

**A STABLE ISOTOPE PALEOCLIMATE RECORD OF THE LATE-
GLACIAL/INTERGLACIAL TRANSITION FROM LOUGH INCHIQUIN,
WESTERN IRELAND**

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A Stable Isotope Paleoclimate Record of the Late-Glacial/Interglacial Transition from Lough Inchiquin, Western Ireland

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Abstract

High-resolution stable isotope analysis of a sediment core from Lough Inchiquin, western Ireland, reveals the presence of the Bølling-Allerød/Younger Dryas/Preboreal post-glacial climate sequence. Changes in lithology and stable isotope values in the carbonate sediment of this small (1.05 km²) lake provide evidence for temperature and aridity variation and a rise of vegetation from a barren post-glacial landscape. The record at Lough Inchiquin indicates a warming of approximately 7.8°C leading into the Bølling and a minimum temperature change of about 7.7 °C associated with the Younger Dryas stadial. Comparisons to δ¹⁸O records from other sites in Ireland and from the GRIP ice core indicate that the environment at Lough Inchiquin changed in response to both local and regional climatic factors. Although a lack of consistent covariation in δ¹³C and δ¹⁸O at this location indicates that different environmental controls affect these isotope ratios, shared cyclicities found by spectral analysis suggest that some factors are influential to both records.

Keywords: Younger Dryas, paleoclimate, Ireland, stable isotopes, lacustrine sediments

Introduction:

The shift from the Last Glacial Maximum to the current Interglacial period is characterized by several climatic oscillations. Most prominent in Western European paleoclimate records and ice cores, the Bølling-Allerød/Younger Dryas/Preboreal sequence of climate shifts is widely recognized in climatic proxy records. In addition to the classical European locations and Greenland, evidence for climatic fluctuations connected to this sequence has been found in Alaska (Engstrom et al., 1990), the Northeastern United States (Anderson et al., 1997), off the California coast (Hendy et al., 2002), in the Tyrrhenian Sea (Kallel et al., 1997), and even asynchronously in Japan (Nakagawa et al., 2003) and Lake Baikal (Prokopenko et al., 1999). However, even the most dramatic and extensively studied of these climate swings, the Younger Dryas, has not been adequately characterized. The apparently global nature of this event has been called into question by Bennett et al. (2000), who report the absence of a Younger Dryas event in a southern hemisphere paleoclimate record.

Even if the Younger Dryas is not global, the wide distribution of sites that indicate climate fluctuation is significant because although they are poorly understood these climate shifts are thought to be closely tied to global ocean and atmospheric circulation (Broecker, 2000). A greater understanding of these abrupt variations, and the impact human activities may be having on global climate (Alley, 2000a), can be gained by studying the magnitude and distribution of these abrupt events. New high-resolution studies of continental paleoclimate in Europe, where the events are most clearly defined, contribute to an overall picture of the conditions that produced climate instability.

The purpose of this study is to develop a paleoclimate record to help illuminate the nature of the glacial-interglacial transition in western Ireland. In July of 2002, I traveled to western Ireland as a member of a KECK Consortium group whose aim was to study post-glacial paleoclimate in lacustrine sediments. Together, the members of this group took sediment cores from various locations for different types of analysis. My focus as a member of this group was a high-resolution stable isotope paleoclimate analysis of sediments representing the transition from the late glacial period to the Holocene.

Background and Setting:

Numerous paleoclimate records using several different proxies have been combined to develop the understanding currently held of paleoclimate since the last glacial maximum. These include: Greenland ice cores (Alley, 2000a; Alley, 2000b; Dansgaard et al., 1993), pollen records (Ahlberg et al., 1996; Isarin et al., 1999; O'Connell et al., 1999; Rundgren, 1995), ocean sediments (Bond et al., 1993), lacustrine stable isotopes (Mayer et al., 1999), beetle parts (Atkinson et al., 1987), midge fossils (Walker et al., 1991) and speleothems (McDermott et al., 2001). Each of these proxies has been found to exhibit part or all of the Bølling-Allerød/Younger Dryas/Preboreal sequence, as follows.

Although minimum glacial temperatures had begun to moderate by the Pleniglacial (Oldest Dryas), the earliest significant postglacial warming is represented by the Bølling-Allerød interstadial. This period began about 14,670 cal. Yr BP (Stuiver et al., 1995) when temperatures increased rapidly. Estimates of summer temperatures during this period, also known as the Windermere interstadial, indicate that peak

temperatures were at least as warm as the modern (Ingolfsson et al., 1997). The Bølling-Allerød is interrupted briefly by the Older Dryas stadial in some high-resolution records (Stuiver et al., 1995).

The Younger Dryas (Loch Lomond) stadial is the most notable and well-studied of the excursions. In less than 200 yrs, summer temperatures in Norway dropped by as much as 5-6° C (Mangerud, 1987), while the summit of Greenland may have been 15±3°C colder than today (Severinghaus et al., 1998). Ahlberg et al. (1996) estimate a temperature decrease of 12°C at Red Bog and Lough Gur in western Ireland. Isarin and Bohncke (1999) suggest a minimum mean July temperature of 11° C in southwestern Ireland during the Younger Dryas. Different studies indicate slightly variable timespans and dates for the Younger Dryas, but climate may have remained cold for as long as 1,500 years. Age estimates for the beginning of the Younger Dryas range between 13,000 cal. yr BP (Smith et al., 1997) and 12,550 yr BP (O'Connell et al., 1999). Most studies place the return to warm conditions at around 11,500 cal. yr. BP (Alley, 2000b; Bjorck et al., 1996; O'Connell et al., 1999; Roberts, 1998).

The Younger Dryas period was succeeded by the Preboreal warming. Once underway, the transition from the Younger Dryas to the Preboreal was very rapid (Stromberg, 1994) taking place in as little as 40 years (Taylor et al., 1997). However, this warming phase was interrupted briefly about three hundred years after it started. At 11,170 yr. BP, Bjorck et al. (1996) show evidence for a brief cold period which they refer to as the Preboreal Oscillation (PBO). After the PBO there was gradual warming until Holocene temperatures were reached (Bjorck et al., 1996). Temperatures following this

event remained warm through the beginning of the Holocene 9,500 cal. yr. BP (Alley et al., 1997).

Some studies have recorded the presence of a Holocene cold event 8,200 yr BP. The cold period, taking place between 8,400 and 8,000 years ago, lasted only about 400 years (Alley et al., 1997; Baldini et al., 2002; Barber et al., 1999; McDermott et al., 2001). Maximum cold of this event lasted only about 100 years with a peak cooling of $6\pm 2^\circ\text{C}$ (Alley et al., 1997). Regardless of scope, this event illustrates that the “stable” Holocene continued to have abrupt climatic shifts.

Western Ireland is an ideal setting for study of climate since the last glacial period using stable isotopes (Ahlberg et al., 1996). Many bogs and fens are near small lakes and overlie carbonate lacustrine sediments from lakes which have filled in, providing ready locations for carbonate cores. Much of Ireland was covered by the last (Wisconsinan) glaciation and sediment records from most lakes begin at the end of the glacial period (Mitchell et al., 1996). The dominant wind direction has remained roughly the same since last glacial period (Ahlberg et al., 1996; Isarin et al., 1997), which is significant because it indicates that moisture came from the same area throughout the time period covered by this study.

Lough Inchiquin, the sample site for this study, is located south of Galway in Western Ireland (Fig. 1). It is found on Upper Carboniferous limestones near the bedrock contact with a unit of upper Avonian shales and sandstones (Fig.1). A small river, not shown on the map, enters the lake on its southwest side but has little influence on processes in the lake. The lake is fairly small (about 1.05 km^2), and continues to deposit carbonate sediment today. Vegetation at the location of the core consists of grasses and



Figure 1: Map of Ireland Showing the locations of Late-glacial core sites discussed in this study. By number, sites are: 1) Lough Inchiquin, 2) Tory Hill, 3) Lough Gur and Red Bog, 4) Belle Lake, and 5) Coolteen. The inset shows a geologic map modified from O'Brian (1962) with a more detailed view of the area around Lough Inchiquin. Location of the LINC-1 core is indicated by a red star.

reeds, although trees grow in other locations around the lake (Fig. 2). The typical post-glacial sequence has been well-illustrated in $\delta^{18}\text{O}$ values and pollen assemblages from sites near Lough Inchiquin in previous studies (Ahlberg et al., 1996; O'Connell et al., 1999).

Methods:

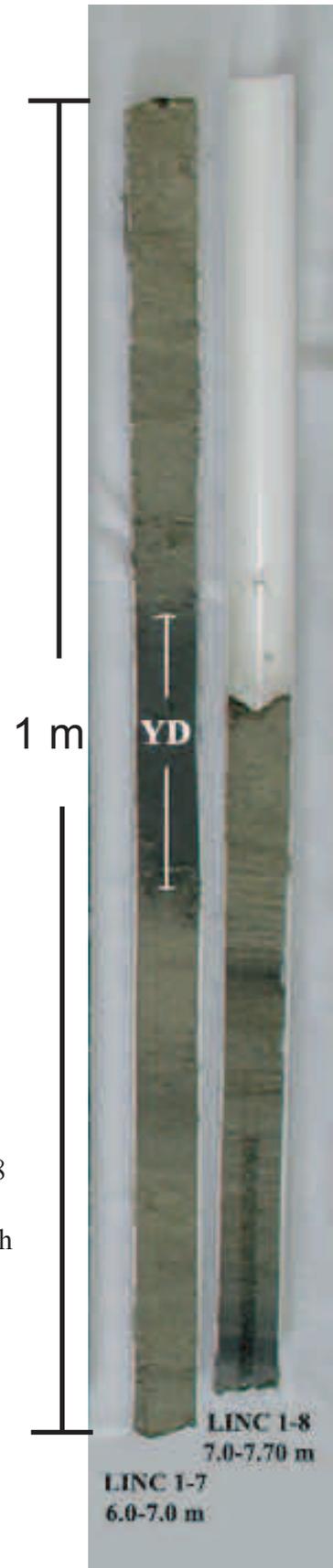
After exploratory coring with a Dutch corer, several piston cores of lake sediments from three locations were taken. In most cases the sediments extended from the present back to the Younger Dryas interval (as indicated by a lithology change from marl to clay) before bedrock was encountered. At Lough Inchiquin, approximately an additional meter of sediment was recovered below the Younger Dryas clay. For my analysis, I used the bottom two core sections (LINC 1-7 and 1-8) from the LINC 1 core (Fig. 3) obtained at the Lough Inchiquin location. These two core sections were selected because they provided approximately 1.5 m of sediment encompassing the Younger Dryas (marked by a distinctive clay-rich section) and sediments from both prior to and after the cold period (Fig. 3). The sediments were sampled on a 10 cm interval for loss on ignition (LOI) and a 2 mm interval for stable isotope analysis.

Along with colleagues from my KECK project, I ran LOI for the sediments at the National University of Ireland, Galway. Samples were burned in a furnace at approximately 500° C to combust organic plant matter (organics) and 1000° C to burn carbonate, and what remained after this step was considered non-combustible (clays, quartz). Before analysis, samples were heated in a drying oven to remove water weight. Values for loss on ignition were calculated and reported as percentages of the total weight of sediment.



Figure 2: The Lough Inchiquin core site. The area with the trampled grasses (red circle) is the core location. The lake's edge is just behind the core where taller grasses are growing.

Figure 3: Core section 1-7 and 1-8 from the LINC 1 core. LINC 1-7 is one meter in length, while LINC 1-8 is only about 0.5 m because bedrock was encountered. The Younger Dryas period is evident in a thick clay rich section (YD) and two other obviously clay-rich intervals occur in LINC 1-8.



This study was principally focused on carbon and oxygen stable isotope analyses. Both of these elements are abundant and climate processes preferentially segregate heavy from light isotopes in the environment, making it possible for past conditions to be deduced from isotope ratios. The abundant presence of these elements in lacustrine marls makes such sediments ideal for estimations of past continental climate (Hays, 1991).

To provide a basis of comparison, stable isotope ratios are typically reported relative to standard values. The difference between the sample value and the standard, known as the delta (δ) value, is given as a per mil (‰) value and calculated according to the equation

$$\delta = \left[\frac{(R_{Sample} - R_{Std})}{R_{Std}} \right] \times 1000$$

where R stands for the ratio of the heavy isotope to the light isotope. Positive delta values, then, indicate that the sample is more enriched in the heavy isotope than the standard while negative delta values indicate that the sample is depleted in the heavy isotope relative to the standard.

Stable isotopes of carbon and oxygen from the LINC 1 core were analyzed at the University of Saskatchewan in Saskatoon, Canada. Prior to analysis samples were heated *in vacuo* at 200 ° C to remove volatiles. Analysis was performed by reacting samples with 103% phosphoric acid at 70 ° C on a Kiel III carbonate preparation device which was coupled to a Thermo Finnigan Mat 253 mass spectrometer. Results have been corrected for ¹⁷O contribution, and have a 0.1 per mil or lower standard deviation. Both

oxygen and carbon stable isotope values have been calculated relative to the VPDB (Vienna Peedee Belemnite) international standard.

Depths in the LINC 1-8 core have been corrected to account for a 20 cm gap between the cores. No dates have yet been obtained directly from the LINC 1 core. However, two different methods were used to obtain a working time scale for this study. Constraints were developed for LINC 1 based on a lithological comparison to a radiocarbon dated core from a study done at nearby Tory Hill (O'Connell et al., 1999). Additionally, using a date of 11,500 Cal. Yr. BP (Roberts, 1998) for the termination of the Younger Dryas, a sedimentation rate was calculated from the top of the core (0 m) to the end of the Younger Dryas (6.404 m) in the LINC 1 core. Assuming a constant sedimentation rate, ages were then calculated for the sediments above and below the Younger Dryas.

To better understand the implications of the stable isotope results, the computer application Matlab was used to analyze data from the section of the core between 6.582 and 7.550 m. The values of duplicate meter levels were averaged prior to analysis. To convert the depth to time, a uniform sedimentation rate between the ages obtained by comparison to the Tory Hill core (O'Connell et al., 1999) was assumed. Carbon and oxygen isotope data were detrended by subtracting their mean values and plotted, the $\delta^{13}\text{C}/\delta^{18}\text{O}$ ratio at each point was plotted against time, and spectral analysis was performed using a fast Fourier transform. In some spectral analyses, a bandpass filter was used to restrict the range of frequencies considered by the program in order to focus on significant peaks. In order to better distinguish significant from inconsequential peaks, a curve was created using two standard deviations from the averages of fast

Fourier transforms of 500 to 1000 randomly generated data sets. This curve (a 95% confidence level of significance) was overlaid on the peaks generated by the LINC 1 data set. Running averages based on depth were also used to smooth the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ curves to make it easier to distinguish trends in the data.

Results:

Correlation of the LINC 1 core with the Tory Hill core (Fig.1) (O'Connell et al., 1999) gives the base of LINC 1 (7.55 m) an age of 15,000 cal. yr BP, (Fig. 4). The Younger Dryas period is also correlated, placing the peak non-combustible values of the LOI results at 12,000 cal. yr BP. O'Connell (1999) gives the entire Younger Dryas a timespan from 11,500-12,500 cal. yr BP. A third age of 8,200 cal. yr BP is correlated to the LINC 1 core at about 4.7 m depth, above the analyzed sediments. Although this does not give a direct age to any part of the section, it indicates that the top of the LINC 1 core section is older than 8,200 cal. yr BP.

Calculated ages based on a constant sedimentation rate are similar to those obtained by comparison to O'Connell's core, but not identical (Table 1). At 4.7 m in the LINC 1 core, which by correlation is placed at an age of 8.2 ka cal. BP, the calculated age is 8.543 ka. At the top of my core section, an age of 10,799 cal. yr BP is obtained. Using calculations carried into the Younger Dryas from the top, the resulting age for the onset of the Younger Dryas is only 11,829 cal. yr BP (Table 1, Column 2). To avoid this problem, ages were calculated with assumed ages for the Younger Dryas (6.594m - 6.404m) of 11,500 cal. yr. BP to 12,500 cal. yr BP (Table 1, Column 3) and 11,500 cal. yr BP to 13,000 cal. yr BP (Table 1, Column 4). These calculation produced ages of 14,158 cal. yr BP. and 14,658 cal. yr BP. respectively for the base of the core.

Table 1

	End of YD at 11,500 cal. yr BP	YD from 12,500-11,500 cal. yr BP	YD from 13,000-11,500 cal. yr BP	O'Connell correlated ages	O'Connell calculated ages
Depth (m)	Age (cal. yr BP)				
4.700	8544	8544	8544	8200	
5.000	9064	9064	9064		
5.500	9932	9932	9932		
6.000	10799	10799	10799		
6.100	10973	10973	10973		
6.200	11146	11146	11146		
6.300	11320	11320	11320		
6.404	11500	11500	11500		
6.500	11667			12000	12000
6.594	11830	12500	13000		12268.57
6.600	11840	12510	13010		12285.71
6.700	12014	12684	13184		12571.43
6.800	12187	12857	13357		12857.14
6.900	12361	13031	13531		13142.86
7.000	12534	13204	13704		13428.57
7.100	12708	13378	13878		13714.29
7.200	12881	13551	14051		14000
7.300	13055	13725	14225		14285.71
7.400	13228	13898	14398		14571.43
7.500	13402	14072	14572		14857.14
7.550	13488	14159	14659	15000	15000

Table 1: Ages calculated with a constant sedimentation rate (columns 2-4) correlated by depth with those obtained from the Tory Hill core (O'Connell, 1999). Ages in column 5 were calculated using a constant sedimentation rate between the O'Connell correlated ages for a conversion of depth to time in spectral analysis.

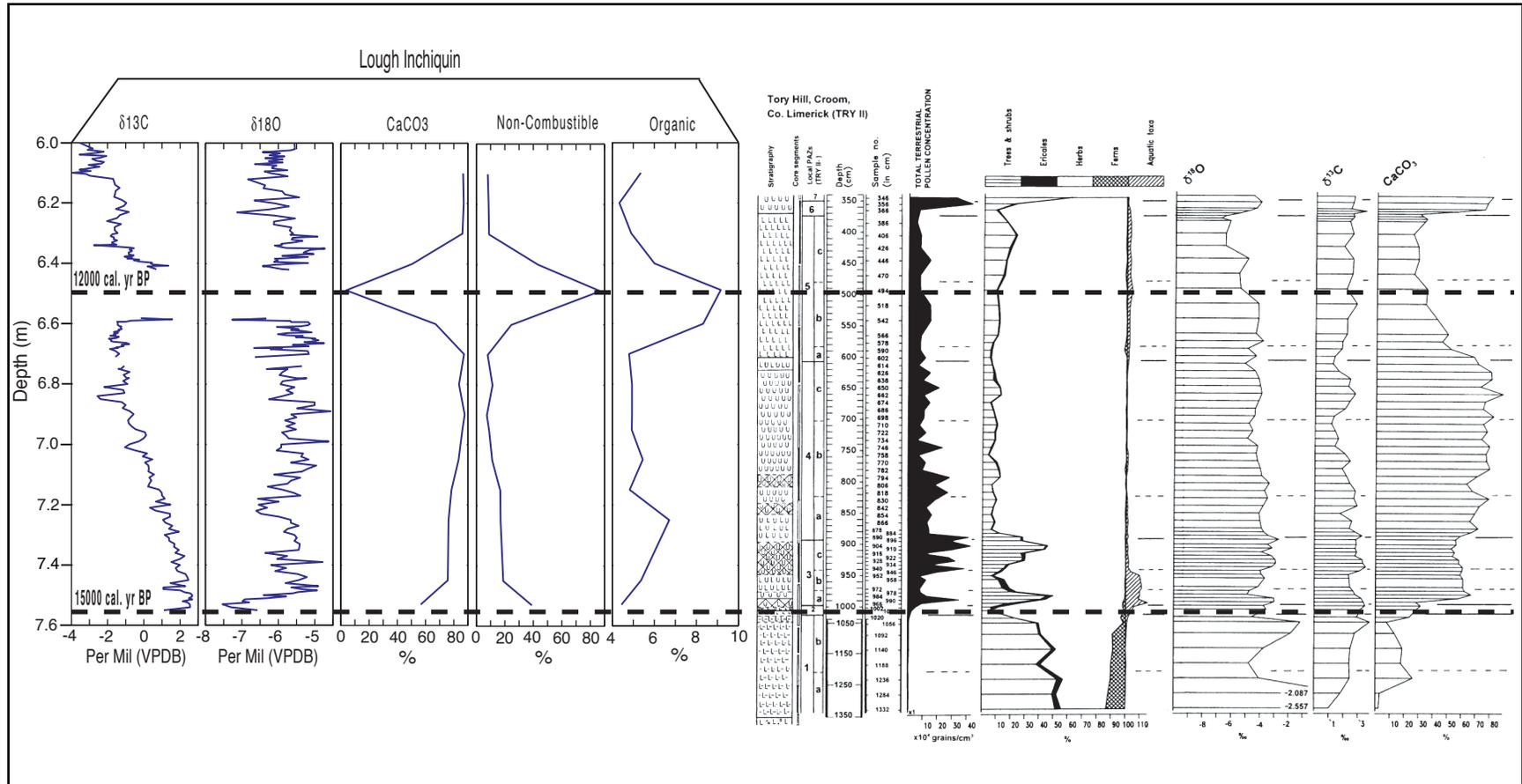


Figure 4: Comparison of LINC 1 stable isotope and LOI results to stable isotope, carbonate content, and pollen data from the Tory Hill sediment core taken by O'Connell et al. (1999). Ages (denoted by dashed lines) are from calibrated radiocarbon dates obtained by O'Connell et al. for the Tory Hill core. Correlation is based on lithological similarity between the cores, illustrated in this figure by similarity between the CaCO₃ curves. δ¹³C and δ¹⁸O show some broad similarities but also numerous small scale variations between the two cores. There does not appear to be a strong correlation between the types of plants indicated by the Tory Hill pollen data and change in δ¹³C from the LINC 1 core.

The general similarity between calculated and correlated dates bolsters confidence in the correlation to Tory Hill. However, there are also strong indications that the assumption of uniform sedimentation rate used in calculations is incorrect. The young ages calculated for the base of the core imply that sedimentation actually occurred more slowly than the uniform sedimentation rate supposed. This is especially true during the Younger Dryas, when calculations with uniform sedimentation rate imply a timespan of less than 400 years for an event that lasted at least a millennium. Based on these clear problems with the calculated dates, the ages from O'Connell (1999) are given preference for the discussion of my data.

Loss on Ignition

Although lithological variations are evident from the color and texture of the core (Fig. 3), quantification of these changes is provided by loss on ignition results (Fig.4). A two-tiered transition from clay-rich to carbonate-rich sediment takes place at the bottom of the core. Between 7.53 and 7.45 m, values shift from 56.6% CaCO₃ and 38.96% non-combustible to 75.74% CaCO₃ and 18.89% non-combustible. Organics show a brief increase at 7.3 m, but return to about 4.83% by 7.15 m. Between 7.05 and 7.15 m, a second carbonate increase from 78.37% to 83.39% takes place. Although the values from the very bottom of the core may have some element of contamination, it is clear that an upward-increasing trend in carbonate is present in the bottom half-meter of the core. At 6.8 m a small clay-rich interval is indicated by increased non-combustible values.

A large increase in non-combustible sediments (from 8.95% to 87.01%) and a dramatic drop in carbonate values (86.14% to 3.84%) takes place between 6.488 and 6.3 m. This shift, the most significant change in the loss on ignition results, distinctly shows

the clay-rich nature of the Younger Dryas interval. The Younger Dryas is also the location of the most dramatic shift in organic values, which rise to a high of 9.16% from a background value of about 5%.

Carbon Isotopes

The most prominent trend in LINC 1 carbon isotopes is a strong overall shift towards more negative values (Fig. 4, 5). A maximum of 2.67 ‰ at the bottom of the core shifts gradually to a minimum of -3.93 ‰ near the top, producing a 6.6 ‰ range of values. The trend in $\delta^{13}\text{C}$ associated with the Younger Dryas period, between 6.608 m and 6.34 m, is also very robust. Although no data were obtained between 6.58 m and 6.43 m, the values surrounding the data gap show that a positive shift in $\delta^{13}\text{C}$ takes place during that period with a rapid onset and abrupt termination. Leading into the gap, between 6.652 m and 6.594 m, $\delta^{13}\text{C}$ increases from -1.93 ‰ to -0.17 ‰ while above the gap, from 6.42 m to 6.34 m, $\delta^{13}\text{C}$ decreases from 0.69 ‰ to -2.75 ‰. Significant smaller scale variation in the $\delta^{13}\text{C}$ are seen in negative excursions between 6.87-6.78 m, 7.05-6.97m, and 6.12-6.07 m. A positive deviation of 1.56 ‰ takes place between the base of the core and 7.5 m, and another variation that could be considered a slight positive excursion exists between 6.29 - 6.12 m.

Oxygen Isotopes

No overall trend is present in the $\delta^{18}\text{O}$ values, although a number of greater than 1‰ shifts take place (Fig.4, 5). The base of the core, between 7.55 m and 7.484 m, shows a dramatic negative peak. The minimum $\delta^{18}\text{O}$ value of this peak (-7.53 per mil at 7.53 m) is the lowest value in the core and has a $\delta^{18}\text{O}$ of 2.59 ‰ lower than the value only 4.6 cm later at 7.484 m. Between 6.604 m and 6.35 m, a negative shift is indicated

Three-point running average of carbon and oxygen stable isotopes

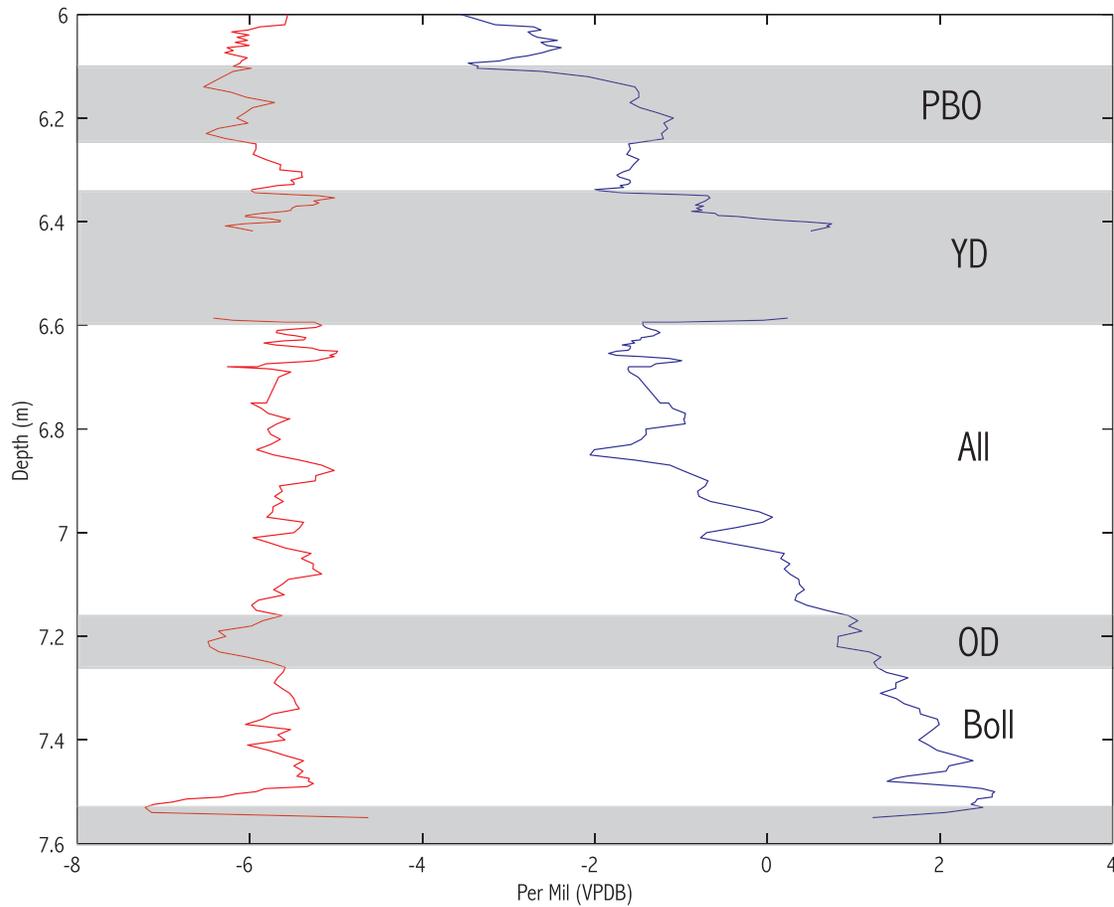


Figure 5: A three-point running average of the isotope values smooths the curve and helps to distinguish trends. $\delta^{18}\text{O}$ (left) is in red while $\delta^{13}\text{C}$ (right) is in blue. Major climatic oscillations in the Lough Inchiquin core are marked: Boll=Bolling, OD=Older Dryas, All=Allerod, YD=Younger Dryas, PBO=Preboreal Oscillation.

leading in and out of the Younger Dryas section of the core associated with a minimum of 2.55 ‰ change. Most of this change is abrupt (between 6.604 m and 6.586 m) leading into the Younger Dryas and the return to higher $\delta^{18}\text{O}$ values following the Younger Dryas is also rapid in the LINC 1 core. Less negative departures from typical values take place between 6.29-6.11 m and 7.27-7.09 m and $\delta^{18}\text{O}$ shows frequent oscillations of 1 ‰ or less throughout the course of the core section.

Spectral Analysis

Spectral analysis found cycles with matching periods in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, but both also showed significant unmatched peaks (Fig. 6). Initial analysis indicated similar dominant peaks at about 4.2 ka, but these were rejected because they represent a longer time period than the data set analyzed. A bandpass filter eliminated this peak, allowing smaller peaks in the carbon isotopes to be distinguished. With the irrelevant long-period peak removed, carbon isotopes showed a dominant spectral peak with a period of approximately 1.7 kyr, and three much smaller peaks with periods of ~1.0 kyr, ~770 yr, and 575 years. $\delta^{18}\text{O}$ showed more numerous peaks than the $\delta^{13}\text{C}$, but included in them were close matches for the peaks previously mentioned. The strongest peaks had periods of ~1.0 kyr and 575 yr, but peaks were also present with periods of ~770 yr, ~400 yr, ~296 yr, and 126 yr. Changing the frequencies of the bandpass filter illustrates that the cycles represented by the periods discussed here are consistently found. Additionally, peaks representing all of these cycles were reliably shown to be above a 95% confidence level-curve. Plotting the filtered $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ curves on top of each other shows that variance is predominantly out of phase although they do occasionally covary (Fig. 7). A

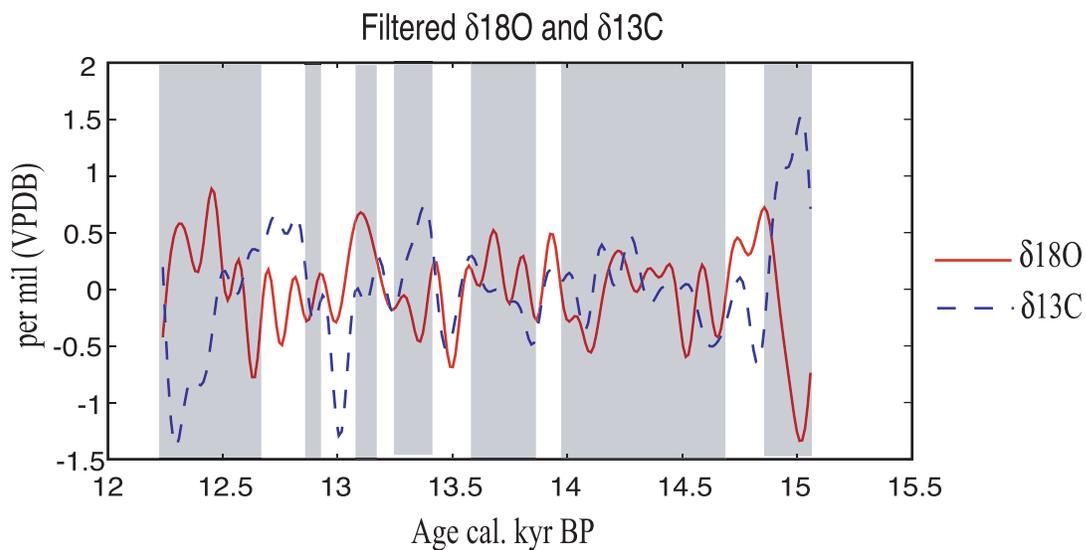


Figure 7: Overlaid plots of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$. The plots have been detrended by subtraction mean values and filtered using a fast fourier transform bandpass filter between 1-0.1 kyr. Areas shaded grey show out of phase relationships between the curves.

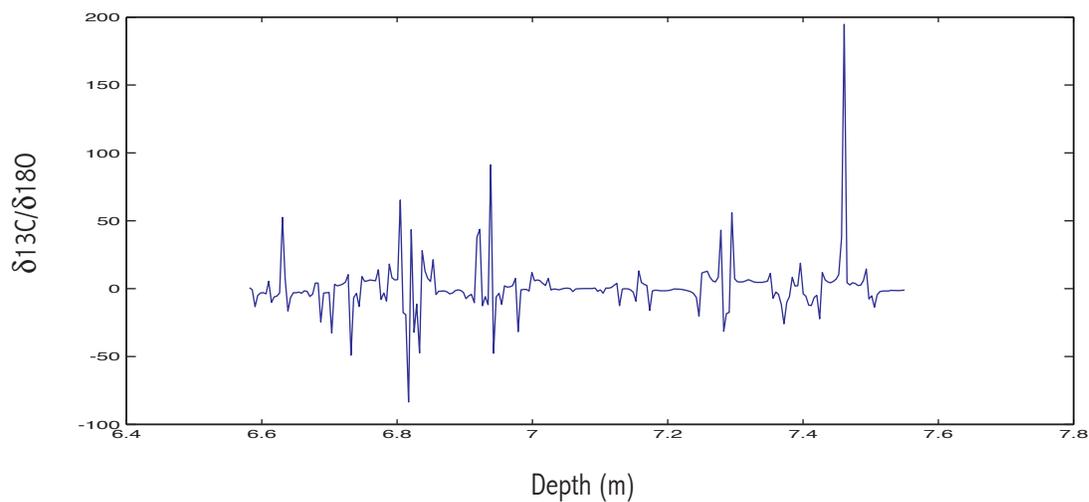


Figure 8: $\delta^{13}\text{C}$ divided by $\delta^{18}\text{O}$ from the LINC 1 record. When values are far from zero, $\delta^{13}\text{C}$ is either much greater or much smaller than $\delta^{18}\text{O}$. Areas where the line is flat imply in-phase relationships while peaks and troughs indicate out of phase conditions.

plot of $\delta^{13}\text{C}/\delta^{18}\text{O}$ with depth also indicates that the two proxies have a primarily out of phase relationship (Fig.8)

Spectral analysis of the portion of data from the GRIP ice core (Dansgaard et al., 1993) that corresponds to the section of LINC 1 that was analyzed yields a number of peaks similar to those found in the LINC analysis. Using a bandpass filter to eliminate irrelevant long-period signals, peaks with periods of 1.766 kyr, ~1.0 kyr, ~770 yr, and ~300 yr are shown to be present in the GRIP record as well as the LINC core (Fig. 9).

Discussion

Considerations for Data Interpretation

Variations in the contents of clay, carbonate, and plant organic matter have implications for the environmental conditions at the time of their deposition. Although the bedrock is limestone, the nearby shales and sandstones provide a ready source for non-combustibles found in the core. During cold, dry periods sparse vegetation would allow clay to be easily washed or blown into the lake from nearby. Additionally, wind can bring clay-size loess particles from a considerable distance when climate is dry and winds are strong. Consequently, high levels of clay (such as during the Younger Dryas) are interpreted as cold, dry environmental conditions.

Most organic matter in the environment is broken down on a (geologically) fast timescale. Therefore, the amount of organic matter remaining in a sediment core may indicate the preservation conditions of that period. Warmer, wetter conditions are generally more conducive to the breakdown of organic matter so increases in organics as indicated by LOI may also denote colder, dryer conditions.

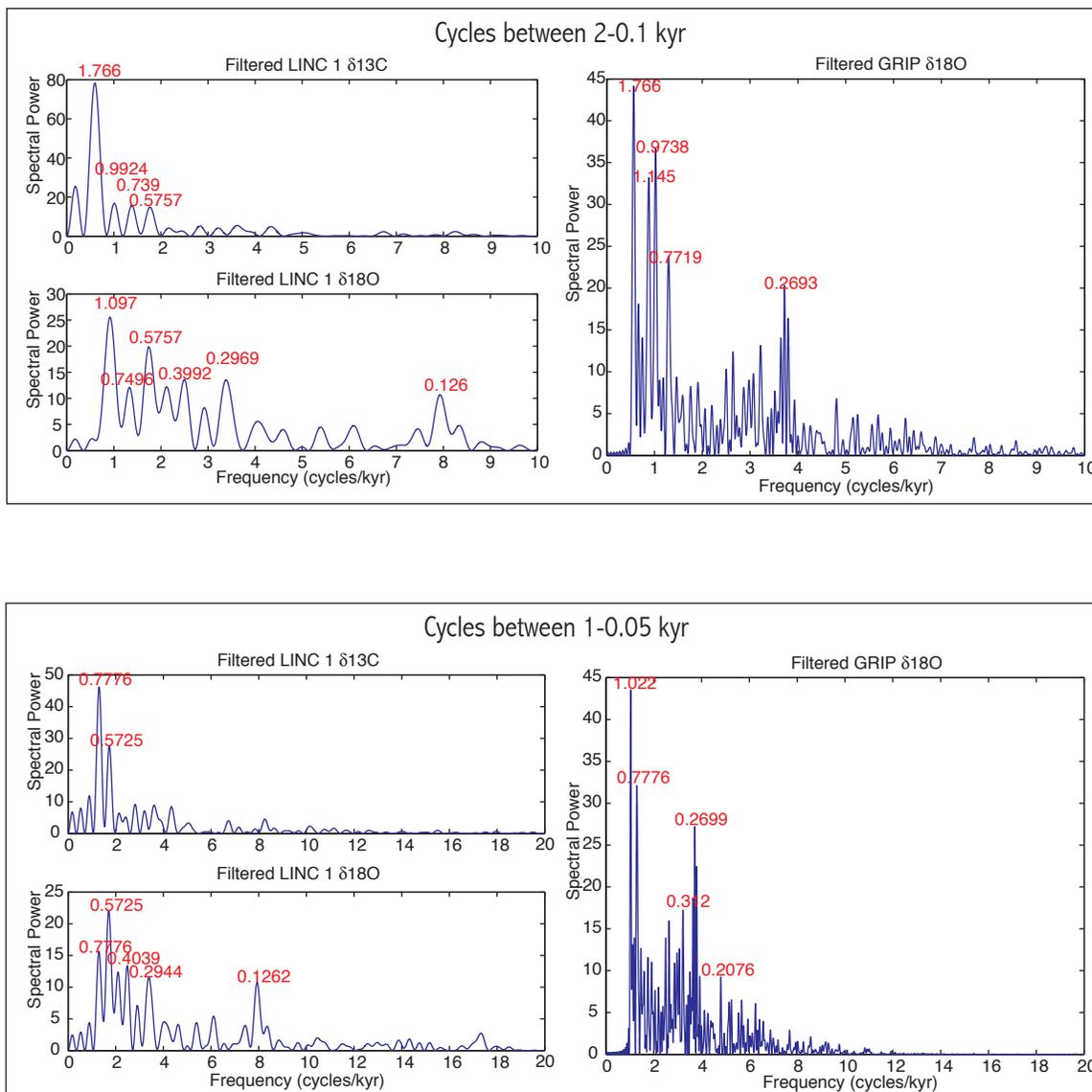


Figure 9: A comparison of spectral analysis from the LINC 1 core to peaks found in the GRIP ice core record (Dansgaard et. al., 1993) shows that a number of the strong cycles in the LINC core are also present in the ice core record. A) shows data filtered for cycles between 2 and 0.1 kyr and illustrates that cycles at 1.766, ~1.0, and between 0.739-0.776 are shared between the cores. B) indicates cycles between 1 and 0.05 kyr and shows correspondence of peaks at 0.7776 and ~0.3 kyr.

Lacustrine carbon isotope records are traditionally thought to be a measure of productivity in the lake (Kirby et al., 2002). When productivity is high, lake organisms draw large amounts of ^{12}C from the water and the carbon remaining for carbonate formation is enriched in ^{13}C . Therefore, high productivity is connected to high $\delta^{13}\text{C}$ values in carbonate sediments. Usually, high productivity is associated with warm, sunny climates which encourage photosynthesis. Alternatively, however, productivity changes may be influenced by changes in aridity. Kirby et al. (2002) have hypothesized that when cloud cover is frequent, productivity is reduced and consequently wet periods result in lowered $\delta^{13}\text{C}$. This interpretation matches well with the LINC 1 core, especially because the most prominent positive excursion is during the YD, which the dust content in Greenland ice cores indicates was a dry period (Mayewski et al., 1993).

A second contributor to $\delta^{13}\text{C}$ values in lacustrine carbonates is the amount of vegetation present in the environment. Because plants preferentially use ^{12}C , organic matter undergoing decay around the lake contributes CO_2 derived from organic carbon with relatively low $\delta^{13}\text{C}$ values (Fig. 10).

This concept can be extended to include the differences in carbon fractionation between C3 and C4 plants. Because these two types of plants produce different levels of $^{13}\text{C}/^{12}\text{C}$ fractionation (C4 plants fractionate less than C3 plants) as a consequence of adaptations to their environments, they contribute carbon with different $\text{C}^{13}/\text{C}^{12}$ ratios. Consequently, a shift in the type of plants present on land between from C3 to C4 plants could increase $\delta^{13}\text{C}$ values. During dry periods, when C4 plants are typically the dominant type of vegetation, an increase in the $\delta^{13}\text{C}$ would be expected (Street-Perrott et al., 1997). However, because organic carbon $\delta^{13}\text{C}$ values are very low (-12 to -25), the

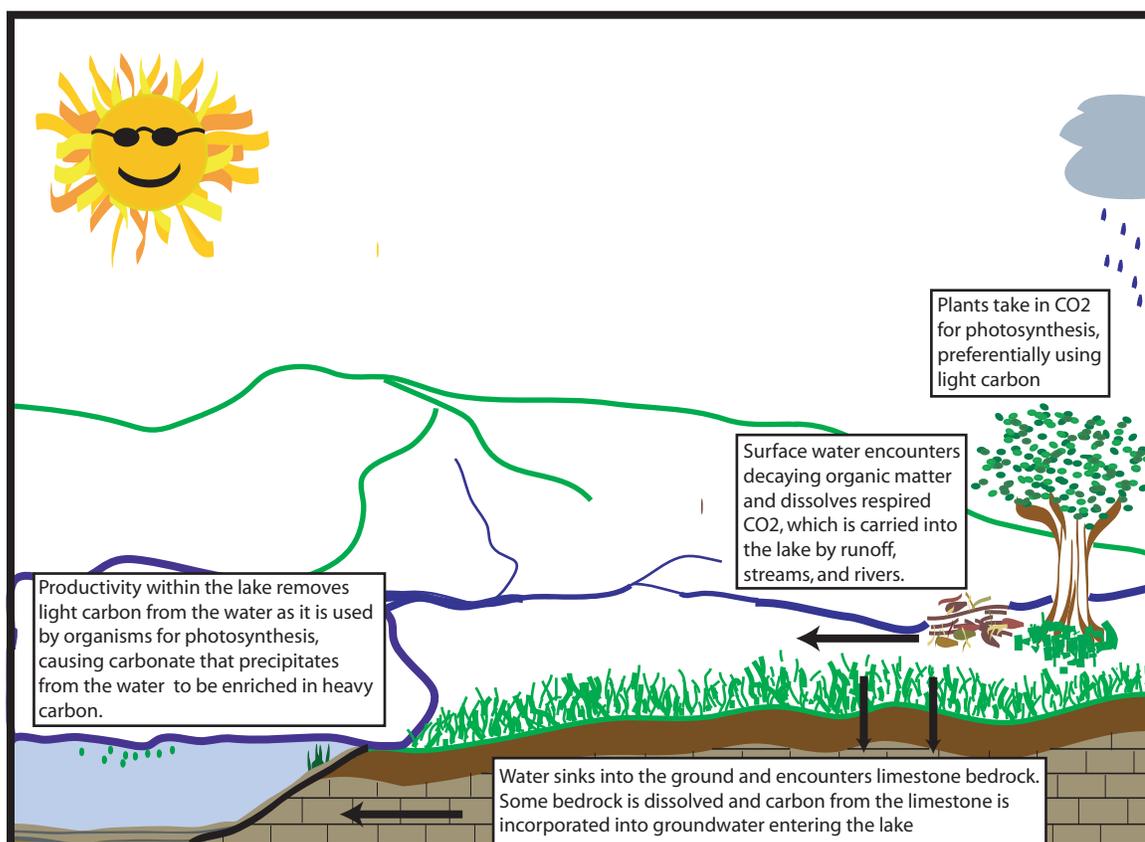


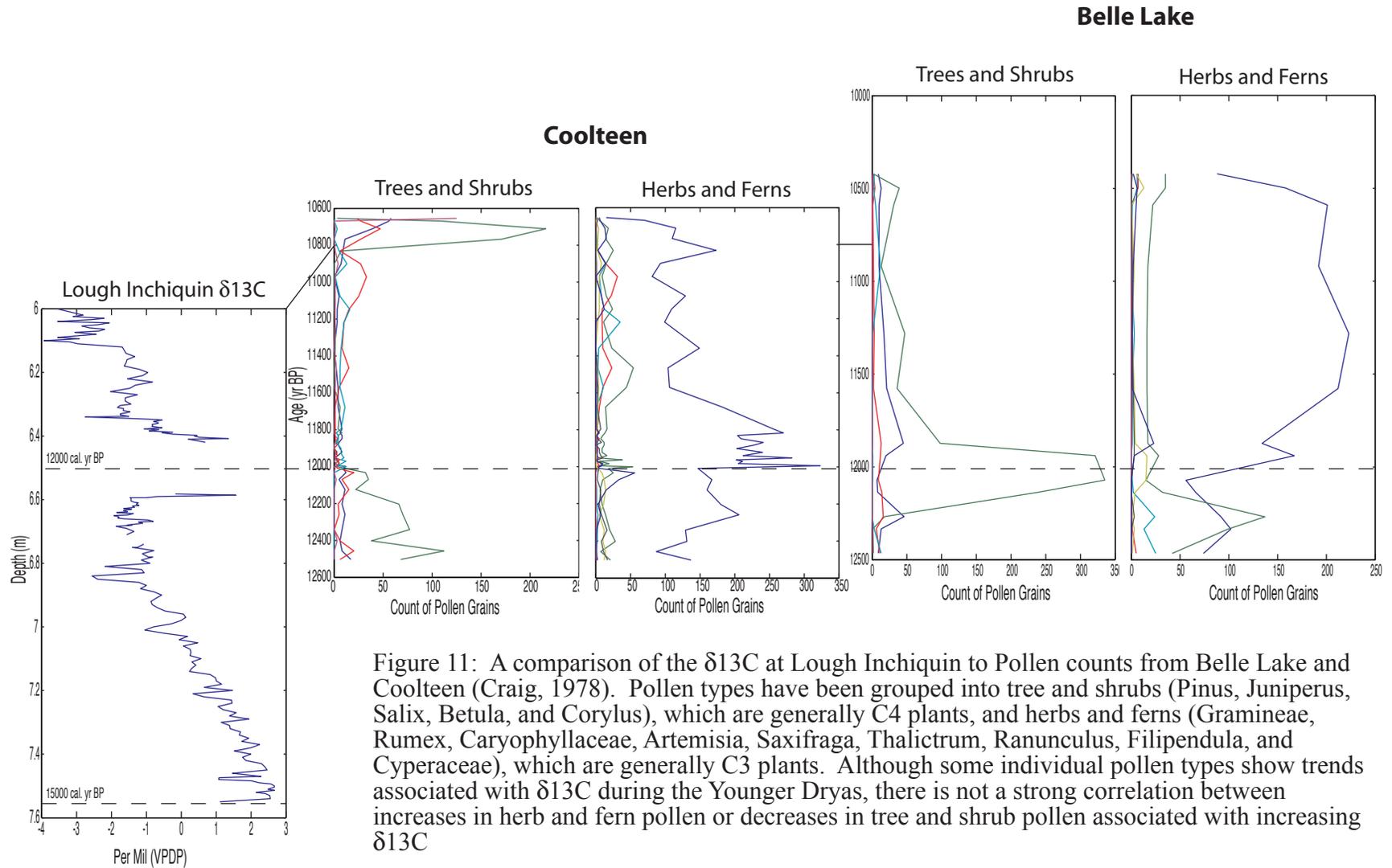
Figure 10: Environmental influences on $\delta^{13}\text{C}$. Water entering the lake can carry dissolved CO₂ derived from organic carbon obtained from decayed plants which is enriched in light carbon, causing $\delta^{13}\text{C}$ to decrease. Groundwater can also bring in carbon dissolved from limestone bedrock, which is low in light carbon and increases $\delta^{13}\text{C}$. Lake productivity affects $\delta^{13}\text{C}$ values in the lake itself. Photosynthetic organisms remove light carbon from the water, raising the $\delta^{13}\text{C}$ of the carbon that remains to be incorporated into sediments. Consequently, high $\delta^{13}\text{C}$ is associated with high lake productivity.

high $\delta^{13}\text{C}$ in the LINC 1 core implies that at Lough Inchiquin the contribution of carbon from decomposed plant matter is small. This would imply that the proportion of C3 to C4 plants is a less significant factor than variations in the total amount of vegetation around the lake.

The conclusion that aridity as represented by plant type is not a strong player at Lough Inchiquin is supported by a comparison of $\delta^{13}\text{C}$ to pollen diagrams from previous studies done by Craig (1978) (Fig 11) and O'Connell (1999) (Fig. 4). This comparison shows only a very rough correspondence of $\delta^{13}\text{C}$ values with pollen counts of herbs/ferns (C4 plants) vs. pollen counts of trees/shrubs (C3 plants). When changes in the types of plants present in the environment correlate with changes in $\delta^{13}\text{C}$, it is often individual species rather than plant groups that appear to show significant change. When plant groups show small changes that appear similar to the carbon isotope record, they lag behind it and therefore cannot be a significant influence on $\delta^{13}\text{C}$ change.

The bedrock underlying Lough Inchiquin is also important to consider when interpreting the carbon stable isotope results. Because there is limestone in such close proximity, groundwater entering the lake may acquire a large "old carbon" signal carbon from the bedrock (Fig. 10). Given that the $\delta^{13}\text{C}$ values of the LINC 1 record are high, it is likely that the limestone of the underlying bedrock did indeed make a large contribution. Positive $\delta^{13}\text{C}$ at the base of the core shows that bedrock may be the primary carbon source for the lake during early sedimentation.

Overall, changing vegetation volume is thought to be the dominant influence on carbon isotopic change at Lough Inchiquin, although it appears that changes in the dominant vegetation type do not play a significant role. Changes in lake productivity



related to cloud cover may also have some importance in this environment, giving the $\delta^{13}\text{C}$ a less prevalent function as an indicator of aridity change.

In the interpretation of lacustrine carbonates, oxygen isotope values are thought to have two primary sources of variation: source water $\delta^{18}\text{O}$ values and temperature. Atmospheric water vapor becomes depleted in ^{18}O during both evaporation and precipitation because evaporation preferentially takes up ^{16}O and precipitation preferentially removes ^{18}O . According to the Rayleigh distillation equation (Broecker et al., 1971), each time an air mass condenses and precipitates water, its $\delta^{18}\text{O}$ gets lower. As temperature increases, however, fractionation factors involved in evaporation and condensation decrease and the Rayleigh effect become less significant (Faure, 1977). The result of these changes is that at warmer temperatures, larger amounts of ^{18}O remain in water vapor and $\delta^{18}\text{O}$ values will be higher. Dansgaard (1964) first demonstrated a positive linear correlation of precipitation $\delta^{18}\text{O}$ values with air temperature (Fig. 12). Though Dansgaard's original work has been revised and refined, the association he illustrated remains significant. A 0.33 ‰/°C rate of change, as discussed by Ahlberg et al. (1996), is used to quantify temperature-based relationships at Lough Inchiquin because it integrates the influence of temperature on isotopic fractionation during carbonate formation.

Water $\delta^{18}\text{O}$ values change in response to variations in precipitation. This can occur in two possible ways; either the $\delta^{18}\text{O}$ value of the precipitation's water source can change or the amounts of precipitation relative to evaporation (P-E) can change. If the $\delta^{18}\text{O}$ values of the oceanic water source are raised by an increase in salinity, then precipitation entering the lake is expected to have a higher $\delta^{18}\text{O}$ value and the water

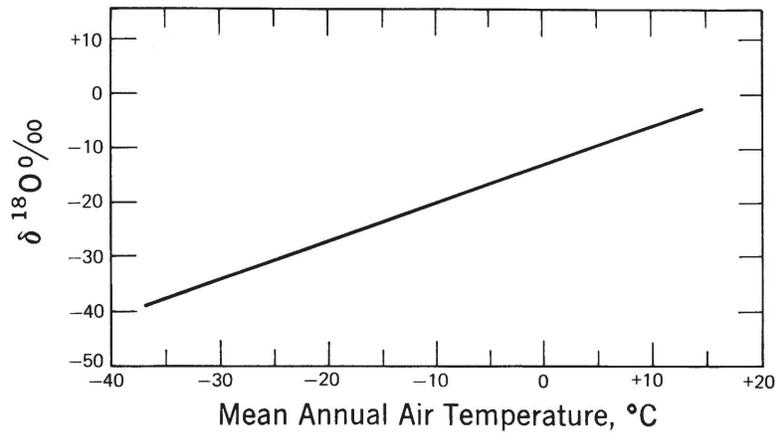


Figure 12: The linear relationship between oxygen isotope values and air temperature as established by Dansgaard (1964). The equation of the line is $\delta^{18}\text{O} = 0.695t - 13.6$, indicating a 0.695 per mil/degree C rate of variation.

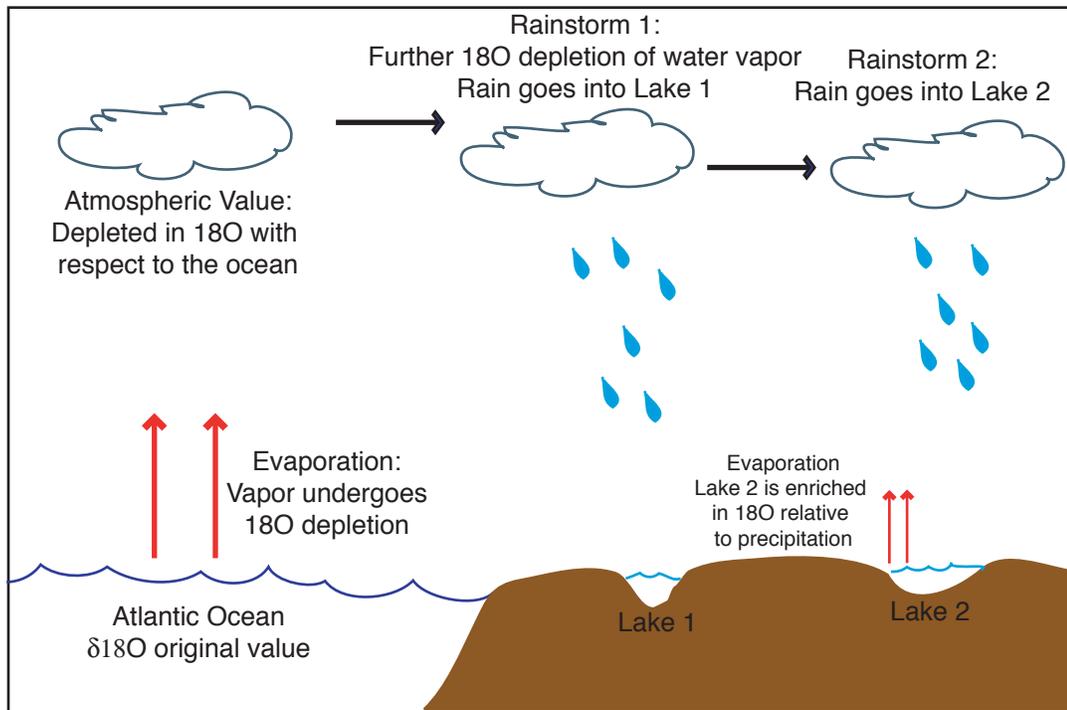


Figure 13: Fractionation relationships for lacustrine $\delta^{18}\text{O}$ values. If the original values in the Atlantic ocean were changed, that change would carry through the system and affect the values in the lake. Distance from the source is important; here Lake 2 would be expected to have lower $\delta^{18}\text{O}$ than Lake 1. However, Lake 2 has undergone evaporation which could enrich ^{18}O and raise $\delta^{18}\text{O}$.

values of the lake will be raised. Additionally, if the location of the water source changes to a more distant area, more heavy oxygen will precipitate out on the way to the lake (Fig. 13) and cause water $\delta^{18}\text{O}$ values to decrease.

Evaporation can produce enrichment in ^{18}O in lakes (Siegenthaler et al., 1986). As a result, the $\delta^{18}\text{O}$ of a lake in a closed basin could be expected to respond to changes in the influx of precipitation, which tends to have low $\delta^{18}\text{O}$, and outflux of evaporation, which depletes the ^{16}O of the remaining lakewater (Fig. 13). This can be considered in terms of precipitation minus evaporation, or P-E. If this is a high value (wet conditions), $\delta^{18}\text{O}$ would be expected to decrease and if it is negative (arid, windy conditions) $\delta^{18}\text{O}$ should increase.

At Lough Inchiquin, the location of the water source is unlikely to have changed significantly because the study site has a close proximity to Ireland's western coast where precipitation from westerly air masses dominated throughout the transition from the last glacial maximum (Ahlberg et al., 1996). Computer models based on aeolian dune records from Europe indicate that winds were from the west/southwest during the Younger Dryas (Isarin et al., 1997), supporting this conclusion.

With an Atlantic water source, changing ocean salinity is the major mechanism for $\delta^{18}\text{O}$ source value variation. Inputs of fresh glacial meltwater decrease salinity and $\delta^{18}\text{O}$ values. $\delta^{18}\text{O}$ values from Atlantic sediment cores show meltwater pulses (as indicated by decreased $\delta^{18}\text{O}$) before and after the Younger Dryas but decreased meltwater influx (slightly increased $\delta^{18}\text{O}$) during the cold period (Fairbanks, 1989; Lehman et al., 1992). Because these changes act in the opposite direction to the changes recorded in the LINC 1 core during major events such as the Younger Dryas, it is concluded that they

play a subordinate role in $\delta^{18}\text{O}$ variation at Lough Inchiquin. If they have any significant effect, it dampens the signal of temperature variation implying that estimates of temperature change are conservative (Ahlberg et al., 1996). Support for the idea that water source has minimal influence in my setting is provided by Rozanski et al. (1993) who suggest that at mid- and high latitudes temperature is the dominant control on the isotopic fractionation of oxygen.

Changes in the aridity of the environment resulting in changed P-E values for the lake likely influence the $\delta^{18}\text{O}$ of the LINC 1 core. However, based on this interpretation under dry conditions $\delta^{18}\text{O}$ in sediments is expected to increase as evaporation removes light oxygen from the water. This negative correlation does not fit well with the LINC 1 data during the Younger Dryas, where despite known dry conditions $\delta^{18}\text{O}$ values are low. Consequently, this effect is considered to be a secondary influence with respect to larger trends at Lough Inchiquin. Because this mechanism acts in the opposite direction as temperature-based $\delta^{18}\text{O}$ change, it is expected to cause underestimation of temperature change over the course of large trends in the LINC 1 data. With respect to more rapid oscillations, however, changes in P-E are a plausible dominant influence on $\delta^{18}\text{O}$.

Testing covariance helps to determine if oxygen and carbon isotopes are the results of the same environmental forcing mechanisms. In the case of my results, visual inspection of graphs shows some cases where there is a negative correlation between changes in the isotope values. However, a plot of carbon vs. oxygen similar to that of Drummond et al. (1995) demonstrates that LINC 1 carbon and oxygen show very little covariance (Fig. 14). Superimposing detrended and filtered $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ curves

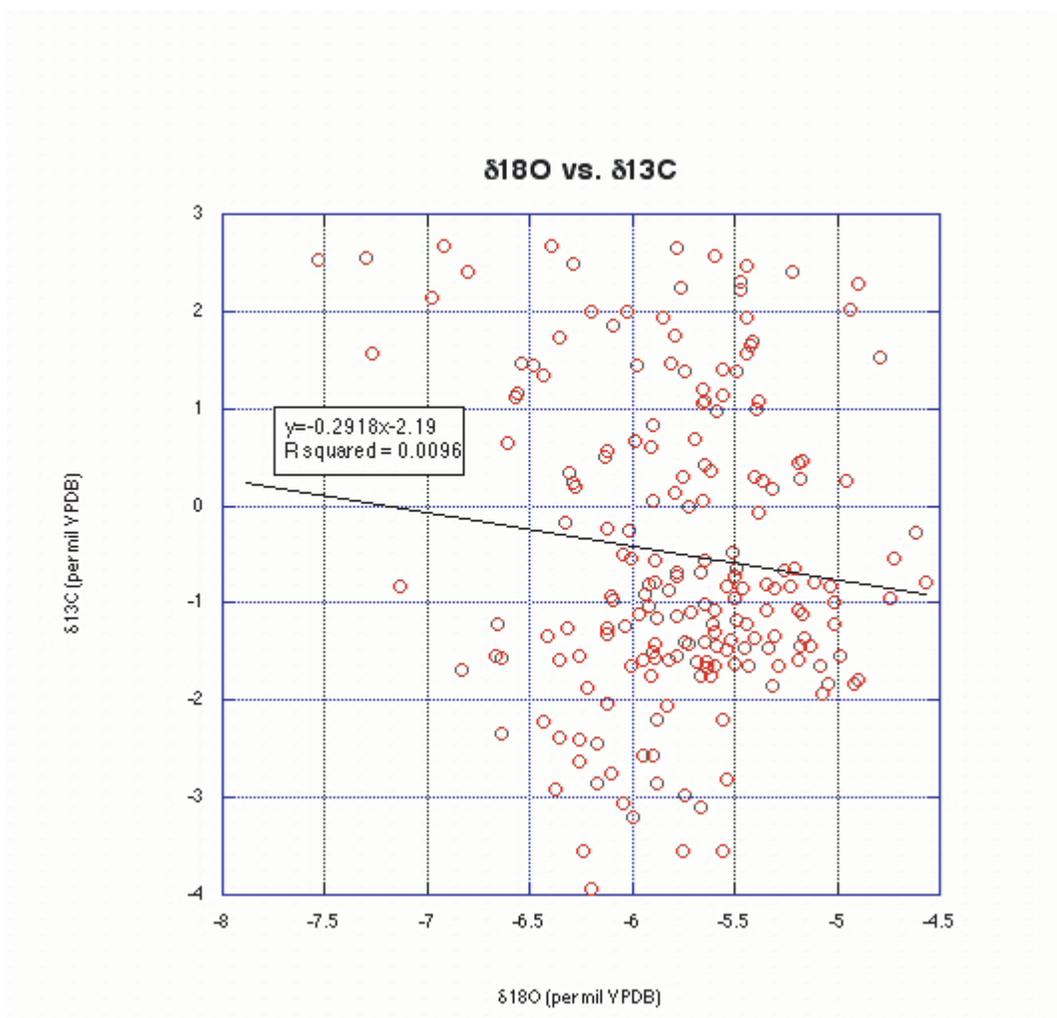


Figure 14: A plot of δ^{18} vs. δ^{13} C clearly shows that the two do not have a consistent relationship as would be expected if they were covariant. A best-fit linear trendline has a very low R squared value, further demonstrating the lack of covariance

indicates that the phase relationships between the two are not consistent, but that during some portions of the record they are in phase (Fig 7). A plot of the $\delta^{13}\text{C}/\delta^{18}\text{O}$ ratio against time (Fig. 8) provides further evidence that while some of the time the isotopes vary concurrently, at other times they change in dramatically different ways. During these times, oxygen and carbon isotopes were either responding to different forces or responding differently to the same mechanisms. This type of result could be produced if a lag time is involved in the response of one isotope value but not the other.

Spectral analysis results make it clear that $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ do have a number of common forcing environmental factors, although their identity is unknown. The consistent occurrence of common peaks (Fig. 6) shows that on some level the changes in isotope values are influenced by the same environmental features. However, both carbon and oxygen show consistent peaks that are not shared, giving an indication that each has environmental influences that do not affect the other. The lack of consistent phase relationships between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ may illustrate the influence of these individual signals. The short individual cycle length of $\delta^{18}\text{O}$ relative to the lengths of $\delta^{13}\text{C}$ individual cycles suggests that environmental changes forcing $\delta^{18}\text{O}$ change operate on a more rapid timescale than those influencing $\delta^{13}\text{C}$. This interpretation fits well with the idea that $\delta^{13}\text{C}$ responds to changes in vegetation, which react more gradually to climate change than influences on $\delta^{18}\text{O}$ such as precipitation and temperature.

Paleoclimate at Lough Inchiquin

Based on comparison to O'Connell et.al (1999), sedimentation in Lough Inchiquin began 15,000 cal. yr BP at the end of the last glacial maximum. A cold climate

at the base of the core is shown by the low $\delta^{18}\text{O}$ values. Correlation to the GRIP ice core implies that the base of the core may be slightly younger than 15,000 (Fig. 15), but still places it prior to the Bølling/Allerød warming. The significant amount of clay and high $\delta^{13}\text{C}$ values in the basal section indicate a cold, dry climate with little or no vegetation present.

Bølling/Allerød warming at Lough Inchiquin is shown to be rapid, recorded in less than 4.6 cm. The $\delta^{18}\text{O}$ indicates that temperatures increased by up to 7.8°C during this short interval. A brief negative $\delta^{18}\text{O}$ shift during the Bølling/Allerød, likely the Older Dryas, suggests that there was a short return to cold and/or dry conditions. A decrease in clays following this interruption implies that the climate became more humid after the inferred Older Dryas. With the exception of the Older Dryas event, $\delta^{18}\text{O}$ temperatures remain relatively high up to the Younger Dryas

$\delta^{18}\text{O}$ leading into and out of the Younger Dryas imply cooling of 7.7°C associated with the event. Interpretation of $\delta^{13}\text{C}$ in the same section implies a dry, low vegetation environment around the lake, consistent with a cold climate conditions. High clay content and relatively high levels of preserved organics in this portion of the core imply a cold, dry climate as well.

Although an abrupt $\delta^{18}\text{O}$ increase marks the end of the Younger Dryas in this core, the $\delta^{18}\text{O}$ values do not maintain the peak values reached at 6.35 m. This might imply that precipitation increased following the initial warming at the beginning of the Preboreal, decreasing $\delta^{18}\text{O}$. Decreasing $\delta^{13}\text{C}$ values at the top of the core support the idea that humidity increased after the Younger Dryas. A brief cold period indicated by the oxygen isotopes at 6.2 m corresponds to a slight increase in $\delta^{13}\text{C}$. This excursion

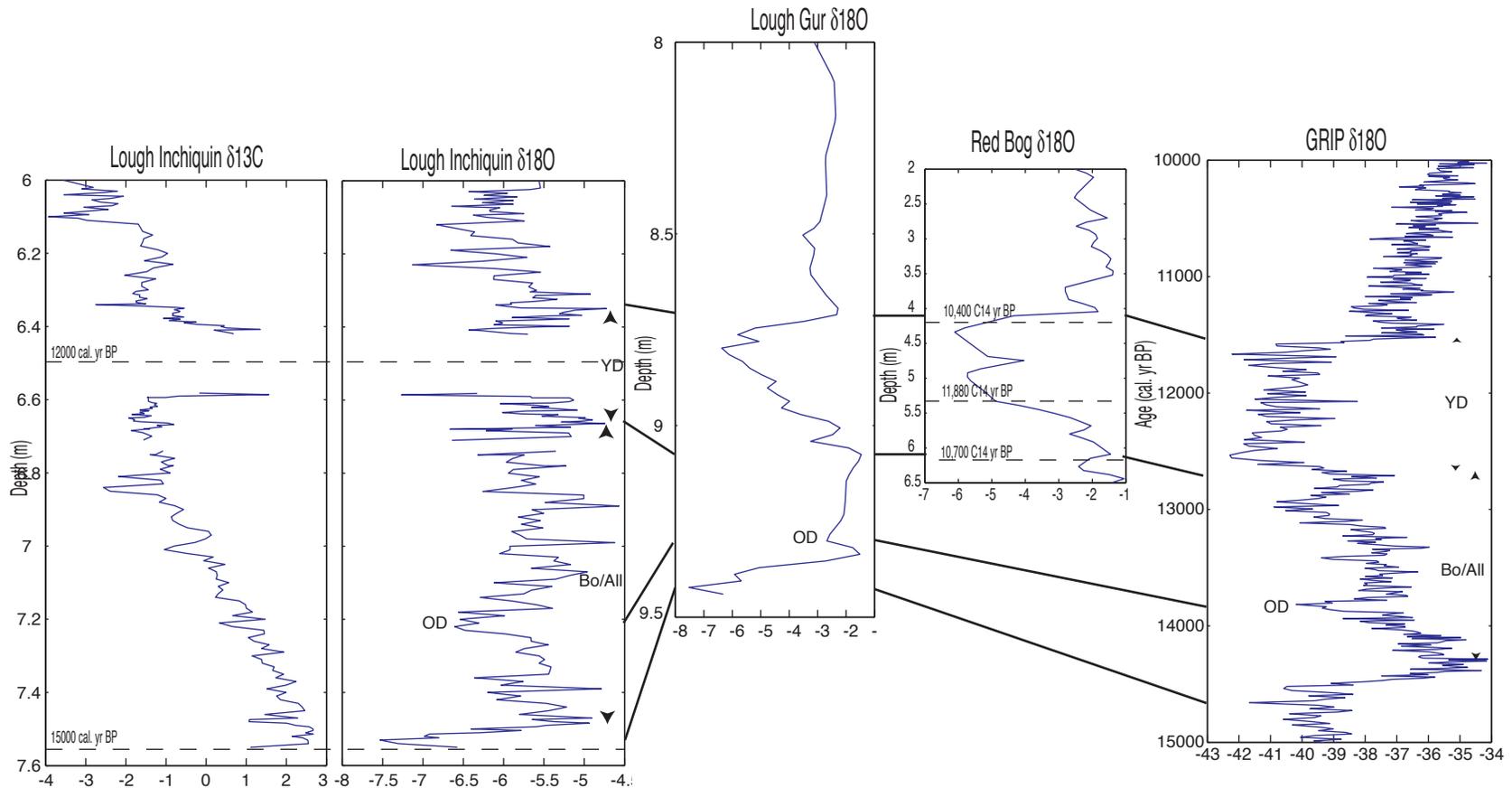


Figure 15: Stable isotope results from Lough Inchiquin compared to other paleoclimate records. Correlations were based on age and curve similarities. Comparison to nearby Lough Gur and Red Bog (Ahlberg et al., 1996) shows that large shifts tend to correlate well but small-scale changes show variability between cores. The GRIP ice core $\delta^{18}\text{O}$ (Dansgaard et al., 1993) also shows a remarkably strong correlation of significant shifts with LINC 1 $\delta^{18}\text{O}$. The Bolling-Allerod (Bo/All), Younger Dryas (YD), and Older Dryas (OD) are marked on the GRIP and LINC 1 $\delta^{18}\text{O}$ records.

hints that a minor cold, dry period followed after the Younger Dryas. The 8,200 yr event is too young to be represented in my core section, but the Preboreal Oscillation correlates well with this event.

The LINC 1 $\delta^{18}\text{O}$ record shows remarkable similarity to the GRIP ice core record (Fig. 9, 15), illustrating the broadly regional nature of the post-glacial sequence trends. Comparison to $\delta^{18}\text{O}$ curves from nearby Lough Gur and Red Bog (Ahlberg et al., 1996), however, shows that significant local variation takes place as well (Fig. 15).

The size of the overall negative shift in $\delta^{13}\text{C}$ makes it clear that a significant environmental change took place between the end of the last glacial maximum and the beginning of the Holocene. Considering the shift as a proxy for the amount of plant life nearby indicates a gradual increase in the vegetation surrounding the lake. Such a change would be consistent with the development of extensive vegetation from a barren post-glacial landscape. Using $\delta^{13}\text{C}$ as a proxy for aridity, it can be inferred that climate at Lough Inchiquin has gradually become wetter since the last glacial period. The presence of a dominant overall shift in $\delta^{13}\text{C}$, a feature not present in $\delta^{13}\text{C}$ records from nearby areas (Fig. 4), implies a strong and highly localized environmental change. A combination of these ideas implies that increasing total vegetation in the area immediately around the lake caused by wetter conditions and warmer temperatures is the most plausible explanation of the change in $\delta^{13}\text{C}$ observed in my record.

Frequent $\delta^{18}\text{O}$ shifts of 1 ‰ or greater magnitude present the possibility of frequent minor environmental changes. Temperature variation may have taken place locally, but at this small scale precipitation patterns contributing to changes in P-E values likely play an important role as well. Precipitation is a factor that can have wide

variations both temporally and spatially, which may help to explain localized $\delta^{18}\text{O}$ variation as shown by the differences between the LINC 1 core and the Tory Hill/Lough Gur/Red Bog cores (Fig.15).

Conclusions:

Stable isotope variation shows that significant environmental change is present in the bottom two meters of the sediment core from Lough Inchiquin. An overall transition from cold, dry, barren glacial conditions to a temperate, moist, vegetation-rich environment took place, although this trend was interrupted by abrupt oscillations. This pattern is considered to be the product of both regional and local climate variation, as is clearly shown by comparison to other cores. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ are affected by both shared and individual environmental influences, and it appears that $\delta^{18}\text{O}$ responds more rapidly to environmental changes than $\delta^{13}\text{C}$.

It is clear that the Lough Inchiquin core contains the well-documented climatic oscillations of the glacial/interglacial transition. Lough Inchiquin oxygen isotopes demonstrate the full post-glacial sequence and are supported by clay/carbonate content and $\delta^{13}\text{C}$ results. Estimates for this study are that cooling associated with the Younger Dryas in this location was a minimum of 7.7°C , and that this returned temperatures to very near those of the late-glacial at the beginning of the core. Large climatic shifts in the LINC 1 record had rapid onset and termination, a feature which is consistent with most previous studies.

Further study of the Lough Inchiquin sediments would help to yield a clearer picture of paleoclimate conditions during their deposition. AMS dates would increase the accuracy of comparisons to other cores, and allow commentary to be made about the

timing of events at Lough Inchiquin relative to their timing in other locations. Although relative temperature change can be discussed in this study, limitations of stable isotope paleoclimate methods make it impossible to obtain absolute paleotemperature measurements based on this data (Ahlberg et al., 1996). Pollen studies of these sediments, however, might allow minimum estimates of absolute temperature based on the plant species present as done by Isarin and Bohncke (1999). Pollen analysis of the LINC 1 core would also be a way to further investigate the relationship between plant types and $\delta^{13}\text{C}$. A comparison with pollen data from Lough Inchiquin would be more accurate than the ones done in this study to cores whose data may reflect local conditions slightly different from Lough Inchiquin's.

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Glaciers Advance
During the Younger Dryas
(Unknown Artist)

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