

UNIVERSITY OF VERMONT, in partial fulfillment of the requirements for the degree of Master of Science

QUANTIFYING THE RESPONSE RATES OF LAKE ECOSYSTEMS TO  
HOLOCENE DEGLACIATION THROUGH THE USE OF STABLE ISOTOPES

A Thesis Presented

by

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to

The Faculty of the Graduate College

of

The University of Vermont

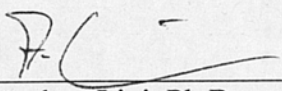
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In Partial Fulfillment of the Requirements  
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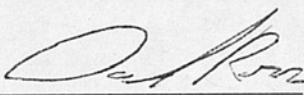
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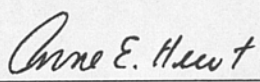
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## ABSTRACT

The analysis of the organic matter preserved in lake sediments can reveal much about the time needed for the establishment of terrestrial and aquatic ecosystems following Holocene deglaciation. Cores from two Vermont lakes (Duck Pond and Lake Morey) were analyzed to determine the stable isotope ratio of carbon and nitrogen and C/N ratios of sedimentary organic matter. These two records were then supplemented by two previously-studied lakes, Sterling Pond and Ritterbush Pond. These analyses enable one to characterize the origin (terrestrial vs. aquatic) of the organic material, as terrestrial and aquatic plants have distinct C/N ratios and often distinct isotopic values. Selected macrofossils were radiocarbon dated in order to establish temporal reference points.

The observed lithological changes at the bottom of each core and the radiocarbon dates suggest that all four cores contain complete post-glacial records. These changes are not the same in each core. In Ritterbush Pond and Duck Pond the lithological change is from gravel to silt, while in Sterling Pond and Lake Morey the sediment changes from silt/clays to gyttja. The lithological change in Lake Morey reflects the draining of glacial Lake Hitchcock, and it was only after this event that Lake Morey could form.

In all cores the organic carbon content increases and the  $\delta^{13}\text{C}$  values decrease following deglaciation. The organic carbon increase suggests that the sediments record the establishment of the local ecosystems as they progressively contribute more organic matter to the lakes. The decrease in  $\delta^{13}\text{C}$  values suggests that the aquatic ecosystems lagged behind the terrestrial ones in colonizing newly deglaciated land. Aquatically-produced organic matter in these lakes has lower  $\delta^{13}\text{C}$  values than watershed plants, and as the aquatic ecosystems evolved and contributed more organic matter to the sediment the bulk  $\delta^{13}\text{C}$  values decreased. The significant decrease (up to -13.5‰) in  $\delta^{13}\text{C}$  values documented in all four lakes is not synchronous (lasting approximately 400 years in Ritterbush Pond to 3100 years in Sterling Pond), ruling out the possibility that the change in isotopic values was due to a change in the concentration of atmospheric carbon dioxide. Instead, local differences between the lakes including elevation, input of terrestrial material, and geomorphologic history influenced the timing and rate of ecosystem establishment.

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Although only my name is on the title page of this thesis, this has certainly not been a solo project. I have received help and guidance during the completion of this thesis from many people, and to them I am most grateful.

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## CHAPTER 1. INTRODUCTION AND BACKGROUND

### 1.1 Statement of Problem

As our appreciation of how today's human population can influence the planet's climate grows, the importance of understanding past climate changes and their effects on the ecosystems also grows. The past response of ecosystems to paleoenvironmental changes may reveal how contemporary ecosystems might respond to present and future environmental changes.

Limnological records have frequently been used to study past changes in ecosystems, as each lake contains a unique record of the environmental changes that affected the local biological communities (e.g. Lin, 1996; Silliman et al., 1996; Ariztegui et al. 1996; Valero Garcés et al.; 1997, Ridge et al. 1999; Meyers and Takemura, 1997; Talbot and Laerdal, 2000). This study focuses on the organic paleolimnological record immediately following Holocene deglaciation in Vermont, which has numerous post-glacial lakes that provide a temporal record of surface and lacustrine processes influenced by climate change. In addition to supplementing previous research on paleoclimatology and paleolimnology, this study also attempts to determine the factors that affect lake ecosystem development in order to allow better comparison between different paleolimnological records. If these factors can be determined and their importance quantified, then the effect of local factors on the ecosystems can be separated from the influences of other factors, such as changes in climate.

## 1.2 History of deglaciation in New England

As this study is concerned with the most recent period of deglaciation and plant colonization, only the previous glacial events will be considered. During the last glacial period, the Wisconsinan, the North American continental ice sheet reached its maximum southern extent approximately 18,000  $^{14}\text{C}$  years ago (Benn and Evans, 1998). In New England, the ice sheet extended as far south as Long Island, New York, and Cape Cod, Massachusetts. After 18,000  $^{14}\text{C}$  years ago, the ice sheet's rate of melting was greater than its rate of advance, and the leading edge of the glacier began to retreat to the north. Although New England began to be covered in less ice, how this process took place is not entirely clear. The entire ice sheet may have retreated as a cohesive unit from the south or there might have been a region-wide downwasting of the ice with remnant ice left in valleys (Ridge et al., 1999). However, it is agreed that the mountaintops were the first areas to be deglaciated, with the lower-elevation valleys being ice-free after the mountains (Ridge et al., 1999).

The variation in deglaciation by elevation complicates simplified discussions of deglaciation history in high relief regions such as Vermont. In such areas, the generalized idea that the land was deglaciated first in the south and later in the north does not hold true. Ridge et al. (1999) show the deglaciation of southern Vermont beginning 12,600  $^{14}\text{C}$  yBP in the Connecticut River valley, but higher-elevation areas north of that were deglaciated sooner. For example, Lin et al. (1995) report a basal date of 12,760  $^{14}\text{C}$  yBP at Sterling Pond, a high-elevation pond in northern Vermont. Additionally, Ridge et al. (1999) lists at least three readvances of the ice sheet to the south, two in the

Champlain Valley and one in the Connecticut Valley, but none of these advances was extensive enough to have affected the study sites. By about 11,500  $^{14}\text{C}$  yBP the ice sheet had retreated north into Quebec and out of the study area (Ridge et al., 1999).

One important feature left after deglaciation was the post-glacial Lake Hitchcock. Lake Hitchcock formed behind a glacial moraine in central Connecticut and eventually covered parts of the modern Connecticut River valley northwards to just beyond St. Johnsbury, Vermont (Ridge et al., 1999). This large lake covered the small basin that would become Lake Morey. While the other three lakes (Ritterbush, Sterling, and Duck) formed after the ice melted, Lake Morey's development was delayed until the draining of glacial Lake Hitchcock. Recent radiocarbon data have established the timing of the draining to be 10,400  $^{14}\text{C}$  years BP (Ridge et al., 1999).

### **1.3 Plant colonization in New England**

Immediately after the glaciers melted, the land did not support any plant life. Plants were colonizing new ground from the south, following the path of the glaciers northward. As the plant communities grew and established themselves, they released pollen into the air, some of which eventually became preserved in lake sediment. This study is not the first to investigate how plants colonized the newly-deglaciated land in New England (e.g. Davis and Jacobson, 1985; Peteet et al., 1990; Maenza-Gmelch, 1997; Jackson et al., 1997). Pollen records have been quite revealing and widely studied, although there are some inherent limitations (such as a lack of spatial resolution, discussed later) to the technique (Peteet et al., 1993).

Pollen studies have been successful in documenting the succession of major types of terrestrial flora following deglaciation in New England (Davis and Jacobson, 1985; Peteet et al., 1990; Peteet et al. 1993; Lin et al., 1995; Jackson et al., 1997). These studies were able to elucidate the transition from cold-loving tundra and boreal forests, immediately after deglaciation, to warmer forests dominated by cold-intolerant species such as oak. However, Peteet et al. (1993) and many others (e.g. Maenza-Gmelch, 1997; Lin et al., 1995) have documented a 1000 year reversal of this trend. The pollen records from this period, from around 11,000 to 10,000  $^{14}\text{C}$  BP, show an expansion of the cold-tolerant boreal forests and a decline in pines and oaks. This period of cooling, termed the Younger Dryas, may have affected the growth rate of plants in the study sites.

While these and other pollen records can reveal large-scale changes, pollens' ability to be widely distributed by the wind does not permit one to study the evolution of plants around one lake (Peteet et al., 1993). Also, not all plants, especially aquatic ones such as algae, produce pollen, and those plants would be absent from the pollen record. One way to determine the difference in establishment rates between terrestrial and aquatic plants is to analyze the total organic matter preserved in lake sediment.

#### **1.4 Origin of lake sediment**

Lake sediment is not a homogenous substance but rather a mixture of different types of material coming from a variety of sources. Separating the different components of lake sediment by their origin yields three categories (Meyers, 1997). First, material can come from directly within the lake, either from the water column or from another

area of the lake bottom. Material also can originate from the surrounding lake watershed, having been washed or blown into the lake. Finally, material can be transported through the air from outside the lake watershed and become incorporated into the sediment. Plant pollen and charcoal from forest fires (and, in certain places, volcanic ash) are two of the most abundant materials transported through the air. However, since pollen and charcoal only comprise a small fraction of lake sediment (Meyers, 1997) they can be ignored, leaving only material coming from either the lake itself or its watershed.

The material coming from aquatic and terrestrial sources is composed of either inorganic or organic matter. Again, this study is mostly concerned with the organic portion of the sediment and its relative amounts and origins, minimizing the importance of the inorganic fraction of the sediment. For this study there remain two important sources for organic material, land plants and aquatic plants. The division between the two plant communities is not sharp, but rather, there is more a continuum between the two, with some plants such as sedges and cattails growing at the terrestrial/aquatic boundary.

### 1.5 Stable Isotopes

Isotopes are atoms that have the same number of protons but different amounts of neutrons. This results in isotopes of the same element having different atomic masses, the atomic mass being the sum of the protons and neutrons in an atom. Stable isotopes, unlike radiogenic ones, do not spontaneously decay into other elements and are stable



under normal conditions. This study focuses on the stable isotopes of carbon and nitrogen.

Carbon has two stable isotopes,  $^{13}\text{C}$  and  $^{12}\text{C}$ . Both have six protons, with  $^{13}\text{C}$  having seven neutrons and  $^{12}\text{C}$  only six.  $^{12}\text{C}$  is the more abundant of the two, comprising 98.9% of the carbon reservoir on earth. The other 1.1% is  $^{13}\text{C}$ . Nitrogen also has two stable isotopes,  $^{15}\text{N}$  and  $^{14}\text{N}$ , both having seven protons and  $^{15}\text{N}$  having one additional neutron. 99.6% of atmospheric nitrogen is  $^{14}\text{N}$ , with the other 0.4% being  $^{15}\text{N}$  (Hoefs, 1997).

The delta notation is used to report differences in isotopic composition between two substances. Delta ( $\delta$ ) in per mil is defined as:

$$\delta \text{‰} = [(R_{\text{sample}}/ R_{\text{standard}}) - 1] \times 1000 \quad (1)$$

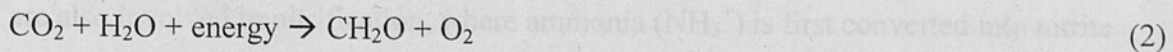
where  $R_{\text{sample}}$  is the ratio of heavy to light isotopes in the measured substance (e.g.  $^{13}\text{C}/^{12}\text{C}$  or  $^{15}\text{N}/^{14}\text{N}$ ); and  $R_{\text{standard}}$  is the ratio of heavy to light isotopes in a standard.

The ratio of heavy to light isotopes in the standard ( $R_{\text{standard}}$ ) is assumed, so the delta notation relates the difference, in per mil, between the measured substance and the standard. A negative  $\delta$  value represents a depletion of the heavy isotope in relation to the standard, while a positive  $\delta$  value represents an enrichment of the heavy isotope. Isotopic measurements are reported this way so that small differences in the isotope ratios of a substance are more easily and accurately related. The standards, by definition, all have a  $\delta$  value of 0‰. The standard for reporting  $\delta^{13}\text{C}$  is PDB, a belemnite from the Pee Dee

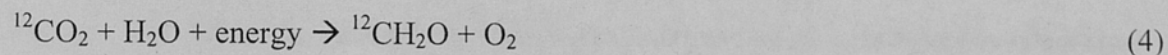
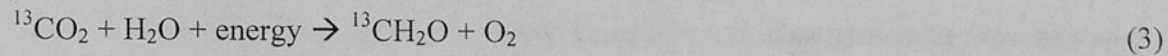
Formation, in South Carolina (Hoefs, 1997), while the standard for  $\delta^{15}\text{N}$  is atmospheric nitrogen (Air). Using internationally recognized standards allows comparison of results between samples from different laboratories, and all  $\delta$  values in this text are reported relative to these two (PDB for  $\delta^{13}\text{C}$  and Air for  $\delta^{15}\text{N}$ ) standards.

In organic materials, the isotope ratio changes mainly through kinetic processes (O'Leary et al., 1992). These kinetic processes favor the incorporation of one isotope over the other, producing isotopic fractionation (a change in the isotope ratio of a material because of a physical, biological, or chemical process) between the source and the product as a result of the reaction. Fractionation can favor either heavy or light isotopes, depending on the direction (reactants to products or vice versa) of the energy transfer. Molecular bonds formed by lighter isotopes are weaker than the bonds formed by heavier isotopes (O'Leary et al., 1992). Organisms fractionate in favor of the molecules with lighter isotopes, as the weaker bonds are easier to break (O'Leary et al., 1992). The most important process for this study, photosynthesis, produces fractionation because of differences in the bond strength of the isotopes of carbon (O'Leary et al., 1992).

Photosynthesis is the process through which plants use the energy from sunlight to produce organic matter. The simplified equation for photosynthesis is (Raven and Johnson, 1992):



Carbon dioxide and water combine with energy from sunlight to produce the basic unit of organic matter, CH<sub>2</sub>O, and oxygen is released to the atmosphere. Considering the effects that <sup>13</sup>C and <sup>12</sup>C have on photosynthesis, produces these two equations:



The first reaction uses carbon dioxide with the heavier <sup>13</sup>C, while the second involves carbon dioxide with the lighter <sup>12</sup>C. The unit of organic matter produced by each reaction differs in the weight of its carbon atom. The second reaction proceeds faster than the first one, because the CO<sub>2</sub> containing the lighter carbon atom requires less energy to break the carbon-oxygen bonds than the CO<sub>2</sub> containing the heavier one. The fractionation of isotopes during photosynthesis produces a large difference (up to 30‰) between the isotopic values of atmospheric carbon dioxide (-8‰) and the plant tissue (Fogel and Cifuentes, 1993).

Most plants are not able to utilize atmospheric nitrogen directly, as they do with carbon dioxide. Instead, they depend on bacteria in the soil to convert nitrogen into forms they can use. Bacteria produce the majority of nitrogen available to plants through nitrogen fixation (N<sub>2</sub> → NH<sub>3</sub>; Raven and Johnson, 1992). These bacteria live symbiotically on the roots of plants (as is the case with legumes) or in the soil. Bacteria are also involved in nitrification, where ammonia (NH<sub>3</sub><sup>+</sup>) is first converted into nitrite

(NO<sub>2</sub><sup>-</sup>) and then into nitrate (NO<sub>3</sub><sup>-</sup>), which can then be used by plants (Fogel and Cifuentes, 1993; Schlesinger, 1997).

## 1.6 Isotopic and elemental composition of plants

Isotopic analyses are able to discriminate between plant groups because of two factors. The first is that different groups of plants do not fractionate isotopes similarly, and the second is that the sources of carbon and nitrogen available to the plants have distinct isotopic values. These two factors create isotopic signatures in plants that can be used, in favorable situations, to identify the sources of sedimentary organic matter.

Plants use one of the three pathways for photosynthesis. These three processes are C<sub>3</sub>, C<sub>4</sub>, and CAM photosynthesis (Bender, 1971). The various pathways differ in the enzyme they use to transport the carbon atom during photosynthesis, resulting in different fractionation values (O'Leary et al., 1992). Plants using the CAM process are succulents such as cacti, which are not present in Vermont or preserved in the sediment of the four study lakes, and will not be discussed further. C<sub>3</sub> plants use the CO<sub>2</sub>-fixing enzyme ribulose, 1,5-bisphosphate carboxylase/oxygenase (Rubisco) (Collatz et al., 1998). The CO<sub>2</sub> reaches the Rubisco enzyme in C<sub>3</sub> chloroplasts only after diffusing through leaf tissue, intracellular space, and the cell wall. Isotope fractionation occurs not only from the use of Rubisco but through this diffusion process, creating a large total fractionation value of about -20‰. By comparison, C<sub>4</sub> plants use a different enzyme (PEP carboxylase) during photosynthesis and actively transport carbon dioxide into the chloroplasts, producing a fractionation of only around -6‰. If the source of the carbon is

the atmosphere (which has a value of  $-8\text{‰}$ ) then  $C_3$  plants will have average isotopic values of  $-28\text{‰}$  (ranging from  $-24\text{‰}$  to  $-34\text{‰}$ ) and  $C_4$  plants of  $-14\text{‰}$  (ranging from  $-6\text{‰}$  to  $-19\text{‰}$ ) (Smith and Epstein, 1971).  $C_4$  plants are mostly grasses that tend to dominate  $C_3$  plants when the climate is hotter or arid and the atmospheric carbon dioxide levels are lower (Collatz et al., 1998). However,  $C_3$  plants can come to dominate areas that would normally favor  $C_4$  plants because almost all trees use the  $C_3$  pathway, and they can shade out the smaller  $C_4$  grasses. Additionally, there are a few tundra grasses that use the  $C_4$  pathway for photosynthesis. Today, the heavily-forested watersheds of the four lakes are composed overwhelmingly of  $C_3$  trees and shrubs.

Plant groups also differ in their elemental composition. These differences are based on the relative amounts of carbon and nitrogen in a plant, and can be expressed as the "C/N ratio" of the plant. The amount of cellulose in a plant is directly proportional to a plant's C/N value (Meyers and Ishiwatari, 1993; Meyers, 1997). Plants use cellulose as a building block, so vascular plants (trees and bushes) have higher amounts of cellulose and higher C/N values, while nonvascular plants (e.g. algae) have less cellulose and lower C/N values (Meyers and Ishiwatari, 1993). Algae also contain more protein, which is rich in nitrogen (Meyers and Ishiwatari, 1993). Since most vascular plants live in terrestrial environments, and most nonvascular plants are aquatic, C/N values are useful in determining the origin of sedimentary organic matter (Meyers and Ishiwatari, 1993). Organic matter derived from terrestrial plants will have C/N values around 20 or higher, while aquatically-derived organic matter will usually have C/N values less than 10 (Meyers and Ishiwatari, 1993). Bulk organic matter with C/N values in between these

two is most likely a mixture of the two sources. However, aquatic macrophytes (e.g. sedges and cattails) are vascular and have similar C/N values to other vascular, terrestrial plants (Meyers, 1997), making conclusions based solely on C/N values often misleading.

### 1.8 Studies using $\delta^{13}C$ , $\delta^{15}N$ , and C/N ratios of organic matter

### 1.7 Isotopic composition of carbon and nitrogen sources

In addition to the mechanisms that plants use to incorporate carbon and nitrogen, the sources of those elements and their different isotopic compositions are important to consider. For terrestrial plants, the ultimate source of both the carbon and nitrogen is the atmosphere. For the atmospheric carbon reservoir, the  $\delta$ -value is  $-8\text{‰}$ , and for nitrogen it is  $0\text{‰}$ .

Some aquatic plants, such as aquatic macrophytes, can also incorporate carbon from the atmosphere, and would thus use the same reservoir as terrestrial plants. These emergent macrophytes would thus have a carbon isotopic signature similar to terrestrial plants. However, other aquatic macrophytes and algae, in particular, take in carbon and nitrogen that is dissolved in the lake water. Without biological activity the dissolved carbon and nitrogen are in isotopic equilibrium with the atmosphere (Hodell and Schelske, 1998). The amount of dissolved carbon is affected by temperature, elevation, and the amount of mixing a lake is subjected to (Horne and Goldman, 1994). The dissolved nitrogen concentrations are determined more by biological factors, as aqueous nitrogen gas is not directly usable by plants but must be converted into a different form of nitrogen first (Horne and Goldman, 1994). However, aquatic plants discriminate against

Takemura, 1997). These two opposite effects from the same action can complicate the

the heavier  $^{13}\text{C}$  and  $^{15}\text{N}$ , and if plant growth is significant, it will affect the isotopic composition of the dissolved carbon and nitrogen reservoirs (Hodell and Schelske, 1998).

### 1.8 Studies using %C, %N, and C/N ratios of organic matter

Total organic carbon and nitrogen in sediment have been used by a number of researchers as a proxy for paleoproductivity (Schelske and Hodell, 1991; Hodell and Schelske, 1998). Willemse and Tornqvist (1999) analyzed sediment from six Greenland lakes and found that fluctuations in the amount of sedimentary organic carbon matched changes observed in the Greenland Ice Core Project (GISP)  $\delta^{18}\text{O}$  record. When the GISP record indicates higher temperatures in Greenland the amount of organic carbon increases concurrently. The studied lakes, having no inlets or outlets, were chosen to minimize fluctuations in the delivery of inorganic sediment to the lake basin. Therefore, the amount of organic carbon in these lakes was most likely controlled by paleoproductivity, as the amount of inorganic material being contributed to the lakes remained fairly constant through time. This record suggests a cause and effect relationship between temperature and paleoproductivity for these lakes.

Changes in the amount of inorganic sediment can affect the relative (%C) amount of organic carbon in two ways. First, detritus can dilute the amount of organic material, lowering the %C and %N values (Meyers and Ishiwatari, 1993). However, inorganic sediment can increase the production of algae within the lake by providing nutrients, boosting the %C after (but not within) the deposition of an inorganic layer (Meyers and Takemura, 1997). These two opposite effects from the same action can complicate the

record, making paleoproductivity estimates based solely on total carbon amounts possible only in special cases.

Ariztegui et al. (1996) analyzed organic matter from sediment from Lake St. Moritz, Switzerland. Their results show a clear increase in the percentage of organic carbon in the sediment immediately following deglaciation. Surprisingly, the authors do not discuss this increase but focus on later fluctuations in the organic record that document supposed periods of increased productivity and/or periods of increased preservation rates. Similar increases in the organic carbon content at the beginning of the Holocene are also found in the sediment records of Lake Ontario (Silliman et al., 1996), Lake Washington (Meyers and Ishiwatari, 1993), Lake Tiberius, Sweden (Hammarlund et al., 1997), Pickerel Lake, South Dakota (Schwalb and Dean, 1998) and Lake Baikal (Qiu et al., 1993).

C/N ratios have been widely used to determine the origin of organic matter (Talbot and Johannessen, 1992; Meyers and Ishiwatari, 1993; Meyers, 1997). Meyers and Takemura (1997) analyzed a 430,000 year long sediment record from Lake Biwa, Japan, and used C/N values to determine the origin of the organic matter. The low C/N values (~ 10 for the majority of the core) suggest that algae produced most of the organic matter. Few periods of higher C/N values correspond with changes in the sediment from fine-grained lacustrine silt and clay to fluvial sand and gravel. These intervals with higher C/N values and coarser sediment were interpreted to contain a higher amount of terrestrial material than the rest of the core, which has lower C/N values and finer sediment (Meyers and Takemura, 1997). The authors also estimated the accumulation



rate of the organic matter, and found that it was much higher during interglacial than during glacial periods. The average grain size also increased during the interglacial period, but the C/N values did not. They concluded that the elevated amount of terrestrial inorganic material (inferred from the larger grain size of the sediment) delivered large quantities of nutrients to the lake, stimulating algal production. Enhanced algal production caused more organic matter to be delivered to the sediment. They further suggested that the increased amount of inorganic material delivered to the lake was due to increased precipitation during the interglacial periods. However, they did not discuss what other landscape changes (e.g. vegetation cover, fire frequency) occurred locally during the interglacial periods that may have also affected the transport of terrestrial inorganic matter.

### **1.9 Studies using C and N stable isotopes**

Although research on the establishment rate of ecosystems following deglaciation is not extensive, other projects using the stable carbon and nitrogen isotopic record from sedimentary organic matter have been completed (e.g. McKenzie, 1985; Talbot and Johannessen, 1992; Ariztegui et al., 1996; Brenner et al., 1998; Hodell and Schelske, 1998). These projects have revealed changes in the paleoproductivity of the lakes and fluctuations in temperature, precipitation, and the vegetation growing in a lake's basin.

Algae growing within a lake preferentially incorporate  $^{12}\text{C}$  over  $^{13}\text{C}$ , which causes an enrichment of  $^{13}\text{C}$  in the dissolved carbon pool of the lake (Brenner et al., 1999). As the algae continue to grow and incorporate carbon, they eventually have to utilize more

$^{13}\text{C}$ . The result is that more positive  $\delta^{13}\text{C}$  values from algae often represent periods of higher algal productivity (Brenner et al., 1999).

Hodell and Schelske (1998) applied this concept to Lake Ontario and found a good correlation between the increase in carbon isotopic values of algal organic matter in the sediment and the dramatic increase in algal production (inferred from organic carbon accumulation rates) following changes in land use in the lake's watershed. During pre-settlement, the lake was phosphorus limited, but after European colonization and clear-cutting in the watershed, more runoff delivered high loads of phosphorus to the lake. Increased phosphorus promoted algal growth and, consequently, the isotopic composition of the algae began to increase. A further increase occurred in the second half of the 20<sup>th</sup> century, when large amounts of phosphorus, in the form of treated sewage and laundry detergents, were delivered to the lake. Once phosphorus input to the lake was reduced in the early 1970's, the carbon isotopic values also decreased as the algal production diminished. This correlation between paleoproductivity levels and carbon isotope values was also found in sediments from lakes in Florida (Brenner et al., 1999).

Ariztegui et al. (1996) analyzed the organic portion of the sediment from Lake St. Moritz for stable carbon isotopes. They found that the isotopic composition of sedimentary organic carbon varied with the source of the organic matter. In sediment with higher amounts of aquatic material, the organic carbon content was higher and carbon isotopic values were lower (-29 ‰). In contrast, when the relative amount of aquatic material declined (and, consequently, the relative amount of terrestrial material increased) the total organic carbon content decreased and the  $\delta^{13}\text{C}$  values increased (-27

‰). The decrease in organic matter was caused by the dilution of the organic matter by terrestrial inorganic material. Ariztegui et al. (1996) suggested that the difference in isotopic values was caused by either a difference in the isotopic composition of the sources (terrestrial vs. aquatic) or by a change in the preservation rates of organic matter on the lake bottom. If the preservation rate is low the isotopic value would increase, as decomposition preferentially mobilizes the lighter isotopes (Ariztegui et al., 1996).

Isotopic studies have also been completed on tropical lakes. Talbot and Johannessen (1992) analyzed sediment from Lake Bosumtwi, Ghana. They found good correlation between the fluctuations in carbon and nitrogen stable isotope values over the 27,000-year length of the core. A dramatic decrease in both carbon (from -12‰ to -28‰) and nitrogen (from 15‰ to 3‰) isotopic values in the sediment recorded a shift in the local vegetation from C<sub>4</sub> to C<sub>3</sub> plants approximately 10,000 year BP. Talbot and Johannessen (1992) stated that the vegetation change was caused by an increase in precipitation, as C<sub>3</sub> plants grow better in wetter climates. Talbot and Laerdal (2000) studied the sediment record of Lake Victoria, East Africa, and inferred similar changes in precipitation, although their conclusions are based more on the sedimentological changes in grain size than on the isotopic record.

Isotopic studies are especially effective on tropical lakes as the isotopic values of the terrestrial vegetation can change drastically, as both C<sub>3</sub> and C<sub>4</sub> plants can grow in the lake watersheds. The isotopic differences between C<sub>3</sub> and C<sub>4</sub> plants, and their different responses to changes in temperature and precipitation, can create large and easily discernible shifts in the isotopic record. Tropical lakes also are less likely to contain a

glacially interrupted record, and both Lake Victoria and Lake Bosumtwi contain Pleistocene sediment.

### 1.10 Diagenesis

An important consideration when studying organic matter in lake sediments is the potential alteration that the organic matter undergoes before it is preserved in the sediment. Organic matter undergoes changes which may alter its original composition, both in the water column and within the sediment.

Meyers and Eadie (1993) investigated the diagenesis of organic matter in the water column and in sediments of Lake Michigan, and also conducted laboratory experiments on decomposing algae. They suspended sediment traps at different depths in the water column to estimate productivity and recycling rates within the lake. From their experiments, they estimated that around 90% of the organic matter produced is recycled within the epilimnion. The organic material that is not recycled undergoes at least partial degradation from oxidizing bacteria, creating the possibility of significant changes in its original elemental and isotopic signatures. However, it appears that these changes are not large enough to erase the original differences between terrestrial and aquatic organic material. Meyers and Eadie's (1993) two sets of sediment traps in Lake Michigan recorded the organic matter's  $\delta^{13}\text{C}$  values increasing 1.1‰ and 1.4‰ over 125 m of depth, and the  $\delta^{15}\text{N}$  values decreasing 0.9‰ over 105 m. Although the total amount of organic matter decreased as it sank through the water column, the C/N values did not. The observed C/N values fluctuated as the organic matter sank and increased slightly

overall, but not enough to eliminate the original difference between the vascular terrestrial plants and the nonvascular aquatic algae. At one of the sediment traps, the C/N value increased by 1.4 (19.8 to 21.2), whereas in the other trap the increase was only 0.2 (13.1 to 13.3) (Meyers and Eadie, 1993). The shift in isotopic and C/N values would most likely be even smaller in the lakes considered for this study, as the depth of Lake Michigan (~ 140 m) is significantly greater than the deepest (Lake Morey, Duck Pond, and Ritterbush Pond, all 14 m deep) of the study lakes. The organic matter in the study lakes would sink through a much shorter water column, exposing it to degradation for a much shorter period of time.

Additional processes affect organic matter preservation once the organic matter has reached the lake bottom. The sediment can be resuspended, returning the organic matter to the water column where it undergoes further degradation. Eadie et al. (1984) estimated that 50% of the sediment in Lake Michigan is resuspended into the hypolimnion. The importance of resuspension increases when the sedimentation rate is lower, and if resuspension is significant, age inversions in the radiocarbon ages of the sediment are expected (Meyers and Ishiwatari, 1993). Resuspension is also greater in large, well-mixed lakes such as Lake Michigan (Eadie et al., 1984), and its effects would be less in the four, smaller lakes in this study, which become stratified in the summer. Stratification isolates the bottom sediments from the mixing action of the wind-stirred water (Horne and Goldman, 1994).

Bioturbation can also affect the preservation rate of organic matter. Bioturbation occurs in the top-most layers of the sediment and acts to homogenize the sediment. For

example, Robbins and Edgington (1975) determined a bioturbation depth of 10 cm in the Great Lakes.

In their study on Lake Michigan sediments, Meyers and Ishiwatari (1993) were able to show that microbial degradation of the sedimentary organic matter can cause a slight decrease of C/N ratios with depth, as the microbes mobilizes a higher percentage of the carbon than the nitrogen. After the decrease of C/N values there were still differences between the two groups (terrestrial vs. aquatic) of plants. The observed decrease does not appear to continue past the most surficial of sediments. There were no consistent shifts in either carbon and nitrogen isotopic values from surface sediment to deeper sediment, and Meyers and Ishiwatari (1993) concluded that diagenesis does not affect the original carbon and nitrogen isotopic signatures in lake sediments.

## CHAPTER 2. METHODS

### 2.1 Coring

The sediments analyzed in this study were collected using a modified Reasoner coring device (Fig. 2.1) (Reasoner, 1993). The device works best in unconsolidated sediments, and is fairly light and portable, two important factors for some of the back-country lakes in the study. The coring work was completed during the winters of 1999 and 2000, using the ice covering the lakes as a stable work platform.

The coring device consisted of a 6 m PVC pipe to collect the sediment. A rubber piston was fitted inside and placed at the bottom of the PVC pipe. After bolting a core head (approximately 1.5 m in length) to the top of the tube, a rope was attached to the core head and a metal cable to the piston (Fig. 2.1). The cable ran down through the center of the pipe and to the piston. The coring device was then lowered through a hole in the ice until the bottom of the tube was resting just on top of the sediment. The tube was lowered using the rope attached to the core head, not the metal cable from the piston, to ensure that the piston remained at the bottom of the pipe.

Once the bottom of the 6 m tube was just resting on the bottom, we secured the metal cable to the ice so that it would not move; the rubber piston remains at the sediment-water interface during the entire process. A weighted-cylinder, with a hole running lengthwise down the center of it, was then lowered through the ice, using the rope attached to the core head as a guide. This cylinder was used as a hammer to slowly tap the PVC pipe into the sediment. As the rubber piston remains at the sediment-water

interface the pipe moves downward past it and into the sediment. After the pipe was hammered 6 m into the sediment and the tube was full (or the hammer was not able to push the pipe further into the sediment) the hammer was lifted up and out of the water. A series of pulleys were connected to the core head rope, and the tube pulled to the surface. The tube was then cut into 1.5 m long sections and the ends sealed with plastic caps and duct tape. Each core section was labeled and then stored at 5° C until it was sampled.

## 2.2 Logging and Sampling

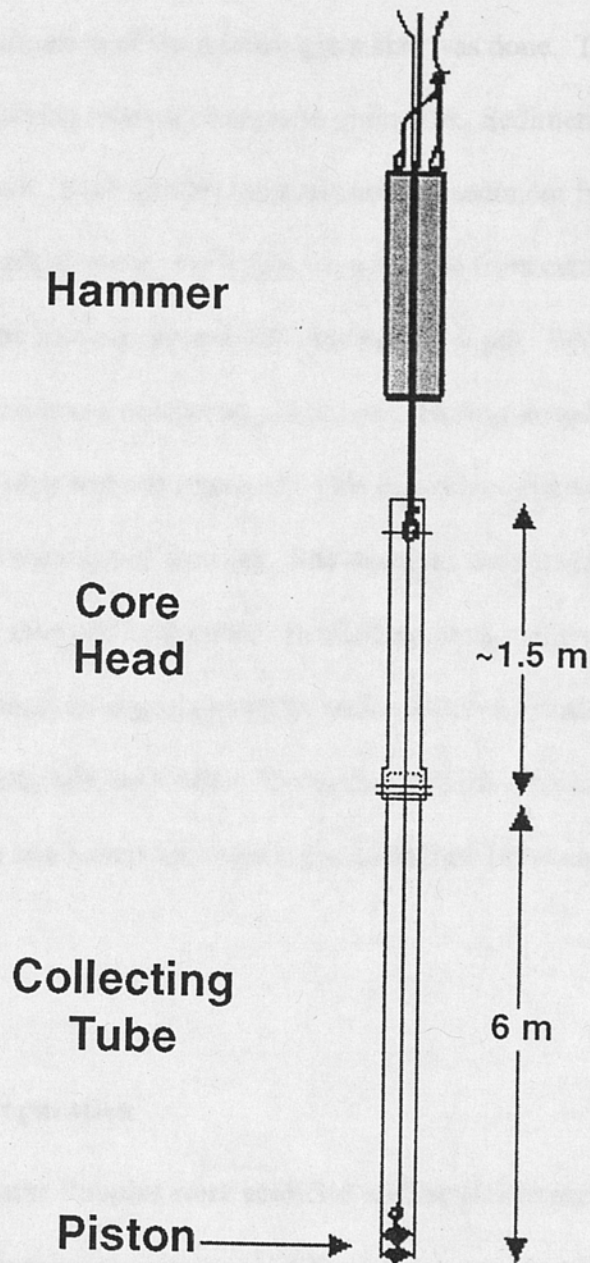
### 2.2.1 Core Splitting

The cores were opened one section at a time, halving them lengthwise. One half of the core was immediately wrapped in saran wrap, sealed within a plastic bag, and returned to cold storage. This sealed half is the archive and serves as a potential source of sediment for remeasuring or to allow future analyses without having to obtain more sediment from a previously cored lake.

### 2.2.2 Logging

Immediately after opening, each section was photographed and a visual down-core log was made, noting any color and grain size changes in the sediment. The photographing and visual logging were completed as soon as the core was opened as the sediments quickly changed colors when exposed to the air. After the visual log was





**Figure 2.1.** A diagram of the modified Reasoner coring device used to collect the cores. The device is shown at the beginning of the coring process before any sediment has entered the 6 m PVC tube. Different length core heads could be attached depending on the situation.

completed, an estimation of the relative grain size was done. This technique was effective in identifying relative changes in grain size. Sediment was sampled at each centimeter of depth. Each labeled interval contains sediment from that centimeter up to the centimeter mark above it. For example, a sample from centimeter 100 contains sediment collected between 99 and 100 centimeters depth. Sediment from each depth interval was placed into a numbered plastic vial. During sampling, the sediment near the edge of the PVC pipe was not removed. This minimizes contamination from the pipe or down-warping of the edge of the core. Sub-samples were taken for some of the layers that were thinner than one centimeter. In addition to the sampled sediment, we removed larger, discrete pieces of organic material such as leaves, cones, and twigs, and placed them in individually labeled bottles. These macrofossil samples were radiocarbon dated by Anders Noren and Karen Jennings at the Lawrence Livermore National Laboratory in California.

### **2.2.3 Sample preparation**

The sediment samples were acidified with approximately 10 milliliters of 0.1 M HCl in order to dissolve any calcite ( $\text{CaCO}_3$ ), which can contribute inorganic carbon to the sediment. The acid converts the calcite to carbon dioxide, which bubbles up and out of the sediment. After allowing the sediment to acidify for 24 hours, each vial was frozen and then placed in a freeze drier for 48 hours. The freeze drier removed the water from the samples, preserving them and allowing the samples to be stored at room temperature. All of the approximately 1,100 samples were processed this way.

## 2.3 Laboratory

### 2.3.1 Elemental analysis (%C, %N, C/N)

After freeze drying the samples, the total amount of carbon and nitrogen (weight %) was determined using a CE NC 2500 elemental analyzer. Accurately weighed amounts of the samples were placed in tin capsules and combusted. The sediment samples used were homogenous to the eye and representative of the rest of the sediment from that interval. If an interval contained coarse inorganic (gravel) or organic (twigs) material, the sediment was first pulverized to obtain a homogenous sample. The samples were initially analyzed at 5 cm intervals in order to provide reasonable resolution. The results obtained from 5 cm analyses match well with the 1 cm LOI analyses done by Anders Noren. The resolution was increased at the bottom of the core. Throughout the analyses, a known standard of ground peach leaves (46.34 %C, 2.94 %N) was also run to calibrate the machine. Duplicates were also analyzed to ensure machine precision, and the precision was approximately 1% (not  $\pm$  1%, but 1% of the quantity measured) for %C and 0.5% for %N.

The amount of carbon and nitrogen present in the sediment was determined prior to isotopic analysis for two reasons. First, the elemental data were used to focus the isotope analyses on the sediment intervals that recorded the initial increase in organic matter content. Second, the data were used to calculate the ideal amount of sediment to use for isotope analysis.

### 2.3.2 Carbon isotopes

The sediment was weighed so that approximately 100 micromoles was available for analysis by the mass spectrometer (Boutton, 1991). For a sample containing 25% carbon, 4.8 milligrams of sediment are needed. The sediment sample was added to a 6 mm outer-diameter quartz tube along with approximately 500 mg copper and 600 mg cupric oxide (Boutton, 1991). The air in the quartz tube was evacuated on a vacuum line and the tubes sealed with a torch. These tubes were then baked in a furnace at 900° C to combust the organic matter completely. After combustion, the tubes contain carbon dioxide, water vapor, and nitrogen gas. The carbon dioxide gas was isolated from the water vapor and nitrogen using liquid nitrogen and a isopropyl/liquid nitrogen cold trap. The carbon dioxide was then analyzed using a VG/Fisons SIRA series II mass spectrometer, with the data reported vs. the VPDB (Vienna Pee Dee belemnite) standard. Expected measurement reproducibility is  $\pm 0.05\%$  based on replicate standards (USGS22 and lab-internal standards).

### 2.3.3 Nitrogen Isotopes

For the nitrogen isotopic analyses, sufficient sediment was measured so that approximately 100 micromoles of nitrogen were produced. A calculated amount of sample was placed inside a 9 mm outer-diameter quartz tube along with 3 g cupric oxide, 2.5 g copper granules, and a smaller tube containing calcium oxide (Kendall and Grim, 1990). After evacuating the tube of air on a vacuum line, the quartz tube was sealed and combusted. The calcium oxide absorbs the carbon dioxide and water released from the

sample, leaving only nitrogen gas to be analyzed by the VG/Fisons SIRA series II mass spectrometer. The results are reported relative to the standard, atmospheric nitrogen, and reproducibility is  $\pm 0.1\%$ , based on replicate standards (IAEA-N<sub>1</sub> and IEAE-N<sub>2</sub>). The mass spectrometer and the elemental analyzer are in the University of Vermont's Environmental Stable Isotope Laboratory, located in the Department of Geology.

## CHAPTER 3. DATA

This chapter presents the data collected from Lake Morey and Duck Pond by the previously described field and laboratory methods. The geographic setting of each lake will be described first, followed by the lithology, the elemental (%C, %N, and C/N ratios) data, and finally the isotopic data.

### 3.1 Lake Morey

#### 3.1.1 Physiographic data

The sediment from Lake Morey, Vermont, was cored in February 1999. Lake Morey is situated 127 m above sea level (Table 3.1) in eastern Vermont (Fig. 3.1 and Fig. 3.2). The core was taken near a large delta on the western edge of the lake (Fig. 3.3). The greatest depth of Lake Morey is 13 m (Table 3.1), and the local drainage basin covers an area of 20.7 km<sup>2</sup>. The drainage basin has a relief of 414 m.

#### 3.1.2 Lithology

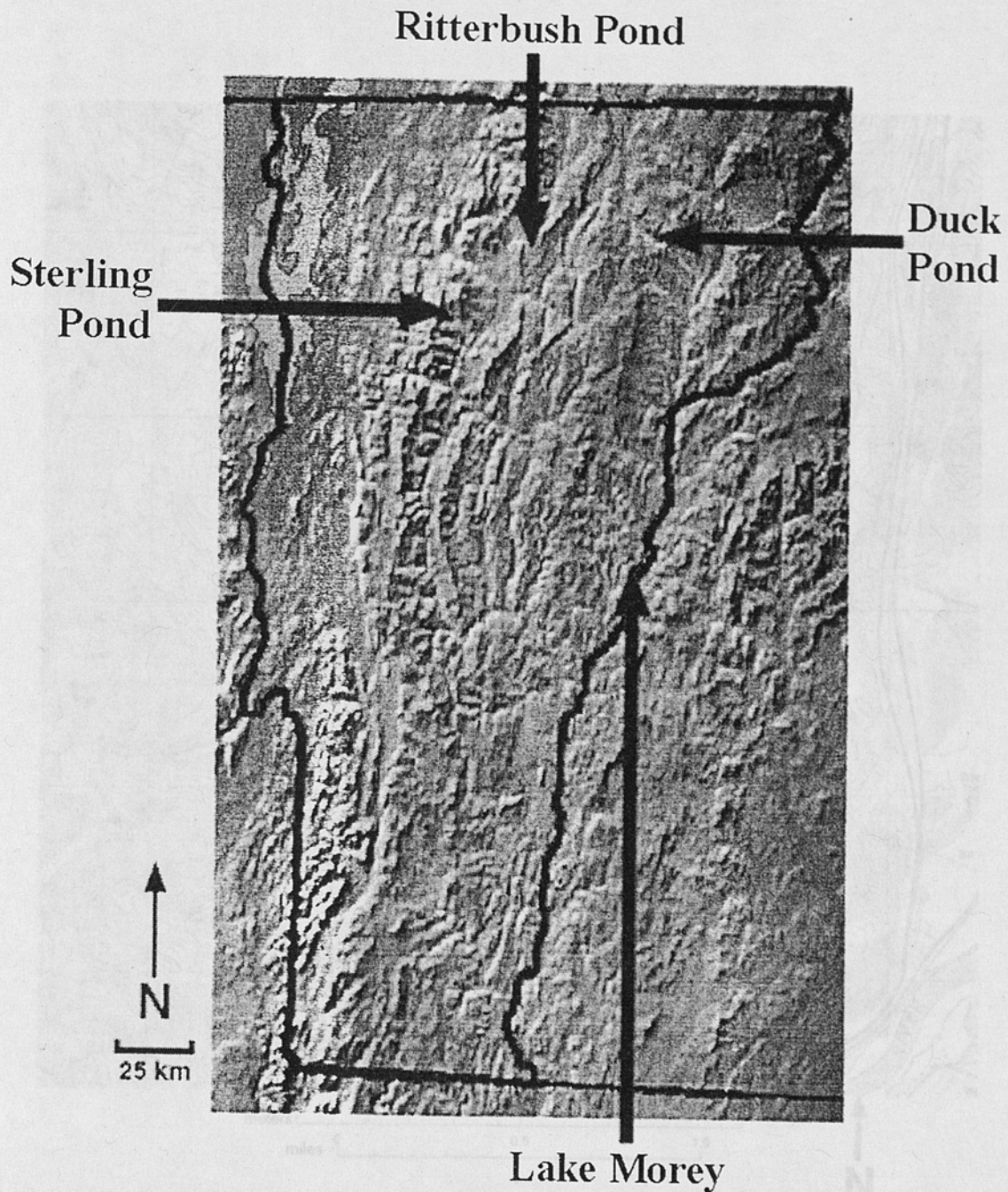
See Appendix A1 for a more detailed description of the Lake Morey core. From the bottom of the core at 560 cm until 472.5 cm the sediment is mostly gray, fine silt interspersed with several darker, coarser layers of fine sand (Fig. 3.4). At 472.5 cm there is a 2.5 cm thick layer of medium sand, the coarsest material found in the core. From 470 cm until 454 cm the sediment is mostly gray silt, although there are multiple darker,

**Table 3.1.** The physiographic data for Lake Morey and Duck Pond.

Lake Name	Surface area (km <sup>2</sup> )	Maximum depth (m)	Elevation (m)	Drainage basin area (km <sup>2</sup> )	Drainage basin relief (m)
Lake Morey	2.22	13	127	20.7	414
Duck Pond	0.03	14	520	0.7	290

**Table 3.2.** The age data from the radiocarbon dating completed on Lake Morey and Duck Pond. (Calibrated years are from Anders Noren, unpublished.)

Lake Name	Core depth (cm)	<sup>14</sup> C yBP	+/-	Calibrated yBP	+/-	Lab #	Sample type
Lake Morey	16	180	50	170	170	57744	Macrofossil
	124	1870	60	1800	70	57745	Macrofossil
	221	3070	40	3300	80	57746	Macrofossil
	305	4300	40	4860	90	57747	Macrofossil
	381	5720	50	6510	110	57748	Macrofossil
	419	7840	40	8610	90	62305	Macrofossil
	431	9450	52	10720	830	62306	Macrofossil
	454	10230	50	11940	350	57749	Macrofossil
	468	10270	11	12040	340	57750	Macrofossil
	501	10370	50	12320	350	57751	Macrofossil
Duck Pond	48	220	11	210	220	57711	Macrofossil
	192	2510	50	2590	150	57712	Macrofossil
	311	4040	40	4490	80	57713	Macrofossil
	480	7530	50	8330	120	57784	Macrofossil
	522	8790	50	9810	290	57714	Macrofossil
	529	8790	50	9810	290	57715	Macrofossil
	558	10900	50	12960	150	57716	Macrofossil
	567	10980	50	13020	110	57717	Macrofossil



**Figure 3.1.** A map of Vermont, showing the approximate locations of the four study lakes.

*Figure 3.2. A topographic map showing the size and elevation of Lake Morey. From the Fairlee (VT, NH) and Ferrisburgh (NH, VT) 1:24,000 USGS quadrangle.*



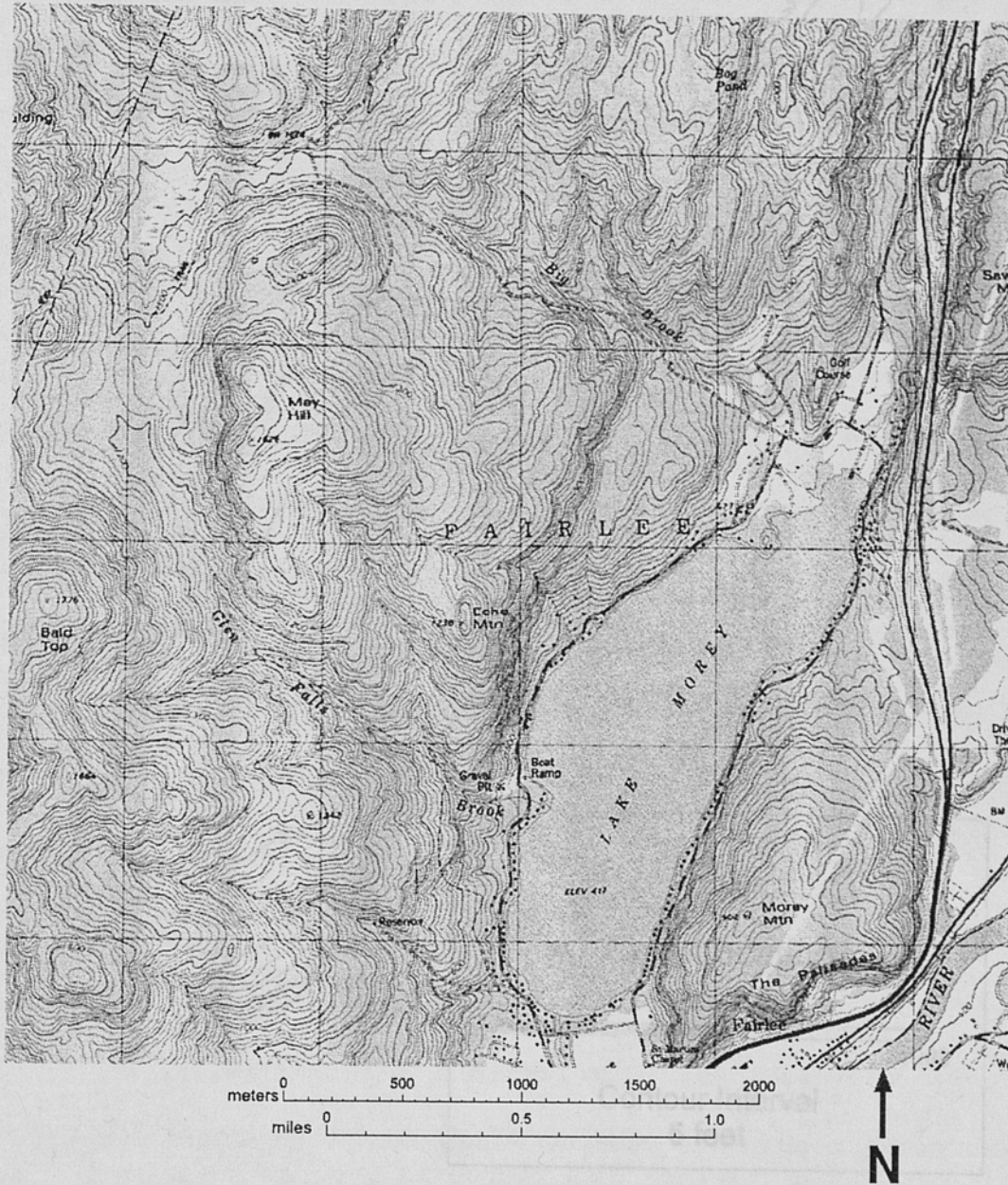
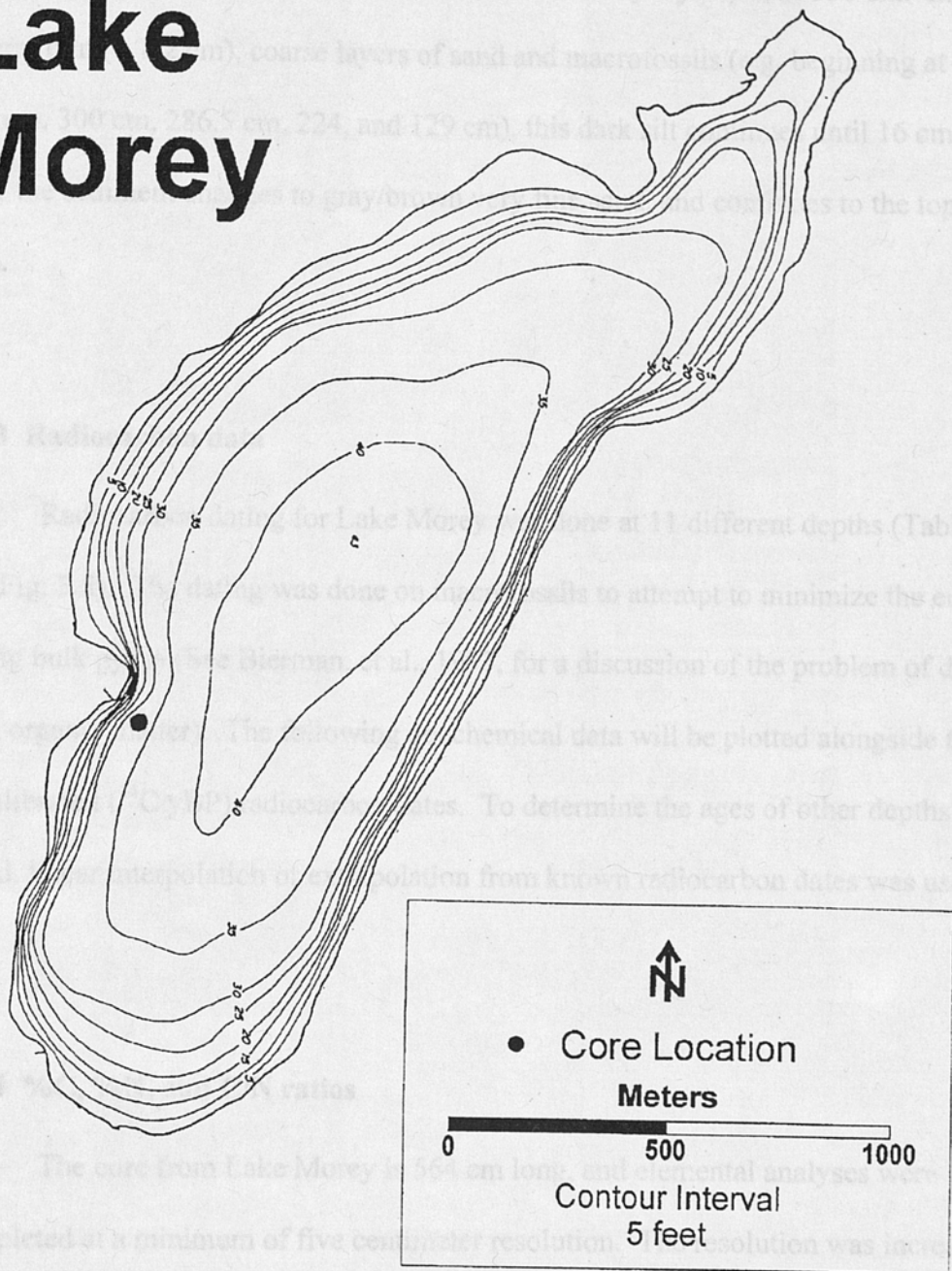


Figure 3.2. A topographic map showing the size and elevation of Lake Morey. From the Fairlee (VT, NH) and Piermont (NH, VT) 1:24,000 USGS quadrangle.

# Lake Morey



**Figure 3.3.** A bathymetry map with the depth of Lake Morey and the location where the core was taken. From the Vermont Agency of Natural Resources Lake Morey bathymetry map.

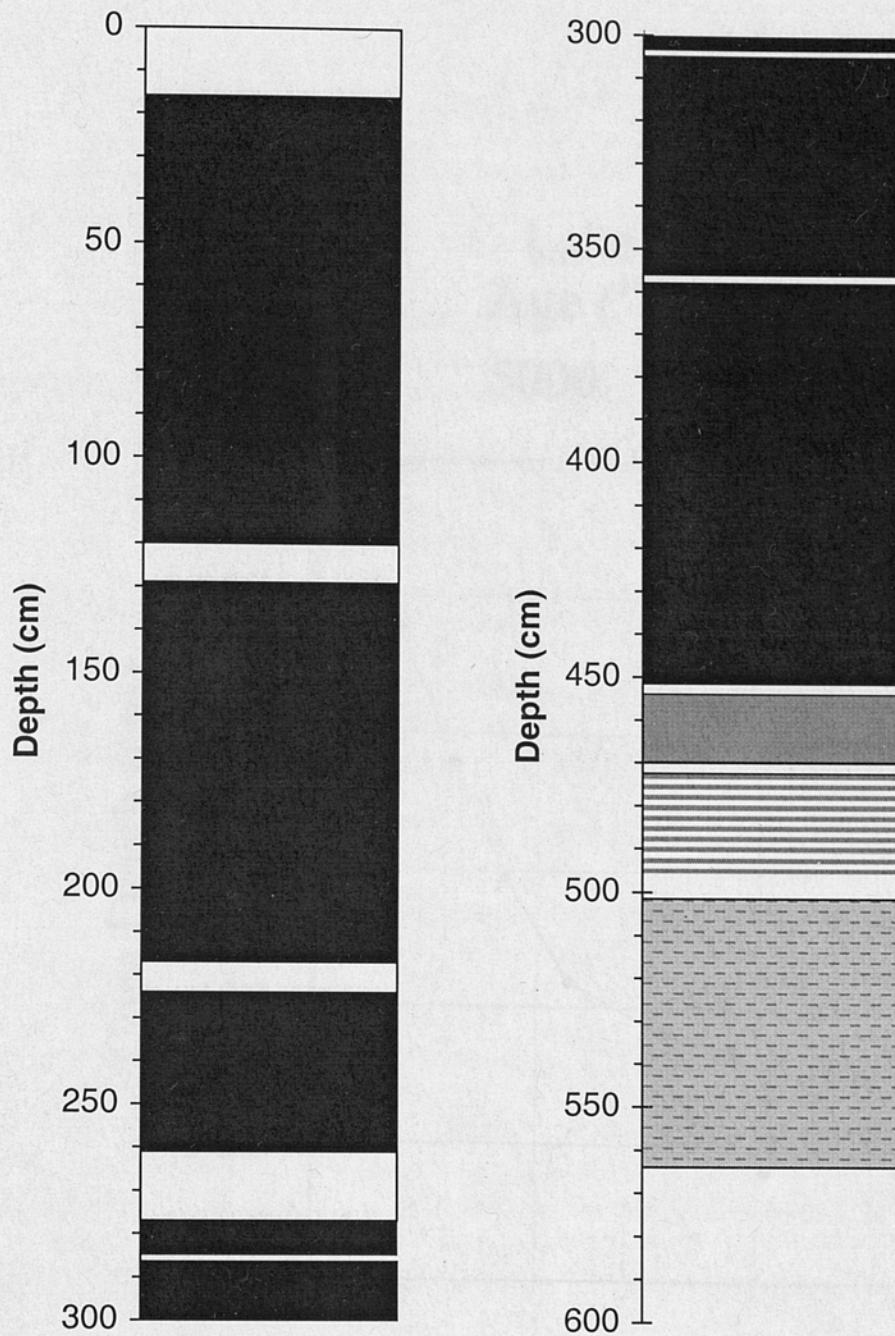
coarser layers within this interval. After 454 cm the sediment darkens to brown/black gyttja, and is mostly silt (except for a few coarser, sandy layers) until 358 cm. Except for several thin (~ 1-2 cm), coarse layers of sand and macrofossils (e.g. beginning at 358 cm, 305 cm, 300 cm, 286.5 cm, 224, and 129 cm), this dark silt continues until 16 cm depth. Here the sediment changes to gray/brown very fine sand, and continues to the top of the core.

### **3.1.3 Radiocarbon data**

Radiocarbon dating for Lake Morey was done at 11 different depths (Table 3.2 and Fig. 3.5). The dating was done on macrofossils to attempt to minimize the error from dating bulk gyttja (See Bierman, et al., 1997, for a discussion of the problem of dating bulk organic matter). The following geochemical data will be plotted alongside the uncalibrated ( $^{14}\text{C}$  yBP) radiocarbon dates. To determine the ages of other depths not dated, linear interpolation or extrapolation from known radiocarbon dates was used.

### **3.1.4 %C, %N, and C/N ratios**

The core from Lake Morey is 564 cm long, and elemental analyses were completed at a minimum of five centimeter resolution. The resolution was increased in the intervals that showed more variation.



**Figure 3.4.** Lithology of the Lake Morey core. Black represents gyttja, white is sand, gray is silt, and the dashed pattern on gray represents thin, alternating silt and clay layers.

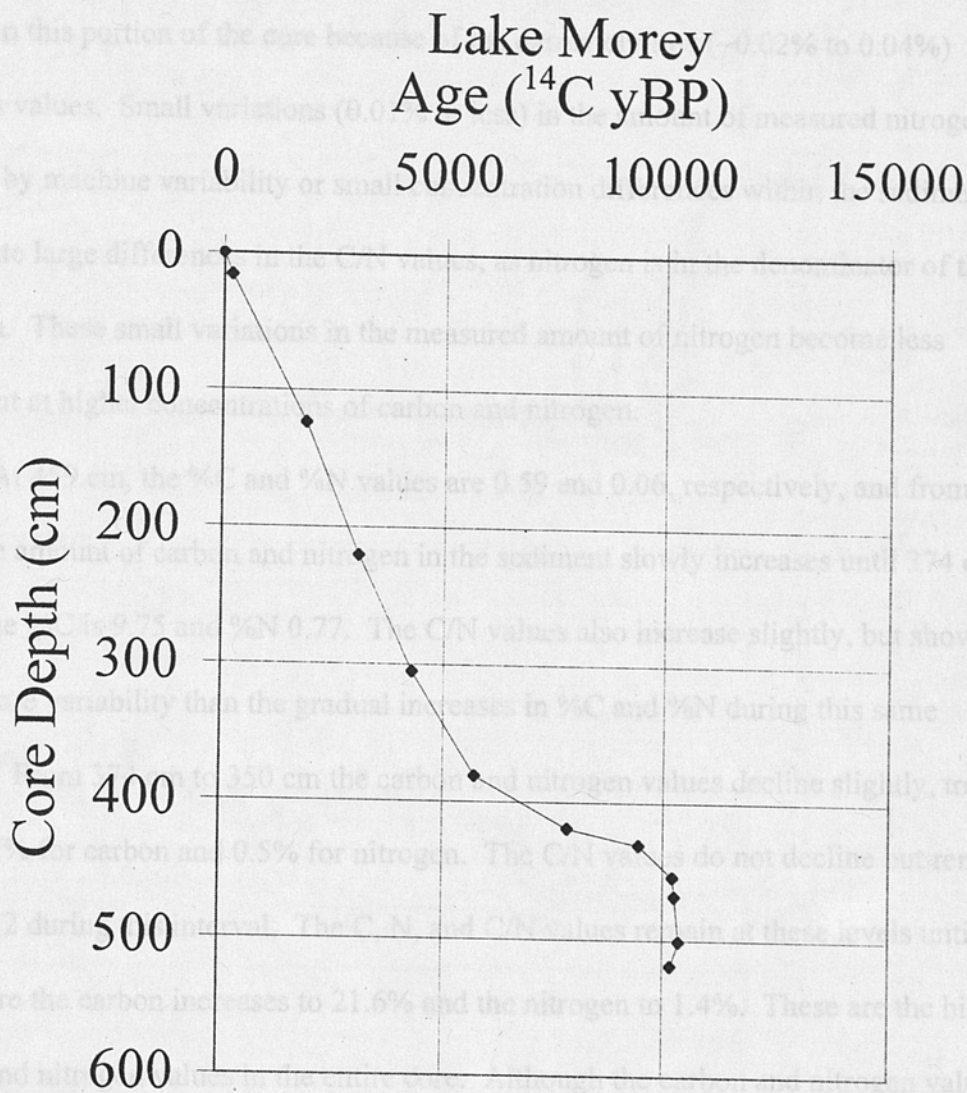
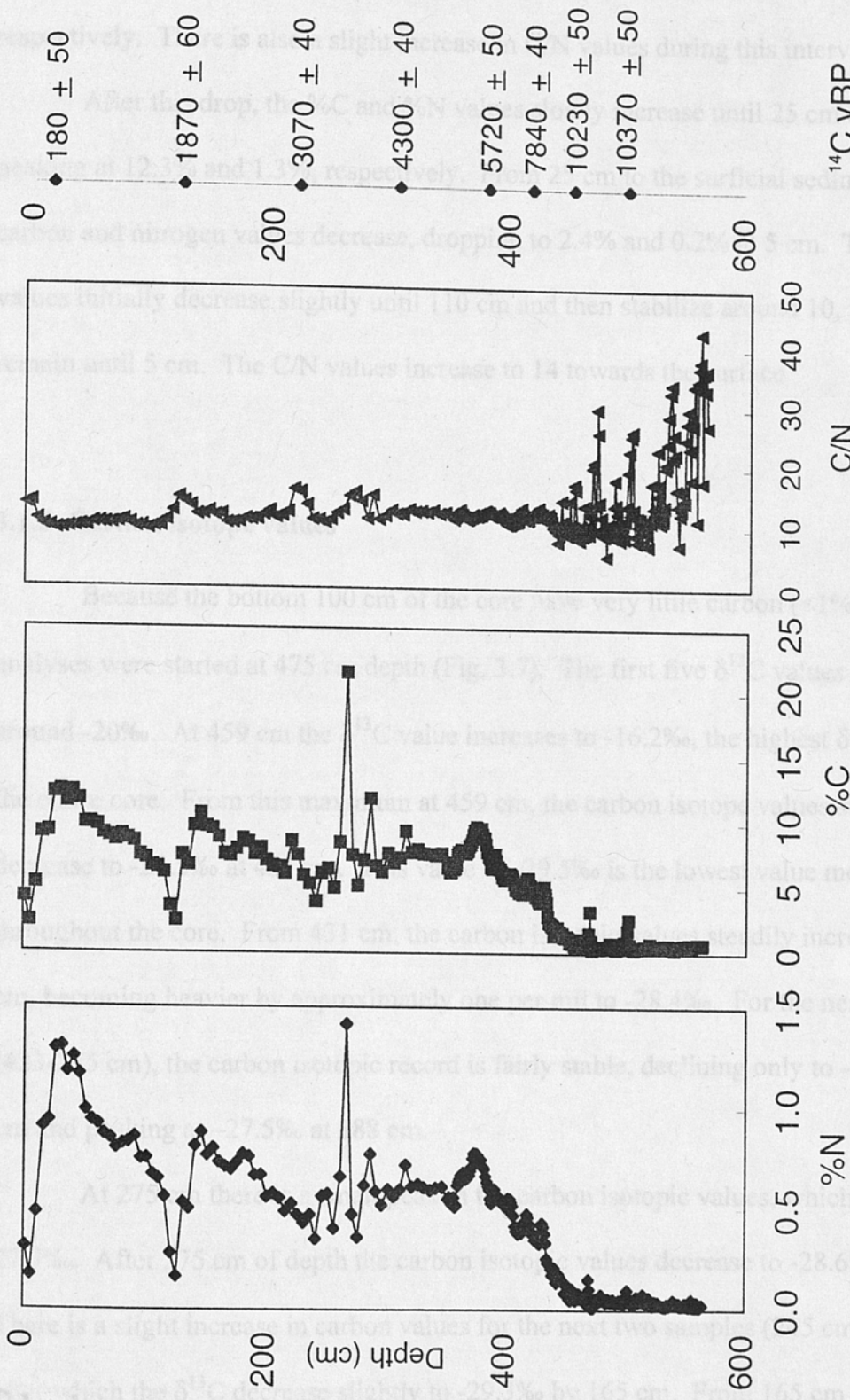


Figure 3.5. Depth vs. age plot for Lake Morey. All ages are in uncalibrated  $^{14}\text{C}$  yBP.

Starting at the bottom of the core, in the first 90 cm the %N and %C values are very low, with carbon values consistently below 1% (Fig. 3.6). The C/N values fluctuate widely in this portion of the core because of the extremely low (~0.02% to 0.04%) nitrogen values. Small variations (0.01% or less) in the amount of measured nitrogen (caused by machine variability or small concentration differences within the sediment) can create large differences in the C/N values, as nitrogen is in the denominator of the equation. These small variations in the measured amount of nitrogen become less important at higher concentrations of carbon and nitrogen.

At 459 cm, the %C and %N values are 0.59 and 0.06, respectively, and from this point the amount of carbon and nitrogen in the sediment slowly increases until 374 cm; where the %C is 9.75 and %N 0.77. The C/N values also increase slightly, but show much more variability than the gradual increases in %C and %N during this same interval. From 374 cm to 350 cm the carbon and nitrogen values decline slightly, to around 7% for carbon and 0.5% for nitrogen. The C/N values do not decline but remain around 12 during this interval. The C, N, and C/N values remain at these levels until 265 cm, where the carbon increases to 21.6% and the nitrogen to 1.4%. These are the highest carbon and nitrogen values in the entire core. Although the carbon and nitrogen values rise at 265 cm, the C/N values do not change as much, rising only to 15.2 from 12. After 265 cm the carbon and nitrogen values vary little around 9% and 0.5% until 125 cm. At



**Figure 3.6.** %N, %C, and C/N values for the core from Lake Morey. Depth in centimeters is along the vertical axes, radiocarbon ages are at the right.

125 cm and 120 cm the carbon and nitrogen values decrease to around 2.5% and 0.2%, respectively. There is also a slight increase in C/N values during this interval.

After this drop, the %C and %N values slowly increase until 25 cm of depth, peaking at 12.3% and 1.3%, respectively. From 25 cm to the surficial sediment the carbon and nitrogen values decrease, dropping to 2.4% and 0.2% at 5 cm. The C/N values initially decrease slightly until 110 cm and then stabilize around 10, where they remain until 5 cm. The C/N values increase to 14 towards the surface.

### 3.1.5 Carbon isotope values

Because the bottom 100 cm of the core have very little carbon (<1%), the  $\delta^{13}\text{C}$  analyses were started at 475 cm depth (Fig. 3.7). The first five  $\delta^{13}\text{C}$  values fluctuate around -20‰. At 459 cm the  $\delta^{13}\text{C}$  value increases to -16.2‰, the highest  $\delta^{13}\text{C}$  value for the entire core. From this maximum at 459 cm, the carbon isotope values steadily decrease to -29.5‰ at 431 cm. This value of -29.5‰ is the lowest value measured throughout the core. From 431 cm, the carbon isotopic values steadily increase until 423 cm, becoming heavier by approximately one per mil to -28.4‰. For the next 148 cm (423-275 cm), the carbon isotopic record is fairly stable, declining only to -28.9‰ at 398 cm and peaking at -27.5‰ at 388 cm.

At 275 cm there is a small peak in the carbon isotopic values, which increase to -27.3‰. After 275 cm of depth the carbon isotopic values decrease to -28.6‰ by 250 cm. There is a slight increase in carbon values for the next two samples (245 cm and 240 cm), after which the  $\delta^{13}\text{C}$  decrease slightly to -29.3‰ by 165 cm. From 165 cm to 130 cm



there is an increase of 1.8‰, and the carbon isotope values remain around -27.5‰ through 120 cm. The carbon values then decline 2‰ to -29.5‰ by 100 cm. The  $\delta^{13}\text{C}$  values fluctuate less than 0.5‰ from 100 cm until 20 cm. At 20 cm depth the carbon isotope values increase, peaking at -26.1‰ at 5 cm. The  $\delta^{13}\text{C}$  value then decreases slightly at 1 cm to -27.0‰.

### 3.1.6 Nitrogen isotope values

Nitrogen isotopes were analyzed starting at 470 cm of depth (Fig. 3.7). The  $\delta^{15}\text{N}$  values increase for the first three samples, from 3.4‰ at 470 cm to 4.5‰ at 459 cm. This value of 4.5‰ is the highest measured in the entire core. From 459 cm to 450 cm the trend reverses, with the  $\delta^{15}\text{N}$  values decreasing to 3.2‰ at 450 cm. The  $\delta^{15}\text{N}$  values then increase to 3.8‰ at 440 cm, and then steadily decline from there to 2.9‰ at 408 cm. After 408 cm the  $\delta^{15}\text{N}$  values remain around 3‰ until 250 cm. From 250 cm to 50 cm the  $\delta^{15}\text{N}$  values decline to 1.7‰. After 50 cm, the nitrogen isotopic values increase to 3.0‰ at the surface of the sediment.

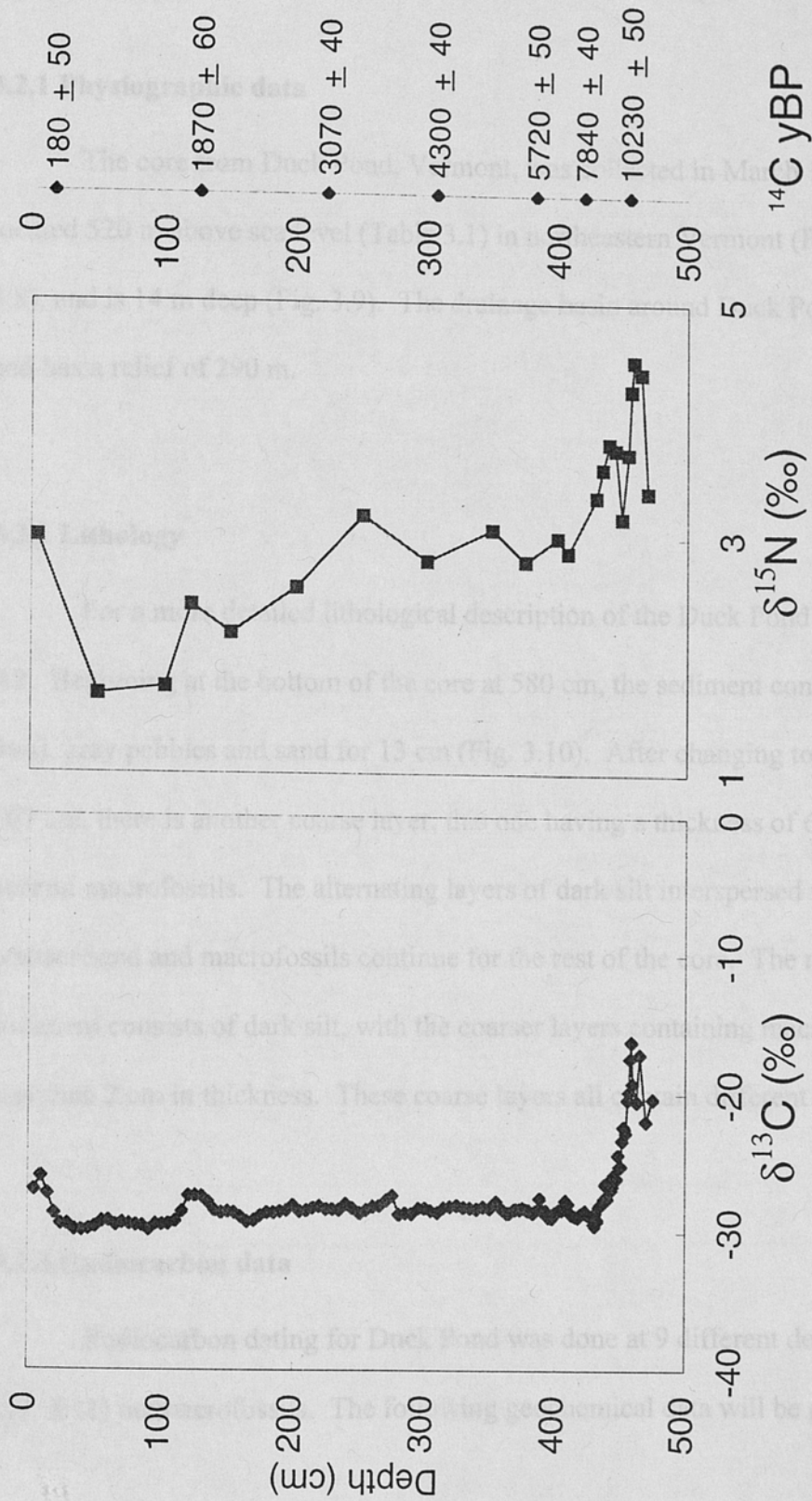


Figure 3.7.  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values for Lake Morey. Radiocarbon ages are at the right.

## 3.2 Duck Pond

### 3.2.1 Physiographic data

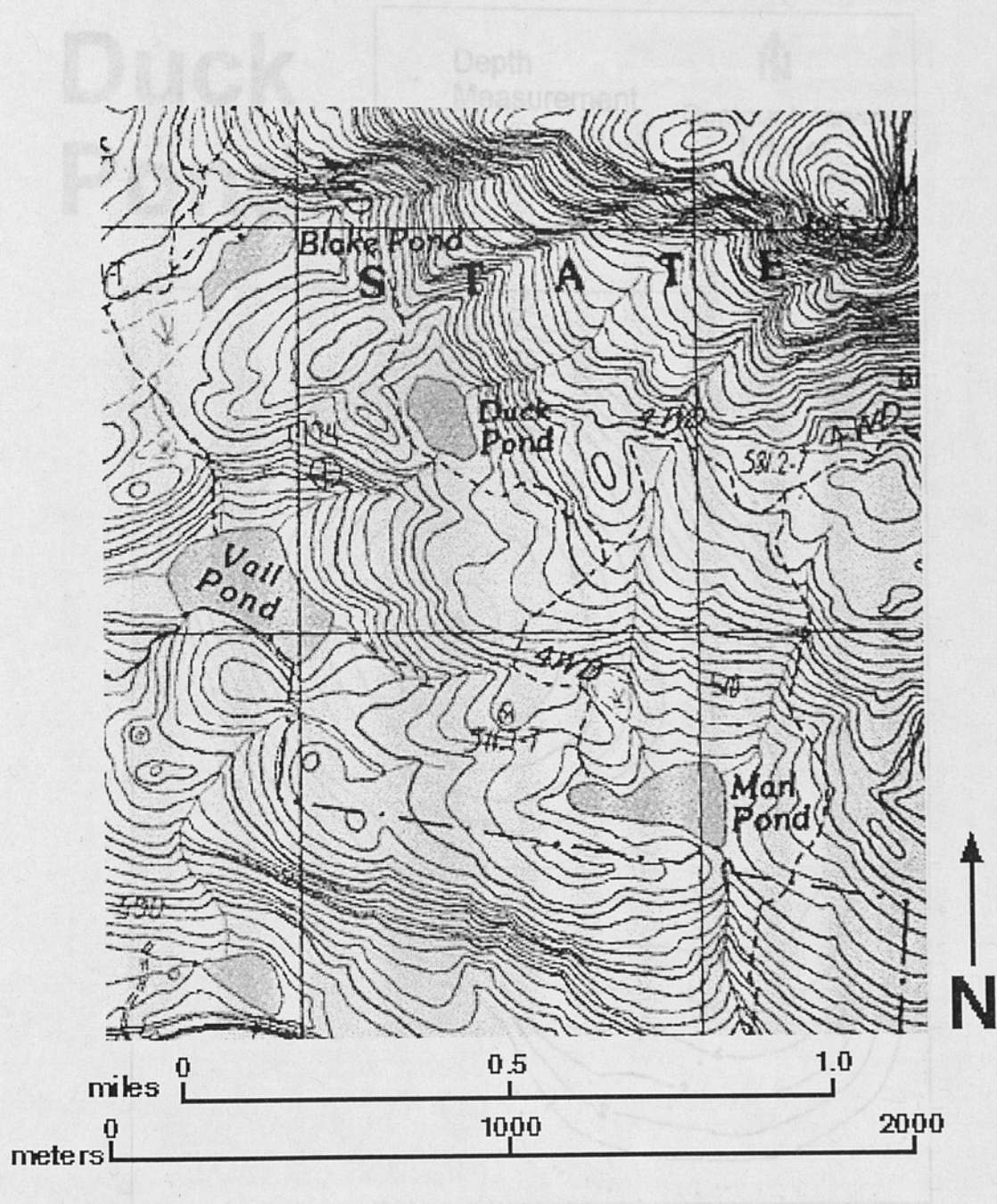
The core from Duck Pond, Vermont, was collected in March 1999. Duck Pond is located 520 m above sea level (Table 3.1) in northeastern Vermont (Fig. 3.1 and Fig. 3.8), and is 14 m deep (Fig. 3.9). The drainage basin around Duck Pond covers 0.7 km<sup>2</sup>, and has a relief of 290 m.

### 3.2.2 Lithology

For a more detailed lithological description of the Duck Pond core see Appendix A2. Beginning at the bottom of the core at 580 cm, the sediment consists of coarse (2-5 mm), gray pebbles and sand for 13 cm (Fig. 3.10). After changing to silt for 11 cm at 567 cm, there is another coarse layer, this one having a thickness of 6 cm and containing several macrofossils. The alternating layers of dark silt interspersed with layers of coarser sand and macrofossils continue for the rest of the core. The majority of the sediment consists of dark silt, with the coarser layers containing macrofossils averaging less than 2 cm in thickness. These coarse layers all contain different grain sizes of sand.

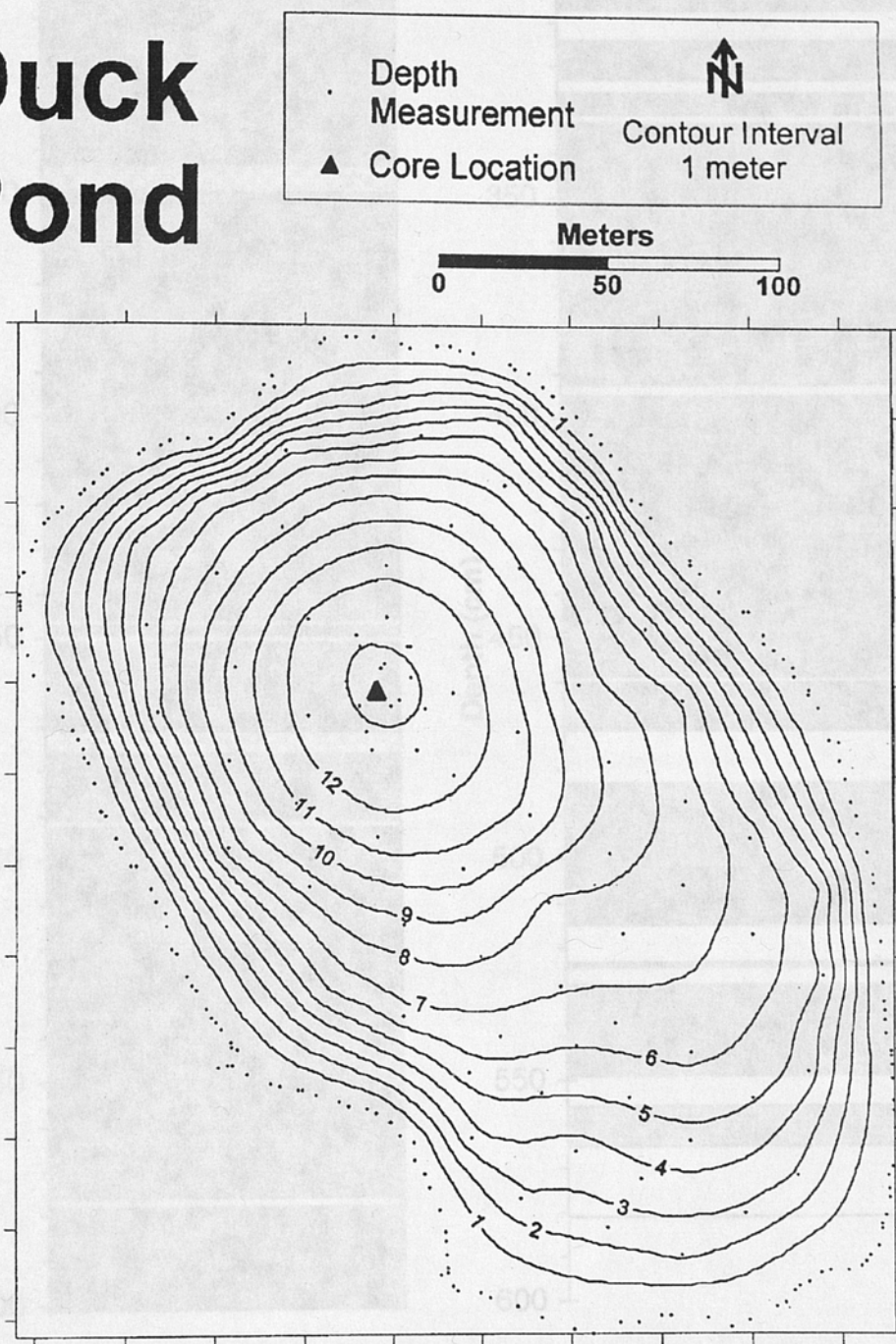
### 3.2.3 Radiocarbon data

Radiocarbon dating for Duck Pond was done at 9 different depths (Table 3.2 and Fig. 3.11) on macrofossils. The following geochemical data will be plotted alongside the

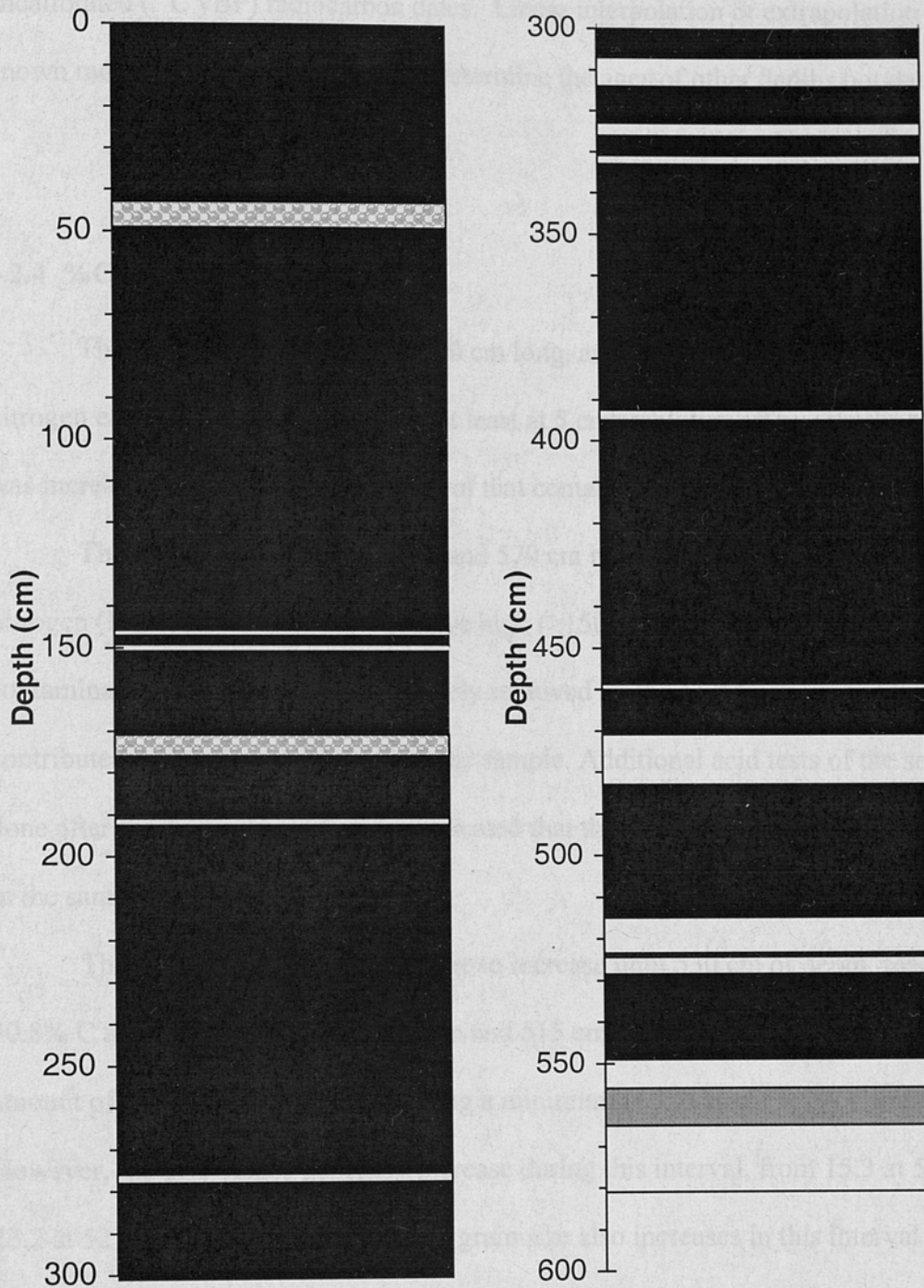


**Figure 3.8.** A topographic map of Duck Pond and its watershed. From the Sutton, VT USGS quadrangle, 1:24,000.

# Duck Pond



**Figure 3.9.** Bathymetry map of Duck Pond with the location of the core and the pond's depth. The small dots indicate where data points (either depth measurements or shoreline outline) were taken.



**Figure 3.10.** The lithology of the core from Duck Pond. The white layers represent sand, the dark layers are gyttja; the gray is silt, and the stippled pattern represents layers of macrofossils (mostly twigs and leaves).

uncalibrated ( $^{14}\text{C}$  yBP) radiocarbon dates. Linear interpolation or extrapolation from known radiocarbon dates was used to determine the ages of other depths not dated.

### 3.2.4 %C, %N, and C/N ratios

The core from Duck Pond is 580 cm long, and total organic carbon (TOC) and nitrogen elemental analysis was done at least at 5 cm resolution. The sampling resolution was increased to 1 cm during the interval that contains the initial increase in %C and %N.

The three samples at 580, 575, and 570 cm have low %C values and almost no nitrogen (Fig. 3.12). These samples have high ( $>150$ ) C/N values, as a result of calcite contamination, which was not completely removed by the acidification, and thus contributed carbon but no nitrogen to the sample. Additional acid tests of the sediment done after the original acidification indicated that there were still small amounts of calcite in the sample.

The %C and %N values continue to increase until 530 cm of depth, reaching 30.8% C and 2.0% N. Between 530 cm and 515 cm there is a general decrease in the amount of carbon and nitrogen, reaching a minimum at 520 cm of 9.7% C and 0.6% N. However, the C/N values generally increase during this interval, from 15.3 at 530 cm to 23.2 at 521 cm. The average sediment grain size also increases in this interval. By 515 cm the %C and %N levels have returned to around 32 and 2.3, respectively. The C/N values decrease to similar levels ( $\sim 15$ ) as before and the grain size also decreases. From

Figure 3.12. Depth vs. age for Duck Pond. All ages are in uncalibrated  $^{14}\text{C}$  yBP.

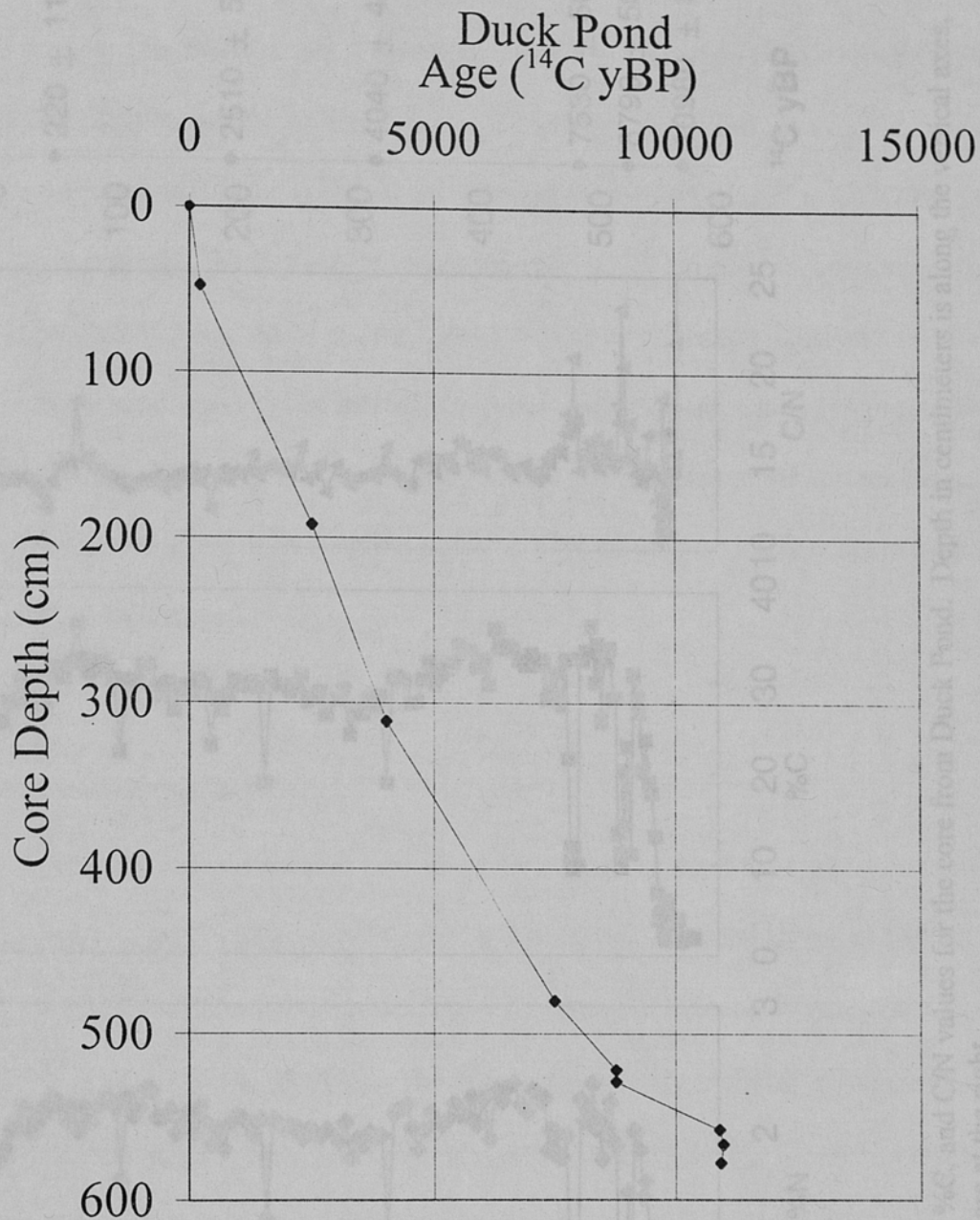


Figure 3.11. Depth vs. age for Duck Pond. All ages are in uncalibrated  $^{14}\text{C}$  yBP.



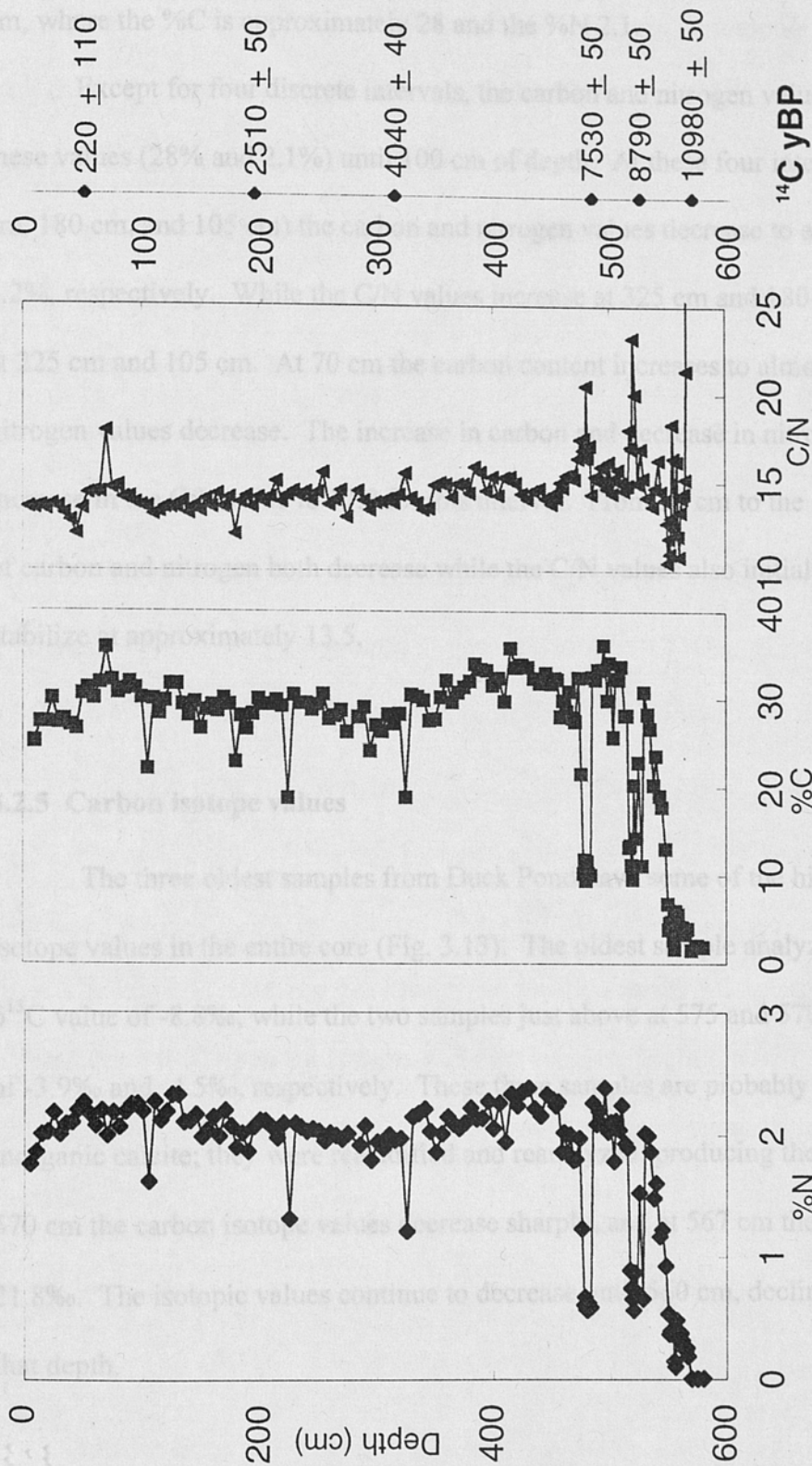


Figure 3.12. %N, %C, and C/N values for the core from Duck Pond. Depth in centimeters is along the vertical axes, radiocarbon ages are at the right.

385 cm a slight but steady decrease in the carbon and nitrogen contents occurs until 350 cm, where the %C is approximately 28 and the %N 2.1.

Except for four discrete intervals, the carbon and nitrogen values remain around these values (28% and 2.1%) until 100 cm of depth. At these four intervals (325 cm, 225 cm, 180 cm, and 105 cm) the carbon and nitrogen values decrease to around 20% and 1.2%, respectively. While the C/N values increase at 325 cm and 180 cm, they decrease at 225 cm and 105 cm. At 70 cm the carbon content increases to almost 36%, while the nitrogen values decrease. The increase in carbon and decrease in nitrogen produces an increase in the C/N value to 17.9 for this interval. From 50 cm to the surface the amount of carbon and nitrogen both decrease while the C/N values also initially decrease but then stabilize at approximately 13.5.

### 3.2.5 Carbon isotope values

The three oldest samples from Duck Pond have some of the highest carbon isotope values in the entire core (Fig. 3.13). The oldest sample analyzed at 580 cm has a  $\delta^{13}\text{C}$  value of -8.8‰, while the two samples just above at 575 and 570 have  $\delta^{13}\text{C}$  values of -3.9‰ and -4.5‰, respectively. These three samples are probably contaminated with inorganic calcite; they were reacidified and reanalyzed, producing the same results. After 570 cm the carbon isotope values decrease sharply, and at 567 cm the  $\delta^{13}\text{C}$  value is -21.8‰. The isotopic values continue to decrease until 560 cm, declining to -30.5‰ at that depth.

After 560 cm the carbon isotopic values increase to  $-24.2\text{‰}$  at 559 cm, but then decrease to  $-31.2\text{‰}$  at 557 cm. At 556 cm there is a sharp increase in isotopic values, with the  $\delta^{13}\text{C}$  value reaching  $-6.7\text{‰}$  at 556 cm. The elevated  $\delta^{13}\text{C}$  values continue somewhat at 555 cm, where the values diminish to  $-18.7\text{‰}$ .

The carbon isotopic values continue decreasing to  $-35.3\text{‰}$  at 535 cm. The  $\delta^{13}\text{C}$  values then increase to  $-28.2\text{‰}$  at 528 cm, after which they decrease back to  $-33.7\text{‰}$  at 525 cm. After 525 cm the  $\delta^{13}\text{C}$  values increase again to  $-28.0\text{‰}$  at 523 cm, and remain around this value for several centimeters, decreasing to  $-30.6\text{‰}$  only at 517 cm. The sediment's isotopic values decline further to  $-35.2\text{‰}$  at 515 cm, after which the values remain relatively constant. By 485 cm the  $\delta^{13}\text{C}$  values have increased only to  $-34.7\text{‰}$ , but from 485 cm to 483 cm there is a sharp increase to  $-28.4\text{‰}$ , which continues to  $-27.7\text{‰}$  at 478 cm. After 478 cm the isotopic values decrease to  $-32.7\text{‰}$  at 473 cm.

Over the next 140 centimeters the  $\delta^{13}\text{C}$  values are mostly stable, increasing only to  $-32.0\text{‰}$  at 330 cm. The carbon isotope values increase to  $-27.5\text{‰}$  at 325 cm, but then decrease back to  $-31.9\text{‰}$  by 320 cm. There is little variation in the carbon isotope values from 320 cm to 110 cm, but there is a small increase to  $-29.3\text{‰}$  at 105 cm. The isotope values return to  $-31.6\text{‰}$  by 100 cm, and then increase slightly to  $-31.2\text{‰}$  at 55 cm depth. After a decrease to  $-32.3\text{‰}$  at 50 cm, they slowly increase towards the top of the core, reaching  $-29.2\text{‰}$ .

### 3.2.6 Nitrogen isotope values

The two oldest samples from Duck Pond analysed for nitrogen isotopes contained the lowest  $\delta^{15}\text{N}$  values in the core (Fig. 3.13). The  $\delta^{15}\text{N}$  values at 567 cm and 563 cm are both 2.3%. After 563 cm, the nitrogen isotopic values increase to 3.0% at 550 cm, but then decrease back to 2.4% by 545 cm.  $\delta^{15}\text{N}$  values increase to 2.7% by 540 cm. The values remain at 2.7% through 535 cm and then increase to 3.0% at 530 cm. From 528 cm to 525 cm, the  $\delta^{15}\text{N}$  values decrease again to 2.2% at 455 cm.

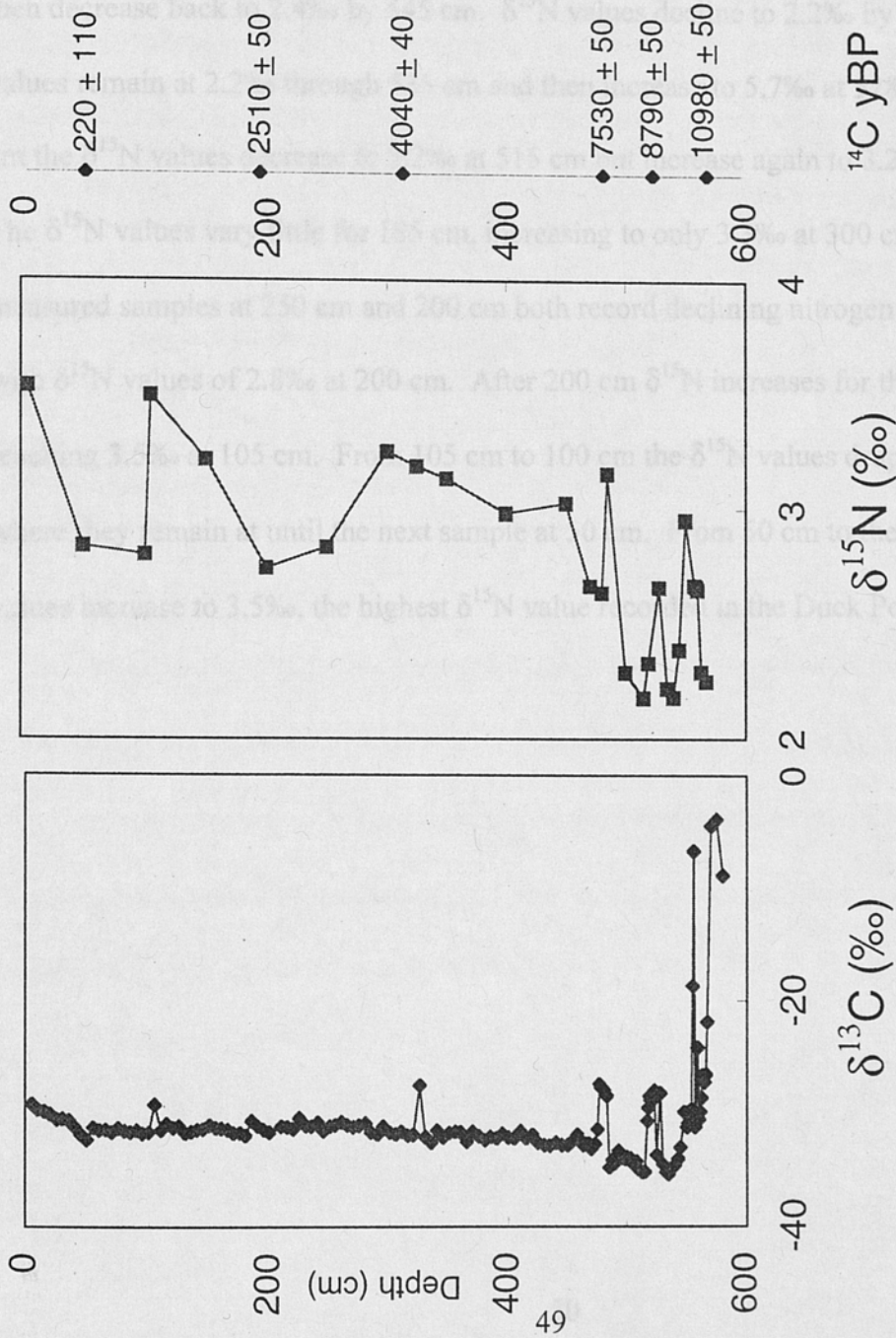


Figure 3.13.  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values for Duck Pond. Radiocarbon ages are at the right.

### 3.2.6 Nitrogen isotope values

The two oldest samples from Duck Pond analyzed for nitrogen isotopes contained the lowest  $\delta^{15}\text{N}$  values in the core (Fig. 3.13). The  $\delta^{15}\text{N}$  values at 567 cm and 563 cm are both 2.3‰. After 563 cm, the nitrogen isotopic values increase to 3.0‰ at 550 cm, but then decrease back to 2.4‰ by 545 cm.  $\delta^{15}\text{N}$  values decline to 2.2‰ by 540 cm. The values remain at 2.2‰ through 535 cm and then increase to 5.7‰ at 528 cm. From 528 cm the  $\delta^{15}\text{N}$  values decrease to 2.2‰ at 515 cm but increase again to 3.2‰ at 485 cm. The  $\delta^{15}\text{N}$  values vary little for 185 cm, increasing to only 3.3‰ at 300 cm. The next two measured samples at 250 cm and 200 cm both record declining nitrogen isotopic values, with  $\delta^{15}\text{N}$  values of 2.8‰ at 200 cm. After 200 cm  $\delta^{15}\text{N}$  increases for the next 95 cm, reaching 3.5‰ at 105 cm. From 105 cm to 100 cm the  $\delta^{15}\text{N}$  values drop sharply to 2.8‰, where they remain at until the next sample at 50 cm. From 50 cm to the surface the  $\delta^{15}\text{N}$  values increase to 3.5‰, the highest  $\delta^{15}\text{N}$  value recorded in the Duck Pond core.

## CHAPTER 4. PREVIOUS REGIONAL RECORDS

Table 4.1. Physiographic data for Sterling Pond and Ritterbush Pond.

The previous studies most comparable to this one were done by Lin (1996), Lini (1997) and Bierman et al. (1997). Parts of these studies were concerned with the organic matter from Sterling Pond and Ritterbush Pond, Vermont. Elemental and isotopic analyses of the sedimentary organic matter from these ponds were completed. The geographic proximity of Sterling Pond and Ritterbush Pond to my study lakes and the similar analyses completed by these studies make these previous records especially relevant. As such, their data will be described below.

Table 4.2. The age data from the radiocarbon dating completed on Sterling Pond and Ritterbush Pond (Bierman et al., 1997).

### 4.1 Sterling Pond

#### 4.1.1 Physiographic data

Sterling Pond is located at 917 m above sea level (Table 4.1) in northern Vermont, along the spine of the Green Mountains (Figs. 3.1 and 4.1). The pond is 9 m deep, and the core was collected from the northeastern basin of the lake (Fig. 4.2). Sterling Pond has a surface area of 0.03 km<sup>2</sup>, and its drainage basin covers 0.3 km<sup>2</sup>. The relief of the drainage basin is only 40 m.

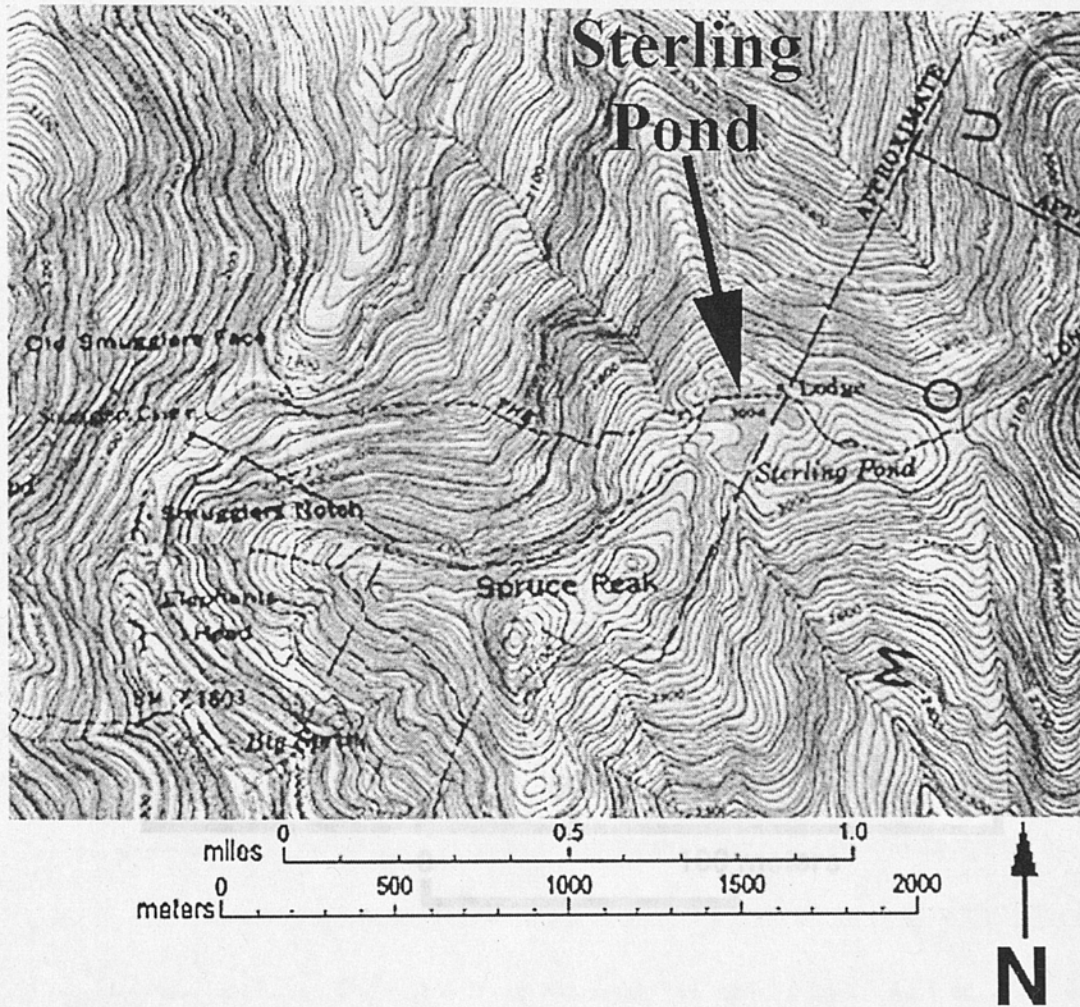
**Table 4.1** Physiographic data for Sterling Pond and Ritterbush Pond.

Lake Name	Surface area (km <sup>2</sup> )	Maximum depth (m)	Elevation asl (m)	Drainage basin area (km <sup>2</sup> )	Drainage basin relief (m)
<b>Sterling Pond</b>	0.03	9	917	0.3	40
<b>Ritterbush Pond</b>	0.05	14	317	2.2	293

**Table 4.2.** The age data from the radiocarbon dating completed on Sterling Pond and Ritterbush Pond (Bierman et al., 1997).

Lake Name	Core depth (cm)	<sup>14</sup> C yBP	+/-	Calibrated yBP	+/-	Lab #	Sample type
<b>Sterling Pond</b>	260	4180	50	4710	120	17893	Gyttja
	420	8600	60	9560	30	18293	Gyttja
	490	11180	60	13100	80	17894	Gyttja
	521	12760	70	15030	560	17895	Gyttja
<b>Ritterbush Pond</b>	96	2570	60	2650	100	33132	Gyttja
	154	2940	70	3100	140	33352	Gyttja
	220	3960	60	4400	110	22993	Gyttja
	348	6540	70	7490	70	20195	Gyttja
	426	9140	60	10310	90	32854	Gyttja
	445	9430	60	10650	90	20196	Gyttja
	464	10380	60	12290	320	33350	Gyttja
	479	11940	90	13950	130	20902	Gyttja

Figure 4.1. A topographic map showing the location and elevation of Sterling Pond. From the Mt. Mansfield (VT) 1:24,000 USGS quadrangle.



**Figure 4.1.** A topographic map showing the location and elevation of Sterling Pond. From the Mt. Mansfield (VT), 1:24,000 USGS quadrangle.

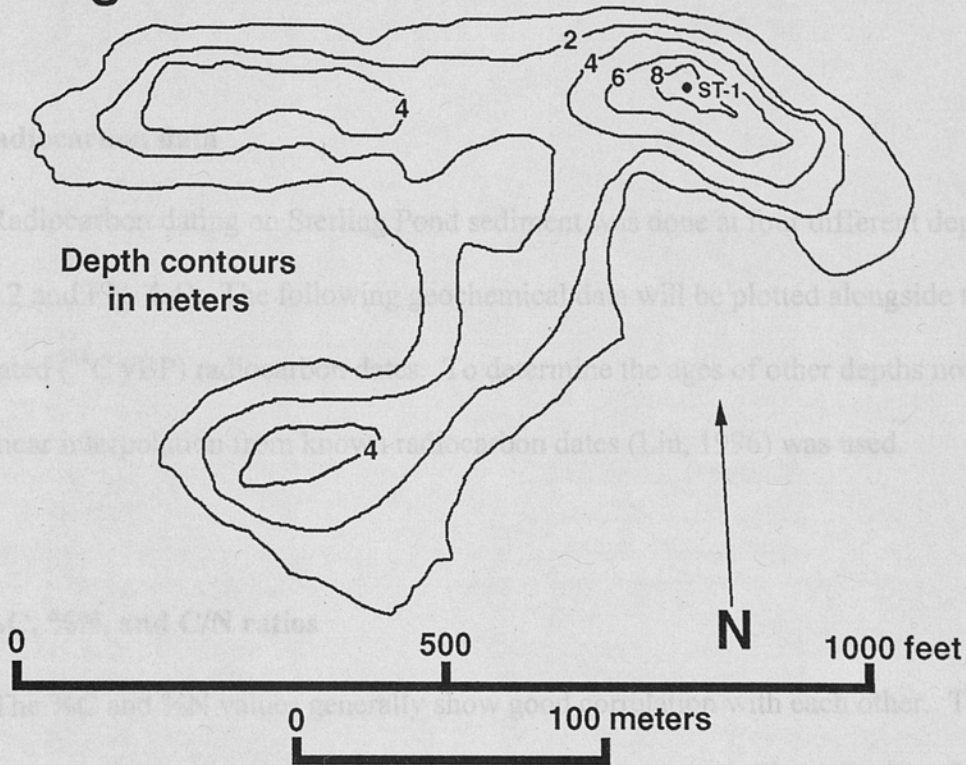


#### 4.1.2 Lithology

See Appendix A3 for a detailed description of the sediment from Sterling Pond.

The lithology of the Sterling Pond core is rather monotonous (Fig. 4.3) (Lin, 1996). The bottom 8 cm of sediment is coarse glacial silt, followed by 35 cm of clay interspersed with layers of pebbles. Layers of clay, silt, and sand then alternate for 23 cm, after which

### Sterling Pond



#### 4.1.3 Radiocarbon

Radiocarbon dating on Sterling Pond sediment cores were at different depths (Table 4.2 and Fig. 4.2). The following chronometric data will be plotted alongside the uncalibrated ( $^{14}\text{C}$  vs P) radiocarbon data to provide the ages of other depths not dated. Linear interpolation between the radiocarbon dates (Lin, 1996) was used.

#### 4.1.4 $\%C$ , $\%N$ , and C/N ratios

The  $\%C$  and  $\%N$  values generally show good correlation with each other. The amount of carbon and nitrogen in the oldest samples is very small (Fig. 4.5). The first three samples (510 cm, 300 cm, 502 cm) all have  $\%C$  values less than 1 and  $\%N$  values at or below 0.1. From 502 to 430 cm, the  $\%C$  and  $\%N$  values steadily increase, reaching 36.6 for  $\%C$  and 2.2 for  $\%N$ . The C/N values also increase during this interval, starting at 9.9 at 502 cm and reaching 17 at 430 cm. After 430 cm, the  $\%N$  values vary little

**Figure 4.2.** Bathymetry of Sterling Pond showing the location of the core (ST-1) used for analysis (Lin, 1996).

#### **4.1.2 Lithology**

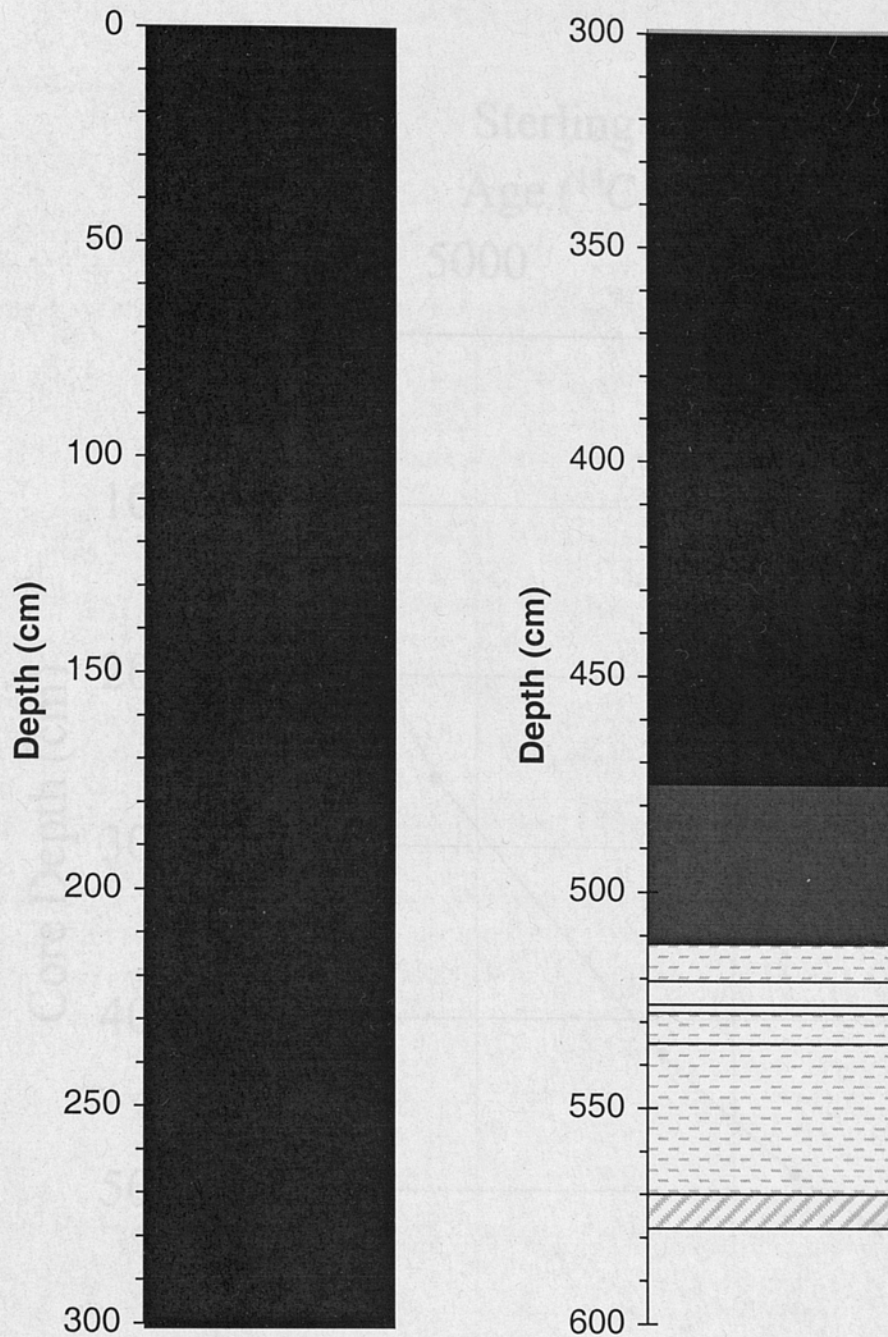
See Appendix A3 for a detailed description of the sediment from Sterling Pond. The lithology of the Sterling Pond core is rather monotonous (Fig. 4.3) (Lin, 1996). The bottom 8 cm of sediment is coarse glacial till, followed by 35 cm of clay interspersed with layers of pebbles. Layers of clay, silt, and sand then alternate for 23 cm, after which the sediment becomes gyttja, which continues to the top of the core.

#### **4.1.3 Radiocarbon data**

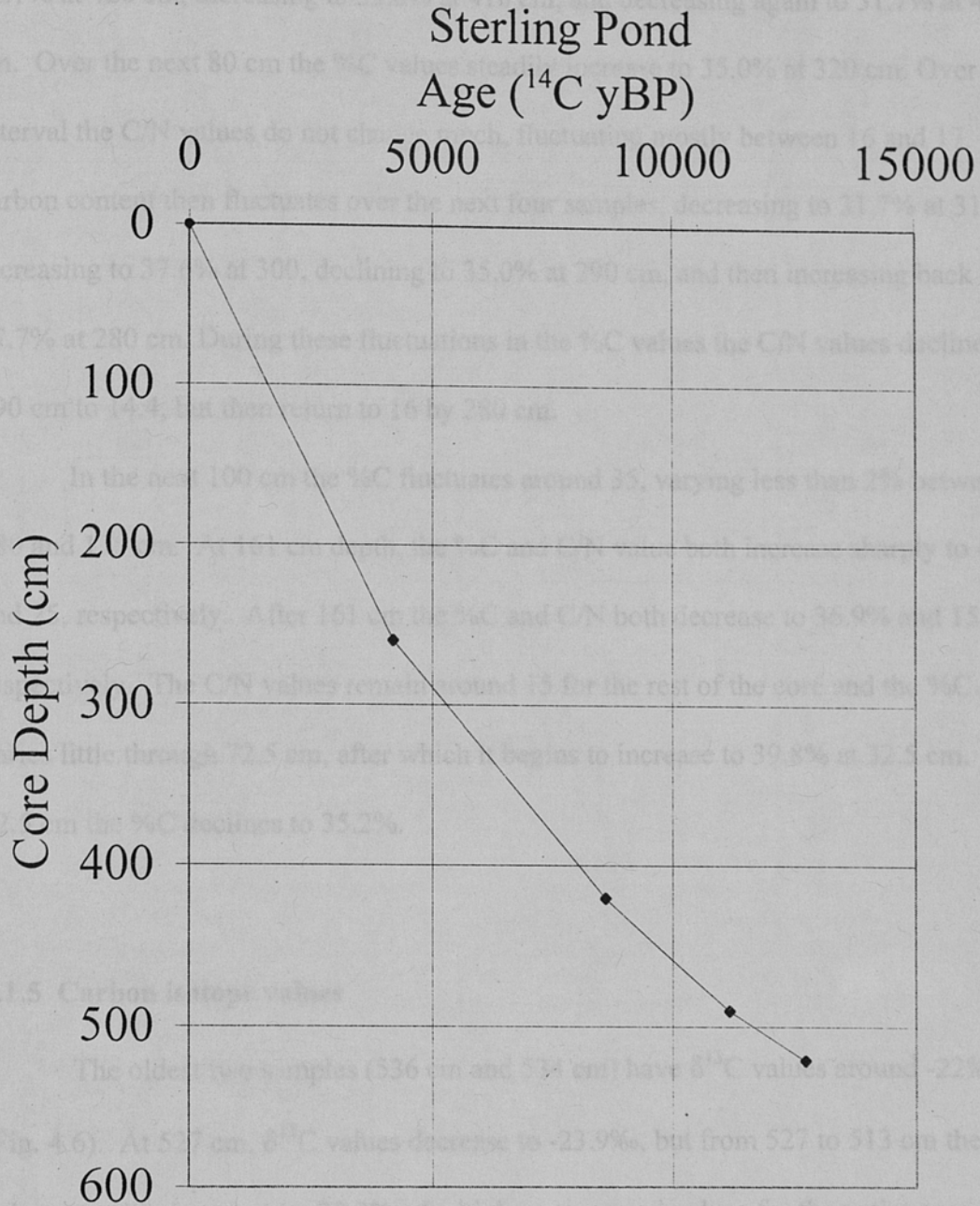
Radiocarbon dating on Sterling Pond sediment was done at four different depths (Table 4.2 and Fig. 4.4). The following geochemical data will be plotted alongside the uncalibrated ( $^{14}\text{C}$  yBP) radiocarbon dates. To determine the ages of other depths not dated, linear interpolation from known radiocarbon dates (Lin, 1996) was used.

#### **4.1.4 %C, %N, and C/N ratios**

The %C and %N values generally show good correlation with each other. The amount of carbon and nitrogen in the oldest samples is very small (Fig. 4.5). The first three samples (510 cm, 506 cm, 502 cm) all have %C values less than 1 and %N values at or below 0.1. From 502 to 430 cm, the %C and %N values steadily increase, reaching 36.6 for %C and 2.2 for %N. The C/N values also increase during this interval, starting at 9.9 at 502 cm and reaching 17 at 430 cm. After 430 cm, the %N values vary little throughout the rest of the core, so only the %C and C/N values will be discussed further.



**Figure 4.3.** The lithology of the sediment from Sterling Pond (Lin, 1996). Black represents gyttja, gray dashes clay, gray diagonals till, and white sand.



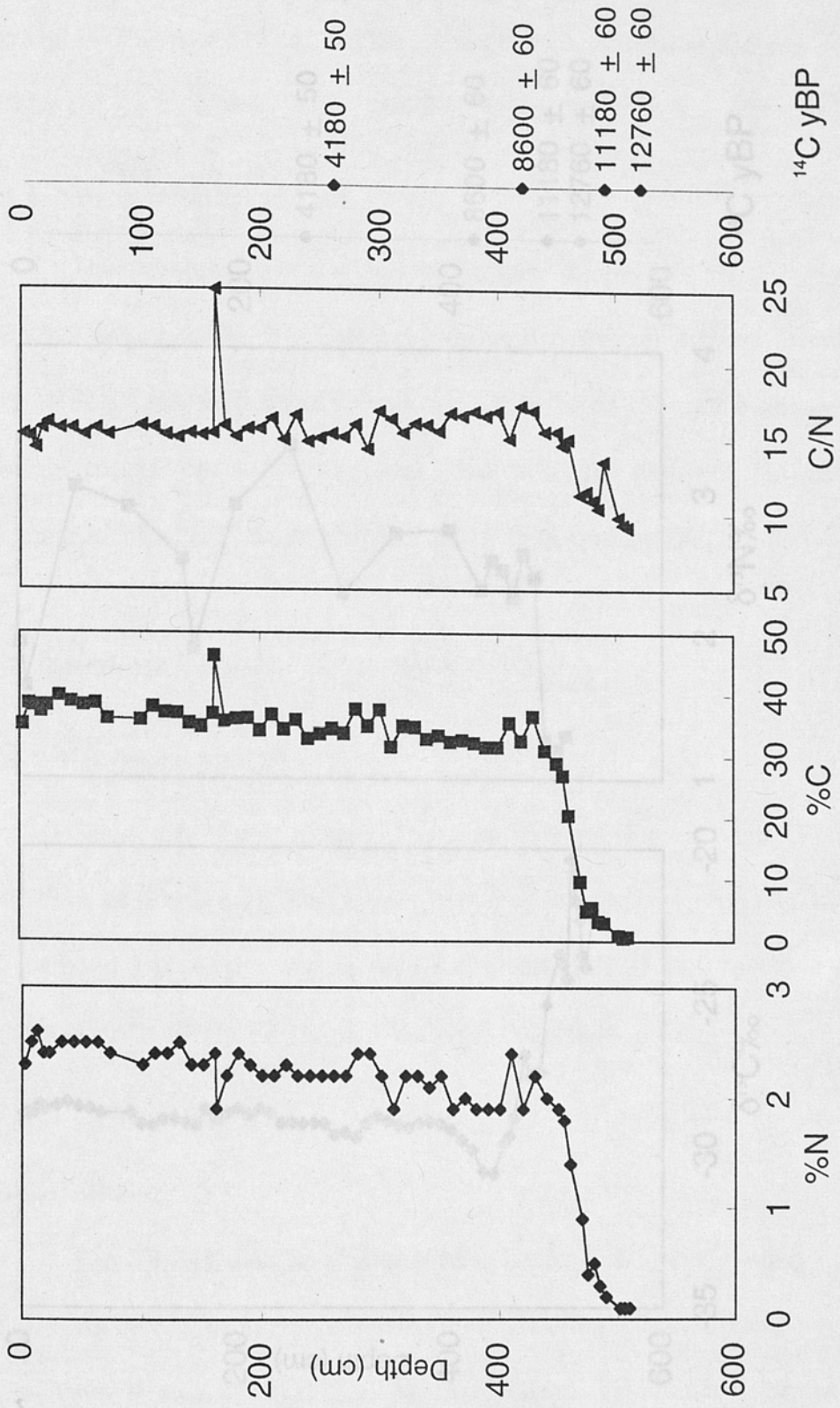
**Figure 4.4.** Depth vs. age for Sterling Pond, with the ages in uncalibrated  $^{14}\text{C}$  yBP (Lin, 1996).

After 430 cm the %C values fluctuate over the next three samples, decreasing to 32.7% at 420 cm, increasing to 35.6% at 410 cm, and decreasing again to 31.7% at 400 cm. Over the next 80 cm the %C values steadily increase to 35.0% at 320 cm. Over this interval the C/N values do not change much, fluctuating mostly between 16 and 17. The carbon content then fluctuates over the next four samples, decreasing to 31.7% at 310 cm, increasing to 37.6% at 300, declining to 35.0% at 290 cm, and then increasing back to 37.7% at 280 cm. During these fluctuations in the %C values the C/N values decline at 290 cm to 14.4, but then return to 16 by 280 cm.

In the next 100 cm the %C fluctuates around 35, varying less than 2% between 280 and 170 cm. At 161 cm depth, the %C and C/N value both increase sharply to 46.4% and 25, respectively. After 161 cm the %C and C/N both decrease to 36.9% and 15.4, respectively. The C/N values remain around 15 for the rest of the core and the %C also varies little through 72.5 cm, after which it begins to increase to 39.8% at 32.5 cm. After 32.5 cm the %C declines to 35.2%.

#### 4.1.5 Carbon isotope values

The oldest two samples (536 cm and 534 cm) have  $\delta^{13}\text{C}$  values around -22‰ (Fig. 4.6). At 527 cm,  $\delta^{13}\text{C}$  values decrease to -23.9‰, but from 527 to 513 cm the  $\delta^{13}\text{C}$  values steadily increase to -20.3‰, the highest measured values for the entire core. The trend reverses after 513 cm, for the carbon isotopic values then decrease from -20.3‰ at 513 cm to -30.6‰ at 440 cm. This is the lowest value obtained throughout the core. From 440 cm  $\delta^{13}\text{C}$  values increase until 380 cm, where they reach -28.9‰. Between 380



**Figure 4.5.** %N, %C, and C/N values for the core from Sterling Pond. Depth in centimeters is along the vertical axes, radiocarbon ages are at the right.

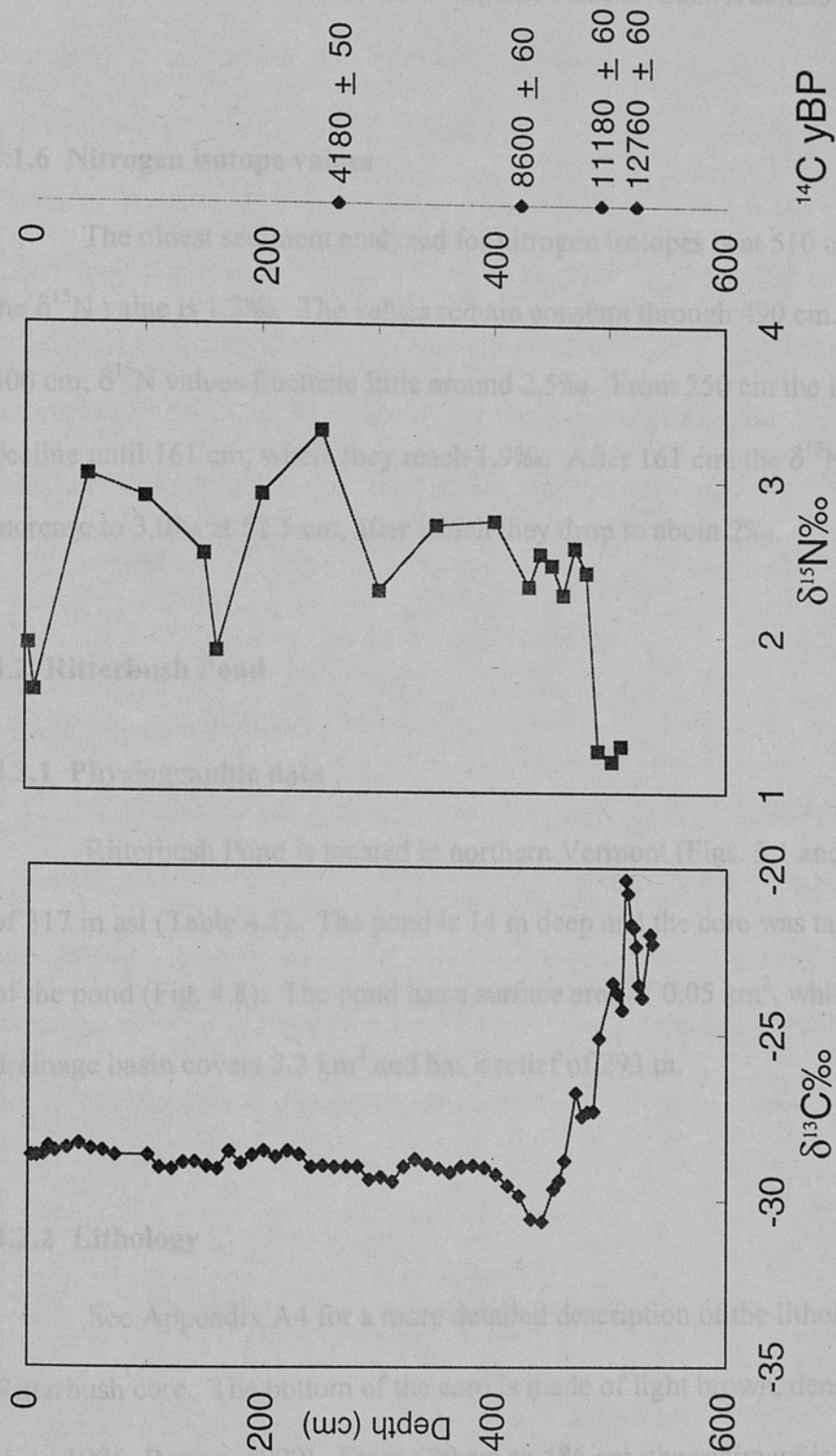


Figure 4.6.  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values for Sterling Pond. Radiocarbon ages are at the right.

and the surface, there is little variation in the  $\delta^{13}\text{C}$  values. The lowest value in this interval is -29.4‰ at 310 cm, and the highest value is -28.3‰ at 42.5 cm.

#### **4.1.6 Nitrogen isotope values**

The oldest sediment analyzed for nitrogen isotopes is at 510 cm (Fig. 4.6), where the  $\delta^{15}\text{N}$  value is 1.3‰. The values remain constant through 490 cm. Between 490 and 300 cm,  $\delta^{15}\text{N}$  values fluctuate little around 2.5‰. From 250 cm the isotope values decline until 161 cm, where they reach 1.9‰. After 161 cm, the  $\delta^{15}\text{N}$  values steadily increase to 3.0‰ at 52.5 cm, after which they drop to about 2‰.

## **4.2 Ritterbush Pond**

### **4.2.1 Physiographic data**

Ritterbush Pond is located in northern Vermont (Figs. 3.1 and 4.7) at an elevation of 317 m asl (Table 4.1). The pond is 14 m deep and the core was taken from the center of the pond (Fig. 4.8). The pond has a surface area of 0.05 km<sup>2</sup>, while the surrounding drainage basin covers 2.2 km<sup>2</sup> and has a relief of 293 m.

### **4.2.2 Lithology**

See Appendix A4 for a more detailed description of the lithology of the Ritterbush core. The bottom of the core is made of light brown, dense clay (Fig. 4.9) (Lin, 1996; Brown, 1999). From 520 cm to 486 cm, the sediment is brown gray to black



banded silt containing several thin (2 to 3mm) black laminae. At 486 cm, the sediment coarsens to sand, which continues for 12 cm. The alternating black and brown layers return at 474 cm, and continue until 424 cm, where a fining-upward sequence of gray sediment begins. This sequence ends at 401.5 cm depth, where the sediment becomes black with several thin (<1 cm) gray layers. The next change is at 349 cm, where another fining upward sequence begins. The top of this layer is at 340.7 cm, after which the sediment is homogenous black gyttja until 153.5 cm. Here, gyttja deposition is interrupted by an 11 cm thick sandy clay layer. Except for three fining-upward layers (142.5 cm to 137.6 cm, 112.6 cm to 110.2, and 100 cm to 96.5 cm) the homogenous gyttja continues to the top of the core.

### 4.2.3 Radiocarbon data

Radiocarbon dating on Ritterbush Pond was done at eight different depths (Table 4.2 and Fig. 4.10). The following geochemical data will be plotted alongside the uncalibrated ( $^{14}\text{C}$  yBP) radiocarbon dates. To determine the ages of other depths not dated, either linear interpolation from known radiocarbon dates or estimations based on the regional deglaciation history were used.

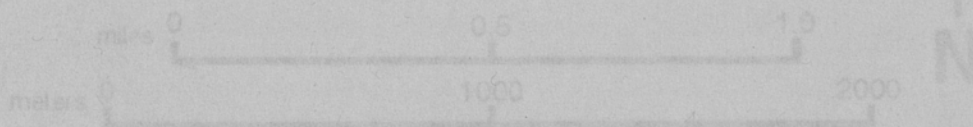
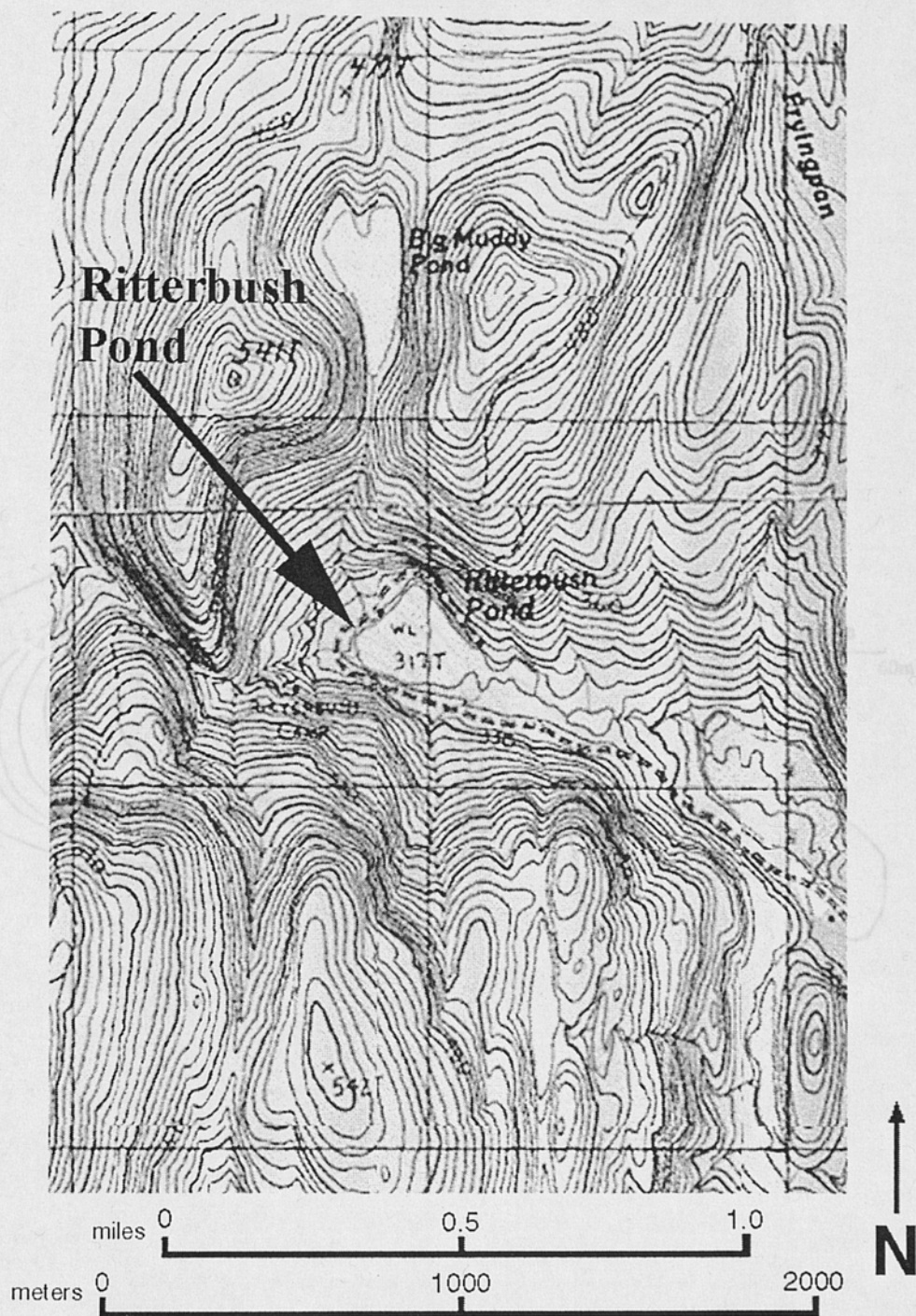


Figure 4.7. Topographic map showing the location and elevation of Ritterbush Pond. From the Eden (VT) 1:24,000 USGS quadrangle map.



**Figure 4.7.** Topographic map showing the location and elevation of Ritterbush Pond. From the Eden (VT) 1:24,000 USGS quadrangle map.

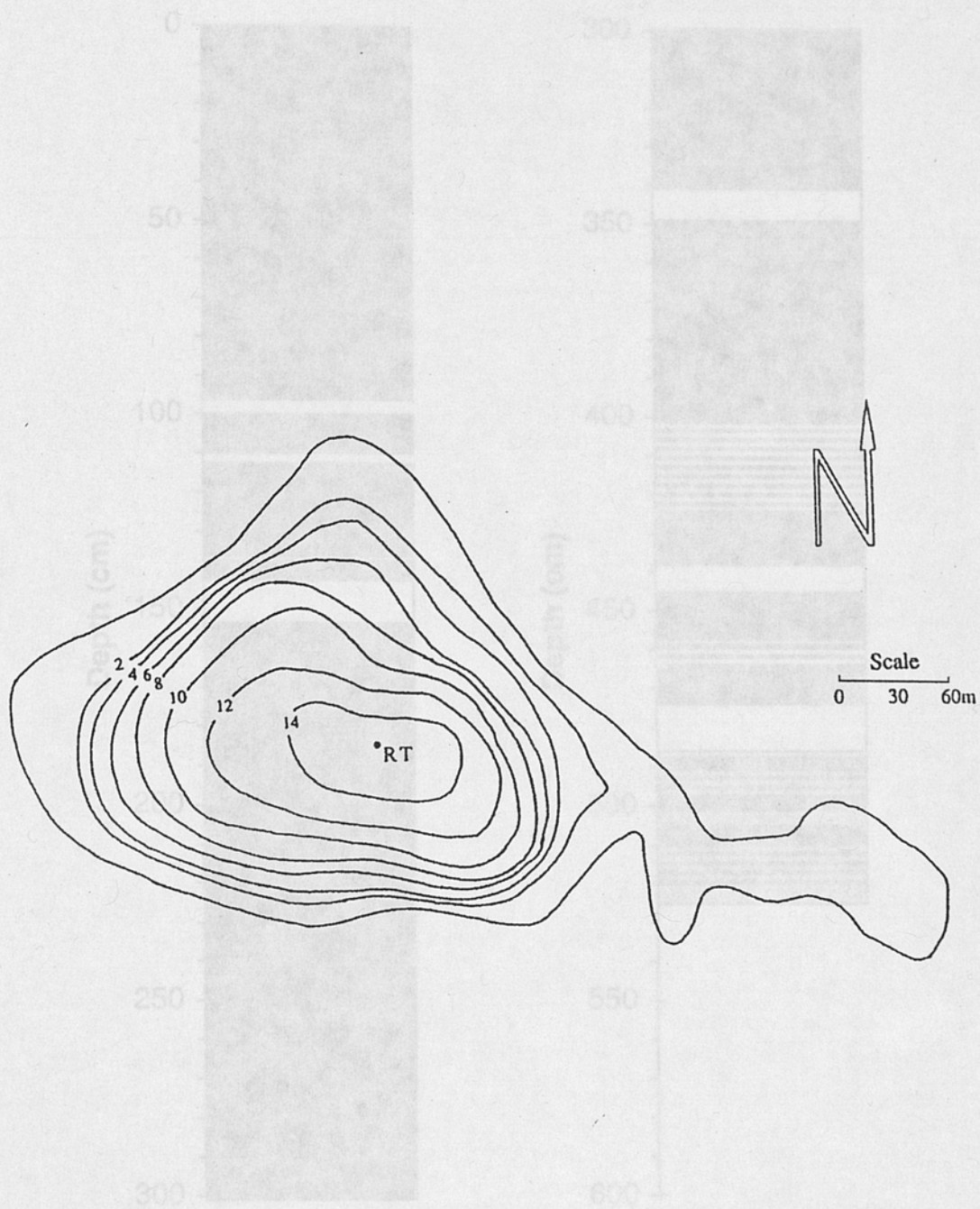
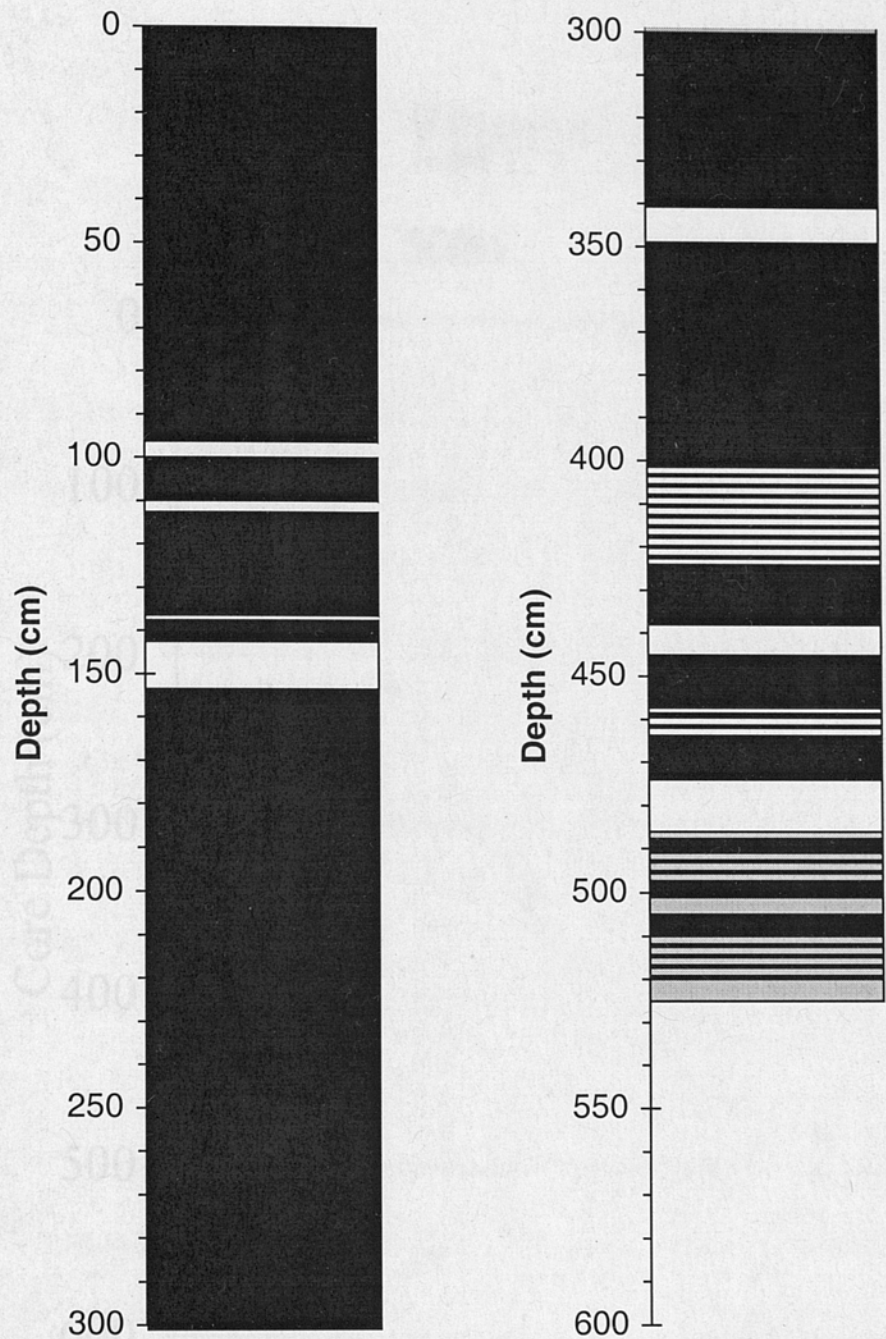
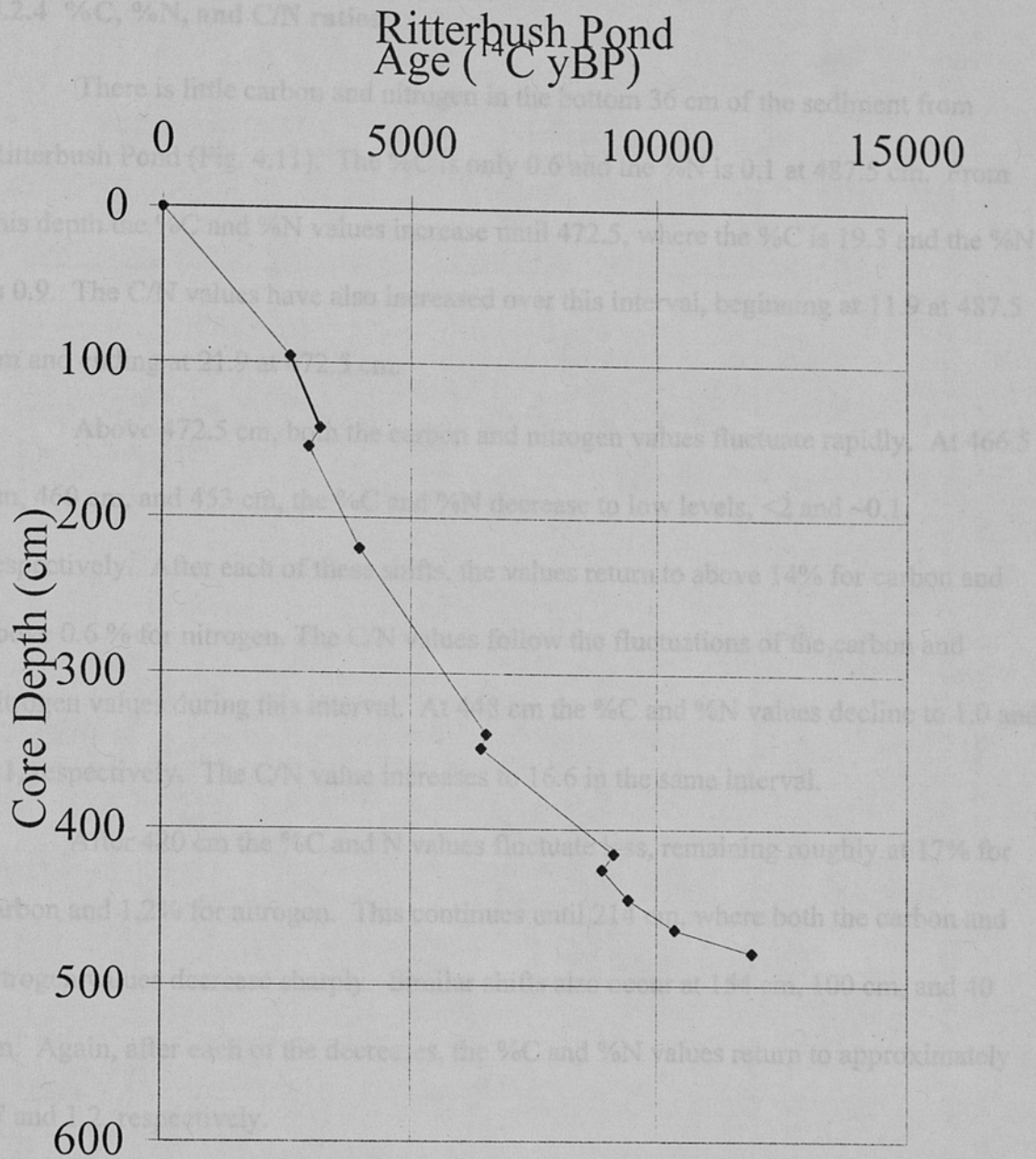


Figure 4.9. The lithology of the core from Ritterbush Pond (Lin, 1996; Brown, 1999). Black represents gyttja, gray is clay, and the white is sand. The striped layers symbolize

**Figure 4.8.** Bathymetry of Ritterbush Pond with the location of the core RT (Lin, 1996).



**Figure 4.9.** The lithology of the core from Ritterbush Pond (Lin, 1996 ; Brown, 1999). Black represents gyttja, gray is clay, and the white is sand. The striped layers symbolize alternating layers of sediments.



**Figure 4.10.** Depth vs. age for Ritterbush Pond, with the ages in uncalibrated  $^{14}\text{C}$  yBP (Lin, 1996).

#### 4.2.4 %C, %N, and C/N ratios

There is little carbon and nitrogen in the bottom 36 cm of the sediment from Ritterbush Pond (Fig. 4.11). The %C is only 0.6 and the %N is 0.1 at 487.5 cm. From this depth the %C and %N values increase until 472.5, where the %C is 19.3 and the %N is 0.9. The C/N values have also increased over this interval, beginning at 11.9 at 487.5 cm and ending at 21.9 at 472.5 cm.

Above 472.5 cm, both the carbon and nitrogen values fluctuate rapidly. At 466.5 cm, 460 cm, and 453 cm, the %C and %N decrease to low levels, <2 and ~0.1, respectively. After each of these shifts, the values return to above 14% for carbon and above 0.6 % for nitrogen. The C/N values follow the fluctuations of the carbon and nitrogen values during this interval. At 448 cm the %C and %N values decline to 1.0 and 0.1, respectively. The C/N value increases to 16.6 in the same interval.

After 420 cm the %C and N values fluctuate less, remaining roughly at 17% for carbon and 1.2% for nitrogen. This continues until 214 cm, where both the carbon and nitrogen values decrease sharply. Similar shifts also occur at 154 cm, 100 cm, and 40 cm. Again, after each of the decreases, the %C and %N values return to approximately 17 and 1.2, respectively.

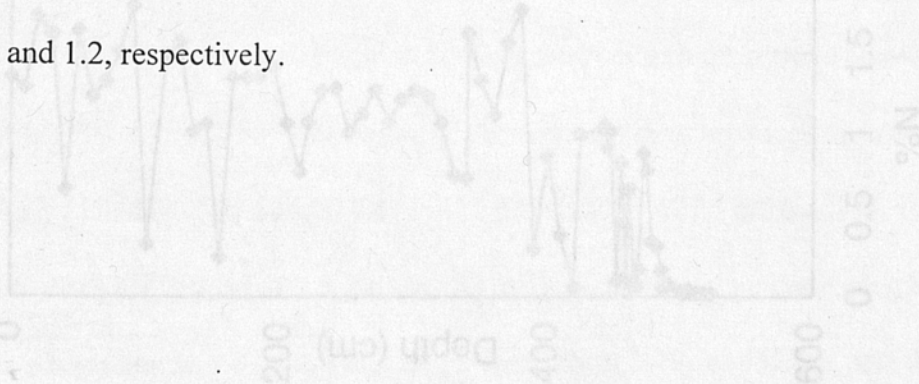


Figure 4.11. %N, %C, and C/N ratios versus depth in Ritterbush Pond (1991).

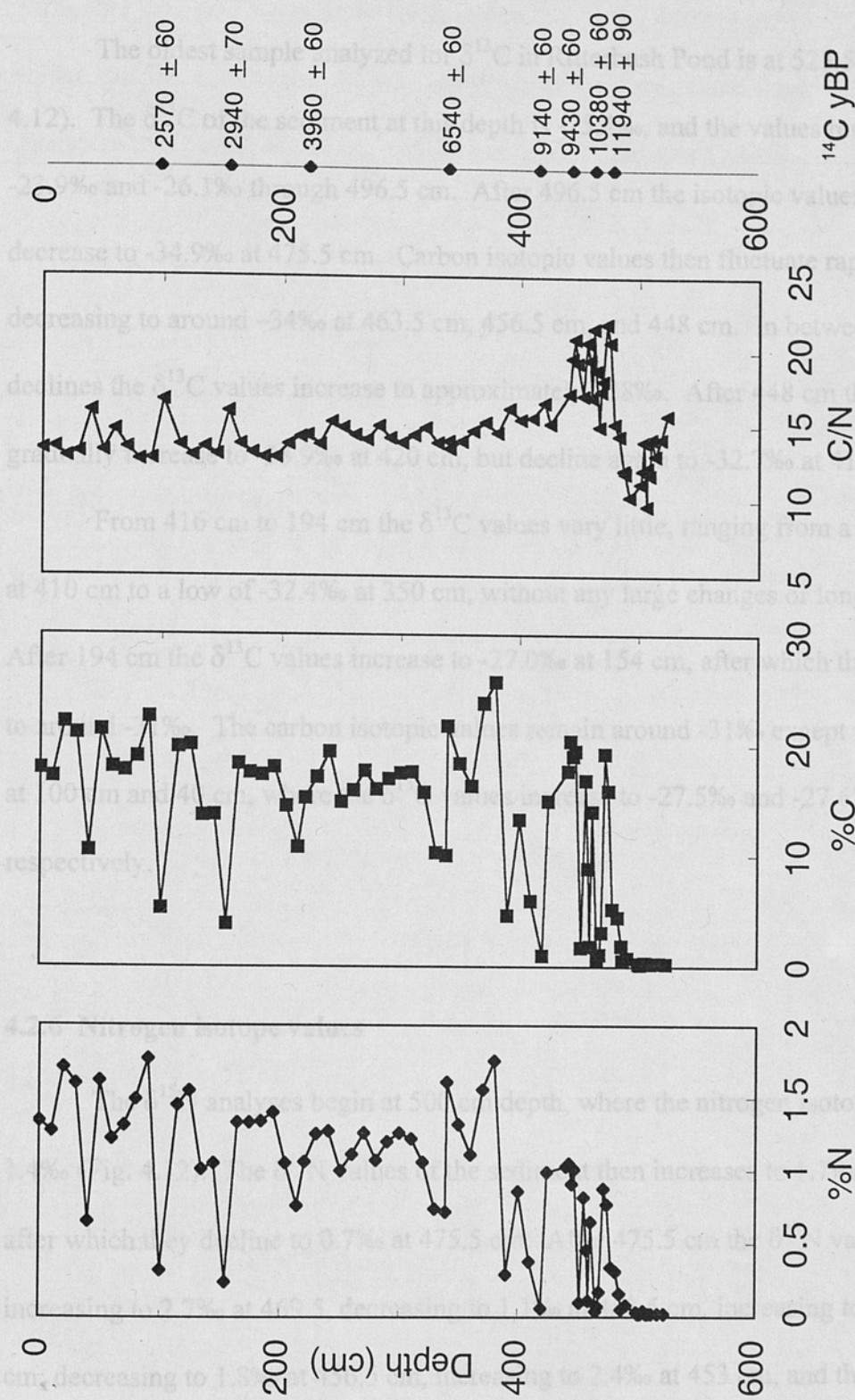


Figure 4.11. %N, %C, and C/N values for the core from Ritterbush Pond. Depth in centimeters is along the vertical axes, radiocarbon ages are at the right (Lini et al., 1999).

#### 4.2.5 Carbon isotope values

The oldest sample analyzed for  $\delta^{13}\text{C}$  in Ritterbush Pond is at 523.5 cm depth (Fig. 4.12). The  $\delta^{13}\text{C}$  of the sediment at this depth is -25.4‰, and the values remain between -23.9‰ and -26.1‰ through 496.5 cm. After 496.5 cm the isotopic values steadily decrease to -34.9‰ at 475.5 cm. Carbon isotopic values then fluctuate rapidly, decreasing to around -34‰ at 463.5 cm, 456.5 cm, and 448 cm. In between these declines the  $\delta^{13}\text{C}$  values increase to approximately -28‰. After 448 cm the  $\delta^{13}\text{C}$  values gradually increase to -25.9‰ at 420 cm, but decline again to -32.7‰ at 416 cm.

From 416 cm to 194 cm the  $\delta^{13}\text{C}$  values vary little, ranging from a high of -30.2‰ at 410 cm to a low of -32.4‰ at 350 cm, without any large changes or long-term trends. After 194 cm the  $\delta^{13}\text{C}$  values increase to -27.0‰ at 154 cm, after which they again decline to around -31‰. The carbon isotopic values remain around -31‰ except for two peaks at 100 cm and 40 cm, where the  $\delta^{13}\text{C}$  values increase to -27.5‰ and -27.6‰, respectively.

#### 4.2.6 Nitrogen isotope values

The  $\delta^{15}\text{N}$  analyses begin at 500 cm depth, where the nitrogen isotopic value is 1.4‰ (Fig. 4.12). The  $\delta^{15}\text{N}$  values of the sediment then increase to 1.7‰ at 487.5 cm, after which they decline to 0.7‰ at 475.5 cm. After 475.5 cm the  $\delta^{15}\text{N}$  values fluctuate, increasing to 2.7‰ at 469.5 cm, decreasing to 1.1‰ at 463.5 cm, increasing to 2.3‰ at 460 cm, decreasing to 1.8‰ at 456.5 cm, increasing to 2.4‰ at 453 cm, and then decreasing



to 1.8‰ at 448 cm. From 448 cm to 420 cm the  $\delta^{15}\text{N}$  values of the sediment increase to 3.0‰, but then decline again to 1.9‰ at 416 cm. After 416 cm the nitrogen isotopic values steadily increase all the way to the top of the core, where the  $\delta^{15}\text{N}$  value is 3.8‰. There are two peaks, at 154 cm and 40 cm, of more positive ( $\geq 4.0\text{‰}$ )  $\delta^{15}\text{N}$  values.

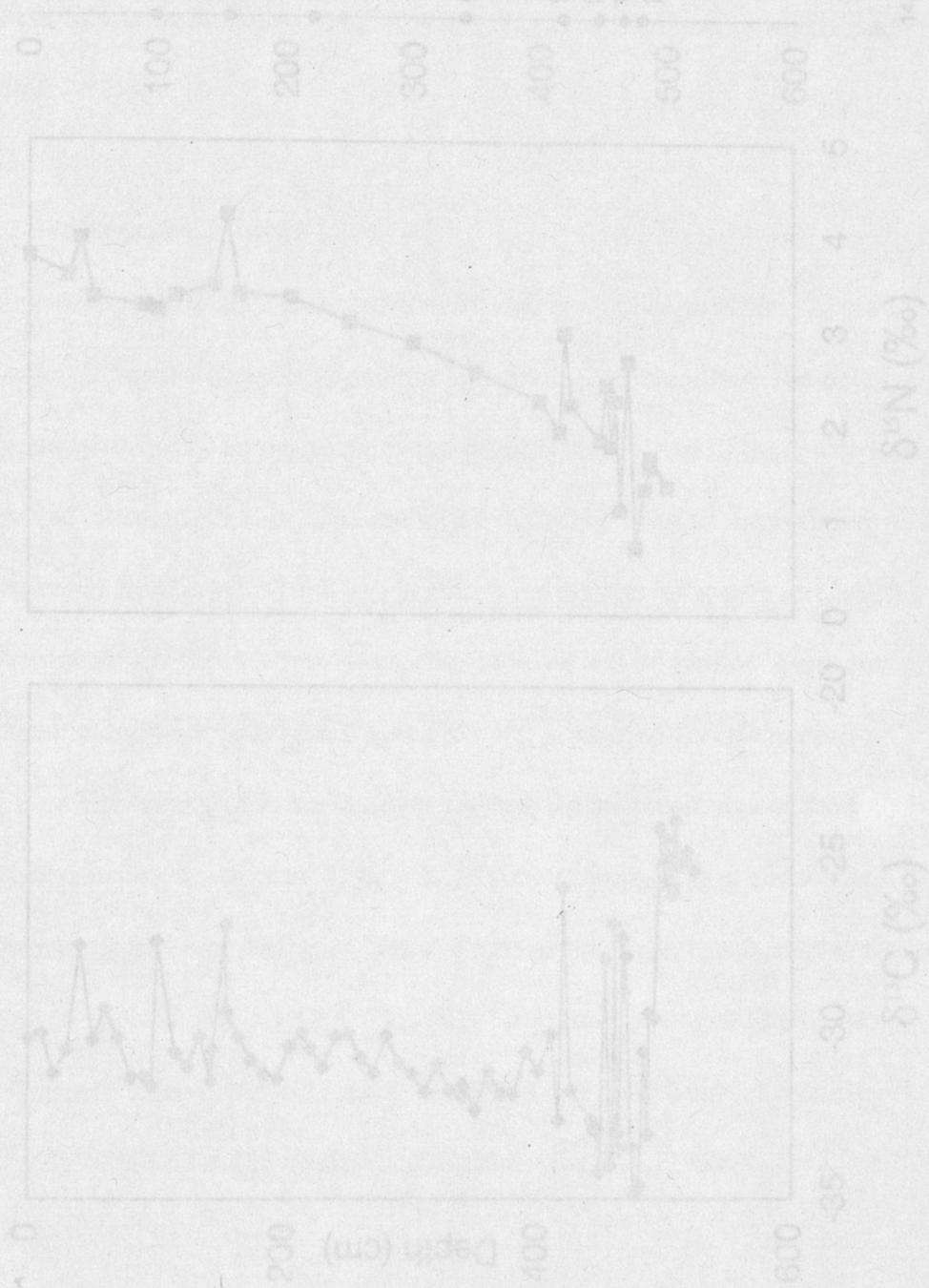


Figure 4.12.  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values for Ritterbusch Pond. Radio-carbon ages are at the right (Linn, et al., 1999)

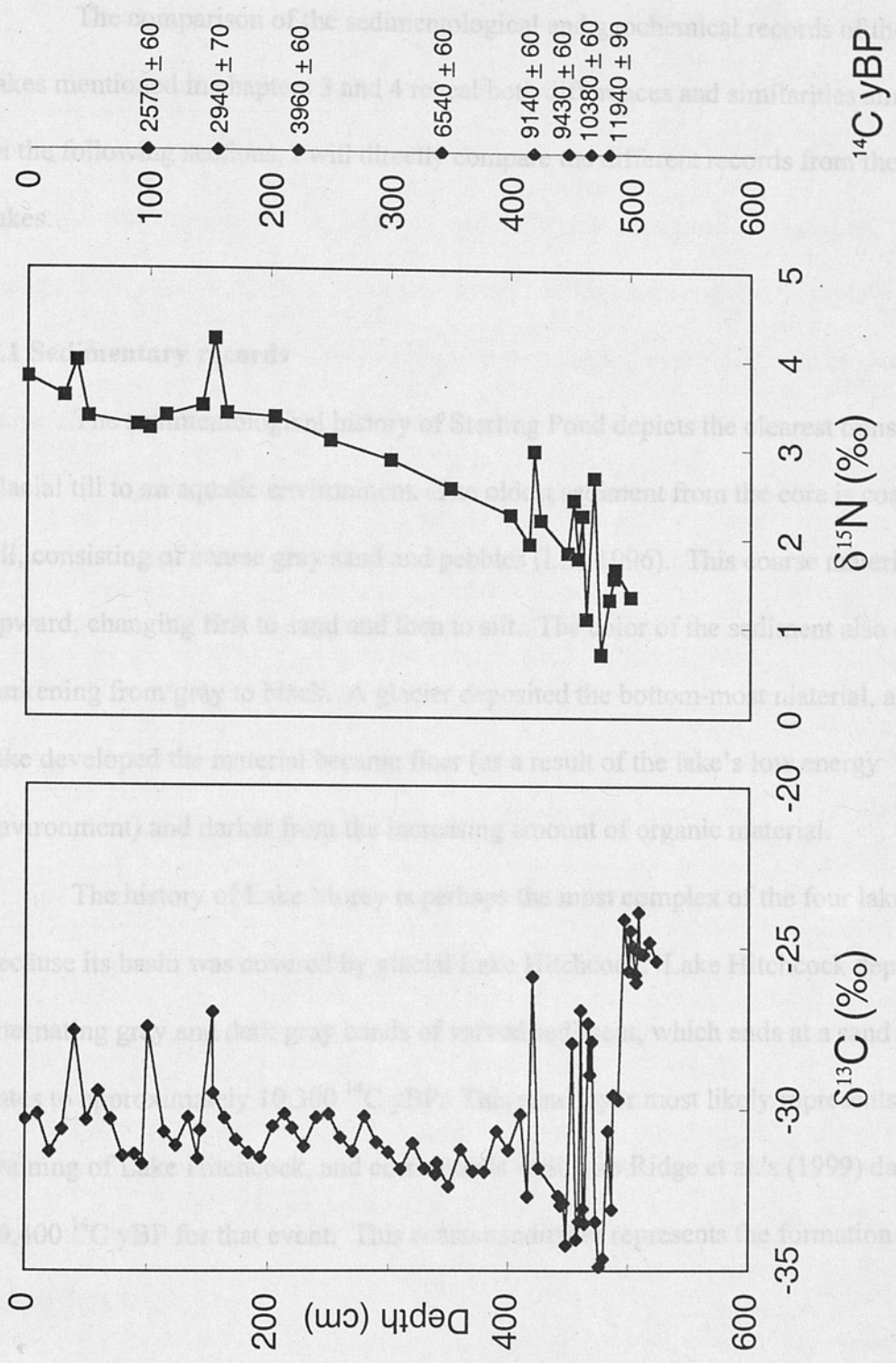


Figure 4.12.  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values for Ritterbush Pond. Radiocarbon ages are at the right (Lini, et al., 1999).

## CHAPTER 5. COMPARATIVE ANALYSIS OF DATA SETS

The comparison of the sedimentological and geochemical records of the four lakes mentioned in chapters 3 and 4 reveal both differences and similarities among them. In the following sections, I will directly compare the different records from the four lakes.

### 5.1 Sedimentary records

The sedimentological history of Sterling Pond depicts the clearest transition from glacial till to an aquatic environment. The oldest sediment from the core is coarse glacial till, consisting of coarse gray sand and pebbles (Lin, 1996). This coarse material fines upward, changing first to sand and then to silt. The color of the sediment also changes, darkening from gray to black. A glacier deposited the bottom-most material, and as the lake developed the material became finer (as a result of the lake's low energy environment) and darker from the increasing amount of organic material.

The history of Lake Morey is perhaps the most complex of the four lakes, mostly because its basin was covered by glacial Lake Hitchcock. Lake Hitchcock deposited alternating gray and dark gray bands of varved sediment, which ends at a sand layer that dates to approximately 10,300  $^{14}\text{C}$  yBP. This sand layer most likely represents the draining of Lake Hitchcock, and corresponds well with Ridge et al.'s (1999) date of 10,400  $^{14}\text{C}$  yBP for that event. This coarser sediment represents the formation of Lake

Morey as an individual body of water. Above this layer the sediment changes to gyttja for the remainder of the core.

The history of Duck Pond is much different from Lake Morey's. The bottom 14 cm of the sediment is coarse sand and pebbles; most likely glacial till. After a gradual change to finer silt, there is another coarse layer. This pattern of coarse sediment alternating with a finer layers of silt is repeated three more time. As a group, these coarse layers fine upwards; each of the coarse layers contains material that is slightly finer than the previous coarse layer. Because these layers contain material that is too coarse to have originated from within the lake, it must have come from the surrounding watershed.

The gradual fining of the coarse layers up core represents an increase of the local landscape's stability resulting from the growth of the local ecosystems. As the plant communities established, they stabilized the hillslopes with root systems, allowing fewer pulses of coarse, inorganic material to wash into the lake basin. The overall upward fining of the sediment represents the transition of the basin from a terrestrial environment to a lacustrine environment.

The sedimentological history of Ritterbush Pond is similar to Duck Pond's history. The oldest sediment is gray from a lack of organic material, and there are frequent layers of sand (Bierman et al., 1997). At first, the darker, organic-rich layers are interspersed within mostly inorganic sediment, but eventually the majority of the sediment is organic-rich silt. The layers of detrital material also become less frequent after the bottom section of the core, as the local plant communities grew and stabilized the basin's hillslopes.

## 5.2 Organic matter and C-isotope records

In each of the four lakes there are two consistent shifts that occur in the oldest sediments. These are a marked increase in organic matter content and a significant decrease in the carbon isotope values. For this discussion, %C values will serve as a proxy for organic matter content. The radiocarbon dates mentioned here are linearly interpolated from the radiocarbon dates obtained from discrete macrofossils.

### 5.2.1 Timing of organic matter increase

The increase in %C in Lake Morey begins at 459 cm depth. The %C increases steadily from 0.5% to 9.8% at 374 cm (Fig 5.1). Interpolating between the known radiocarbon dates (Table 3.2) the starting and ending dates of the increase in %C can be calculated. The %C increases from 0.5% to 9.8% over 4600  $^{14}\text{C}$  years, from 10,200  $^{14}\text{C}$  to 5600  $^{14}\text{C}$  yBP.

In Duck Pond, %C increases from 4.1% at 567 cm to 30.9% at 530 cm (Fig. 5.2). The radiocarbon age for 567 cm is  $10980 \pm 50$   $^{14}\text{C}$  yBP, and the  $^{14}\text{C}$  age for 530 cm is 8900  $^{14}\text{C}$  yBP. The radiocarbon date at 567 cm was measured, while the age for the sediment at 530 cm was interpolated between calculated ages at 567 cm and 522 cm ( $8790 \pm 50$   $^{14}\text{C}$  yBP). The increase in the %C of the sediment at the bottom of Duck Pond lasted about 2100  $^{14}\text{C}$  years.

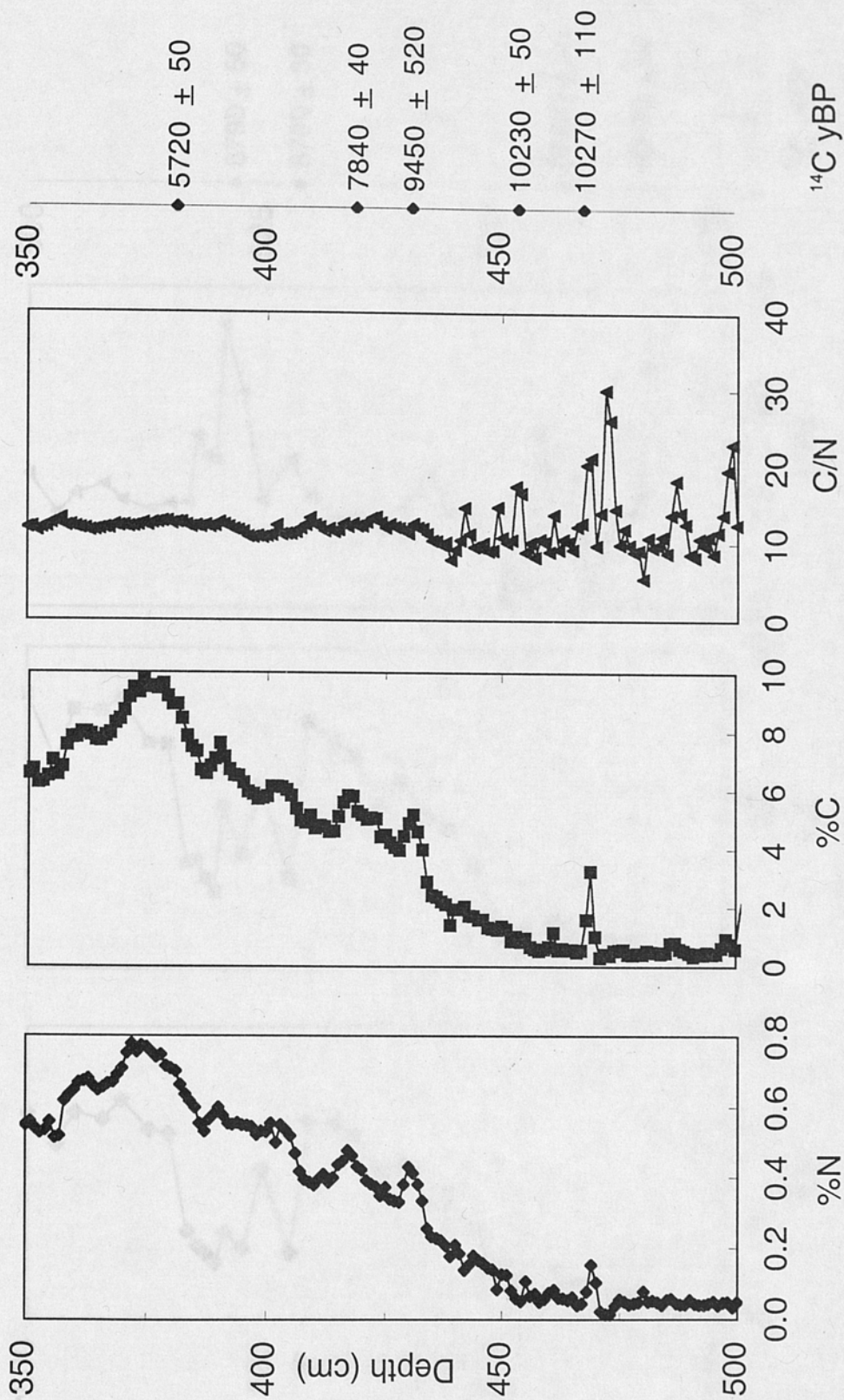
In Sterling Pond, the increase in %C begins at 513 cm (from the LOI record, Lin, 1996) and continues until 430 cm (Fig. 5.3). The oldest sample analyzed for %C is 510

cm (0.7%), and the %C increases to 36.6% at 430 cm. Using radiocarbon dates at 521 cm ( $12,760 \pm 70$   $^{14}\text{C}$  yBP), 490 cm ( $11,180 \pm 60$   $^{14}\text{C}$  yBP), and 420 cm ( $8600 \pm 60$   $^{14}\text{C}$  yBP) from Lin (1996), the starting and ending dates of the increase in %C are approximately 12,400  $^{14}\text{C}$  yBP and 9000  $^{14}\text{C}$  yBP, an interval of approximately 3400  $^{14}\text{C}$  years.

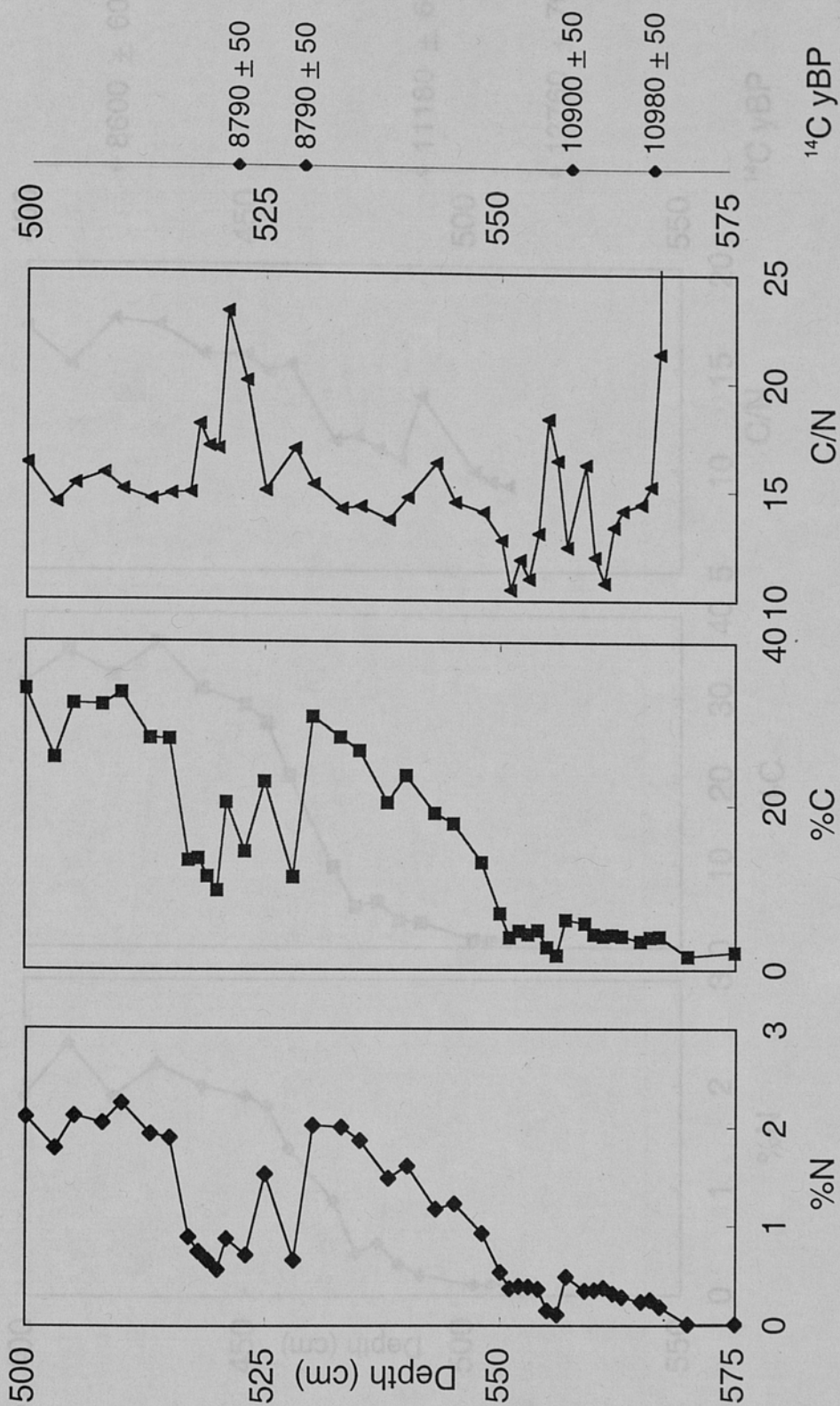
In the Ritterbush Pond sediment, the %C increases from 0.6% at 487.5 to 19.3% at 472.5 cm (Fig. 5.4). From calculated radiocarbon dates at 479 cm ( $11940 \pm 60$   $^{14}\text{C}$  yBP) and 464 cm ( $10380 \pm 60$   $^{14}\text{C}$  yBP) (Bierman et al., 1997), the interpolated radiocarbon date of 472.5 cm is 11,300  $^{14}\text{C}$  yBP. Linear extrapolation past the known date of 11,940  $^{14}\text{C}$  yBP at 479 cm is not accurate, as the sedimentation rate after deglaciation is often high (e.g. Duck Pond). Thus, dating the depth of 487.5 cm is problematic. A minimum age of 12,000  $^{14}\text{C}$  yBP can be assumed for 487.5 cm, given the radiocarbon age of 479 cm. This results in a duration of at least 700  $^{14}\text{C}$  years for the %C increase in Ritterbush Pond.

### 5.2.2 Timing of C-isotope shifts

The negative shift in  $\delta^{13}\text{C}$  values in Lake Morey occurs over a much shorter interval of core than the increase in %C values (Fig. 5.5). The decrease in  $\delta^{13}\text{C}$  values begins at 459 cm depth, where the  $\delta^{13}\text{C}$  value of the sediment is -16.3‰. From 459 cm the  $\delta^{13}\text{C}$  values decrease 13.2‰ to -29.5‰ at 431 cm. The shift begins at 10,200  $^{14}\text{C}$  yBP and ends at 9500  $^{14}\text{C}$  yBP, a total of 700  $^{14}\text{C}$  years. These dates were interpolated using the radiocarbon dates at depths 468 cm, 454 cm, and 431 cm.



**Figure 5.1.** The bottom portion of the Lake Morey core, showing the %C, %N, and C/N values. Notice the increase in %C and %N beginning at 459 cm.



**Figure 5.2.** The bottom portion of the Duck Pond core, showing the %C, %N, and C/N values. Notice the increase in %C and %N beginning at 567 cm.



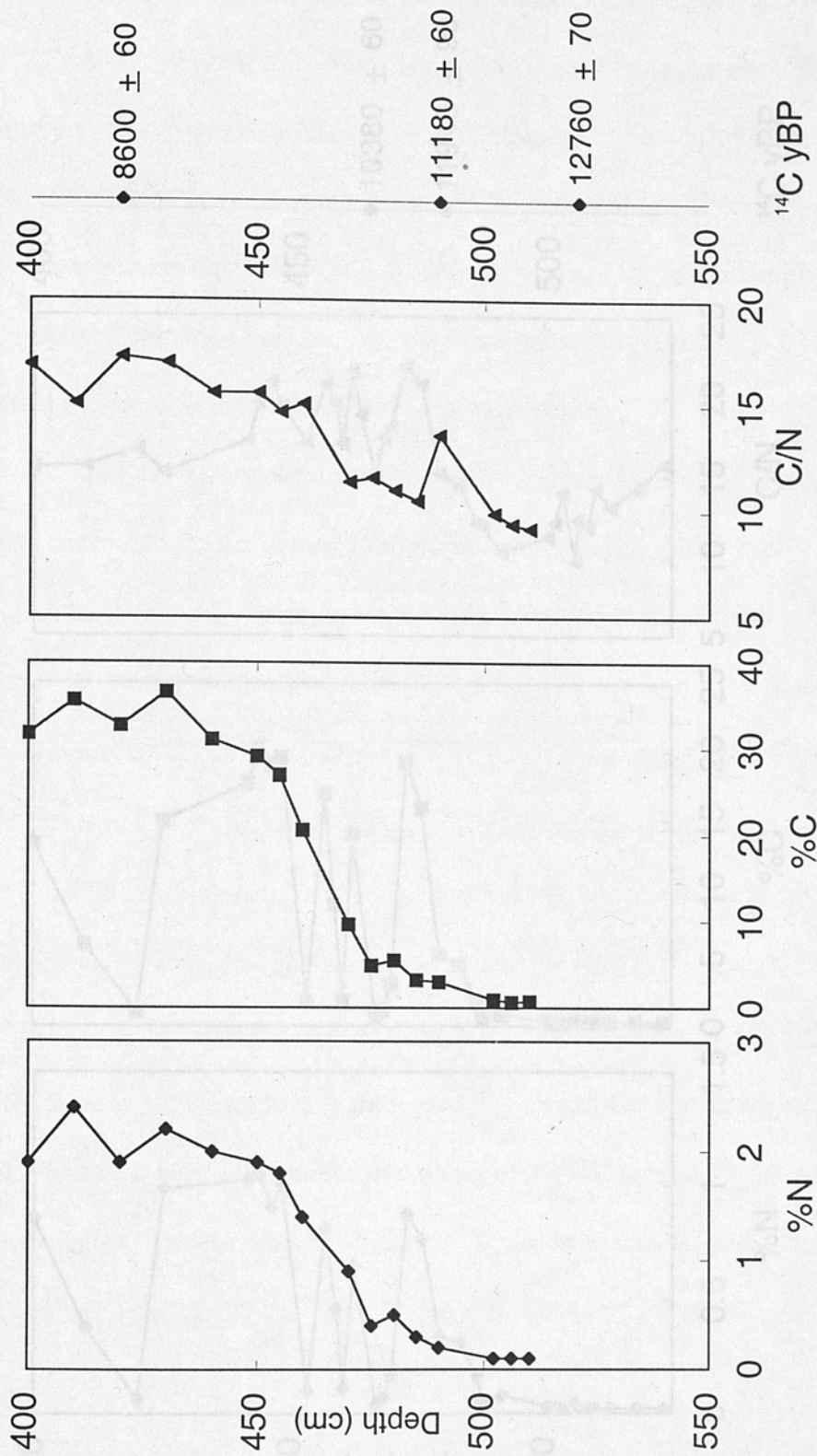
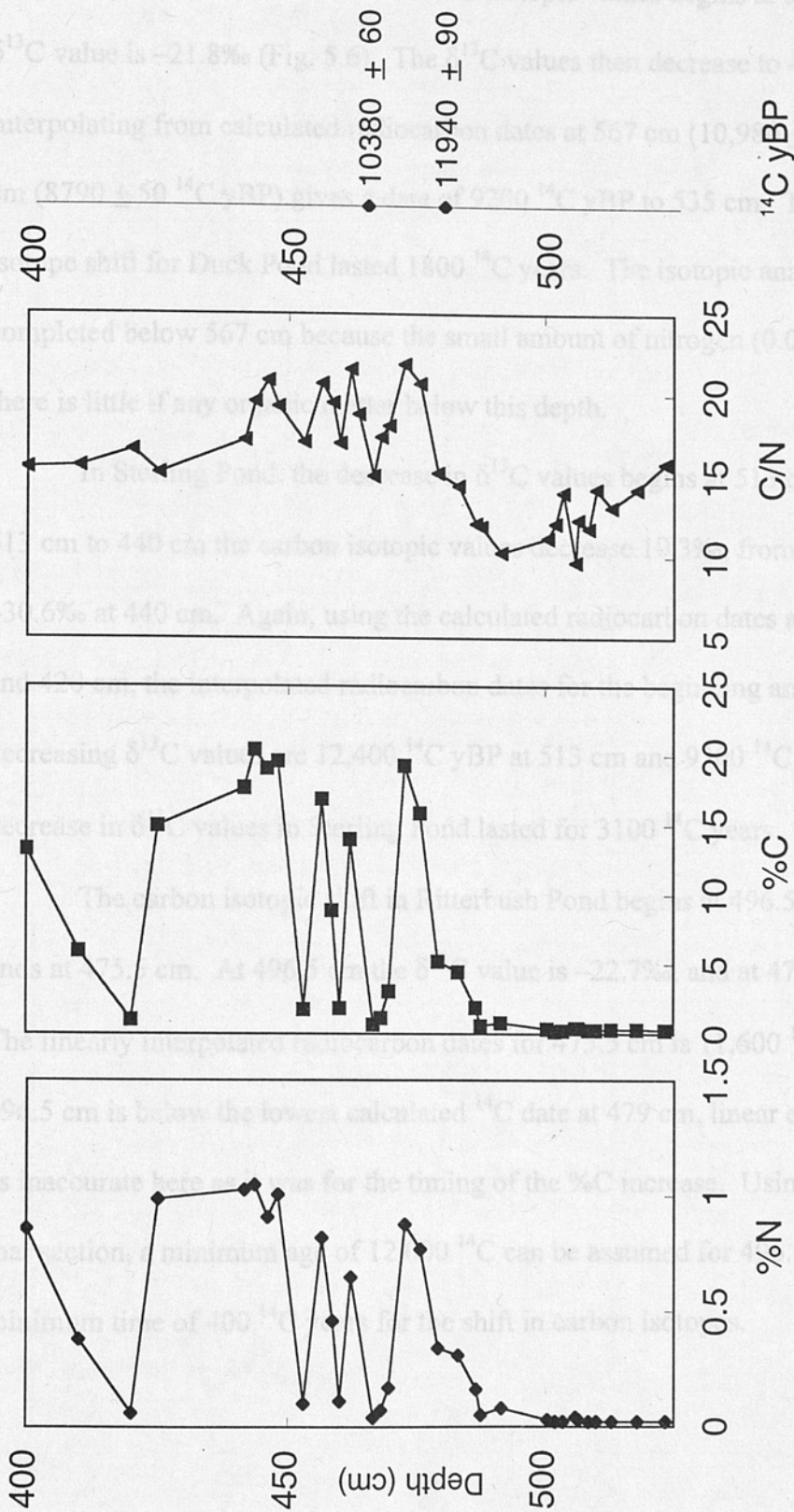


Figure 5.3. The bottom section of the Sterling Pond core, highlighting the increase in %C and %N starting at 502 cm.

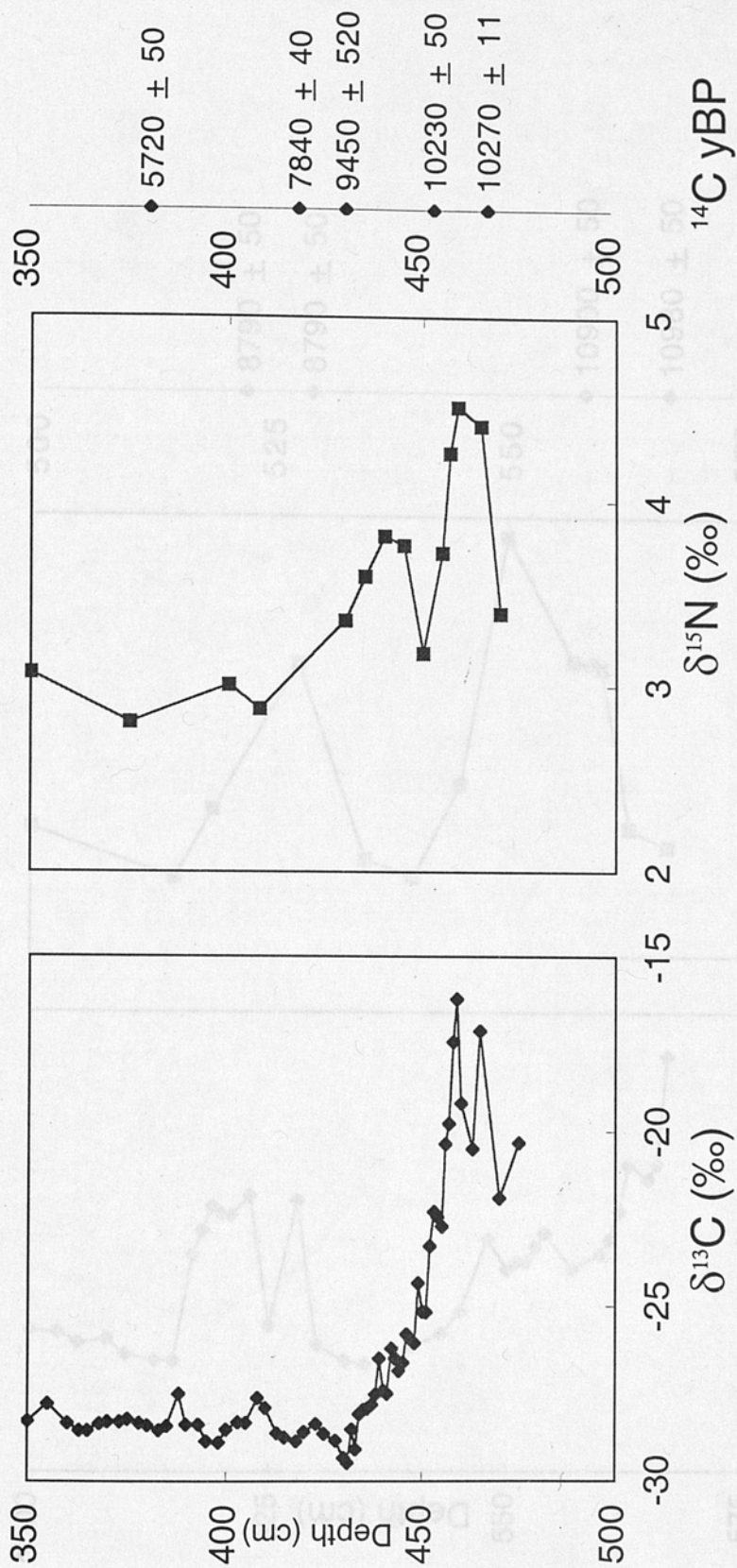


**Figure 5.4.** The bottom section of the Ritterbush core with the %C, %N, and C/N values. Note the increase of the %C and %N values at 487.5 cm.

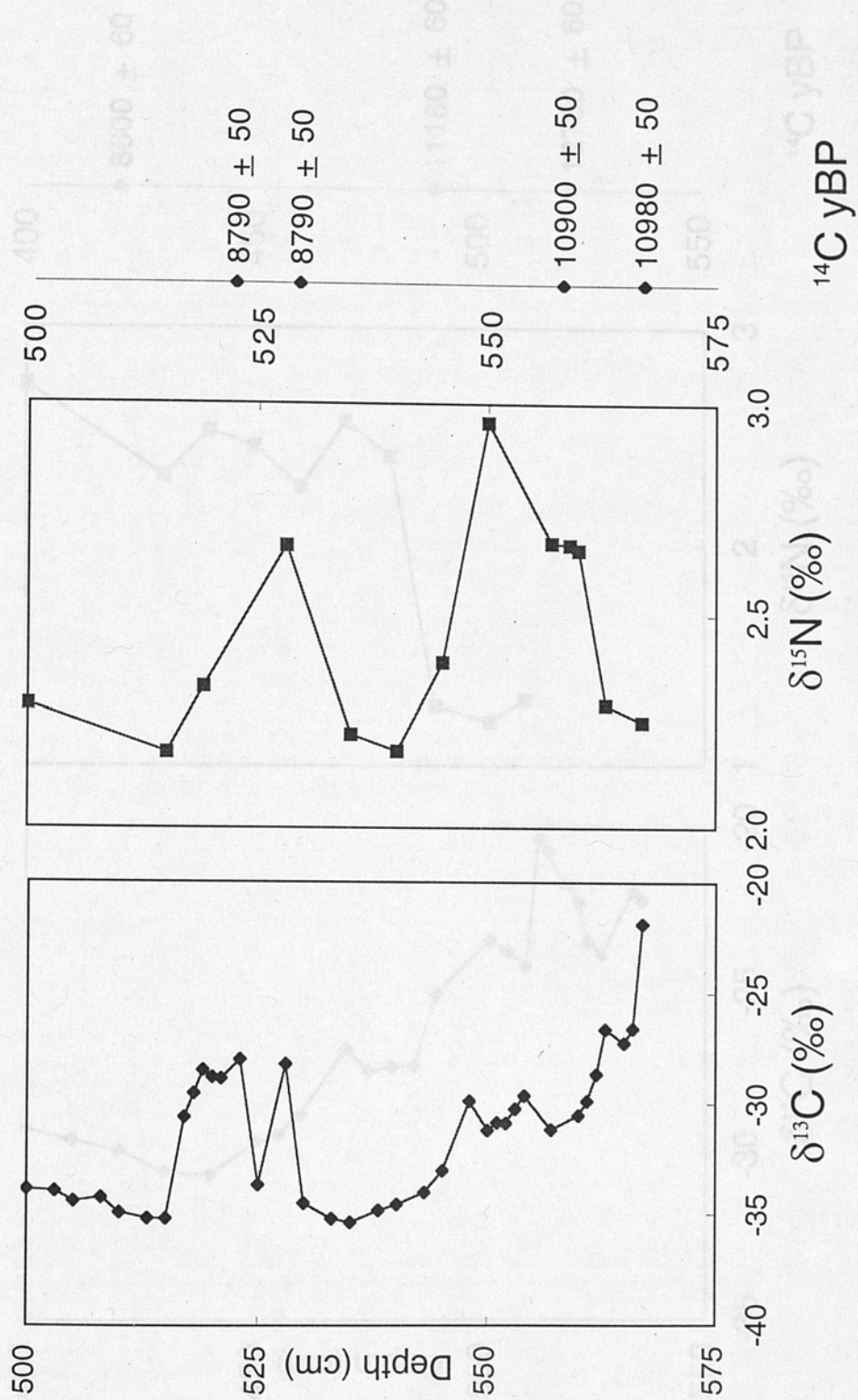
In Duck Pond, the shift in carbon isotopic values begins at 567 cm, where the  $\delta^{13}\text{C}$  value is  $-21.8\text{‰}$  (Fig. 5.6). The  $\delta^{13}\text{C}$  values then decrease to  $-35.3\text{‰}$  at 535 cm. Interpolating from calculated radiocarbon dates at 567 cm ( $10,980 \pm 50$   $^{14}\text{C}$  yBP) and 522 cm ( $8790 \pm 50$   $^{14}\text{C}$  yBP) gives a date of 9200  $^{14}\text{C}$  yBP to 535 cm. The initial carbon isotope shift for Duck Pond lasted 1800  $^{14}\text{C}$  years. The isotopic analyses were not completed below 567 cm because the small amount of nitrogen (0.01%) suggests that there is little if any organic matter below this depth.

In Sterling Pond, the decrease in  $\delta^{13}\text{C}$  values begins at 513 cm (Fig. 5.7). From 513 cm to 440 cm the carbon isotopic values decrease 10.3‰, from  $-20.3\text{‰}$  at 513 cm to  $-30.6\text{‰}$  at 440 cm. Again, using the calculated radiocarbon dates at 521 cm, 490 cm, and 420 cm, the interpolated radiocarbon dates for the beginning and ending of decreasing  $\delta^{13}\text{C}$  values are 12,400  $^{14}\text{C}$  yBP at 513 cm and 9300  $^{14}\text{C}$  yBP at 440 cm. This decrease in  $\delta^{13}\text{C}$  values in Sterling Pond lasted for 3100  $^{14}\text{C}$  years.

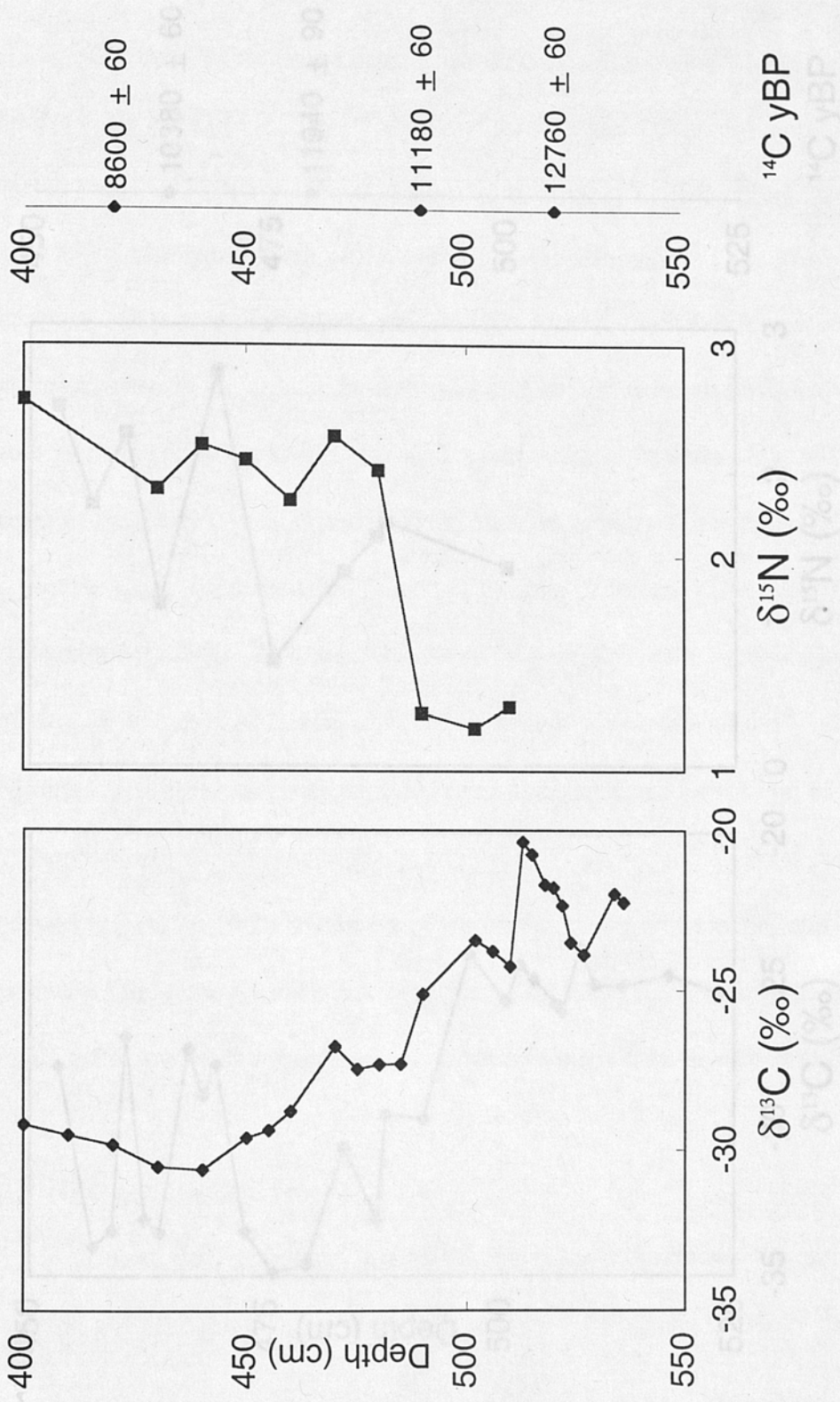
The carbon isotopic shift in Ritterbush Pond begins at 496.5 cm (Fig. 5.8), and ends at 475.5 cm. At 496.5 cm the  $\delta^{13}\text{C}$  value is  $-22.7\text{‰}$ , and at 475.5 cm it is  $-34.9\text{‰}$ . The linearly interpolated radiocarbon dates for 475.5 cm is 11,600  $^{14}\text{C}$  yBP. However, as 496.5 cm is below the lowest calculated  $^{14}\text{C}$  date at 479 cm, linear extrapolation would be as inaccurate here as it was for the timing of the  $\delta^{13}\text{C}$  increase. Using the same logic as in that section, a minimum age of 12,000  $^{14}\text{C}$  can be assumed for 496.5 cm, producing a minimum time of 400  $^{14}\text{C}$  years for the shift in carbon isotopes.



**Figure 5.5.** The  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values from the bottom of the Lake Morey core. Note the decreasing carbon isotopic values beginning at 459 cm.



**Figure 5.6.** The  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values from the bottom of the Duck Pond core. Note the decreasing carbon isotopic values beginning at 567 cm.



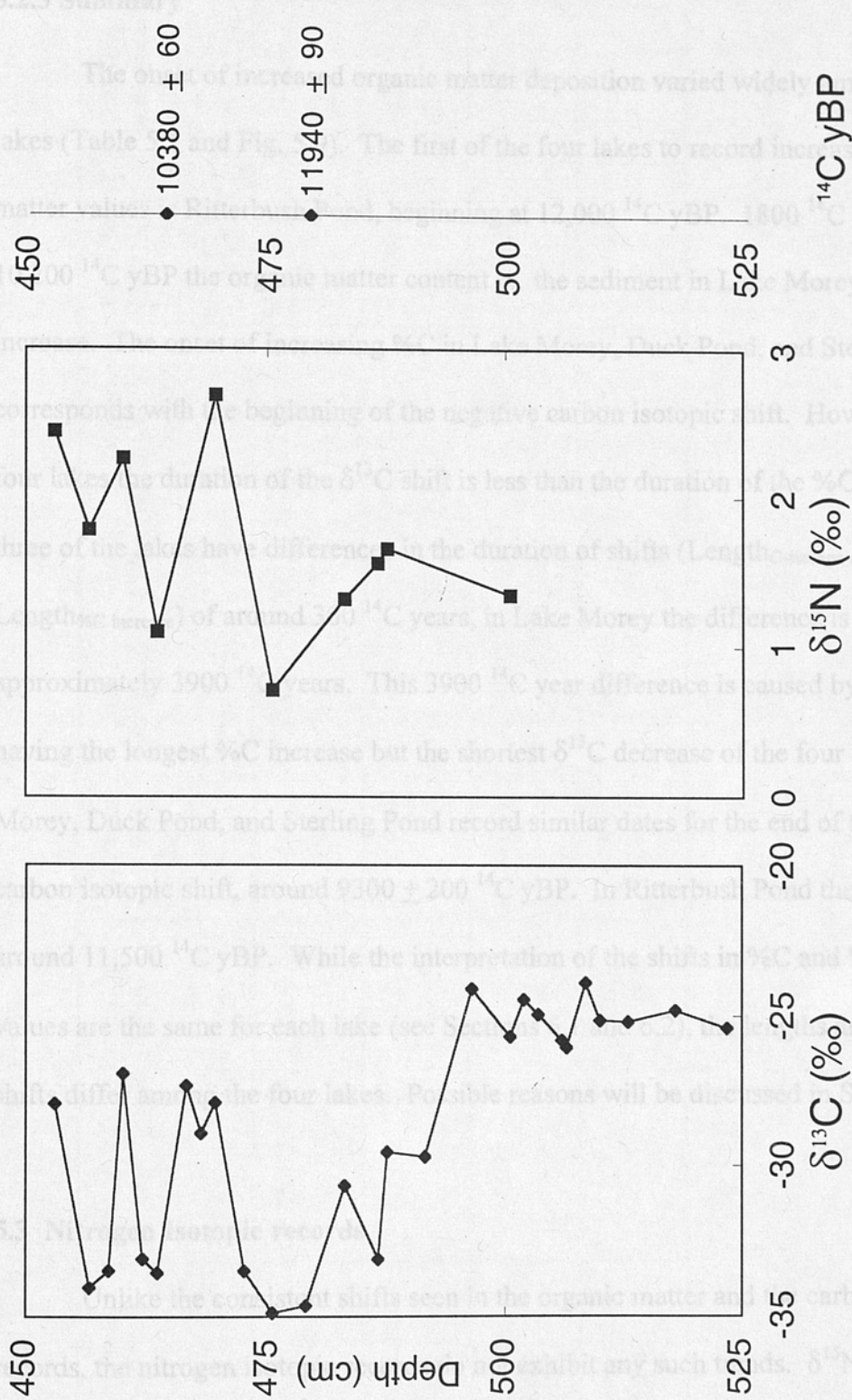
**Figure 5.7.** The  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values from the bottom of the Sterling Pond core. Note the decreasing carbon isotopic values beginning at 502 cm.

5.2.3 Summary

The onset of increased organic matter deposition varied widely among the four lakes (Table 5.1 and Fig. 5.8). The first of the four lakes to record increased organic matter values was Ritterbush Pond, beginning at 12,000  $^{14}\text{C}$  yBP. 1800  $^{14}\text{C}$  years later at 10,200  $^{14}\text{C}$  yBP the onset of the matter content in the sediment in Lake Morey began to increase. The onset of increasing  $\delta^{15}\text{N}$  in Lake Morey, Duck Pond, and Sterling Pond corresponds with the beginning of the negative carbon isotopic shift. However, in all four lakes the duration of the  $\delta^{13}\text{C}$  shift is less than the duration of the  $\delta^{15}\text{N}$  shift. While three of the lakes have different durations of shifts (lengths of shifts) of approximately 1000 years, in Lake Morey the duration of shifts is longer, approximately 1900 years. This 1900 year difference is caused by Lake Morey having the longest  $\delta^{13}\text{C}$  increase but the shortest  $\delta^{13}\text{C}$  decrease of the four lakes. Lake Morey, Duck Pond, and Sterling Pond record similar dates for the end of the negative carbon isotopic shift, around  $9300 \pm 200$   $^{14}\text{C}$  yBP. In Ritterbush Pond the values are around 11,500  $^{14}\text{C}$  yBP. While the interpretation of the shifts in  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values are the same for each lake (see below), the reasons for the shifts in  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values differ among the four lakes. Possible reasons will be discussed below.

5.2.3.1 Nitrogen isotopic ratios

Unlike the constant shifts seen in the organic matter analysis, the nitrogen isotopic ratios exhibit any such shifts.  $\delta^{15}\text{N}$  values have



**Figure 5.8.** The  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  values from the bottom of the Ritterbush Pond core. Note the decreasing carbon isotopic values beginning at 496.5 cm.

### 5.2.3 Summary

The onset of increased organic matter deposition varied widely among the four lakes (Table 5.1 and Fig. 5.9). The first of the four lakes to record increasing organic matter values is Ritterbush Pond, beginning at 12,000  $^{14}\text{C}$  yBP. 1800  $^{14}\text{C}$  years later at 10,200  $^{14}\text{C}$  yBP the organic matter content of the sediment in Lake Morey began to increase. The onset of increasing  $\%C$  in Lake Morey, Duck Pond, and Sterling Pond corresponds with the beginning of the negative carbon isotopic shift. However, in all four lakes the duration of the  $\delta^{13}\text{C}$  shift is less than the duration of the  $\%C$  shift. While three of the lakes have differences in the duration of shifts ( $\text{Length}_{\text{C-isotope decrease}} - \text{Length}_{\%C \text{ increase}}$ ) of around 300  $^{14}\text{C}$  years, in Lake Morey the difference is much longer, approximately 3900  $^{14}\text{C}$  years. This 3900  $^{14}\text{C}$  year difference is caused by Lake Morey having the longest  $\%C$  increase but the shortest  $\delta^{13}\text{C}$  decrease of the four lakes. Lake Morey, Duck Pond, and Sterling Pond record similar dates for the end of the negative carbon isotopic shift, around  $9300 \pm 200$   $^{14}\text{C}$  yBP. In Ritterbush Pond the shift ends around 11,500  $^{14}\text{C}$  yBP. While the interpretation of the shifts in  $\%C$  and  $\%N$  and  $\delta^{13}\text{C}$  values are the same for each lake (see Sections 6.1 and 6.2), the lengths and timing of the shifts differ among the four lakes. Possible reasons will be discussed in Section 6.3.

### 5.3 Nitrogen Isotopic records

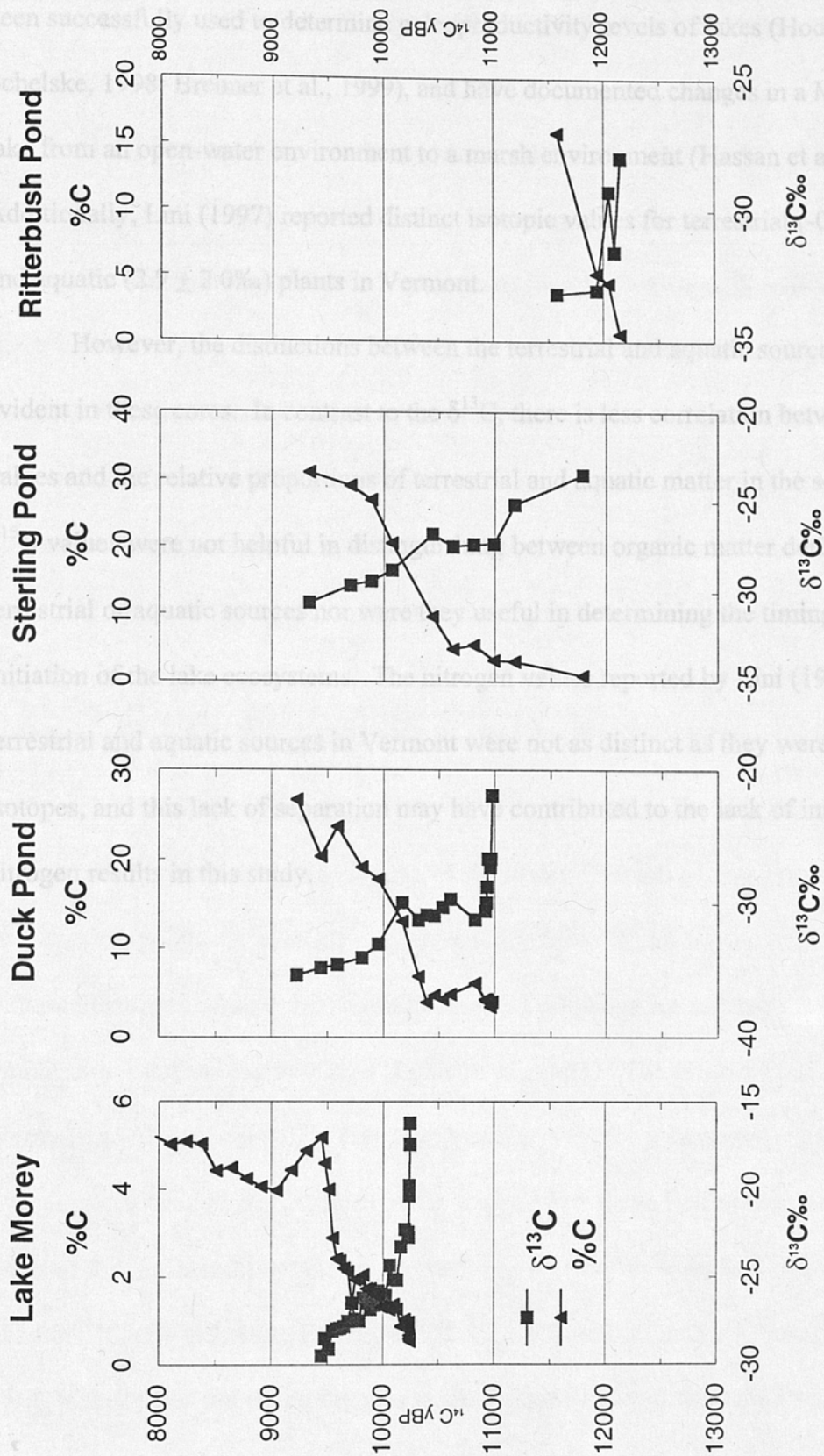
Unlike the consistent shifts seen in the organic matter and the carbon isotopic records, the nitrogen isotopic records do not exhibit any such trends.  $\delta^{15}\text{N}$  values have



**Table 5.1** Summary of lengths of shifts (in uncalibrated  $^{14}\text{C}$  years) for  $\delta^{13}\text{C}$  and  $\delta^{13}\text{C}$  values for Lake Morey, Duck Pond, Sterling Pond, and Ritterbush Pond.

Lake name	Start of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ yBP)	End of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ yBP)	Duration of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ y)	Start of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ yBP)	End of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ yBP)	Magnitude of $\delta^{13}\text{C}$ shift (‰)	Duration of $\delta^{13}\text{C}$ shift ( $^{14}\text{C}$ y)	~Difference in duration of shifts ( $^{14}\text{C}$ y)
Lake Morey	10200	5600	4600	10200	9500	13.2	700	3900
Duck Pond	11000	8900	2100	11000	9200	13.5	1800	300
Sterling Pond	12400	9000	3400	12400	9300	10.3	3100	300
Ritterbush Pond	>12000	11300	>700	>12000	11600	12.2	>400	~300

Figure 5.9. Graphical summary of lengths of shifts in  $\delta^{13}\text{C}$  and  $\delta^{13}\text{C}$  values for the four lakes. Vertical axes are  $^{14}\text{C}$  yBP.



**Figure 5.9.** Graphical summary of lengths of shifts in %C and  $\delta^{13}\text{C}$  values for the four lakes. Vertical axes are  $^{14}\text{C}$  yBP.

been successfully used to determine paleoproductivity levels of lakes (Hodell and Schelske, 1998; Brenner et al., 1999), and have documented changes in a Midwestern lake from an open-water environment to a marsh environment (Hassan et al., 1997). Additionally, Lini (1997) reported distinct isotopic values for terrestrial ( $-0.8 \pm 1.3\%$ ) and aquatic ( $2.9 \pm 2.0\%$ ) plants in Vermont.

However, the distinctions between the terrestrial and aquatic sources were not evident in these cores. In contrast to the  $\delta^{13}\text{C}$ , there is less correlation between the  $\delta^{15}\text{N}$  values and the relative proportions of terrestrial and aquatic matter in the sediment. The  $\delta^{15}\text{N}$  values were not helpful in distinguishing between organic matter derived from terrestrial or aquatic sources nor were they useful in determining the timing of the initiation of the lake ecosystems. The nitrogen values reported by Lini (1997) from terrestrial and aquatic sources in Vermont were not as distinct as they were for carbon isotopes, and this lack of separation may have contributed to the lack of interpretable nitrogen results in this study.

## CHAPTER 6. INTERPRETATION AND DISCUSSION

The following sections will discuss possible causes for the two consistent shifts, the increase in %C and the decrease in  $\delta^{13}\text{C}$ , documented in the oldest sediment of the four lakes. The %C increase will be discussed first, beginning with comparing the basal dates of the four lakes with the established deglaciation history of the region. The %C increases will then be compared to similar records from other lakes. The decrease in  $\delta^{13}\text{C}$  values will then be considered, along with possible causes of the shift. The suggested causes are a change in  $\text{pCO}_2$ , a shift from  $\text{C}_4$  to  $\text{C}_3$  plants in the lake basin, and finally a change in the relative contributions of terrestrial and aquatic plants to the lake sediment.

### 6.1 Organic matter records

At the peak of the last glacial period the area encompassing the four study lakes was completely covered by ice (Ridge et al., 1999). The advancing glaciers destroyed any existing preglacial plant communities in the area. When the ice retreated, the newly-exposed land was initially barren and lifeless. Following the melting glaciers, were plant communities colonizing new land (Petee et al., 1993). The deglaciation chronology in Vermont and New Hampshire (see Ridge et al., 1999, for a summary) has been established by dating the organic matter produced by these first plants in the area. The dates of the initial increase in sedimentary organic matter deposition in Duck Pond (11,000  $^{14}\text{C}$  yBP), Sterling Pond (11,800  $^{14}\text{C}$  yBP), and Ritterbush Pond (>12,000  $^{14}\text{C}$  yBP) indicate that the oldest organic matter in the lakes was deposited soon after

deglaciation. The onset of organic matter deposition in Lake Morey (10,300  $^{14}\text{C}$  yBP) corresponds well with Ridge et al.'s (1999) date of 10,400  $^{14}\text{C}$  yBP for the draining of glacial Lake Hitchcock, which uncovered Lake Morey and allowed plant communities to develop around and within the lake.

Similar records of post-glacial increases in sedimentary organic matter also exist for Lake Ontario (Silliman et al., 1996), Lake St. Moritz, Switzerland (Ariztegui et al. 1996), and Moon Lake, North Dakota (Valero Garcés et al., 1997). Other examples of lakes recording increased sedimentary organic matter at the start of the Holocene (e.g. Lake Victoria, Africa (Talbot and Laerdal, 2000) and Lake Biwa, Japan (Meyers and Takemura, 1997)) exist, but these lakes were not glaciated. These lakes had preexisting ecosystems contributing organic matter to the sediment during the Pleistocene; the increases of sedimentary organic matter recorded in these unglaciated lakes do not represent the establishment of plant ecosystems on formerly barren ground. Although the researchers studying Lake Ontario, Lake St. Moritz, and Moon Lake did not discuss the mechanisms that produced increased organic matter after deglaciation, it was also probably caused by the establishing local ecosystems.

$\delta^{13}\text{C}$  value decrease (10.3‰) in the lakes

## 6.2 C-isotopic shift

In addition to the increase in organic matter in each lake's sediment there is also a roughly concurrent negative shift in the  $\delta^{13}\text{C}$  values of the sediment. Previous studies (See Table 6.1 for a summary of previous interpretations of  $\delta^{13}\text{C}$  shifts occurring at the

more than 13‰. The timing of the shifts varied among the four lakes, with at least 1500

glacial-postglacial transition) have proposed three processes to explain other records of decreasing carbon isotopic values at the beginning of the Holocene.

One proposed explanation for the decreasing  $\delta^{13}\text{C}$  values of the organic matter is a change in atmospheric  $\text{pCO}_2$ . Raynaud et al. (1993) documented an increase in  $\text{pCO}_2$  from 200 ppmv to 280 ppmv at the beginning of the Holocene. Increasing  $\text{pCO}_2$  induces increased carbon isotopic fractionation in plants, producing organic matter with more negative  $\delta^{13}\text{C}$  values (O'Leary and Paneth, 1992). Crayton and Epstein (1977) documented a 2‰ decrease in  $\delta^{13}\text{C}$  values from wood samples from North America from 12,000 to 10,000  $^{14}\text{C}$  years BP. Isotopic changes attributed to post-glacial increases in  $\text{pCO}_2$  are also found in organic matter in marine sediments ( $\sim -3\%$ ; Jasper and Hayes, 1990) and lake sediments from Kenya ( $\sim -14\%$ ; Street-Perrott et al., 1997) and Japan ( $\sim -4\%$ ; Meyers and Horie, 1993) (Table 6.1).

While it is possible that the increase in  $\text{pCO}_2$  caused some of the initial decline observed in the  $\delta^{13}\text{C}$  values of the organic matter from the four study lakes it can not explain the full magnitude of the shifts. The magnitudes of the carbon isotopic shifts believed to be caused by the increase in  $\text{pCO}_2$  are less by a factor of five than the smallest  $\delta^{13}\text{C}$  value decrease (10.3‰) in the lakes.

Additionally, if the post-glacial global change in  $\text{pCO}_2$  were the driving force behind the negative carbon isotopic shift, then not only should the magnitudes of the shifts be similar, but their timing should also be synchronous. Neither is the case. The magnitudes of the shifts varied more than 3‰ among the four study lakes, from 10.3‰ to more than 13‰. The timing of the shifts varied among the four lakes, with at least 1800

**Table 6.1.** Magnitude, duration, and interpretation of glacial-postglacial carbon isotopic shifts reported in the literature (Lini, unpublished).

Post-glacial lakes	+/-	Magnit. (%)	Duration $^{14}\text{C}$ y	Interpretation	Reference
Chesney Bog, Canada	+	11 (-26 to -15)	1500	> lake production Algae using $\text{HCO}_3^-$	Aravena et al., 1992
Toboggan Lake, Canada	-	10 (-26 to -36)	2500	> lake production	Aravena et al., 1992
Horse-Shoe Lake, Minnesota	+	11 (-28 to -17)	3700	Change in water hardness	Stuiver, 1975
<b>Unglaciatiated lakes</b>					
Lake Bosumtwi, Africa	-	21 (-10 to -31)	1400	$\text{C}_4 \rightarrow \text{C}_3$ ecosystem, forests replacing grassland	Talbot and Johannesen, 1992
Lake Barombi Mbo, Africa	-	7 (-26 to -33)	4000?	$\text{C}_4 \rightarrow \text{C}_3$ plants	Giresse et al., 1994
Lake Victoria, Africa	-	5 (-19 to -24)	1400	Shift to algal dominated production	Talbot and Lærdal, 2000
Lake Albert, Africa	-	7 (-19 to -26)	5050	Increased algal production	Beuning et al., 1997
Sacred Lake, Africa	-	15 (-15 to -30)	2200	$\text{C}_4 \rightarrow \text{C}_3$ plants Effect of increased $\text{pCO}_2$	Street Perrott et al., 1997
Lake Kimilili, Africa	-	14 (-7 to -21)	6000	Algae? Effect of increased $\text{pCO}_2$	Street Perrott et al., 1997
Lake Biwa, Japan	-	4 (-22 to -26)	?	Effect of increased $\text{pCO}_2$	Meyers and Horie, 1993
Lake Jih Tan, Taiwan	+	7 (-24 to -17)	3000	Change in water hardness and/or productivity	Stuiver, 1975
Lake Baikal, Siberia	-	8 (-20 to -28)	?	Increased algal production	Qiu et al., 1993

$^{14}\text{C}$  years separating the initiation of the carbon isotopic shift in Ritterbush Pond and Lake Morey, and 2100  $^{14}\text{C}$  years between the ending of the shifts in Ritterbush Pond and Duck Pond (Table 5.1 and Fig. 5.9).

A second possible cause for the negative  $\delta^{13}\text{C}$  shift is a change from  $\text{C}_4$  to  $\text{C}_3$  plant communities growing in the lake's watershed and contributing organic matter to the lake sediment (Table 6.1). The different biochemical pathways that these two plant groups use during photosynthesis produce isotopic differences among them, with  $\text{C}_4$  plants having higher  $\delta^{13}\text{C}$  values (average: -8‰) than  $\text{C}_3$  plants (average: -28‰) (Fogel and Cifuentes, 1993).  $\text{C}_4$  plants are mostly grasses that dominate over  $\text{C}_3$  plants during conditions of low  $\text{pCO}_2$  (Collatz et al., 1998). Negative shifts in the  $\delta^{13}\text{C}$  values of sedimentary organic matter have been attributed to the shift from  $\text{C}_4$  to  $\text{C}_3$  plants in Africa (Talbot and Johannessen, 1992; Street-Perrott et al., 1997; Talbot and Laerdal, 2000) (Table 6.1).

It is possible that in the study lakes some of the initial post-glacial organic matter was from tundra  $\text{C}_4$  grasses that were eventually replaced by  $\text{C}_3$  forests. Pollen records from various New England lakes do show a transition from tundra to more boreal plant communities following deglaciation (e.g. Peteet et al., 1993; Jackson et al., 1997; Maenza-Gmelch, 1997). However, if  $\text{C}_4$  plants initially produced the majority of the organic matter then the initial  $\delta^{13}\text{C}$  values should be closer to that of  $\text{C}_4$  plants. The least negative  $\delta^{13}\text{C}$  values measured at the beginning of the carbon isotopic shift are -16.3‰ in Lake Morey and -22.7‰ in Ritterbush Pond. These  $\delta^{13}\text{C}$  values are more negative than what expected if  $\text{C}_4$  plants originally produced most of the sedimentary organic



matter. Thus, while the possibility of a shift from C<sub>3</sub> to C<sub>4</sub> plants can not be eliminated, the importance of this process in causing the initial negative shift in carbon isotopic values is minimal.

The third possible cause for the observed decrease in carbon isotopic values is a change in the source of the organic matter (Table 6.1). If the organic matter originating from terrestrial and aquatic sources has different isotopic composition and the relative contributions from these two sources change over time, then the overall isotopic value of the sediment will also change. An increased contribution of aquatically-produced organic matter has been cited as the cause of negative carbon isotopic shifts in lake sediments in Africa (Talbot and Johannessen, 1992; Talbot and Laerdal, 2000) and Asia (Qiu et al., 1993) (Table 6.1). While aquatically-produced organic matter is not always isotopically distinct from terrestrial organic matter (Meyers, 1997), in the case of the four studied lakes a change in the relative proportion of organic matter from the two sources represents a plausible explanation for the initial negative shift in carbon isotopic values.

If this last mechanism is the cause of the negative shift documented in the four Vermont lakes, then it is important to establish that the terrestrial and aquatic sources of organic material to these lakes are isotopically distinct. Sedimentological evidence from the cores suggests that they do. In lake sediments, increased grain size is an indicator of a greater proportion of terrestrial material (Silliman et al., 1996; Meyers and Takemura, 1997; Bierman et al. 1997; Brown et al., 2000). There are several layers in the cores that contain sand, such as at 40 cm, 100 cm, and 154 cm in Ritterbush Pond and at 325 cm and 477-483 cm in Duck Pond. These layers frequently contain identifiable terrestrial

macrofossils, including spruce cones, leaves, and twigs. Isotopic analyses (see Bierman et al., 1997, for Ritterbush Pond) indicate that each one of these layers with abundant terrestrial material also has higher  $\delta^{13}\text{C}$  values than the gyttja above and below it. In Sterling Pond, isolated terrestrial macrofossils analyzed for  $\delta^{13}\text{C}$  display less negative values than the surrounding gyttja (Lini, unpublished).

In addition to the differences in the terrestrial and aquatic organic material observed within the sediment, modern plant matter collected in and around Vermont lakes also exhibits differences. Lini (1997) analyzed modern terrestrial and aquatic flora in Vermont and found that the aquatic plants had generally lower  $\delta^{13}\text{C}$  values than their terrestrial counterparts.

A decrease in the relative proportion of less negative land-derived organic matter (and, consequently, an increase in the amount of more negative aquatic organic matter) in the lake sediment would cause the  $\delta^{13}\text{C}$  values of the sediment to also decrease. This is the most plausible hypothesis to explain the initial  $\delta^{13}\text{C}$  shift observed in the four Vermont lake cores. The concurrent increases in sedimentary organic matter suggest that both land and aquatic ecosystems were growing and contributing organic matter to the sediment, but the decreasing carbon isotopic values indicate that the terrestrial ecosystems developed prior to the aquatic ones. As the aquatic ecosystem grew and became an important source of organic matter, the carbon isotopic values of the sediment decreased. Once the proportion of terrestrially and aquatically produced organic matter stabilized, the negative shift in the  $\delta^{13}\text{C}$  values of the sediment ceased. The stabilization of the carbon isotopic values occurred before the increase in carbon content ceased in all

four lakes, implying that the terrestrial and aquatic ecosystems continued to grow after their relative contributions to the sediment stabilized.

### 6.3 Differences in the timing of ecosystem establishment

Although the same general pattern of increasing organic matter content and decreasing carbon isotopic values is observed in the oldest sediment of each lake, there are certain differences among them. In this section, I will discuss possible scenarios that produced the documented differences in the timing of ecosystem establishment.

Lake Morey was the last of the four lakes to form, although the area was free of ice around 12,400  $^{14}\text{C}$  years ago (Ridge et al., 1999). Lake Morey did not exhibit signs of ecosystem development for approximately 2000  $^{14}\text{C}$  years after deglaciation because it was covered by the large, post-glacial Lake Hitchcock in the Connecticut River valley. From varve chronologies and radiocarbon dating, Ridge et al. (1999) estimated that the draining of Lake Hitchcock occurred approximately 10,400  $^{14}\text{C}$  yBP. This correlates well with the basal date of 10,300  $^{14}\text{C}$  yBP for the initiation of Lake Morey. It was only after Lake Hitchcock drained that ecosystems could develop in the basin that became Lake Morey.

While the presence of glacial Lake Hitchcock is a plausible explanation for why Lake Morey was the last of the four lakes to form, there is not a similarly good explanation for why the sediment in Lake Morey continues to record increasing organic matter values long after the other three records. The sediment in Lake Morey shows increasing  $\%C$  values for over 4600  $^{14}\text{C}$  years after the initial increases (Table 5.1 and

Fig. 5.9). This is 1200  $^{14}\text{C}$  years longer than the next longest shift of 2800  $^{14}\text{C}$  years in Sterling Pond. However, even after this long increase, the sediment still contains the lowest %C values of the four lakes. Sterling Pond, with its high elevation producing a short growing season and a relative lack of terrestrial inputs that would have provided nutrients, should have the slowest increases in sedimentary organic matter. The plants colonizing Lake Morey, the lowest-elevation lake of the four, may have had the advantage of arriving from established plant populations in nearby lakes and rivers that were not covered by Lake Hitchcock, instead of advancing piecemeal after the melting glacier. This probably explains why Lake Morey displays the shortest shift in  $\delta^{13}\text{C}$  values, but the long increase in %C values is puzzling. The C/N values above 434 cm depth (9600  $^{14}\text{C}$  yBP) are between 11 and 12. The magnitude of the C/N values indicates that the source of the organic matter is mostly aquatic, and the lack of variation in C/N values suggests that little terrestrial material was entering the lake after 9600  $^{14}\text{C}$  yBP. Also, the sedimentation rate during the period of increasing sedimentary organic matter is lower than it is either before or after in the core (Fig. 3.5). A low organic matter accumulation rate would explain the low %C values in Lake Morey.

The sediment in Sterling Pond records the second longest record of increasing organic matter content and the longest record of decreasing  $\delta^{13}\text{C}$  values (Table 5.1 and Fig. 5.9) of the four lakes. There are two likely reasons for the slow establishment of the ecosystems at Sterling Pond. These are the elevation of the pond, and the amount of terrestrial material being delivered to the lake basin from the watershed. Sterling Pond is the highest of the four study lakes, situated at 917 m above sea level, along the spine of

the Green Mountains. The high elevation produces shorter growing seasons and colder temperatures than at the other three lakes. The shorter growing season allows less time per year for the ecosystems to grow, as the lake and its watershed are covered in ice and snow more days of the year than the other three lakes.

The second reason for the delay in the establishment of the ecosystems in and around Sterling Pond is the small amount of terrestrial material being input into the lake compared to the other three lakes. Unlike the other three lakes, the sediment at the bottom of Sterling Pond lacks the layers of terrestrial inorganic material interspersed in the gyttja. These detrital layers frequently provide nutrients such as nitrates and phosphates (Meyers and Takemura, 1997) to the lake, and the development of the aquatic ecosystem in Sterling Pond could have been slowed by a lack of nutrients. This lack of inputs is most likely caused by the low relief (40 m) of the Sterling watershed.

Though the establishment of the aquatic ecosystem in Sterling Pond was delayed by a lack of terrestrial nutrients, the aquatic plants in Ritterbush Pond developed quickly. Ritterbush has the shortest recorded length of increasing organic matter content and the second shortest decrease in  $\delta^{13}\text{C}$  values (after Lake Morey), at 900 and 1100  $^{14}\text{C}$  years, respectively (Table 5.1 and Fig. 5.9). Ritterbush Pond differs from Sterling Pond in the amount of terrestrial material that was delivered to the lake early in its history (Brown et al., 2000). The oldest sediment of Ritterbush records numerous pulses of terrestrial material, with larger grain sizes and a higher amount of inorganic material than the surrounding gyttja. Each of these events brought terrestrial material to the lake and would have stocked the lake with nutrients such as phosphates and nitrates, allowing the

ecosystems to establish quickly. Ritterbush Pond is also 600 m lower in elevation than Sterling Pond, which produces a longer growing season at Ritterbush Pond.

7.1.5 Duck Pond's history is similar to that of Ritterbush Pond. There are several layers of terrestrially derived inorganic sediment at the bottom of the core that might represent periods of increased nutrient delivery to the lake. Each time an inorganic layer was deposited, nutrients would be added to the lake, promoting ecosystem growth. These layers, however, do not occur as frequently as they do in Ritterbush Pond, and Duck Pond is also situated 200 m higher than Ritterbush Pond. The fewer detrital layers and higher elevation probably caused the increase in organic matter content (2100  $^{14}\text{C}$  years) and decrease in  $\delta^{13}\text{C}$  values (1800  $^{14}\text{C}$  years) to be longer in Duck Pond than in Ritterbush Pond (Table 5.1 and Fig. 5.9). The ecosystems established themselves at Duck Pond faster than they did at Sterling Pond because of the more frequent nutrient-providing terrestrial inputs and lower elevation when compared to Sterling Pond.

## CHAPTER 7. CONCLUSIONS

### 7.1 Summary of findings

Two consistent trends are recorded in the sediment from all four lakes, the initial increase in organic matter content and the initial decrease in carbon isotopic values. These shifts reflect the establishment of the terrestrial and aquatic ecosystems following deglaciation, with the decrease in  $\delta^{13}\text{C}$  values indicating that the development of the aquatic ecosystem was slower than development of the terrestrial ecosystem. The growing terrestrial and aquatic ecosystems caused the increase in carbon and nitrogen content, as they contributed more and more organic matter to the sediment of the lakes over time. At the end of the increase in sedimentary organic matter, the local ecosystems were fully established.

Initially, the terrestrial ecosystems grew faster than their aquatic counterparts and, consequently, delivered more organic material than the aquatic plants. As the aquatic ecosystems evolved, they contributed more negative  $\delta^{13}\text{C}$  organic material to the sediment, thus causing the overall decrease in carbon isotopic values of the sediment. Once the shifts in  $\delta^{13}\text{C}$  values ended, the current proportions of terrestrial and aquatic organic material were presumably established. It is difficult to exactly quantify the prior relative proportions of terrestrial and aquatic matter in the sediment, as presumably the early-Holocene algae's carbon isotopic composition varied as much ( $\sim 20\%$ ) as the modern ones do in Vermont (Lini, 1997). The isotopic composition of the terrestrial organic matter can be deduced from the composition of the land-derived layers, but

unless the carbon isotopic value of the algae were determined, then the relative contribution of the two sources can not be accurately calculated. For example, the  $\delta^{13}\text{C}$  of the original material in the detrital layers from Ritterbush Pond averages  $-27\text{‰}$  while the majority of the sediment is approximately  $-31\text{‰}$ . If the algae's isotopic composition is estimated to be  $-33\text{‰}$ , then the relative contributions from terrestrial and aquatic sources would be approximately equal. However, if the  $\delta^{13}\text{C}$  of the algae were  $-37\text{‰}$  instead, then the organic matter would be roughly 40% aquatic and 60% terrestrial. Thus, the initial estimate of the  $\delta^{13}\text{C}$  value of the algae determines the estimate of the relative contributions of the two sources. This method would not be feasible in Sterling Pond or Lake Morey, which lack detrital layers. However, isolated terrestrial microfossils could be analyzed to arrive at an estimate of the composition of the terrestrial organic matter.

Although stable proportions of organic matter derived from the terrestrial and aquatic ecosystems were established by the end of the carbon isotopic shifts (9500  $^{14}\text{C}$  yBP for Lake Morey, 9200  $^{14}\text{C}$  yBP at Duck Pond, 9300  $^{14}\text{C}$  yBP at Sterling Pond, and 11,600  $^{14}\text{C}$  yBP for Ritterbush Pond), these ecosystems continued to develop together and contribute additional organic matter to the sediment. The additional growth is indicated by the continued increase in organic matter content well after the end of the decrease in carbon isotopic values.

The amount of terrestrial inputs to the lake, the elevation, and the glacial history of the landscape appear to be the main factors influencing the timing and length of the  $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$  shifts. The input of terrestrial material affects the amount of nutrients available for plant growth, the elevation determines the length of the growing season, and



post-glacial landforms affected the development of one of the lakes. Sterling Pond had the longest recorded decrease in  $\delta^{13}\text{C}$  values, and it had the least terrestrial input and the highest elevation. Duck Pond, with higher inputs of terrestrial material and lower elevation, had briefer shifts in organic matter content and carbon isotopic values than Sterling Pond. Ritterbush Pond, which is both lower in elevation and was subject to higher terrestrial influx than Duck Pond, had even shorter organic matter and carbon isotopic shifts than those recorded in Duck Pond. Glacial Lake Hitchcock delayed the development of Lake Morey, but once Lake Morey became established, the relative proportions of terrestrial and aquatic organic matter stabilized faster than they did in any of the four other lakes. However, this stabilization of the isotopic record is in contrast to the continued increase in organic content in Lake Morey, which lasted longer than in any of the other three lakes.

The nitrogen isotope records were more complex than the carbon isotope records. The oldest sediments in the lakes did not show clear trends in their nitrogen isotope composition, nor was there any correlation between the  $\delta^{15}\text{N}$  values and the source of the sediment; when the relative amounts of terrestrial and aquatic organic matter changed, the  $\delta^{15}\text{N}$  did not exhibit as clear a pattern of changes as the  $\delta^{13}\text{C}$  values did.

## 7.2 Future Research

There are several future directions to take to expand and verify the results of this study. First, more complete post-glacial cores from a larger geographic area need to be obtained. The larger data set will assist in determining the true relevance of elevation and

nutrient input on the rate of ecosystem establishment. Cores should continue to be collected from a range of elevations and from a variety of geomorphic settings to include diverse levels of inorganic inputs into the lake. Another core from a lake once-covered by glacial Lake Hitchcock would provide interesting comparison to the record from Lake Morey. Also, it would be interesting to determine if the documented climatic cooling during the Younger Dryas period (~13,000-11,500 calendar yBP) had any effect on the rate of ecosystem development. The ecosystems at Lake Morey, Duck Pond, and Sterling Pond were growing during the Younger Dryas, and the lower temperatures may have slowed their growth rate. To accomplish this, complete post-glacial cores from lakes that were deglaciated earlier than the four lakes in this study are needed. These cores would include the sediment from ecosystems that were established before the onset of the Younger Dryas, and would most likely come from more southern locations in New England that were deglaciated earlier than the four in this study.

Second, while the sediment should continue to be analyzed for organic matter content and carbon isotopes, additional laboratory techniques may provide further insight into the history of ecosystem development. One of these techniques is pigment characterization. Interpretation of previous phytoplankton composition has been made using total carotenoid concentrations (Leavitt, 1993), and might be helpful in quantifying the amount of aquatic organic material present in the sediment. Another possibly helpful analysis would be Rock-Eval pyrolysis, which determines the hydrogen index (HI) of the organic matter. Changes in the HI have been associated with changes in the source (terrestrial vs. aquatic) of lacustrine organic matter (Ariztegui et al., 1996). These

additional techniques might help to more accurately determine the relative proportions of terrestrial and aquatic material in the sediment.

While nitrogen isotopic analyses were not helpful in the analysis of these lakes, the analyses done on modern data imply that differences between terrestrial and aquatic organic matter may be detected with  $\delta^{15}\text{N}$ . However, until more conclusive results on their usefulness are obtained, the resolution of the nitrogen analyses will not approach that of the  $\delta^{13}\text{C}$ , and should be a secondary priority.

Finally, the importance of accurate radiocarbon dating is paramount in any future studies involving lake sediments. As the sedimentation rate can change quickly following deglaciation, the use of linear interpolation between distant radiocarbon dates becomes less certain. This is especially true when the oldest radiocarbon date is not from the oldest organic matter. Accurate and reliable comparisons of ecosystem development rates between lakes are only possible when the ages of the lake sediments are well known.

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## APPENDIX A. DETAILED LITHOLOGICAL DESCRIPTIONS

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### A.1 Lake Morey

From 360 cm to 501.5 cm the sediment is mostly gray, fine silt (Fig. 4.6). Interspersed in this material are several dark brown/gray layers with coarser (very fine sand) layers. These thin (0.5 cm or less), coarser layers have sharp top and bottom contacts. From 501.5 cm to 495.5 cm is a thicker layer with the same darker, coarser material as the previous layers. After 495.5 cm, the sediment is again mostly gray, fine silt interspersed with 13 thin (12 layers <0.5 cm, 1 at 1.5 cm) layers containing darker, very fine sand. This continues until 473.5 cm, where there is a 2.5 cm thick layer of gray/dark gray medium sand, the coarsest material found in this core. The bottom contact of this coarse layer is sharp while the top contact is gradational.

From 470 cm until 454 cm the sediment is finer and lighter, changing to gray silt. In this section there are three layers with coarser material, the thickest layer being 1 cm thick. There are also several color changes (gray to brown/gray) here that do not have

## APPENDIX A. DETAILED LITHOLOGICAL DESCRIPTIONS

If not stated otherwise, contacts in the lithology of Lake Morey and Duck Pond can be assumed to be sharp for all layers. A sharp contact is a change within approximately 1mm, a gradational contact between 1 mm and 0.5 cm, and a diffuse contact occurs over a distance of greater than 0.5 cm. Grain-size was determined tactilely, so that these descriptions should be assumed to be more qualitative than quantitative.

### A.1 Lake Morey

From 560 cm to 501.5 cm the sediment is mostly gray, fine silt (Fig. 4.6). Interspersed in this material are several dark brown/gray layers with coarser (very fine sand) layers. These thin (0.5 cm or less), coarser layers have sharp top and bottom contacts. From 501.5 cm to 495.5 cm is a thicker layer with the same darker, coarser material as the previous layers. After 495.5 cm, the sediment is again mostly gray, fine silt interspersed with 13 thin (12 layers <0.5 cm, 1 at 1.5 cm) layers containing darker, very fine sand. This continues until 472.5 cm, where there is a 2.5 cm thick layer of gray/dark gray medium sand, the coarsest material found in this core. The bottom contact of this coarse layer is sharp while the top contact is gradational.

From 470 cm until 454 cm the sediment is finer and lighter, changing to gray silt. In this section there are three layers with coarser material, the thickest layer being 1 cm thick. There are also several color changes (gray to brown/gray) here that do not have

corresponding changes in grain size. From 454 cm to 451.5 cm the grain size coarsens to fine sand and the sediment shades to dark gray. This 2.5 cm thick section does have a thin (3 mm), brown layer in the middle of the same grain size (fine sand) as the surrounding material. Except for two thin layers, the sediment from 451.5 cm to 358 cm is all fine grained. There is a 0.7 cm thick layer from 449 cm to 448.3 cm and a 0.5 cm layer from 441.5 cm to 441 cm that contain very fine sand, but the rest of the sediment is silt. Again, there are multiple color changes (from dark gray to brown and back) within this interval, but the grain size does not change except during the two thin layers. These color changes cease around 430 cm. A diffuse contact around 404 cm marks the beginning of a consistent brown color, which continues to 358 cm.

At 358 cm there is a gradational contact where the sediment coarsens to very fine sand and is slightly darker. This layer ends at a diffuse contact around 356 cm. From 356 cm to 305 cm the sediment is consistently brown, silt, until it gradationally changes to very fine sand 305 cm. This 1.5 cm thick layer is darker and contains many macrofossils. After a gradational contact at 303.5 cm the sediment changes back to brown silt until another gradational contact at 300 cm. This is another darker layer of very fine sand that contains many macrofossils. Another gradational contact terminates this layer at 295.5 cm. From 295.5 cm until 286.5 cm the sediment is again brown silt, but coarsens gradationally at 286.5 cm to very fine sand. This 1.5 cm thick layer is darker and contains many macrofossils. After a diffuse contact at 285 cm the sediment fines back to brown silt for the next 8 cm, until 277 cm. At 277 cm the sediment darkens

with macrofossils. This layer continues until a diffuse contact around 275 cm. After 275

and coarsens to very fine sand. This layer also contains multiple macrofossils. A diffuse contact around 261 cm ends this layer, after which the sediment is again the brown silt.

The brown silt continues until a gradational contact at 224 cm, where the sediment coarsens to very fine sand and contains several macrofossils. The brown silt returns at 217 cm, and continues until a gradational contact at 129 cm, where the sediment becomes dark brown, fine sand rich in macrofossils. A gradational contact 9 cm higher at 120 cm depth ends this layer, after which the sediment is brown silt until a gradational contact at 16 cm. There the sediment becomes gray/brown very fine sand, which continues all the way to the top of the core.

## **A.2 Duck Pond**

The bottom most sediment in Duck consists of coarse (2-5 mm), gray pebbles and sand (Fig. 4.11). The top of this layer also contains several macrofossils such as leaves, twigs, and cones. At 567 cm there is a gradational contact, and the grain size fines to silt while the color changes to dark brown. At 556 cm there is a sharp contact and the sediment again coarsens to coarse (3-4 mm), gray pebbles and sand. This layer and the layer from 580 cm to 567 cm contain the coarsest sediment found in the core. There is a diffuse contact around 550 cm, after which the sediment fines to silt and turns to dark brown. The brown silt continues to 528.5 cm, where there is a 4 cm thick layer containing medium sand and many macrofossils. There is then a thin layer of dark brown silt from 524.5 cm to 523.5, after which the sediment returns to the medium sand rich with macrofossils. This layer continues until a diffuse contact around 515 cm. After 515

cm, the sediment is dark brown silt until 483 cm, where the sediment becomes medium sand with many macrofossils. This coarse layer continues until a diffuse contact at 471 cm. After this contact the sediment becomes dark, brown silt until 460 cm, where there is a 1 cm thick layer containing many macrofossils.

After this thin layer the sediment is once again dark, brown silt, until 395 cm. After a gradational contact the sediment is slightly coarser, darker, and contains more macrofossils than the surrounding sediment. This layer ends at 392.5 cm, after which the sediment is again dark brown silt. The silt continues until 333 cm, where there is a slightly coarser layer containing frequent macrofossils. This layer has gradational contacts at the top and bottom, and ends at 331 cm. The dark, brown silt returns at 331 cm and continues until a gradational transition at 326 cm. From 326 cm to 323 cm the grain size increases slightly and the color changes to red/brown. After 323 cm the sediment is again dark, brown silt for the next 9 cm. At 314 cm the grain size increases slightly, after a gradational transition. This layer also has more macrofossils than the nearby sediment, and ends at a diffuse contact around 307 cm. After 307 cm is dark, brown silt until 278 cm, where there is a diffuse contact. From 278 cm to 276 cm there is a layer containing many macrofossils and having a coarser (very fine sand) grain size. After a gradational contact the sediment reverts to dark, brown silt for the next 83.5 cm. At 192.5 cm there is a gradational transition to very fine sand, which continues for 1.5 cm.

At 191 cm the sediment returns to the dark, brown silt. A diffuse contact around 176 cm marks the start of a section containing many macrofossils. This section continues

until a diffuse contact at 171 cm, after which the sediment is the dark, brown silt for the next 20.5 cm. A gradational contact at 150.5 cm marks the beginning of a 1 cm thick layer with slightly larger (very fine sand) grain size and slightly darker color. The sediment is again dark, brown silt from 149.5 cm until a gradational contact at 147 cm, where the sediment changes to darker, very fine sand for 1 cm. A gradational contact at 146 cm ends this coarser layer, and the next 97 cm of sediment is the dark, brown silt. A gradational contact at 49 cm is the beginning of a darker layer with the same grain size (silt) as the sediment below it but containing more macrofossils. This layer is 6 cm thick, ending at a gradational contact at 43 cm. From 43 cm to the top of the core the sediment, once again, the dark, brown silt.

### **A.3 Sterling Pond**

The following lithological description was compiled using data from a previous thesis (Lin, 1996). Starting at the bottom of the Sterling Pond core, the first 8 centimeters from 578 cm to 570 cm consists of coarse glacial till. At 570 cm the sediment changes to gray clay that contains pebbles interspersed within it. This clay continues until 535 cm, where the sediment becomes coarser, containing more silt than the previous layer. This layer from 535 cm to 528 cm also has small fragments of schist. A 2 cm thick layer of coarse, gray sand is next, followed by 5 cm of fine sand. At 521 cm the sediment is still gray but becomes finer, decreasing to clay. This gray clay layer continues until 512.5 cm, where the grain size increases to silt and the color darkens to brown. This section

also contains several thin, black layers within it. After 475 cm the sediment is black, homogenous gyttja all the way to the top of the core.

This section continues until 449.5 cm, where the sediment again darkens to mostly black with four thin, gray layers in it. This darker layer ends at 445 cm, where the

#### **A.4 Ritterbush Pond**

The following lithological description was compiled using data from a previous thesis (Lin, 1996). The core from Ritterbush Pond begins at 525 cm, where the sediment is light brown, dense clay. After 520 cm the clay is interbedded with thin layers of black silt. The alternating clay and silt layers continue until 509 cm, where the sediment changes to mostly black silt. This layer is 4 cm thick, after which the sediment becomes mostly gray and lacks the black silt layers. The gray clay continues until 501 cm, where the sediment changes to a brown-black color for 1 cm. From 501 cm to 500 cm the interbedded black and gray layers return, and from 500 cm to 498 cm the sediment is mostly black. There alternating black and gray layers beginning at 498 cm and continue until 491 cm, where the sediment is black until 488.5 cm. The lithology then fluctuates again between black and gray layers until 486 cm. Here the sediment becomes coarser, gray sand with a few black layers interspersed within the section. The sand continues until 474 cm, where the sediment is brown gray with few black layers. After 10 cm the sediment becomes darker, changing to black with a few thin gray layers. From 464 cm to 458.5 cm the sediment is the opposite of the previous section, with the sediment being mostly gray with a few black layers. At 458.5 cm the sediment switches again to mostly

black with several gray layers. This section ends at 456.5 cm, where the color changes to brown gray.

This section continues until 449.5 cm, where the sediment again darkens to mostly black with four thin, gray layers in it. This darker layer ends at 445 cm, where the sediment is a lighter gray for the next 7 cm. At 438 cm the sediment darkens to a gray black, and then gradationally to just black. This layer ends at 424 cm, at which point begins a fining-upward layer of gray sediment that continues until 401.5 cm. The sediment is then mostly black with several thin (<1 cm) gray layers for the next 52.5 cm until 349 cm of depth. Another fining-upward sequence of sediment begins at 349 cm and lasts until 340.7 cm. The sediment is then mostly black and homogenous until 153.5 cm, where there is an 11 cm thick sandy clay layer. The black sediment returns from 142.5 cm until 137.6 cm, where there is another fining-upward layer that continues until 137.6 cm. From 137.6 cm to 136.7 cm there is another fining upward layer, after which the black, homogenous sediment returns until 112.6 cm. Here there is another fining upward sequence of sediment, which continues until 110.2 cm. The next 10.2 cm consist of the black sediment, after which is a 3.5 cm thick layer of fining upward sediment that ends in clay. After this layer ends at 96.5 cm the rest of the sediment is black gyttja up to the top of the core.



**Appendix B. Laboratory Data**

**B.1 Lake Morey**

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
1	0.31	4.18	13.6	-28.90	
5	0.17	2.36	13.9	-27.04	2.99
10	0.48	5.29	11.0	-26.05	
15	0.91	9.14	10.1	-27.27	
20	0.94	9.26	9.9	-28.69	
25	1.29	12.28	9.5	-29.44	
30	1.30	12.45	9.5	-29.45	
35	1.18	11.58	9.8	-29.83	
40	1.25	12.40	9.9	-29.89	
45	1.17	11.78	10.1	-29.81	
50	0.99	9.93	10.0	-29.62	1.66
55	0.94	9.96	10.6	-29.25	
60	0.92	9.81	10.6	-29.22	
65	0.87	9.16	10.5	-29.48	
70	0.85	8.94	10.5	-29.33	
75	0.81	8.65	10.7	-29.43	
80	0.82	9.03	11.0	-29.49	
85	0.83	8.78	10.5	-29.56	
90	0.86	8.57	10.0	-29.85	
95	0.75	7.53	10.1	-29.40	
100	0.75	7.28	9.7	-29.47	1.73
105	0.67	6.72	10.0	-29.42	
110	0.65	6.57	10.1	-29.20	
115	0.56	6.18	11.1	-28.51	
120	0.28	3.56	12.6	-27.38	2.43
125	0.16	2.46	14.9	-27.39	
130	0.53	7.51	14.1	-27.42	
135	0.50	6.77	13.4	-27.85	
140	0.81	9.75	12.0	-28.41	
145	0.88	10.72	12.2	-28.57	
150	0.77	9.70	12.6	-28.51	2.18
155	0.80	9.53	11.9	-28.59	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
160	0.73	8.86	12.1	-28.90	
165	0.71	7.60	10.7	-29.25	
170	0.70	7.47	10.7	-29.14	
175	0.73	7.98	10.9	-28.81	
180	0.77	8.78	11.4	-28.57	
185	0.76	8.54	11.3	-28.58	
190	0.67	7.61	11.4	-28.49	
195	0.67	8.06	12.1	-28.42	
200	0.56	7.09	12.7	-28.07	2.58
205	0.58	6.72	11.7	-28.48	
210	0.55	6.63	12.0	-28.40	
215	0.49	6.15	12.6	-28.24	
220	0.53	8.61	16.3	-28.13	
225	0.47	7.53	16.0	-28.31	
230	0.43	5.95	13.8	-28.33	
235	0.47	5.36	11.4	-28.36	
240	0.35	3.99	11.3	-27.95	
245	0.56	6.12	10.9	-28.26	
250	0.57	6.51	11.4	-28.60	3.19
255	0.40	5.02	12.4	-28.36	
260	0.66	8.53	13.0	-28.11	
265	1.42	21.59	15.2	-28.06	
270	0.47	7.55	16.0	-27.67	
275	0.36	5.19	14.4	-27.34	
280	0.62	7.16	11.6	-28.69	
285	0.77	11.92	15.4	-28.58	
290	0.59	6.72	11.5	-28.65	
295	0.52	6.28	12.1	-28.26	
300	0.60	7.70	12.9	-28.26	2.80
305	0.56	6.98	12.4	-28.43	
310	0.60	7.32	12.3	-28.54	
315	0.72	9.41	13.1	-28.18	
320	0.61	7.78	12.7	-28.11	
325	0.62	7.99	12.8	-28.03	
330	0.61	7.69	12.7	-28.17	
335	0.59	7.66	12.9	-28.19	
340	0.61	7.54	12.5	-28.29	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
345	0.61	7.70	12.6	-28.13	
350	0.55	6.61	12.0	-28.26	3.07
351	0.56	6.74	12.0		
352	0.54	6.28	11.6		
353	0.53	6.33	12.0		
354	0.54	6.51	12.2		
355	0.56	7.04	12.6	-27.78	
356	0.51	6.57	12.8		
357	0.52	6.83	13.1		
358	0.62	7.58	12.2		
359	0.64	7.83	12.3		
360	0.65	7.84	12.1	-28.33	
361	0.67	7.99	11.9		
362	0.68	7.99	11.8		
363	0.68	7.91	11.6	-28.53	
364	0.66	7.74	11.7		
365	0.65	7.72	11.9	-28.53	
366	0.66	7.87	12.0		
367	0.67	7.97	11.9		
368	0.67	8.30	12.4	-28.34	
369	0.69	8.44	12.2		
370	0.71	8.72	12.3	-28.29	
371	0.75	9.16	12.2		
372	0.78	9.49	12.2		
373	0.76	9.35	12.3	-28.29	
374	0.77	9.75	12.6		
375	0.77	9.51	12.4	-28.22	2.81
376	0.76	9.56	12.6		
377	0.74	9.48	12.8		
378	0.75	9.63	12.9	-28.33	
379	0.72	9.20	12.9		
380	0.71	8.94	12.6	-28.39	
381	0.70	8.93	12.7		
382	0.66	8.45	12.7		
383	0.64	7.85	12.3	-28.54	
384	0.62	7.51	12.2		
385	0.60	7.36	12.3	-28.42	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
386	0.55	6.71	12.1		
387	0.53	6.59	12.4		
388	0.57	6.85	12.1	-27.49	
389	0.58	7.11	12.3		
390	0.60	7.63	12.7	-28.37	
391	0.58	7.16	12.4		
392	0.55	6.73	12.1		
393	0.55	6.55	11.9	-28.38	
394	0.56	6.52	11.7		
395	0.55	6.30	11.5	-28.86	
396	0.55	5.98	10.9		
397	0.54	5.91	10.9		
398	0.52	5.73	10.9	-28.88	
399	0.53	5.78	10.9		
400	0.53	5.83	11.0	-28.51	3.01
401	0.56	6.17	11.1		
402	0.50	6.16	12.3		
403	0.55	6.14	11.2	-28.31	
404	0.54	6.04	11.3		
405	0.52	5.85	11.3	-28.33	
406	0.47	5.41	11.5		
407	0.42	5.00	12.0		
408	0.40	4.93	12.5	-27.62	2.88
409	0.39	5.17	13.3		
410	0.38	4.73	12.4	-27.90	
411	0.40	4.82	12.1		
412	0.41	4.79	11.6		
413	0.39	4.62	11.9	-28.62	
414	0.40	4.64	11.6		
415	0.44	5.14	11.8	-28.75	
416	0.45	5.61	12.6		
417	0.48	5.84	12.2		
418	0.46	5.83	12.6	-28.86	
419	0.43	5.30	12.3		
420	0.42	5.20	12.3	-28.58	
421	0.39	5.03	12.8		
422	0.38	5.10	13.3		

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
423	0.38	5.05	13.4	-28.36	
424	0.35	4.44	12.7		
425	0.37	4.50	12.1	-28.65	
426	0.34	4.25	12.5		
427	0.34	4.07	12.1		
428	0.33	3.98	12.0	-28.84	
429	0.38	4.45	11.7		
430	0.43	4.89	11.4	-29.33	3.35
431	0.41	5.19	12.6	-29.47	
432	0.38	4.61	12.1	-28.51	
433	0.33	4.00	12.0	-29.08	
434	0.26	2.89	11.3	-28.07	
435	0.23	2.43	10.4	-27.95	3.59
436	0.23	2.37	10.3	-27.92	
437	0.22	2.21	9.9	-27.81	
438	0.21	2.12	10.2	-27.51	
439	0.18	1.40	7.9	-26.48	
440	0.21	1.94	9.2	-27.40	3.81
441	0.19	1.95	10.4	-27.50	
442	0.14	2.05	14.7	-26.20	
443	0.15	1.74	11.3	-26.48	
444	0.18	1.73	9.6	-26.84	
445	0.17	1.61	9.6	-26.61	3.76
446	0.16	1.60	10.1	-25.80	
447	0.15	1.32	9.1	-25.95	
448	0.14	1.30	9.2	-26.04	
449	0.08	1.25	14.7	-24.33	
450	0.13	1.37	10.8	-25.15	3.18
451	0.12	1.27	10.2	-25.17	
452	0.08	0.87	10.6	-23.28	
453	0.06	1.03	17.5	-22.28	
454	0.05	0.85	16.6	-22.45	
455	0.11	0.96	8.9	-22.72	3.72
456	0.07	0.67	10.2	-20.34	
457	0.07	0.59	8.2	-19.77	4.27
458	0.05	0.53	10.6	-17.41	
459	0.06	0.59	10.2	-16.20	4.52

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
460	0.07	0.67	9.0	-19.19	
461	0.08	1.16	13.8		
462	0.06	0.56	9.3		
463	0.06	0.62	10.6	-20.49	
464	0.05	0.55	10.2		
465	0.06	0.58	9.3	-17.11	4.42
466	0.04	0.50	12.3		
467	0.04	0.56	12.8		
468	0.08	1.59	20.3		
469	0.15	3.26	21.2		
470	0.10	1.02	9.8	-21.90	3.39
471	0.02	0.30	14.0		
472	0.01	0.42	30.0		
473	0.01	0.39	26.1		
474	0.04	0.55	14.5		
475	0.06	0.55	9.9	-20.32	
476	0.05	0.59	12.0		
477	0.04	0.42	9.9		
478	0.04	0.40	9.0		
479	0.05	0.44	9.2		
480	0.08	0.43	5.5		
481	0.05	0.54	10.8		
482	0.05	0.51	9.6		
483	0.05	0.48	9.9		
484	0.04	0.44	10.9		
485	0.06	0.49	8.9		
486	0.06	0.79	13.8		
487	0.04	0.82	18.3		
488	0.04	0.59	14.1		
489	0.04	0.53	12.6		
490	0.05	0.47	8.7		
491	0.04	0.35	8.3		
492	0.04	0.44	10.7		
493	0.04	0.40	9.8		
494	0.04	0.48	11.0		
495	0.05	0.43	8.7		
496	0.04	0.45	11.5		

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
497	0.05	0.64	13.8		
498	0.05	0.94	19.6		
499	0.03	0.78	23.0		
500	0.05	0.60	12.6		
501	0.07	1.81	25.6		
502	0.10	2.65	26.4		
503	0.06	0.76	13.3		
504	0.05	0.44	9.1		
505	0.06	0.44	8.0		
506	0.03	0.37	12.1		
507	0.05	0.49	10.6		
508	0.04	0.52	11.9		
509	0.04	0.41	11.2		
510	0.06	0.51	7.9		
511	0.04	0.45	10.6		
512	0.05	0.50	10.4		
513	0.04	0.47	10.7		
514	0.05	0.44	9.4		
515	0.08	0.58	7.3		
516	0.04	0.51	12.0		
517	0.04	0.49	11.4		
518	0.03	0.58	17.7		
519	0.04	0.55	13.6		
520	0.06	0.79	12.7		
521	0.03	0.48	13.9		
522	0.03	0.50	15.8		
523	0.02	0.58	23.2		
524	0.03	0.55	17.7		
525	0.04	0.48	11.2		
526	0.04	0.44	12.4		
527	0.04	0.45	11.7		
528	0.04	0.51	14.0		
529	0.03	0.60	20.4		
530	0.03	0.60	21.1		
531	0.02	0.54	24.5		
532	0.02	0.64	28.9		
533	0.02	0.66	33.0		

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
534	0.02	0.66	34.2		
535	0.03	0.61	22.0		
536	0.02	0.62	26.0		
537	0.03	0.61	22.7		
538	0.03	0.56	16.7		
539	0.04	0.51	14.0		
540	0.05	0.36	7.4		
541	0.04	0.45	12.5		
542	0.03	0.46	13.4		
543	0.03	0.49	14.8		
544	0.03	0.58	16.8		
545	0.03	0.50	14.7		
546	0.02	0.58	25.2		
547	0.02	0.63	27.9		
548	0.02	0.68	30.0		
549	0.03	0.64	23.7		
550	0.03	0.61	18.4		
551	0.02	0.67	28.4		
552	0.03	0.71	27.5		
553	0.02	0.72	34.1		
554	0.02	0.70	34.1		
555	0.05	0.59	11.5		
556	0.02	0.68	34.9		
557	0.02	0.67	37.0		
558	0.02	0.70	38.2		
559	0.02	0.73	42.9		
560	0.04	0.66	17.9		
561	0.02	0.67	36.4		
562	0.02	0.66	33.3		
563	0.02	0.66	36.7		
564	0.03	0.73	27.4		



## B.2 Duck Pond

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
5	1.77	23.96	13.6	-29.22	3.54
10	1.87	25.25	13.5	-29.65	
15	2.01	27.42	13.6	-29.84	2.75
20	2.04	27.39	13.5	-30.07	
25	2.19	30.02	13.7	-30.50	
30	2.02	27.37	13.6	-30.43	
35	2.10	27.67	13.2	-30.63	
40	2.11	27.08	12.8	-31.20	
45	2.21	26.64	12.1	-31.88	
50	2.27	30.62	13.5	-32.34	2.84
55	2.18	31.30	14.3	-31.23	
60	2.09	30.11	14.4	-31.48	
65	2.21	31.89	14.4	-31.42	2.81
70	2.01	35.88	17.9	-31.63	
75	2.19	32.14	14.7	-31.31	
80	2.08	30.72	14.8	-31.55	
85	2.20	31.10	14.1	-31.65	
90	2.26	31.92	14.1	-31.47	
95	2.26	31.24	13.8	-31.73	
100	2.21	30.15	13.7	-31.62	2.80
105	1.63	22.18	13.6	-29.21	3.50
110	2.28	30.08	13.2	-31.48	
115	2.13	28.44	13.3	-30.92	3.25
120	2.20	29.87	13.6	-31.25	
125	2.34	31.74	13.6	-31.05	
130	2.34	31.76	13.6	-31.71	
135	2.11	29.45	13.9	-31.57	
140	2.13	28.19	13.2	-31.43	3.19
145	2.18	29.69	13.6	-31.34	
150	2.00	26.73	13.4	-31.03	3.22
155	2.05	28.82	14.1	-31.24	
160	2.16	29.39	13.6	-31.33	
165	2.00	28.72	14.4	-31.40	
170	2.21	30.09	13.6	-31.72	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
175	2.08	29.47	14.1	-31.67	
180	1.91	22.99	12.1	-31.95	
185	2.03	28.13	13.9	-30.61	
190	1.87	26.64	14.2	-31.10	
195	2.08	28.36	13.6	-31.43	
200	2.13	30.08	14.1	-31.64	2.75
205	2.12	29.42	13.9		
210	2.08	29.47	14.2	-31.11	
215	1.98	29.90	15.1	-31.18	2.9
220	2.10	29.24	13.9	-31.36	
225	1.32	18.90	14.3	-30.37	
230	2.10	30.50	14.5	-31.10	
235	2.07	29.56	14.3	-31.12	
240	2.04	29.59	14.5	-30.80	
245	2.06	28.84	14.0	-31.52	
250	2.04	29.45	14.5	-31.14	2.84
255	1.96	30.57	15.6	-31.00	
260	2.02	27.85	13.8	-30.75	
265	2.01	28.12	14.0	-31.03	
270	1.98	28.79	14.6	-31.20	
275	2.02	26.32	13.0	-31.08	
280				-30.86	
285	1.93	27.99	14.5	-31.61	
290	2.08	29.13	14.0	-31.83	
295	1.80	24.24	13.5	-31.14	
300	1.95	27.10	13.9	-31.65	3.25
305	1.90	26.37	13.9	-31.80	2.8
310	2.01	28.32	14.1	-31.56	
315	1.97	27.24	13.9	-31.94	
320	1.98	28.38	14.3	-31.87	
325	1.22	18.94	15.5	-27.49	3.19
330	2.16	30.53	14.2	-32.03	
335	2.19	30.31	13.9	-32.55	
340	2.22	29.88	13.5	-31.45	
345	2.04	27.64	13.6	-32.11	
350	1.87	27.73	14.9	-31.71	3.14
355	2.02	30.28	15.0	-31.59	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
360	2.16	32.18	14.9	-31.63	
365	2.10	29.73	14.2	-32.48	
370	2.06	30.78	14.9	-31.59	
375	2.11	31.41	14.9	-31.96	
380	2.23	32.10	14.4	-32.30	
385	2.14	34.02	15.9	-32.01	
390	2.15	33.46	15.6	-32.17	
395	2.29	33.28	14.6	-31.75	
400	2.06	31.64	15.3	-32.08	2.98
405	2.27	32.44	14.3	-32.13	
410	1.95	29.78	15.3	-31.51	
415	2.34	35.89	15.3	-32.15	
420	2.25	33.67	15.0	-31.84	
425	2.36	33.99	14.4	-32.56	
430	2.39	33.71	14.1	-32.76	
435	2.23	32.42	14.6	-32.83	
440	2.17	31.85	14.7	-32.52	
445	2.34	33.03	14.1	-32.75	
450	2.30	31.79	13.8	-32.76	3.03
453	2.25	32.44	14.4	-32.51	
455	2.28	32.26	14.1	-32.24	
458	1.87	28.07	15.0	-31.84	
460	2.03	29.93	14.7	-32.37	
463	1.86	28.70	15.4	-32.77	
465	2.01	30.90	15.4	-32.68	
468	1.80	27.97	15.5	-32.67	
470	1.81	27.58	15.2	-33.10	2.67
473	1.98	32.46	16.4	-32.65	
475	1.24	21.67	17.5	-31.34	
477	0.61	10.33	16.9	-27.38	
478	0.68	11.90	17.5	-27.69	
479	0.58	9.55	16.4	-27.66	
480	0.57	11.69	20.5	-27.80	2.64
481	0.63	10.36	16.5	-27.98	
483	0.59	10.17	17.1	-28.41	
485	2.27	32.55	14.3	-34.72	3.16
488	2.10	32.25	15.4	-34.15	

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
490	2.20	33.37	15.2	-34.39	
493	2.10	33.45	15.9	-33.25	
495	2.39	36.20	15.1	-33.91	
498	2.04	29.90	14.7	-33.99	
500	2.11	34.19	16.2	-33.85	2.29
503	1.79	25.82	14.4	-33.93	
505	2.12	32.43	15.3	-34.38	
508	2.05	32.32	15.8	-34.21	
510	2.25	33.79	15.0	-34.89	
513	1.94	28.25	14.6	-35.17	
515	1.90	28.20	14.8	-35.16	2.18
517	0.90	13.34	14.9	-30.60	
518	0.75	13.55	18.0	-29.51	
519	0.66	11.32	17.0	-28.46	2.33
520	0.57	9.65	16.9	-28.80	
521	0.88	20.39	23.2	-28.85	
523	0.72	14.38	20.0	-27.98	
525	1.53	22.94	15.0	-33.66	
528	0.66	11.23	16.9	-28.21	2.67
530	2.02	30.86	15.3	-34.50	
533	2.00	28.39	14.2	-35.20	
535	1.87	26.73	14.3	-35.34	2.22
538	1.49	20.36	13.7	-34.80	
540	1.61	23.70	14.7	-34.55	2.18
543	1.18	19.14	16.3	-34.01	
545	1.23	17.85	14.5	-33.01	2.39
548	0.93	13.10	14.0	-29.86	
550	0.54	6.89	12.8	-31.18	2.96
551	0.38	3.96	10.5	-30.82	
552	0.40	4.75	11.9	-30.86	
553	0.39	4.32	11.0	-30.21	
554	0.37	4.84	13.1	-29.62	
555	0.15	2.75	18.3		
556	0.11	1.78	16.4		
557	0.49	6.10	12.4	-31.16	2.67
559	0.35	5.68	16.2		2.67
560	0.36	4.32	12.0	-30.50	2.66

Depth (cm)	%N	%C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
561	0.38	4.12	10.8	-29.86	
562	0.32	4.28	13.4	-28.66	
563	0.29	4.10	14.1	-26.60	2.29
565	0.24	3.46	14.4	-27.23	
566	0.26	3.96	15.2	-26.58	
567	0.19	4.06	21.4	-21.84	2.25
570	0.01	1.63	163.0	-4.45	
575	0.01	2.14	214.0	-3.87	
580	0.01	1.97	197.0	-8.83	

### B.3 Sterling Pond

From Lini (unpublished)

Depth (cm)	% N	% C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
2.5	2.30	35.20	15.3	-28.68	1.94
7.5	2.50	38.50	15.2	-28.68	1.64
12.5	2.60	38.20	14.4	-28.64	
17.5	2.40	37.20	15.7	-28.41	
22.5	2.40	38.20	16.1	-28.54	
32.5	2.50	39.80	15.7	-28.43	
42.5	2.50	38.90	15.7	-28.30	
52.5	2.50	38.30	15.3	-28.46	3.03
62.5	2.50	38.60	15.7	-28.52	
72.5	2.40	36.10	15.3	-28.64	
100	2.30	35.90	15.9	-28.64	2.89
110	2.40	38.10	15.8	-29.01	
120	2.40	37.10	15.3	-29.05	
130	2.50	37.00	15.1	-28.87	
140	2.30	35.50	15.4	-28.84	
150	2.30	34.90	15.3	-28.98	2.52
160	2.40	36.90	15.4	-29.05	
161	1.90	46.40	25.0		1.90
170	2.20	35.70	15.9	-28.54	
180	2.40	36.10	15.2	-28.89	
190	2.30	36.20	15.7	-28.65	
200	2.20	34.30	15.7	-28.52	2.91
210	2.20	36.80	16.4	-28.70	
220	2.30	34.40	15.0	-28.52	
230	2.20	35.90	16.6	-28.63	
240	2.20	33.00	14.9	-29.00	
250	2.20	33.70	15.1	-28.95	3.32
260	2.20	34.50	15.4	-29.00	
270	2.20	33.80	15.2	-28.98	
280	2.40	37.70	16.0	-28.98	
290	2.40	35.00	14.4	-29.36	
300	2.20	37.60	16.9	-29.30	2.29
310	1.90	31.70	16.4	-29.43	

Depth (cm)	% N	% C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
320	2.20	35.00	15.5	-28.97	
330	2.20	34.80	16.1	-28.75	
340	2.10	33.00	16.0	-28.91	
350	2.20	33.50	15.6	-29.04	2.71
360	1.90	32.50	16.8	-29.14	
370	2.00	32.80	16.6	-28.98	
380	1.90	32.30	16.9	-28.94	
390	1.90	31.70	16.6	-28.99	
400	1.90	31.70	16.9	-29.21	2.74
410	2.40	35.60	15.1	-29.53	
420	1.90	32.70	17.3	-29.84	
430	2.20	36.60	17.0	-30.53	2.32
440	2.00	31.10	15.6	-30.62	2.53
450	1.90	29.10	15.6	-29.62	2.46
455	1.80	27.00	14.7	-29.38	
460	1.40	20.60	15.1	-28.79	2.27
470	0.90	9.70	11.4	-26.75	2.57
475	0.40	4.90	11.6	-27.46	
480	0.50	5.50	11.0	-27.32	2.41
485	0.30	3.20	10.5	-27.30	
490	0.20	3.00	13.6	-25.10	1.27
502	0.10	0.90	9.9	-23.43	1.20
506	0.10	0.60	9.4	-23.76	
510	0.10	0.70	9.2	-24.24	1.30
513				-20.34	
515				-20.75	
518				-21.68	
520				-21.77	
522				-22.33	
524				-23.50	
527				-23.87	
534				-21.97	
536				-22.25	

#### B.4 Ritterbush Pond

Depth (cm)	% N	% C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
0	1.34	17.90	13.4	-30.31	3.79
10	1.27	17.16	13.5	-30.15	
20	1.71	22.09	12.9	-31.31	
30	1.60	21.11	13.2	-30.63	3.58
40	0.65	10.45	16.0	-27.58	3.97
50	1.62	21.43	13.3	-30.33	3.35
60	1.22	18.04	14.7	-29.45	
70	1.31	17.74	13.5	-30.30	
80	1.49	18.98	12.7	-31.45	
90	1.77	22.65	12.8	-31.40	3.26
96				-31.69	
100	0.32	5.27	16.7	-27.45	3.22
114	1.45	19.88	13.7	-30.70	3.36
124	1.55	20.12	13.0	-31.12	
134	1.01	13.67	13.6	-30.25	
142				-31.52	
144	1.05	13.72	13.1	-30.64	3.47
154	0.24	3.85	16.0	-27.00	4.22
154.5				-29.50	
164	1.33	18.37	13.8	-30.27	3.39
174	1.33	17.59	13.2	-30.93	
184	1.34	17.35	12.9	-31.34	
194	1.40	18.06	12.9	-31.49	
204	1.05	14.51	13.8	-30.51	3.35
214	0.76	10.79	14.3	-30.14	
220	1.06	15.28	14.4	-30.57	
230	1.25	17.15	13.8	-31.13	
240	1.27	19.45	15.3	-30.24	
250	1.00	14.90	14.9	-30.14	3.09
260	1.11	15.97	14.4	-30.86	
270	1.26	17.74	14.1	-31.29	
280	1.07	16.03	15.0	-30.21	
290	1.20	17.01	14.2	-31.00	



Depth (cm)	% N	% C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
300	1.26	17.50	13.9	-31.31	2.88
310	1.22	17.62	14.4	-31.85	
320	1.06	15.80	14.9	-31.05	
330	0.74	10.33	13.9	-31.83	
339	0.72	10.05	14.0	-32.02	
340	1.61	21.84	13.5	-31.65	
350	1.32	18.41	14.0	-32.39	2.56
360	1.11	16.39	14.7	-31.25	
370	1.56	23.91	15.3	-31.92	
380	1.76	25.78	14.6	-31.93	
390	0.29	4.65	16.2	-30.68	
400	0.86	13.30	15.5	-31.26	2.27
410	0.38	5.96	15.6	-30.16	
416				-32.70	1.94
420	0.06	1.01	16.6	-25.88	2.98
425	0.99	14.99	15.2	-31.81	2.21
441.5	1.03	17.72	17.2	-32.69	
443.5	1.05	20.45	19.6	-32.97	
446	0.91	19.08	20.9	-32.91	
448	1.01	19.60	19.3	-34.22	1.84
453	0.10	1.75	17.1	-27.94	2.44
456.5	0.82	16.92	20.7	-34.04	1.78
458.5	0.46	8.90	19.5	-33.47	
460	0.11	1.85	17.0	-26.93	2.26
462	0.65	14.03	21.6	-33.08	
463.5			18.8	-33.55	1.10
466.5	0.04	0.67	15.0	-27.32	
468	0.07	1.13	17.4	-28.92	
469.5	0.17	3.08	18.1	-27.89	2.69
472.5	0.88	19.31	21.9	-33.49	
475.5	0.77	15.94	20.7	-34.88	0.71
479	0.34	5.23	15.2	-34.64	
483	0.31	4.49	14.4	-30.66	1.32
486.5	0.16	1.88	12.1	-33.10	1.56
487.5	0.05	0.58	11.9	-29.55	1.66
491.5	0.08	0.78	10.3	-29.70	
496.5				-24.10	

Depth (cm)	% N	% C	C/N	$\delta^{13}\text{C}$	$\delta^{15}\text{N}$
500.5	0.03	0.34	11.2	-25.70	1.35
502	0.02	0.19	12.1	-24.46	
503.5	0.02	0.24	14.0	-24.94	
506	0.04	0.39	9.7	-25.82	
506.5	0.03	0.39	12.3	-26.05	
508.5	0.02	0.26	11.8	-23.88	
510	0.02	0.25	14.2	-25.15	
513	0.02	0.31	13.1	-25.19	
518	0.02	0.30	14.2	-24.82	
523.5	0.02	0.26	15.8	-25.39	