EVOLUTION RATES OF POST GLACIAL LAKE ECOSYSTEMS IN NORTHERN
NEW ENGLAND: A GEOCHEMICAL STUDY USING LAKE SEDIMENTS

A Thesis Presented

by

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ABSTRACT

Geochemical analysis of sedimentary records from post-glacial lakes yields insight into how surface and lacustrine processes have been influenced in the past by both climatic and environmental changes. Sediment cores were retrieved from five small (0.38 to 1.43 km^2) post-glacial lakes in northern New England (New Hampshire and Maine) to determine differences in the timing of early Holocene ecosystem establishment on a local to regional scale. The lower section of each core documents the transition from organic-poor sands, silts, and clays to organic-rich gyttja-type sediments. The amount of sedimentary organic matter (OM) is a function of lacustrine productivity and the magnitude of organic inputs from the revegetating watershed.

Multiple studies have examined the response of lake ecosystems to glacial-postglacial climate change; however, most of the previously studied lakes already existed during the last Glacial period. Thus, little is known about the biologic response of a newly deglaciated landscape and how rapidly aquatic plant communities develop in lakes surrounded by barren, carbon and nutrient-poor watersheds.

Laboratory analytical methods used in this study include measurement of organic carbon content (%C), carbon/nitrogen (C/N) ratios, stable carbon isotopes (δ^{13}C), and hydrogen and oxygen indices (HI, OI) from Rock-Eval pyrolysis. These proxies aid in determining the relative contribution of allocthonous vs. autochthonous OM input to lakes. Radiocarbon dates were obtained for macrofossils to provide estimates of sediment age.

The %C records and Rock-Eval data show increases that correlate with primary succession of terrestrial vegetation (tundra → mixed woodlands → mature forest) and the warming that occurred after the Younger Dryas cooling period. Differences are apparent in both the timing and rate of ecosystem establishment in all of the study lakes, suggesting that lake elevation, frequency and magnitude of terrigenous inputs, and proximity to the receding Laurentide ice sheet are key factors controlling nutrient availability to the lake and revegetation rates. Negative shifts in bulk OM δ^{13}C values occur in the older sediments of all five lakes. These are interpreted as reflecting changes in OM source resulting from both a climate driven switch from C_4 and C_3 tundra shrubs to C_3 woodland plants and increased aquatic productivity.

The data collected for the lakes in New Hampshire and Maine were compared with geochemical records from previous studies on Vermont lakes by Lini (1999) and Galster (2001). Similarities in lake sedimentary records across northern New England include increases in OM content up-core, and a negative trend in δ^{13}C values in the older sediments. However, distinct differences are apparent throughout the study lakes. Primarily, the Vermont lake sediments display higher %C values and more negative δ^{13}C values than the lakes in New Hampshire and Maine. In addition, the correlation between the first documented increase in OM content and the decrease in δ^{13}C values, as observed in the Vermont lakes, is not apparent in the New Hampshire and Maine lakes.

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Chapter 1. INTRODUCTION

1.1 Introduction

Upon recession of the Laurentide ice sheet during the late Pleistocene/early
Holocene eras, post-glacial lakes were formed on the barren landscape of northern New
England. As ecosystems developed in and surrounding these lakes, alterations in the
aquatic and terrestrial biologic communities were recorded in the sediments deposited
onto the lake bottom. These sedimentary records provide detailed information on how
surface and lacustrine processes were influenced by climatic changes, such as those that
occurred during the early Holocene.

The majority of previous studies dealing with lacustrine sedimentary records of
similar age document the response of lakes to climate change in vegetated watersheds in
non-glaciated areas. Thus, little is known about how rapidly plant communities develop
in newly formed lakes surrounded by barren, carbon and nutrient-poor watersheds. This
study seeks to investigate both the rates at which lacustrine ecosystems were established
in these newly created lakes, and the relationship between watershed and lake ecosystem
evolution.

This thesis investigates ecosystem development rates of lakes in northern New
England using geochemical analyses of sedimentary records. Sediment records from
three lakes in New Hampshire (Stinson Lake, South Pond, and Crystal Lake) and two
lakes in western Maine (Johns Pond and Worthley Pond) were analyzed. The
sedimentary organic matter (OM) is characterized via the use of organic carbon content

(%C), carbon to nitrogen (C/N) ratios, stable carbon isotopes (δ^{13}C), and hydrogen and oxygen indices (HI, OI) from Rock-Eval pyrolysis.

1.2 Overview of Thesis

This research concentrates on sedimentary records from post-glacial lakes in New Hampshire and Maine, and compares them with previous studies done in Vermont (Lini, 1999; Galster, 2001), in order to achieve the following objectives:

- To investigate the onset of primary productivity in newly formed, post-glacial lakes in northern New England.

- To determine and compare ecosystem development rates in different lakes.

- To investigate which parameters influence the rate at which a lake ecosystem develops.

The document begins with a comprehensive literature review (Chapter 2), followed by a description of the study area, and the field and lab methods used throughout the course of the project (Chapter 3). The second half of the thesis consists of a paper to be submitted for publication to the Journal of Paleolimnology (Chapter 4). The paper presents the bulk of the results of this study, as well as a detailed discussion and comparison of the data across northern New England. Finally, a summary of the findings and suggestions for future research are presented in Chapter 5. The appendices contain the laboratory data, age models, bathymetric and topographic maps, and aerial photographs of the studied lakes.

Chapter 2. COMPREHENSIVE LITERATURE REVIEW

2.1 Deglaciation History of Northern New England

The most recent glacial period in North America is known as the Wisconsinan era. The continental ice sheet reached its glacial maximum approximately 18,000 ^{14}C years before present (^{14}C yr BP) (21,000 - 22,000 calibrated yr BP), when it extended as far south as Long Island, NY (Benn and Evans, 1998).   Following the glacial maximum, the ice sheet began to retreat across New England towards northern Canada.

The timing and sequence of deglaciation in northern New England has been debated over the past 60 years and more studies and systematic mapping are still needed (Ridge et al., 1999; Thompson, 1999). Ridge et al. (1999) estimate that deglaciation began in southern Vermont and New Hampshire 12,600 ^{14}C yr BP (15,200 - 15,500 yr BP), and that the ice sheet receded into southern Quebec about 11,500 ^{14}C yr BP (13,700 - 13,800 yr BP). Numerous glacial lakes existed in northern New England during the recession of the ice sheet, however none of the New Hampshire and Maine lakes in this study were located within these glacial lake basins. A readvance of the ice sheet occurred in the Littleton-Bethlehem area of northwestern New Hampshire approximately 12,000 ^{14}C yr BP (14,000 - 14,100 yr BP) (Ridge et al., 1999; Thompson et al., 1999). The Littleton-Bethlehem area is located geographically north or west of the three lakes cored in New Hampshire. The ice sheet withdrew completely from the northwestern White Mountains in New Hampshire between 13,000 and 11,000 ^{14}C yr BP (13,000 - 15,900 yr BP) (Thompson et al., 1999).

The ice sheet in western Maine receded in a general northward direction from the coast of Maine to the Canadian border from about 13,500 to 11,500 ^{14}C yr BP (16,400 - 13,700 yr BP) (Thompson, 2001). In central western Maine, near Worthley Pond, striations indicate that the ice sheet receded to the north or north-northeast. Ice margin retreat in northwestern Maine, near Johns Pond, occurred in both northern and western directions, and the ice receded over the Canadian border by 11,500 ^{14}C yr BP (13,800 - 13,700 yr BP) (Thompson, 2001).

During the glacial-interglacial transition, a brief cooling period occurred, known as the Younger Dryas (YD) (Peteet et al., 1990). In the northeastern United States, the timing of the YD was approximately 11,000 to 10,000 ^{14}C yr BP (13,100 - 11,500 yr BP). During this period, the vegetation switched back to species typical of colder climates, as documented in pollen studies by an expansion and then rapid decline of boreal species around 10,000 ^{14}C yr BP (11,500 yr BP) (Peteet et al., 1990).

2.2 Revegetation Patterns of Deglaciated Terrains

The type of plant cover to first colonize recently deglaciated terrain depends on the species of propagules (spores and seeds) available, the climate, and the ground surface conditions (Pielou, 1991). Biota (such as mosses and lichen) play a dominant role in the soil development, and soils develop slowly until enough vegetation is present to provide abundant organic matter (Matthews, 1992). The vascular vegetation that is first established along glacial boundaries is grown from seeds carried by the wind and deposited into the barren terrain (Pielou, 1991). Since tundra plants are typically in closest proximity to glacial boundaries, the first pioneering seeds are likely to be tundra

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pollen (Matthews, 1992). The amount of time necessary for significant vegetation to develop depends on the development rates of soils. Nutrient-carrying soils are blown in by the wind, deposited onto the barren landscape, and vegetation begins to grow (Matthews, 1992). The vegetation enriches the soils, and serves as a barrier; catching other blowing soils, and eventually developing the landscape (Pielou, 1991).

Many studies have been conducted on revegetation patterns of recently deglaciated terrain. Near the Klutlan Glacier, Yukon Territories, mosses grew on the terrain that had been deglaciated for at least nine years, followed by lichen, then by flowering plants (Birks, 1980). The deglaciated terrain near the Franz Joseph Glacier, New Zealand, was first colonized by herbs and mosses, followed by shrubs 10 years after deglaciation (Burrows, 1990).   In Glacier Bay, Alaska, a matted layer of 'black crust' on the soil surface, containing blue-green algae, developed within a few years of deglaciation (Lawrence, 1979).

Another study on Glacier Bay showed that rapid re-vegetation of the newly deglaciated landscape (surrounding lakes formed within the past 50 years) altered the flux of nutrients into the adjacent lake, changing the aquatic communities and production within the lake ecosystem (Engstrom et al., 2000). For example, the development of nitrogen-fixing alder thickets caused a buildup of nitrogen in the surrounding soils, and was responsible for a rise and subsequent decline of total nitrogen within the lake water. However, older lakes in the area that lack evidence of an alder vegetation phase did not experience a rise of nitrogen in the waters (Engstrom et al., 2000).

Revegetation in Northern New England

After the last glacial period, vegetation in northern New England began as tundra, changed to mixed woodlands, and eventually became mature forest (Davis and Jacobson Jr, 1985). Typical tundra plants in the area consisted of sedge, willow, grasses, sage, dwarf birch, alder, fern, and mosses; woodlands consisted of poplars, jack pine, spruce, and small amounts of birch, ash, elm, balsam fir, and larch; and early forests were dominated by spruce, balsam fir, birch, poplar, ash, small pine, and larch (Davis and Jacobson Jr, 1985). The landscape changed from tundra to woodland to forest in a northeastward and northward direction, with succession controlled by elevation gradients and deglaciation patterns (Davis and Jacobson Jr, 1985). By estimating establishment rates using the results from Davis and Jacobson, Jr. (1985), it is evident that the vegetation in the watersheds in my study established at different times and rates (Table 2.1).

Anderson et al. (1986) studied late and post-glacial vegetation around Upper South Branch Pond, Maine, which is located to the northeast of the two Maine lakes in my study. They found that the first plants to grow around Upper South Branch Pond after deglaciation were herb-shrub tundra plants, such as sedges, grasses, and mosses that typically colonize on rock or mineral soils. Woodland vegetation occupied the watershed from ~ 10,500 to 9,000 ^{14}C yr BP (10,500 - 8800 yr BP), and the abundance of mosses decreased during this time period. The establishment of present-day (Holocene) moss communities correlates with an increase in OM found in the sediment and the development of the first forest communities (Anderson et al., 1986).

Table 2.1. Approximate dates of deglaciation and time periods of vegetation change in the areas surrounding the studied lakes. Estimated dates of deglaciation are based on data in Thompson et al. (1999), Ridge et al. (1999), and Thompson (2001). Estimated vegetational changes are based on pollen data from Davis and Jacobson, Jr. (1985).

Lake Name	Deglaciation (^{14}C yr BP)	12,000 ^{14}C yr BP	11,000 ^{14}C yr BP	10,000 ^{14}C yr BP	9,000 ^{14}C yr BP
Stinson Lake	~12,300	Tundra	Mixed Woodlands	Mixed Woodlands/ Forest	Forest
South Pond	~12,200	Tundra	Tundra /Mixed Woodlands	Mixed Woodlands/ Forest	Forest
Crystal Lake	~12,500	Tundra /Mixed Woodlands	Mixed Woodlands/Forest	Forest	Forest
Johns Pond	~12,000	Tundra	Tundra	Mixed Woodlands/ Forest	Forest
Worthley Pond	~12,500	Tundra /Mixed Woodlands	Mixed Woodlands/Forest	Forest	Forest

Aquatic and terrestrial insects are a significant component of the OM in the sediment cores from lakes in the northern White Mountains of New Hampshire (Dorian, 2000). The older sections of the cores contain indicators of herbaceous ground cover such as sedges, rush, buttercup, and mosses, as well as nitrogen-fixing and woody shrubs (artic cranberry and blueberry). Evidence also suggests that a few small trees were present in the tundra landscape including dwarf birch, willow, and spruce (Dorian, 2000).

2.3 Sources of Organic Matter to Sediment

Lake sediments are composed of a mixture of both organic and inorganic materials. These are either produced within the lake itself, or are transported into the lake from the surrounding watershed. Although OM only accounts for a small portion of lake sediments, it plays an important role in determining past rates of productivity and terrestrial OM contribution (Meyers and Ishiwatari, 1993; Meyers, 1997; Meyers and Lallier-Verges, 1999). Since this study is mainly concerned with the organic fraction of sediments, the inorganic portions will not be discussed in detail.

Organic matter enters sediments from three main pathways; as plant particles washed into the lake, as algae and macrophytes growing within the lake itself, or from eolian sources. However, eolian OM contribution is small compared to that from lacustrine or terrestrial sources (Meyers and Ishiwatari, 1993).

2.4 Organic Matter Content (%C)

Organic matter contains approximately 50% C, thus the concentration of carbon in lake sediments is directly related to the amount of OM within the sediment (Meyers and Lallier-Verges, 1999). The concentration of OM in sediments is influenced by plant

production within the lake, terrestrial plant production surrounding the lake, and the amount of inorganic material transported into the lake (Meyers, 1997). The %C values of lake sediments may be altered via degradation within the water column during sinking (see Section 2.8) (Meyers, 1997). The amount of organic carbon (C_{org}) present in sediments may also be decreased due to dilution by detrital inorganic material in runoff or by minerals that precipitated out of the water column (Meyers and Lallier-Verges, 1999).

Multiple studies have focused on how %C values in lake sediments change due to differences in productivity and terrestrial input. For example, in Lake St. Moritz, Switzerland, the C_{org} content in the sediments shows a clear relationship with productivity levels: %C increases as the productivity within the lake increases, and decreases when productivity is reduced (Ariztegui et al., 1996a). Low amounts of C_{org} (in the sediments dated ~15,500 - 13,200 ^{14}C yr BP or 17,500 - 15,200 yr BP) in Lake Victoria, East Africa, were interpreted as the result of increased input of terrestrial material into the lake, combined with unfavorable conditions for the preservation of OM; however, increased rates of primary production and burial of OM during the glacial post-glacial transition (approximately 12,000 - 10,500 ^{14}C yr BP or 13,600 - 11,150 yr BP) caused an increase in C_{org} content (Talbot and Laerdal, 2000).

2.5 C/N Ratios

An estimate of the source of OM in lake sediments can be obtained by comparing the concentration of C and N (the C/N ratio). The C/N ratio is directly related to the amount of cellulose in a plant. Non-vascular plants, such as algae, have low amounts of

cellulose and C/N ratios ranging from 4 to 10; vascular plants contain higher amounts of cellulose and show C/N ratios of 20 or greater; and aquatic macrophytes tend to have C/N ratios below 20 (Meyers and Ishiwatari, 1993; Meyers, 1997; Meyers and Lallier-Verges, 1999). If a lake is highly productive and receives little OM from the surrounding watershed, the sediments will have a relatively low C/N ratio; conversely, lakes that receive more terrestrial than lacustrine OM will display higher C/N ratios in the sediment (Meyers and Ishiwatari, 1993; Meyers and Lallier-Verges, 1999).

Meyers and Lallier-Verges (1999) and Meyers and Ishiwatari (1993) compiled sediment C/N ratios from various studies on lakes throughout the world. They suggest that C/N ratios of 13-14 in the surface sediments of lakes such as Lake Bosumtwi in Africa are the result of a relatively equal mixture of algal and terrestrial plant contributions to the sediment. In Lake Biwa, Japan, C/N ratios fluctuated between 5 and 10 throughout the past 430,000 years, yielding the conclusion that OM inputs were dominated by algal sources (Meyers, 1997; Meyers and Takemura, 1997). In a sediment core from the Rochester Basin in Lake Ontario, the C/N ratios increase from the bottom glacial sediments toward the more recent sediments. This was interpreted to be the result of a switch from algal to terrestrial OM dominance (Silliman et al., 1996).

2.6 Stable Carbon Isotopes (δ^{13}C)

δ^{13}C of Plants

Plants take up carbon through the process of photosynthesis, using energy provided by the sun to produce OM. Water and atmospheric carbon dioxide are combined to form OM (CH_2O) and oxygen is released into the atmosphere. The simple equation for photosynthesis is shown in Equation 1 (Schlesinger, 1997).

$$H_2O + C\ O_2 \rightarrow O_2 + CH_2O \qquad (1)$$
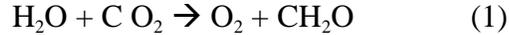
There are three main pathways by which plants perform photosynthesis: C_3, C_4, and CAM (O'Leary et al., 1992). Plants with the C_3 pathway use the enzyme RuBP carboxylase (Rubisco) to reduce CO_2. Rubisco preferentially incorporates $^{12}CO_2$ over $^{13}CO_2$, thus C_3 plants have low (more negative) $\delta^{13}C$ values (-20 to -32 ‰, with an average of -27‰; Boutton,1991a). C_4 plants use the enzyme PEP carboxylase to reduce CO_2. Although PEP carboxylase discriminates against ^{13}C, it does so less than Rubisco does. As a result, C_4 plants have higher (less negative) $\delta^{13}C$ values (-9 to -17‰, with an average of -13‰; Boutton, 1991a). CAM plants use the crassulacean acid metabolism to fix CO_2, and tend to have $\delta^{13}C$ values that range between those of C_3 and C_4 plants (Boutton, 1991a). Most CAM plants are members of the succulent and desert plant varieties, and will not be considered further in this study since there is no evidence for the existence of desert environments in northern New England. Most of the species currently surrounding the five studied lakes in northern New England are C_3 plants, although C_4 plants may have been in the area during the time period of tundra vegetation that existed shortly after deglaciation.

Phytoplankton and freshwater macrophytes use the C_3 pathway for carbon fixation and have $\delta^{13}C$ values that reflect their source of carbon (LaZerte and Szalados, 1982; Keeley and Sandquist, 1992). In oligotrophic lakes, phytoplankton utilize dissolved CO_2 in equilibrium with the atmosphere and fractionate against ^{13}C during photosynthesis (Brenner et al., 1999; Meyers and Lallier-Verges, 1999). However, during periods of high production the supply of $^{12}CO_2$ in the water may be depleted, and some algae and aquatic plants will switch to HCO_3 (bicarbonate) within the water for their carbon source,

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resulting in ^{13}C enriched OM and higher $\delta^{13}C$ values (Stuiver, 1975; Hakansson, 1985; Brenner et al., 1999; Wetzel, 2001).

If dissolved inorganic carbon (DIC) is used as the source of C, the $\delta^{13}C$ value of OM produced is influenced by the $^{13}C/^{12}C$ ratio of the DIC (Stuiver, 1975; LaZerte and Szalados, 1982). The isotopic signature of DIC in the water column is controlled by the rate of CO_2 exchange with the atmosphere, the type and decomposition rate of OM in the watershed, whether HCO_3 or CO_2 is used for primary production, and rates of photosynthesis and respiration within the lake itself (Brenner et al., 1999). Freshwater DIC incorporates CO_2 from terrestrial C_{org}, yielding $\delta^{13}C$ values lower than those of atmospheric CO_2 (- 8‰) (LaZerte and Szalados, 1982). Aquatic macrophytes range in $\delta^{13}C$ values based on whether they are submergent and use DIC as their main C source, or if they are emergent and use both DIC and atmospheric CO_2.

A change in lake trophic state can alter the $\delta^{13}C$ of aquatic plants (McKenzie, 1985). As lakes become more eutrophic, and produce greater amounts of phytoplankton, the $\delta^{13}C$ in the surface water will become more positive. However, if primary production switches from being dominated by macrophytes to being dominated by algae, the $\delta^{13}C$ will decrease, since phytoplankton generally have lower $\delta^{13}C$ values than macrophytes (Hakansson, 1985).

Since both terrestrial plants and surface waters obtain CO_2 from the atmosphere, changes that occur to atmospheric CO_2 content (pCO_2) can affect the $\delta^{13}C$ of plants. Significant changes in the concentration of pCO_2 occurred after 1840, when the industrial revolution began. This change has been documented by a negative shift of ~1.5‰ in the

$\delta^{13}C$ values of atmospheric CO_2 (Keeling et al., 1989), and may have caused similar negative shifts in the $\delta^{13}C$ of surface waters and terrestrial plants.

Since $\delta^{13}C$ values of terrestrial and aquatic plants differ, they can aid in distinguishing whether sedimentary OM originated as C_3 or C_4 terrestrial material or is of lacustrine origin. Plotting the $\delta^{13}C$ value against the C/N ratio provides a good estimate of the origin of the OM in the sample (Figure 2.1) (Meyers, 1997; Meyers and Lallier-Verges, 1999).

$\delta^{13}C$ of Sediments

The $\delta^{13}C$ value of sediments is directly influenced by the amount of allochthonous and autochthonous material in the sediment, productivity levels, and changes that may occur as a result of alterations in the chemical composition of OM as it moves through the water column and/or as it rests in the surface sediments (see Section 2.8). Many studies have used stable carbon isotopes to identify changes in both productivity and origin of sedimentary OM. Stable carbon isotopes were used to investigate past changes in productivity that have occurred since the mid-1800s in Lake Ontario (Schelske and Hodell, 1991; Hodell and Schelske, 1998). Consistent increases in $\delta^{13}C$ of the sediments over time were interpreted as the result of historical increases in nutrient input (phosphorus) into Lake Ontario. The added nutrients allowed for greater productivity within the surface waters, resulting in higher $\delta^{13}C$ values (-24 to -25 ‰ vs. -27‰) than those measured in the older pre-eutrophication sediments (Schelske et al., 1983; Schelske and Hodell, 1991; Hodell and Schelske, 1998).
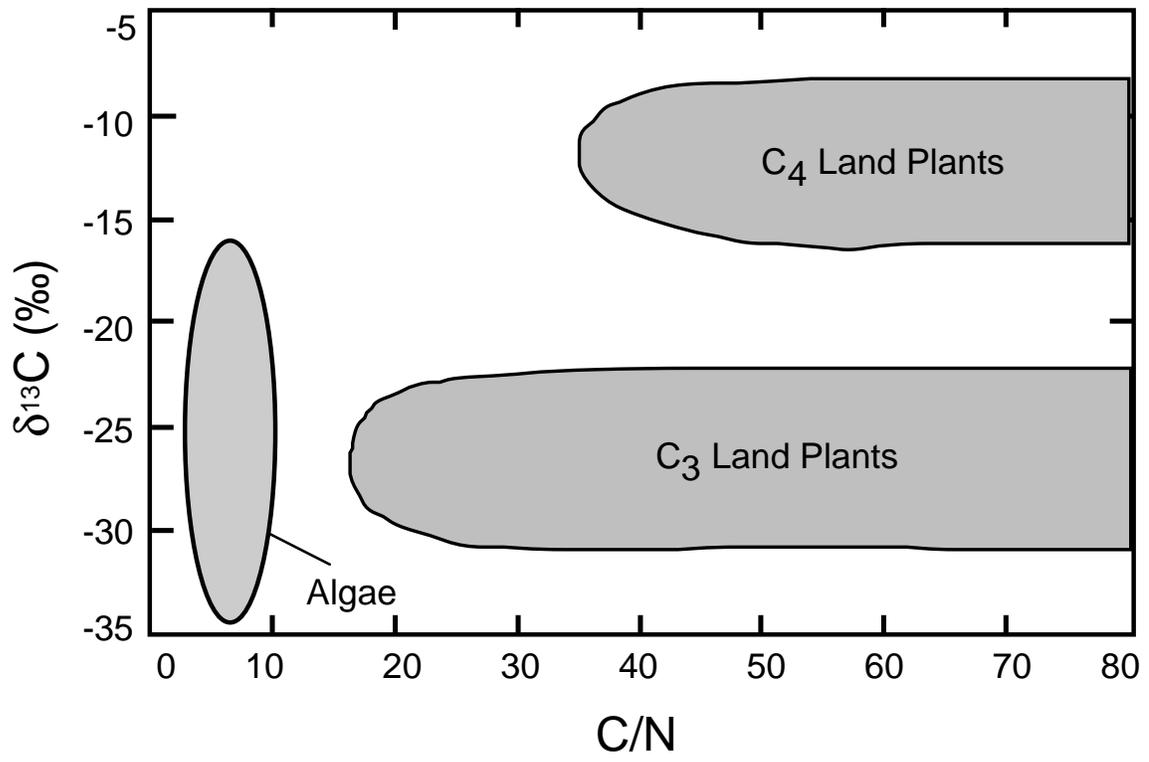
Figure 2.1. C/N and $\delta^{13}C$ values of algae, C_3 land plants, and C_4 land plants (modified after Meyers, 1997).

A core from Quechee Lake, VT, revealed 10,000 ^{14}C year (11,500 yr BP) old sediments with low (-30‰) δ^{13}C values, interpreted to be due to algal OM contribution (Stuiver, 1975). Samples from the same core with higher δ^{13}C values were interpreted as the result of an increase in terrestrial OM reaching the lake, combined with an increase in submerged macrophytes.

Hydrologic changes along the boreal tree-line can also affect the δ^{13}C in adjacent lake sediments (Wolfe et al., 1996). For example, mid-Holocene sediments enriched in ^{13}C in Toronto Lake, Northwest Territories, Canada, were associated with increased aquatic production and decreasing amounts of CO_2 in closed basin conditions. A declining trend of δ^{13}C values throughout the Holocene was attributed to recycling of ^{13}C-depleted carbon (Wolfe et al., 1996).

Talbot and Laerdal (2000) published a detailed 17,500 year paleolimnological record of Lake Victoria, East Africa, showing relatively high δ^{13}C values from the time period of ~15,500 through 13,200 ^{14}C yr BP (17,500 through 15,200 yr BP). These are thought to represent deposition of terrestrial C_4 plant material with δ^{13}C values > -20‰. However, as the lake began a period of rapid expansion between ~13,000 and 12,000 ^{14}C yr BP (15,000 and 13,600 yr BP), the δ^{13}C values began to decrease, suggesting an increase of primary production within the lake itself. A short period of deep mixing between the surface waters and the isotopically depleted deep waters followed, causing the δ^{13}C values to decrease further; however, once the mixing episode ended, the δ^{13}C values remained relatively stable, suggesting only gradual changes in the production and burial of OM in the sediments (Talbot and Laerdal, 2000).

2.7 Rock-Eval Pyrolysis

Rock-Eval pyrolysis was originally developed for use in the petroleum industry to evaluate hydrocarbon potential of petroleum source rocks (Espitalie et al., 1985; Talbot and Livingstone, 1989). Although the use of Rock-Eval pyrolysis on lake sediments is still a relatively new practice, it has proven to be a useful tool to estimate whether the sedimentary OM in sediments is of lacustrine or terrestrial origin (Meyers and Lallier-Verges, 1999).

Rock-Eval pyrolysis yields the hydrogen index (HI) and oxygen index (OI) of a sample. The HI (mg HC/g TOC) serves as a proxy for the H/C ratio of the OM in the sediments. The OI (mg CO_2/g TOC) is a proxy for the O/C ratio of the OM. These two indices are plotted against each other in a Van Krevelen-type diagram (Tissot and Welte, 1984) to determine whether the OM in the sediments originated from lacustrine or terrestrial sources (Figure 2.2). Samples with a high HI and a low OI consist mainly of lacustrine OM, while those with a low HI and a high OI contain more terrestrial OM (Talbot and Livingstone, 1989).

In a study of Lake St. Moritz, Switzerland, the bulk sediment HI values were used as an indicator of variations in sedimentary OM composition (Ariztegui et al., 1996a). The most recent sediments, which document a period of high productivity within the lake, show HI values up to 490. However, throughout the rest of the core, the sediments contain more terrestrial than algal OM, and have HI values below 400 (Ariztegui et al., 1996a; Ariztegui et al., 1996b).
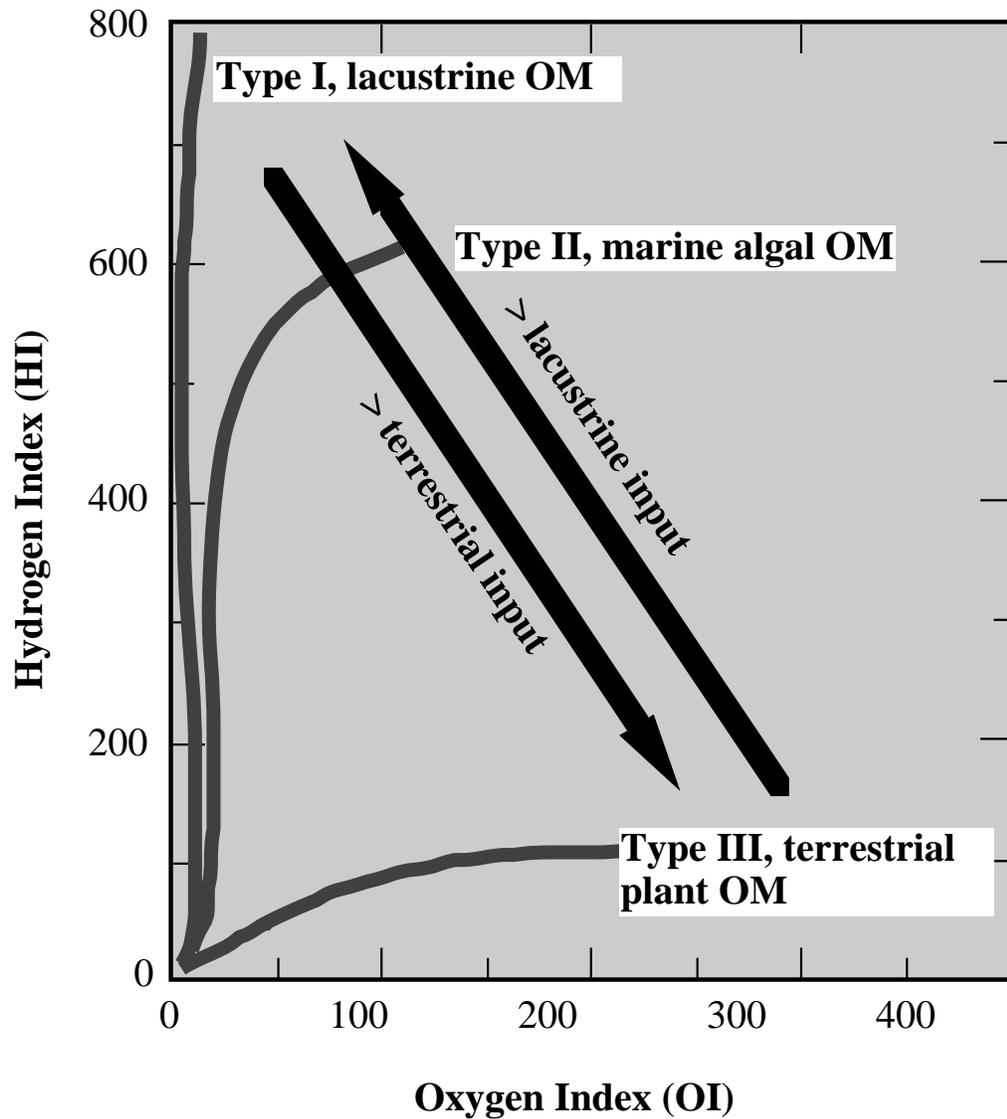
Figure 2.2. Van Krevelen-type diagram showing HI and OI values of lacustrine, marine, and terrestrial OM. Arrows represent shifts in relative amounts of lacustrine and terrestrial input in relation to the HI and OI values. The HI (mg HC/g TOC) is a proxy for the H/C ratio, the OI (mg CO_2 /g TOC) is a proxy for the O/C ratio of the OM.

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Talbot and Laerdal (2000) correlated low HI values in late Pleistocene sediments of Lake Victoria, East Africa with a time period of abundant terrestrial input into the lake. Around ~13,200 ^{14}C yr BP (15,000 yr BP), the sediment HI values began to increase due to a shift to deeper water conditions in the lake, by then dominated by algal production (Talbot and Laerdal, 2000).

2.8 Diagenesis and Bioturbation

Diagenesis occurs when the chemical composition of OM is altered during travel through the water column or during deposition in the surface sediments. However, as evidenced by numerous studies on bulk sediments, minimal diagenetic alteration of OM occurs after sedimentation (Meyers and Lallier-Verges, 1999). OM may also be consumed in the bioturbated surface layers of sediments (Meyers, 1997). The OM that sinks to the bottom of the water column is the foundation of the benthic food chain, and these benthic organisms may alter the chemical composition of the OM. In addition, bacteria living below the sediment surface appear to be capable of consuming and/or altering the composition of the OM after sediment burial (Meyers, 1997). However, in small lakes, such as those in this study, the sedimentation rates are high and the OM is often buried quickly, allowing for little to no bioturbation.

Diagenetic processes in the water column and top sediment layers have been found to reduce the amount of C_{org} in sediments (Meyers and Eadie, 1993; Schelske and Hodell, 1995). Sedimentary C_{org} concentrations in Lake Michigan diminished by a factor of 10 between the plankton stage and settlement on the lake bottom, but little difference

was found between the initial and final C/N ratios (Meyers and Eadie, 1993). These changes were attributed to a loss of algal components in the upper water column and apparent lateral input of algal OM near the surface sediments. Ertel and Hedges (1985) found vascular plant debris with C/N ratios between 30 and 40, removed from sediment having a C/N value of 15; evidence that little diagenesis occurred within the water column.

Many studies have shown that diagenesis has little overall affect on the $\delta^{13}C$ composition of sediments. During diagenesis ^{12}C is preferentially removed via consumption by bacteria and inorganic oxidation, resulting in ^{13}C enriched OM; however the resulting change in $\delta^{13}C$ is no more than 1-2 ‰ (Talbot and Livingstone, 1989). In oligotrophic lakes, where %C values are less than a few percent, no diagenetic isotopic effects have been observed (Talbot and Livingstone, 1989).

2.9 Similar Studies Conducted in Vermont

Previous studies have examined post-glacial landscape evolution in Vermont using a similar geochemical approach (Lini et al., 1998; Lini, 1999; Galster, 2001). Others have conducted research on pollen and inorganic sediments to study the paleo-vegetation and paleo-climate in Vermont following deglaciation (Li, 1996; Brown, 1999; Noren, 2002). Sediment cores collected from four Vermont lakes (Ritterbush Pond, Sterling Pond, Duck Pond, and Lake Morey) display significant lithological and geochemical changes, which reflect watershed re-vegetation patterns and rates. The $\delta^{13}C$ records in these lakes are characterized by a significant negative shift (up to 11‰) in the bottom sections of the cores (Figures 2.3, 2.4). The shift correlates with increasing OM

content in the sediment. Isotopic and C/N data from present-day biological communities near the study sites and other Vermont lakes (Lini et al., 2000), suggests that δ^{13}C in the sediments reflects the gradual establishment of aquatic primary producers after these post-glacial lakes were formed. Furthermore, biomarker data (Kruge and Lini, 2002) and biological indicators (pollen and diatoms) show substantial shifts associated with lake water chemistry changes and catchment-vegetation processes (Lini et al., 2002).

Although isotopic shifts of comparable magnitude are found in all four Vermont lakes, the age data indicate that these shifts are neither synchronous nor of the same duration (Galster, 2001). The duration of the shifts is variable, ranging from less than 500 ^{14}C years (Ritterbush Pond and Duck Pond) to about 3000 ^{14}C years (Sterling Pond). The data suggests that the response of terrestrial vegetation and aquatic biota to deglaciation might be more immediate in lower altitude lakes. However, the difference between the δ^{13}C records cannot be attributed to elevation alone. The most rapid development of aquatic primary producers occurs in the lakes that have larger inputs of coarse-grained terrestrial sediments, yielding greater wash-in of nutrients (Galster, 2001).
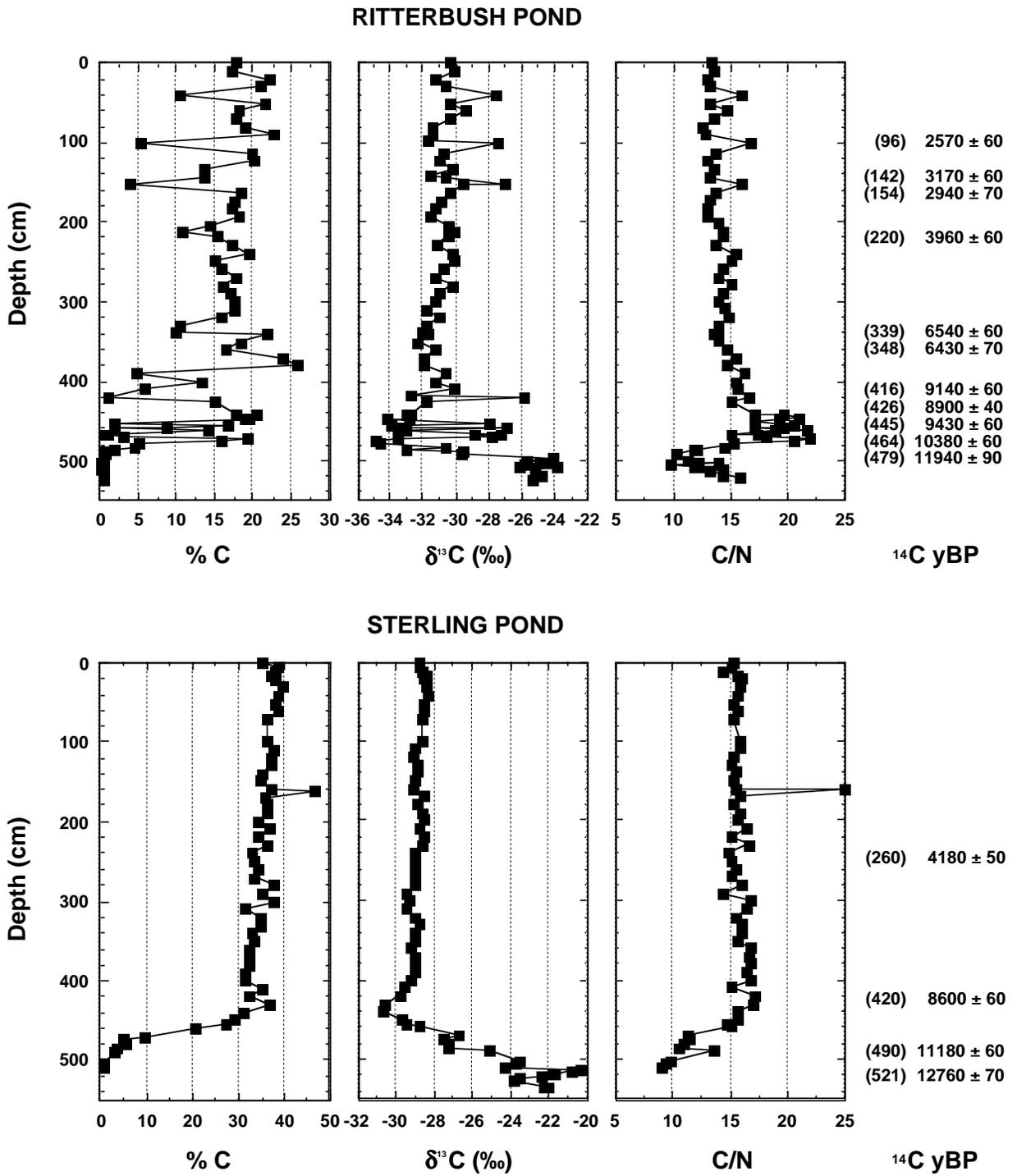
RITTERBUSH POND

(96)	2570 ± 60
(142)	3170 ± 60
(154)	2940 ± 70
(220)	3960 ± 60
(339)	6540 ± 60
(348)	6430 ± 70
(416)	9140 ± 60
(426)	8900 ± 40
(445)	9430 ± 60
(464)	10380 ± 60
(479)	11940 ± 90

STERLING POND

(260)	4180 ± 50
(420)	8600 ± 60
(490)	11180 ± 60
(521)	12760 ± 70

Figure 2.3. %C, δ^{13}C, C/N ratios, and ^{14}C AMS dates for Ritterbush Pond and Sterling Pond, VT (modified after Lini, 1999).

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DUCK POND

LAKE MOREY

Figure 2.4. %C, δ^{13}C, C/N ratios, and ^{14}C AMS dates for Duck Pond and Lake Morey, VT (modified after Lini, 1999).

22
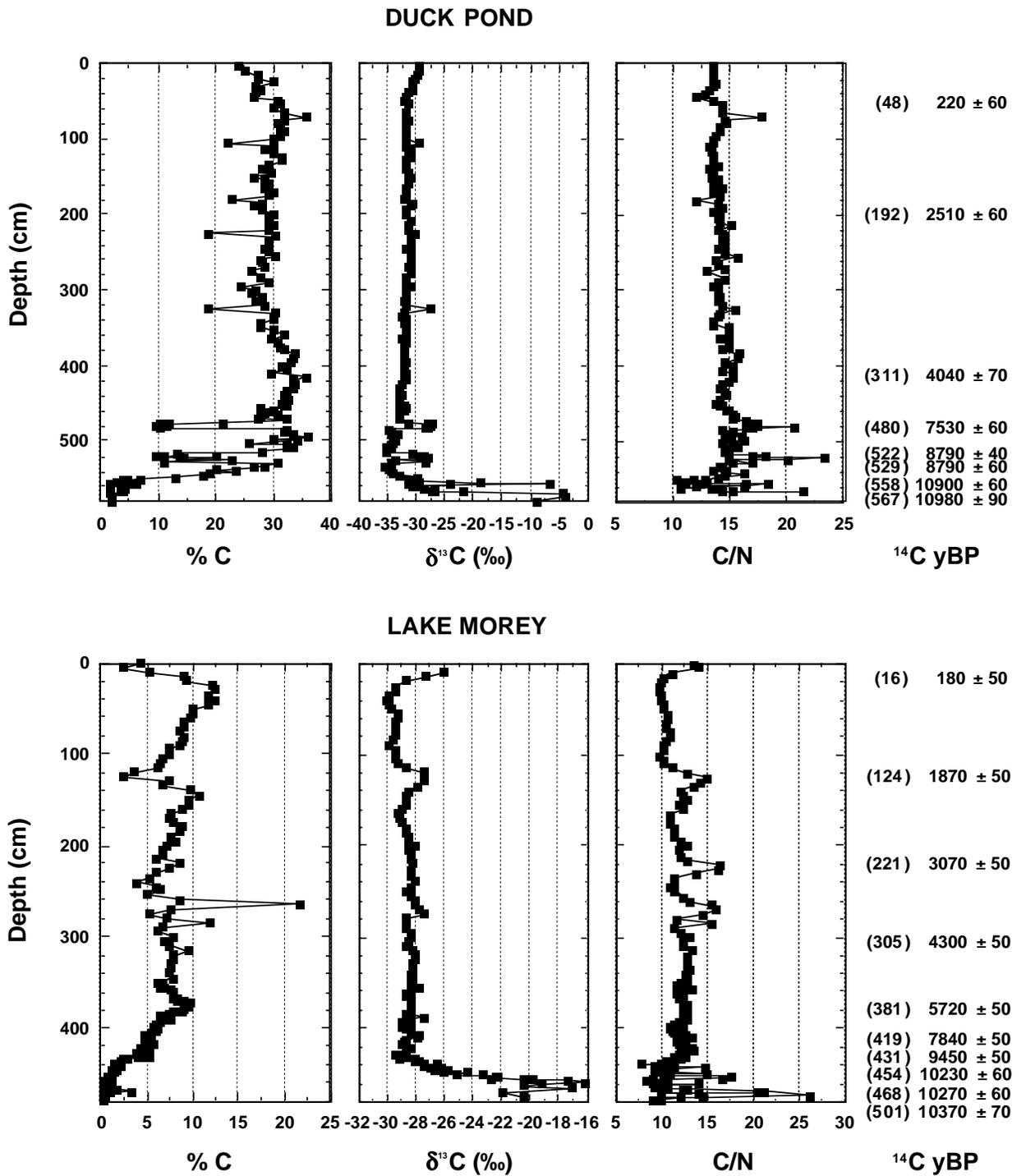
Chapter 3. STUDY AREA AND METHODS

3.1 Study Area

A total of five lakes were analyzed for this study: Stinson Lake, South Pond, and

Crystal Lake in New Hampshire, and Johns Pond and Worthley Pond in Maine (Figure

3.1). The surficial cover in all five watersheds consists of glacial till, and all of the lakes

are situated in carbonate-free bedrock (Sollar and Packard, 1998).

The following observations were made during fieldwork conducted in the summer of

2001. The Stinson Lake watershed consists of a well-forested area with both deciduous

and coniferous vegetation. Few macrophytes were observed along the shoreline. The

area surrounding South Pond is a mixed deciduous and coniferous forest. Bedrock is

exposed in small sections and few macrophytes were observed along the shoreline.

Crystal Lake is surrounded by mixed deciduous and coniferous vegetation. Macrophytes

were present along the shoreline, which is heavily wooded except in the areas of the

public sandy beaches. A 1975 report by the U.S. Geological Survey lists both

submergent and emergent vegetation as being scant, and the lake water as colorless for all

three of the New Hampshire lakes (Cowing and Lash, 1975). All lakes in this study

appear to be oligotrophic (i.e., have low aquatic production).

The vegetation in the Johns Pond watershed consists primarily of evergreens and

the shoreline is heavily wooded. Boulders and some macrophytes were observed along

the shoreline. The area surrounding Worthley Pond is comprised of deciduous trees with

a few evergreens, and is well developed with cabins and homes. Macrophytes were
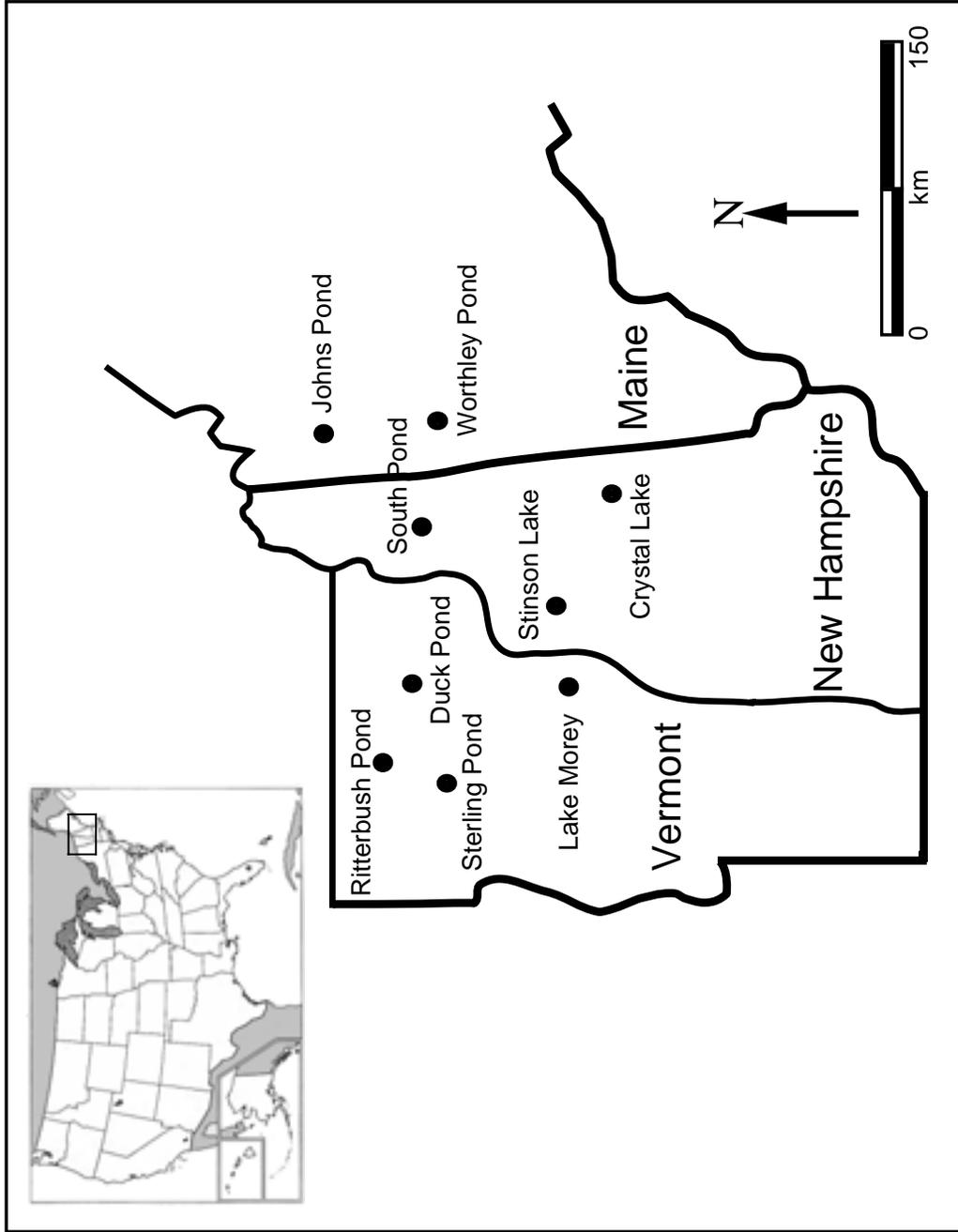
Figure 3.1. Locations of the studied lakes in northern New England. Those in Vermont are from previous studies by Lini (1999) and Galster (2001).

observed near the beach areas, and blue-green algae were visible on the water surface. Other physical characteristics of the lakes are shown in Table 3.1.

3.2 Field Methods

Winter Coring

One sediment core was obtained from each study lake. Cores were collected using a modified Reasoner coring device (Reasoner, 1993) (Figure 3.2), and ranged in length from 4 to 8 meters. Coring took place from January through March 2001, when the lakes were ice covered, to take advantage of the ice for use as a stable platform.

Sediment was collected using a 6 meter (m) PVC pipe fitted with a piston. A core head was bolted to the top of the pipe, and the apparatus was lowered through a hole augured in the ice. The PVC pipe was slowly pounded into the sediment using a 50 pound steel cylinder hammer, which was attached to a rope and lowered through the ice. Each core tube was pounded into the sediment until full or until met with refusal from clay or bedrock layers. In the case of Crystal Lake, two core tubes were used in succession to retrieve the entire Holocene record, yielding a total core length of approximately 8 m. For ease of transport, each full core tube was cut into 1.5 m sections, capped, and sealed in the field. All cores were stored at 5°C until processed.

Summer Coring

Summer coring was conducted in July 2001. A Glew gravity coring device (Figure 3.3) (Glew, 1989) was used to obtain short (24 to 40 centimeter) cores in order to

Table 3.1 Physical characteristics of the studied lakes. Lakes in New Hampshire and Maine were cored as part of this study. Vermont lakes were studied by Lini (1999) and Galster (2001) and are listed for comparison.

Lake Name	Surface Area (km^2)	Maximum Depth (m)	Elevation (m)	Drainage Basin Area (km^2)	Drainage Basin Relief (m)	Latitude	Longitude
New Hampshire:							
Crystal Lake	0.38	18	146	15.0	353	N 43°54'	W 71°05'
Stinson Lake	1.40	22	396	20.7	655	N 43°52'	W 71°48'
South Pond	0.70	26	340	7.4	427	N 44°36'	W 71°22'
Maine:							
Johns Pond	1.08	15	533	18.2	384	N 45°04'	W 70°46'
Worthley Pond	1.43	15	174	13.5	344	N 44°24'	W 70°26'
Vermont:							
Duck Pond	0.03	14	520	0.7	290	N 44°42'	W 72°04'
Lake Morey	2.22	13	127	20.7	414	N 43°55'	W 72°09'
Ritterbush Pond	0.05	14	317	2.2	293	N 44°45'	W 72°36'
Sterling Pond	0.03	9	917	0.3	40	N 44°33'	W 72°47'

Figure 3.2. Modified Reasoner coring device used to collect the 4-8 m sediment cores.

recover undisturbed sections of the top (recent) sediments. The cores were extruded in

the field at 1 centimeter (cm) intervals using an Glew core extruding device (Glew,

1988). Each centimeter of sediment was scraped into a plastic vial, frozen, and then

placed in a freeze-drier for 48 hours to remove all water within the sample. Since this

study concentrates on the OM content of the older sediments in the core, the recent

sediments will be used in future studies and are not discussed further in this thesis.

Biological Sampling

Samples of terrestrial plants, macrophytes, and phytoplankton were collected

from the five lakes to determine the elemental and isotopic composition of the present-

day vegetation within each lake and its watershed. Individual plant samples were stored

in Ziploc bags and frozen. Macrophyte and terrestrial plant samples were identified by

common names (Appendix A).

For phytoplankton sampling, one 20 liter carbuoy was filled with water from each

lake. The water was pre-filtered through a 250 μm sieve to remove any zooplankton,

then filtered through a 1.2 μm quartz fiber filter (Whatman GF/C) in the field to collect

the phytoplankton. Each filter was placed in an oven at 60 °C until dry, and stored in a

plastic case.

3.3 Laboratory Methods

Each 4 to 8 m sediment core was opened and split in half lengthwise. One half was

immediately wrapped and sealed in plastic, then placed in cold storage for archive
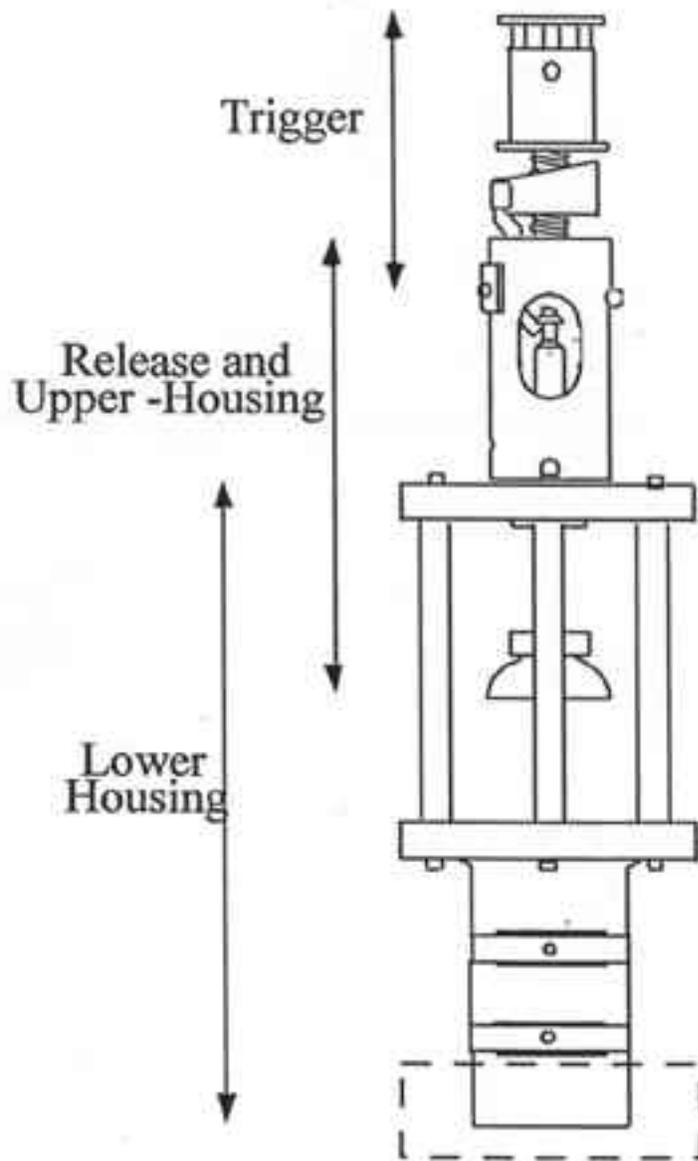
Figure 3.3. Glew gravity coring device used to obtain the 24-40 cm sediment cores. The dashed box at the base of the lower housing is the attach-point for a clear PVC tube, used to collect the sediment.

purposes. The other half was photographed and a visual log was made noting changes in color and grain size. The sediments were sampled at a 1 cm resolution, with each sample placed into a plastic vial, frozen, and then placed in a freeze-drier for 48 hours to remove all water within the sample. Any observed macrofossils were collected from the cores during sampling. Approximately 320 macrofossils were removed for possible use in radiocarbon dating of the sediment layers.

%C, %N, C/N Analyses

The total amount of carbon (%C) and nitrogen (%N) in each sediment sample was analyzed at a 1 to 10 cm resolution for each core. Between 5 and 200 milligrams (mg) of each sample were placed into a tin capsule, sealed, and analyzed on a CE Instruments NC 2500 elemental analyzer (EA). To confirm EA calibration, a standard, NIST-1547 (Peach Leaves: 46.34 %C, 2.94 %N), was run every 15 samples throughout the analyses. The precision of the analyzer was approximately 1% of the quantity measured for %C, and 0.5% for %N. The amount chosen for analysis was dependent on the estimated organic matter (OM) content obtained from loss-on-ignition records. C/N ratios were calculated from the %C and %N data.

%C, %N, and C/N ratios were also determined for the collected macrophytes, phytoplankton, and watershed plants. All plant samples were first washed and then dried in an oven at 60°C. The dried plants were milled into a powder and placed in plastic vials. Between 3 and 5 mg of each sample were placed into a tin capsule, sealed, and analyzed on a CE Instruments NC 2500 elemental analyzer. For phytoplankton samples, 3/4 of each filter was placed in a tin capsule and run as described above.

Stable Carbon Isotopes ($\delta^{13}C$)

Carbon has two non-radioactive isotopes, carbon-12 (^{12}C) and carbon-13 (^{13}C). Stable carbon isotope analysis compares the ratio of ^{13}C in the form of carbon dioxide ($^{13}CO_2$) to ^{12}C ($^{12}CO_2$) in a sample, and the value is reported using the $\delta^{13}C$ notation (Equation 2). The ratio (R) of $^{13}C/^{12}C$ in the sample is compared with that of a standard and reported in units of per mil (‰) as follows:

$$\delta^{13}C \text{ (‰)} = [(R_{sample} / R_{standard}) - 1] * 1000 \qquad (2)$$

Carbon isotope compositions are reported as deviation from the hypothetical Vienna-PDB standard (V-PDB). V-PDB replaces the exhausted PDB standard, which was the original standard material for carbon isotope analysis (Hoefs, 1997).

Stable carbon isotope analyses were performed on all of the cores at a 1 to 10 cm resolution. Between 5 and 2000 mg of each sample were placed in a quartz tube with 600 mg of cupric oxide (CuO) and 500 mg of copper (Cu), and the tube was sealed on a vacuum line. The samples were baked at 900°C in a furnace to combust the OM, yielding carbon dioxide, water vapor, and nitrogen in the quartz tube (Boutton, 1991b). The carbon dioxide was separated from the water vapor and nitrogen on a vacuum line using liquid nitrogen and isopropyl/liquid nitrogen cold traps. A stable isotope ratio mass spectrometer (VG Sira II) was used to analyze the CO_2, yielding the measured $\delta^{13}C$ values. As stated above, the data are reported versus the V-PDB standard. Precision of measurements is $\pm$ 0.05‰ based on replicate standards (USGS-22 and internal-lab standards).

Isotope analyses were also performed on the macrophytes, phytoplankton, and watershed plant samples collected from each lake. Three to 5 mg of each milled plant sample were placed in a quartz tube with 600 mg of CuO and 500 mg of Cu, and prepared for stable carbon isotope analysis as described above. In the case of phytoplankton samples, one quarter of a GF/C filter was rolled and placed into a 9 mm quartz tube.

Rock-Eval Pyrolysis

Rock-Eval pyrolysis was conducted on sediment samples at a 20 to 30 cm resolution. Samples were sent to the Geochemical and Environmental Analyses Laboratory at the University of Neuchatel, Switzerland for analysis.

Up to 100 mg of each milled sample were used for analysis during the two-stage process. During the pyrolysis stage, each sample was heated isothermally at 300°C for three minutes, then heated to 650°C. During the oxidation phase, each sample was heated isothermally at 400°C for three minutes, then heated to 850°C. As the sediments are heated, the hydrocarbons (HC) within the sample decompose, and four resulting main peaks are produced (Figure 3.4). The S1 peak represents the hydrocarbons released during the isothermal phase, the S2 peak represents the hydrocarbons produced between 300°C and 650°C, the S3 peak represents the CO_2 resulting from the pyrolysis of OM up to 400°C, and the S4 peak (not shown on Figure 3.4) represents the CO_2 released from residual OM below 550°C during the oxidation step. The peaks are used to calculate the amount of total organic carbon (TOC), hydrogen index (HI = mg HC/g TOC), and oxygen index (OI = mg CO_2/g TOC) for each sample (Espitalié et al., 1985). The HI and
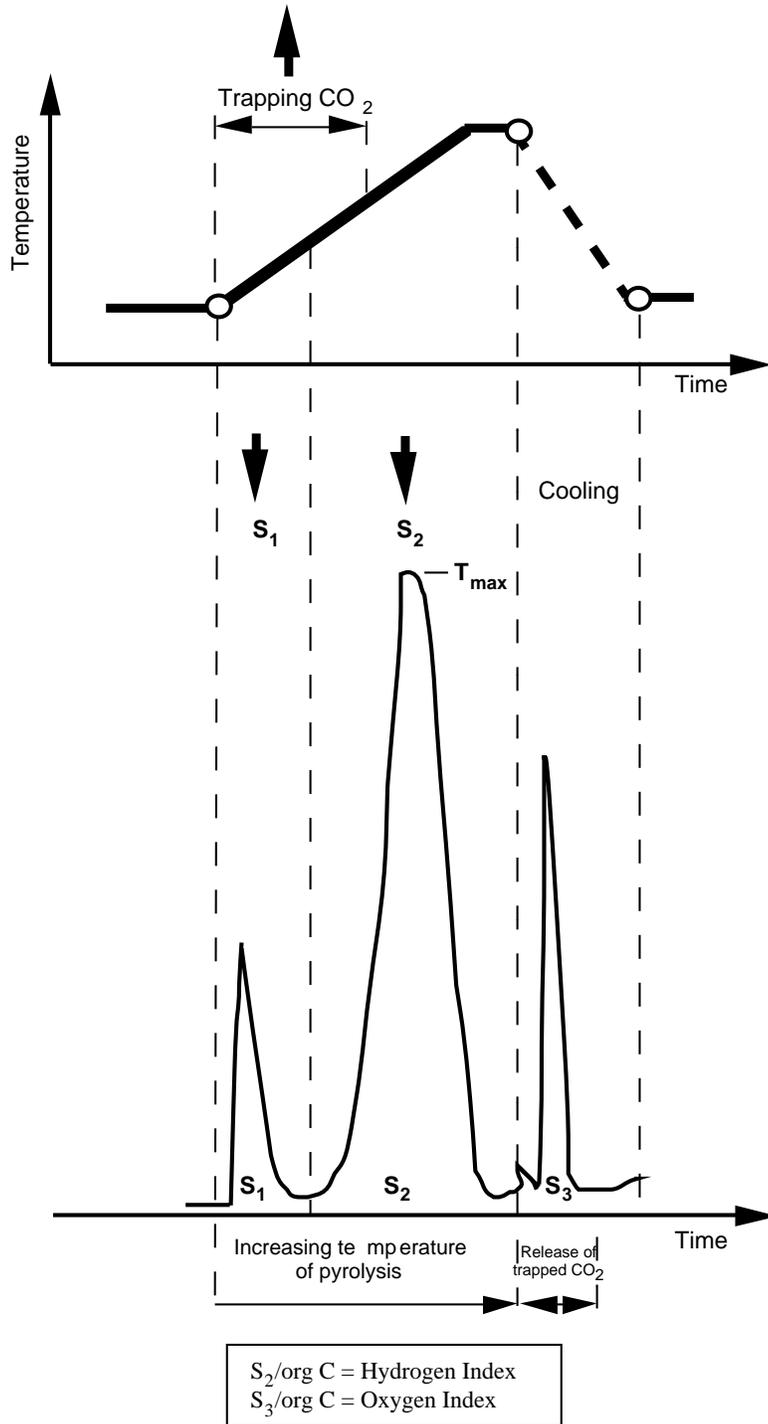
Figure 3.4. S1, S2, and S3 peaks generated during Rock-Eval pyrolysis (modified after Tissot and Welte, 1984).

OI for each sample are plotted on a Van-Krevelen type diagram (Tissot and Welte, 1984)

to determine whether the OM is from a terrestrial or lacustrine origin (Meyers and

Lallier-Verges, 1999).

Radiocarbon Dating

Carbon 14 (^{14}C) is produced in the upper atmosphere, and taken up by plants via

photosynthesis. The ^{14}C isotope decays with a half-life of 5730 years, yielding the

possibility of using the ^{14}C to determine the age of organic materials (Faure, 1986). In

the case of this study, macrofossils removed from the cores were used to date the

sediment layers in which they were located.

Radiocarbon dating was conducted at the Lawrence Livermore National

Laboratory in Livermore, CA in August 2001 and January 2002, under the guidance of

John Southon and Brian Krantz. Forty-five macrofossils obtained from the sediment

cores were dated. Macrofossils were chosen from the sediments as close to the glacial-

postglacial transition as possible, as well as from areas of the core showing fluctuations

in sediment lithology.

All macrofossils received a standard acid-base-acid treatment (HCL and NaOH)

to remove any carbonate and humic acids (Olsson, 1991). Samples were then placed in a

quartz tube with cupric oxide (CuO), and combusted to produce CO_2. The CO_2 was

isolated and graphitized on a vacuum line using liquid nitrogen and isopropyl/liquid

nitrogen cold traps, packed into targets, and analyzed on an accelerated mass

spectrometer. Uncertainties ranged from 35 to 330 ^{14}C years. The larger errors were the

result of small amounts of carbon (0.03 to 0.1 mgC) in some of the macrofossil samples.

Radiocarbon ages do not directly correspond to true calendar year ages due to changes that occurred in the production rate of atmospheric ^{14}C and changes in global ocean ventilation rates over time (Faure, 1986; Stuiver and Brazunias, 1993). Therefore, the radiocarbon dates were calibrated to calendar years by Adam Parris using CALIB rev4.3 (Stuiver and Reimer, 1993; Stuiver et al., 1998).

Age Model Calculation

Since ^{14}C dates were only obtained for those sediment layers with datable macrofossils, an age model was created for each studied lake (Appendix B). The models were calculated by first plotting the actual ^{14}C dates versus depth on a scatter plot. Best-fit lines were drawn between points, yielding an equation for each line. The formula was put into a spreadsheet, with the depth serving as the variable, and approximate ages for each centimeter were calculated. In most cases, more than one best-fit line was drawn for each lake, as sedimentation rates differed throughout the individual sedimentary records. Age models were also created for the calibrated year dates, reported in this thesis as "yr BP".

Chapter 4. ARTICLE FOR SUBMISSION TO THE JOURNAL OF PALEOLIMNOLOGY

Evolution of post-glacial lake ecosystems in northern New England: a geochemical study using lake sediments

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Key Words: Holocene, organic matter, stable carbon isotopes, deglaciation

Abstract

Sediment cores were retrieved from five small (0.38 to 1.43 km^2) post-glacial lakes in northern New England to determine differences in the timing of early Holocene ecosystem establishment on both a local and regional scale. The lower section of each core documents the transition from organic-poor sands, silts, and clays to organic-rich gyttja-type sediments, the combined result of landscape revegetation and the onset of primary production within the lakes. Using geochemical tools, such as organic carbon content (%C), carbon/nitrogen (C/N) ratios, stable carbon isotopes (δ^{13}C), and hydrogen and oxygen indices (HI, OI) from Rock-Eval pyrolysis, the relative contribution of allocthonous vs. autochthonous organic matter input through time can be estimated. The %C records and Rock-Eval data show distinct correlations with primary succession of terrestrial vegetation and the end of the Younger Dryas climate oscillation. Differences are apparent in both the timing and rate of ecosystem establishment in all of the study lakes, suggesting that lake elevation, frequency and magnitude of terrigenous inputs, and proximity to the receding Laurentide ice sheet are key factors controlling revegetation rates. Negative shifts in bulk organic matter (OM) δ^{13}C values occur in the older sediments of all five lakes. These are interpreted as reflecting changes in OM source resulting from both a climate driven switch from C$_4$ and C$_3$ tundra shrubs to C$_3$ woodland plants and increased aquatic productivity.

Introduction

Geochemical analyses of sedimentary records from post-glacial lakes in northern New England yield insight into the development of paleoecosystems on newly deglaciated barren landscapes. Alterations in the aquatic and terrestrial biological

communities in and surrounding these lakes, the result of climate and environmental changes, are recorded in the sediments deposited onto the lake bottom.

Although previous studies on lacustrine sedimentary records of similar age document the response of lakes to climate change in vegetated watersheds in non-glaciated areas (Talbot and Johannessen, 1992; Giresse et al., 1994; Talbot and Laerdal, 2000), little is known about how rapidly plant communities develop in newly formed lakes surrounded by barren, carbon and nutrient-poor watersheds. We cored lakes in New Hampshire and Maine that were deglaciated between 12,500 ^{14}C yr BP to 11,500 ^{14}C yr BP (Ridge et al., 1999; Thompson, 2001), to investigate the rates at which lacustrine ecosystems were established in newly created post-glacial lakes, and the relationship between watershed and lake ecosystem evolution.

This study focuses on the bulk organic fraction of the collected sediments. The sedimentary OM is characterized via the use of organic carbon content (%C), carbon to nitrogen (C/N) ratios, stable carbon isotopes (δ^{13}C), and hydrogen and oxygen indices (HI, OI) from Rock-Eval pyrolysis. The results are compared with those from studies done in Vermont by Lini (1999) and Galster (2001) to see if relationships exist in how post-glacial lake ecosystems developed across northern New England. Comparing the individual geochemical records of these lakes yields insight into the geomorphic, hydrologic, and biological parameters that determine the rates at which an ecosystem develops in a newly formed landscape.

Study Sites

We studied five lakes in northern New England, chosen for their watershed characteristics and ease of access (Figure 4.1). Stinson Lake is located in Rumney, Grafton County, NH. Local geology is dominated by granite typical of New Hampshire plutonic rocks (Lyons et al., 1997). South Pond, also referred to as Percy Pond, is located in Stark, Coos County, NH. The underlying bedrock consists of granite, syenite, and rhyolite (Lyons et al., 1997). Crystal Lake, also known as Robertson Pond, is located in Eaton Center, Carroll County, NH. The local bedrock is composed of schist and quartzite (Lyons et al., 1997). Johns Pond (Davis Township, Franklin County, ME) and Worthley Pond (Peru, Oxford County, ME) are both situated in epidote to amphibolite phase metasediments (Osberg et al., 1985). The surficial cover in all five watersheds consists of glacial till (Sollar and Packard, 1998). Lakes range in depth (15 to 26 m), elevation (146 to 533 m), drainage basin area (7.4 to 20.7 km^2), and steepness of drainage basin (344 to 655 m) (Table 4.1). All lakes appear to be oligotrophic and have heavily forested shorelines, with both deciduous and coniferous vegetation.

Materials and Methods

One sediment core (4 to 8 meters (m) long) was collected from each lake using a modified Reasoner piston coring device (Reasoner, 1993). Coring took place from January through March 2001 using the ice-cover as a stable platform.   For ease of transport, each full core tube was cut into 1.5 m sections, capped, and sealed in the field. All cores were stored at 5°C until processed. The cores were sampled at the University of Vermont, at a 1 cm resolution, with each sample placed into a plastic vial, frozen, and

then placed in a freeze-drier for 48 hours to remove all water within the sample. Any

observed macrofossils were collected from the cores during sampling for use in

radiocarbon dating of the sediment layers. Samples of terrestrial plants, macrophytes,

and phytoplankton were collected from the five lakes to determine the elemental and

isotopic composition of the present-day vegetation within each lake and its watershed.

%C, %N, C/N Analyses

The amount of organic carbon (%C), total nitrogen (%N), and the C/N ratio were

determined at a 1 to 10 cm resolution for each core using a CE Instruments NC 2500

elemental analyzer (EA). %C, %N, and C/N ratios were also determined for the collected

macrophytes, phytoplankton, and watershed plants.

Stable Carbon Isotopes ($\delta^{13}C$)

Stable carbon isotope analyses were performed on all of the cores at a 1 to 10 cm

resolution. Since none of the lakes is located in carbonate bedrock, it was not necessary

to pretreat the samples with acid to remove carbonates. Samples were converted to CO_2

gas according to conventional methods (Boutton, 1991b), and run on a VG Sira II stable

isotope ratio mass spectrometer. Results are expressed using the $\delta^{13}C$ notation, and

reported in units of per mil (‰) relative to V-PDB (Hoefs, 1997). Precision of

measurements is ± 0.05‰ based on replicate standards (USGS-22 and internal-lab

standards). Isotope analyses were also performed on the macrophytes, phytoplankton,

and watershed plant samples collected from each lake.

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Rock-Eval Pyrolysis

Rock-Eval pyrolysis was conducted on sediment samples at a 20 to 30 cm resolution. Samples were sent to the Geochemical and Environmental Analyses Laboratory at the University of Neuchatel, Switzerland. Up to 100 mg of each milled sample were used for analysis. Two important parameters are determined during pyrolysis, the hydrogen index (HI = mg HC/g TOC) and the oxygen index (OI = mg CO_2/g TOC). The HI and OI for each sample are plotted on a Van-Krevelen type diagram (Tissot and Welte, 1984) to determine organic matter (OM) source (Meyers and Lallier-Verges, 1999).

Radiocarbon Dating

Accelerated mass spectrometer (AMS) radiocarbon dating of 45 macrofossils was conducted at the Lawrence Livermore National Laboratory in Livermore, CA. Macrofossils were chosen from sediments as close to the glacial-postglacial transition as possible, as well as from areas of the core showing marked fluctuations in sediment lithology. Age uncertainties range from 35 to 330 ^{14}C years. The larger errors were the result of small amounts of carbon (0.03 to 0.1 mgC) in some of the macrofossil samples. All numerical ages in this paper are discussed in ^{14}C years, with corresponding calibrated year dates on Figures 4.2 through 4.6, and in Table 4.2. Since ^{14}C dates were only obtained where datable macrofossils were found, an age model was created for each lake.

Measured ^{14}C dates were plotted versus depth, and best-fit lines were drawn to determine sedimentation rates and to calculate model ages.

Results

The basal-most radiocarbon date obtained for each core varied between 10,000 ^{14}C yrs in Johns Pond to 12,300 ^{14}C yrs BP in Stinson Lake and Worthley Pond. However, similar trends in %C and δ^{13}C values and lithologic transitions from inorganic sands, silts, and clays to organic-rich gyttja-type sediments were observed in all five cores. In this section, the lakes are discussed geographically from West to East, and results are presented in reference to the accompanying figures.

Stinson Lake

Carbon content in the Stinson Lake sediments remains negligible from the bottom of the core to cm 337 (10,000 ^{14}C yr BP), then steadily increases reaching a maximum value of 10.3% at the top of the core (Figure 4.2). The C/N values are variable (8.1 to 16.5) in the lowermost 146 cm of the core, but show little change above cm 262 (7800 ^{14}C yr BP), where they fluctuate between 12.7 and 14.2 (Figure 4.2). The δ^{13}C values show a general decreasing trend throughout the core, but most markedly from the bottom sample (- 22.8‰) up to cm 100 (- 26.8 ‰) (3300 ^{14}C yr BP), after which they continue to decrease to -27.0‰ at the top of the core (Figure 4.2). The Rock-Eval data show a distinct increase in HI values from the bottom of the core to cm 247 (7400 ^{14}C yr BP), where the HI is 232. Similarly, the OI values change during the same interval, where

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they sharply decrease from 377 to 280. Both the HI and OI values show little change after cm 247 (Figure 4.2).

South Pond

In general, only minor amounts (< 1.0%) of OM are present in the bottom 210 cm of sediment in South Pond. However, the %C values increase rapidly to 11.0 by cm 184 (8800 ^{14}C yr BP) (Figure 4.3). From cm 184 to the top of the core, the %C values fluctuate between 8 and 14, with a generally increasing trend. The lowermost 40 cm of the core have C/N values below 10. Starting at cm 390, C/N ratios increase significantly and show marked fluctuations (minimum 11, maximum 23). The top 220 cm are characterized by lower C/N ratios ranging from 11 to 14 (Figure 4.3). The δ^{13}C values show a decreasing trend from the base of the core (- 21.7‰) to cm 184 (8800 ^{14}C yr BP), where they reach a minimum of - 27.6‰. Little change is observed in the rest of the core, with δ^{13}C values are between - 27.6‰ and - 26.9‰ (Figure 4.3). Overall, HI values increase from the bottom to the top of the core, while the OI values decrease. However, these two records show significant fluctuations: HI values range from 22 to 258, and OI values range from 68 to 420 (Figure 4.3).

Crystal Lake

The lowermost 125 cm of the Crystal Lake core contain insignificant amounts (~ 0.2%) of C, however %C values increase markedly between cm 690 and 560. A maximum of 16% C is reached at cm 371 (5800 ^{14}C yr BP). From cm level 660 (10,900 ^{14}C yr BP) up to the top of the core, C/N values remain constant (~ 13.5), but the C/N

values in the lowermost, C-poor sediments decrease sharply from 34 to 6 (Figure 4.4).

The $\delta^{13}C$ values decrease gradually from the bottom sediments through cm 560 (-28.5‰)

(9100 ^{14}C yr BP), then show little change through the top of the core (Figure 4.4). The

$\delta^{13}C$ values for cm 815 (-7.1‰) and 800 (-14.2‰) are much less negative than those

observed in any of the other studied lakes. Repeated treatment with hydrochloric acid

(HCl) did not alter the isotopic composition of these samples; therefore, they indicate the

presence of ^{13}C-enriched OM. The HI values increase from 200 to 350 from the bottom-

most sample up to cm 425, then decrease to 275 and show little change through the top of

the core. OI values decrease from the bottom sample up to cm 425 (190 to 128), after

which they gradually return to values ~180 (Figure 4.4).

Johns Pond

Small amounts of C (< 0.5%) are present in the lower 175 cm of the core; the %C

increases markedly to 10% by cm 150 (4100 ^{14}C yr BP) and then remains fairly consistent

(Figure 4.5). The C/N values are variable (7.6 to 13.5) from the base of the core through

cm 220, then display a sharp increase up to cm 180 (15.3), and remain relatively stable

through the top of the core (Figure 4.5). The oldest AMS date (10,360 ± 130 ^{14}C yr BP)

for Johns Pond corresponds to cm 240. However, the observed changes in C/N occur

much lower in the core, and are thus undatable. We assume an age of ~12,000 ^{14}C yr BP

for the oldest sediments, in correlation with the estimated deglaciation date of the area

(Davis and Jacobson Jr, 1985; Thompson, 2001). The $\delta^{13}C$ values progressively decrease

from the bottom to the top of the core. In the lowermost sediments, the values decrease

steadily from - 23.0‰ to - 25.4‰ in cm 225 (Figure 4.5). Between cm 225 and cm 208

the δ^{13}C values rapidly decrease to - 27.0‰ then increase back to - 25.8‰, correlating

with similar trends in the %C and C/N records. After cm 208 the δ^{13}C values decrease

gradually to - 27.1‰ in the top of the core. The HI values increase up-core from the

bottom sediments to cm 174 (282), then show little change through the top of the core,

while the OI values decrease up to cm 200 (170), then vary little through the surface

sediments (Figure 4.5).

Worthley Pond

The OM content in the bottom 70 cm of sediments in Worthley Pond is

negligible (< 0.3%), followed by a gradual increase up-core between cm 380 and cm 310

(2.1%), and a more marked increase by cm 285 (7.5%) (9300 ^{14}C yr BP). The %C values

vary between 5.3% and 7.8% in the top 285 cm (Figure 4.6). The C/N ratios show a

decreasing trend from the bottom of the core to cm 365 (8.6), but show little change from

cm 365 through the top of the core, varying between 12 and 13 (Figure 4.6). The δ^{13}C

values display a marked decrease in the bottom 85 cm, starting at - 21.6‰ and reaching

- 28.3‰ in cm 365 (Figure 4.6). The shift is followed by a rapid 2‰ increase through

cm 355, and then a gradual decrease to - 27.9‰ at the top of the core. The HI values

increase from the bottom-most measured sample (cm 390) to 374 in cm 230, then

decrease slightly, and show little change through the top of the core. The OI values

display a small increase in the bottom sediment samples, then decrease to 207 in cm 270,

showing little variation through the top of the core.

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Discussion

Relationship between patterns of primary succession and organic matter records

The geochemical records produced for the five study lakes display clear trends that can be related to the revegetation history of their watersheds, as illustrated by the %C records. All five lakes display a sharp increase in OM content, with maximum %C values ranging from 7.5 (Worthley Pond) to 17 (Crystal Lake). The earliest onset of increased OM deposition occurred in Worthley Pond and Crystal Lake (approximately 12,000 ^{14}C yr BP) (Table 4.3). However, a prominent increase in %C occurs in all five lakes shortly after 10,000 ^{14}C yr BP, correlating with the end of the Younger Dryas (YD) period in northern New England (Peteet et al., 1990). During the YD, a period of brief but severe cooling from approximately 11,000 to 10,000 ^{14}C yr BP, vegetation assemblages were dominated by boreal species typical of colder climates (Peteet et al., 1990). The marked %C increases occurring after the YD are the result of climate warming, yielding greater terrestrial and aquatic production. This is well-defined by the %C records in Crystal Lake and Worthley Pond, where the %C values began to increase prior to 11,000 ^{14}C yrs BP, then leveled out during the YD, and began to increase again after the conclusion of the YD.

Pollen data show that post-glacial terrestrial vegetation communities in northern New England evolved from tundra to woodlands to mature forest in a northeastward and northward direction (Davis and Jacobson Jr, 1985). The pattern of primary succession is related to changes in the amount of OM within the sediment cores (Figure 4.7). In all

five of the study lakes, the tundra/ice cover periods correlate with the primarily inorganic bottom sediments. Increasing amounts of OM are evident with the establishment of a mixed woodland environment surrounding Worthley Pond. In Stinson Lake, South Pond, Crystal Lake, and Johns Pond the most marked increase in %C correlates with the onset of mature forest. Thus we assume that sedimentary %C increases occurred only after enough vegetation had grown in and around the lakes to alter the geochemical signal preserved in their sediments. This happened earlier in Worthley Pond and Crystal Lake, most likely due to their lower elevation (Table 4.1), which allowed for longer growing seasons for the plant communities surrounding the lakes, and therefore earlier contribution of OM to the sediments.

The succession of terrestrial vegetation corresponds to lithological changes in the cores defined by variations in color and sediment grain size. Sediments from Stinson Lake and South Pond have two distinct lithological transitions, whereas sediments from the other lakes have only one. The older transition is marked by alternating light brown and light gray laminae, grain size changing from silts in the lower sediments to very fine sand, and can be linked to the shift from tundra to woodland vegetation. The lithological change occurring further up-core in all five lakes is marked by distinct changes from gray to brown fine sediments. These younger changes in lithology correlate with the transition from woodland to forest vegetation in all locations except Stinson Lake, where the sediment lithology changes only after the development of mature forest.

The observed lithological changes and the occurrence of discrete terrigenous layers in some of the cores suggest a relationship between terrigenous inputs and

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establishment rates of aquatic communities. Sediments of larger grain size typically

indicate deposition of higher amounts of terrestrial material into a lake (Silliman et al.,

1996; Bierman et al., 1997; Meyers and Takemura, 1997; Brown, 1999; Brown et al.,

2000; Brown et al., 2002; Noren et al., 2002). Nutrients such as N and P are washed into

the lake along with the inorganic material, thus favoring aquatic OM development

(Meyers and Takemura, 1997).

Aquatic production is tracked via the covariance between the HI and %C records,

as well as the decreasing trends in C-isotope values (see *Post-glacial C-isotope shift*

section). When using Rock-Eval pyrolysis data, OM of Type I, with HI values up to 800,

tends to be of lacustrine origin.   Type II OM, with HI values up to 500-600, can be from

either marine or lacustrine origin, depending on the dominant algal community (Ariztegui

et al., 1996b). Type III OM tends to be of terrestrial origin, with HI values typically

between 150-300. In all of the study sites except Crystal Lake, the HI values in the oldest

sediments are very low (10-83), but increase roughly in correspondence with the onset of

increased OM deposition, reaching values as high as 370 in Worthley Pond. OI values

show an opposite trend; they are highest in the oldest sediments (350-420) and generally

decrease up-core. The low HI values observed at the bottom of the cores probably reflect

the presence of highly degraded OM in the predominantly silty-sandy inorganic

sediments. As both greater aquatic productivity and OM preservation occur, the HI

values increase (Patience et al., 1996; Ariztegui et al., 1996a; Ariztegui et al., 1996b).

Increased aquatic productivity is also suggested by the inverse relationship of the HI and

OI values. This reflects a change from hydrocarbon-rich to hydrocarbon-poor OM, likely

the result of increased algal contribution (Ariztegui et al., 1996b).

Timing and duration of ecosystem establishment

Since %C in sediments reflects both the amount of OM entering the lake basin

from the watershed and production in the lake itself (Meyers, 1997; Meyers and Lallier-

Verges, 1999), the onset of significant OM deposition is likely related to the timing and

rates of terrestrial and aquatic ecosystem development. Thus, we can estimate rates of

land and aquatic ecosystem establishment via the duration of the %C shifts (Table 4.3).

Shift duration is defined as the time interval between the onset of the %C increase and the

point where the %C values become relatively constant. These intervals vary from 800

(South Pond) to 5200 ^{14}C years (Johns Pond). Since the %C values in Stinson Lake

increase continuously from 10,000 ^{14}C yrs BP through the top of the core, a shift duration

was not estimated for this lake. We relate differences in establishment rates to the

diverse physical characteristics of the study lakes. For instance, Johns Pond is located at

the highest elevation (533 m) and shows the longest transition from inorganic to organic-

rich sediments. Vegetation at higher elevations has a shorter growing season, and lower

mean annual temperatures diminish net aquatic productivity.

The onset of increased OM deposition occurs in conjunction with a stabilization

of the δ^{13}C and C/N values. We interpret the variations in δ^{13}C and C/N values prior to

the %C increase to be the result of periodically changing ratios of terrestrial vs. aquatic

OM deposition, tracking the development of aquatic and land plants and reflecting a time

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of unstable landscape. In all five lakes, the variable C/N ratios correlate with changes in sediment grain size as measured by Parris et al. (2002): higher C/N values are found in the coarse grained (very fine - to medium sand) intervals, while lower C/N values are observed in the fine grained (clay - silt) intervals. We interpret the coarser grained sediments as inputs from the surrounding watershed, depositing layers of inorganic sediment and terrestrial OM, thus diluting the amount of aquatic OM in the measured sample. Overall, Crystal Lake, Johns Pond, and Worthley Pond (Figures 4.4, 4.5, and 4.6 respectively) display a decreasing trend in C/N values until the increase in %C occurs. In Stinson Lake (Figure 4.2) and South Pond (Figure 4.3) C/N ratios fluctuate significantly in the lower portion of the cores, with the highest values corresponding to silt and sand layers containing plant macrofossils. The very low C/N values obtained in the bottom-most, clay-rich and OM-poor (< 0.3 %) sediments of South Pond are likely due to inorganic N absorption on the clays (Meyers, 1997). Once the proportions of terrestrial and aquatic OM entering the sediment stabilize, little change is seen in the $\delta^{13}C$ and C/N values; however, the continued %C increase is evidence of sustained growth within the plant communities.

Post-glacial C-isotope shift

Trends toward more negative $\delta^{13}C$ values are apparent in the older sediments of all five study lakes. We attribute this trend primarily to changes in the relative contributions of terrestrial vs. algal OM to the sediments during early stages of lake development. Although a distinction between algal and terrestrial matter can not always

be made using $\delta^{13}C$ values alone (Meyers, 1997), algal OM has often been found to have more negative $\delta^{13}C$ values than its terrestrial counterparts (Talbot and Johannessen, 1992; Talbot and Laerdal, 2000). In northern New England, terrigenous layers in Ritterbush Pond, VT, show less negative $\delta^{13}C$ values (- 27‰) than the surrounding gyttja layers (- 31‰) (Bierman et al., 1997), and isotopic analyses on modern terrestrial and aquatic plants throughout Vermont generally show more negative $\delta^{13}C$ values in aquatic plants (Lini et al., 2000). Stable isotope analyses on modern plants surrounding the five study lakes in NH and ME display similar results (Table 4.4).

Since the $\delta^{13}C$ values in the oldest sediments are less negative than those further up-core, it appears that the terrestrial ecosystems surrounding the lakes became established prior to the aquatic vegetation. The growth of land plants, soil development, and subsequent sediment deposition of land plant and soil material contributing N and P nutrients (Meyers and Takemura, 1997) enhanced the productivity of the aquatic ecosystem. This "fertilization" jump started the aquatic ecosystems, resulting in increasingly depleted sediment $\delta^{13}C$ values over time. Scant variation in $\delta^{13}C$ values further up-core indicates fully established aquatic and terrestrial ecosystems. The C/N records provide additional evidence for the shifting proportions of terrestrial OM (C/N > 20) vs. aquatic OM (C/N < 10; (Meyers and Ishiwatari, 1993; Meyers, 1997; Meyers and Lallier-Verges, 1999) over time. Consistent with the negative $\delta^{13}C$ shift, the decreasing C/N values as lakes age reflect greater input of algal OM into the sediments as aquatic productivity increased.

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During the transition period from tundra to mixed woodland vegetation, both C_4 and C_3 plants might have coexisted in the watersheds, resulting in the carbon stable isotope values around -22.5‰ we measured in the oldest sediments of all five lakes. Similar values have been reported for the late glacial early post-glacial time interval in several African lakes, where C_4 plants were gradually replaced by C_3 plants as the climate became more humid (Talbot and Johannessen, 1992; Giresse et al., 1994; Street-Perrott et al., 1997). The change to C_3 plant dominated woodlands that occurred in northern new England, combined with additional influence from a more productive algal community, would explain the trend over time toward more negative $\delta^{13}C$ values (up to - 28‰), similar to what has been documented in studies of African and Siberian lakes (Talbot and Johannessen, 1992; Qui et al., 1993; Street-Perrott et al., 1997; Talbot and Laerdal, 2000). If we exclude the two oldest samples from Crystal Lake, all five lakes display negative $\delta^{13}C$ shifts of magnitudes (4‰ to 7‰) that are comparable to those reported for other lakes during the early Holocene (Stuiver, 1975; Meyers and Horie, 1993; Giresse et al., 1994; Talbot and Laerdal, 2000).

The duration of the negative $\delta^{13}C$ shifts in NH and ME lakes is variable, and ranges from 2200 ^{14}C years (South Pond) to 3200 ^{14}C years (Johns Pond) (Table 4.3). As with the %C trend, the $\delta^{13}C$ values in Stinson Lake decrease continuously from 12,400 ^{14}C yrs BP through the top of the core, and a shift duration was not estimated for this lake. The shifts of shorter duration in South Pond, Crystal Lake, and Worthley Pond are attributed to more frequent, episodic inputs of terrigenous sediment carrying nutrients

into these lakes and allowing for a more rapid development of the aquatic ecosystems than in the other two lakes.

We attribute the negative $\delta^{13}C$ shift mostly to changing contributions of terrestrial and aquatic OM to the sediment; however, the sediment C-isotope values may have been influenced by additional factors. The input of ^{13}C-depleted carbon from CO_2 produced during soil respiration and plant OM decomposition, as well as HCO_3^- derived from dissolution of silicate bedrock can shift the $\delta^{13}C$ composition of dissolved inorganic carbon (DIC) in a lake towards more negative values (Hakansson, 1985; Hammarlund et al., 1997). This shift would be reflected by the isotopic composition of aquatic plants, which utilize DIC for photosynthesis. Changes in the isotopic composition of marine OM (Jasper and Hayes, 1990; Rau et al., 1991; Fontugne and Calvert, 1992) and terrestrial plants (Krishnamurthy and Epstein, 1990; Leavitt and Danzer, 1992; Leavitt, 1993) have been attributed to increased concentrations (from 200 ppmv to 280 ppmv; Raynaud et al., 1993) and decreasing isotopic values (Yapp and Epstein, 1977) of atmospheric CO_2 during the glacial-postglacial transition. However, the magnitude of the negative isotopic shifts found in the marine and land plant records is significantly smaller (1-3‰) than the shifts observed in most lakes, and an increase in pCO_2 cannot explain the large positive shifts found in some hard-water lakes (Stuiver, 1975; Aravena et al., 1992). Furthermore, if a change in atmospheric pCO_2 was indeed the main trigger for the isotopic shifts, these would be more or less synchronous and of similar duration, which they are not.

It is unlikely that diagenetic processes are the cause for the observed changes in sediment geochemistry. Although these processes are at work during the sinking of OM through the water column, reducing the amount of OM reaching the sediments (Meyers and Eadie, 1993; Schelske and Hodell, 1995), C/N ratios and OM δ^{13}C values typically do not display large changes, especially in oligotrophic, relatively shallow lakes such as those in our study (Ertel and Hedges, 1985; Talbot and Livingstone, 1989; Meyers and Eadie, 1993).

Comparison across northern New England

To track trends in post-glacial revegetation patterns across northern New England, we compare the geochemical records of the lakes in NH and ME with those of lakes previously studied in Vermont: Duck Pond, Lake Morey, Ritterbush Pond, and Sterling Pond (Figures 4.1, 4.8, and 4.9) (Bierman et al., 1997; Lini, 1999; Galster, 2001).

The onset of OM deposition in Crystal Lake, Worthley Pond and three of the Vermont lakes (Duck Pond, Sterling Pond, and Ritterbush Pond) occurs earlier (> 10,000 ^{14}C yr BP) than in Lake Morey, Stinson Lake, South Pond, and Johns Pond (9000 to 10,200 ^{14}C yr BP) (Table 4.3). This might be attributed to transitions from tundra to woodland vegetation, occurring earlier in the former lakes (Davis and Jacobson Jr, 1985). Lake Morey constitutes a special case since it is situated within the basin of former Glacial Lake Hitchcock, which did not drain until ~10,400 ^{14}C years BP (Ridge et al., 1999). Thus, the timing of OM -rich sediment deposition and forest revegetation around Lake Morey is closely tied to the drainage of Lake Hitchcock (Galster, 2001). The areas around Stinson Lake, South Pond, and Johns Pond were deglaciated later (12,300 to

12,000 ^{14}C yrs BP) than the other lakes in this study (Ridge et al., 1999; Thompson et al., 1999; Thompson, 2001), and may have received high volumes of glacial melt water and inorganic sediment due to being in closer proximity to the retreating ice sheet. This is compounded by the timing of a glacial re-advance in the Littleton-Bethlehem area of NH ~12,000 ^{14}C yrs BP (14,000 - 14,100 yr BP) (Ridge et al., 1999; Thompson et al., 1999), to the northwest of Stinson Lake and southwest of South Pond.

Duck Pond, Ritterbush Pond, and Sterling Pond display much higher %C values (> 30%) than Lake Morey and the lakes in NH and ME (maximum ~18%). We attribute the lower %C values in Lake Morey to greater amounts of glacial lake clays being washed in from the former catchment basin of Lake Hitchcock. Stinson Lake, South Pond, and Johns Pond also received a large amount of inorganic detrital material during the early Holocene due to their close proximity to receding glaciers, and thus, display lower %C values in the oldest sediments. The drainage basin relief of the lakes in NH and ME is higher than those surrounding the VT lakes, and we ascribe the overall lower %C values to these steeper drainage basins, delivering larger amounts of terrestrial material to the lakes. Although terrestrial OM may supply nutrients to the lake as described earlier, large amounts of terrestrial inorganic matter can depress the OM signal in the sediments.

The NH and ME lakes display negative δ^{13}C shifts of smaller magnitude (maximum of 7‰ vs. 15‰) and reach less negative δ^{13}C values (- 28.5‰ vs. - 35.3‰) than the VT lakes. The Vermont DEC lists waters of the VT lakes as mesotrophic to eutrophic (VT Department of Environmental Conservation, 2002), as opposed to the

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oligotrophic conditions we observed in the NH and ME lakes. Although we cannot be certain that the trophic states of the lakes were the same in the early Holocene as they are today, smaller contributions of phytoplankton-derived OM to the sediments of the NH and ME lakes could explain the smaller magnitude of the observed $\delta^{13}C$ shift.

In general, the negative $\delta^{13}C$ shifts lasted longer in the NH and ME lakes. The geochemical records in the VT lakes also show a clear covariance between decreasing $\delta^{13}C$ values and increasing %C values (Galster, 2001). However, this correlation was not observed in the NH and ME cores and clearly deserves further investigation.

Conclusions

In five post-glacial northern New England lakes, the change from inorganic to OM-rich sediments correlates with the primary succession of vegetation in the watersheds, in particular with the establishment of mature forest ecosystems following the end of the Younger Dryas cooling period. However, the timing and rates of ecosystem development vary as a result of differences in elevation, proximity to the receding ice margin, and the frequency and magnitude of terrigenous inputs contributing nutrients to the developing lakes.

Fluctuations in the $\delta^{13}C$ and C/N records in the oldest, mostly inorganic, sediments correlate with changes in sediment lithology and reflect episodic changes in OM source (terrestrial vs. aquatic) in a not yet fully vegetated landscape. The negative trends in $\delta^{13}C$ documented in the lakes are also attributed to changes in OM source, at first via a climate-induced switch from mixed C_4 and C_3 tundra grasses to C_3 vegetation,

followed by increased aquatic primary productivity. Processes such as increased input of isotopically depleted DIC from developing watershed soils, changes in the concentration and isotopic composition of atmospheric CO_2 during the early Holocene, and diagenesis might be attributed to the negative shifts in $\delta^{13}C$, but their role was probably minor.

The results of this study imply that even on a local scale the response of different lakes to climate change is not identical. This is supported by the documented differences in timing and rate of post-glacial lake evolution. We thus suggest that records of geochemical proxies such as $\delta^{13}C$ and %C, often used for regional and global correlation, may in fact be the result of biological rather than contemporaneous climatic factors.

Acknowledgements
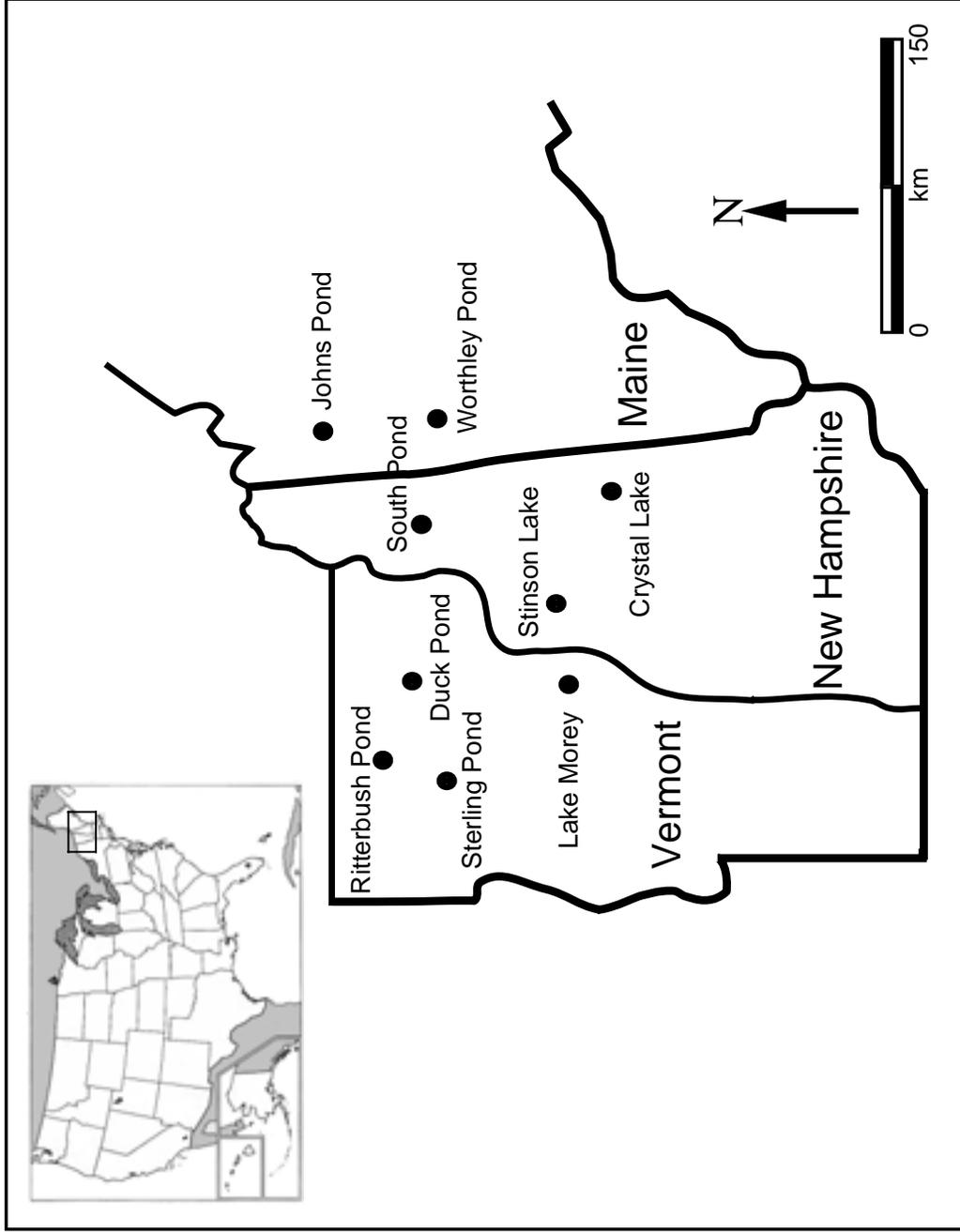
Figure 4.1. Locations of the studied lakes in northern New England. Those in Vermont are from previous studies by Lini (1999) and Galster (2001).
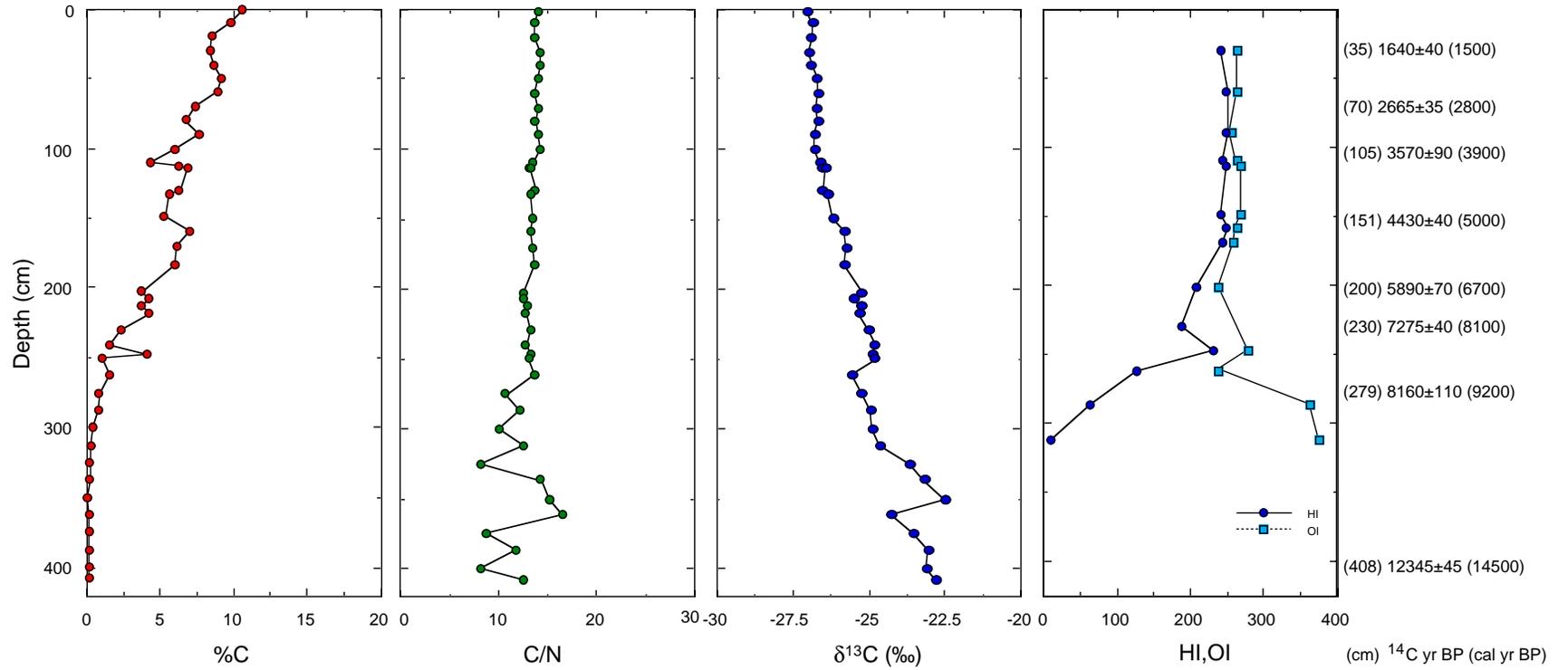
Figure 4.2. %C, C/N, δ^{13}C, and Rock-Eval data for Stinson Lake, NH. Dates are presented as both ^{14}C yrs BP and calibrated yrs BP.

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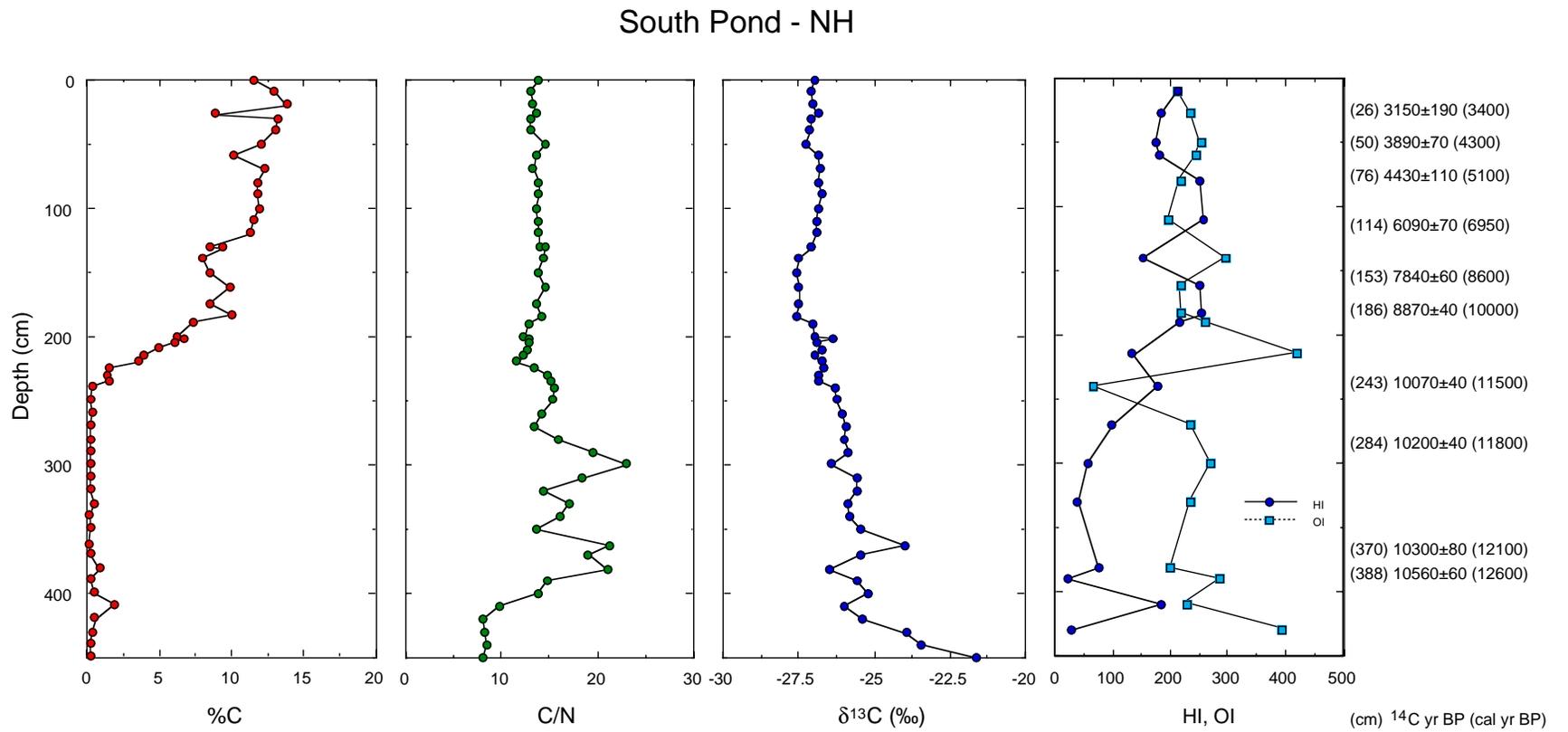
South Pond - NH

Figure 4.3. %C, C/N, δ^{13}C, and Rock-Eval data for South Pond, NH. Dates are presented as both ^{14}C yrs BP and calibrated yrs BP.
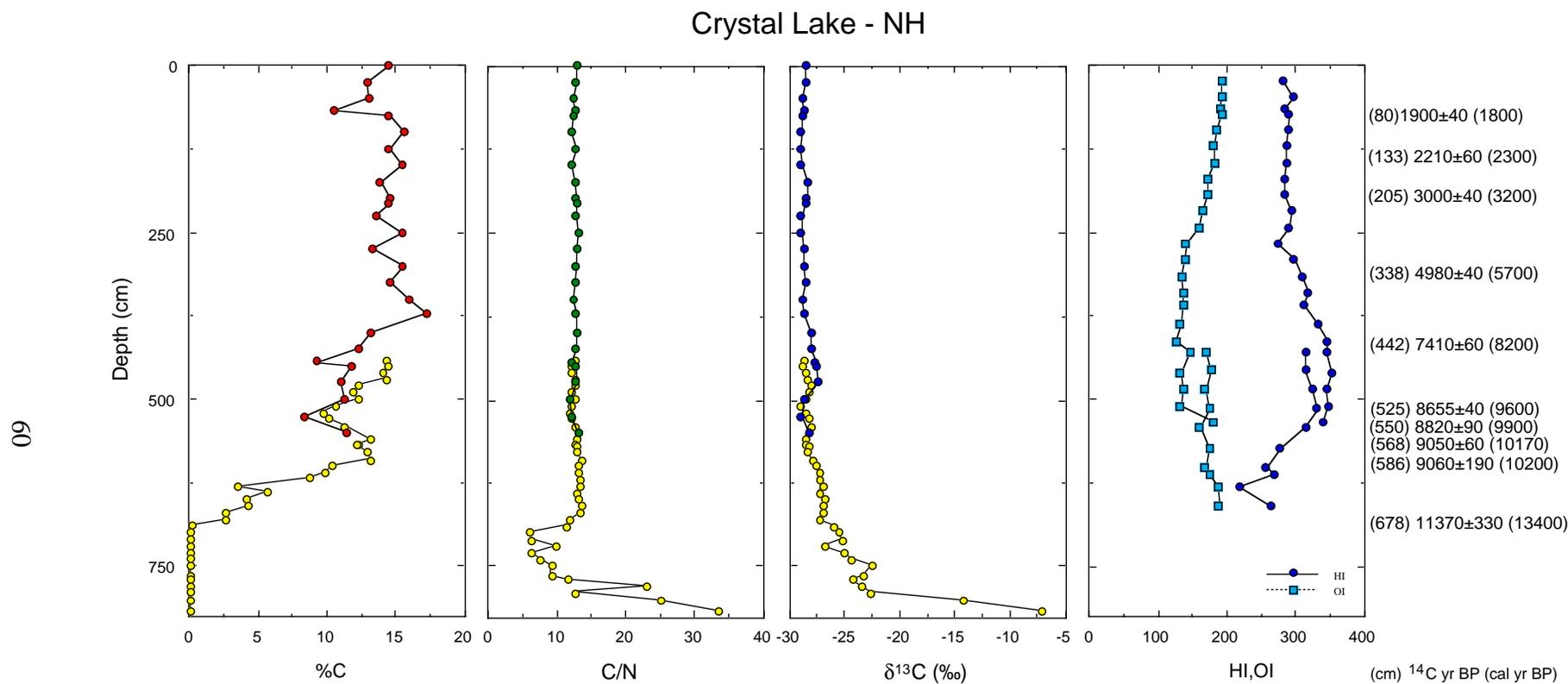
Figure 4.4. %C, C/N, δ¹³C, and Rock-Eval data for Crystal Lake, NH. Dates are presented as both ¹⁴C yrs BP and calibrated yrs BP. In order to obtain the entire early Holocene record, two cores were taken from the same location, resulting in overlap in the sediment records.
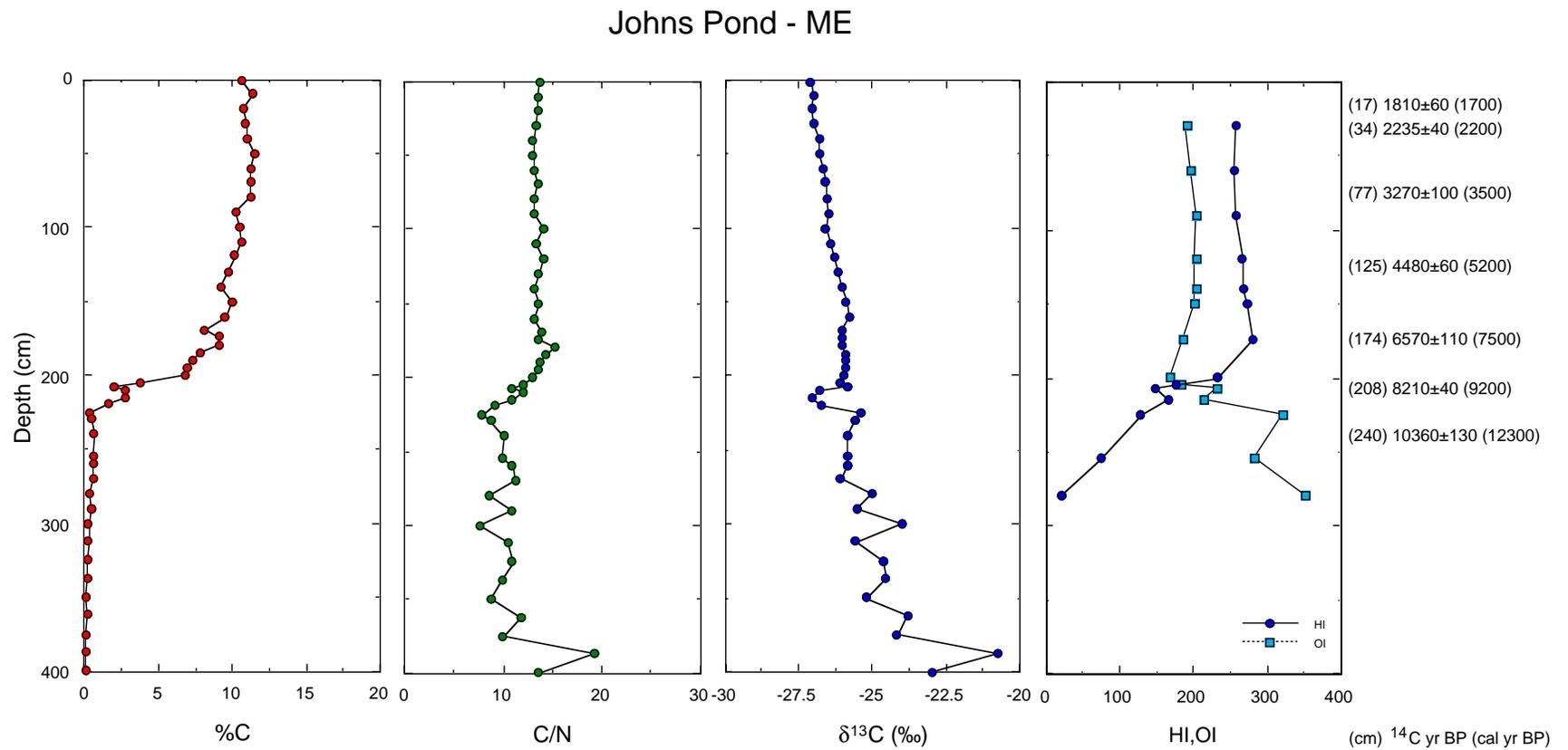
Figure 4.5. %C, C/N, δ¹³C, and Rock-Eval data for Johns Pond, ME. Dates are presented as both¹⁴C yrs BP and calibrated yrs BP.
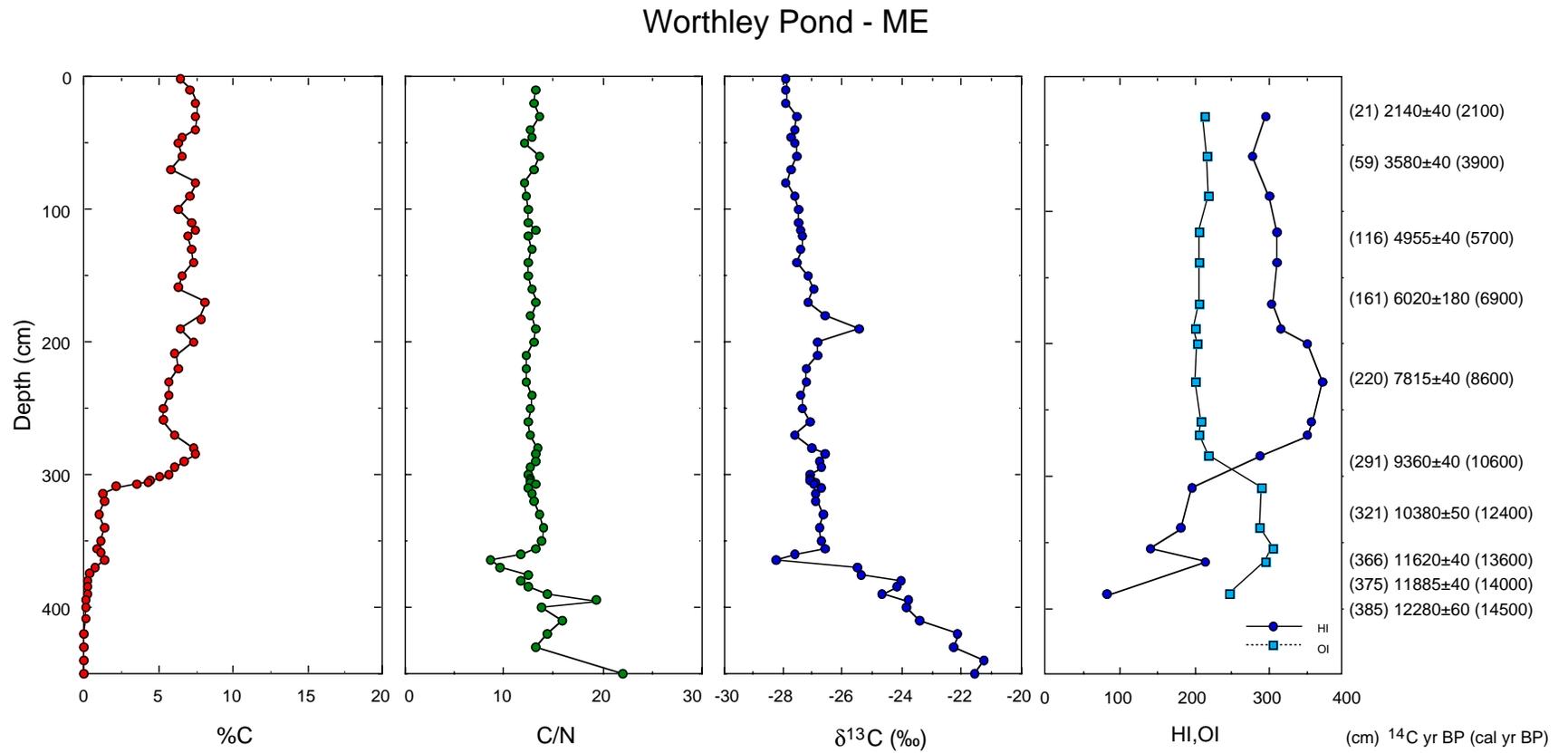
Figure 4.6. %C, C/N, δ¹³C, and Rock-Eval data for Worthley Pond, ME. Dates are presented as both ¹⁴C yrs BP and calibrated yrs BP.
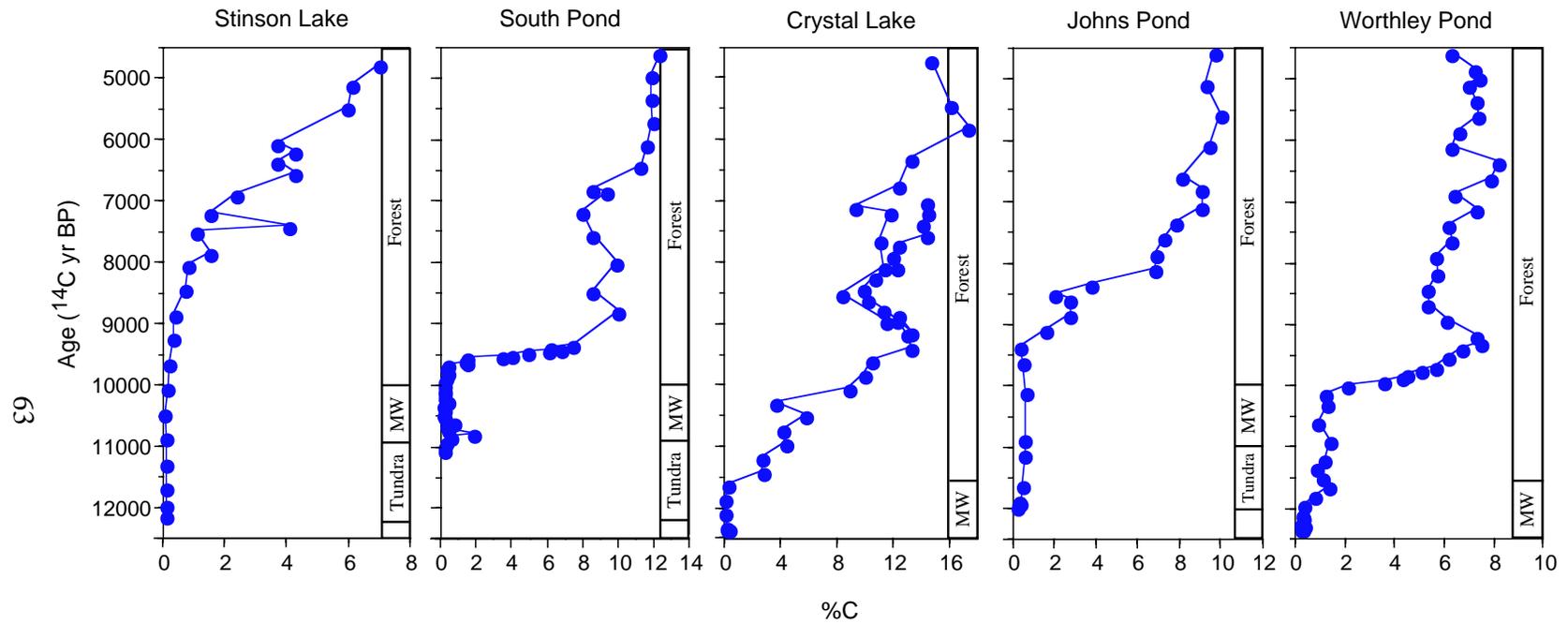
Figure 4.7. %C records, ^{14}C ages, and primary succession of vegetation stages. MW=mixed woodlands. Crystal Lake data shows an overlap of two core sections. Vegetation zones based on pollen data from Davis and Jacobson, Jr. (1985).
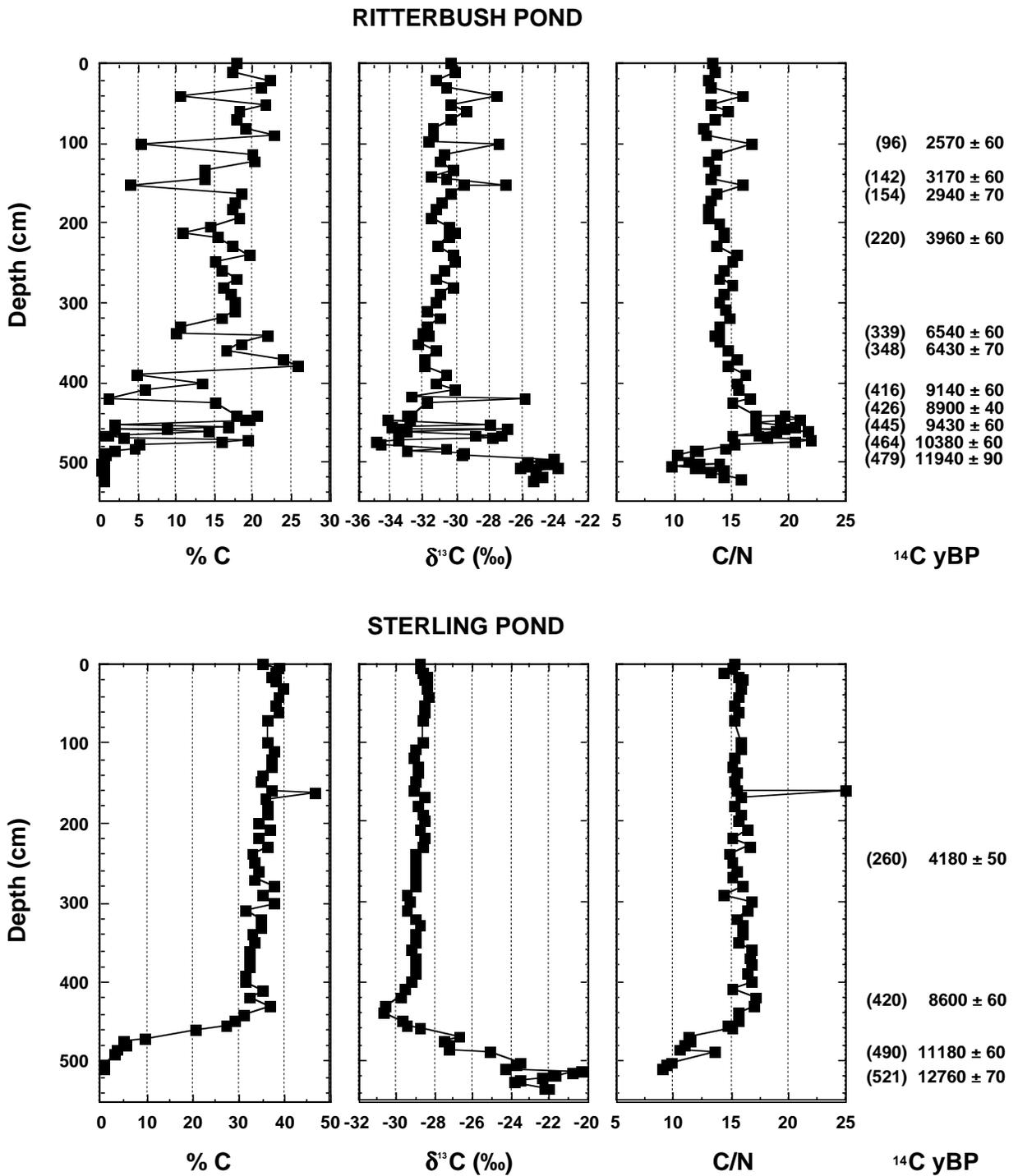
Figure 4.8. %C, δ¹³C, C/N, and ¹⁴C AMS dates for Ritterbush Pond and Sterling Pond, VT (modified after Lini. 1999).
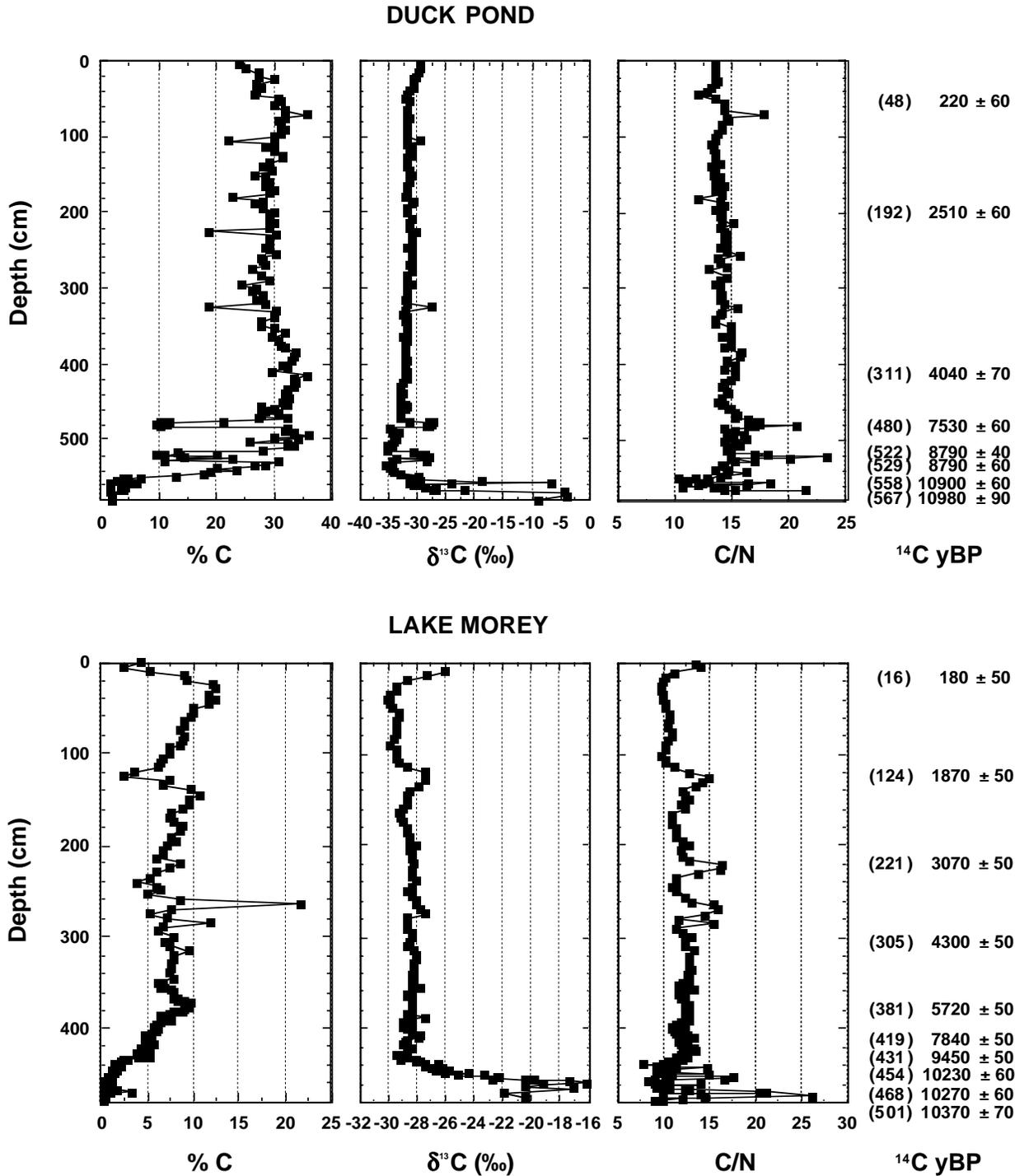
DUCK POND

LAKE MOREY

Figure 4.9. %C, δ^{13}C, C/N ratios, and ^{14}C AMS dates for Duck Pond and Lake Morey, VT (modified after Lini,1999).

Table 4.1. Physiographic characteristics of the studied lakes. Data for Vermont lakes from Galster (2001).

Lake Name	Surface Area (km²)	Maximum Depth (m)	Elevation (m)	Drainage Basin Area (km²)	Drainage Basin Relief (m)	Latitude	Longitude
New Hampshire							
Crystal Lake	0.38	18	146	15.0	353	N 43°54'	W 71°05'
Stinson Lake	1.40	22	396	20.7	655	N 43°52'	W 71°48'
South Pond	0.70	26	340	7.4	427	N 44°36'	W 71°22'
Maine							
Johns Pond	1.08	15	533	18.2	384	N 45°04'	W 70°46'
Worthley Pond	1.43	15	174	13.5	344	N 44°24'	W 70°26'
Vermont							
Duck Pond	0.03	14	520	0.7	290	N 44°42'	W 72°04'
Lake Morey	2.22	13	127	20.7	414	N 43°55'	W 72°09'
Ritterbush Pond	0.05	14	317	2.2	293	N 44°45'	W 72°36'
Sterling Pond	0.03	9	917	0.3	40	N 44°33'	W 72°47'

Table 4.2. Summary of the ^{14}C age determinations from the lake cores. Conversion to calibrated years is via Stuiver et al. (1998). All ^{14}C dates were measured from macrofossils.

Core	Depth (cm)	^{14}C Date	Calibrated Years
Stinson Lake	35	1640 ± 40	1500
Stinson Lake	70	2665 ± 35	2800
Stinson Lake	105	3570 ± 90	3900
Stinson Lake	151	4430 ± 40	5000
Stinson Lake	200	5890 ± 70	6700
Stinson Lake	230	7275 ± 40	8100
Stinson Lake	279	8160 ± 110	9200
Stinson Lake	408	12345 ± 45	14500
South Pond	26	3150 ± 190	3400
South Pond	50	3890 ± 70	4300
South Pond	76	4430 ± 110	5100
South Pond	114	6090 ± 70	6950
South Pond	153	7840 ± 60	8600
South Pond	186	8870 ± 40	10000
South Pond	243	10070 ± 40	11500
South Pond	284	10200 ± 40	11800
South Pond	370	10300 ± 80	12100
South Pond	388	10560 ± 60	12600
Crystal Lake	80	1900 ± 40	1800
Crystal Lake	133	2210 ± 60	2300
Crystal Lake	205	3000 ± 40	3200
Crystal Lake	388	4980 ± 40	5700
Crystal Lake	442	7410 ± 60	8200
Crystal Lake	525	8655 ± 40	9600
Crystal Lake	550	8820 ± 90	9900
Crystal Lake	568	9050 ± 60	10170
Crystal Lake	586	9060 ± 190	10200
Crystal Lake	678	11370 ± 330	13400
Johns Pond	17	1810 ± 60	1700
Johns Pond	34	2235 ± 40	2200
Johns Pond	77	3270 ± 100	3500
Johns Pond	125	4480 ± 60	5200
Johns Pond	174	6570 ± 110	7500
Johns Pond	208	8210 ± 40	9200
Johns Pond	240	10360 ± 130	12300
Worthley Pond	21	2140 ± 40	2100
Worthley Pond	59	3580 ± 40	3900
Worthley Pond	116	4955 ± 40	5700
Worthley Pond	161	6020 ± 180	6900
Worthley Pond	220	7815 ± 40	8600
Worthley Pond	291	9360 ± 40	10600
Worthley Pond	321	10380 ± 50	12400
Worthley Pond	366	11620 ± 40	13600
Worthley Pond	375	11885 ± 40	14000
Worthley Pond	385	12280 ± 60	14500

Table 4.3. Summary of the shifts in %C and δ^{13}C observed in the study lakes. Data for the Vermont lakes from Galster, 2001.

Lake Name	Start of %C Shift (^{14}C yr BP)	End of %C Shift (^{14}C yr BP)	Duration of %C Shift (^{14}C yr BP)	Start of δ^{13}C Shift (^{14}C yr BP)	End of δ^{13}C Shift (^{14}C yr BP)	Duration of δ^{13}C Shift (^{14}C yr BP)	Magnitude of δ^{13}C Shift ‰
New Hampshire							
Stinson Lake	10,000	top of core	N/A	12,400	top of core	N/A	4
South Pond	9,600	8,800	800	11,000	8,800	2,200	7
Crystal Lake	11,600	9,100	2,500	13,800	11,400	2,400	5
Maine							
Johns Pond	9,300	4,100	5,200	12,000	8,800	3,200	7
Worthley Pond	12,000	9,300	2,700	14,000	11,600	2,400	7
Vermont							
Lake Morey	10,200	5,600	4,600	10,200	9,500	700	13
Duck Pond	11,000	8,900	2,100	11,000	9,200	1,800	14
Sterling Pond	12,400	9,000	3,400	12,400	9,300	3,100	10
Ritterbush Pond	>12,000	11,300	>700	>12,000	11,600	>400	12

Table 4.4. $\delta^{13}C$, %C, %N, and C/N ratio data for modern plants from the study lakes.

Lake	Plant Type	$\delta^{13}C$	%C	%N	C/N
Stinson Lake	Macrophytes	-26.16	31.42	1.71	18.42
	Phytoplankton	-28.41			8.55
	Terrestrial Plants	-32.05	47.86	1.09	43.72
		-27.91	48.32	1.81	26.65
		-29.59	49.12	1.87	26.33
South Pond	Macrophytes	-27.50	43.18	3.67	11.77
		-24.45	43.87	3.34	13.13
	Phytoplankton	-29.29			9.24
	Terrestrial Plants	-30.35	49.61	1.91	26.01
		-29.53	48.97	1.10	44.57
Crystal Lake	Macrophytes	-28.64	37.94	1.35	28.11
		-26.00	44.01	4.08	10.78
		-28.00	43.65	2.14	20.43
		-26.35	42.11	2.26	18.63
	Phytoplankton	-28.54			9.61
	Terrestrial Plants	-25.78	47.79	3.99	11.98
Johns Pond	Macrophytes	-27.37	41.76	3.75	11.15
		-25.73	36.32	1.60	22.70
	Phytoplankton	-29.97			8.96
	Terrestrial Plants	-28.81	48.65	0.72	67.46
		-27.84	48.74	1.20	40.69
		-25.36	52.63	3.24	16.25
		-29.87	46.00	1.41	32.53
Worthley Pond	Macrophytes	-28.38	42.02	2.92	14.37
		-28.94	37.46	1.52	24.72
		-24.12	43.02	3.14	13.71
		-30.33	13.09	0.72	18.06
		-24.56	43.27	2.43	17.80
		-26.44	40.26	2.44	16.47
	Phytoplankton	-29.43			7.73
	Terrestrial Plants	-28.40	39.62	0.94	41.97
		-27.92	44.52	1.95	22.78
		-26.41	47.54	0.94	50.53
		-26.22	48.56	0.50	97.23
		-29.25	44.96	1.56	28.80
		-26.98	47.79	1.89	25.33
		-27.42	50.52	2.34	21.55
		-28.40	36.53	1.60	22.87
		-30.58	47.47	1.68	28.26

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Chapter 5. CONCLUSIONS

5.1 Summary

The purpose of this study was to compare the timing and rate of ecosystem establishment in newly formed post-glacial lakes in New Hampshire and Maine. Paramount to comparing ecosystem establishment rates is determining the parameters that influence the rate at which ecosystems develop. The following are the main conclusions developed during the course of this research:

1. The transition from inorganic to OM-rich sediments correlates with primary succession of vegetation in the watershed, either the onset of mixed woodlands or mature forest ecosystems. Inorganic sections of the core correlate with the tundra time periods. The OM records also track the warming at the end of the Younger Dryas in northern New England, at ~10,000 ^{14}C yrs BP, indicating a marked increase in productivity in both the terrestrial and aquatic environments.

2. The timing of ecosystem development in the five study lakes varies as a result of differences in elevation, proximity to the receding ice sheet, and contribution of terrigenous material.

3. Variations in the δ^{13}C and C/N records in the older sediments correlate with changes in sediment lithology and are the result of changing OM source (terrestrial vs. aquatic).

4. The documented negative trends in δ^{13}C values are primarily the result of changes in the source of OM, via both climate induced changes from C_4 and C_3 tundra grasses and shrubs to C_3 plants, and increased aquatic primary productivity.

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Processes such as increased DIC production in the watershed soils, increases in

atmospheric pCO_2 during the early Holocene, and diagenesis may also contribute

to the negative shifts in $\delta^{13}C$ values.

5. The studied New Hampshire and Maine lakes contain less sedimentary OM than

the Vermont lakes. This can be attributed to Stinson Lake, South Pond, and Johns

Pond being in closer proximity to the receding ice margin during the early

Holocene, and steeper drainage basins in New Hampshire and Maine. The

Vermont lakes display higher algal productivity than the New Hampshire and

Maine lakes, resulting in more negative $\delta^{13}C$ values in the Vermont lakes.

5.2 Suggestions for Future Research

Additional paleolimnological research is needed to gain further information on the

processes that affected lakes and their surrounding watersheds across northern New

England during the early Holocene. One such technique is analysis of the sediments for

biologic indicators. Specifically, diatom analysis in the lake sediments would provide

further details about the changing conditions within the lake waters, and help determine

how changes in water chemistry affect the %C, $\delta^{13}C$, C/N, and Rock-Eval records.

Furthermore, a detailed analysis of the macrofossils found in the sediments would reveal

what types of plants were growing in and around the lakes during the early Holocene.

The effect of bedrock geology should be investigated in more detail. Granite and

quartzite in New Hampshire, metasediments in Maine, and metamorphic schist in

Vermont may have had different effects on lake water chemistry, resulting from

variations in the composition of the overlying soils and vegetation. The difference in

weathering rates between quartzites, granites, and schists leads to inorganic sediment deposition of varying magnitude in the newly formed post-glacial lakes, as well as different levels of nutrient contribution to the lakes due to the distinct mineral content of the bedrock. These differences in lake water chemistry would influence the growth of the ecosystems in and surrounding the lakes.

Cores from additional locations within each of the study lakes should be obtained to observe what affect core location has on the geochemical records. In our five study lakes, the cores were taken in close proximity to the lake depocenter, and near areas where deltaic deposits were formed from inflowing streams. Expanding the study sites within each lake basin would aid in determining how the geochemical records are affected by their proximity to terrestrial OM sources.

Furthermore, the study area should continue to be expanded across New England. In particular, lakes located in mid to southern New England would provide data on areas with earlier documented deglaciation. Examining sediment records from lakes in eastern New York would allow for further comparison with the Vermont lakes. Finally, lakes with diverse physical characteristics should continue to be cored to determine in more detail how factors such as lake elevation, depth, surface area, and size and relief of the drainage basin affect the timing and rates of ecosystem establishment.

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Appendix A. Modern Plant and Phytoplankton Data

	Plant Name	δ¹³C	%C	%N	C/N
Stinson Lake	*Macrophytes*				
	spargium	-26.16	31.42	1.71	18.42
	Terrestrial Plants				
	paper birch	-32.05	47.86	1.09	43.72
	red maple	-27.91	48.32	1.81	26.65
	aspen	-29.59	49.12	1.87	26.33
	Phytoplankton	-28.41			8.55
South Pond	*Macrophytes*				
	spargium	-27.50	43.18	3.67	11.77
	water lily	-24.45	43.87	3.34	13.13
	Terrestrial Plants				
	mountain holly	-30.35	49.61	1.91	26.01
	hemlock	-29.53	48.97	1.10	44.57
	Phytoplankton	-29.29			9.24
Crystal Lake	*Macrophytes*				
	eleocharis	-28.64	37.94	1.35	28.11
	water lily	-26.00	44.01	4.08	10.78
	pickeral weed	-28.00	43.65	2.14	20.43
	potamogeton	-26.35	42.11	2.26	18.63
	Terrestrial Plants				
	button bush	-25.78	47.79	3.99	11.98
	Phytoplankton	-28.54			9.61
Johns Pond	detritus	-26.87	17.09	0.46	36.91
	Macrophytes				
	water lily	-27.37	41.76	3.75	11.15
	horsetail	-25.73	36.32	1.60	22.70
	Terrestrial Plants				
	spruce	-28.81	48.65	0.72	67.46
	fir	-27.84	48.74	1.20	40.69
	sugar maple	-25.36	52.63	3.24	16.25
	birch	-29.87	46.00	1.41	32.53
	Phytoplankton	-29.97			8.96

	Plant Name	$\delta^{13}C$	%C	%N	C/N
Worthley Pond	detritus	-28.67	42.61	0.89	47.66
	Macrophytes				
	pickeral weed	-28.38	42.02	2.92	14.37
	spike rush	-28.94	37.46	1.52	24.72
	water lily	-24.12	43.02	3.14	13.71
	quillwort	-30.33	13.09	0.72	18.06
	water lily	-24.56	43.27	2.43	17.80
	burreed	-26.44	40.26	2.44	16.47
	Terrestrial Plants				
	white pine	-28.40	39.62	0.94	41.97
	birch	-27.92	44.52	1.95	22.78
	cedar	-26.41	47.54	0.94	50.53
	hemlock	-26.22	48.56	0.50	97.23
	scarlet pine	-29.25	44.96	1.56	28.80
	red maple	-26.98	47.79	1.89	25.33
	paper birch	-27.42	50.52	2.34	21.55
	fern	-28.40	36.53	1.60	22.87
	white birch	-30.58	47.47	1.68	28.26
	Phytoplankton	-29.43			7.73

Appendix B. Age Models

B.1 Calculated ¹⁴C and calibrated ages

Stinson Lake			South Pond			Crystal Lake		
Depth (cm)	¹⁴C yrs BP	cal yrs BP	Depth (cm)	¹⁴C yrs BP	cal yrs BP	Depth (cm)	¹⁴C yrs BP	cal yrs BP
1	931	630	1	2018	2403	1	719	939
10	1147	899	10	2351	2757	25	1013	1193
20	1387	1198	20	2721	3150	50	1319	1459
30	1627	1497	27	2980	3425	67	1528	1639
40	1867	1796	30	3091	3543	75	1626	1724
50	2107	2095	40	3461	3936	100	1932	1989
60	2347	2394	50	3831	4329	125	2238	2255
70	2587	2693	60	4201	4722	150	2545	2520
80	2827	2992	70	4571	5115	175	2851	2785
90	3067	3291	80	4941	5507	200	3157	3051
100	3307	3590	90	5311	5900	206	3231	3222
110	3547	3889	100	5681	6293	225	3464	3595
113	3619	3979	110	6051	6686	250	3770	4087
114	3643	4009	120	6421	7079	275	4076	4579
130	4027	4487	130	6791	7472	300	4383	5070
132	4075	4547	131	6828	7512	325	4689	5562
149	4453	4881	140	7161	7865	350	5418	6054
159	4753	5268	150	7531	8258	371	5788	6467
170	5083	5694	162	7975	8730	400	6299	7037
183	5473	6197	175	8456	9241	425	6739	7529
202	6043	6932	184	8789	9594	444	7073	7903
207	6193	7126	190	9310	10521	450	7179	8021
212	6343	7319	200	9376	10625	475	7619	8512
218	6523	7552	202	9389	10646	500	8059	9004
230	6883	7783	205	9409	10677	525	8499	9496
240	7183	8154	210	9442	10729	550	8939	9987
247	7393	8414	215	9475	10782	2nd Core		
250	7483	8525	220	9509	10834	441	7020	7844
262	7843	8970	225	9542	10886	450	7179	8021
275	8019	9453	230	9575	10938	460	7355	8217
287	8408	9898	235	9608	10990	470	7531	8414
300	8829	10380	240	9641	11042	480	7707	8611
312	9218	10825	250	9707	11147	490	7883	8807
325	9639	11308	260	9773	11251	500	8059	9004
337	10028	11753	270	9840	11356	510	8235	9201
350	10449	12235	280	9906	11460	520	8411	9397
362	10838	12680	290	9972	11564	530	8587	9594
375	11259	13163	300	10038	11669	540	8763	9869
387	11648	13608	310	10105	11773	560	9115	10024
400	12069	14090	320	10171	11877	567	8849	10079
408	12328	14387	330	10237	11982	570	8916	10102
			340	10303	12086	580	9140	10180
			350	10370	12190	590	9364	10308
			363	10456	12326	600	9588	10654
			370	10502	12399	610	9812	11000
			382	10581	12524	620	10036	11345

Stinson Lake			South Pond			Crystal Lake		
Depth (cm)	¹⁴C yrs BP	cal yrs BP	Depth (cm)	¹⁴C yrs BP	cal yrs BP	Depth (cm)	¹⁴C yrs BP	cal yrs BP
			390	10634	12608	630	10260	11691
						640	10484	12037
						650	10708	12382
						660	10932	12728
						670	11156	13073
						680	11380	13419

B.1 Calculated ¹⁴C and calibrated ages (continued)

Johns Pond			Worthley Pond		
Depth (cm)	¹⁴C yrs BP	cal yrs BP	Depth (cm)	¹⁴C yrs BP	cal yrs BP
1	1441	1137	1	1378	1527
10	1657	1427	10	1719	1858
20	1898	1748	20	2098	2226
30	2139	2069	30	2477	2594
40	2379	2391	40	2856	2962
50	2620	2712	46	3083	3183
60	2861	3033	50	3235	3330
70	3102	3355	60	3606	3698
80	3342	3676	70	3847	4066
90	3583	3998	80	4088	4434
100	3824	4319	90	4329	4802
110	4064	4640	100	4570	5170
120	4305	4903	110	4811	5538
130	4547	5386	116	4960	5650
140	5051	5868	120	5062	5762
150	5554	6350	130	5318	6043
160	6058	6833	140	5574	6324
170	6562	7315	150	5830	6605
174	6763	7508	160	6086	6886
180	7065	7797	170	6342	7167
185	7317	8038	180	6598	7448
190	7569	8280	190	6854	7729
195	7820	8521	200	7110	8010
200	8072	8762	210	7366	8291
205	8324	9003	220	7622	8572
208	8475	9170	230	7878	8853
210	8576	9363	240	8134	9134
215	8828	9844	250	8390	9415
220	9079	10325	260	8646	9696
225	9331	10806	270	8902	9977
230	9583	11288	280	9158	10258
240	10086	12250	285	9286	10399
			290	9359	10761
			295	9510	10954
			300	9660	11146
			302	9720	11223
			304	9780	11300
			306	9841	11377
			308	9901	11454
			310	9961	11531
			315	10112	11724
			320	10262	11916
			330	10563	12301
			340	10864	12686
			350	11165	13071
			355	11316	13264
			360	11466	13456
			365	11617	13649
			370	11767	13841
			375	11918	14034
			385	12219	14419

B.2 Measured ¹⁴C and calibrated dates vs. depth

Stinson Lake

South Pond

Crystal Lake

Johns Pond
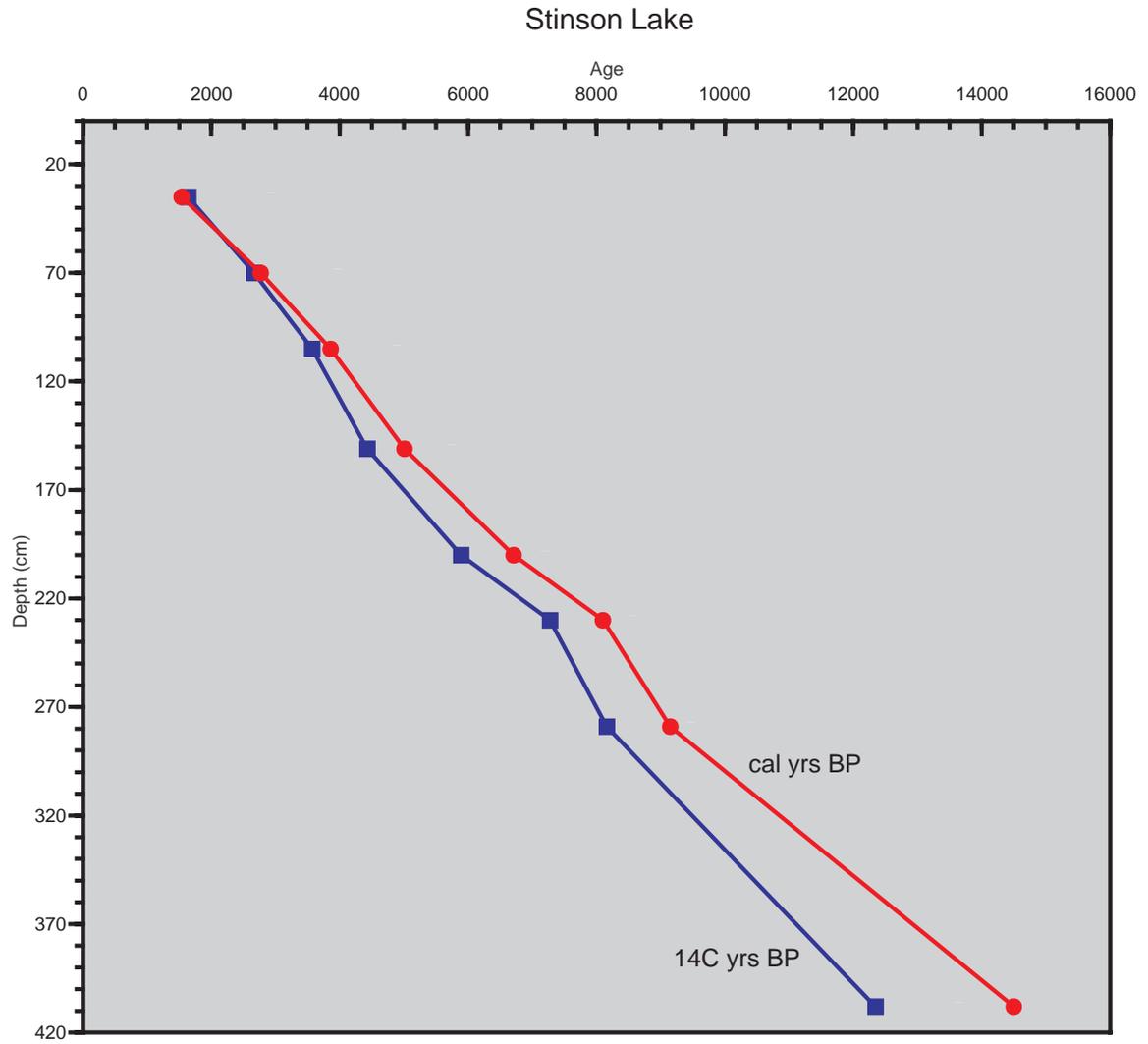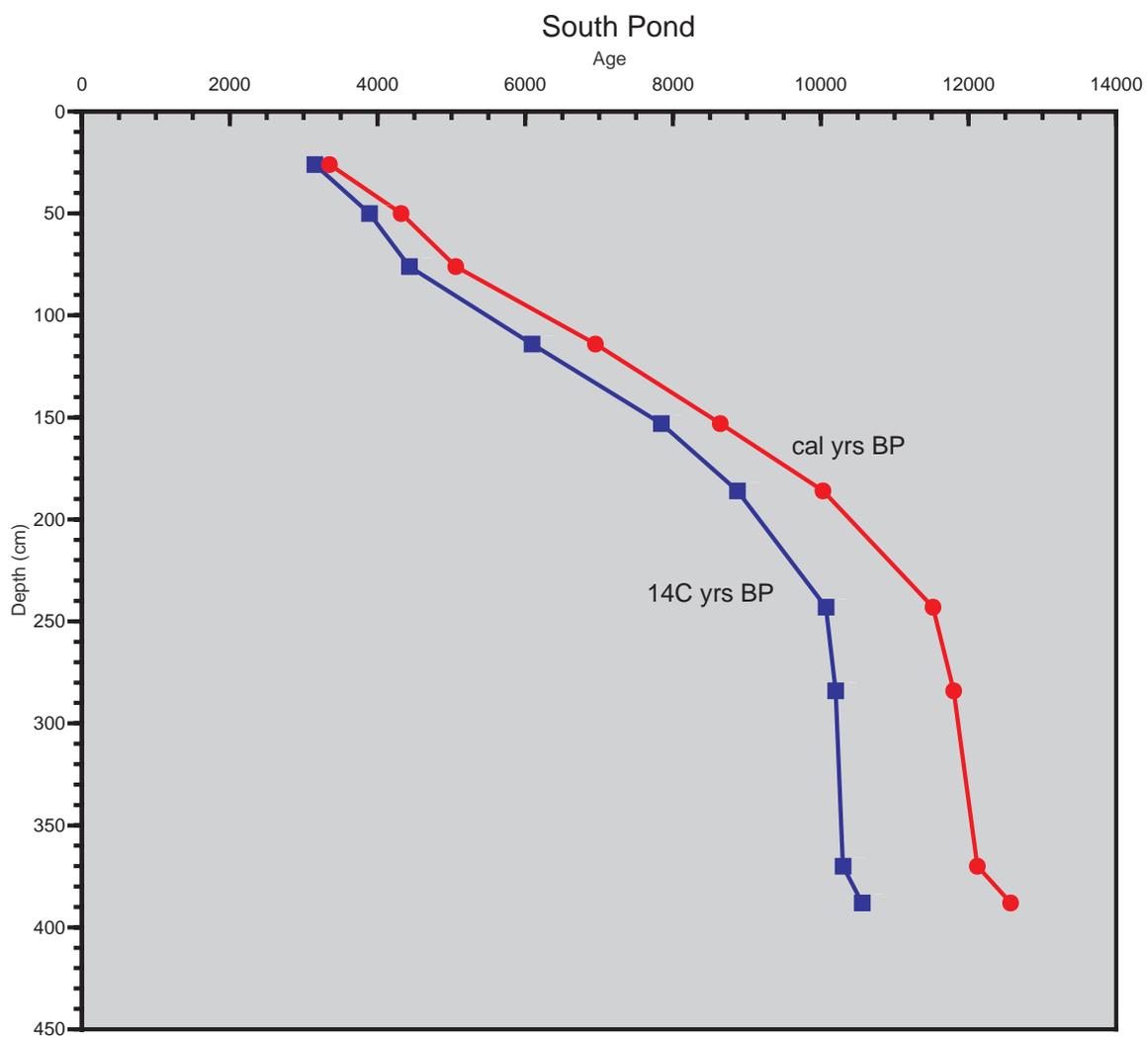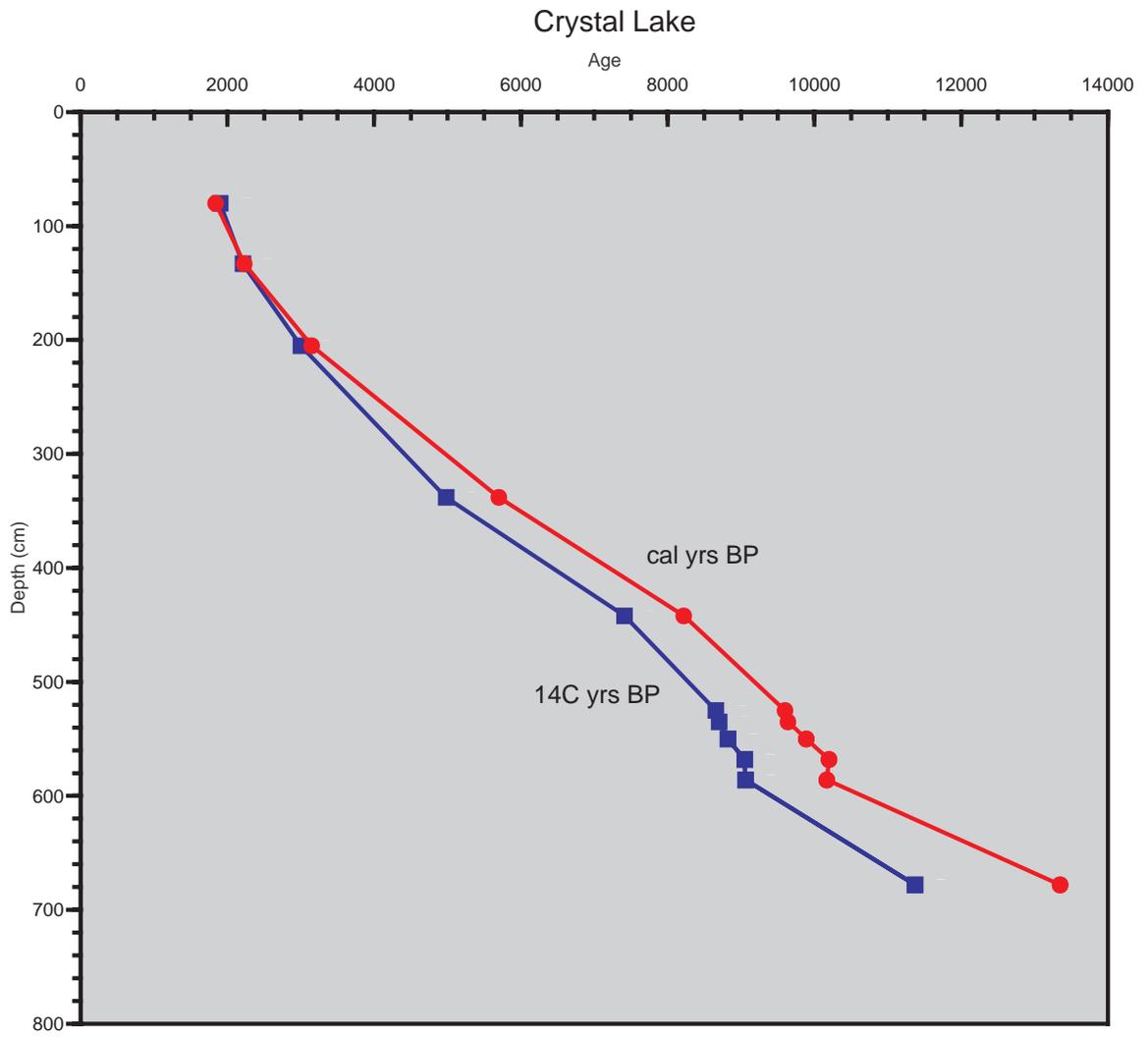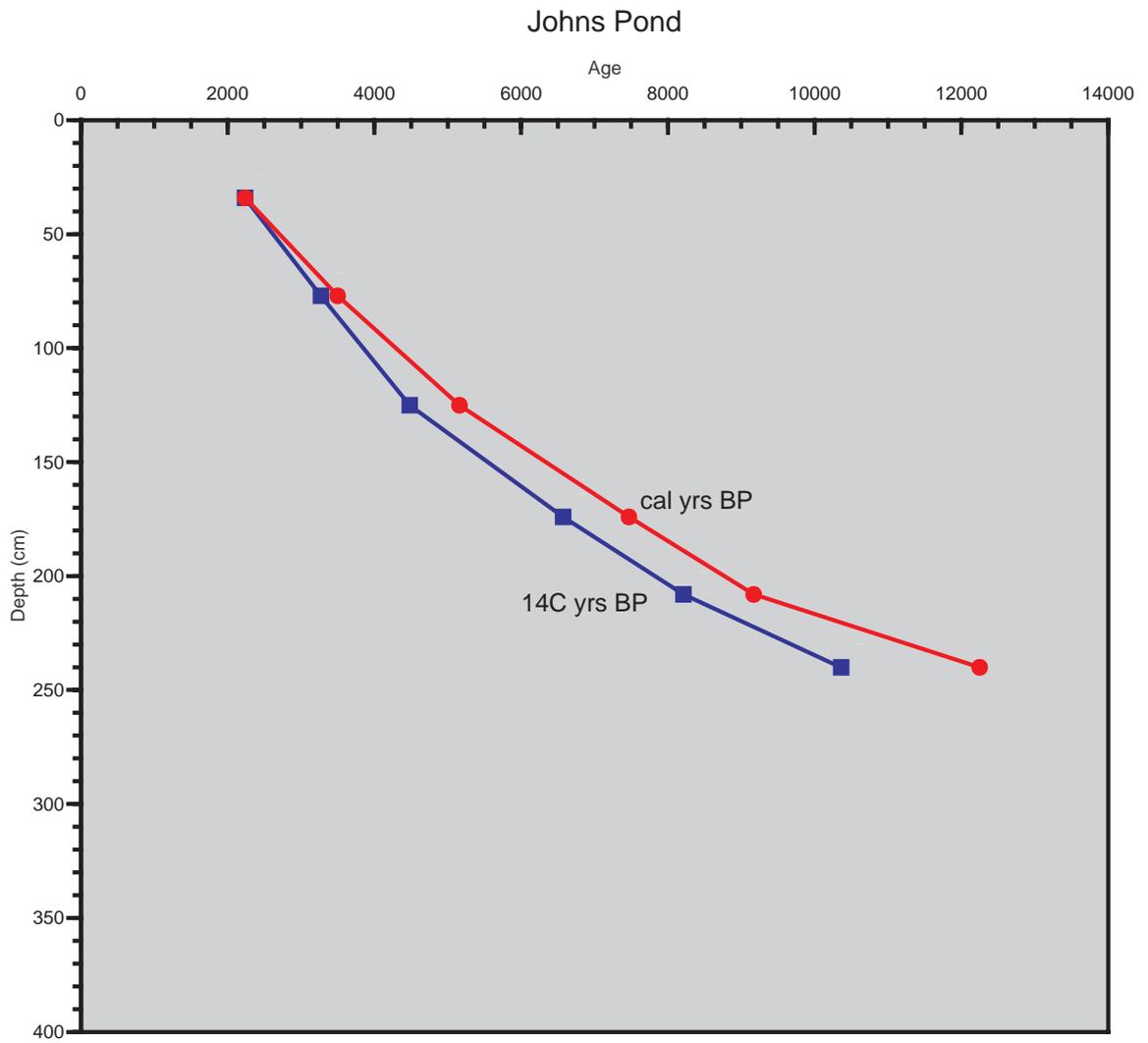
Worthley Pond
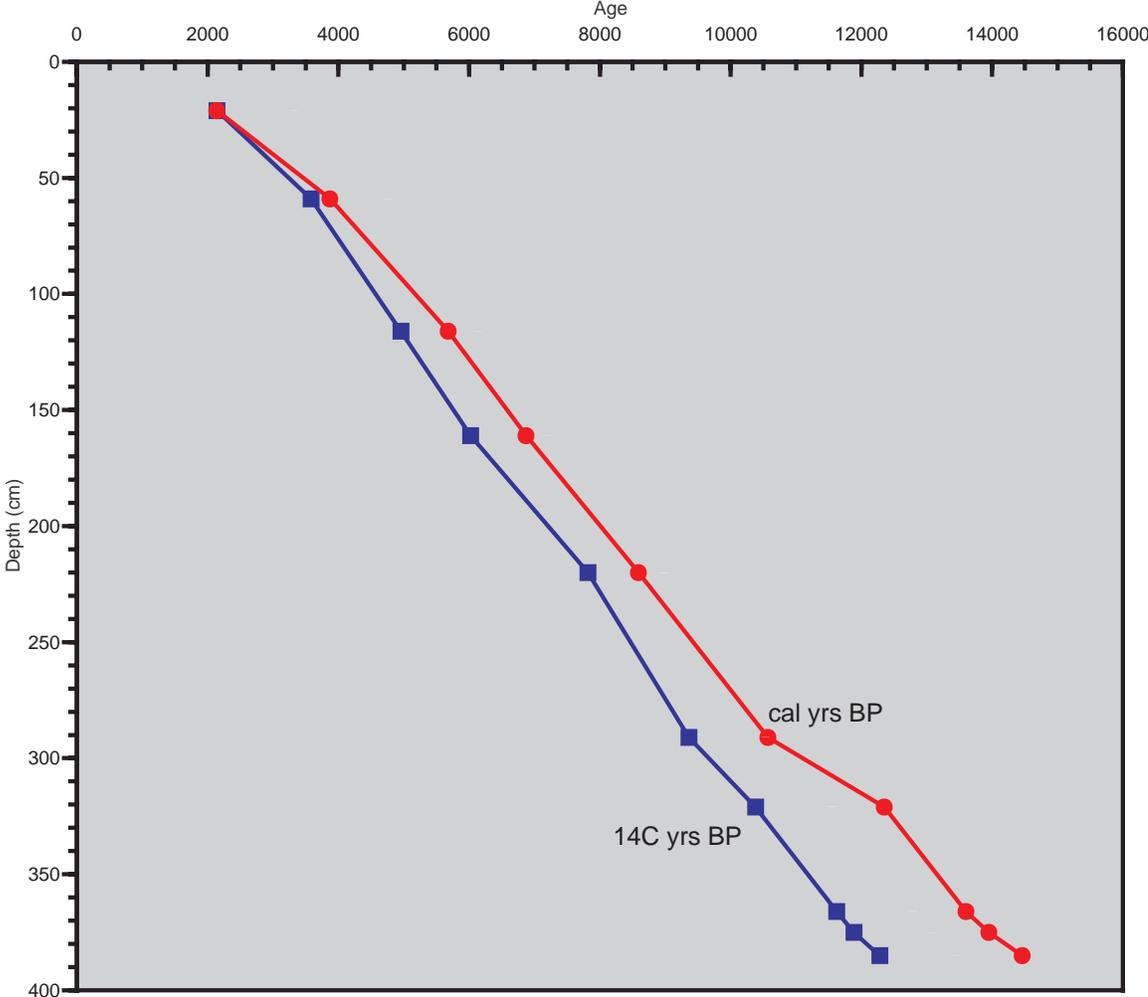
Appendix C. Laboratory Data

C.1 Stinson Lake, NH

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
1	-27.05	10.58	0.76	13.95		
10	-26.88	9.77	0.72	13.64		
20	-26.91	8.50	0.63	13.59		
30	-27.01	8.41	0.59	14.14	241	266
40	-26.89	8.59	0.61	14.19		
50	-26.72	9.12	0.65	14.07		
60	-26.68	8.86	0.65	13.55	251	265
70	-26.78	7.44	0.53	14.02		
80	-26.67	6.78	0.49	13.79		
90	-26.80	7.57	0.54	13.95	250	256
100	-26.80	5.87	0.42	14.11		
110	-26.63	4.35	0.32	13.50	243	263
113	-26.52	6.32	0.48	13.10		
114	-26.45	6.86	0.52	13.20	249	269
130	-26.52	6.25	0.45	13.80		
132	-26.38	5.58	0.42	13.30		
149	-26.23	5.31	0.40	13.40	242	270
159	-25.82	7.00	0.53	13.30	251	263
170	-25.74	6.10	0.46	13.40	245	261
183	-25.85	5.97	0.44	13.60		
202	-25.26	3.68	0.29	12.50	209	240
207	-25.50	4.27	0.34	12.50		
212	-25.29	3.69	0.29	12.90		
218	-25.33	4.24	0.33	12.80		
230	-25.07	2.33	0.18	13.20	187	
240	-24.82	1.49	0.12	12.70		
247	-24.91	4.05	0.31	13.20	232	280
250	-24.83	1.05	0.08	13.10		
262	-25.58	1.49	0.11	13.60	126	240
275	-25.22	0.79	0.07	10.70		
287	-24.94	0.69	0.06	12.00	63	364
300	-24.90	0.33	0.03	10.00		
312	-24.70	0.30	0.02	12.50	10	377
325	-23.67	0.15	0.02	8.20		
337	-23.18	0.13	0.01	14.30		
350	-22.51	0.03	0.00	15.10		
362	-24.33	0.08	0.01	16.50		
375	-23.56	0.08	0.01	8.70		
387	-23.05	0.07	0.01	11.80		
400	-23.17	0.07	0.01	8.10		
408	-22.82	0.04	0.00	12.50		

C.2 South Pond, NH

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
1	-26.98	11.53	0.83	13.92		
10	-27.07	12.93	0.99	13.07	214	215
20	-27.03	13.86	1.05	13.26		
27	-26.85	8.91	0.65	13.66	183	235
30	-27.09	13.26	1.02	13.04		
40	-27.10	13.01	1.00	13.06		
50	-27.24	11.99	0.83	14.51	175	253
60	-26.83	10.14	0.75	13.60	182	246
70	-26.78	12.32	0.92	13.33		
80	-26.86	11.82	0.85	13.90	252	219
90	-26.73	11.82	0.85	13.90		
100	-26.85	11.92	0.88	13.60		
110	-26.88	11.58	0.84	13.75	258	199
120	-26.89	11.25	0.81	13.92		
130	-27.08	8.55	0.61	14.06		
131	-27.05	9.37	0.65	14.51		
140	-27.46	7.96	0.55	14.39	154	294
150	-27.55	8.51	0.62	13.80		
162	-27.49	9.90	0.68	14.50	250	218
175	-27.48	8.52	0.63	13.60		
184	-27.58	9.96	0.70	14.20	254	222
190	-27.02	7.43	0.58	12.90	216	262
200	-26.99	6.20	0.50	12.40		
202	-26.34	6.78	0.53	12.70		
205	-26.90	6.10	0.50	12.70		
210	-26.73	4.90	0.40	12.60		
215	-26.96	4.00	0.30	12.30	135	420
220	-26.72	3.50	0.30	11.50		
225	-26.64	1.50	0.10	13.50		
230	-26.83	1.40	0.10	14.90		
235	-26.87	1.50	0.10	15.10		
240	-26.29	0.40	0.03	15.60	180	68
250	-26.22	0.30	0.02	15.20		
260	-26.05	0.40	0.03	14.20		
270	-25.92	0.30	0.02	13.50	101	238
280	-25.99	0.26	0.02	15.90		
290	-25.84	0.24	0.01	19.50		
300	-26.39	0.26	0.01	23.10	56	272
310	-25.55	0.27	0.01	18.40		
320	-25.59	0.23	0.02	14.30		
330	-25.89	0.46	0.03	17.00	41	238
340	-25.79	0.19	0.01	16.10		

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
350	-25.42	0.25	0.02	13.60		
363	-23.99	0.11	0.01	21.20		
370	-25.44	0.27	0.01	19.00		
382	-26.50	0.79	0.04	21.10	78	201
390	-25.58	0.30	0.02	14.70	22	289
400	-25.20	0.46	0.03	13.70		
410	-25.98	1.85	0.19	9.90	185	228
420	-25.40	0.57	0.07	8.00		
430	-23.95	0.34	0.04	8.30	28	395
440	-23.43	0.27	0.03	8.50		
450	-21.65	0.24	0.03	8.00		

C.3 Crystal Lake, NH

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
Crystal	(core 1)					
1	-28.62	14.54	1.11	13.10		
25	-28.62	12.97	1.00	12.90	282	193
50	-28.78	13.10	1.05	12.50	296	193
67	-28.75	10.60	0.84	12.60	285	191
75	-28.85	14.53	1.17	12.50	291	193
100	-28.91	15.69	1.27	12.30	289	186
125	-29.01	14.55	1.13	12.90	288	182
150	-28.97	15.54	1.26	12.30	288	184
175	-28.31	13.94	1.10	12.70	284	173
200	-28.43	14.65	1.14	12.90	286	172
206	-28.64	14.58	1.12	13.10		
225	-28.91	13.63	1.07	12.80	294	165
250	-28.98	15.51	1.18	13.20	292	160
275	-28.70	13.41	1.02	13.10	275	141
300	-28.73	15.55	1.22	12.80	299	140
325	-28.65	14.70	1.16	12.70	310	136
350	-28.87	16.06	1.30	12.40	318	137
371	-28.76	17.39	1.36	12.80	314	138
400	-27.99	13.25	1.01	13.10	334	134
425	-28.06	12.37	0.97	12.80	347	128
444	-27.64	9.28	0.76	12.20	344	147
450	-27.56	11.83	0.92	12.90		
475	-27.47	11.05	0.86	12.80	355	134
500	-28.68	11.34	0.95	11.90	345	137
525	-28.92	8.35	0.69	12.20	348	133
550	-28.21	11.53	0.87	13.20	342	181
Crystal	(core 2)					
441	-28.76	14.43	1.12	12.90	316	171
450	-28.86	14.50	1.19	12.20		
460	-28.56	14.10	1.15	12.20		
470	-28.35	14.38	1.13	12.80	316	178
480	-28.11	12.34	0.98	12.60		
490	-28.28	11.98	0.98	12.30		
500	-28.59	12.32	0.97	12.70	324	168
510	-28.94	10.69	0.87	12.30		
520	-28.58	9.89	0.82	12.00		
530	-28.25	10.22	0.84	12.20	331	176
540	-28.06	11.33	0.88	12.80		
560	-28.54	13.30	1.02	13.00	315	162
567	-28.49	12.34	0.96	12.90		
570	-28.25	12.29	0.95	13.00		

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
580	-28.34	12.96	0.99	13.10		
590	-27.87	13.28	0.97	13.70		
600	-27.56	10.47	0.80	13.20		
610	-27.32	9.95	0.75	13.30		
620	-27.29	8.86	0.66	13.50	258	169
630	-26.93	3.68	0.27	13.60	268	176
640	-27.31	5.80	0.44	13.10		
650	-26.89	4.20	0.32	13.30	219	188
660	-26.93	4.36	0.32	13.70		
670	-26.98	2.73	0.20	13.50		
680	-27.34	2.76	0.23	12.00	263	190
690	-25.99	0.31	0.03	11.50		
700	-25.55	0.13	0.02	6.10		
710	-25.25	0.11	0.02	6.40		
720	-26.83	0.20	0.02	10.00		
730	-25.06	0.11	0.02	6.30		
740	-24.42	0.12	0.02	7.70		
750	-22.42	0.11	0.01	9.40		
765	-23.31	0.16	0.02	9.30		
770	-24.36	0.24	0.02	11.80		
780	-23.40	0.19	0.01	23.20		
790	-22.65	0.12	0.01	12.60		
800	-14.23	0.20	0.01	25.20		
815	-7.08	0.22	0.01	33.60		

C.4 Johns Pond, ME

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
1	-27.09	10.60	0.77	13.79		
10	-27.01	11.37	0.84	13.47		
20	-27.02	10.77	0.80	13.48		
30	-27.00	10.89	0.81	13.36	259	192
40	-26.81	11.05	0.85	13.04		
50	-26.82	11.51	0.88	13.04		
60	-26.68	11.27	0.85	13.20	257	199
70	-26.58	11.21	0.83	13.46		
80	-26.55	11.15	0.84	13.21		
90	-26.50	10.27	0.78	13.23	259	206
100	-26.58	10.49	0.74	14.10		
110	-26.45	10.64	0.80	13.30		
120	-26.33	10.15	0.73	14.00	267	205
130	-26.16	9.73	0.72	13.60		
140	-26.04	9.29	0.70	13.20	270	205
150	-25.88	10.02	0.74	13.60	274	202
160	-25.79	9.48	0.72	13.20		
170	-26.02	8.16	0.59	13.90		
174	-26.05	9.09	0.67	13.60	282	187
180	-26.02	9.10	0.59	15.30		
185	-25.95	7.89	0.55	14.30		
190	-25.90	7.30	0.53	13.80		
195	-25.91	6.90	0.51	13.40		
200	-26.00	6.88	0.53	13.00	235	170
205	-26.12	3.80	0.32	11.90	178	184
208	-25.83	2.07	0.19	11.00	149	234
210	-26.77	2.77	0.23	12.00		
215	-27.02	2.78	0.25	10.90	168	215
220	-26.73	1.64	0.18	9.20		
225	-25.40	0.38	0.05	7.90	130	324
230	-25.60	0.49	0.06	8.80		
240	-25.85	0.63	0.06	10.10		
255	-25.85	0.59	0.06	9.80	77	285
260	-25.86	0.59	0.05	10.90		
270	-26.05	0.50	0.04	11.30		
280	-25.01	0.40	0.05	8.60	23	352
290	-25.51	0.48	0.04	10.80		
300	-24.02	0.28	0.04	7.60		
312	-25.54	0.25	0.02	10.40		
325	-24.62	0.15	0.01	10.90		
337	-24.54	0.14	0.01	9.90	218	189
350	-25.22	0.12	0.01	8.80		

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
362	-23.80	0.17	0.01	11.80		
375	-24.19	0.09	0.01	10.00		
387	-20.77	0.04	0.00	19.30		
400	-22.96	0.05	0.00	13.50		

C.5 Worthley Pond, ME

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
1	-27.90	6.45	0.46	14.00		
10	-27.88	7.12	0.54	13.20		
20	-27.90	7.54	0.58	13.10		
30	-27.54	7.55	0.56	13.60	295	214
40	-27.61	7.40	0.58	12.80		
46	-27.71	6.65	0.51	13.00		
50	-27.56	6.32	0.52	12.10		
60	-27.52	6.64	0.49	13.60	279	218
70	-27.70	5.76	0.44	13.10		
80	-27.88	7.51	0.62	12.10		
90	-27.64	7.08	0.57	12.30	301	220
100	-27.48	6.31	0.50	12.50		
110	-27.48	7.22	0.57	12.60		
116	-27.44	7.42	0.56	13.30	312	208
120	-27.33	6.99	0.56	12.50		
130	-27.39	7.28	0.56	13.00		
140	-27.54	7.36	0.59	12.50	310	208
150	-27.15	6.58	0.52	12.60		
160	-26.97	6.32	0.49	12.90		
170	-27.16	8.21	0.61	13.40	304	207
180	-26.61	7.86	0.62	12.80		
190	-25.45	6.43	0.48	13.30	318	201
200	-26.85	7.31	0.56	13.10	351	204
210	-26.85	6.19	0.50	12.30		
220	-27.21	6.29	0.51	12.30		
230	-27.21	5.66	0.46	12.30	374	203
240	-27.38	5.72	0.44	13.00		
250	-27.33	5.35	0.42	12.80		
260	-27.12	5.33	0.43	12.50	358	210
270	-27.63	6.09	0.48	12.70	351	207
280	-27.04	7.31	0.54	13.50		
285	-26.56	7.51	0.56	13.40	289	220
290	-26.77	6.71	0.51	13.30		
295	-26.74	6.16	0.48	12.70		
300	-27.12	5.64	0.45	12.60		
302	-27.06	5.12	0.40	12.70		
304	-27.06	4.51	0.35	12.80		
306	-26.86	4.32	0.34	12.70		
308	-26.99	3.57	0.27	13.20		
310	-26.74	2.13	0.17	12.60	197	292
315	-26.92	1.25	0.10	12.90		
320	-26.88	1.30	0.10	13.10		

Depth (cm)	$\delta^{13}C$	%C	%N	C/N	HI	OI
330	-26.64	0.92	0.07	13.60		
340	-26.80	1.41	0.10	14.10		
350	-26.76	1.17	0.08	14.00		
355	-26.56	0.85	0.06	13.20	141	307
360	-27.57	1.12	0.09	11.90		
365	-28.29	1.38	0.16	8.60	216	295
370	-25.55	0.78	0.08	9.80		
375	-25.35	0.38	0.03	12.50		
380	-24.05	0.29	0.02	11.90		
385	-24.13	0.24	0.02	12.60		
390	-24.65	0.24	0.02	14.50	83	249
395	-23.82	0.12	0.01	19.50		
400	-23.87	0.11	0.01	13.80		
410	-23.40	0.10	0.01	15.90		
420	-22.12	0.06	0.00	14.40		
430	-22.25	0.06	0.00	13.20		
440	-21.26	0.02	0.00	-		
450	-21.60	0.03	0.00	22.00		
370	-25.55	0.78	0.08	9.80		
380	-24.05	0.29	0.02	11.90		
390	-24.65	0.24	0.02	14.50		
400	-23.87	0.11	0.01	13.80		
410	-23.40	0.10	0.01	15.90		
420	-22.12	0.06	0.00	14.40		
430	-22.25	0.06	0.00	13.20		
440	-21.26	0.02	0.00	-		
450	-21.60	0.03	0.00	22.00		

Appendix D. Topographic Maps, Aerial Photographs[1], and Bathymetric Maps of the Study Lakes

D.1 Stinson Lake

Stinson Lake
Rumney, NH

N

0 ⊢————————⊣ 1Km

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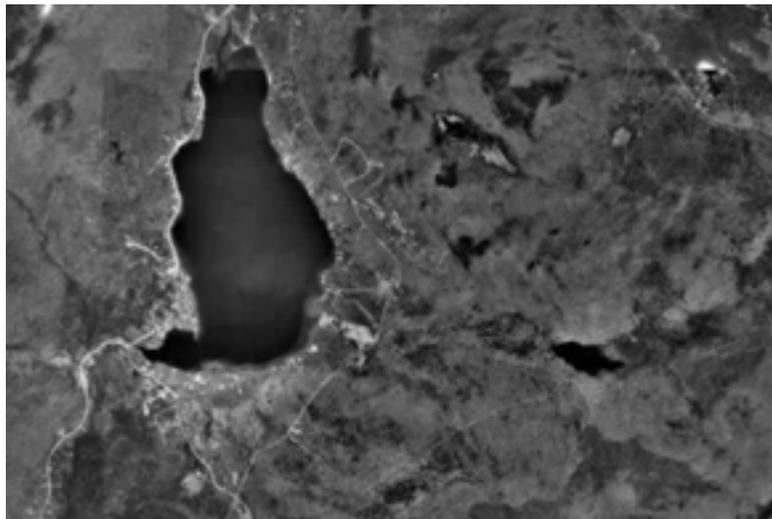
[1] Topographic maps and aerial photographs obtained from the United States Geological Survey via the Microsoft® TerraServer website (www.terraserver.homeadvisor.msn.com)

Stinson Lake

Core Location

Contour Interval
20 feet

0   meters   500

N
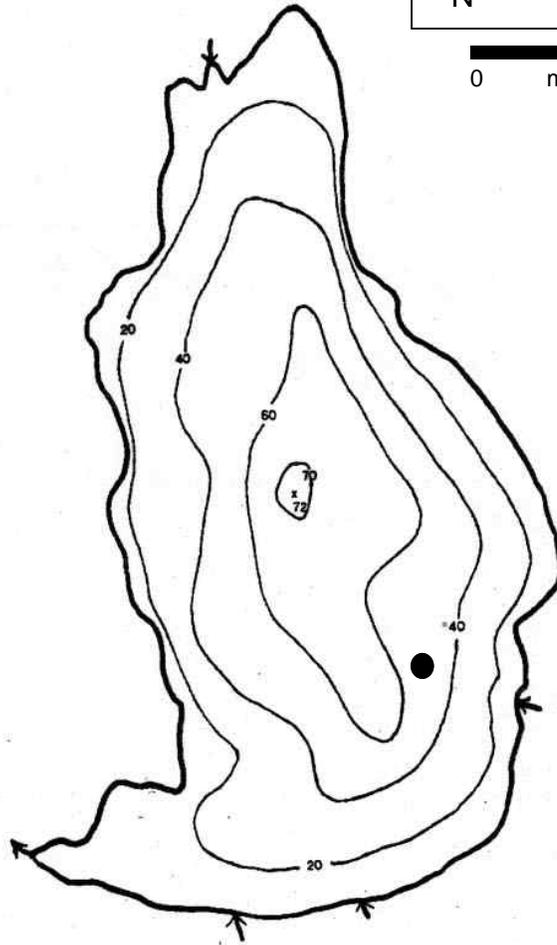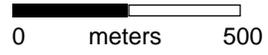
D.2 South Pond

South Pond
Stark, NH

N

South Pond

N

● Core Location

Contour Interval
20 feet

0   meters   500
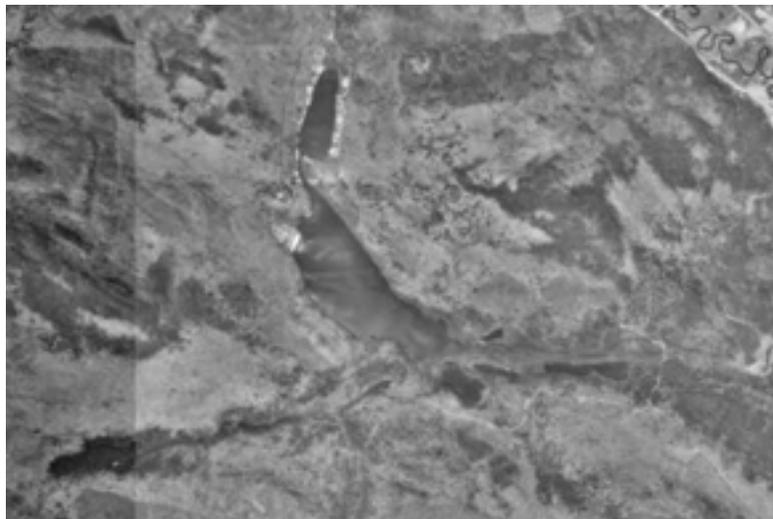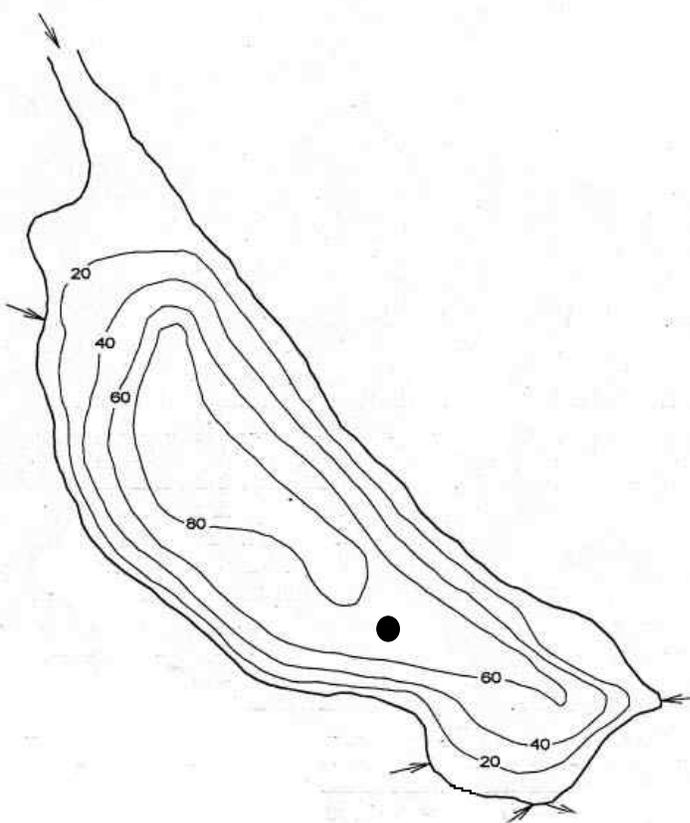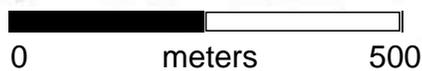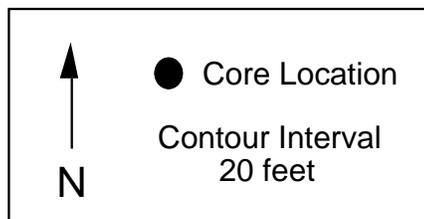
Crystal Lake
Eaton Center, NH

N

Crystal Lake

Johns Pond
Davis Township, ME

N

Johns Pond

N
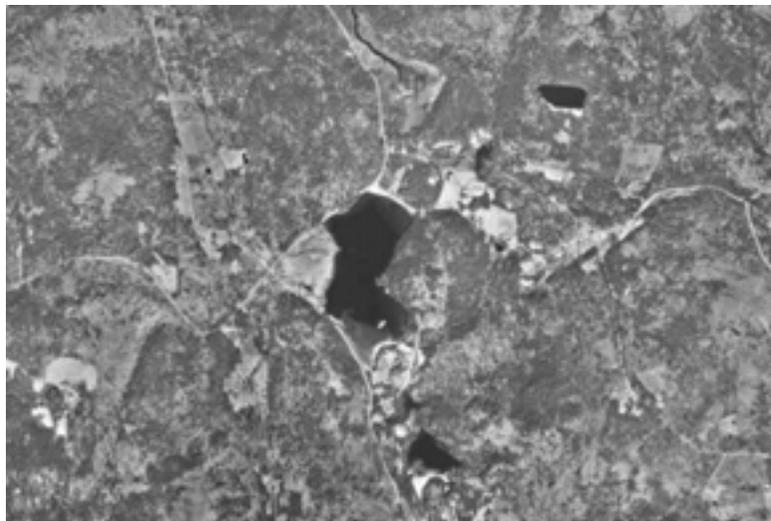
● Core Location

Contour Interval
10 feet

0　tenths of mile　3

40

30

20

10

109
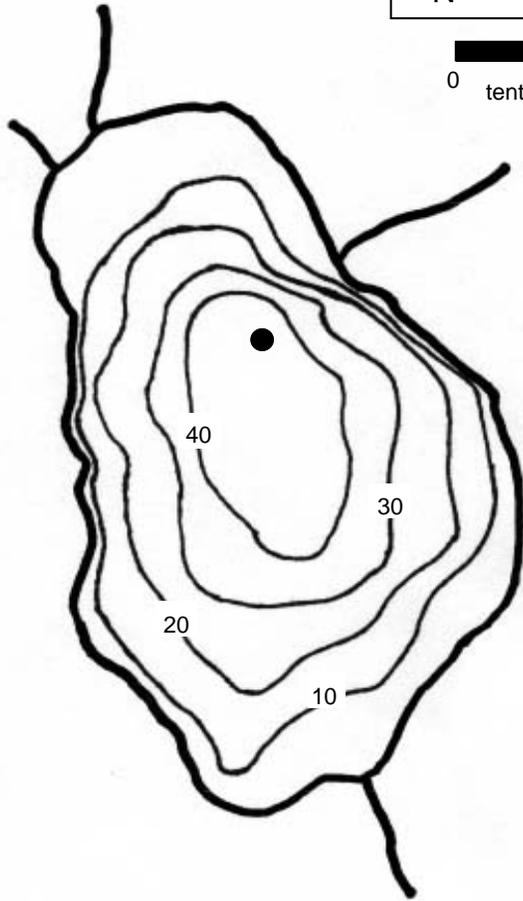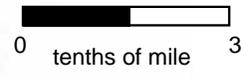
Worthley Pond
Peru, ME

N
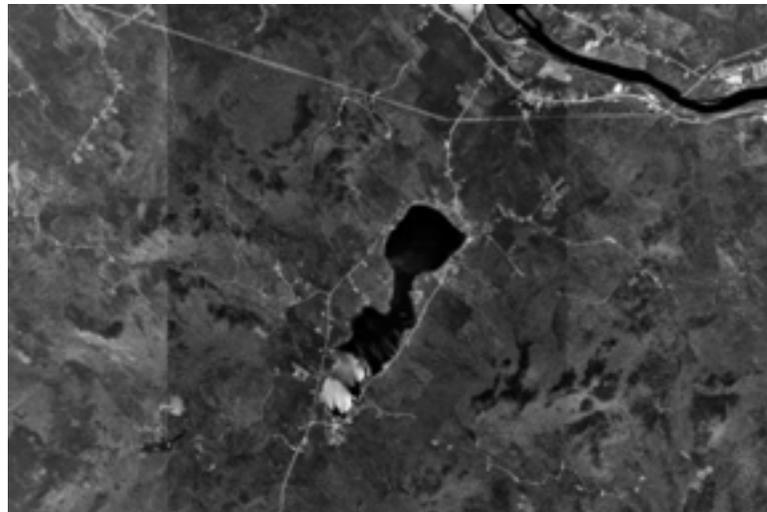
Worthley Pond

N

● Core Location

Contour Interval
20 feet

0   Tenths of mile   10

20

40