A Mid-Late Holocene Multi-Proxy Paleoenvironmental Reconstruction of Northern Finnmark Using a Sediment Core from the Island of Ingøy, Norway

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A Mid-Late Holocene Multi-Proxy Paleoenvironmental Reconstruction of Northern Finnmark Using a Sediment Core from the Island of Ingøy, Norway

Bates College Geology Department Thesis

Presented to the Faculty of the Department of Geology, Bates College, in partial fulfillment of the requirements for the Degree of Bachelor of Science

by

Claire Lockwood Markonic

Lewiston, Maine
April 7th, 2017
Abstract

The Arctic is currently responding to the current warming to a greater degree than any other region on the planet (Serreze and Barry, 2011). For accurate predictions to be made regarding the response of the Arctic to current and future fluctuations in climate, a compilation of high-resolution marine and terrestrial paleoenvironmental records encompassing the entire region must be compiled. Through a multi-proxy approach, this study aims to reconstruct past environmental conditions of northern Finnmark using combined geochemical, physical and biological analyses. A 65 cm sediment core was scanned for downcore elemental profiles using an ITRAX-XRF core scanner at 500µm resolution, while stable carbon and nitrogen isotopes (δ\textsuperscript{13}C and δ\textsuperscript{15}N), C/N molar ratios, percent carbon (%C) and percent nitrogen (%N), percent loss on ignition (%OM), chlorophyll, magnetic susceptibility and grain size were subsampled for at 1 cm resolution.

An age-depth model spanning 10,500 cal years BP was developed using four AMS radiocarbon dated terrestrial macrofossils. However, based on the well-documented sea level history of the island of Ingøy, and the failure to identify a marine-lacustrine boundary within Core #1 the age-depth model was deemed invalid. The lack of a reliable age-depth model prevented the reconstructed regional paleoclimate and paleoenvironmental conditions of northern Finnmark from being linked to large-scale patterns in climate variability.

Paleoenvironmental conditions within the lake based on the combination of biological and geochemical analyses were characterized by high rates of primary productivity. High percent organic content sourced from C\textsubscript{3} terrestrial plant material makes up the majority of the sediment composition. Higher counts of Fe/Mn seen in the bottom half of Core #1 suggests that reducing conditions, resulting in anoxic waters and greater counts of elemental sulfur, dominated the later half of the sediment record. Deposits of eolian transported sand and mica detritus periodically interrupt the gyttja matrix and subsequently dilute the biological and geochemical organic signals.
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Introduction

Paleoclimate reconstructions spanning the Holocene have become abundant in the scientific literature due to recent concerns surrounding the potential effects of anthropogenic climate change (Wanner et al., 2008). In order to differentiate between anthropogenic forcing and natural climate variability a historical record of climate that extends prior to the instrumental period must be established. The late Holocene epoch, specifically the past 1,500 years, has been characterized by intervals of climate anomalies including the Little Ice Age (1,400-1,700 A.D.) and Medieval Climate Anomaly (950-1,250 A.D.) (Mann et al., 2009). Although, the effects of these events are well documented on regional as well as global scales, the specific mechanisms that operate to produce such dramatic fluctuations in climate are still poorly understood (Marcott et al., 2013; Mayewski et al., 2004; Wanner et al., 2008). For the sake of future generations, it is important to continue to develop what is currently known about the various feedback mechanisms contributing to climate variability. The establishment of a comprehensive record of scientific evidence can be used to guide public response to the changing climate, not only in the years to come but, more importantly, in the present day. One such region that has been drawing special attention due to its recent response to present day climate dynamics is the Arctic.

The Arctic is currently responding to the current warming to a greater degree than any other region on the planet (Serreze and Barry, 2011). This Arctic amplification is caused by a number of feedbacks that operate within the climate system. The most prominent of these include the albedo effect, retreat and expansion of sea ice cover, cloud cover, aerosol loading, soot covered snow, and changes in atmospheric and oceanic circulation patterns (Serreze and Barry, 2011). Arctic amplification is expected to become stronger as these various processes operate in conjunction with one another. For example, an increase in the circulation of Atlantic Water (AW) into the northern latitudes, entering the Arctic along the coast of Norway as well as through the Fram Straight, has been observed over recent decades. The North Atlantic Current (NAC) provides the main transport of warm water into the Arctic Ocean, as it is an extenuation of the Gulf Stream. Recently, an
unprecedented heat flux as well as an increase in volume of AW has been recorded in the Arctic (Hald et al., 2011; Polyakov et al., 2013; Polyakov et al., 2010; Spielhagen et al., 2011). Effects of this trend include a reduction in the ice extent of the Polar Ice Cap. According to the National Snow and Ice Data Center (NSIDC) on March 7th of this year sea ice reached its maximum extent at 14.42 million km$^2$, the lowest ever recorded in the 38-year satellite record. This maximum is 97,000 million square kilometers below the previous lowest maximum that occurred on February 25th, 2015.

Due to the extreme climate variability of the Arctic region, and the fact that it has demonstrated some of the strongest feedback mechanisms recorded in comparison to any other area on Earth, it is an ideal setting to research past as well as recent climate change. The goal of this study is to create a paleoenvironmental record of the late Holocene through a multi-proxy analysis of a lake sediment core taken from an isolation basin located on the island of Ingøy, Norway. The Arctic island is situated just off the northernmost coast of Finnmark County at approximately 71°N latitude (Figure 1.1). Due to its unique hydrographic location, the climate of the Ingøy is influenced by a series of complex mechanisms that underlie regional climate variability (Allen et al., 2007).

**Regional Hydrography**

As mentioned above, Ingøy is located within a dynamic region of the North Atlantic. The occurrence of three ocean current systems is observed off of the northern coast of Norway. These include the North Atlantic Current (NAC), the Arctic Current and the Norwegian Coastal Current (NCC), which propagate and mix within the Barents Sea (Figure 1.6) (Polyakov et al. 2013).
Figure 1.1: Map of the oceanographic setting of the island of Ingøy (yellow star). Red arrows depict the path taken by the NAC warm water current, while the Arctic Waters are shown in blue (Polyakov et al. 2013).

The NCC is a surface water current, while the Arctic Current and NAC are part of a deep water system that surrounds Ingøy (Slagstad et al., 1990). The interaction of these currents is dynamic and changes in circulation can result in shifts in patterns of ocean oscillations.

One such oscillation is known as the Atlantic Multi-decadal Oscillation (AMO). Multi-decadal in scale and based on changes in SST, the AMO is believed to be driven by fluctuations in surface heat flows, wind driven ocean currents, fluctuations in sea ice and, more specifically, changes in the Atlantic Meridional Overturning Circulation (AMOC) (Pinto and Raible, 2012; Slagstad et al., 1990). The AMOC is characterized by a northward flow of warm water that exists as a surface current and the return of cold, deep-water to the south. As warm, saline water moves north it loses heat and increases in density, which causing it to sink, also resulting in a decrease in surface salinity. The
overturning circulation occurs in the Northern Hemisphere between the boundary of the NAC and the Arctic Current (Naafs et al., 2010). Because the AMO is characterized by changes in SST, AMO- phases refer to cold temperatures while AMO+ phases refer to warm temperatures.

Determining correlations between multiple proxy records can serve to strengthen chronological constraints and reduce uncertainty when reconstructing past climate. This study focuses on terrestrial and marine ecosystem response to climate forcings and feedbacks in the late Holocene to recent times. A mid-late Holocene paleoenvironmental reconstruction is presented developed from a multi-proxy analysis of the physical sedimentology, geochemical analysis, and biological record of a sediment core recovered from a coastal isolation basin located on the island of Ingøy, northern Finnmark County, Norway.

Ecological Response to Climate Variability

Although many land-based paleoclimate records exist for the Arctic, marine-based proxies have been increasingly employed in an attempt to better understand the relationship between oceanic and atmospheric circulation patterns, the combination of which exhibit a profound impact on the climate variability of the region. The most recent work in this area has utilized the marine bivalve, *Arctica islandica*, collected off the coast of Ingøy and Rolvsøya, to reconstruct a marine ecosystems response to climate change (Mark, 2016; Mette et al., 2016; Savage, 2016; Wanamaker et al., 2011).

Through the investigation of the relationship between shell growth, sea surface temperatures (SST) and δ\(^{18}\)O isotopic composition, a 113-year master shell growth chronology was developed in order to strengthen correlations between external climate forcings and biological responses of the bivalves (Mette et al., 2016). It was discovered that shell growth displays a strong inverse relationship to modern day sea surface temperatures of the North Atlantic as well as shell δ\(^{18}\)O isotopic composition.
oceanographic influence of the NAC on the benthic community was found to correspond to changes in Atlantic Meridional Overturning Circulation (AMOC), Atlantic Multidecadal Oscillation (AMO) and the North Atlantic Oscillation (NAO). Narrower growth increments were observed during warmer years while the bivalves were recorded to have grown more during colder years (Mette et al., 2016). This phenomenon was attributed to an increase in zooplankton abundance during warmer periods or AMO positive phases. An increase in the primary consumer of phytoplankton, *Arctica*’s main food source, is inferred to have led to the subsequent depletion of food availability for the bivalve during the summer, eventually resulting in smaller increments of growth. Therefore, an ecological response to changing climate as a result of variability in ocean and atmospheric circulation patterns has been documented for the region. However, the terrestrial response to such variability is still poorly recognized.

The goal of this current study is to develop a terrestrial-based multi-proxy paleoclimate and paleoenvironmental reconstruction from lake sediments on the island of Ingøy, which will establish the basis for comparison with late-Holocene shell-based marine constructions (Mette et al., 2016, Wanamaker et al., 2011). The development of a more complete understanding of the Arctic climate through the late Holocene is imperative in order to evaluate future impacts of increased warming, not only on marine and terrestrial ecosystems, but our climate system as a whole.

**Using Lake Sediments to Infer Paleoclimate**

*Multi-Proxy Reconstructions*

Lake sediments are commonly used to produce high-resolution records of past climate in environments around the world. Lake sediments contain materials originating from the surrounding landscape, the lake itself as well as the atmosphere. The deposition and subsequent preservation of these various inputs can be used to infer local, regional and global paleoenvironmental and paleoclimatological conditions (Williamson et al., 2009).
Because a number of complex interactions take place within lake ecosystems, one way to strengthen paleolimnological interpretations inferred from a single proxy is to combine as many records as possible in a technique known as a multi-proxy approach (Birks and Birks, 2006; Bradley, 2015). For example, Jiang et al., (2011) were able to perform a detailed multi-proxy analysis on a lake sediment core from Ny-Ålesund, Svalbard, reconstructing a record of environmental changes for the past 3,000 years. Proxies that were examined in their study included organic matter content, C/N.ratios, CaCO$_3$, magnetic properties, organic carbon and nitrogen isotopes, sediment pigments and diatom assemblages. Each record demonstrated consistent trends throughout the LIA cold period (A.D. 1,420-1,850), during which algal production diminished and led to a decrease in lake primary productivity (Jiang et al., 2011).

Although multi-proxy analysis can be used to produce valuable, high-resolution paleoclimate data, the complex interactions that occur within an ecosystem may be heavily influenced by regional effects, such as seasonal changes, which could lead to results that are misrepresentative of the climate system as a whole. For example, a multi-proxy reconstruction spanning the past 15,000 years was achieved using lake sediments from Lake Nattmålsvatn, northern Norway (Janbu et al., 2011). A large amount of uncertainty was determined to surround the climate signals inferred from $\delta^{13}$C and $\delta^{15}$N, due to the fact that ice cover would isolate the lake for large parts of the year. Therefore, the sensitivity of the lake to changing environmental conditions was reduced (Janbu et al., 2011).

Geologic History of Northern Finnmark

*Bedrock Geology*

The bedrock geology of northern Norway is part of the Kalak Nappe Complex (KNC) (Kirkland et al., 2007). This includes a series of nappes, or highly metamorphosed and eastward thrusting allochthonous rocks. The KNC was recently discovered to consist of two separate successions distinguished as the “upper” and “lower” nappes. The island of
Ingøy is part of the upper nappe complex, which is comprised of the Sørøy-Seiland Nappe and Havvatnet Imbricate Stack. These metasediments were affected by the Porsanger Orogeny around 840 Ma during the second succession in the late Precambrian (Kirkland et al., 2007). The bedrock of my study area on Ingøy is classified as the Storelv Schist (Figure 1.2). Understanding the bedrock geology of the study site is important as the different sources of sediment that enter the isolation basin are considered and can be identified through various analyses including ITRAX X-ray fluorescence, isotope analysis, and magnetic susceptibility.

**Figure 1.2:** Bedrock geology of Finnmark, Northern Norway. The bedrock type for the island of Ingøy, as part of the upper nappe complex or Sørøy Sucession (S), is shown in red (adapted from Kirkland et al., 2007).

**Glacial History**

The island of Ingøy has been subjected to numerous periods of glacial fluctuation over the past 40 ka. The most recent peak in global ice volume occurred around 23,000-21,000 cal yr BP, during what is known as the Last Glacial Maximum (LGM) (Hughes et al., 2015). At the time of the LGM the Svalbard-Barents-Kara Ice Sheet (SBKIS), extending over the Barents Sea continental shelf, the Fennoscandian or Scandinavian Ice Sheet...
(SIS), comprising the nucleus of the ice sheet, and British-Irish Ice Sheet (BIIS), overlying the British Isles, were conjoined to form the Eurasian Ice Sheet (Hughes et al., 2015; Winsborrow et al., 2010) (Figure 1.3). Glacial retreat became very rapid from 19,000-14,000 years BP, culminating at the onset of the Bølling chronozone. Radiocarbon dates of mollusk shells and macroalgae from lake sediments have constrained the timing of deglaciation of the northernmost coast of Finnmark to be around 14,600 cal yr BP (Romundset et al., 2011) (Figure 1.4).

Uniquely situated at the boundary of the SBKIS and SIS, the island of Ingøy was influenced by events including isostatic rebound and eustatic sea level rise, which occurred throughout the Holocene. As glacial retreat continued, two distinct post-glacial shorelines were created that are continuous across northern Finnmark and can be correlated across all of western Norway. The first shoreline formed during the Younger Dryas cool period (11,000 cal yr BP) and is known as the ‘Main’ shoreline.
The rapid cooling event led to a decrease in sea level as the SIS expanded. The Main shoreline is visible across the island of Ingøy and is situated at approximately 30 meters absl. Sea level during the Younger Dryas was significantly higher than it is today due to the fact that the recent glaciation had caused isostatic depression of the earth’s crust.

The second paleoshoreline formed during what is known as the ‘Tapes Transgression, which occurred between 6,000-8,000 cal yr BP (Donner et al., 1977; Romundset et al., 2011). Sea level had been steadily dropping due to isostatic rebound that began to take place following the Younger Dryas. However, this was reversed when a sudden rise in sea level occurred as a result of an increase in the input of meltwater during the deglaciation of the Laurentide Ice Sheet as well as Antarctica (Wohlfarth et al., 2008). This global increase in eustatic sea level led to the formation of large beach ridges that accumulated during the peak of the mid-Holocene Transgression (Romundset et al., 2011).

Figure 1.4: Deglaciation of the Eurasian Ice Sheet from 19 ka to 14 ka BP. The SBKIS, SIS, and BIIS have completely separated by 15 ka (Hughes et al., 2015).

Paleoshorelines that can be traced along the coast of northern Norway have been examined and dated extensively as evidence of glacial and post glacial phenomenon (Donner et al., 1977; Hald and Vorren, 1983; Lambeck et al., 1990; Long et al., 2011; Marthinussen, 1962; Sollid et al., 1973). Most recently, Romundset (2011) conducted a study in which the deglaciation history of the outer coast of Finnmark was examined through the development of a number of sea level curves. According to this study, the Tapes paleoshoreline on Rolvsøya was observed at approximately 4 m absl based on
research on several isolation basins. However, new ages from *A. islandica* shell material obtained from paleoshorelines on the island of Ingøy establish the maximum elevation of the transgression at 9-10 m absl and the marine limit at 30 m absl (Retelle et al., 2016, *unpublished*) (Figure 1.5). Following the Tapes Transgression, sea level continued to fall to its present day level as isostatic rebound of the landscape resumed.

**Figure 1.5:** Relative sea-level curve developed by Retelle et al. 2016 for the island of Ingøy, Norway. The Tapes transgression is shown as the period between 6,000-8,000 cal yr BP (9-10 m asl) and a green diamond marks marine limit at approximately 14,600 cal yr BP (30 m asl).
Methods

Field Methods

A set of 3 cores were recovered from a small isolation basin located to the north of Sanden Bay, just west of Saeterbukta Cove, on the island of Ingøy, Norway (Figure 2.1). The first core (Core #1, GPS location: 71.07616ºN, 24.085667ºW) was retrieved on August 20th and the second and third cores (Core #2 GPS: 71.067556ºN, 24.085861ºW and Core #3 GPS: 71.067333ºN, 24.085667ºW) were retrieved on August 23rd using a universal percussion corer. A small rowboat was used to navigate the shallow lake while water depth was measured using a plumb line marked in meters. Pockets of dense vegetation were visible from the surface within the southernmost area of the lake and water depth remained relatively shallow (<1m) throughout.

The sediment recovered in each core using a universal percussion corer averaged approximately 60 cm. The universal corer includes a 1 meter-long polycarbonate core barrel that attaches to a core head via a rubber boot coupler. The core head contains a valve that opens once the core is inserted into the sediment, allowing for water and mud to flow through the core head without disrupting the sample. A cylindrical slide hammer weight is moved up and down the aluminum holobar extension rod via a non-stretch rope to gradually drive the core into the sediment. As the core is pulled free from the lakebed the core head valve seals and creates a vacuum, effectively holding the sample inside of the core barrel. The core is capped with a poly end cap in the water before being lifted into the boat to ensure no sediment is lost and then sealed with electrical tape.

In order to prepare the cores for transport back to the U.S., excess water was decanted from the top of the cores by drilling a hole approximately 5-8 cm above the sediment-water interface into the polycarbonate tube. The hole was taped shut and the superabsorbent polymer, sodium polyacrylate, was poured into the remaining layer of water in order to ensure the sediment and sediment-water interface would remain undisturbed during transport to Bates College.
Figure 2.1: Top: A map of the island of Ingøy with the field site outlined by a yellow circle. Bottom left: Satellite image of the isolation basin from which Core #1 was recovered. Bottom Right: Image of the isolation basin with the coring location denoted by a yellow diamond.
Laboratory Methods

Core Preparation

Each core was split using a table saw calibrated to cut through the polycarbonate tubing without disturbing the sediment. After each tube was sliced on two adjacent sides a fishing-line was inserted into the slits and pulled lengthwise through the core, beginning at the base, in order to evenly split it into two halves: an archive half for nondestructive lab analyses and a working half for destructive analyses. These were then covered with plastic wrap, labeled and sealed in a plastic sleeve, and placed in plastic D-tubes to be stored in a refrigerator core locker at Bates College. After splitting the cores it was decided that only Core #1 would be analyzed due to the fact that it was the longest of the three cores recovered.

Subsampled Analyses

Loss on Ignition

A measurement of the organic matter content (%OM) was obtained through loss-on-ignition (LOI) analysis. Core #1 was sampled at 1 cm intervals using a 1cc hollow aluminum cube. The cube was pushed into the split core surface until the top was flush with the sediment surface and removed using stainless steel tools shaped to facilitate the extraction of the sample. The sampled sediment was placed into pre-weighed porcelain crucibles. Weighing these crucibles and subtracting the weight of the empty crucible yielded the mass of the wet sediment or wet bulk density (g/cc). The wet samples were then placed into a Fisher Isotemp 100 series oven at 100°C to dry overnight. The next day, the samples were removed from the oven, cooled, and reweighed to yield dry bulk density (g/cc). The dry crucibles were then placed into a Thermolyne 6000 muffle furnace for 1 hour at 550°C, during which any organic matter contained within the sediment was oxidized to carbon dioxide and ashed. These samples were cooled in a dessicator and weighed a final time so that the percent organic matter loss-on-ignition
could be calculated using the equation below. All calculations used to determine the
downcore values of percent organic content were performed in Microsoft Excel and then
transferred to Sigma Plot 13.0 where they were plotted.

Equation 2.1: %LOI Determination

\[
\%LOI = \frac{(Dry\ Mass - Roasted\ Mass)}{(Dry\ Mass)} \times 100
\]

Grain Size

Grain size was analyzed using a Beckman Coulter LS 13 320 Laser Particle Size
Analyzer (LPSA) (Figure 2.2) located at Bates College. The working halves of Core #1
was sampled at 1 cm intervals using a 1cc hollow aluminum cube. The subsampled
sediment was placed into a 50 mL Oak Ridge centrifuge tube and covered with 5 mL of
30% Hydrogen Peroxide (H\textsubscript{2}O\textsubscript{2}), which was pipetted into each tube in order to
chemically disintegrating any organic material contained within the sample. Distilled
water was added to bring the tubes to approximately 10 mL and a glass stirring rod was
used to break up aggregated sediment. The centrifuge tubes were kept uncapped in a rack
and placed a Precision Water Bath Model 183 set to 50ºC for five hours. After five hours
had passed, the water bath was tuned off and the samples were left to sit overnight.

The next morning, deionized water was added to fill the tubes completely. The tubes
were placed in an IEC CU-5000 centrifuge at 2000 rpm for 10 minutes and the
supernatant was decanted. Distilled water was again added to fill the tubes. Each tube
was placed on a Vortex Genie and dimembrated using a Fisher Scientific 60 Sonic
Dismembrator for 30 seconds. The centrifuge process was then repeated and supernatant
decanted. Next, 15 mL of a 0.7 g/mL sodium metaphosphate dispersant solution was
added to each sample tube, which was again placed on the Vortex Genie and sonically
dismembrated for 30 seconds. Samples were centrifuged once more at 2000 rpm for 10
minutes, supernatant was decanted and 5 mL of dispersant was added to each tube. Prior
to loading into the LPSA, each sample was shaken on the Vortex Genie and sonically dismembrated for 1 minute in order to deflocculate any particles that could produce a misrepresentative grain size reading.

The prepared samples were immediately loaded into the sample well of the Coulter Particle Size Analyzer. The Coulter LSPA measures a range of grain sizes from 0.4 to 200 microns through the analysis of diffracted laser light. Each sample was run three times and the final run was used to characterize the mean grain size of the sampled sediment.

![Diagram of the Beckman Coulter LS 13 320 Laser Particle Size Analyzer (LPSA) located in the Bates College Quaternary Laboratory.](image)

**Figure 2.2:** Diagram of the Beckman Coulter LS 13 320 Laser Particle Size Analyzer (LPSA) located in the Bates College Quaternary Laboratory.

**Stable Isotope and Elemental Analysis**

Stable isotope measures in organic matter can be used as indicators of paleoclimate variation and paleoproductivity within and around lake systems (Janbu et al., 2011; Torres et al., 2012). C/N ratios in lacustrine organic matter are altered and controlled by terrestrial vs. aquatic plant sources. Terrestrial organic matter will have a C/N ratio of >20 while lacustrine algae will display a ratio between 4-10 (Meyers and Lallier-vergés, 1999).
Core #1 was sampled at 1 cm intervals using a 1cc hollow aluminum cube starting from
the top of the core (0-1cm) to a depth of 65 cm. Samples were dried in porcelain crucibles
for four hours at 70ºC in a Precision Oven, model no. 5122133. After drying, samples
were homogenized within their crucibles using an agate pestle. The homogenized
sediment corresponding to each cm depth interval was measured out on a microbalance in
5x9 mm tin cups. These were then folded into themselves in order to prevent sediment
loss during analysis, which was conducted using a Costech Elemental Analyzer (EA)
interfaced via combustion to a Delta V Advantage Isotope Ratio Mass Spectrometer in
the Bates College Environmental Geochemistry Lab. Between measurements all stainless
steel utensils and instruments were wiped clean with Kimwipes® to avoid cross-
contamination. The EA measures relative peaks in nitrogen and carbon content for each
sample. Standards of known isotopic composition including freeze-dried cod muscle
tissue, acetonilide, and caffeine were prepared and were run interspersed throughout the
sediment samples along with a blank tin cup for the first run. Approximately 2-5 mg
subsamples of sediment were used for organic rich material and 12-18 mg were used for
organic poor material.

Data obtained from this analysis were presented in micromoles of percent total nitrogen
(\%N) and percent total carbon (\%C), $\delta^{13}$C and $\delta^{15}$N. C/N molar ratios were also
calculated from %N and %C. All calculations were performed in Excel and plotted in
SigmaPlot 13.0.

**Chlorophyll Analysis**

Total chlorophyll-\(\alpha\) pigments for Core #1 were analyzed using a Turner Design 10-AU
fluorometer. The fluorometer utilizes blue wavelengths of light in order to produce a
fluorescence of chlorophyll-\(\alpha\) in the red part of the electromagnetic spectrum. Core #1
was subsampled at 1 cm resolution using a 1cc hollow aluminum cube during which
special care was taken to prevent the sediment from being exposed to light.
To extract pigments, 10 mL of 100% acetone was added to each sample. The samples were sonicated for 30 seconds and then capped and stored in the freezer overnight. After 12 hours, samples were centrifuged at 2,000 RPM for 5 minutes, after which approximately 9 mL of pigment extract was transferred to a glass cuvette. A cuvette containing 100% acetone was used to blank the Fluorometer and the absorbance of each sample was read at 663 nm. In order to degrade active pigments to phaeopigments so that total pigment concentrations could be calculated samples were acidified with 3 drops of 6M HCl after the initial reading. Chlorophyll-α and phaeopigments were determined using the following equations:

**Equation 2.2: Active chlorophyll-α pigment concentration determination**

\[
\text{Chlorophyll-α (µg/L)} = ((r/r-1) \times (F_0 - F_a)) \times (\text{dilution factor}) \times (\text{acetone vol.}) \times (\text{sediment vol. (mL)})
\]

**Equation 2.3: Phaeopigment concentration determination**

\[
\text{Phaeopigments (µg/L)} = ((F_a \times r) - F_0) \times F_s \times (r/(r-1))
\]

Where \( r \) is the acidification coefficient \( (F_0/F_a) \) for pure Chlorophyll-α (1.9848), \( F_0 \) is the reading before acidification, \( F_a \) is the reading after acidification, \( F_s \) is the slit setting conversion factor (1), the dilution factor is 100 and the acetone volume is 8 mL (.008 L).

Once the total pigment value for each sample were determined by taking the sum of the concentrations of both Chlorophyll-α and phaeopigments, they were converted to µg/g using the average bulk density of Core #1. Because 1 cc of sediment is equal to one cm³, the total pigment concentrations were converted by multiplying average bulk density to obtain values in µg/g.
Geochronology

AMS Radiocarbon

Radiocarbon ($^{14}$C) is produced in the atmosphere through neutron bombardment of nitrogen atoms. The $^{14}$C atoms become oxidized to $^{14}$CO$_2$ and mixes in with the already existing carbon dioxide, effectively entering into all pathways of the biosphere (Bradley, 1999). Over geologic time, the rate of $^{14}$C production in the upper atmosphere has become equilibrated with the rate of its decay in the global carbon reservoir. Thus, the total weight of global $^{14}$C remains constant. For example, plants uptake CO$_2$ through photosynthesis, which also serves to incorporate $^{14}$C into their tissues. Their tissues then become balanced with the $^{14}$C content of the atmosphere as cells die and are replaced. When the organism dies, this process comes to a halt and the $^{14}$C content declines. This decline occurs at a negative exponential rate and can be described as a function of time. From this a relative age for organic matter can be determined as all living things assimilate $^{14}$C into their tissues (Bradley, 1999).

Plant macrofossils were collected from the working half of Core #1 as they were revealed through subsampling and logging. A total of 4 samples were collected. The samples were observed under a microscope, placed in vials labeled with their corresponding depths in cm and sent to DirectAMS in Seattle, WA to be AMS radiocarbon dated.

Lead-210 ($^{210}$Pb) and Caesium-137 ($^{137}$Cs)

After the archive half of Core #1 was sampled for chlorophyll analysis the remaining top 30 cm of sediment was collected in 1 cm increments, placed in a 250 mL glass beaker and kept in a Fisher Isotemp 100 series oven at 100ºC to dry overnight. After drying, the samples were bagged, labeled and sent to Mark Baskaran at Wayne State University to be analyzed for Pb-210 and Cs-137.
Non-Destructive Analysis

*Downcore Stratigraphy*

Photographs of the archive half of Core #1 were taken with a NIKON D800E digital camera in the Imaging Center at Bates College. Descriptions of the downcore stratigraphy, including color, lithology and corresponding depth intervals were compiled. A symbolized stratigraphic column was produced in Sigma Plot 13.0 and placed alongside the high-resolution image of Core #1. The classification scheme outlined by Schnurrenberger et. al., (2003) and the Munsell Soil Chart number 5Y were used to guide descriptions of the lacustrine sediment.

*Magnetic Susceptibility*

Measurements of sediment magnetism through magnetic susceptibility (MS) analysis can be used to infer the input of minerogenic material into the lake. Although MS provides information regarding the concentration of iron-bearing magnetic minerals, other analyses must also be employed in order to fully reconstruct the past depositional environment of the catchment. MS was measured using a Bartington MSE2 sensor in the Bates College Quaternary Lab.

The split archive half of each core was taken out of the cold room and left in the lab overnight to warm the to room temperature, which prevented temperature fluxes from having an influence on the magnetic properties of the sample (Oldfield, 1999). The plastic wrap in which the cores were initially stored was marked at 0.5 cm intervals with a sharpie marker. Measurements were taken every 0.5 cm starting at the top of the core with the point sensor set to 0.1 SI units. The compiled measurements were corrected for drift using the Multisus software. All results were plotted in Sigma Plot 13.0.
**ITRAX-XRF Analysis**

The relative elemental composition of Core #1 was determined through X-ray fluorescence (XRF) analysis using a Cox ITRAX XRF Core scanner (Figure 2.3) located at the Ronald B. Gilmore XRF Laboratory, University of Massachusetts Amherst. Non-destructive, high resolution XRF analysis can provide useful geochemical data from marine and terrestrial sediment records that can serve to guide sample selection for further destructive analyses (Balascio et al., 2011; Croudace et al., 2006; van der Bilt et al., 2015).

To begin, Core #1 was positioned to enter the machine top first. After the length was defined the machine conducted an initial scan, taking an RGB image, a radiograph and a laser triangulated topographic profile of the split core surface. The radiograph was obtained at a voltage setting of 60 kV, a current setting of 40 mA and an exposure time of 2000 ms. The topographic surface profile obtained through laser triangulation is used in order to ensure that contact is not made between the surface of the core and the ITRAX XRF detector and to maintain a constant distance between the two (Croudace et al., 2006).

Following the first scan the core was again placed at the loading position and covered with a plastic film in order to prevent the sediment from drying out and cracking. Start and end points corresponding to the top and bottom of the core were selected from the RGB images obtained from the first scan. The second scan was then run in which XRF data was collected using a flat, rectangular-section micro-X-ray, which served to irradiate the sample to produce incremental downcore profiles of specific elements (Croudace et al., 2006). X-rays were produced with a voltage of 30 kV and a current of 55 mA through a molybdenum (Mo) tube while XRF measurements were made at 500-micron steps with an exposure time of 10 seconds. Geochemical data on a total of 36 individual elements was collected.
Figure 2.3: Image depicting a split core section being loaded into the ITRAX XRF Core Scanner (BOSCORF, 2012).
Results

AMS Radiocarbon Dating

Four terrestrial macrofossils were radiocarbon dated by DirectAMS in Seattle, WA. Starting at the top of the core, $^{14}$C analysis yielded ages of $1560 \pm 30$ (9 cm), $2110 \pm 20$ (14.5 cm), $1920 \pm 25$ (22 cm) and $4850 \pm 25$ (40 cm). Using the program CALIB 7.1 $^{14}$C ages were converted to calendar years BP (Stuiver and Reimer, 1993). The conversion of radiocarbon years to calendar years yielded downcore ages of $1489$ (9 cm), $2087$ (14.5 cm), $1882$ (22 cm), and $5595$ (40 cm). The age of the macrofossil collected at 22 cm was inverted, yielding an age younger than the previous sample by 190 radiocarbon years.

An age-depth model was developed with the converted ages using a linear interpolation, which was performed on the points between each consecutive age. Using this method, a model of the sedimentation rate in the isolation basin over time can be inferred. The highest sedimentation rate was observed over the 393(13cm) years between $1489(9cm)$ and $1882(22cm)$ cal yrs BP. From $1882$ to $5595(40cm)$ cal yrs BP it was observed to slow significantly. After $5595$ cal yrs BP the rate of sedimentation was assumed to remain constant from $5595$ cal yrs BP until the base of the core (10,500 cal yrs BP) (Figure 3.1).

The results obtained through linear interpolation were plotted against the downcore multi-proxy records so that depth was replaced with sediment age in calendar years BP. Due to the high sedimentation rate between $1489$-$1882$ cal years BP it is difficult to discern specific variability across multi-proxy records during this time period (Figures 3.3 and 3.4).
**Figure 3.1:** Age-depth model created using four AMS ages converted to calendar years using CALIB7.1. Inverted age and projected sedimentation rate is shown in red. The black lines indicate the projected sedimentation rate if the inverted age was ignored.

**210Pb and 137Cs Geochronology**

Downcore values of lead-210 and caesium-137 in decays per minute per gram (DPM/g) are plotted in Figure 3.3. Lead-210 displays a peak of 16.92 DPM/g at 0.5 cm depth. Values are observed to immediately decrease to 4.91 DPM/g at 1.5 cm, after which a slight increase is seen to 5.90 DPM/g at 2.5 cm depth. A gradual decrease is observed from 2.5 cm to 10.5 cm depth until values increase a final time from 14.41 DPM/g to
16.71 DPM/g at 12.5 cm depth. Cs-137 also displays a peak at .5 cm (3.53 DPM/g). This is followed by a gradual decrease in downcore values to 0 DPM/g recorded at 12.5 cm.

Figure 3.2: Downcore profiles of the Pb-210 and Cs-137 decays per minute per gram (DPM/g) values for Core #1. Note the peaks in Cs-137 and Lead-210 observed within the first 0.5 cm.

Visual Stratigraphy

Note to reader: The symbolization of the downcore stratigraphy for Core #1 is as follows: Black (2.5/1-2.5/2) gyttja is represented by a white and black cross-hatched rectangle, light gray (7/1-7/2) sand layers are represented by a yellow speckled rectangle and areas where mica fragments are observed in high concentrations are represented by a black speckled rectangle. Figure 3.3 shows the symbolized visual stratigraphy log developed for Core #1.
The majority of Core #1 is characterized as a dark brown and black gyttja. Sediment color darkens slightly from the top to the base of the core. Macrofossils of terrestrial vegetation (>3 mm) are observed in large concentrations throughout the upper quarter of the core and decrease in concentration with depth. Due to the fact that gyttja is the main stratigraphic unit, changes in the downcore stratigraphy are characterized as depositional interruptions and it is to be assumed that gyttja is present in between each of the units described. Slight changes in texture and color are observed throughout the downcore stratigraphy. At approximately 15.5-16.5 cm a thin, light gray, fine-grained sand layer occurs. Sand and silt size mica grains exist in higher concentrations within the gyttja at approximately 35-47 cm. At approximately 41.5-42 cm a second thin light gray sand layer is observed within the mica-rich unit, which is followed by a thicker light gray sand layer located at 43.5-44.5 cm downcore. High mica concentrations are again observed from 57-63 cm.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Lithology</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-15.5 cm</td>
<td>Black (2.5/1-2.5/2) rich gyttja makes up the majority of the matrix. Terrestrial macrofossils visible throughout. Possibly stem and leaf fragments of autochthonous organic matter.</td>
<td></td>
</tr>
<tr>
<td>15.5-16.5 cm</td>
<td>Light gray (7/1-7/2) fine-grained sand layer. Deposition likely the result of aeolian transportation.</td>
<td></td>
</tr>
<tr>
<td>35-47 cm</td>
<td>High concentration of mica fragments observed throughout the gyttja.</td>
<td></td>
</tr>
<tr>
<td>41.5-42 cm</td>
<td>Light gray sand deposits seen between 41.5 cm and 44.5 cm depth. Color similar to first observed sand layer.</td>
<td></td>
</tr>
<tr>
<td>43.5-44.5 cm</td>
<td>Fewer macrofossils seen throughout the gyttja. Sediment displays a less porous, more muddy texture.</td>
<td></td>
</tr>
<tr>
<td>44.5-65 cm</td>
<td>Micaceous layer observed at the base of the core. Slightly lower concentrations of mica fragments compared to the first observed mica-rich layer.</td>
<td></td>
</tr>
<tr>
<td>57-63 cm</td>
<td>Micaceous layer observed at the base of the core. Slightly lower concentrations of mica fragments compared to the first observed mica-rich layer.</td>
<td></td>
</tr>
</tbody>
</table>

**Figure 3.3:** A visual stratigraphy log of Core #1. Figure includes a high-resolution image paired with the symbolized core lithology, depth intervals and visual description of each
stratigraphic unit that is observed downcore. Sampling locations of AMS radiocarbon dated terrestrial macrofossils are also marked.

Percent Loss on Ignition

Organic content values for Core #1 range from 5% to 75% (Figure 3.3). At 0-3 cm depth %OM values increase from 27% to approximately 60% in Core #1 and are maintained until 9 cm. A sudden decrease is observed at 9 cm to 37%, which then increases to 75% at 12 cm. Upon reaching the sand layer %OM drops to 13% at 14-16 cm depth. An increase to 58% from 16 cm is seen immediately following the sand layer where %OM values ranging from 59% to 35% are maintained until 35 cm depth. A decrease is observed at 35-46 cm depth with %OM displaying a gradual decrease within the mica rich stratigraphic layer, eventually reaching a minimum value of 5% that coincides with a thick sand deposit seen at 44 cm. %OM again gradually increases to 40% at 48 cm before dropping suddenly to 23% at 50 cm. Immediately after an increase to 51% at 51 cm depth, %OM values are observed to range from 51% to 43% from 51-58 cm. Another trough is observed at the base of the core within a second mica rich deposit that occurs from 58-62 cm depth where %OM values decline to a low of 18% (Figure 3.4).

Stable Isotope and Elemental Analysis

Core #1 was sampled at 1 cm intervals for the analysis of %C, %N, C/N ratios, and carbon and nitrogen stable isotopes, δ¹³C and δ¹⁵N. The top 21 cm of the core show large variability with δ¹³C values ranging from -29.05‰ to 26.25‰. A gradual increase is observed from 21-56 cm depth. A peak occurs from 30-43 cm with a maximum value of -26.50‰. A trough is observed at the base from 56-62 cm with a recorded minimum value of -26.88‰. δ¹⁵N values display a slight increase from the top of the core to 20 cm depth (1.6-3.29‰). A gradual decrease to a minimum value of .761‰ occurs from 20-35 cm. A peak is seen at 40 cm with a value of 2.40‰. Finally, a decrease corresponding to the micaeous deposit is found at the base of the core from 56-62 cm (1.01‰) (Figure 3.4).
Figure 3.4: Downcore profiles for nitrogen and carbon stable isotopes, %C and \%N, C/N, loss on ignition and total pigments.
The ratio of C/N shows an increase from the top of the core to 14 cm (29) followed by a decrease at 16 cm corresponding to the occurrence of the sandy stratigraphic unit. Values continue to show a general decrease with a major peak observed at 43 cm depth (25). A decrease is observed following this peak where values stabilize around 17.5 until the base of the core (Figure 3.4).

Pigment Analysis

The greatest total pigment values are recorded from 50-61 cm depth, with a maximum value of 13 µg/g occurring at 61 cm and from 1-0 cm with a maximum of 19 µg/g. Decreases are observed in conjunction with micaceous and sandy stratigraphic units with the lowest recorded value of 3 µg/g found within the fine-grained sand deposit at 44.5 cm. Total pigments appear to fluctuate around the mean from 35-11 cm before decreasing at the top of the core from 11-1 cm depth (Figure 3.4).

Grain Size

Measurements of grain size yielded values ranging from 22 µm to 1023 µm. The majority of grain size values fell within the silt/clay fraction (<63 µm). A number of peaks were observed to exceed 62 µm. However, no grain sizes were recorded greater than 2 mm. Therefore, the sediment of Core #1 is predominantly silt/clay (<62 µm) and sand (<2 mm >62 µm). Values within the silt/clay range are consistent around 22 µm throughout the length of the core. Small peaks in grain size are observed at 9 cm, 14 cm, 17 cm and 21 cm with values of 118 µm, 150 µm, and 189 µm and 102 µm respectively. An increase in grain size is seen at the occurrence of the mica rich stratigraphic unit from 36 cm-42 cm with values ranging from 886-1023 µm. Values dropped off upon reaching the sandy layer at 42 cm back down to 22 µm. The next peak is observed at 56 cm depth with a value of 916 µm that occurs within the second mica-rich deposit at the base of the core. Values drop once more to 142 µm at 58 cm, which is then followed by final peak at 427 µm at the base of the mica deposit (Figure 3.5).
Figure 3.5: Downcore ITRAX-XRF elemental profiles, MS, grain size and %LOI.
Magnetic Susceptibility

The magnetic susceptibility measurements for Core #1 yield predominantly negative values. Values less than zero correspond to areas where gyttja makes up the majority of the composition in the visual stratigraphy log. Areas where values exceed zero are found at 2 cm (.4 SI), 16.5 cm (.2 SI), 38-44.5 cm (.4-.3.2 SI), and 61.5-63 cm (.9-.3.3 SI). There are two peaks that represent the highest susceptibility measurements recorded for Core #1 at 44.5 cm (3.2 SI) and 63 cm (3.3 SI). The areas of positive magnetic susceptibility values correspond to the light gray sand layers and mica-rich areas as seen in the visual stratigraphy log (Figure 3.5).

ITRAX-XRF Analysis

Downcore elemental profiles obtained through ITRAX-XRF analysis for Core #1 are shown in Figure 3.5. It has been well established in XRF analysis that X-ray energy detection limits can vary substantially across a variety of machine parameter settings and that organic-rich sediments are often associated with less efficient excitation of elements. (Croudace et al., 2006). Therefore, only the elements that were more easily detected and displayed a high overall XRF sensitivity, such as potassium (K), titanium (Ti), calcium (Ca), manganese (Mn), iron (Fe), and sulfur (S) were analyzed.

Potassium (K) ranges from .01-2.09 kcps with an average value of .17 kcps. An overall increase in values is observed within the mica-rich section of the core from 35-41 cm with the largest peak observed at 41 cm. Values remained low from 48-54 cm until another slight increase is observed at the base, reaching a maximum value of .8 kcps at 59 cm.

Titanium (Ti) values range from .02-2.67 kcps with an average of .34 kcps. Three distinct peaks occur at 28 cm, 40 cm, and 43 cm with values of 2.04 kcps, 2.67 kcps and 1.67 kcps respectively. The second two peaks take place within the mica-rich layer where an overall increase in values is observed.
Calcium (Ca) values range from .09kcps to 2.84kcps with a mean of .64kcps. Values start high at the top of the core and decrease steadily until a peak at 16cm (1.56kcps). Two larger peaks are seen at 42cm and 44cm with values of 2.1kcps and 2.84kcps respectively. The increase in values is correlated to the occurrence of two distinct sand units observed in the visual stratigraphy.

Manganese (Mn) ranges from .08-1.02kcps with an average of .144kcps. Distinct peaks are observed at the top of the core from 0.8-1.6cm. A maximum value of 1.02kcps occurs at 16cm. From 36cm-45cm values are seen to increase, reaching a maximum of .88kcps. Mn counts then decrease until the base of the core where another increase is observed from 54cm-64cm. The greatest values are observed in correspondence with the sand deposit at ~16cm depth and the two mica-rich deposits recorded in the visual stratigraphy.

The downcore profile of the ratio of Iron to Manganese (Fe/Mn) corresponds to the visual stratigraphy with lower values being observed within the mica rich deposit from around 12cm-45cm depth. The lowest values correspond to a thick sand layer found at the base of the first mica-rich deposit from 44cm-45cm (61.5kcps). The greatest values are observed within the organic rich gyttja and increase downcore to a maximum value of 5162.42kcps at 56cm depth before tapering off at the base.

Iron (Fe) values range from 1.26-204.92kcps with an average of 36.44kcps. Values are relatively stable around 15kcps until 37cm depth where a gradual increase is observed to culminate at the base of the core.

Sulfur (S) ranges from 0.0025-.35kcps with an average of .1kcps. Values are stable until 35cm depth where an increase is observed at 42cm (.253kcps). Following this peak, values decrease below the mean until an increase with a maximum value of .35kcps is seen at 49cm. Values are again observed to decrease to below the mean at 57cm, which is followed by an increase in values in the bottom 5cm of the core.
Discussion

Geochronology and Implications

The shoreline and sea level history of the island of Ingøy constructed by Retelle (2016) predicts that the isolation basin from which Core #1 was recovered would have been inundated by the sea at approximately 8,000 cal years BP. With the sea level reaching a maximum height of approximately 9 meters during the Tapes Transgression the isolation basin, situated at 7 meters above sea level, should have been inundated by marine waters sometime between 8,000-6,000 cal years BP and emerged from the sea ca., 5000 cal yrs BP. Marine influence within the catchment would have been recorded in the sediment record contained in Core #1 as fluctuations in sea level are known to affect water column chemistry, organic matter input and source of sediment (Balascio et al., 2011). The inundation of the isolation basin by the sea would have been identified through a variety of proxies including elemental profiles of calcium and sulfur (Croudace et al., 2006). δ^{13}C stable isotope values greater than -20‰ and a C/N ratio between 4-10 also serve as indicators of marine influence and were not recorded in Core #1 (Torres et al., 2012). This suggests that the sediment record does not include marine sediments associated with the mid-Holocene Tapes Transgression (ca. 6,000-8,000 cal yrs BP). Due to the fact that no evidence of marine influence could be identified, the age-depth model developed using the three AMS radiocarbon dates was deemed unacceptable for use. It is inferred that the macrofossil of woody vegetation yielding an age of 5,595 cal yrs BP had to have been transported into the catchment from the surrounding watershed rather than produced in situ. Therefore, it is not likely that the sediment record of Core #1 extends past 6000 cal yrs BP (Figure 4.1).

With this discovery, it is acknowledged that care must be taken in future studies where radiocarbon dating of material is employed to ensure any ages that are obtained fit within the known paleoenvironmental and paleoclimatological history of the area. Although a marine-lacustrine boundary was not observed in the 65 cm of Core #1, the recovery of a long core would likely display evidence of the isolation basin having been inundated by
the sea during periods of advancing sea level. Based on the well-documented sea level history of the island of Ingøy, evidence of marine influence could provide a reliable chronological constraint in the development of an age-depth model.

**Figure 4.1**: Sea level curve developed by Retelle (2015) for the island of Ingøy using 10 AMS radiocarbon dated marine shells recovered from a series of raised beaches. Green diamond indicates the marine limit at approximately 30 m asl and age of deglaciation from Romundset et al., 2011. Red dashed line represents the 7-meter elevation of the isolation basin from which Core #1 was recovered.

**Sedimentation Rate**

Through the analysis of Caesium-137 it was concluded that sediment accretion rates within the catchment are very low. The apparent $^{137}$Cs accumulation since 1954 is recorded in the top 12.5 cm of Core #1. Due to the fact that a peak in $^{137}$Cs is observed within the topmost 0.5 cm of the core and over 60 years have passed between its release into the atmosphere and this present study, the sedimentation rate into the catchment is inferred to be minimal. In past studies, variations in sediment grain size and $^{137}$Cs activity
have been linked, where fine-grained particles including clays and organic material appear to display greater retention of atmospheric fallout (Frissel and Penders, 1983). Because the upper 12.5 cm of Core #1 are characterized by highly variable amounts of sand, evidenced by increased values in XRF profiles of Ti, Ca, Mn, and K, the $^{137}$Cs profile could not be used for absolute dating. The downward translocation of $^{137}$Cs over 12 cm depth is significantly more than what is expected for arctic lake sediments in which reduced biological activity and bioturbation is generally associated (Hasholt et al., 2000). Due to the fact that Core #1 was recovered from a shallow, non-varved, organic rich isolation basin, variable eolian deposition into the catchment coupled with high biological activity leads to a large amount of uncertainty surrounding the construction of an age-depth model through the use of $^{137}$Cs fallout.

Lead-210 is another example of a fallout radionuclide that has been widely used for dating lake sediment cores (Appleby and Oldfield, 1978). $^{210}\text{Pb}$ is a naturally occurring radioactive element and is part of the uranium-238 radioactive decay series. With a half-life of $4 \times 10^9$ years, uranium is considered to be present in unchanging concentrations in the earth's crust over time. Uranium-238 slowly undergoes radioactive decay into uranium-234 (half-life of 248,000 yrs), which decays into uranium-234. The decay series continues with uranium-234 decaying into thorium-230 (half-life of 80,000 yrs) and lastly, radium-226 (half-life of 1620 yrs). The final decay product, radium-226, is found at low, constant levels in soils everywhere. When radium-226 decays, it produces radon-222 (half-life of ~3.8 days), a radioactive inert gas that can escape into the atmosphere via the air-soil interface (Appleby, 2001). Radon-222 naturally decays into polonium-218 after a period of several days residence time in the atmosphere, during which it attaches to aerosols and is eventually transported to the earth via dust and rain. The decay of polonium-218 into lead-210 (half-life of 22.3 yrs) occurs following its deposition on the globe's surface. The lead-210 that falls into a lake or ocean attaches to suspended matter and becomes permanently fixed in the sedimentation record through deposition (Turekian, 1977).
Although a downcore profile for Lead-210 was compiled for the top 12.5 cm of Core #1, it could not be used in the construction of an age-depth model. The fundamental equations relating excess lead-210 concentration in sediment to its fluctuation in the sediment surface and mass accumulation rate assume both constant sediment accumulation and fluxes in lead-210 (Sanchez-Cabeza and Ruiz-Fernández, 2012). Therefore, the downcore profile of lead-210 concentration should display an exponential curve. However, the activity versus depth record of lead-210 seen in Core #1 deviates from the expected exponential relation. The linear, non-exponential decrease in lead-210 that is observed downcore is most likely the result of bioturbation or sediment mixing due to extreme wind effects.

**Multi-Proxy Approach**

Based on trends observed across multiple proxies, four distinct zones were identified and labeled as Zone I, Zone II, Zone III, Zone IVa and Zone IVb. Zone I is characterized in Figure 3.4 by decreasing values observed throughout all records of biological activity. A slight increase in the detrital signal obtained through ITRAX-XRF analysis is also observed. Zone II is characterized by large variability across all organic proxies where no obvious trends are perceived. However, the detrital signal in Zone II is stable with little fluctuation observed. Zone III is characterized by another increase recorded in almost all of the organic proxy records and a second stabilization phase of the detrital signal where, again, little fluctuation is observed. Zone IVa and IVb are both distinguished by a general decrease in organic proxy records and a significant increase in the detrital input observed through ITRAX-XRF analysis (Figure 3.4 and 3.5).

Through the implementation of a multi-proxy analysis discrepancies between the age-depth model and paleoclimatological evidence were identified. Stable isotope values of $\delta^{13}$C and C/N ratios showed no evidence of marine influence in the sediment record. Elemental profiles of sulfur and calcium obtained through XRF analysis also neglected to show any evidence for the inundation of the isolation basin by the sea. Therefore, it was concluded that the AMS date of 5,595 cal yrs BP was not representative of the overall
sedimentation rate and could not be used in the construction of an accurate age-depth model. Had analyses on stable isotopes of $\delta^{13}C$ and XRF not been conducted and conclusions were drawn solely from magnetic susceptibility, loss on ignition and grain size data, inconsistencies between proxy records and the radiocarbon-dated terrestrial macrofossils may have been overlooked. Once again the multi-proxy approach is confirmed to be one of the most reliable methods utilized in the reconstruction of detailed paleoenvironmental conditions.

**Stable Isotope Analysis**

Paleoenvironmental proxies including $\delta^{13}C$, and C/N serve as reliable indicators of sediment source material. $\delta^{13}C$ values ranging from -9‰ to -15‰ have been associated with C$_4$ plants while values ranging from -23‰ to -33‰ are associated with C$_3$ plants (Janbu et al., 2011; Meyers, 2003; Meyers and Lallier-vergès, 1999) (Table 4.1). Due to the fact that $\delta^{13}C$ values observed in Core #1 ranged from -25‰ to -29‰, the sedimentary organic matter contained within the catchment was determined to consist exclusively of C$_3$ vegetation. In past studies, C/N values have been used to distinguish particulate organic carbon (POC), such as freshwater phytoplankton, from C$_3$ land-plants due to the fact that they display a similar range of $\delta^{13}C$ values (Deines, 1980). C$_3$ vascular vegetation has been observed to display C/N ratios greater than ~20 while phytoplankton tends to fall between 5 and 7. The fact that C/N ratios never fall below 16 indicates that Core #1 contains organic matter dominantly derived from C$_3$ terrigenous vegetation.

Complexities in nitrogen cycling, such as an abundance of nitrogen-fixing cyanobacteria or shifts in planktonic and heterotroph assemblages, can make interpretations of organic matter preserved in lake sediments difficult (Meyers, 2003). Dissolved NO$_3$, which is commonly used by algae as a form of dissolved inorganic nitrogen (DIC), typically displays values 7-10‰ greater than atmosphere-derived N$_2$, which becomes incorporated in the soil via nitrogen fixers and made available to land plants. The difference between atmospheric nitrogen $\delta^{15}N$ and DIN is preserved in the $\delta^{15}N$ of algae and C$_3$ plants, which is recorded as +8.5‰ and +0.5‰ respectively (Meyers, 2003). Because Core #1 displays
\( \delta^{15}N \) values ranging from 1.6-3.3‰ the primary source of nitrogen to the catchment is inferred to be atmospherically-derived \( \text{N}_2 \). Future research into the variety of factors that may be complicating the \( \delta^{15}N \) signal in Core #1 is required before any additional conclusions can be drawn.

<table>
<thead>
<tr>
<th>Proxy</th>
<th>Environmental Interpretation</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>K</td>
<td>Increased detrital input, drier conditions</td>
<td></td>
</tr>
<tr>
<td>Ti</td>
<td>Increased detrital input, eolian deposition, flood layers</td>
<td></td>
</tr>
<tr>
<td>Ca</td>
<td>Marine influence, primary productivity, detrital carbonates</td>
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<tr>
<td>Mn</td>
<td>Increased detrital input, oxygenation of bottom waters</td>
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<tr>
<td>Fe/Mn</td>
<td>Reducing conditions</td>
<td>Croudace and Rothwell, 2015</td>
</tr>
<tr>
<td>Fe</td>
<td>Increased detrital inputs, redox conditions</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td>Marine influence, soil derived S from leaching</td>
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<tr>
<th>Stable Isotopes</th>
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<tr>
<td>( \delta^{13}C )</td>
<td>Proportions of ( \text{C}_3 ) and ( \text{C}_4 ) plants (( \text{C}_3 ) plants = -33‰ to -23‰, ( \text{C}_4 ) plants = -15‰ to -9‰)</td>
<td>Meyers, 2003</td>
</tr>
<tr>
<td>( \delta^{15}N )</td>
<td>Proportions of algal and land-plant organic matter, enhanced algal productivity, nitrogen fixation</td>
<td>Meyers, 2003</td>
</tr>
<tr>
<td>C/N</td>
<td>Proportions of algal and land-plant organic matter (Algae = less than ~10, land plants = larger than ~12)</td>
<td>Meyers, 2003</td>
</tr>
</tbody>
</table>

**Table 4.1**: Table of ITRAX-XRF elements and elemental ratios listed with their associated environmental interpretations outlined by Croudace and Rothwell, 2015. Stable isotopes \( \delta^{15}N \), \( \delta^{13}C \) and C/N are also listed alongside their environmental interpretations as discussed by Meyers, 2003.

**XRF Analysis**

ITRAX XRF analysis served as a valuable tool for obtaining high-resolution downcore elemental profiles. However, as Balter (2014) discussed, the results should only be interpreted relatively and not quantitatively. This is due to the fact that characteristics of the sediment matrix, including particle size, water content, mineralogy, density and
changes in in surface topography, can have an effect on XRF readings (Croudace et al., 2006).

XRF analysis did serve to provide valuable insight regarding the input of detrital material to the catchment as well as past biologic conditions of the lake. For example, the ratio of Fe/Mn serves as an indicator of reducing conditions or the limitation of oxygen within the water column (Croudace et al., 2006) (Table 4.1). The general increase that is observed downcore is interrupted at the occurrence of mica rich (35-40cm) and fine-grained sand layers located within Zone IVa and IVb. Coupled with an increase in Mn, it can be inferred that oxygenation of the water column took place at these intervals through an influx of allochthonous material into the catchment. The record of an increase in detrital material is compiled through grain size analysis and magnetic susceptibility measurements, which highlight changes in the overall core stratigraphy, characterized predominantly as organic rich gyttja. The fact that the occurrence of micaceous grains is seen to increase from 35-45cm depth suggests weathering and erosion of the surrounding bedrock, which has been classified as a micaceous schist (Kirkland et al., 2007).

Although variability in sulfur corresponding to detrital input is seen within the bottom half of the downcore XRF elemental profile, fluctuations are interpreted to be related to the trophic state of the isolation basin rather than marine influence. Sulfate reduction occurs in sediments of most lakes where microbially reduced sulfur is incorporated into the organic matter via diagenesis (Urban et al., 1999). Sulfur enrichment of organic matter can occur in lake sedimentary environments containing very low sulfur concentrations. Sulfate reducing bacteria may thrive by using sulfide and other forms of reduced sulfur to fix carbon dioxide. In high arctic lakes sulfate reduction has been found to be limited to anoxic conditions where any sulfide that is produced is either reoxidized to sulfate and lost within the water column or is bound to Fe and organic matter retained in the sediment (Drevnick et al., 2010). Therefore, Fe and organic matter (%C, %N) were determined to be good predictors of within-lake reduction of inorganic sulfur. The downcore XRF profile of elemental sulfur for Core #1 displays a strong correlation to that of Fe and Mn. Due to the fact that high XRF counts of Fe have been attributed to
reducing conditions in lacustrine sediment systems, increases in sulfur within the bottom half of Core #1 have been inferred to correspond to anoxic conditions. Increased organic content (%C and %OM), higher total pigment concentrations, higher counts of Fe/Mn and less negative $\delta^{13}$C values provide evidence of an anoxic paleoenvironment.

Although the dynamic cycling of sulfur in its numerous reduced and oxidized forms is complex, processes such as an increase in sulfur content within eutrophic lake systems that typically display anoxic environments can serve as valuable indicators of paleoenvironmental conditions (Urban et al., 1999). The fact that increased counts of elemental sulfur were attributed to reducing conditions within the lake rather than marine influence provides further evidence supporting the lack of a marine-lacustrine boundary observed in Core #1.

Conclusions

The combination of biogeochemical and physical analyses allowed variability to be observed within the sediment record and changes in the paleoenvironmental record to be identified. The organic-rich gyttja was periodically interrupted by increases in minerogenic inputs including wind-blown sand and mica fragments. Such intervals have been interpreted as being associated with drier climates, which could have increased the amount of eolian deposition into the catchment. However, because an accurate age model could not be developed for Core #1, changes in the multi-proxy paleoclimate record cannot be directly correlated with larger scale climate forcings or climate anomalies such as the Neoglacial or the Holocene thermal maximum at this time.

The relationship between organic proxies and minerogenic inputs need to be examined further in order to determine the potential mechanisms that drive shifts in the multi-proxy record. Linking the paleoenvironmental record reconstructed from Core #1 to external climate forcings is nearly impossible without the aid of a reliable age-depth model. Hypotheses can be drawn regarding the environmental conditions of the lake based on the combination of biological and geochemical analyses. However, the identification of
specific mechanisms resulting in the climate variability that is recorded in Core #1 requires further investigation into well-dated paleoenvironmental records of the region.
References


Marcott, S. A., Shakun, J. D., Clark, P. U., and Mix, A. C., 2013, A Reconstruction of Regional and Global Temperature for the Past 11,300 Years: Science, v. 339, no. 6124, p. 1198.


1.) Summary figure of stable isotope analysis, loss-on-ignition and total pigment analysis plotted against calendar years BP.
2.) Summary figure of ITRAX-XRF analysis, grain size, magnetic susceptibility and loss on ignition analyses plotted against calendar years BP.