflected in the Gorton Pond proxy record by pronounced changes in multiple proxies (Figure 1.17). This interval is marked by a stabilization of magnetic susceptibility and density values, which indicate a cessation of high-energy transport of larger grains. Concomitant with the leveling out of the aforementioned proxies is a slight decrease in organic carbon deposition, and these values deviate from the increasing trend observed in Environment A. Sulfur and nitrogen concentrations stay relatively stable throughout the Younger Dryas.

The largest deviations associated with the Younger Dryas lie within the isotope records. To begin with, the δD_{BA} paleotemperature proxy suggests a sharp decrease in surface air temperatures at the onset of the YD. This decrease is consistent with accepted Younger Dryas signals in the northeast, and an overall decrease of 2.8°C to a YD value of 7.7°C reflect conditions $\sim 4^{\circ}$ colder than modern values. The $>20\text{‰}$ drop in δD_{BA} is comparable to the decrease seen in hydrogen isotope delta values at Crooked Pond, MA (Huang et al., 2002), which suggest an average temperature $>5^{\circ}$ less than modern values.

Secondly, a large magnitude decrease in $\delta^{13}\text{C}$ is observable in the transition into the YD stadial. Values drop rapidly and parallel the observed decrease in temperature. This large shift towards more negative isotope values is likely a result of decreased aquatic productivity, and a concomitant increase in C/N ratios indicates that

the organic matter source was shifting back toward the terrestrial domain. Increased seasonal ice cover during the Younger Dryas may have led to limited light penetration, and thus lower aquatic productivity. The large increase in $\delta^{15}\text{N}$ may be due to an increase in terrestrial organic matter input, as soil organic matter has a typical $\delta^{15}\text{N}$ value of (5–6‰).

Along with the decrease in $\delta^{13}\text{C}$, a sharp decline in $\delta^{34}\text{S}$ marks the onset of the YD. Sulfur is primarily supplied to lake environments in the form of inorganic sulfates and S-bearing compounds, and sulfur sedimentation is largely controlled by the supply of carbon and reactive metals, such as iron. The conversion of dissolved sulfate to hydrogen sulfide gas is achieved through sulfur-reducing bacteria, and if this H_2S gas reacts with iron or organic compounds, sedimentation will occur. However, the rate of this reaction is controlled by the availability of organic matter and sulfate. Limitation of organic matter and/or reactive metals will result in a slower sulfate-reduction reaction and greater isotopic fractionation, leading to larger depletions in ^{34}S . Alternatively, a sufficient supply of OM and reactive metals, or a limitation of sulfate, will result in a faster reaction and minimal fractionation of original sulfate. The overall increasing trend in the sulfur isotope values throughout Gorton Pond's record (Figure 1.11) would suggest that sulfate is becoming more limited over time, resulting in higher $\delta^{34}\text{S}$ ratios. Deviations from this increase, such as at the onset of the Younger Dryas, may be due to limitations of either organic matter, or reactive metals. Because in lake environments iron is rarely in short supply (Cohen, 2003), the sharp decrease in $\delta^{34}\text{S}$ ratios at the onset of the Younger Dryas is likely a result of a decrease in the organic matter content that occurs during this period.

Although the proxies measured in Gorton Pond reflect the accepted characteristics of the Younger Dryas, it seems the timing of occurrence is earlier in our interpretation. The onset of YD chronozone is well constrained at 12,900 – 11,600 years BP for lakes in southern New England (Shuman et al., 2004; Ellis et al., 2004; Menking et al., 2012). Hou et al. (2007) measured δD_{BA} of lake sediments at Blood Pond in south-central Massachusetts, and interpreted the YD onset to have occurred at approximately 13.0 kyr BP, which is 500 years after the observed time at Gorton Pond. However, the start of the YD in the Gorton Pond record is controlled by the original moraine-bracketed basal age estimation (~18.0 kyr BP) of sediment core GP13, and the associated assumptions mentioned in Section 1.3.3. After analysis of the proxy changes, it is likely that this basal age is too old by several centuries, as an estimate for ice-block melting rates was not assumed in the calculation of the timing of Gorton Pond's genesis, and it is likely that the entire sediment record was not penetrated. After the observation of multiple proxy changes in the sediment character, a basal age of 16.0 kyr is more appropriate. A linear sedimentation rate from 507 (12,100 cal. years BP) to 670 centimeters (16,000 cal. years BP) would place the YD onset at 13,000 years BP, consistent with other paleoclimate reconstructions (Stuiver et al., 1995; Shuman et al., 2004; Ellis et al., 2004; Menking et al., 2012; Hou et al., 2007). However, a lack of datable material at the bottom on the core results in a basal age estimate that was tuned to observed proxy shifts assumed to be the YDC. The termination of the YD is constrained by a radiocarbon sample, and can be assumed to indicate a precise age boundary.

In summary, the YD is observed in the Gorton Pond proxy record, and indicates a shift back to cold, glacial conditions. The hydrogen isotope ratios indicate rapid cooling, and match similar trends found in other lakes in the region. Decreases in organic carbon and $\delta^{13}\text{C}$ values suggest decreased lake productivity; perhaps due to increased duration of ice cover. The decrease in the sulfur isotope signal is likely related to the low organic matter production. The timing of the beginning of the YD led use to tune our basal age determination to a younger absolute age of 16,000 years before present, which put the onset of the YDC at 13.0 kyr BP, a value consistent with other paleolimnological studies (Shuman et al., 2004; Hou et al., 2007) as well as the GISP2 ice record of Stuiver et al. (1995).

4.3 Environment C: Post-Younger Dryas Warming Period (11.8 kyr BP – 9.6 kyr BP)

Although the Younger Dryas chronozone indicates great change in the environment surrounding Gorton Pond, it is the post-YD (Environment C) period that shows the most pronounced proxy variations in the entire record (Figure 1.17). Shifts in multiple proxies from 11.8 kyr BP to 9.6 kyr BP suggest that the time period following the Younger Dryas was one of significant climatic changes, and seem to indicate a clear transition from the Late Pleistocene epoch to Holocene-type conditions. Moreover, comparisons with other paleoclimatic studies in the region show very interesting similarities and discrepancies.

The post-YD chronozone at Gorton Pond is characterized by increases in nearly every proxy measured. Only the GRAPE density values showing a decreasing trend, as sediments were becoming more organic rich. Carbon, nitrogen, sulfur, and

hydrogen isotope ratios all show increasing trends throughout this interval, as do organic carbon, nitrogen, and sulfur percentages. While the overall trends for these proxies are increasing, the chronozone shows two distinct sub-environments. Early after the termination of the YD (11.8 – 10.9 kyr BP), a sharp positive increase in $\delta^{13}\text{C}$ occurs, bringing ratios back to pre-YD values. Organic carbon percentages remain relatively similar to those found at the end of the YD, as do nitrogen and sulfur compositions, and $\delta^{15}\text{N}$ ratios. A slight increase in magnetic susceptibility is observable early in the post-YD Environment C1. The $\delta\text{D}_{\text{BA}}$ values indicate temperatures similar to the cool, pre-YD values observed in late Environment A.

A significant environmental change occurs at *ca.* 10.8 kyr BP, evidenced by rapid increases in $\delta\text{D}_{\text{BA}}$, and $\delta^{13}\text{C}$. The hydrogen isotope ratio data suggests significant warming within Environment C; an increase from 10.6° to 15.8°C is observed in the last millennium of this time interval. These $\delta\text{D}_{\text{BA}}$ shifts coincide with Hou et al. (2006) $\delta\text{D}_{\text{BA}}$ shifts found at Blood Pond, MA (Figure 1.19). Coupled with a shift toward more positive carbon isotope ratios and OC preservation, the data shifts suggest a period of very warm, productive conditions in the Gorton Pond region, and this interpretation is consistent with the work done at Shawagunk Ridge in southeastern NY (Menking et al., 2013), and southern New England (Shuman et al., 2004). Furthermore, a spike in magnetic susceptibility, ARM, and IRM may suggest a lowering of lake level during this time, but more evidence is needed to make this interpretation.

This period of the early Holocene in southern New England has been shown to have warm, drought-like conditions (Shuman et al., 2004). Pollen assemblages at Rogers Lake show abundances in higher temperature tolerant pollen zone “B” (*Pinus*)

taxa, and decreases in *Picea*, which prefer cooler temperatures (Davis, 1969; Shuman et al., 2004). The presence of shallow water emergent macrofossils at Allamuchy Pond in NJ suggests similar dry conditions (Peteet et al., 1993). Ground-penetrating radar and sediment studies at Davis Pond, western MA, also evidence drought conditions from 13.4 – 10.9 kyr BP, and 9.2 kyr BP (Newby et al., 2011).

Despite this consistency with proximal lake studies in the region, studies at Seneca Lake (Ellis et al., 2004) and Fayetteville Green Lake (Kirby et al., 2002a) in western NY suggest cold and wet conditions following the Younger Dryas stadial. Both studies attribute this difference in regional climate to a change in the location of the winter circumpolar vortex following the Younger Dryas. Kirby et al. (2002b) modeled the position of the winter circumpolar vortex and determined that a significant expansion occurred from 11.6 – 10.3 kyr BP, and suggested a post-glacial renewal of thermohaline circulation (THC) as a forcing mechanism. Renewal of THC would increase the transport of warm air up the eastern coast of North America, and the intensified gradient may have caused a semi-permanent cold air trough in the western NY region (Kirby et al., 2002b). This explanation could justify the asynchronicity seen in Gorton Pond's Environment C, and could also provide insight into the regional differences observed within the western and eastern paleoclimate study locations.

Multiple proxy analysis provides evidence of a post-YD warm period in the region of Gorton Pond. The transition from full glacial to inter-glacial conditions has long been characterized by unstable climate conditions in NE North America, and the shifts evident in Gorton Pond's Environment C support this characterization. A change

in multiple proxies at 10.8 kyr BP could be explained by the interpretation of Kirby et al. (2002b), in which an expansion of the winter circumpolar vortex may have resulted in the cold/wet conditions observed in western NY, while warm/dry conditions are observed in SE New England.

4.4 Early-Middle Holocene Stable Period (9.6 kyr BP – 3.6 kyr BP)

Following the rapid warming trend observed in the two millennia long post-Younger Dryas interval, the Gorton Pond sediment record reflects relatively stable conditions. From 9.6 – 3.6 kyr BP, values for most proxies begin to show smaller amplitude variations relative to the environments observed earlier in Gorton Pond's history. Increasing OC% and sediment record-high carbon isotope ratios would suggest increased productivity throughout this interval.

The relative consistency in the proxy record at Gorton Pond throughout this period relatively long (6.3 kyrs) is not necessarily reflected in other literature sources. Shuman et al. (2004) interprets this period as spanning three distinct New England regional climate regimes (Figure 1.17, bottom), and Menking et al. (2013) interprets no less than 6 different environmental changes preserved in the sediments from two sites in the Sky Lakes region of NY (Figure 1.18). New England pollen zones shift from pine-dominated zone B, to the oak-hemlock beech (C1) zone at 8.1 kyr BP, and then to an oak-hickory (C1) zone at approximately 5.0 kyr BP, suggesting significant climate change in the region over this period. However, the proxy record does not show significant changes in the carbon, nitrogen, or sulfur elemental or isotopic proxies.

However, of particular interest is the behavior of the hydrogen-isotope ratios during Environment D. While values generally reflect changes observed in Blood Pond (Figure 1.19), Gorton Pond's record displays higher amplitude variations in δD_{BA} . It is possible that this may be due to the region's high sensitivity to changes in air mass sources. Gorton Pond is located in a region that is influenced by both cold, northern air masses, as well as warmer air masses from the Gulf of Mexico, on a seasonal and decadal basis (Yarnal and Leathers, 1988). Although the location of Gorton Pond in respect to Blood Pond is not that great a map distance (Figure 1.18), our site may have higher sensitivity to changes in atmospheric conditions. This being said, there are many other local effects that could cause different isotopic fractionation effects on hydrogen, including changes in precipitation/evaporation budgets, influence and source of groundwater, lake residence time, and/or dominant watershed vegetation. Despite the difference in amplitude, it is interesting that the overall shifts between the two sites are preserved in the δD_{BA} ratios, but as of now a single definitive explanation for this relationship cannot be determined.

4.5 Environment E: Late Holocene Cooling Period (3.6 kyr Bp – 0.25 kyr BP)

The onset of Environment E occurred at *ca.* 3600 years BP, and is characterized by initial warm conditions, though cooling occurs through this period. Changes in environmental magnetism and lake productivity are evident in Environment E. Although a large void in core material exists (~900 years), interpretations can be confidently made about the overall lake conditions that existed in the Late Holocene.

Organic carbon content is highest in the Late Holocene than in any other part of the record, but $\delta^{13}\text{C}$ values show a significant shift toward more negative numbers, and are comparable to values that occurred during the Younger Dryas. This part of the record is very interesting, as increases in organic matter would suggest increasing productivity, but the carbon isotope ratios do not show the typical ^{13}C enrichment common to highly productive conditions. This negative covariance between elemental composition and isotopic ratios may be explained by a shift in the mixing regimes in the lake. Stratification causes deep waters to become depleted in ^{13}C , but upwelling events will return these depleted waters to the surface, and isotopic ratios will show a decreasing $\delta^{13}\text{C}$ trend. This period of Holocene history has been shown to be a very wet climate interval (Shuman et al., 2004) and increasing lake depth may result in changes in mixing regimes.

In Environment E, elemental nitrogen percentages show a significant increase, but nitrogen isotope ratios show a shift toward more negative values (Figure 1.13). The increase in productivity and nitrogen concentrations coupled with a decrease in heavy nitrogen isotope values may best be explained by a change in the dominant algal biology in Gorton Pond. A shift to a dominance of nitrogen fixers such as cyanobacteria would reflect a decrease in the ^{15}N isotopes, as these biota draw nitrogen directly from the atmosphere, and almost no isotopic fractionation occurs. The shift toward nitrogen isotope ratios toward those of atmospheric N_2 would support this hypothesis.

4.6 Environment F: Anthropogenic Overprint (0.25 kyr BP – Modern)

Our goal for this work is to analyze and interpret the natural variations in the watershed at Gorton Pond. However, we have determined that the natural signal began to be influenced by anthropogenic changes to the watershed approximately 250 years before present. The changes have been displayed in Figure 1.13, and can be characterized by sharp increases in magnetic susceptibility, increases in the nitrogen isotope ratios, and decreases in the sulfur isotope ratios. These changes are likely resulting from significant land-use alterations over the past two-and-a-half centuries (Table 1.1). The increase in susceptibility and is likely due to the atmospheric deposition of fine grained magnetic material from the burning of fossil fuels, and the increase in $\delta^{15}\text{N}$ tends to be observed in areas that have increased run-off of nitrates from fertilizers and animal waste.

4.7 Organic Carbon Cycles in the Holocene

Many climate-forced factors influence lacustrine primary productivity, including temperature, sunlight, pH, CO_2 concentration, water column stratification and turbidity, cloud cover, and nutrient supply (Wetzel, 1983). However, if we assume that the concentration of deposited organic matter is directly related to aquatic primary productivity, we may begin to investigate patterns in the dataset that may be attributed to climate cycles.

Periodicities in organic matter deposition at Gorton Pond were analyzed through spectral analysis of the Holocene-aged loss-on-ignition dataset using the multi-taper method and auto-regressive (1) red-noise hypothesis (Mann and Lees, 1996) (Figure 1.16). Significant spectral peaks were observed at 802-, 750-, 568-, 537-

, and 439-year frequencies, with 99% confidence. Unfortunately, our relatively sparse sampling rate ($n = 75.5$ yrs) limits our ability to resolve any significant decadal or multi-decadal periodicities. However, the ~ 550 year cycle has been shown to be related to Holocene-aged ocean circulation patterns in the North Atlantic (Chapman and Shackleton, 2000), and atmospheric conditions over Greenland (Stuiver, 1995). This result would suggest that our organic matter deposition record might be controlled by the same ocean-atmosphere climate forcings.

Conclusions

- A paleoenvironmental history of Gorton Pond, RI, was developed using multiple proxy analysis of a composite sediment core taken from the deepest hole in the lake basin. Based on multiple lines of physical, chemical, and magnetic sedimentary evidence, the record at Gorton Pond consists of six different major environmental zones, spanning from *ca.* 16,000 yrs BP to modern day
- *Environment A–Late Pleistocene Deglacial Period: 16.0–13.0 kyr BP:* Conditions in the oldest sediments indicate significant input of glacier derived material, and low organic matter content suggests that the lake was slowly increasing in primary productivity. Hydrogen isotope ratios provide evidence of fluctuating cold/cool temperatures, and possibly point to an Older Dryas/Bølling-Allerød transition. Moraine ages in the area would suggest a basal age of ~18.0 kyr BP at Gorton Pond, which suggests that we may not have penetrated the entire sediment record.
- *Environment B–Younger Dryas Stadial: 13.0–11.8 kyr BP:* A return to cold, glacial conditions is indicated by a sharp decrease in temperature, a stalling of organic carbon deposition, and a sharp shift to more negative carbon isotope ratios. The decrease in sulfur isotope ratios is likely due to the limited organic matter availability. Multiple proxy shifts in this part of the record formed the

basis for using a basal age estimate of 16.0 kyrs BP, as the YDC is well constrained in both paleolimnological and paleoclimatological studies.

- *Environment C–Early Holocene Warming Period: 11.8–9.8 kyr BP:* This environment reflects rapid warming that occurred following the YDC. Increases in nearly every proxy occur during this warming period, and warm productive conditions are interpreted at our site, and correlate well with records from southeastern NY, southern New England, and Massachusetts. However, a period of cold and wet conditions is proposed at multiple sites in western NY (Kirby et al. 2002a; Ellis et al. 2004). Kirby et al. (2002b) proposes a shift in the winter circumpolar vortex at 10.3 kyr BP that would explain the regional differences in climate. This could be an explanation for the asymmetry observed in the proxy record at Gorton Pond, and validate the interpretation of two sub-environments.
- *Environment D–Middle Holocene Stable Period: 9.8–3.5 kyr BP:* This environment demonstrates low amplitude variations in most proxies, and indicates a period of relative lake stability. Interesting shifts in Gorton Pond's δD_{BA} record parallel the same proxy shifts in observed in Blood Pond, MA, but the Gorton Pond record displays higher amplitude changes. It is unclear as to what would cause the higher amplitude variations, but the air/moisture sources and/or local effects should be investigated further.

- *Environment E–Late Holocene Cooling: 3.5–0.25 kyr BP:* A significant cooling trend is observed throughout the Late Holocene Environment E. A decrease in carbon isotope ratios may be due to the increasing lake levels interpreted by other studies. Increased organic matter production coupled with decreases in the $\delta^{15}\text{N}$ isotope ratios may be an indication that a shift in the dominant algal community from phytoplankton to N-fixing cyanobacteria may have occurred during Environment E.
- *Environment F–Anthropogenic Influence: 0.25–Modern:* A clear anthropogenic signal is observed in the sediment record at Gorton Pond. Increased susceptibility, nitrogen isotope ratios, and sulfur isotope ratios are all likely due to land-use changes in the watershed over the past two-and-a-half centuries.
- *Climate change mechanisms:* Spectral analysis of the organic carbon record from Gorton Pond revealed significant centennial-scale periodicities. A dominant 550-year cycle may indicate that organic matter preservation at Gorton Pond may be related to changes in ocean-atmosphere dynamics. Potential multi-decadal or decadal periodicities could not be resolved due to a sparse sampling rate of 75.5 years.

Figure 1.1.) Aerial image of Gorton Pond and the surround urban landscape. T.F. Green Airport is visible in the top right, and the Pawtuxet River runs through the top left.

Figure 1.2.) Bathymetric map of Gorton Pond. Red marker refers to the location of sediment core GP13. Black marker refers to the location of freeze core GPRI-90. The survey was completed in feet, and the contour interval is 3 feet. The deepest hole is at 45 feet (13.7 meters) depth. Adapted from Guthrie and Stolgitis (1987).

Figure 1.3.) Recessional moraines in southern New England. Adapted from Hubeny (2006). Ages are in cal. kyBP and are based on the work of Balco and Shafer (2006) for recessional moraines, and Ridge (2003;2004) for varve work done at the Lower Quinnipiac Valley and Rocky Hill Dam sites.

Figure 1.4.) Showing the specific tie-points used for sediment core correlation at Gorton Pond. A 14-cm downward shift of GPRI-90 correlates the cores.

GP13 Environmental Magnetic Proxies

Figure 1.5.) Showing the various environmental magnetic proxies used at Gorton Pond. Concentration parameters are shown in black lines. Grain size parameters are shown in blue, and mineralogy indicators are shown in green. Environmental zones indicated within the mineralogy box on the right.

Figure 1.6.) A.) Showing GP13 magnetic susceptibility, B.) GRAPE density, and C.) stratigraphic column. Key proxy shifts and are demarcated in red, and approximate magnitude is noted.

Figure 1.7.) A.) Showing GP13 organic carbon content, B.) carbon isotope ratios, and C.) stratigraphic column. Key proxy shift values are indicated in black, with magnitudes noted. Peaks marked with arrows.

Figure 1.8.) A.) Showing GP13 elemental nitrogen content, B.) isotope ratios, and C.) stratigraphic column. Key proxy peaks are marked with arrows, and magnitudes are provided.

GP13 Organic Carbon to Nitrogen Ratio

Figure 1.9.) Showing organic carbon (from elemental analyzer) to nitrogen ratio values.

Figure 1.10.) A.) Showing GP13 elemental sulfur content, B.) stable isotope ratios, and C.) stratigraphic column. Key proxy peaks are marked with arrows, and magnitudes are provided.

Organic Carbon to Sulfur Ratio

Figure 1.11.) Showing organic carbon (EA) to sulfur ratios.

Figure 1.12.) Showing GP13 δD ratios of behenic acid. More negative values indicate colder lake water temperature, while less negative values indicate warmer temperatures. Inferred temperature (Hou et al. 2006) is in the middle panel.

Figure 1.13.) Showing all proxies at Gorton Pond versus depth. Six distinct environments are interpreted based on shifts in multiple proxies. Environments are labeled “A” through “F” on the right hand side of the figure.

Figure 1.14.) Showing the sources of sediment organic matter at Gorton Pond through time (Meyers and Teranes, 2001). Organic matter is a mixture of C₃ terrestrial plants and lacustrine algae throughout most of the sediment record.

Figure 1.15.) The multi-disciplinary age-depth model for Gorton Pond. The upper 212 cm were constrained by a 2nd-order polynomial. Linear regressions were used to connect each radiocarbon date.. The sediment record between 507 cm and the core base isn't well constrained, but an assumed basal age of 16 cal. kyBP is used.

GP13 Spectral Analysis of Organic Matter

Spectral Peaks

Number	Frequency	Periodicity (yrs.)	Confidence
1	0.00399679	250	90%
2	0.00350528	285	90%
3	0.00227649	439	99%
4	0.00186258	537	99%
5	0.00175910	568	99%
6	0.00133226	750	99%
7	0.00124629	802	99%
8	0.00113184	884	95%
9	0.00049151	2035	95%

Figure 1.16.) Spectral analysis plot of the GP13 organic matter time series. Time series consists of organic matter values calculated from loss-on-ignition analyses within the depth interval 507 – 212 cm (12.1 – 1.8 kyr. BP). Table below shows the significant spectral peaks, with 99% confidence intervals in bold.

Gorton Pond Environmental Proxy Change Record

Figure 1.17.) Showing multiple proxies versus age at Gorton Pond. Results suggest that there are 5 significant environmental changes recorded in the sediment recorded at Gorton Pond. Bottom panels show comparison to other N.E. vegetation and hydroclimate reconstructions. Gray boxes alternate for visualization purposes.

*adapted from Deevey (1939) & Davis (1969) **adapted from Shuman et. al (2004)

Figure 1.18.) Locations of sites mentioned in this thesis chapter. Abbreviations are as follows: SL = Seneca Lake (Ellis et al. 2004); FGL = Fayetteville Green Lake (Kirby et al. 2002); AP = Allamuchy Pond and LP = Linsley Pond (Peteet et al. 1993); MiL = Lake Minnewaska and MoL = Lake Mohonk (Menking et al. 2012); DP = Davis Pond (Newby et al. 2011) RL = Rogers Lake (Davis 1969; Shuman et al. 2004); BP = Blood Pond (Hou et al. 2006); CP = Crooked Pond (Huang et al. 2002) WP = Winneconnet Pond (Suter 1985; Shuman et al. 2004); and GP = Gorton Pond.

Figure 1.19.) Showing the δD_{BA} paleotemperature curves for Gorton Pond and Blood Pond, as well as pollen analysis from Blood Pond. Gorton Pond environmental zones are listed at the top of the figure. Paleotemperature pattern changes agree fairly consistently, but the amplitude of change is much higher at Gorton Pond than at Blood Pond. Vegetation changes at Blood Pond correlate with the pollen zones of Davis (1969). Figure modified from Hou et al. (2006). The Younger Dryas Chronozone is demarcated with a blue rectangular box, and pollen shifts are demarcated with alternating green boxes.

Table 1. The History of Apponaug Village (from King et al. 1993).

1600	Narragansett Indians occupy land around Gorton Pond and a major Indian trail, the Pequot Trail, existed at the present location of Highway 1 on the east shore of Gorton Pond.
1643	Warwick settled by Samuel Gorton. Initial settlement concentrated at Warwick Neck.
1676	Original settlement destroyed in King Phillip's War.
1677	Settlement rebuilt.
1696	The original village of Apponaug established at John Micarters' fulling mill (textiles) on Kekamewit Brook.
1708	Population of Warwick 480. Seaport on Apponaug Cove established about this time.
1730	Population of Warwick 1,178.
1774	Population of Warwick 2,438. Post Road, a major thoroughfare, occupies the location of the Pequot Trail.
1805	Cotton factory opened.
1830	Stonington Railroad built through Apponaug Cove.
1834	Warwick municipal government formed. First printing and dyeing of textiles relocated to Apponaug Village.
1850	Oriental Printworks opened.
1860	Apponaug Company founded.
1892	New City Hall built.
1900	Population of Warwick 21,316.
1920	The arrival of the automobile. Population of Warwick and West Warwick 28,942.
1927	US Rt 1, the major paved route between New York and Maine, built.
1958	Apponaug Company Mill closed.
1965-1968	Construction and opening of Interstate Highway I- 95.
1969	Apponaug Mill destroyed by fire.
1975	New Police Department building constructed on the south shore of Gorton Pond.
1980	Population of Warwick and West Warwick 114,149.

Table 1.2. Sediment core inventory and associated information for Gorton Pond.

Core	Sections	Section Depth (cm)	Age Control	Proxies Run	Proxy Resolution	Elemental Analysis Standards	IRMS Standards
GP13	GP13BC1 GP13LC2D2 GP13LC2D3 GP13LC2D4 GP13LC2D5 GP13LC2D6	0 - 115 1.92 - 2.78 2.84 - 3.77 3.82 - 4.77 4.81 - 5.77 5.75 - 6.70	4 AMS- ¹⁴ C	Susceptibility	1 cm		
				GRAPE density	1 cm		
				Loss-on-ignition	2 cm		
				NRM, ARM, IRM	1 cm		
				%Carbon	8 cm	Sulfanilimide: C:41.81%	
				%Nitrogen		N:16.25%	
				%Sulphur		S:18.62% H:4.65%	
GPRI-90	1 section	0 - 77	8 ²¹⁰ Pb/Cs ¹³⁷ & <i>Ambrosia</i> Horizon	δ ¹³ C	8 cm		USGS-40 & USGS-41
				δ ¹⁵ N			
				δ ³⁴ S	8 cm		IAEA S-2 & IAEA S-3
				δD	15 cm (avg)		
GPRI-90	1 section	0 - 77	8 ²¹⁰ Pb/Cs ¹³⁷ & <i>Ambrosia</i> Horizon	Susceptibility	1 cm		
				%Carbon %Nitrogen	6 cm (avg)		

Table 1.3. Chronological data for Gorton Pond.

Type	Depth (cm)	¹⁴ C Lab code	δ ¹³ C (‰)	¹⁴ C date ¹⁴ C (yr BP)	Cal age range (2σ) (cal yr BP)	Relative Area Under Curve	Age* (pre-2013)
Surface	0-1						0
¹⁴ C	212	BETA-363083	-27.4	1840 ± 30	1883 - 1708 1865 - 1840	0.933996 0.066004	1859 ± 87
	335	BETA-363084	-27.7	6140 ± 30	7157 - 6950	1	7117 ± 103
	448	BETA-363085	-25.4	9570 ± 40	11110 - 10736	1	10986 ± 182
	507	BETA-363086	-24.0	10310 ± 40	12225 - 11978	0.816269	
					12381 - 12257	0.183731	12165 ± 123
²¹⁰ Pb/ ¹³⁷ Cs**	15						25
	18						31
	21						39
	22						40
	24 (¹³⁷ Cs peak)						50
	27						63
	30						80
	33						100
<i>Ambrosia</i> **	61						338

*median probability

**taken from core GPRI-90

**Reconstructions of the Late Quaternary Landscapes Observed in the Seismic
Record at Greenwich Bay, Rhode Island**

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Abstract

The study of underwater archaeological sites has been of importance to researchers for decades, but increased interest in near shore construction projects (i.e. Cape Wind; Deep Water Wind) has made the identification of submerged potential cultural sites even more imperative. Greenwich Bay is an area that has been the location of many Paleo-Indian (up to 10,000 years BP) artifact findings, and is a prime location for developing best practices methodology for the determination of archaeological sensitivity. A geophysical survey was conducted using acoustic sub-bottom (CHIRP) seismic reflection techniques, and 24 lines covering 40 km were processed and interpreted. Four distinct seismic units were interpreted, spanning from the Pleistocene-aged deglacial sediments to modern day estuarine sediments. A basal till/bedrock surface marks the limit of seismic penetration, and overlying thick (up to 42 m) varved proglacial lake sediments suggest that the area was part of Glacial Lake Narragansett (GLN). Paleochannel stream cuts unconformably lie above the proglacial lake sediments, indicating a draining of GLN and a period of subaerial exposure in which glacial meltwater formed an organized tributary system in Greenwich Bay. A GIS-based Local Polynomial Interpolation model was used to create a representative surface of the stream-dissected paleolandscape that existed prior to marine inundation. While useful, denser CHIRP coverage (100 m line spacing minimum) is recommended in order to take stress off of the interpolation. The proximity of channels to archaeological finds at Cedar Tree Beach would suggest that they might have been important resources for ancient inhabitants. Comparisons to the USGS East Greenwich Quadrangle Surficial Geology Map (Smith, 1955) shows significant kame terrace and

other glacial meltwater deposits that may indicate a paleo-drainage of the modern day Pawtuxet river into the paleolandscape observed in the Greenwich Bay seismic record.

Introduction

2.1.1. Statement of the Problem

In recent years, there has been increased interest in sediment deposits in Greenwich Bay, Rhode Island, due to the geometric relationships they have with older (> 6,000 years BP) paleo-tributaries that may have been key resources for ancient Native American tribes (Leveillee, 2003). The recent discovery of paleo-Indian artifacts at Cedar Tree Beach (Figure 2.1A.) suggests that there is high potential for a submerged paleocultural site located in Greenwich Bay. In regards to early cultural histories, it has been hypothesized that the characteristics of the natural environment are of greater significance to the location behavior of humans than are the characteristics of the social environment (Kvamme, 2006). Therefore, the location of a potential paleocultural site in Greenwich Bay would be best predicted by the reconstruction of the natural paleolandscapes on which ancient people lived, and subsequent application of an archaeology-based prediction model. However, paleolandscape reconstructions of submerged environments require dense, multi-disciplinary datasets, and a thorough understanding of both the regional geologic framework and sea level history.

In the following chapter, a preliminary paleolandscape reconstruction will be developed through the interpretation of seismic reflection profiles. A number of maps will be provided, and hopefully form the basis for a geospatial paleocultural site location model in the future.

2.1.2. Justification for and Significance of the Study

The study of underwater archaeological sites has been of importance to researchers for decades (Gagliano et al., 1982; Benjamin, 2010; Evans and Keith, 2011), but increased interest in near shore construction projects relating to alternative energy development (i.e. Cape Wind; Deep Water Wind) has made the identification of potential submerged cultural sites of utmost importance. While the Bureau of Ocean Energy Management (BOEM) is the federal agency in charge of leasing potential sites for offshore energy development on the outer continental shelf (OCS), they are also required by law to consider the full effects that development may have on submerged cultural resources. To act in accordance with these various statutes (Outer Continental Shelf Lands Act [OCSLA]; National Environmental Policy Act [NEPA]; and National Historic Preservation Act [NHPA]) BOEM has urged leaseholders against altering any archaeological resources that may be found during site development. However, standardized scientific methodology for the assessment of archaeological sensitivity in submerged environments is lacking. The absence of set protocols has been a concern for historic preservation officers at federal, state, and tribal levels, and has made the process of decision-making more difficult.

Fortunately, recent advances in marine research technology have lessened the degree of logistical difficulty for shallow-water archaeological and geophysical surveying methods, and have increased spatial and temporal efficiency without compromising safety. Decades of seafloor exploration and research in coastal regions by governmental, academic, and private organizations have led to an improvement in the accessibility of necessary data, as well as an overall increase in the pool of multi-

disciplinary scientific knowledge. Furthermore, the passage of the Energy Policy Act of 2005 granted BOEM jurisdiction of some types of renewable energy development on the OCS, and their internal Environmental Studies Program has assumed responsibility for implementation and management of relevant research.

Each of these developments has led to renewed efforts for the development of key protocols necessary for the identification of archaeological sites in submerged environments, and the reconstruction of marine paleocultural landscapes. It is for this reason that BOEM has funded the University of Rhode Island to spearhead a 4-year research project with the two main objectives listed below.

- 1.) Enhance and refine our understanding of submerged paleocultural landscape distribution on the Atlantic OCS, especially landscapes of tribal significance; and
- 2.) Understand and identify paleocultural landscapes of importance to regional Tribes through collaborative research.

A critical element that makes this study even more unique is that it will involve Tribal representatives in all aspects of the project, from planning and fieldwork to decision-making and mitigation.

The focus of the Paleocultural Landscapes Project will take place at four different submerged coastal locations along a N-S transect in Rhode Island waters (Figure 2.1A.). These sites were chosen based on both the existence of available

geoenvironmental databases and also the accessibility and ease of archaeological excavation.

The research presented in this chapter focuses on the Greenwich Bay area, which was chosen as the initial study site for the overall project due to the abundance of paleo-Indian artifacts previously recovered at the site (Figure 2.1B). In addition, previously-collected geophysical and geological data was available for initial interpretations, and Greenwich Bay's protected and relatively shallow water environment provided an ideal location for developing effective surveying methods."

2.1.3. Study Site and Geologic Context

Greenwich Bay is located in the mid-western passage of Narragansett Bay (Figure 2.2.) and is a shallow embayment (average depth of 2.6m) covering approximately 12 square kilometers. The area is of archaeological importance due to the presence of pre-European settlements that were situated around the coast of the bay in the past (Rhode Island Historic Preservation & Heritage Commission [RIHPHC] site files, accessed 2012), and also due to the large number of ancient projectile points that have been found on the shores and shallow water areas surrounding Greenwich Bay. Dates of these projectile points have been estimated to span over 10,000 years, based on the shape and style of their construction (Dr. Timothy Ives, personal communication). Moreover, significant evidence exists for other archaeological sites proximal to Greenwich Bay (RIHPHC, accessed 2012).

Bedrock Geology

The bedrock geology of Greenwich Bay is described by Quinn (1952) in the East Greenwich Quadrangle bedrock geologic map. Due to the existence of thick glacial deposits in the region, the characteristics of the bedrock geology is of minor importance in this study. However, the lower elevation of the Rhode Island Formation bedrock in modern day Narragansett Bay provided natural pathways for the ice lobes of the Laurentide Ice Sheet (J.H. Smith, 1955). The depth of the bedrock valley in the East Greenwich Quadrangle is approximately 35 meters below mean sea level, and the lowest elevation of estuarine deposits has been estimated at ~12 meters below mean sea level (Upson and Spencer, 1964). A detailed summary of the bedrock geology in this chapter's study site has been presented previously in section 1.1.3 of this thesis.

Quaternary Geology

The bedrock geology of the East Greenwich Quad underlies thick deposits of unconsolidated Quaternary-age sediment cover. These deposits consist largely of Late Pleistocene ground moraines and sequential outwash deposits associated with the retreat of the Laurentide Ice Sheet (LIS), although recent swamp, alluvium and shoreline deposits exist (J.H. Smith, 1955). The deglaciation chronology of southern New England is extensively covered in scientific literature archives, and a detailed summary is given in thesis section 1.1.4.

Boothroyd and August (2008) divide the Quaternary geologic deposits of Rhode Island into four general glacial provinces: (1) thick stratified deposits, (2) granitic/gneissic gravelly till upland, (3) compact till upland, and (4) complex till and

stratified material. The geology surrounding Greenwich Bay is best described as thick stratified glacial material, deposited prior to and during glacial retreat. The eastern side of the Bay consists of till and the western side of ice-marginal sand and gravel, while glacial deltas exist in both the north and south shores (Boothroyd and McCandless, 2003). The modern estuarine floor of Greenwich Bay is largely silt and sandy-silt associated with low energy conditions (Oakley et al., 2012).

During the Quaternary epoch, the geomorphology of New England landscapes were largely altered by glacial processes, and resulting deposition of glacial material makes up much of the modern geology. Large glacial lake complexes were common following deglaciation and prior to isostatic uplift. Modern day Narragansett Bay was occupied by Glacial Lake Narragansett, and the majority of sedimentary structures prior to marine inundation are lacustrine floor, lacustrine delta, and alluvial fans structures (Boothroyd and August, 2008). A notable delta exists in Warwick north of Greenwich Bay, and drained into what was most likely a glacial lake. Although the mechanism for the draining of Glacial Lake Narragansett is not fully understood, one school of thought centers around catastrophic draining due to isostatic uplift (Urish et al., 2001; Oakley et al., 2013).

Post-glacial sedimentation began following the drainage of Glacial Lake Narragansett approximately 19,500 years BP (Uchipi et al., 2001). Initially, a freshwater fluvial drainage surface was established on the exposed lake floor. Subsequently, during the Holocene, a marine transgression produced estuarine sedimentation on top of the older glacial lacustrine deposits (Boothroyd and August,

2008). The estuarine sedimentation rate for Narragansett Bay is estimated at approximately 0.65 – 1.3 mm/yr (Peck and McMaster, 1991).

Research conducted in the lower Western passage of Narragansett Bay interprets three distinct depositional environments since deglaciation (Peck and McMaster, 1991). The first sequence, a glacial depositional sequence, consists largely of lower till and bedrock, glaciolacustrine fan material, and glaciolacustrine prodelta slope/lakefloor. Overlying unconformably above these deposits is a fluvial unconformity and fluvial deposits associated with freshwater drainage systems. Lastly, an estuarine depositional sequence covers the bay floor, deposited after marine transgression.

Sea Level History

The sea level history of Narragansett Bay has been studied extensively (McMaster, 1983; McMaster, 1984; Peck and McMaster, 1991). McMaster (1984) performed an extensive seismic reflection survey of Narragansett Bay, and determined a postglacial inundation history based on the sea-level curve of Oldale and O'Hara (1980), bathymetric and acoustic sub-bottom data, and backstripping of Holocene estuarine sediments. McMaster's interpretation estimates that marine inundation of Narragansett Bay occurred at approximately 10,200 years BP, when sea level reached 30 m below modern. Using the sea level curve of Oldale and O'hara (1980), McMaster determined reference shorelines for 5-meter depth increments, providing both a timing of inundation and estimate of estuarine sedimentation for the entire

Narragansett Bay. McMaster's interpretations suggest that marine inundation of Greenwich Bay began at approximately 5,500 years BP.

Comprehension of the Late Pleistocene and Holocene geologic history of the Greenwich Bay area is an essential step in the process of reconstructing the landscapes that may have existed during these epochs. When working to distinguish distinct landforms from geological and geophysical data, it is necessary to fully understand the depositional and erosional mechanisms that may have helped shape the paleolandscape. I hypothesize that the sediments within Greenwich Bay represent a history that spans from the retreat of the Laurentide Ice Sheet in the Late Pleistocene, through the Holocene epoch, and into modern day. Therefore, a thorough understanding of the glacial and deglacial history, as well as the timing of sea level rise in the region, is necessary for the interpretation of potential submerged paleolandscapes.

Methods

2.2.1 Sub-bottom seismic reflection profile acquisition, processing, and interpretation

Compressed High-Intensity Radar Pulse (CHIRP) acoustic sub-bottom data was acquired in the spring of 2006 using a pole-mounted Edgetech 670c full spectrum CHIRP profiler, operated at 1.0 – 10.0 kHz. Twenty-four survey lines from Greenwich Bay were incorporated in this study, and can be found in (Table 2.1). Line spacing was 50 meters in the inner Bay (lines 1-9, Figure 2.2.), and approximately 300 meters in the mid and outer Bay.

Seismic lines were converted to SEG-y format and processed using Chesapeake Technology Inc.'s SonarWeb Pro. Sediment bottom tracking and reflector mapping was completed with this software, and profile depths were calculated using an assumed acoustic velocity of 1500 meters per second. Geospatial attributes were exported to Microsoft Excel spreadsheets as decimal degree latitudes (Y) and longitudes (X), and depth below towfish (Z). Sediment thicknesses were calculated using the “Seismic Reflectors Thickness” tool in SonarWeb. Thicknesses are calculated in the program by taking the algebraic difference between specific reflectors, and XYZ values are provided for every shotpoint between the reflectors of interest. Seismic images were saved in JPEG format, and imported into Adobe Photoshop CS5 for sharpening and reflector drawing.

Sub-bottom profiles were examined for seismic units using seismic analysis of reflector characteristics, terminations, and configurations (Mitchum et al., 1977). Depositional environment changes were interpreted in the context of the prior work

done in the area (McMaster, 1984; Boothroyd and Sirkin, 2001; Boothroyd and August, 2002; Oakley, 2011). When inferring sedimentary facies, borehole records of similar deposits in the Western Passage of Narragansett Bay were used as ground truthed records (Peck and McMaster, 1991).

2.2.2 Geospatial data analysis, interpolation, and visualization

Seismic reflector points were exported to Excel spreadsheets, organized into five specific “bins”, and imported into ESRI’s ArcMap 10 as X, Y, and Z coordinate point data. Overlapping seismic bins were used in order to decrease the geospatial extent of interpolation, which showed irregularity and imprecision when attempting to interpolate the entire geophysical dataset. GIS-based interpolation methods responded much better to smaller datasets. Inspection of each interpolation was possible by visually comparing the output surfaces with seismic images, in both two- and three-dimensions.

Following the organization of the geospatial points into bins, they were converted into shapefiles and ArcMap’s geostatistical analysis package was used to interpolate the Z values (as sub-bottom depth) between seismic lines. A Local Polynomial Interpolation (LPI) Model was used, with a 1st order constant Kernel function. The output surface prediction was based on a minimum of 10 nearest neighbor points, and a maximum of 1000, in an 8-sector search ellipsoid. A separate LPI was developed under the same model parameters for each seismic data bin.

The Local Polynomial Interpolation model was chosen due to the high freedom the user has to change model parameters, the relatively fast processing speed, and

because spatial autocorrelation of the input data is not needed with this method. This interpolation model works by fitting many polynomials through the data surface, each within specified overlapping neighborhoods. A linear regression is fit through a user-defined number of nearest neighbors, and the value of predicted points is calculated by where this regression intersects with a polynomial's center.

Interpolation model parameters such as the number of included neighbors, major and minor semi-axis length, and bandwidth were altered during the development of the method. Due to the differences in reflector coverage and number of data points used associated with each bin, different model parameters were used for each LPI. ArcMap provides optimized model parameters that minimize root mean square prediction errors, and small variations from these baseline values were used. Interpolations between seismic lines conformed well to the seismic images; however, the largest areas of error in this method showed on the edges of each bin, where the LPI used data from just the single outermost line. When rasterized, the edges of each bin had to be manually cut from the data before mosaicking, as the output pixel values were obviously erroneous. The exclusion of erroneous data is a simple procedure in ArcMap, but it introduces user bias and another source of procedural error. Even so, visual inspection of interpolated output surfaces with seismic images was used for quality control, and provided the basis for qualifying the limitations of this method.

LPI results were exported to filled contour vector data files for use in ArcMap's 3D Analyst package. Triangulated irregular networks (TINs) were created in 3D Analyst, and then converted to raster datasets using the "TIN to Raster" tool. Individual raster bins were visually inspected for quality control, and anomalous

pixels were clipped from the bin by using ArcMap's "Extract by Mask" tool. Once the quality of the individual bin rasters was found to be satisfactory, they were mosaicked into a single raster layer using the "Mosaic" tool.

All geospatial data was originally displayed in a geographic coordinate system (GCS) with the World Geodetic System (WGS) standard of 1984 as a horizontal datum, and the North American Vertical Datum of 1988 as the vertical datum. However, the calculation of raster statistics is best completed when the data is displayed in a projected coordinate system (PCS). When measurements were necessary, all interpolated layers were projected into the Rhode Island State Plane Meters PCS.

2.2.3. Additional datasets and software

Interpolated output surfaces were viewed in ArcMap 10.0 as two-dimensional raster datasets. However, three-dimensional visualization was completed through the use of QPS Inc. Fledermaus IVS7 software. The interpolated surfaces were imported into Fledermaus as gridded raster data directly from an ArcMap geodatabase file, and topographic data in the study area was displayed in Fledermaus by importing the United States Geological Survey's 1/3 arc-second National Elevation Dataset (Gesch, 2007; Gesch et al., 2002). In most cases, a 15x vertical exaggeration of 3-D data was used in order to aid in the visualization of subtle changes in Z values. Seismic images were georeferenced and displayed within the interpolated surfaces by using the "Vertical Curtain" tool. In addition to seismic images, the USGS Quaternary surficial

cover map (Smith, 1955) was georeferenced in ArcMap and displayed in Fledermaus to aid the visual association of the present day topography with glacial deposits.

Results

2.3.1. Seismic Analysis

The seismic analysis of Greenwich Bay shows several significant reflectors and sedimentary units. The naming convention for reflectors is d1, d2, d3 (for “discontinuity”), with 1 being the oldest reflector. The ravinement surface is denoted as R1, and acoustic bottom as “AB”. Seismic units are referred to as 1 – 4, with seismic unit 1 being the oldest, and are denoted as U1, U2, U3, and U4. Any potential paleochannel structures are denoted as “PC”. Representative seismic lines are provided in Figures 2.3 – 2.7, and are seismic interpretations for survey lines 9, 12, 13, 16, and 19, respectively. Referrals to these lines in the text are in the context of the aforementioned figures, and figure numbers will not always be referred to. Lines were chosen due to the quality of the seismic data, and also the location in Greenwich Bay.

CHIRP acoustic sub-bottom sonars provide high-resolution records, but cannot achieve the same level of penetration as other conventional seismic sources. Due to the shallow water column, and relatively thin sediment record (~50 meters) in Greenwich Bay, the CHIRP sediment profiler provided excellent seismic images. However, the presence of gas pockets in the deep channel of the bay caused significant acoustic disturbance (Figure 2.7), and poor acoustic penetration due to sandy bottom sediment is evident in the northwest regions of seismic lines 17 and 19 – 24. Despite these difficulties, the ravinement surface could be distinguished and mapped in the majority of seismic lines (Table 2.1.).

AB – Acoustic Basement

The “acoustic basement” represents the deepest, and oldest, reflector reached through the CHIRP penetration. In general, acoustic basement can represent a bedrock surface, till, glacial outwash, gas pockets due to decaying organic matter, or poor CHIRP penetration. Whatever the case, acoustic basement demarcates the limit of visible sediment in our record, and in some places in Greenwich Bay can be confidently determined to mark the lowermost sediments in the Bay (Figure 2.6; Figure 2.7). However, due to acoustic disturbance, the AB reflector may not always be confidently used as a stratigraphic marker (demarcated by red boxes in figures). Careful seismic analysis is necessary when interpreting the usefulness of the acoustic basement reflector, and care has been taken to only include the AB data that can be confidently used as a sediment record base in Greenwich Bay.

Discontinuity d1 and seismic unit 1

The boundary of d1 represents the base of seismic unit 1, and in most parts of the Bay d1 occurs at the seismic acoustic basement. This discontinuity is not observed in the Inner Bay, and neither is seismic unit 1 until survey line 10. Discontinuity d1 is continuous from line 10 through line 24, and generally lies approximately 25 – 30 meters below the sediment surface. In survey lines 15 and 16, discontinuity d1 is much shallower than in lines 10 – 14 or 17 – 24, due to the presence of glacial deltas in the north and south of the mid-Bay.

Seismic unit 1 is the thickest sedimentary unit throughout Greenwich Bay, and can be found from line 10 to line 24. The unit shows parallel and rhythmic sub-

reflectors throughout its entire depth, and throughout the entire Bay. Seismic unit 1 is thickest in Line 12 at >20 meters, and thins out through the Mid Bay just off of Sally Rock point on the Potowomut peninsula. It is only ~10 meters thick in lines 15 and 16. Further east at line 19, U1 thickens again to ~15 meters at its deepest point.

Discontinuities d2 and d3, and seismic units 2 and 3

Discontinuity d2 is very interesting due to its irregularity throughout the Bay. This reflector overlies seismic unit 1 in line 12 as a very clear reflector but is almost unobservable in survey line 13 (Figure 2.5). The discontinuity is unmistakable, as the sediments change from the regular, rhythmic character in U1, to an acoustically-transparent facies in seismic unit 2. This discontinuity is interpreted as a conformable contact, as no significant erosion is evident in either the lower or upper reflectors bounding U2 (Figure 2.6; Figure 2.7). However, seismic unit 2 appears to have undergone significant erosion in the mid-Bay, as the beds are cut by paleochannel structures in lines 12, 13, and 16. Line 19 is the only location in the reference images that seismic unit 2 is not eroded by paleochannels, but it is observable in lines 20 through 23 as well. Seismic unit 2 is thickest in the outer bay, averaging approximately 5 meters, and thins toward the inner bay, averaging 2–4 meters in lines 12 and 13.

Discontinuity d3 overlies seismic unit 2, and is only definitely shown in the northeast ends of lines 12 and 16, and in line 19. It is demarcated as a dashed reflector in the NE end line 13, because it is unclear whether or not the reflector shown is d2 or d3. However, it is confidently observed in lines 17 and 20 through 23. Overlying d3 is

seismic unit 3, observed in the northeast ends of reference lines 12, 13, and 16. Unit 3 is approximately 4 – 7 meters thick in these lines, which is considerably thinner than survey lines in the outer bay. It is observed as a continuous facies in lines 19 through 23, averaging over 8 meters thick in this part of the Bay. The rhythmic sub-reflectors in seismic unit 3 are considerably less hummocky than those observed in seismic unit 1. Figure 2.7 best displays this characterization, as the SW side of U1 that overlying d1/AB shows a rhythmic, albeit irregular package of sediments, while the rhythmites above d3 are more parallel.

Paleochannel structures, discontinuity R1, and seismic unit U4

The seismic record in Greenwich Bay shows several paleochannel structures that cut through lower seismic units. Incised paleochannels are shown as blue reflectors in the reference figures, and 19 total structures were identified in Greenwich Bay. Line 9 has two paleochannel cuts, one in the SE and one in the NW, and the sedimentary infill is laminated with several sub-reflectors visible. The middle of Line 9 shows what is interpreted as paleo-lake deposits, not paleochannel deposits (Figure 2.3). Two channels were also found in line 13 and also line 17. A single paleochannel was interpreted in lines 11, 12, 14, 16, and 18–23.

Discontinuity R1, otherwise known as the ravinement surface, is observed in nearly every survey line in the Greenwich Bay seismic dataset. It is interpreted to be an unconformity created by tidal and wave action, and demarcates the chronostratigraphic point of marine transgression in this part of Narragansett Bay. The ravinement surface was mapped in 21 out of 24 seismic survey lines (Table 2.1), for a

total coverage of over 21 kilometers. The depth of the ravinement fluctuates slightly throughout the Bay. Areas proximal to shorelines tend to have a shallower R1 return, and areas further from the coast a deeper return. The average depth of R1 is 7–8 meters. However, areas of paleochannel structures tend to show deeper ravinement surfaces; Line 16 has an 11-meter R1 depth in the mid-Bay (Figure 2.6). Deeper ravinement reflectors in these areas are likely due to the use of the incised paleochannels as pathways for seawater inundation during marine transgression.

Seismic unit 4 overlies the ravinement surface everywhere in Greenwich Bay. It represents all sedimentation that has occurred since marine transgression in the region, and the thickness of U4 averages 7-8 meters. Seismic unit 4 is acoustically transparent throughout Greenwich Bay, but a darker seismic return is observed in the lower half of this unit, which may reflect changes in sediment characteristics due to increasing depths during marine transgression (Vinhateiro et al., 2007; Oakley, 2012).

Figure 2.8 shows specific reflector coverage, as well as information on the seismic data bins used.

2.3.2. Interpolation Model Results

Two interpolated maps were created to best understand the Late Quaternary paleoenvironments. These maps are named “Depth to Glacial Surface” (Figure 2.9), and “Depth to Pre-Inundation Surface” (Figure 2.10). Although similar methods were employed, these two maps differ in the quality of the output surface, especially in the areas where seismic data bins were mosaicked. These differences are likely a function of the overall character of the landscapes, as the gradual changes observed in the

Depth to Glacial map may be easier to interpolate than the sharp changes observed in the Depth to Pre-Inundation Surface map.

It is important to keep in mind as a reader that the depth scale in the provided seismic images is meters below towfish, but the color bars representing depth in the interpolated surfaces are sediment sub-bottom depth. This offset must be kept in mind when comparing depths, as the values in the seismic images will seem deeper due to the overlying water column thickness.

The Depth to Glacial Surface map was created by interpolating the acoustic basement reflectors in the seismic dataset, and shows the results of the five LPI models stitched together with the Mosaic tool. The transitions from bin to bin are largely seamless, and the overall depth spectrum covered by the interpolation matches well when the seismic images are consulted. The transition between seismic bins 1 and 2 (Figure 2.8A) shows irregularity, and a small hole exists in the output surface (Figure 2.9). An obvious smearing of pixels is also observed on the east side of the deepest part of the record, near lines 11 and 12. While visually discontinuous, this problem area does not alter the overall character of the paleolandscape. In general, the Depth to Acoustic Basement interpolated surface captures the general trend of the bedrock/till seismic unit, and provides meaningful depth values to this reflector.

The Depth to Pre-Inundation Surface interpolation also shows a valuable two-dimensional visualization of the paleolandscape associated with the combined R1 and PC datasets (Figure 2.10). However, this surface displays more pixel irregularity than the Depth to Glacial Surface interpolation. In particular, seismic bin 3 shows outlier pixel values in the areas of survey track lines. Areas of sharp pixel value contrast exist

on the actual headings of lines 10, 11, 12, and 13. These values deviate from surrounding interpolated values, and are an artifact of the interpolation model. These erroneous pixels are linear in character and span across the Mid-Bay, but the area covered by each pixel group is very small, and does not alter the overall character of the output surface.

Of more concern are the existences of relatively large polygon areas that are interpolated to be the same depth. Paleochannel areas in line 14 and 16 in particular show extensive areas of flat paleo-topography in the deepest part of the interpolation (Figure 2.10.). It is unclear as to why these large dimensions of pixel areas are being represented as flat areas, but it may be related to the spacing and orientation of seismic lines (300 meters, parallel and ungridded). It may also be an artifact of organizing the seismic data into bins. One of the problem regions is line 16, which is an area of overlapping bins, and in this example the flatness may be a result of mosaicking the individual LPI datasets. It remains unclear as to which part of the procedure is problematic, but there is an ongoing effort to try and resolve this issue.

Despite these data processing and interpolation issues, the LPI models used in Greenwich Bay provided a useful means for visualizing reflectors of interest as a paleo-surface in both two-dimensions and three-dimensions. Comparisons with seismic images indicate that the process is not perfect, but the general overall geometries of the paleolandscapes are preserved. As with all interpolated data, these output surfaces should be understood as data prediction products, and should be used with discretion.

Discussion

2.4.1. Seismic Sequence Analysis

Analysis of twenty-four seismic CHIRP lines in Greenwich Bay has revealed a number of discontinuities, and five seismic facies were interpreted. By incorporating the principles of seismic sequence and seismic facies analysis, three distinct depositional environments are observable in the Greenwich Bay sediment record. These interpretations will be discussed in the following section, and comparisons between the interpretations here, and those of McMaster (1984), Peck and McMaster (1991), Boothroyd and August (2001), and Oakley (2012) will be discussed. Paleolandscape visualizations are provided and discussed in the context of usefulness in the determination of an archaeological site prediction model for the Paleocultural Landscapes Project.

2.4.2. Depositional Environment 1: Proglacial Lake Setting

Acoustic basement (AB) was observed as a hard seismic return in nearly all seismic lines (Figure 2.8B). Depth to acoustic basement was interpolated and mapped in Figure 2.9. The shallow depth to acoustic basement in the mid-Bay are due to the progradational glacial fans extending north off of Potowomut peninsula and south from the Warwick plains delta. The greater depths to AB to the east and west of these mid-Bay progradational fans are likely indicative of depths to bedrock, although a thin till veneer may still cover areas of the bedrock thalweg (Upson and Spencer, 1964). The thickest stratified deposits observable are found in these deep bedrock/till basins, with up to 42 meters of sediment in the northern Mid-Bay, in the northeast ends of

Lines 11, 12, and 13. The southern end of the Outer Bay region off of Sally Rock Point in Potowamut has deposits up to 35 meters thick.

Unit 1 lies unconformably over acoustic bottom, and consists of rhythmic reflectors in a thick sedimentary sequence. This unit is interpreted as proglacial delta/lakefloor deposits, and has been observed in many other regions in Narragansett Bay (Pack and McMaster, 1991; Oakley, 2012). This unit is very thick throughout Greenwich Bay, at >30 meters on average, and overlies the till/bedrock surface from Line 9 to Line 24. Peck and McMaster (1991) found a similar unit in their work in the western passage of Narragansett Bay, and classify the sequence as rhythmically-layered dark gray silt that was deposited as proximal glacial varves, according to the criteria of Ashley et al. (1985).

Unit 1 shows some spatial heterogeneity in terms of varve characteristics. Line 12 best shows the variation in varved sediments, but differences in vertical thicknesses and spacing of these sub-reflectors can be observed throughout Unit 1. We interpret these varve configuration changes due to the manner in which they were deposited. Areas of hummocky, irregular varve configurations were probably deposited as proximal lacustrine fans by sub-glacial meltwater and density underflows (Gustuvson and Boothroyd, 1987; Oakley, 2012). Perhaps the best area to observe the sediments associated with this type of depositional mechanism is in Line 19. The Unit 1 varved sediments in the SW end of the line are very irregular, but laterally continuous and less disturbed further into the middle of the line. Despite these changes in varve configuration, attempts to resolve differences in types of glacial lakefloor sediments was not the focus of this study, and Unit 1 is interpreted as a glacial lakefloor unit. The

internal reflectors of Unit 1 would suggest proximal glacial varve sediments dominate this unit.

Unit 2 lies conformably above Unit 1, and is an acoustically transparent seismic facies deposited between the parallel sub-reflector facies of Unit 1 and Unit 3. Unit 2 is bound by conformable contacts, indicating that minimal or no erosion occurred between units. The U2 facies is interpreted as a continuation of the glacial lakefloor record. The vertical resolution of the CHIRP system used was 1-10 kHz, which provides a vertical resolution of approximately 20 centimeters. Internal reflectors less than 20 centimeters thick would not be resolved in seismic images. We interpret Unit 2 as being distal glacial varves, deposited as the ice margin retreated further north.

Unit 3 lies conformably above U2 and shows similar reflector characteristics as U1, but reflectors are less wavy and hummocky in Unit 3 than in Unit 1. The internal seismic reflector configuration would suggest that a return to thicker glacial varve sedimentation occurred, likely due to a readvance of the Laurentide Ice Sheet into the Greenwich Bay region. This interpretation of the U3 facies would suggest that sediments were still being deposited on a glacial lakefloor setting.

The upper termination of seismic Unit 3 is characterized by erosional features associated with paleochannel cuts, unconformity R1, and/or collapse features (Figure 2.7.) caused by the melting of abandoned ice. This unit is observed in the northeast ends of lines 10–16, as fluvial and post-inundation erosion has destroyed the record in the southern ends of the Mid-Bay. However, Unit 3 is very thick (>10 meters) in the southwest ends of Lines 17 and 19–24. The northeast ends of these records were not

observed in the seismic images, as the sand/gravel bottom-type of Greenwich Bay (Oakley, 2012b) resulted in poor seismic penetration.

The seismic sequence and facies analysis of Units 1, 2 and 3 has led us to interpret this sequence as a single depositional environment: a proglacial lake setting. The deposits within this environment show variations in inferred lithologic character, but all facies would be found in a proglacial setting. Very thick proglacial lake deposits have been found in various areas of Narragansett Bay (Oakley, 2012; McMaster, 1984; Peck and McMaster, 1991), and a single proglacial lake named Glacial Lake Narragansett (GLN) likely covered all of Narragansett Bay following the retreat of the Laurentide Ice Sheet (Oakley, 2012). Analysis of bridge borings from the western passage of Narragansett Bay directly sampled the sediments associated with GLN, and up to 38 meters of dark gray silt and clay rhythmite couplets approximately 5 centimeters thick characterize these deposits. The proglacial lake environment observed in Greenwich Bay probably consists of similar lithologies; however, the rhythmitic reflectors observed are likely groupings of sedimentary couplets, rather than individual varves.

Glacial interpretations resulting from the seismic data presented in this thesis have been used to quantify the lateral and vertical extent of observed proglacial lake sediments in Greenwich Bay (Figure 2.11). The thickest deposits are greater than 34 meters of sub-bottom depth, and lie within the deeper till/bedrock basins observed in the Depth to Glacial interpolation product. Paleo-lake deposits were found in the upper reaches (east of line 10) of Greenwich Bay, but as of now are interpreted as paleo-lake

deposits not necessarily related to GLN. Our analysis of the thickness and extent of GLN in Greenwich Bay conforms well to determinations made by Oakley (2012).

Age control for glacial deposits is difficult to obtain due to the lack of radiocarbon-datable material available during these times. Due to disagreements in the existing literature about the constrained age of Glacial Lake Narragansett, data in this thesis is not used to quantify the age boundaries of the proglacial lake depositional environment in Greenwich Bay. While a seismic sequence analysis of the proglacial lake environment has been provided in this thesis, the main focus lies in the landform that existed following the draining of GLN. Therefore absolute age constraints are not a concern for this work.

2.4.3. Depositional Environment 2: Post-Glacial Fluvial Setting

Overlying the thick, stratified sediments deposited within the proglacial lake environment are a number of erosional unconformities associated with paleochannel structures. These paleochannels vary in width and depth, but are laterally continuous through Greenwich Bay (Figure 2.8D), and are interpreted to have been part of the Narragansett Bay stream-dissected surface originally proposed by McMaster (1984). An alternative to the paleo-tributary interpretation was proposed by Boothroyd and August (2008), in which they interpreted these PC structures to be due to groundwater spring sapping associated with high postglacial water tables. However, the lateral extent and geomorphology of these reflectors lead us to conform to McMaster's interpretations, and discuss the paleolandscape in the context of an organized tributary system.

Paleochannels were identified in nearly every survey line, and are best represented in the provided seismic images. There is evidence of two distinct paleochannels in line 9 within the inner bay, both approximately 150 meters in width (Figure 2.3.). These structures extend westward to line 13, where evidence of a third paleochannel reflector is observed in the southeast (Figure 2.5.), proximal to modern day Greenwich Cove. By line 14, these three distinct channels appear to converge into a single, wide (~700 meter) channel. This channel passes through line 15 as a thinner (~250 meter) structure, but by line 16 increases again to over 300 meters width. The paleochannel system continues eastward through the remaining lines, following the location of the modern deep bathymetric channel, and likely continues south through the western passage of modern day Narragansett Bay. Peck and McMaster (1991) confirmed the existence of a fluvial unconformity in the western passage of Narragansett Bay within bridge boring samples. They describe the associated sediments as fluvial sands and marsh sediments indicative of stream deposits. It is likely that the paleochannels observed in Greenwich Bay were geographically connected to those found by Peck and McMaster (1991) and the deposits are similar to those found within the Jamestown Bridge boring samples.

As stated, the original proposal of a postglacial tributary system was made by McMaster in 1984, based on a Narragansett Bay-wide seismic survey. He suggested that following the drainage of GLN the lakefloor deposits were sub-aerially exposed, and the formation of dendritic stream networks were established. I interpret the paleochannel reflector configuration in Greenwich Bay to have been part of this model. After investigation of the interpolated Depth to Pre-Inundation Surface, I

interpret at least two paleo-streams existed in Greenwich Bay following the draining of GLN and prior to marine inundation. The first paleo-stream flowed from northwest to southeast from the mouth of Apponaug Cove and just south of modern day Cedar Tree Beach. The second paleo-stream flowed from eastward, and connected with the first paleochannel in the mid-Bay area, but it is unclear whether or not this channel originated just north of Chepiwanoxet Point, or if it came from Greenwich Cove (Figure 2.10). Additional sonar lines in the area surrounding this location may help to determine the genesis of proposed paleo-streams.

2.4.4. Depositional Environment 3: Post-Inundation Estuarine Setting

Discontinuity R1 lies unconformably atop the post-glacial fluvial deposits interpreted as paleo-streams channels. The seismic facies lying above R1, also referred to as the ravinement surface, is acoustically transparent and consists of estuarine sediments deposited following marine inundation.

McMaster (1984) used the sea level curve of Oldale and Ohara (1980) and the depths of the ravinement surface throughout Narragansett Bay (what he refers to as a Holocene basal unconformity) to infer stages of Holocene shoreline advancement during marine transgression. He interpreted marine inundation at the mouth of Greenwich Bay to have occurred approximately 5,500 years before present. However, work by Vinhateiro (2007) would suggest that this age is slightly young. An AMS-dated sediment core from western Greenwich Bay suggests that inundation occurred prior to 6500 years before present, and shallow estuarine conditions were established in the inner bay by ~3500 years ago (Vinhateiro, 2007). This disagreement between

absolute age chronologies could be investigated further through sediment core analysis, but available literature brackets the age of the ravinement at approximately 6000 years BP. Therefore, the stream-dissected fluvial surface landscape is suggested to be older than 6000 years.

2.4.5. Archaeological Context and Significance, and Visualization Techniques

Kvamme (2006) hypothesized that the behaviors of less complex cultural societies would depend more on the natural environment versus the social environment. If this hypothesis is correct, then it is through the study of paleolandscapes that predictions can be made about the locations of ancient cultural habitation sites. By incorporating the principles of seismic sequence analysis, and through the use of interpolation models in ArcGIS software, a better understanding of paleolandscape surface geomorphology can be attained. This understanding of the physical landscape can help archaeologists develop prediction models concerning potential paleocultural habitation sites.

The previous discussion sections explained the interpretation of three depositional environments inferred from the seismic record. While this thesis presents the entire sediment package characterizing Greenwich Bay, the main focus from an archaeological standpoint is within the pre-inundation paleolandscape that existed prior to the modern estuarine setting. The location of a paleo-stream within modern day Greenwich Bay would have been an excellent resource for ancient peoples, and the recent numerous archaeological findings near Cedar Tree Beach would suggest that it was an inhabited site for an extended period of time. If the interpretations of

Greenwich Bay's marine inundation can be bracketed at approximately 6000 years BP, and estimates for the oldest Paleo-Indian artifacts are accurate (~10,000 years), then at least 4,000 years of human history occurred while the fluvial paleolandscape was exposed in Greenwich Bay. Therefore, understanding this landscape is necessary for determining where potential habitation sites would have been, and this information will be used when predicting where they might be submerged today.

While it is useful to visualize landscapes in two-dimensions, the use of three-dimensional visualization software can provide the next step in understanding an underlying paleolandscape. IVS7 Fledermaus software was used to render three-dimensional paleolandscape visualizations for the interpolated Depth to Glacial map as well as the Depth to Pre-Inundation Surface map. The use of three-dimensional visualizations will not only help to better understand the limitations of our seismic data interpolations, but is also valuable for portraying seismic data to non-geologists (i.e. archaeologists).

2.4.6. A Proposed Connection Between the Modern and Paleo-Landscape

The focus of this work is to reconstruct and identify paleolandscapes that are currently submerged, but were previously sub-aerially exposed, and could have potentially been areas supporting human habitation. However, by investigating the characteristics of the modern landscape, a connection between the modern terrestrial environment and the submerged paleolandscape can be hypothesized. The Quaternary surficial cover map of Smith (1955) was used to suggest possible drainage routes that may have been the source for the paleo-stream observed in the Pre-Inundation Surface

landscape south of Cedar Tree Beach (Figure 2.14.). Within the topographic lows between modern day Pawtuxet River and Apponaug Cove are significant glacial meltwater-deposited kame terrace beds (Qktx), as well as undifferentiated sand and gravel deposits (Qsu) that are associated with local or temporary drainage patterns. Although it is speculative at the present, the topography and Quaternary sediment cover may indicate that the paleo-drainage course of the Pawtuxet River might have been the source of water that shaped the pre-inundation landscape discussed in this chapter.

Conclusions

- Twenty-four seismic sub-bottom (CHIRP) lines were investigated in Greenwich Bay, RI, in order to determine paleolandscape changes since deglaciation. Four seismic discontinuities were found and mapped in each seismic image, and five seismic facies were interpreted. These facies made up three distinct depositional environments observable in the seismic record.
- Depositional environment 1 was interpreted as a proglacial lake environment. Sediments in this environment were thick (>40 meters) proximal and distal glaciolacustrine varves. These interpretations are in agreement with other literature, and this environment has been mapped extensively throughout Narragansett Bay in other studies. These sediments were likely laid down within Glacial Lake Narragansett, which formed following retreat of the Laurentide Ice Sheet from modern day Narragansett Bay.
- Depositional environment 2 is interpreted as a stream-dissected landscape that occurred following the catastrophic draining of GLN. The upper boundary of the proglacial lake environment is truncated by a fluvial erosional unconformity. Paleochannel structures and deposits are observed throughout Greenwich Bay, and we interpret that a number of paleo-streams drained on top of the subaerially-exposed glacial lakefloor sediments. Of particular interest is a paleochannel structure that passes to the south of modern day Cedar Tree Beach, as that is the

location of a number of Paleo-Indian projectile artifacts that date back to the age of this paleolandscape. These results will be used in the development of an archaeological site prediction model, for the purpose of identifying potential submerged paleocultural habitation sites.

- Depositional Environment 3 is an estuarine environment that resulted from marine transgression into Greenwich Bay. A marine transgressional unconformity, known as the ravinement, marks the upper boundary of the fluvial paleochannel deposits. Estuarine sediments average ~7-8 meters in thickness throughout the bay.
- A Local Polynomial Interpolation model in ArcMap 10 was used to create two-dimensional and three-dimensional surfaces of the proglacial lakefloor paleolandscape, as well as the paleolandscape surface that existed just prior to marine inundation. The methods presented in this thesis were utilized successfully to create a preliminary paleolandscape reconstruction for Greenwich Bay prior to marine inundation. However, the 300m trackline spacing used for sub-bottom profiling necessitated significant interpolation to visualize the paleolandscape surface between survey lines. The LPI model performed well in that it captured the general geometry of the landscape observed in the seismic images. However, additional method development needs to be conducted to understand the areas where large flat polynomials were output by the LPI, and further investigation into the organizational binning of the geospatial data. Based on the research presented

in this thesis, 100 meter grid trackline spacing is recommended for additional surveying.

- An investigation of the Quaternary surficial cover in the East Greenwich Quadrangle may help connect the exposed topographic landscape to the submerged pre-inundation landscape. Kame terrace deposits and undifferentiated sand/gravel beds are observed in topographic lows between the Pawtuxet River and our interpolated pre-inundation surface, which might suggest a paleo-drainage pattern of the Pawtuxet in the Late Pleistocene/Early Holocene.

Figure 1.) A.) Showing the four sites associated with the Paleocultural Landscapes Project; inset is the Gorton Pond/Greenwich Bay site; B.) Showing a sample of the projectile points discovered in the region.

Figure 2.2.) A. Showing the regional location of Greenwich Bay, and B. the location in reference to Narragansett Bay; C. Satellite image of Greenwich Bay with seismic survey lines superimposed, and line numbers indicated.

Figure 2.3.) Greenwich Bay Line 9 seismic reflection profile and interpretation of seismic reflectors. Depth in meters based on water and sediment velocity of 1500 m/s.

Figure 2.4.) Greenwich Bay Line 12 seismic reflection profile and interpretation of seismic reflectors. Depth in meters based on water and sediment velocity of 1500 m/s. See text for explanation of unconformities.

Figure 2.5.) Greenwich Bay Line 13 seismic reflection profile and interpretation of seismic reflectors. Depth in meters based on water and sediment velocity of 1500 m/s. See text for explanation of interpreted unconformities.

Figure 2.6.) Greenwich Bay Line 16 seismic reflection profile and interpretation of seismic reflectors. Depth in meters based on water and sediment velocity of 1500 m/s. See text for explanation of interpreted unconformities.

Figure 2.7.) Greenwich Bay Line 19 seismic reflection profile and interpretation of seismic reflectors. Depth in meters based on water and sediment velocity of 1500 m/s. See text for explanation of interpreted unconformities.

Figure 2.8.) Showing the extent of seismic reflector coverage. Panel A shows all CHIRP coverage, bathymetry of Greenwich Bay, and the seismic data bins used. Panel B shows all CHIRP line numbers and location of Acoustic Bottom (AB) reflector coverage. Panel C shows CHIRP coverage and all locations of ravinement (R1) coverage. Panel D shows all locations of paleochannel cuts superimposed on CHIRP line coverage. All panels in 1 : 90,000 map scale. Bathymetry dataset comes from rasterized NOAA datasets, and basemaps are standard in ArcMap 10.0.

Figure 2.9.) Showing the sub-bottom depth to acoustic basement (AB) surface constructed by digitally stripping all sediments penetrated by the CHIRP signal.

Figure 2.10.) Showing the sub-bottom depth to ravinement surface constructed by digitally stripping all post-marine inundation sediments from the early Holocene fluvial surface. Interpreted paleo-stream thalwegs are indicated by red lines. Rectangles indicate the “large polynomial” problem mentioned in text. Linear artifacts mentioned in text are also indicated.

Figure 2.11.) Showing the extent and thickness of mapped sediments in the proglacial lake depositional environment.

Figure 2.12.) Fledermaus scene showing a three-dimensional rendition of the Pre-Inundation Surface interpolation. A georeferenced image of seismic line 16 is draped within the scene. The large polynomial interpolation problem is shown in this figure, as the blue region in the forefront of line 16 is an incorrect 9.5 meters depth for ~300 meters width. Despite this problem, the LPI product still captures useful horizontal changes in the paleolandscape.

Figure 2.13.) Fledermaus scene showing a three-dimensional rendition of the area between the Depth to Glacial and Depth to Pre-Inundation Surface interpolations. The large polynomial problem in the interpretation model is shown well in three-dimensions, as the purple area indicated by the white arrow does not properly capture the paleochannel reflector in line 16. The acoustic basement reflector is captured appropriately, as is the ravinement surface outside of the paleochannel area.

Figure 2.14.) Fledermaus scene showing a georeferenced Quaternary surficial cover map (Smith, 1955) in relation to our interpolated Depth to Pre-Inundation Surface. Kame terrace (Qktx) deposits are represented by very light green, and local/temporary drainage sands/gravels (Qsu) are represented by light purple. A suggested exposed-to-submerged landscape is outlined in red.

Table 2.1. Seismic line information for CHIRP data in Greenwich Bay.

Line #	SB File Name	Seismic Bin(s)	Heading	Length (meters)	Mapped Reflector Extent	
					R1 (%)	AB (%)
1	SB_009000	1	NW	226.5	0	0
2	SB_008000	1	SE	464.1	0	0
3	SB_007000	1	NW	425.8	0	0
4	SB_006000	1	NW	486.8	94.8	89.5
5	SB_005000	1	SE	485.6	76.7	97.2
6	SB_004000	1	NW	711.5	98.7	97.4
7	SB_003000	1	SE	778.4	98.6	98.6
8	SB_002000	1, 2	NW	781.0	90.0	97.9
9	SB_001000	1, 2	SE	825.1	80.0	87.8
10	SB_015000	2,3	SE	1185.2	88.6	96.5
11	SB_016000	2,3	NW	1413.5	62.4	38.6
12	SB_017000	2,3	SE	1568.9	73.4	82.5
13	SB01000	3	NW	2272.0	93.1	99.3
14	SB02000	3	SE	2268.1	93.7	96.3
15	SB03000	3,4	NW	1829.2	56.8	98.5
16	SB04000	3,4	SE	1388.5	71.1	97.0
17	SB05000	4,5	NW	1843.9	58.4	84.6
18	SB008000	3	ESE	1599.2	50.2	86.2
19	SB007000	4,5	SE	2213.3	47.7	75.1
20	SB006000	5	NW	2417.4	46.0	62.2
21	SB005000	5	SE	2474.2	42.3	54.4
22	SB003000	5	NW	2396.6	38.7	39.5
23	SB002000	5	SE	2308.9	29.2	24.2
24	SB001000	5	NW	2421.0	63.3	29.4

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