Geology, Limnology and Paleoclimatology of Green Lakes State Park, New York

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Introduction

Green Lakes State Park, located ~15 km east of Syracuse, New York near the village of Fayetteville, is a stunning geologic and limnologic wonderland (Fig. 1). Located along the axis of a narrow glacial meltwater channel, the Park consists of two small, moderately deep (maximum water depth of ~53 m) lake basins (Green Lake and Round Lake), protected along their flanks by steep walls of Silurian bedrock and separated from each other by a small modern wetland (Figs. 2 and 3).

These lakes are unusual, not only for their geologic setting, but also their water characteristics which have attracted the attention of limnologists for decades. In fact, limnologically, Green Lake is one of the best studied lakes in the world (Torgersen et al., 1981; Thompson et al., 1990). There are two aspects of the limnology of Green Lake which make it so interesting. Foremost is the fact that it is a meromictic lake (as is Round Lake) which means that when the water column turns over seasonally, it does so only above a strong chemically-controlled density boundary or chemocline at a water depth of ~18 m (Fig. 4).[1] Because of this, bottom waters >18 m deep are permanently stratified and devoid of oxygen (anoxic). The second unusual limnologic feature of these lakes is that they annually precipitate large amounts of calcium carbonate (calcite), both from the open, wind-mixed surface waters of the lake (epilimnion) and along their shallow, near shore (littoral zone) margins in the form of "reefs" or bioherms.

The annual precipitation of calcite from the epilimnion coupled with the fact that deep bottom waters are anoxic, results in the deposition and preservation (no bioturbation) of annual sediment laminations or true varves. Because of this, Green Lake and Round Lakes have enormous potential from unraveling decadal to annual scale climate change (paleoclimatology) for at least the entire Holocene (past 10,000 years). With our current interest and concerns about possible global warming, the meromictic lakes of Green Lakes State Park may well provide a long-term, high-resolution template of natural climatic variability in which to view future environmental change.

The purpose of this fieldtrip is to acquaint you with the basic geology, limnology, and preliminary paleoclimatology of Green Lakes State Park. We will draw upon an extensive published literature to discuss: (1) the geologic origin of the lakes; (2) their fundamental limnology; (3) the annual, open water precipitation of calcium carbonate which will explain why "Green Lake is green"; and, (4) the origin of their "fringing reefs"; we will then draw upon new, unpublished data that will: (5) document the natural history of relative lake level fluctuations, and thus climate change, throughout the Holocene; and finally, (6) examine in high resolution, the natural and anthropogenic environmental changes which have occurred over the past 2,500 years.

This fieldtrip will be a leisurely half-day walking tour around both Green and Round Lakes. Scientific information will be provided via a series of large, colorful posters supplemented by on site examination of sediment cores. There will be ample time for discussion, and if the weather cooperates, we will be treated to a spectacular feast of fall foliage against the backdrop of aqua lakes; so, bring an open mind and your cameras!

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[1] That part of the water column which mixes during turnover is referred to as the mixolimnion whereas that which does not mix is the monimolimnion.

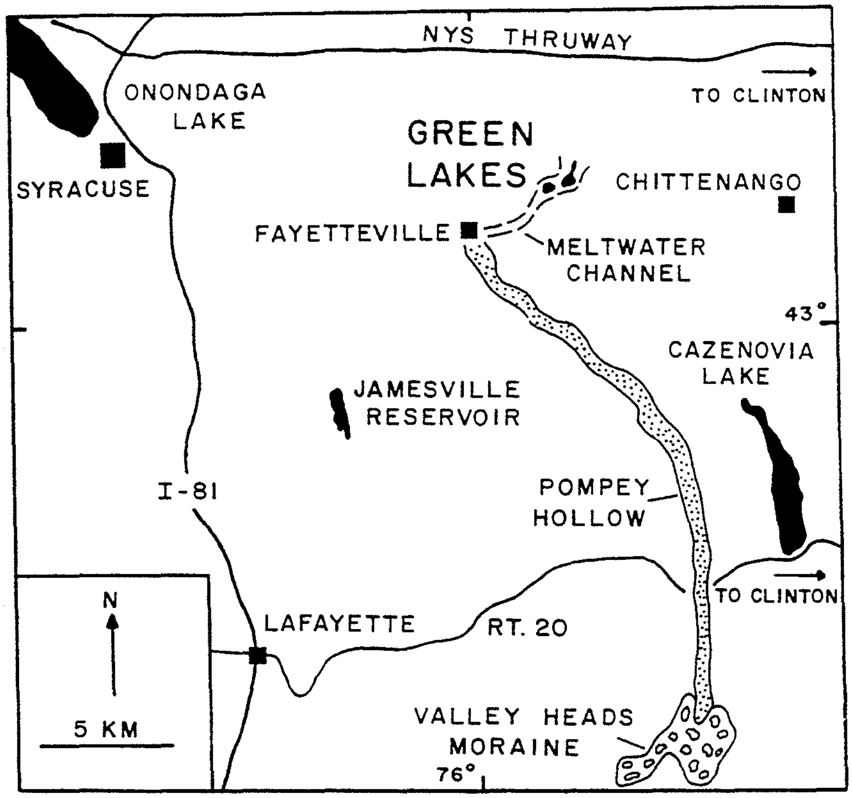


Fig. 1: General index map of Green Lakes State Park region of central New York, illustrating location of Green Lakes relative to Pompey Hollow (glacial trough) and Valley Heads Moraine (~14.4 ka). Major highways also shown. To reach the Park from Hamilton College in Clinton, ~~go~~ north and take the N.Y.S. Thruway west to exit 34. From Canastota take Rt. 13 to Chittenango and then Rt. 5 to Green Lakes State Park. Alternatively, head south from Clinton and take Rt. 20 west to Cazenovia and then Rt. 92 to Manlius. Follow signs to Fayetteville and then signs to the Park. We will meet at the boathouse on the northwest corner of Green Lake.

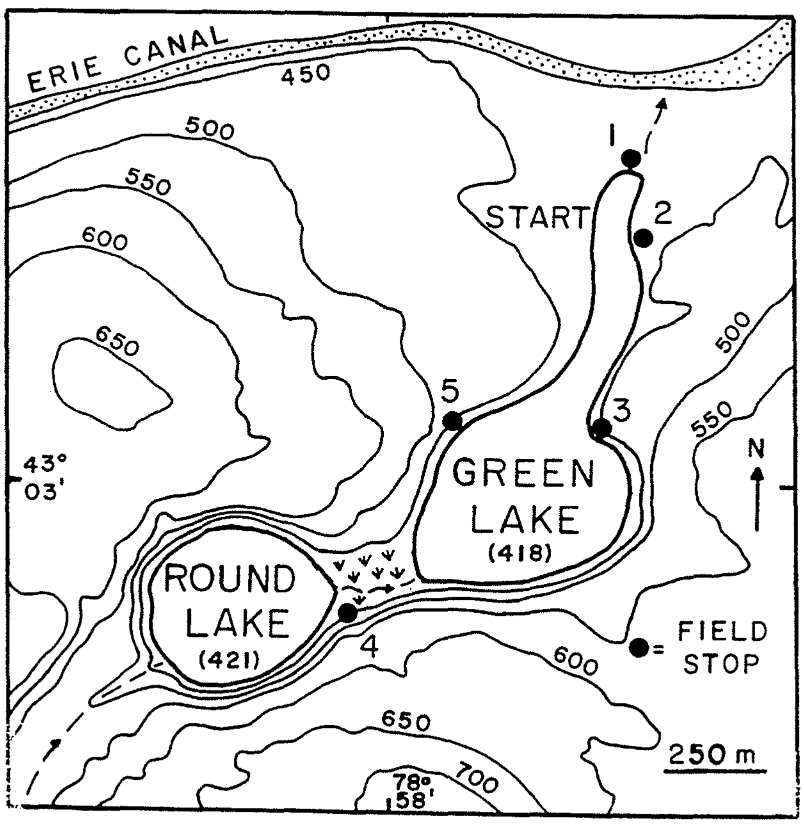


Fig. 2: Location map for Green Lake State Park illustrating local topography. Note that both Green and Round Lake are located in a SW-NE trending "cross channel". Numbered circles correspond to field trip stops. Elevations in feet above sea level (100'= 30.5 m).

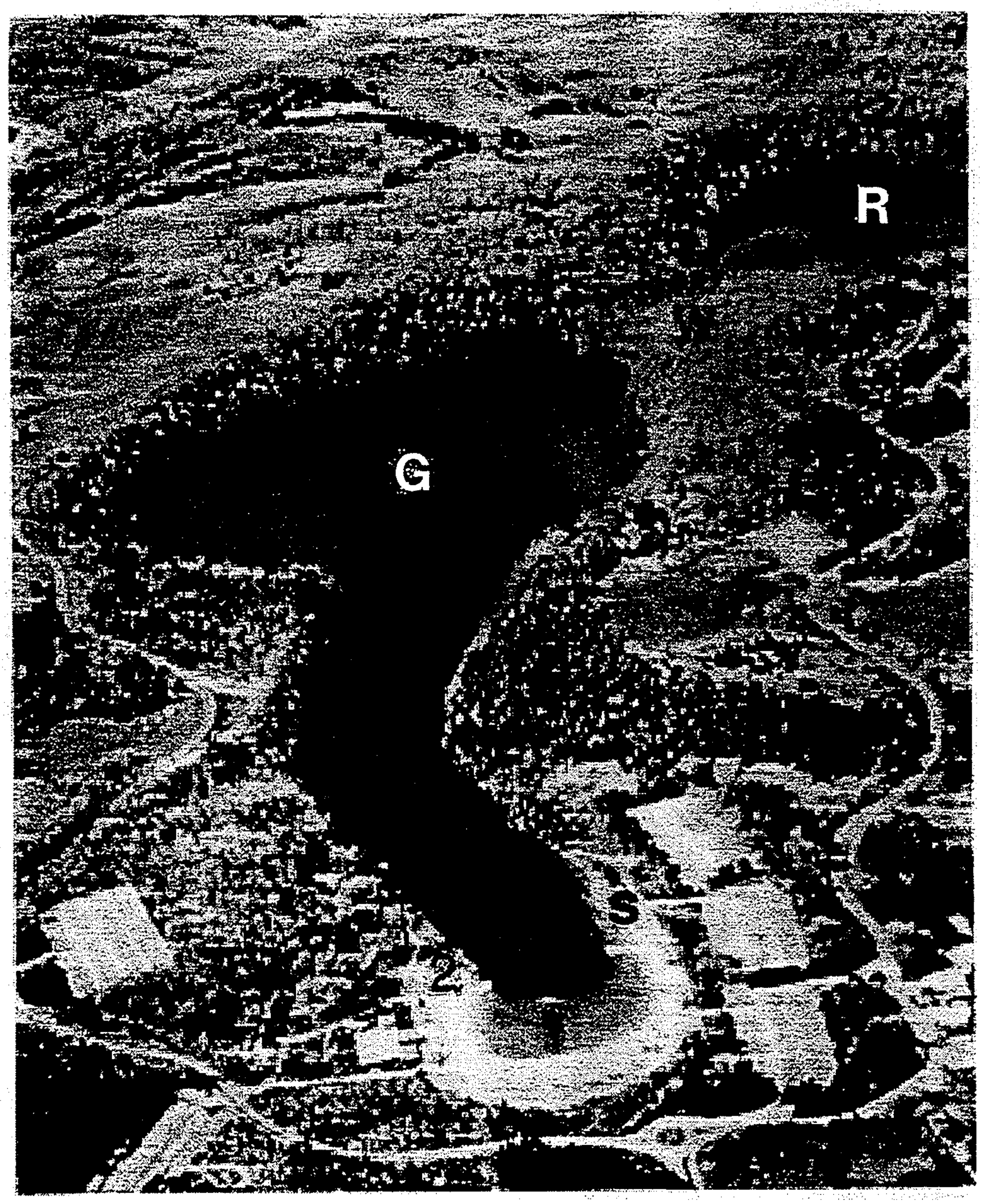


Fig. 3: Aerial photograph (looking south) of Green Lakes State Park illustrating main basins of Green (G) and Round (R) Lake and the "neck" of Green Lake, as well as the wetland (W) between the two lakes. Field trip stops highlighted by numbers; S= start.

Field Trip Stops

This field trip will begin at the boathouse near the northwest corner of Green Lake. We will then proceed clockwise around both Green Lake and Round Lake before returning to the boathouse. There will be a total of five stops to discuss the geology, limnology and paleoclimatology of Green Lakes State Park and to observe the scenic beauty of this local treasure.

Stop #1: Geologic Setting and Origin of the Lakes

Our discussion of the geology of Green Lakes State Park will take place on the swimming beach at the north end of Green Lake. From here we will have an unobstructed view to the south along the axis of the glacial meltwater channel in which Green and Round Lakes reside.

As we look to the south, steep bedrock walls can be seen rising from the surface of Green Lake at an elevation of 418 ft. (127 m) to at least 550 ft. (168 m) above sea level, although the highest summit (South Hill) adjacent to the park extends to an elevation of 788 ft. (240 m) (Fig. 2)(Muller, 1967). Total bedrock relief in the area, as measured from the top of South Hill to the bottom of Green Lake is 543 ft. (166 m). However this value should be considered a minimum because the thickness of sediments beneath the floor of Green Lake is not known, although Muller (1967) reported that drilling "just north" of Green Lake penetrated 138 ft. (42 m) of unconsolidated sediment without reaching bedrock.

The rocks that make up the walls of the meltwater channel in Green Lakes State Park are part of the Salina Group of Silurian (~440 to 410 million years ago) age (Muller, 1967). The oldest stratigraphic unit in the Park is the Vernon Shale which consists largely of easily eroded red and green shale as well as gypsum ($CaSO_4 \cdot H_2O$). The top of the Vernon Shale extends upward to within ~40 ft. (12 m) of the surface of Green Lake (Fig. 4; Muller, 1967). Overlying the Vernon Shale is more shale and argillaceous dolostone of the Syracuse Formation which contains evidence ("hopper casts") of evaporite minerals such as halite (NaCl) and gypsum. This middle unit extends upward to about the present day surface elevation of Green Lake (Fig. 4). The remainder of Silurian stratigraphy here consists of a relatively resistant dolostone member of the Syracuse Formation (Fig. 4) and the overlying Camillus Shale capped by a dolostone of the Bertie Formation which forms the summits of surrounding hills (Muller, 1967). The lake basins, however, rest within shales and evaporites of the Syracuse Formation and underlying Vernon Shale (Fig. 4).

Both Green Lake and Round Lake are located in an east-northeast "cross channel" (Fairchild, 1909; Sissons, 1960) which cuts obliquely across the Onondaga Escarpment at the northern edge of the Appalachian Plateau (Fig. 1 and 2). Today, this cross channel is drained by a "markedly underfit stream" (Muller, 1967) and near its southwestern end there is an abrupt topographic drop of ~100 ft. (30.5 m) from a dolostone member of the Syracuse Formation to the surface of Round Lake (Fig. 2).

As noted by Muller (1967), Green Lake (which was originally known as Lake Sodom) was first depth sounded by Vanuxem (1839) and subsequently by Miner (1933), who also sounded (via lead-line) Round Lake. Both lakes consist of steep-sided, flat-floored, semi-circular basins with maximum water depths slightly >50 m. However, Green Lake also contains an elongated "neck" or "handle" which shoals to the north (Fig. 2). As also noted by Muller (1967), the surface elevations of Green Lake (418 ft; 127 m) and Round Lake (421 ft.; 128 m) are significantly (~6 m) below the highest stand of pro-glacial Lake Iroquois which existed from ~12.5 to 11.4 ka ^{14}C (Muller and Prest, 1985; Anderson and Lewis, 1985). The surface of Green Lake is also ~6 ft. (2 m) lower in elevation then the water surface of the nearby Erie Canal (Fig. 2).

Green Lakes State Park is located north of the Valley Heads Moraine which was deposited ~14.4 ka ^{14}C (Muller and Calkin, 1993) during a brief phase of Laurentide ice sheet instability coincident with Heinrich event H-1 (Mullins et al., 1996). Thus, modern Green Lake and Round Lake must be less than 14,400 ^{14}C years in age (Fig. 5). The meltwater channel in which both lakes reside connects in the village of Fayetteville with Pompey Hollow, a large glacial trough whose origin is likely linked with the nearby Finger Lakes (Mullins et al., 1996). The Green Lakes meltwater channel is but one of many east-west orientated glacial drainages in the Syracuse-Oneida region (Sissons, 1960).

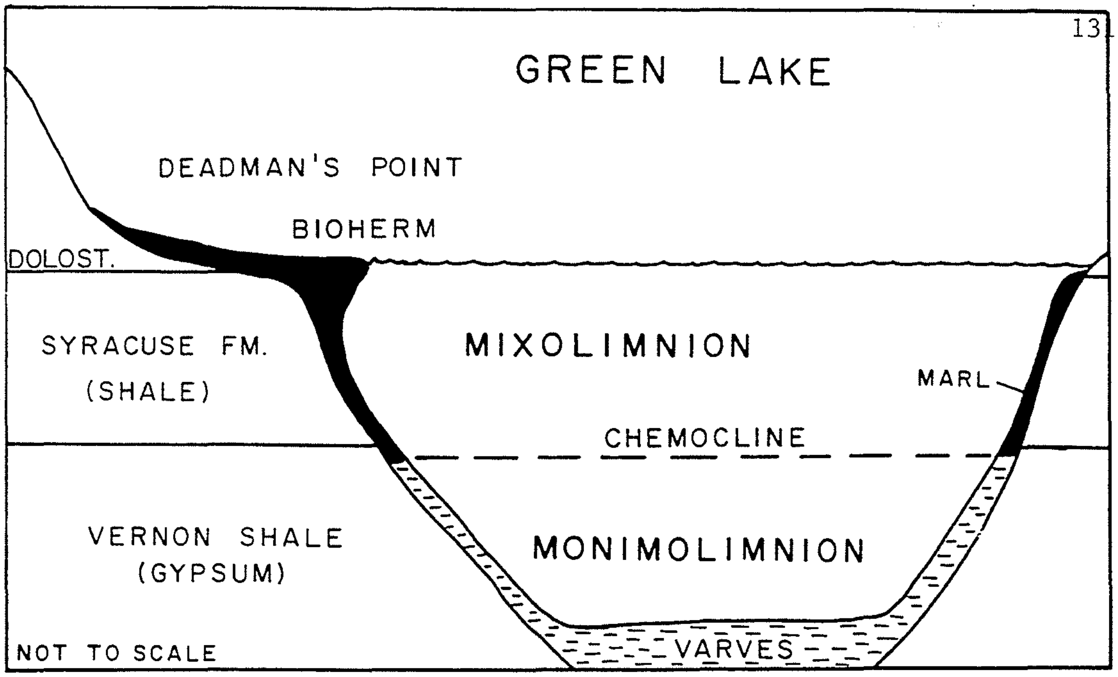


Fig. 4: Stratigraphic cross-section across main basin of Green Lake illustrating bedrock lithology, water depth and position of chemocline, which divides lake into an upper mixolimnion and a lower monimolimnion. Based on Dean and Fouch (1983).

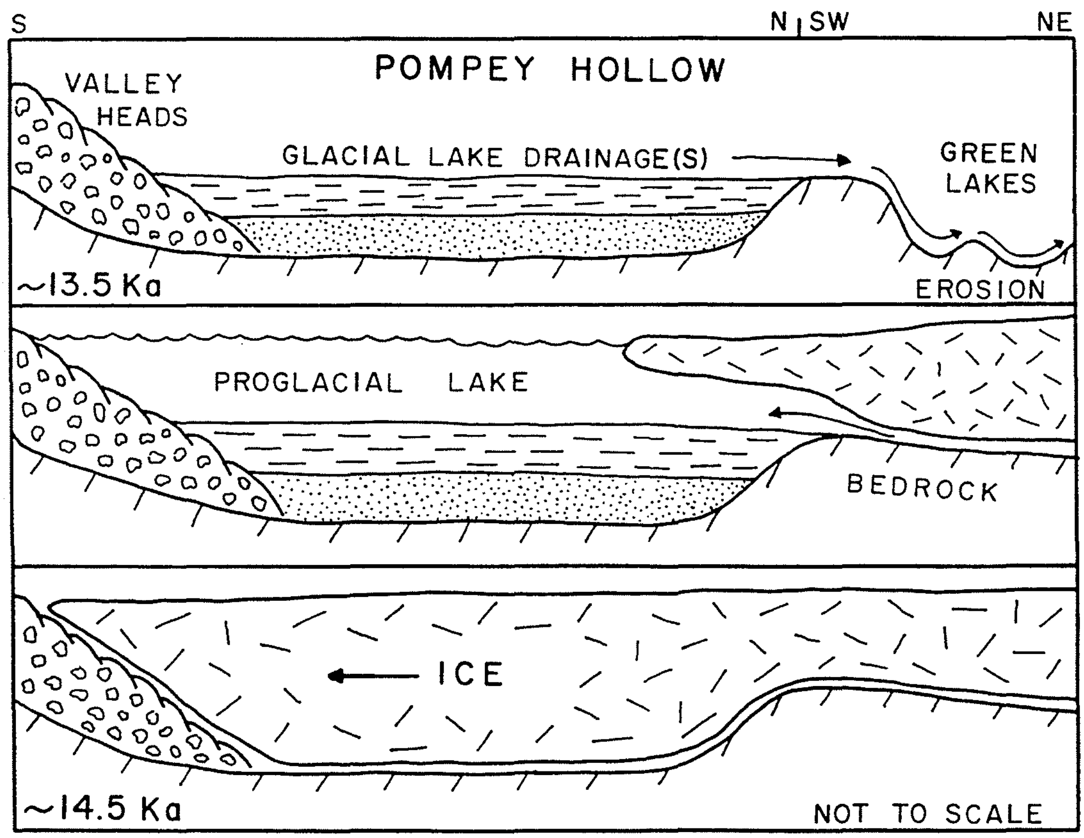


Fig. 5: Schematic cartoon illustrating possible source of meltwater for the Green Lakes plunge-pools. Panels are sequential from bottom (~14.5 ka) to top (~13.5 ka ^{14}C). Lowest panel illustrates surge of ice through Pompey Hollow and deposition of Valley Heads Moraine. Middle panel depicts retreating ice margin and the development of a proglacial lake with bottom sediment. In upper panel, ice withdraws to expose lower elevation outlet along Green Lakes meltwater channel and catastrophic outflow to erode plunge-pool basins. Note change in direction of panels from S-N to SW-NE. Glacial trough relationships based on Mullins et al. (1996).

The origin of the Green Lake and Round Lake basins is still uncertain although a number of hypotheses have been proposed (Muller, 1967), including: (1) plunge-pools; (2) solution sinkholes; and, (3) subglacial meltwater "potholes". Native American legend of "Kayahao" originally suggested that the lakes were volcanic in origin although there is no evidence to support it. This legend speaks of a Native American woman who infuriated the spirit of the lake which then "...spit forth fire and flame, (and) caused a great commotion of the waters..." (Gibson, 1925, p. 21).

The most widely accepted hypothesis is that these lake basins were scoured out at the base of large, late-glacial waterfalls. Based on stratigraphic relationships and morphology, Miner (1933) developed the hypothesis that the lakes are glacial plunge-pool basins (Fig. 6) which followed Vanuxem's (1839) earlier suggestion that they were eroded by a "whirlpool of great magnitude". However, problems with this hypothesis have centered around the source of water for the waterfalls plus the fact that there are two ("twin") lake basins.

Fairchild (1899) suggested that water impounded in highstanding proglacial lakes to the south of the Park may have cut the Green Lakes meltwater channel, whereas Sissons (1960) argued that water flowed from the Great Lakes in a complex system of marginal, subglacial and englacial conduits. If Fairchild's (1899) view of a proglacial source of water is correct, then Pompey Hollow, which connects with the Green Lakes meltwater channel at an elevation of ~530 ft. (162 m), is a likely source. Pompey Hollow, a 1 km wide, flat-floored (~780 ft., 238 m) trough, with as much as 520 ft. (159 m) of relief, extends south of the Park for ~20 km (Fig. 1). As ice receded from the Valley Heads Moraine at Delphi Falls shortly after 14.4 ka ^{14}C proglacial lake waters would likely have been impounded in Pompey Hollow, dammed by the Valley Heads Moraine to the south and the Laurentide ice sheet to the north (Fig. 5). In the nearby Finger Lakes, high-standing proglacial lakes existed until ~13.9 ka ^{14}C when the ice sheet retreated north of the Onondaga Escarpment opening up lower elevation outlets to the east which resulted in catastrophic drops of proglacial lake levels, cutting of gorges (glens), and a drainage reversal (Mullins et al., 1996). Muller (1967) has argued that the Green Lakes Channel originated as a pre-late Wisconsin feature. If correct, then as the northward retreating Laurentide Ice Sheet uncovered the precursor Green Lakes Channel impounded proglacial lake waters in Pompey Hollow (hundreds of feet higher in elevation) may have catastrophically poured through the Green Lakes meltwater channel (Fig. 5). Today, unlike the Finger Lakes, Pompey Hollow is a dry, flat-floored lake valley.

Such a scenario may have been similar to that documented for the Syracuse meltwater channels which experienced catastrophic floods with waterfalls the size of Niagra Falls when a 600 ft. (183 m) deep proglacial lake in the Onondaga-Tully trough drained to the east (Hand and Muller, 1972; Hand, 1978) creating a plunge-pool basin at Clark Reservation. Hand (1978) estimated that such catastrophic floods may have had water discharge as high as 7,000,000 ft.^{3}/sec. (~2.1x10^6 m^3/sec.) and flow velocities exceeding 50 ft./sec. (~15 m/sec.)! Similar catastrophic floods from Pompey Hollow, flowing over relatively resistant dolostone caprock as a waterfall in Green Lakes Channel may have eroded underlying shales to produce plunge pool basins which today we know as Green and Round Lakes.

The second problem which must be overcome for the plunge pool hypothesis to be viable, is the origin of "twin basins" which has previously been discussed by Muller (1967). Because simple headward erosion of a waterfall should result in a single gorge, Fairchild (1909) proposed a simple solution; that the Green and Round Lake basins do occupy a single bedrock trough but have been isolated by post-glacial sediment accumulation (Fig. 6). If correct, then the depth to bedrock beneath the wetland between Green and Round Lakes should be similar to that beneath the basins (i.e. >50 m). However, we have collected sediment cores only up to 11 m in length from this wetland (see Stop #4) which represent ~9,100 ^{14}C years. Also, one of our cores from the center of the wetland penetrated only 4.8 m and appeared to bottom on bedrock. Thus, there may be a bedrock high or sill separating Green and Round Lakes. A land-based seismic reflection/refraction profile along the axis of this wetland would likely resolve the question of whether or not the lakes are separated by a bedrock sill.

A second possibility is that the lake basins formed sequentially, with Green Lake having formed first. This would necessitate a phase of rapid waterfall retreat to form the elongated "neck" of Green Lake followed by a pause in the rate of retreat to form the main basin of Green Lake. A second phase of rapid waterfall retreat would then be required followed by a second pause to excavate the Round Lake basin. Such a scenario could be reconciled if the outpouring of proglacial lake waters from Pompey Hollow occurred in the two discrete phases, perhaps as a result of glacial oscillation (Muller, 1967).

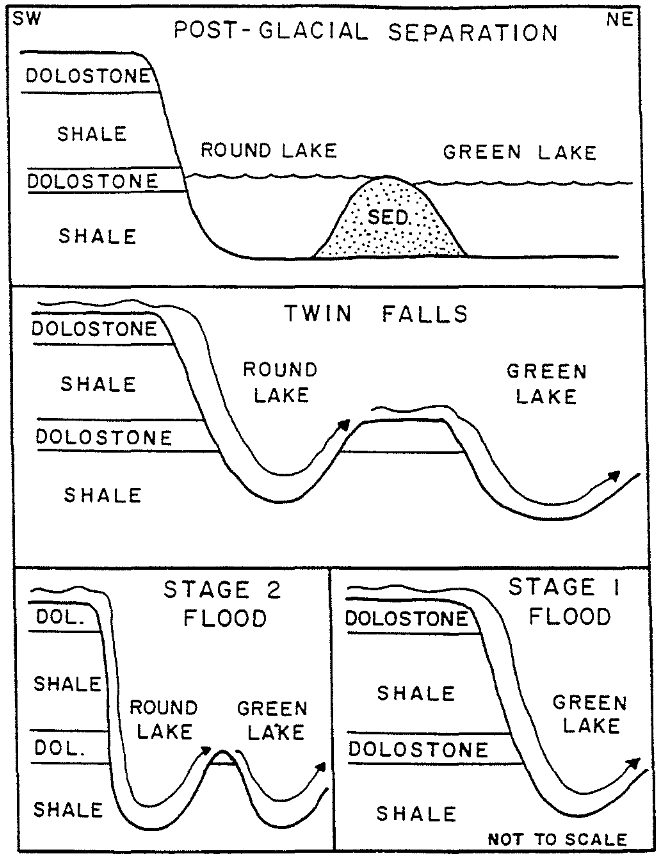


Fig. 6: Schematic cartoons illustrating possible explanations for "twin" plunge-pool basins at Green Lakes State Park. (Top) Post-glacial separation caused by sediment accumulation (Fairchild, 1909). (Middle) Twin falls flowing over two dolostone caprocks (Muller, 1967). (Bottom) Two stage outflow of proglacial lake waters impounded in Pompey Hollow.

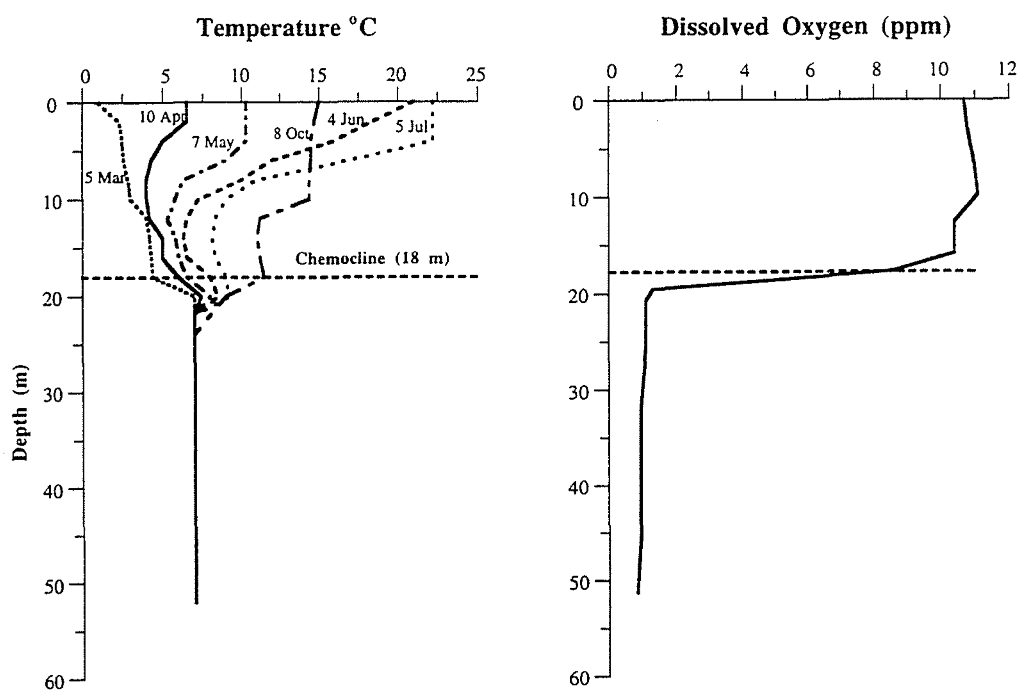


Fig. 7: Vertical distribution of temperature (left) and oxygen (right) versus water depth in Green Lake. Temperature data are seasonal, but note that water below chemocline is a constant 7 °C (from Culver and Brunskill, 1969). Note "anoxic" nature of water below the chemocline versus oxygenated water above (from Turano and Rand, 1967).

A third hypothesis, advocated by Muller, 1967), is that there were twin waterfalls which migrated headward simultaneously. For this mechanism to work there must have been some stratigraphic (or structural) control to initiate twin falls during a single meltwater episode. Muller (1967) suggests that both basins were controlled by water falling over dolostone caprocks; Green Lake basin by the "Middle Dolomite" of the Syracuse Formation and Round Lake basin by the "Upper Dolomite". The fact that there is no headwall scarp at the southwest end of Green Lake (as there is at Round Lake) would have to be explained by either incision and erosion of the lower caprock or by "burial" of the scarp by subsequent impounding of waters in Green Lake (Muller, 1967).

It is apparent from the above discussion that the origin of Green and Round Lakes continues to be uncertain. However, morphologically, they do appear to be plunge-pool basins but satisfactory explanations for the source of the water and why there are "twin basins" must be acceptable. Most useful would be the acquisition of subsurface data, both geophysical and drill core.

Stop #2: Limnologic Setting of Green Lake

From our vantage point at Stop #1 we will progress clockwise and walk a few hundred meters along the lake path to an open area just south of the diving platform. Here we will continue to have an open view of Green Lake and have an opportunity to present and discuss it's basic limnology (characteristics of the water column). Limnologically Green Lake is probably the best studied meromictic lakes in the world (Thompson et al., 1990); thus, there are considerable data from which to draw upon.

According to Thompson et al. (1990), Clark (1849) was the first to report (quoting L.C. Beck) that there was something unusual about the limnology of Green Lake: "Water drawn from a depth of one hundred and sixty-eight feet (51 m) was found to be strongly charged with sulphurated hydrogen... (which) blackened silver powerfully, and gave copious precipitates...". In 1931 Eggleton also noted the "foul odor of deeper mud deposits" and speculated that Green Lake might be stagnate, characterized by "incomplete spring and fall overturns" and perhaps a complete lack of whole-lake circulation. Eggleton's (1931) account of Green Lake is believed to be the first of a meromictic lake anywhere in North America (Thompson et al., 1990). Following his early assessment, Eggleton (1956) also reported results of a 25-year study of Green Lake concluding that "for all practical purposes, there is no significant amount of dissolved oxygen in the water below 19-20 m at any time of the year" (Thompson et al., 1990).

More recent studies have confirmed Eggleton's pioneering work on Green Lake, while enhancing our understanding of why Green Lake is meromictic. Of particular importance are publications by Deevey et al. (1963), Turano and Rand (1967), Takahashi et al. (1968), Brunskill and Ludlam (1969) and Torgensen et al. (1981). Based on a study of sulfur and carbon isotopes in Green Lake, Deevey et al. (1963) noted unusual quantities of sulphate which they suggested were derived from nearby outcrops of Silurian gypsum entering the lake as surface waters and perhaps "gypsiferous springs". Deevey et al. (1963) concluded that geochemically, Green Lake is more akin to Gulf Coast salt dome systems than to most temperate lakes.

In 1965, Turano and Rand (1967) conducted a series of measurements on the temperature and water chemistry of Green Lake from late winter (February 25) to early spring (April 18). They found that waters of the mixolimnion are colder (3-5 °C) than waters below the chemocline (~7 °C). However, surface water temperature of Green Lake varies greatly (3-22 °C) as a function of season whereas the monimolimnion maintains a relatively constant temperature of ~7 °C (Fig. 7; Brunskill and Ludlam, 1969).

Turano and Rand (1967) also found that dissolved oxygen concentrations are highest in the mixolimnion ($\geq$ 10 ppm) of the lake, decrease abruptly between 16 and 19 m and then remain relatively stable to the bottom of the lake (Fig. 7), consistent with previous results. However, unlike Eggleton (1956) who reported that the monimolimnion of Green Lake was completely devoid of oxygen, Turano and Rand (1967) did detect very low levels of oxygen (~1 ppm). Based on these data they concluded that the monimolimnion of Green Lake is not completely stagnant but rather must have some exchange with other waters, perhaps groundwater. However, Turano and Rand (1967) are the only investigations who have reported some oxygen below the chemocline.

Turano and Rand (1967) also found that both sulfate (oxidized) and suphide (reduced) increase below the chemocline (Fig. 8) but that only sulphate is present above the chemocline. In the monimolimnion, sulphide concentrations are high (up to 50 ppm) representing a "very considerable quantity" of H_2S (Turano and Rand, 1967). Maximum concentrations of sulphide occur near the bottom of the lake suggesting a source via the degradation of organic matter in the sediment primarily by sulfur reducing bacteria (Turano and Rand, 1967). However, sulphide is also oxidized by purple, photosynthetic bacteria that inhabit the chemocline in Green Lake (Culver and Brunskill, 1969). The abundance of sulphide in the monimolimnion and absence in the mixolimnion, of Green Lake, though, is directly related to oxygen concentrations.

Carbon dioxide and pH values in Green Lake are mirror images of each other (Fig. 9): CO_2 concentrations are relatively low in the mixolimnion and high in the monimolimnion whereas just the opposite is true for pH (Turano and Rand, 1967). CO_2 concentrations are controlled largely by photosynthesis above the chemocline and decomposition of organic matter below the chemocline. pH (a negative measure of hydrogen ion concentration) is inversely controlled by CO_2 concentrations. However, pH values are neutral or above (~7.2-8.0) throughout the entire water column of Green Lake (Fig. 9).

Turano and Rand (1967) also used data on water hardness (Ca, Mg concentrations), alkalinity, pH and temperature to calculate the saturation state of Green Lake with respect to calcium carbonate, both above and below the chemocline. They found that both water layers of the lake are "considerably supersaturated" with respect to calcium carbonate which should promote the precipitation of large quantities of marl.

In 1968 Takahashi et al. published their geochemical study of Green Lake. Although Green Lake had traditionally been believed to be meromictic because of "its great depth... and small area..." (Deevey et al., 1963), Takahashi et al. (1968) concluded that Green Lake's meromixis is a result of two different water types entering the lake. Below ~18 m groundwater entering the basin must be nearly twice as saline as surface waters, regardless of temperature (Takahashi et al., 1968). They found that surface waters are $4x10^{-4}$ g/cm^3 less dense than deep water. Thus, the permanent stratification of Green Lakes can be fully explained by salinity differences, which is contrary to most lakes which stratify due to temperature-controlled differences in density. This is verified by conductivity (a gross measure of salinity) measurements in Green Lake which display an abrupt increase at ~18 m water depth (Fig. 10; Brunskill and Ludlam, 1969). Thus, the meromixis of Green Lake is the consequence of a chemically-controlled density boundary or chemocline. However, the salinity of Green Lake is controlled not by Na and Cl as in the oceans, but rather by Ca and SO^4 derived from gypsum in surrounding bedrock (Takahashi et al., 1968).

That the chemocline in Green Lake is the result of the inflow of saline groundwater is further supported by chemical studies of interstitial waters recovered from bottom sediments. Brunskill and Harriss (1969) found that interstitial waters in a core from the deep (>50 m) central basin of Green Lake show continuous increases of Na, Mg, Sr, Ca, Cl and Br with depth indicating the mixing of monimolimnetic waters with more saline groundwater diffusing upwards into the bottom sediment.

Takahashi et al. (1968) also estimated the residence time (the amount of time a water molecule resides in the lake before being removed) of waters in Green Lake. It had previously been known that the surface outflow of Green Lake exceeds the surface inflow by about an order of magnitude (Turano and Rand, 1967); thus, there must be appreciable inflow of groundwaters. Because of this, plus the meromictic nature of the lake, considerable differences in residence times might be expected. Takahashi et al. (1968) indicated a residence time on the order of only 2 years for surface waters whereas deep waters are replaced no less than once every 35 years. Brunskill and Ludlam (1969) concurred with the short surface water residence times estimated by Takahashi et al. (1968), and further argued that surface inflow supplies <50% of water flow into Green Lake. Based on ^3H-^3He water mass ages in Green Lake, Torgensen et al. (1981) have confirmed the short residence times of the mixolimnion. However, they concluded that the residence time of water in the monimolimnion is considerably less than previously reported, ranging up to only ~7 years. Torgensen et al. (1981) also present evidence for a secondary chemocline at ~33m where there is a significant increase in the concentration of S, CH_4, CO_2, Na, and Cl. They explain both the short residence time of the monimolimnion and the secondary chemocline as a consequence of diffuse groundwater input.

Stop # 3: Precipitation of Calcium Carbonate

From Stop #2, near the dive platform of Green Lake, we will continue our clockwise walk around the lake to Deadman's Point, located at the eastern junction of the "neck" and main basin of Green Lake. Here we will first

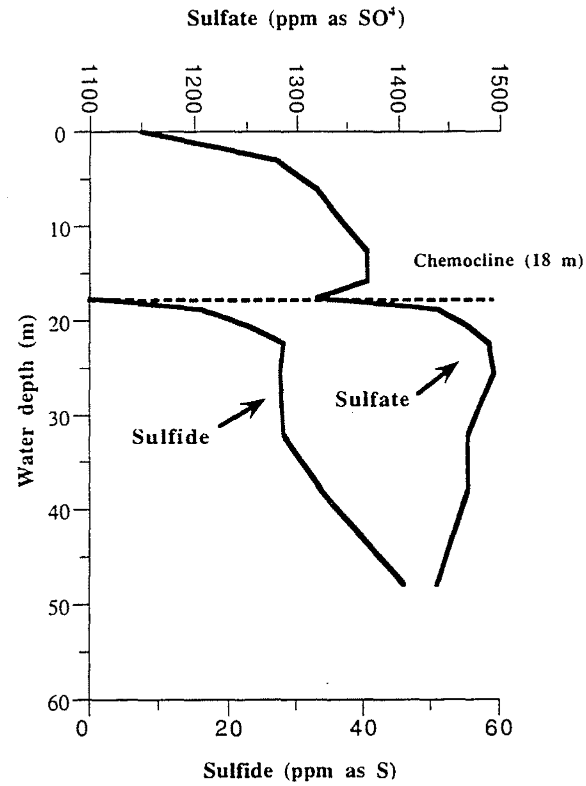


Fig. 8: Distribution of sulphate (oxidized) and sulphide (reduced) versus water depth in Green Lake. Note that sulphate is present both above and below chemocline whereas sulphide occurs only below ~18 m water depth. Decrease in sulphate concentration along chemocline is due to presence of purple sulphate reducing bacteria (from Turano and Rand, 1967).

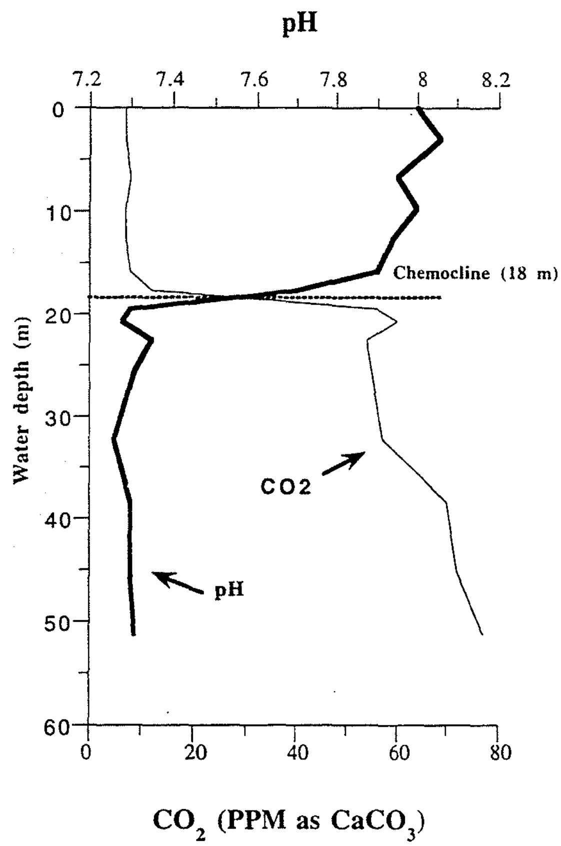


Fig. 9: Vertical distribution of carbon dioxide and pH versus water depth in Green Lake. Note abundance of CO_2 below chemocline and a concomitant decrease in pH. However, all pH values in Green Lake are above neutral (i.e.->7; from Turano and Rand, 1967).

discuss the seasonal, open water precipitation of calcium carbonate ("whitings") from Green Lake, followed by a discussion of littoral zone precipitation and the formation of the prominent "reef" or bioherm that we will see at Deadman's Point.

Whiting Events

One of the most commonly asked questions about Green Lake is "why is it green"? Well, the truth is that the lake is not always green. During winter (when not frozen) and early spring months, open waters of the lake tend to be blue, whereas they become green from late spring through fall, when most people visit the Park.

The color of water in a lake varies largely as a function of the amount and type of particulate matter in the water column. Oligotrophic (low nutrient) lakes with little particulate matter, like Skaneateles Lake, tend to display a blue color, whereas those lakes with high concentrations of particulate matter from primary biological productivity, like Oneida Lake, tend to be green, especially during the summer. The reason for this is that clear water scatters light predominantly from the blue portion of the electromagnectic spectrum, whereas when the amount of particulate matter increases in the water column longer wavelength radiation is preferentially scattered, and the lake appears green (Wetzel, 1975).

Green Lake is an oligotropic lake (Thompson et al., 1997) with annual primary productivity estimated at ~290 g C/m^2 (Culver and Brunskill, 1969). However, ~83% of this annual primary production in Green Lake occurs within the primary chemocline (~18 m water depth) due to the photosynthetic activity of purple sulfide oxidizing bacteria. Obviously, an alternative explanation for the seasonal green color of Green Lake is required. A hint comes from Wetzel's (1975) statement that lake sediments rich in light-colored calcium carbonate (marl) reflect more light than non-marl sediments, and that lakes with large amounts of suspended calcium carbonate backscatter energy from the green spectrum of visible light.

The abundance of calcium carbonate deposits in Green Lake has long been known (Clark, 1849; Bradley, 1929, 1963; Howe, 1932; Eggleton, 1956; Brunskill, 1969). Takahashi et al. (1968) as well as Brunskill (1969) documented that the entire water column of Green Lake is supersaturated with respect to calcium carbonate. However, based on $\delta^{13}C$ data, Takahashi et al. (1968) concluded that calcite precipitates only from the surface waters of Green Lake. They showed (via data from Deevey et al., 1963) that calcite crystals from the sediment of Green Lake have a $\delta^{13}C$ (a measure of the ratio of $^{13}C/^{12}C$) value of -4 ‰, which is much more similar to that of surface waters (-7 ‰) than to waters below the chemocline (-18 ‰). By independent means Brunskill (1969) also showed that a 3-4 fold increase in calcite supersaturation occurs in the mixolimnion of Green Lake from May through August, which is the time of massive precipitation of calcite at rates as high as 2 g/m^2/day. From sediment trap studies, Brunskill (1969) found calcite crystal loads in the water column of 35 g/m^2 in June and July of 1967 (Fig. 11). He argued that the precipitation of calcite in Green Lake is an inorganic process controlled largely by seasonal temperature change- "...temperature is the direct causal factor in the initiation of calcite precipitation... with photosynthesis playing only a secondary, and probably minor, role" (Brunskill, 1969, p. 844).

However, this concept of a purely inorganic origin for calcite precipitation in Green Lake has recently been challenged by Thompson and Ferris (1990) and Thompson et al. (1997). Based on experimental and transmission electron microscopy results of waters collected from Green Lake, Thompson and Ferris (1990) concluded that calcite precipitation is the direct result of the photosynthetic activity of the cyanobacterial picoplankton (0.2-2.0 μm) *Synechococcus* which occurs at concentrations of ~10^5 cells/ml of water in the mixolimnion of Green Lake. They convincingly demonstrated that calcite (as well as a gypsum and perhaps magnesite) precipitates epicelluarly around *Synechococcus* cells (Fig. 12). The reason for this is that the photosynthetic metabolism of *Synechococcus* directly results in the alkalinization (increase pH) of the microenvironment immediately surrounding the cell because of its ability to use bicarbonate (HCO_3^-) as a carbon source which results in the production of hydroxyl (OH^-) ions (Thompson and Ferris, 1990). Calcium (Ca^{2+}) ions which concentrate on the surface of *Synechococcus* cells serve as nucleation sites for the precipitation of calcite and gypsum (Fig. 13).

This new view of Green Lake "whiting events" (direct precipitation of fine-grained calcite from the open water column) was further tested by direct monthly field measurements from May 1989 to April 1990 (Thompson et al., 1997). Secchi disk data (Fig. 14) clearly document the onset of the annual Green Lake whiting in May, when water clarity diminishes dramatically (Secchi disk readings decrease from 18 to 8 m). Minimum water clarity (Secchi disk readings of 4.5 m) occurred during July and August and then gradually improved through fall, winter and early spring

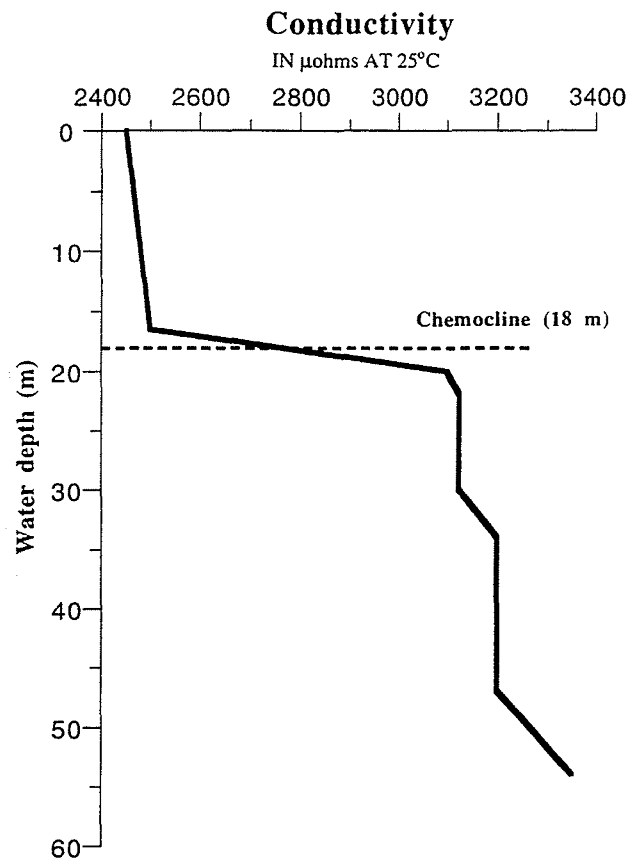


Fig. 10: Vertical distribution of conductivity versus water depth in Green Lake. Conductivity is rough measure of salinity. Note abrupt increase along chemocline at ~18 m (from Brunskill and Ludlam, 1969).

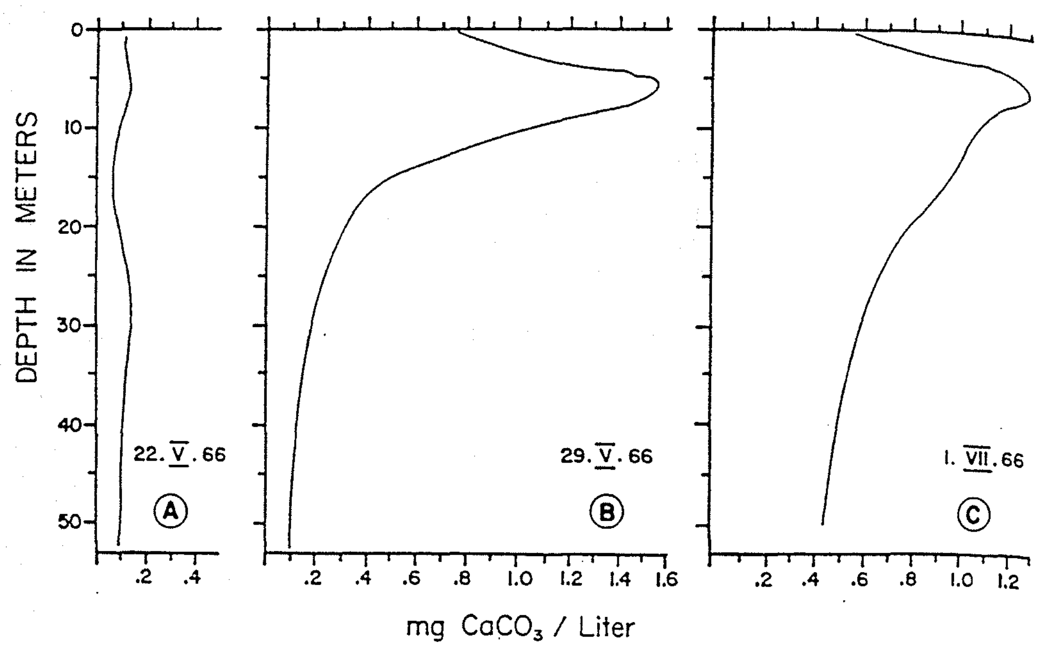


Fig. 11: Distribution of suspended calcite crystals versus water depth in Green Lake on May 22, 1966 (left), May 29, 1966 (center), and July 1, 1966 (right). Note peak abundance (~1.6 mg $CaCO_3$/l) at ~5 m water depth on May 29, 1966 due to spring whiting event (from Brunskill, 1969).

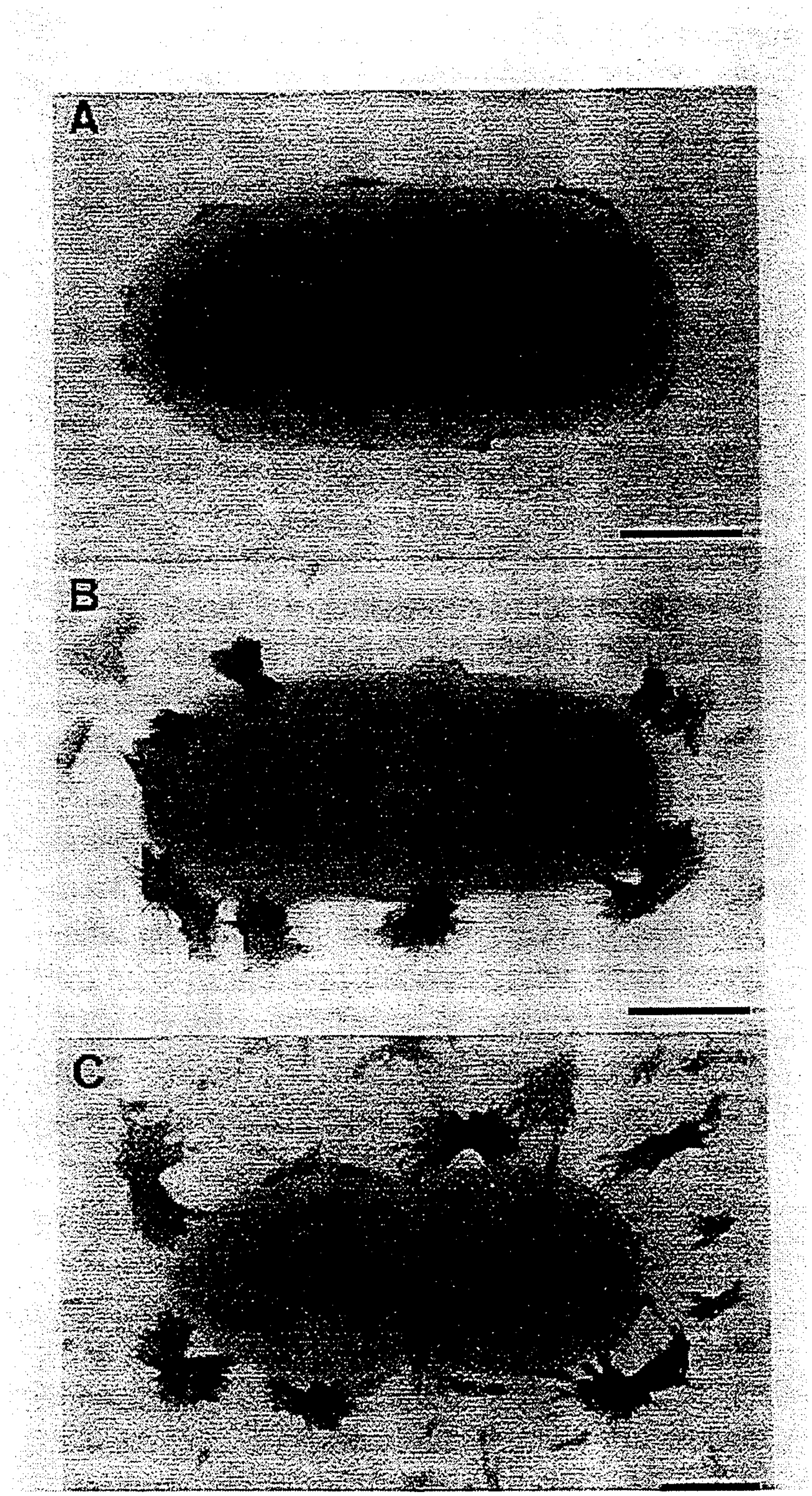


Fig. 12: Transmission electron microscope (TEM) photographs illustrating progressive (A to C) mineral deposition on surface of *Synechococcus* cell. Cell beginning to divide in C; bar scale =500 nm (from Thompson and Ferris, 1990).

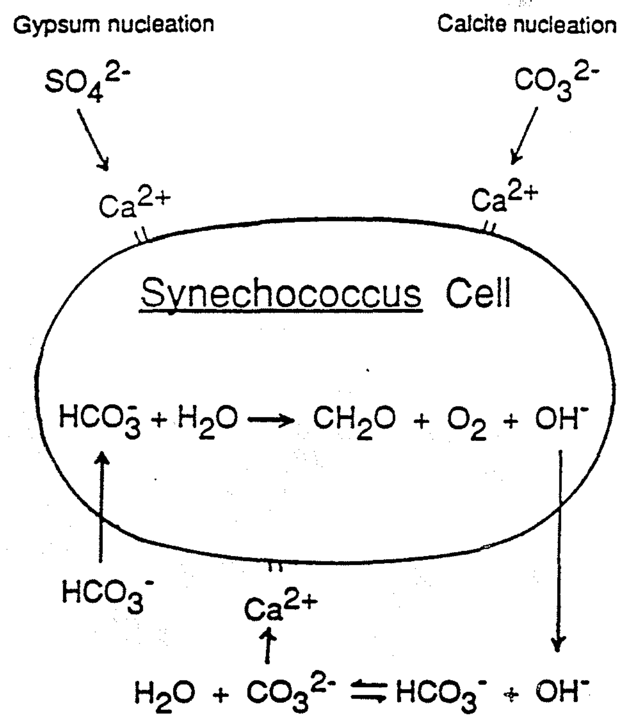


Fig. 13: Schematic cartoon illustrating basic biogeochemical reactions both on and within idealized *Synechococcus* cell during mineral precipitation (from Thompson and Ferris, 1990).

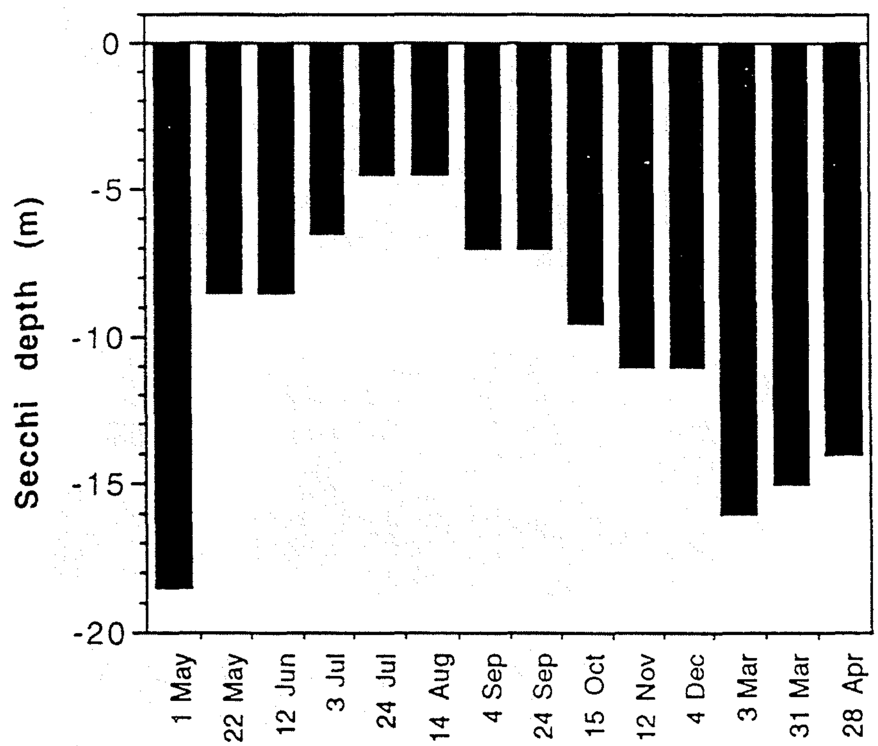


Fig. 14: Seasonal (1989) Secchi Disk data for Green Lake documenting low visibility (i.e.- low water clarity) during late spring/summer months due to suspended calcite crystals precipitated during whiting event (from Thompson et al., 1997).

(Thompson et al., 1997). During this whiting event the color of Green Lake changed from bluish-green to greenish-white (Thompson et al., 1997).

The onset of this whiting event was directly related to the spring bloom of *Synechococcus* triggered by increasing temperatures and light intensities (Fig. 15; Thompson et al., 1997). That *Synechococcus* is directly responsible for the precipitation of calcite is supported by $\delta^{13}C$ data which reveal that whiting calcite (as well as sediment calcite) in Green Lake is enriched by ~3-5 ‰ relative to the $\delta^{13}C$ of dissolved inorganic carbon (DIC) of summer surface waters (Thompson et al., 1997). The reason for this isotopic enrichment is the preferential uptake of relatively light ^{12}C by the photosynthetic picoplankton (Fig. 16). According to Thompson et al. (1997) individual cells of *Synechococcus* can precipitate calcite crystals a number of times per season because they periodically shed their outer "S-layer" once it is mineralized.

To summarize, the reason that Green Lake is green (particularly during spring and summer) is because of the widespread precipitation of calcium carbonate. Calcite crystals precipitated around picoplankton result in an increase of particulate matter in the water column which preferentially scatters back green light. In addition, the accumulation of calcium carbonate in littoral zone waters of Green Lake provides a permanent surface of reflection for wavelengths within the green spectrum of visible light.

Bioherms

Here at Stop #3 we can literally stand on a massive, *in situ* accumulation of calcium carbonate or bioherm. Although the best developed bioherms in Green Lake are here at Deadman's Point, they are also found across the lake along the south-facing shoreline (Fig. 17). These unusual littoral zone accumulations of massive carbonate in Green Lake have been known for many decades (Walcott, 1914; Bradley, 1929; Howe, 1932; Eggleton, 1956).

The bioherms of Green Lake have traditionally been viewed as algal in origin, a concept most recently advocated by Dean and Eggleston (1975), Eggleston and Dean (1976), Dean (1981), and Dean and Fouch (1983). In fact, Dean and Eggleston (1975) compared these freshwater bioherms of Green Lake to marine algal "cup" reefs of Bermuda.

However, Thompson et al. (1990) have reinterpreted the "algal bioherms" of Green Lake as <u>thrombolitic microbialites</u> which simply means a massive carbonate accumulation precipitated by microbes (rather than algae). SCUBA observations here at Deadman's Point revealed that the microbialites have grown up and out from dolostone beds within the Syracuse Formation. They are lobate features that protrude as overhangs into Green Lake and are as much as 10 m thick (Fig. 18). The undersides of the overhangs are colonized by aquatic mosses as well as sponges, and have laminated stromatolitic caps (Thompson et al., 1990). The internal structure of these features is characterized by a non-laminar, clotted texture. The active, outer growing portion of these bioherms is heavily colonized by a benthic variety of *Synechococcus* (Thompson et al., 1990). They believe that this benthic species of cyanobacteria, rather than algae, is responsible for the *in situ* development of the bioherms in Green Lake. *Synechococcus* is specifically adapted for life in oligotrophic, hardwater lakes like Green Lake because of its ability to directly use bicarbonate (HCO_3^-) as a carbon source and its very low nutrient requirements (Thompson et al., 1990).

Benthic forms of *Synechococcus* colonize most any hard surface in the lake as evidenced by the accumulation of calcite on human artifacts such as bottles and cans (Fig. 19; Dean and Fouch, 1983). It is also common to see a white calcite coating on trees that have fallen into the lake. However, if one assumes that the bioherms at Deadman's Point originated shortly after Green Lake formed (~11-12 ka) their growth rate would only be ~1 mm/year (Thompson et al., 1990).

Before leaving Deadman's Point for our next stop, please take a few minutes to examine the marl deposit which extends to ~1 m <u>above</u> present day lake level. This calcite-rich deposit contains an abundance of lacustrine, littoral zone gastropods, thus providing clear evidence for a previous highstand of Green Lake (Eggleton, 1956). Harmon (1970) assumed that the level of Green Lake dropped when the Erie Canal (which is located just north of Green Lake; Fig. 1) was built in the early 1800's. However, as we will see at our next stop, there is a well preserved record of natural lake level fluctuations beneath the wetland which connects Green and Round Lakes.

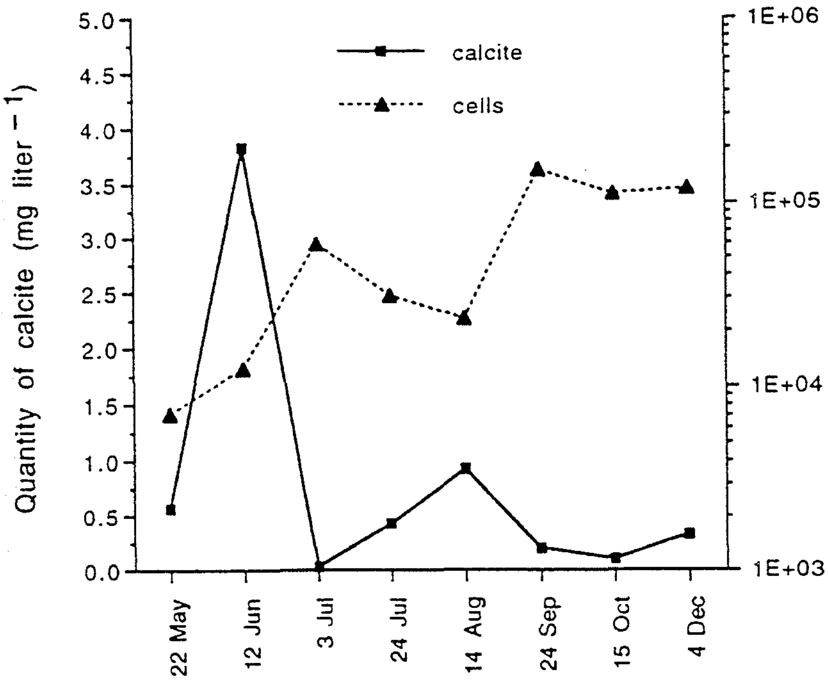


Fig. 15: Seasonal (1989) abundance of suspended calcite crystals and *Synechococcus* cells in Green Lake. Note late spring peak (~4 mg/l) of calcite crystals at 4 m water depth is coincident with the seasonal rise in the abundance of cyanobacterial picoplankton (from Thompson et al., 1997).

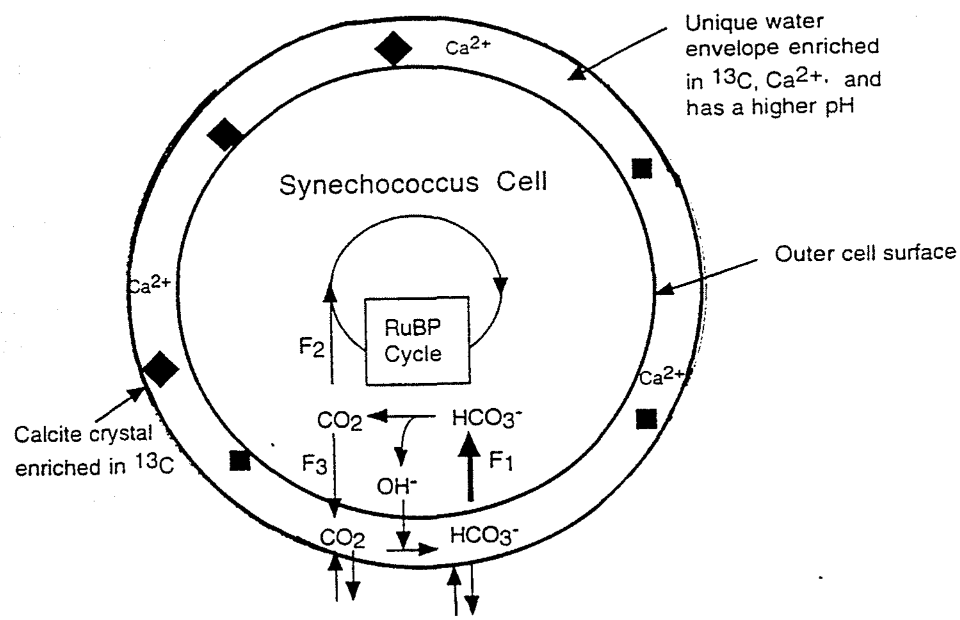


Fig. 16: Schematic cartoon illustrating *Synechococcus* metabolism reactions and epicelluar precipitation of calcite crystals in surrounding microenvironment. $\delta^{13}C$ enrichment of calcite due to preferential uptake of ^{12}C by the picoplankton (from Thompson et al., 1997).

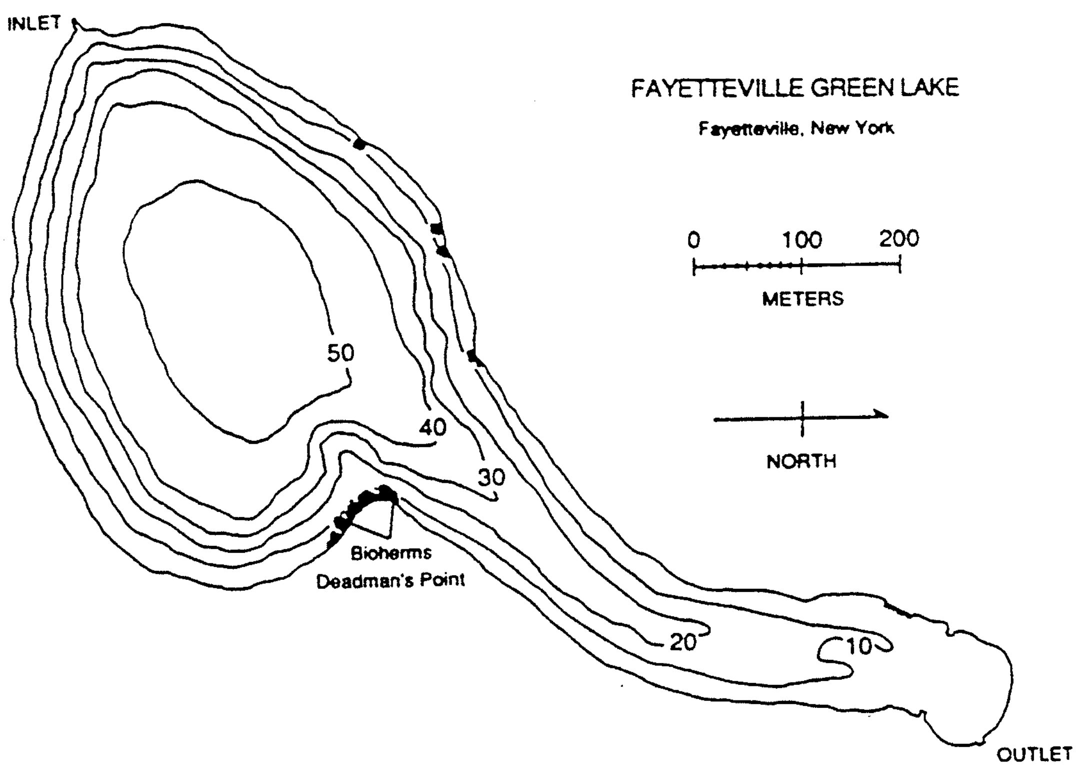


Fig. 17: Distribution of bioherms along the shore of Green Lake. Note major development at Deadman's Point (Stop #3) as well as lesser development along south-facing littoral zone (from Thompson et al., 1990). Contours are in meters.

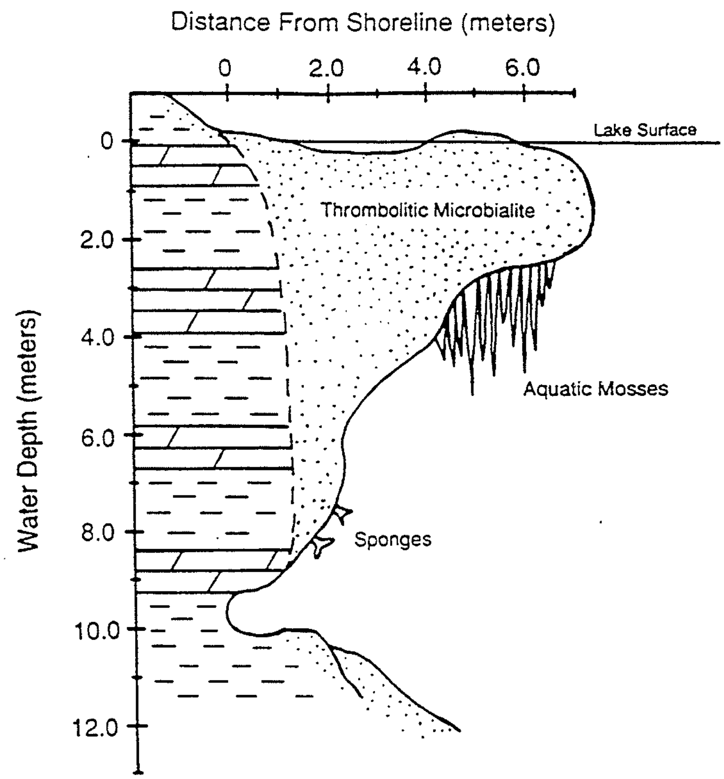


Fig. 18: Vertical cross-section of "bioherm" (thrombolitic microbialite) at Deadman's Point (from Thompson et al., 1990).

Stop #4: Wetland Stratigraphy and Natural Lake Level Fluctuations

From Deadman's Point we will continue our clockwise journey to the southern end of Green Lake. Here the path will divide and we will bear to the left. This will take us along the intervening wetland between the two lake basins (Figs. 2 and 3). We will stop at the next juncture of the path at the east end of Round Lake to present the subsurface stratigraphy of the wetland and discuss its implications for natural lake level fluctuations as well as regional paleoclimate change over the past ~8,000 years. Sediment cores previously recovered from the wetland will be available for viewing and discussion at this stop.

During the summer of 1996 we collected sediment cores, up to 11.2 m long, from three sites in the wetland between Green and Round Lakes (Fig. 20). There is at least 11 m of sediment beneath both the eastern and western ends of the wetland (Fig. 20). However, at the middle site only ~5 m of sediment was recovered before terminating on what appeared to be bedrock. If correct, there is a very shallow bedrock sill separating the Green and Round Lake basins.

Most of the sediment recovered at all three sites is fossiliferous marl consisting of 60-90% calcium carbonate (Figs. 21 and 22). The abundance of lacustrine fossils indicates deposition in relatively shallow, oxic waters (i.e.- littoral zone). However, a number of organic-rich (up to 90%) peat layers, representing various wetland environments, were also recovered and radiocarbon dated (Fig. 21). Based on radiocarbon data, the cores extend back to ~9.1 ka ^{14}C. Average sediment accumulation rates for the long cores are on the order of ~1.2 m/1000 years, although intervals of marl have accumulated at rates of up to 2.6 m/1000 years (Fig. 23).

Of particular interest is the series of marl-peat cycles (N=4) recovered at core site GLWL-3 nearest to Green Lake (Fig. 22). The four peat layers have been radiocarbon dated (University of Texas Radiocarbon Lab #'s TX-9191, TX-9136 through TX-9133), from bottom to top, at 7,570±70 years, 5,620±70 years, 4,545±65 years, and 2,169±50 years. Calculation of an average period for these cycles yields an average of ~1800 years (range 1075 to 2385).

The marls obviously represent open lacustrine connections between Green and Round Lakes whereas the peats represent wetland environments similar to those that exist today. The modern surface peat has developed during the past ~2,000 years. Thus, it seems highly unlikely that the marl outcrop that we saw ~1 m above modern lake level at Deadman's Point was stranded by an anthropogenic lowering of lake level less than 200 years ago during construction of the Erie Canal. This is supported by tree-ring counts from the wetland which indicate that living trees predate the construction of the Erie Canal (Brunskill and Ludlam, 1967).

The marl-peat cycles most likely record natural fluctuations of relative lake level approximately every 2,000 years over the past 8,000 ^{14}C years. If we interpret such natural lake level fluctuations (Fig. 24) as products of regional changes in precipitation minus evaporation, the data further imply relatively dry climatic conditions at ~ 8 ka, 6 ka, 4 ka, and 2 ka with relatively wet climates inbetween. This is consistent with results from a recent series of cores collected from the Montezuma wetlands (north end of Cayuga Lake) which also revealed relatively dry events at ~8 ka, 6 ka, and 4 ka (Mullins, in prep.). Thus, the relative lake level fluctuations recorded from the wetland in Green Lakes State Park are likely a response to regional climate changes rather than local drainage basin effects.

The ~ 8 ka ^{14}C relative lake lowstand at Green Lakes coincides with the end of a cool, dry period (10.1-8.2 ka ^{14}C) defined by stable isotope data from Seneca Lake (Anderson et al., 1997). This event followed the well-known Younger Dryas cold interval (~10.8-10.3 ka ^{14}C), and was correlative with a rapid phase of melting of the Laurentide ice sheet which poured large volumes of cold, isotopically light meltwater into the Great Lakes (Anderson et al., 1997). It also occurred quite close (~7.5 ka ^{14}C) to a recently discovered North Atlantic cool/dry event defined in Greenland ice cores (Alley et al., 1997).

The ~6 ka ^{14}C relative lake lowering defined at Green Lakes may also be a reflection of broader scale climate changes. In her review of natural lake levels throughout eastern North America, Harrison (1989) discovered that lakes here in general reached their lowest levels ~6 ka ^{14}C, implying widespread drought for at least a short interval of time. Cores from Montezuma wetlands also display a marked drop in the percentage of calcium carbonate at this time, implying relatively dry and/or cool conditions (Mullins, in prep).

The ~4 ka ^{14}C relative lake drop at Green Lakes coincides with widespread regional peat development, which effectively ceased marl deposition in the Finger Lakes region (Mullins, in prep.) Approximately 4 ka also marks the

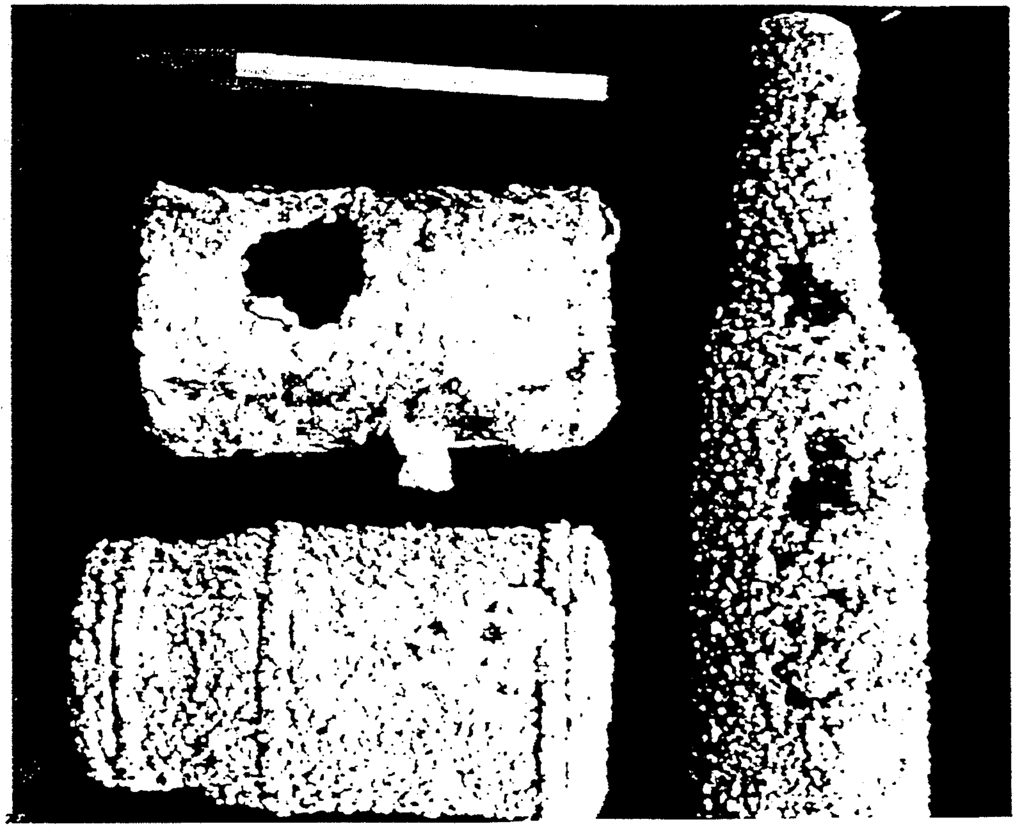


Fig. 19: Photograph of bottle and cans recovered from Green Lake with a thick coating of calcium carbonate documenting rapid rate of precipitation (from Dean and Fouch, 1983).

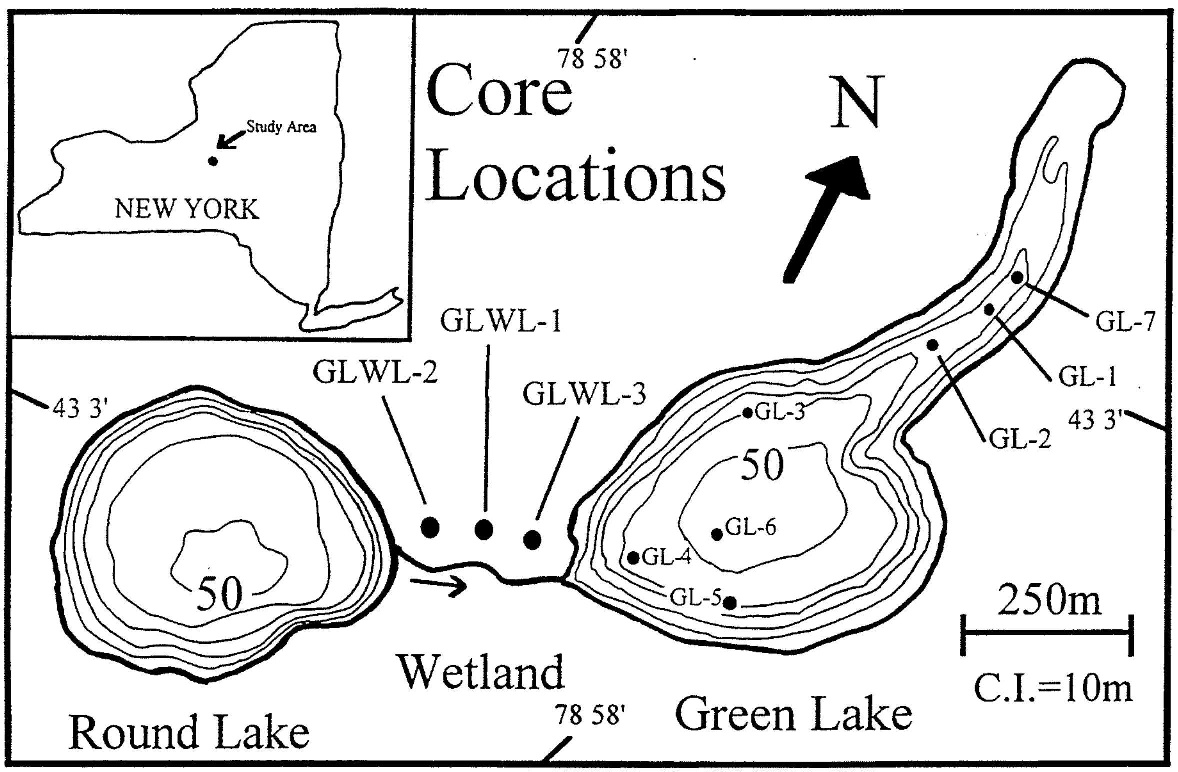


Fig. 20: Bathymetry of Green and Round Lakes including location of Green Lake (GL) sediment cores as well as cores recovered from the Green Lake wetland (GLWL) that will be discussed at stops #4 and 5.

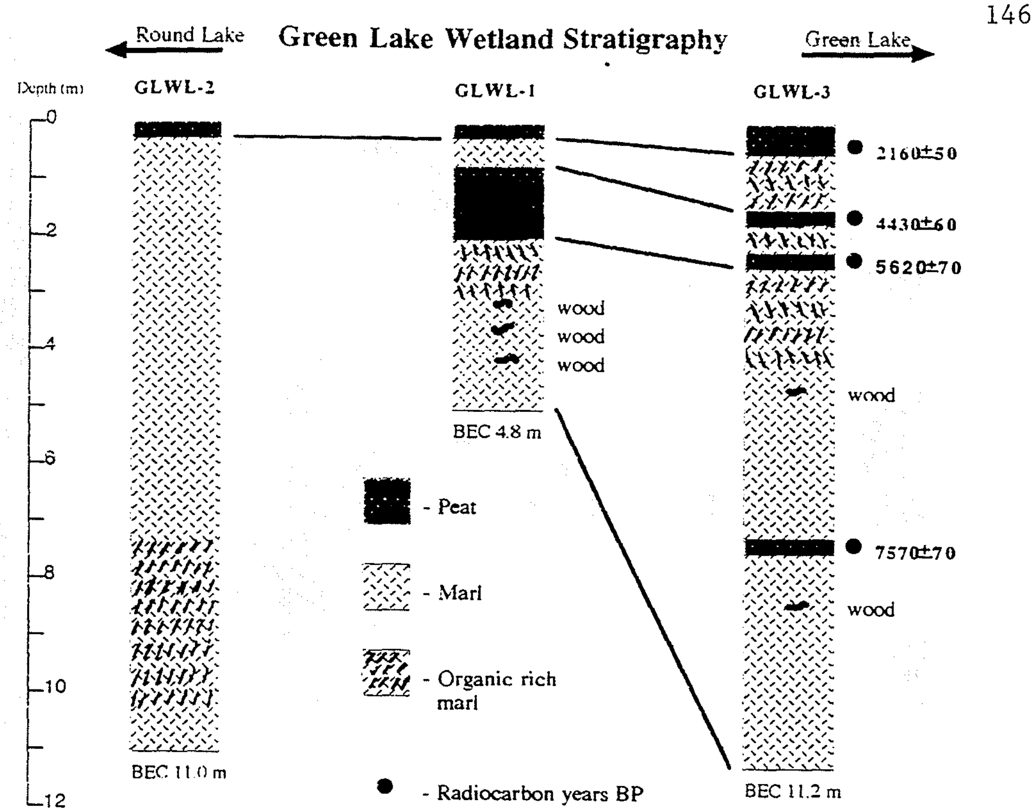


Fig. 21: Lithostratigraphy of the three wetland cores (see Fig. 20) and radiocarbon results. Note abundance of open-lacustrine marl (calcium carbonate rich deposits) with intervening peat (wetland) layers. BEC= bottom entire core.

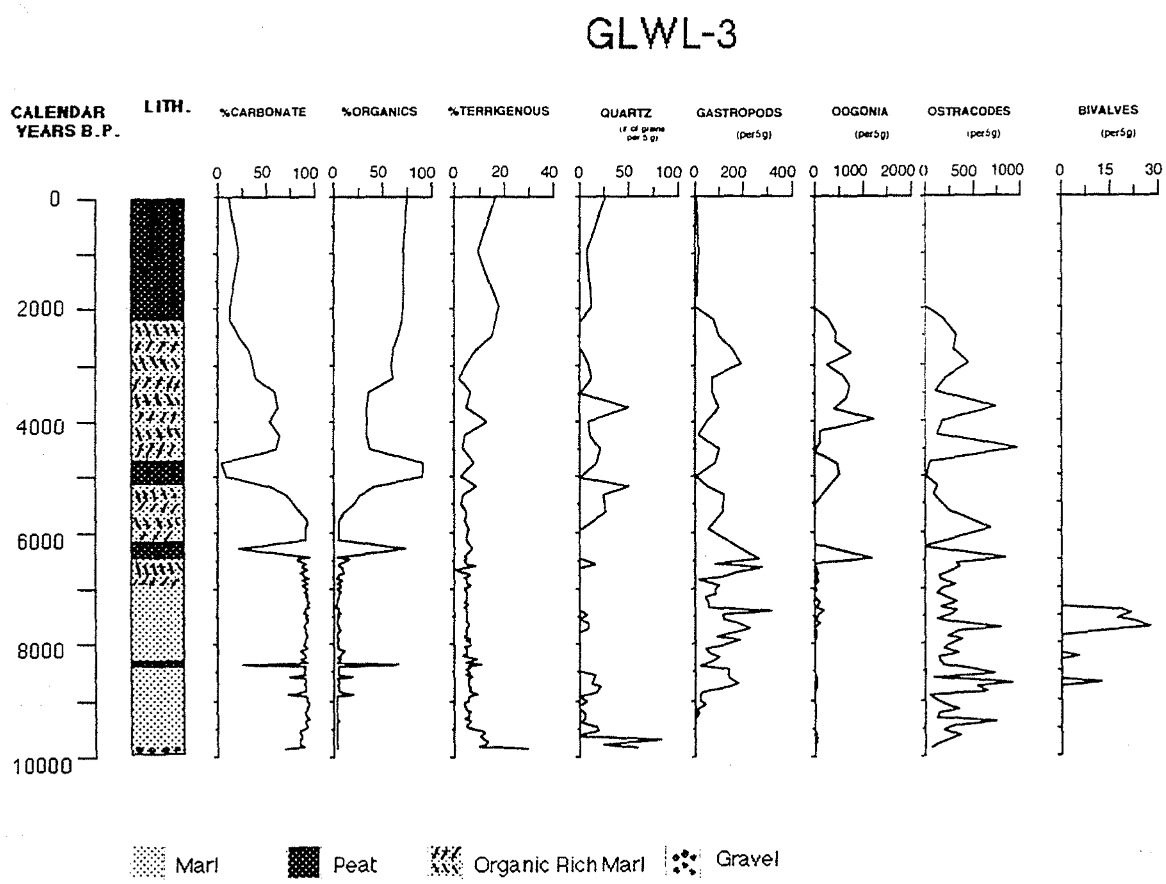


Fig. 22: Loss-on-ignition data versus depth for Green Lake wetland (GLWL) core 3. TOM= total organic matter. Note high percentage of calcium carbonate in marl layers (~90%) and high TOM values (up to 90%) for peats. Also note relatively high percentage of terrigenous material (~30%) at base of core where shale chips were also recovered, suggesting that core bottomed on bedrock.

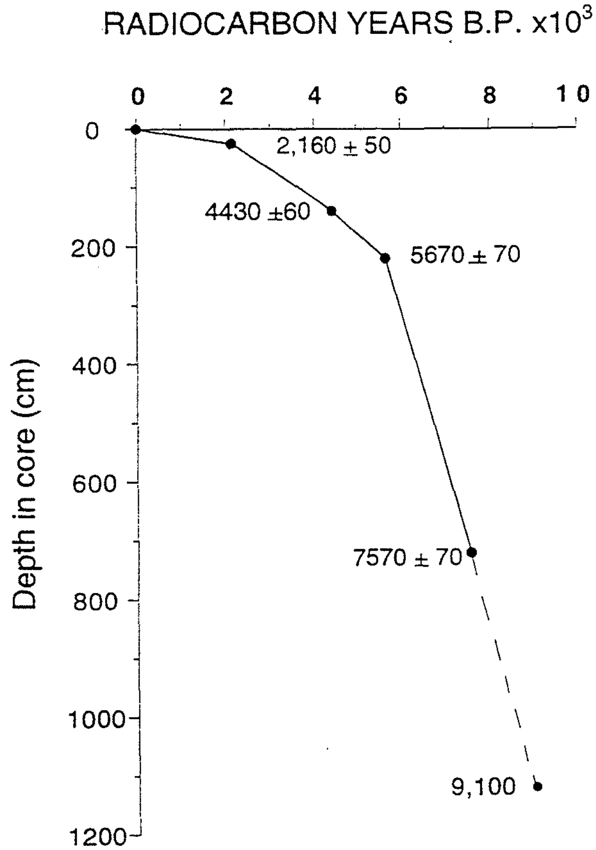


Fig. 23: Age versus depth curve for Green Lake wetland (GLWL) core 3 based on radiocarbon results. Note relatively rapid sediment accumulation between ~9.1 ka and 5.6 ka ^{14}C followed by a gradual decline, suggesting the rapid infill of a bedrock depression.

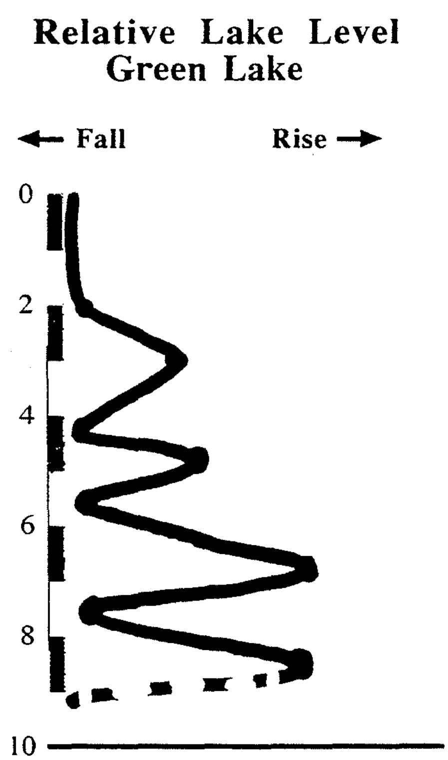


Fig. 24: Relative lake-level curve for Green Lake based on radiocarbon dates on peat layers contained in GLWL-3. Note relative lake lowstands at ~7.6 ka, 5.6 ka, 4.4 ka, and 2.2 ka ^{14}C.

end of the mid-Holocene Hypsithermal (~9-4 ka ^{14}C) in central New York State, which was mostly relatively warm and wet (Dwyer et al., 1996).

Why relative lake level fluctuations in Green Lakes should occur every ~2000 years is open to speculation. Perhaps they are related to regional climate changes caused by changes in the production of North Atlantic Deep Water (and thus heat transport) which presumably caused the Younger Dryas climate reversal (Broecker et al., 1989); or, perhaps they are related to shifts in the mean position of the jet stream and atmospheric pressure systems (Dwyer et al., 1996). However, there may also be an intriguing relationship between lake level fluctuations at Green Lake and natural variations in solar irradiance caused by sunspot activity. Wigley and Kelly (1990) have developed a proxy for prehistorical solar variation based on the difference between tree-ring counts and ^{14}C age, which they as well as Karlen and Kuylenstierna (1996), have used to explain Holocene climate changes. The peats dated from the Green Lake wetland match-up well with relatively high amounts of solar irradiance at ~8 ka, 6 ka, 4 ka and 2 ka ^{14}C (Fig. 25). Greater amounts of solar energy at these times may have led to increased rates of evaporation and thus, relative lake-level lowerings. Although this potential linkage between solar variability and lake levels is speculative, it is certainly deserving of further thought and investigation.

Stop #5: Environmental Changes During The Past 2,500 Years: Short Cores From The Deep Lake

From our previous vantage point of the wetland between Green and Round Lakes we will continue to proceed clockwise around Round Lake. During the course of this part of the fieldtrip keep a sharp eye out for littoral zone bioherms similar to (but smaller than) those at Deadman's Point as well as marl benches indicative of a previous high stand(s) of lake level. At the southwest end of Round Lake we will also be able to observe the inlet to the lake as well as the ~100 ft. (30.5 m) high cliff over which a late glacial waterfall presumably flowed to carve out this plunge-pool basin. We will then walk, uninterrupted, around all of Round Lake, past the north-side of the wetland, and back to Green Lake before our next stop. Our final stop (#5) will be along the path on the northwest side of Green Lake (across from Deadman's Point) which will afford us an open view of the circular, deep water basin of Green Lake (Figs. 2 and 3).

Because of the seasonal precipitation of calcium carbonate (whitings) coupled with the anoxic nature of waters below the chemocline, sediments in the deep main basin (and part of the "neck") of Green Lake are characterized by distinct annual laminations or varves (Fig. 26). Using sediment trap data Brunskill (1969) was able to document the summer accumulation of a light tan calcite laminae and the winter accumulation of a dark homogenous laminae of more organic-rich sediment; thus, proving the annual nature of these laminations (i.e.-varves). Ludlam (1969) further documented that the light colored laminae contain an average of ~82% calcium carbonate whereas dark laminae contain only ~55%. Average thickness of the more recent varves is on the order of ~0.7 mm.

Ludlam (1969) also noted that not all sediments below the chemocline are varved; there is also an abundance of massive or "unlaminated" sediment. Because some of these massive layers (mm to cm scale) are graded, have erosional bases and contain coarse littoral zone grains, they have been interpreted as turbidites (Fig. 26; Ludlam, 1969, 1974). We have also recovered deep lake sediments from Green Lake which contain coarse gastropod shells and other material which has obviously been derived and transported from much shallower depths. There is also very good photographic evidence (Fig. 27) for the deformation of varves via subaqueous slumping (Ludlam, 1974; Dean and Fouch, 1983). Thus, there is little question that mass wasting and sediment gravity flows are processes that operate in Green Lake.

However, Ludlam (1969, 1974) has interpreted all massive layers (even those ~1 mm thick) as turbidites and estimated that 40-65 % of basinal sediments in Green Lake were deposited as gravity flows. We do not agree with this assessment which evolved largely in the late 1960's and early 1970's when turbidity current theory was being developed and widely applied. Many of the massive layers we have examined display no objective evidence for deposition by turbidity currents; they are not graded, do not contain displaced grains, and do not have erosional bases. In fact our core 6 form the central basin of Green Lake contains relatively few unequivocal turbidites (Fig. 28). Although this does not mean that all the massive layers are not turbidites, it does highlight a need for additional hypotheses. An alternative interpretation is that these massive layers (particularly the thin 1-2 mm thick ones) are simply thicker than normal winter layers of annual couplets (varves). In fact, Ludlam (1981) stated that "...thin turbidites... can not be separated from the annual layers" (p. 85). At the time of this writing, we are

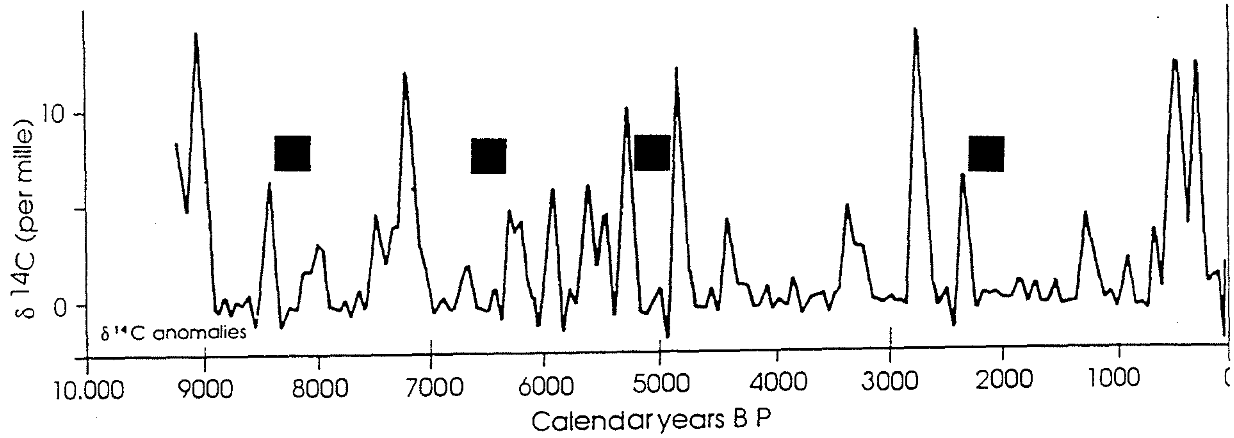
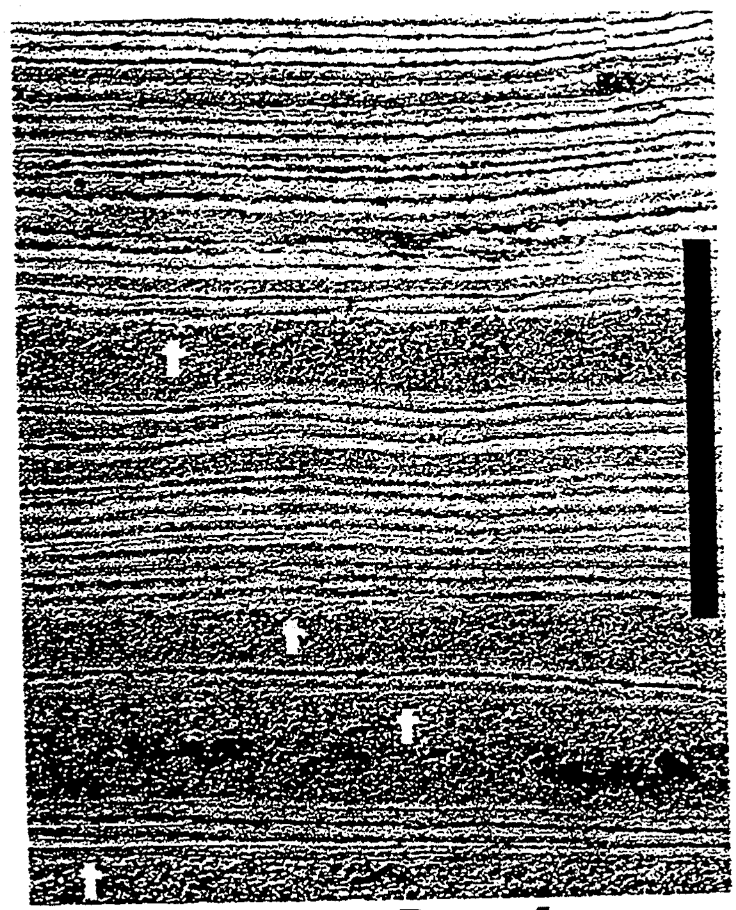


Fig. 25: Solar irradiance curve for the Holocene based on $\delta^{14}C$ anomalies. High $\delta^{14}C$ values indicate relatively low solar irradiance due to a low sunspot activity. Black squares denote ages of peat layers from core GLWL-3. Note correlation of ages of Green Lake peat layers calibrated to calendar years with low $\delta^{14}C$ (i.e.- relatively high solar irradiance) suggesting possible evaporative drawdown for lake-level lowerings. Solar irradiance curve based on Wigley and Kelley (1990): taken from Karlan and Kuylenstierna (1996).



Bar = 5 cm

Fig. 26: Photograph of dried Green Lake sediment core recovered from below the chemocline. Note mm-scale light/dark couplets which are true annual varves as well as coarser-grained massive layers interpreted as turbidites (from Dean and Fouch, 1983).

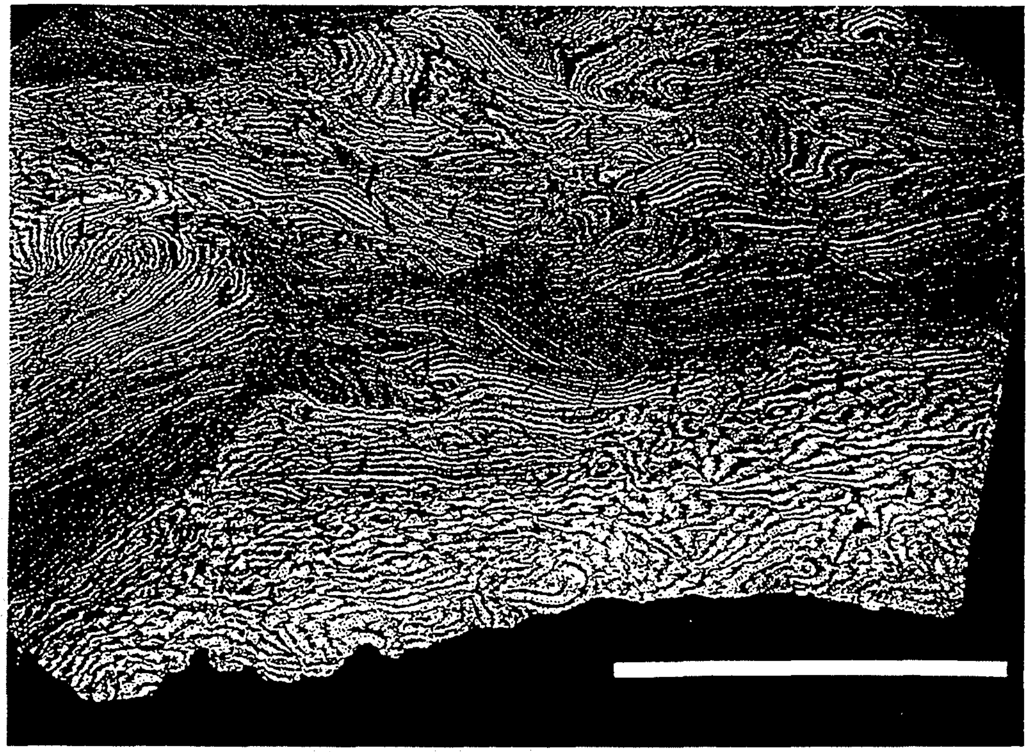


Fig. 27: Photograph of dried slab of Green Lake sediment collected from below the chemocline, illustrating contorted varves interpreted as a slump sequence. Bar scale = 5cm; from Dean and Fouch (1983).

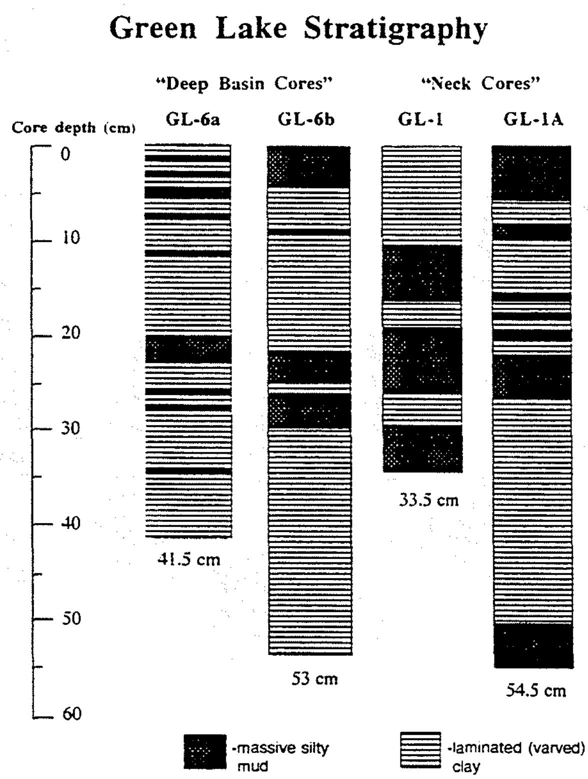


Fig. 28: Lithostratigraphy of short cores recovered from below the chemocline in Green Lake, illustrating the relative abundance of annual laminations (varves) and massive layers (turbidites). Note relative paucity of massive layers in core GL-6 from the central basin of Green Lake.

conducting compositional analyses and detailed varve versus radiocarbon chronologies in an attempt to develop objective criteria for the origin of many of these "massive layers" in Green Lake.

The accumulation rate of profundal sediment in Green Lake has also been found to be variable, both temporally and spatially. Ludlam (1981) noted a 48% increase in sedimentation rates between the late 1800's and the 1970's. Ludlam (1984) also discovered that sedimentation rates in the "neck" of Green Lake are ~27% higher than they are in the main basin, perhaps due to the reworking and focusing of sediment from shallow water.

During the summer of 1996 we collected short (<50 cm) gravity cores at seven sites in Green Lake (Fig. 20) with the objective of using the annual deposits (varves) as high-resolution recorders of relatively recent environmental change. We have focused on a pair of cores recovered at site 6 from the floor of the central basin because they contain the lowest ratio of varved sediments to gravity flow deposits (i.e.-turbidites). Because massive zones are interpreted as instantaneous events, they were subsequently subtracted from the final varved sediment record, thus yielding a total varved section of 32.5 cm. Dating of the core is based on four approaches: (1) varve counts; (2) a radiocarbon date on terrestrial organic matter; (3) geochemical correlation with a core dated by ^{210}Pb techniques (Whalen and Lewis, 1980); and, (4) recognition of a pink varve deposited in 1963 (Ludlam, 1969; Ludlam, 1984). The base of the varved section is ~2500 years B.P. (504 B.C.; Fig. 29) and is thus a record of both natural and anthropogenic processes. Sediment accumulation rates between 2500 years B.P. and 167 years B.P. (1830 A.D.) averaged ~10 cm/1000 years versus ~70 cm/1000 years over the past 167 years.

Loss-on-ignition results for total calcium carbonate content in core 6 display a marked decrease at the same depth (18.5 cm) where accumulation rates increase by nearly an order of magnitude (Fig. 30). Average total carbonate values drop from ~80% prior to 1830 A.D. to ~60-65% after 1830 A.D. (Fig. 30). In contrast, average total organic matter contents in core 6 remain relatively constant at ~10% both prior to, and after, 1830 A.D. These data indicate that starting in ~1830 A.D., the flux of terrigenous material to Green Lake increased significantly resulting in increased rates of sediment accumulation (Fig. 30). This is supported by mineralogical data (determined by X-ray diffraction) which show an abrupt increase in the relative abundance of detrital quartz and dolomite at ~1830 A.D. (Fig. 31). Prior to 1830 A.D. dolomite is generally absent from core 6 and quartz is only a minor component. Trace element data (determined by direct current plasma emission spectrometry) of the carbonate fraction in core 6 also display a marked change at ~18.5 cm depth. Concentrations of Pb, Mg, Mn, Cu, and Fe all increase above this depth, whereas Ca and Sr decrease. Lead (a common anthropogenic element) is of particular interest (Fig. 32). Prior to 1830 A.D. lead concentrations range from 5 to 55 ppm with an average ~25 ppm; whereas, after 1830 A.D. Pb concentrations range from 35 to 105 ppm with an average of ~70 ppm (Fig. 32).

Collectively, these results argue for a significant anthropogenic influence on sedimentation in Green Lake since ~1830 A.D. Whalen and Lewis (1980) have previously proposed that deforestation of the region in the early to middle 1800's increased surficial runoff and the flux of terrigenous material into Green Lake. Our results are consistent with their suggestion. Pollution of surface waters by heavy metals at this time might also explain the observed increases of Pb, Cu and Fe since 1830 A.D., although these metals may have also entered the Green Lake basin via the atmosphere as a by-product of coal burning starting with the industrial revolution.

We have also sampled (N=48) varves from core 6 for stable isotope analysis in order to evaluate both natural and potential anthropogenically-influenced climate changes over the past 2500 years. These new data represent the first stratigraphic stable isotope results for Green Lake since a brief mention of results by Stuiver (1968, 1970). $\delta^{18}O$ (a measure of the $^{18}O/^{16}O$ ratio) values vary by as much as ~2 ‰ over the length of the core (Fig. 33) which is 1 ‰ greater than the variability indicated by Stuiver (1968). $\delta^{18}O$ values also display a gradual depletion from -10.2 ‰ to -10.7 ‰ between 500 B.C. and 200 A.D. If this entire depletion (-0.5 ‰) is due solely to temperature change (0.34 ‰/1°C; Rozanski et al., 1993; Anderson et al., 1997) it would imply regional spring/summer cooling of up to 1.5 °C.[2] However, some or all of this isotopic change could also be due to shifts in atmospheric moisture sources. In this case more depleted $\delta^{18}O$ values would imply an increase of moisture from Pacific, Canadian or North Atlantic air masses, which all tend to be cooler than the warm, isotopically heavy air masses which advect into central New York State from the Gulf of Mexico (Rozanski et al., 1993).

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[2] In terrestrial records, such as ice cores or lake sediments, $\delta^{18}O$ depletions imply relative coolings due to latitudinal fractionation of ^{18}O. This general relationship is opposite that for marine records.

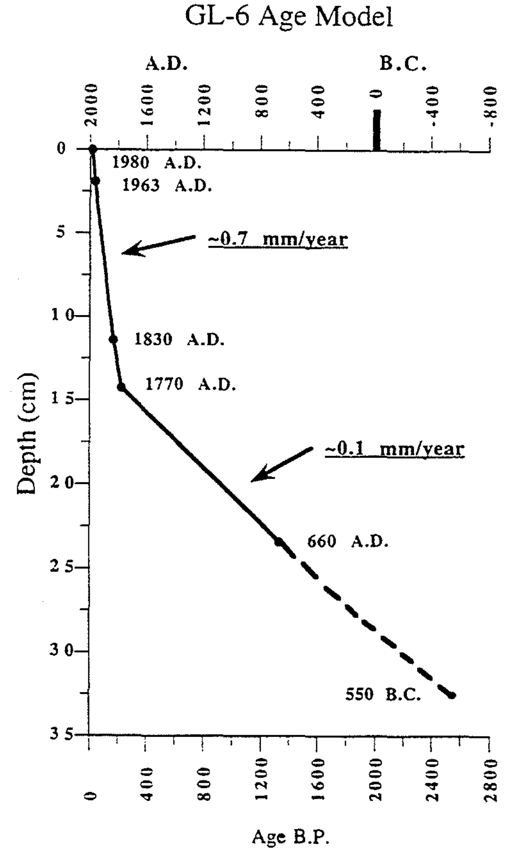

Fig. 29: Age versus depth curve for Green Lake core GL-6. Surface age of 1980 A.D. based on the recogntion of a distinctly colored 1963 varve (Ludlam, 1969, 1984) as well as replicate varve counts. Date of 1830 A.D. based on geochemical correlation with a ^{210}Pb-dated core and the 660 A.D. date is based on AMS radiocarbon results of a cedar leaf fragment calibrated to calendar years. Basal age of ~2500 years B.P. (504 B.C.) is based on linear extrapolation. Note large increase in sediment accumulation rates within the past 150-200 years.

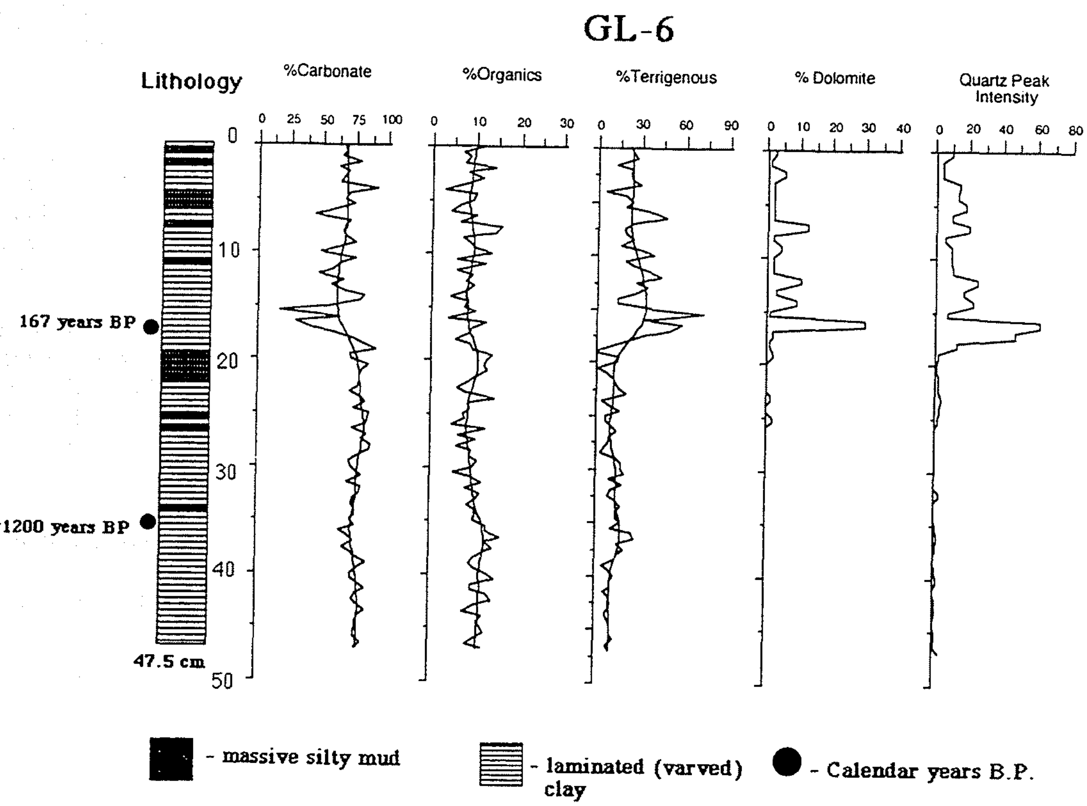

Fig. 30: Loss-on-ignition data from Green Lake core GL-6. Note high calcium carbonate contents (~60-90%) and relatively low total organic matter (TOM) values (~10%). Also note decrease in total carbonate content at 18-19 cm (~167 years B.P.; 1830 A.D.).

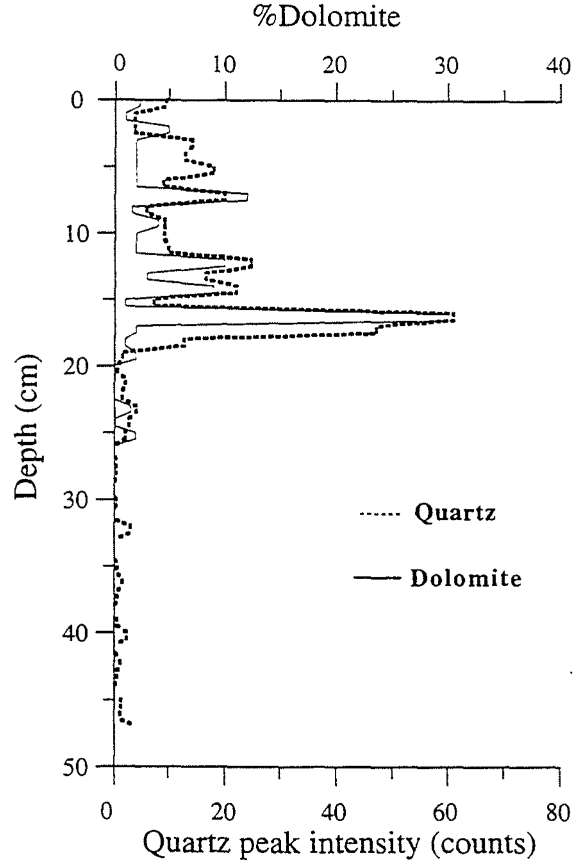

Fig. 31: Relative quartz peak intensity (based on XRD results) and percent dolomite (in the carbonate fraction) versus depth in Green Lake core GL-6. Note abrupt increase of both detrital quartz and dolomite at 18-19 cm (~1830 A.D.) coincident with increase in sediment accumulation rate.

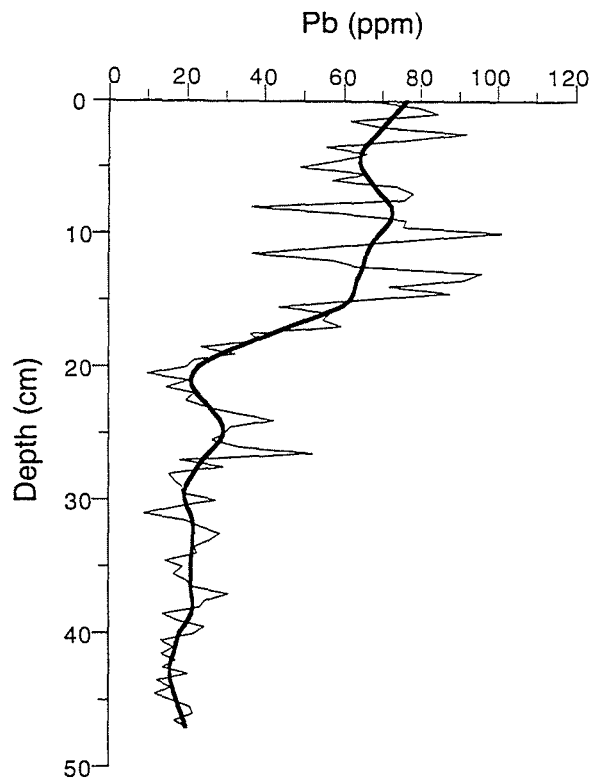

Fig. 32: Concentration of lead in the calcium carbonate fraction of Green Lake core GL-6. Note abrupt increase of Pb at 18-19 cm (~1830 A.D.) coincident with increase in sediment accumulation rates.

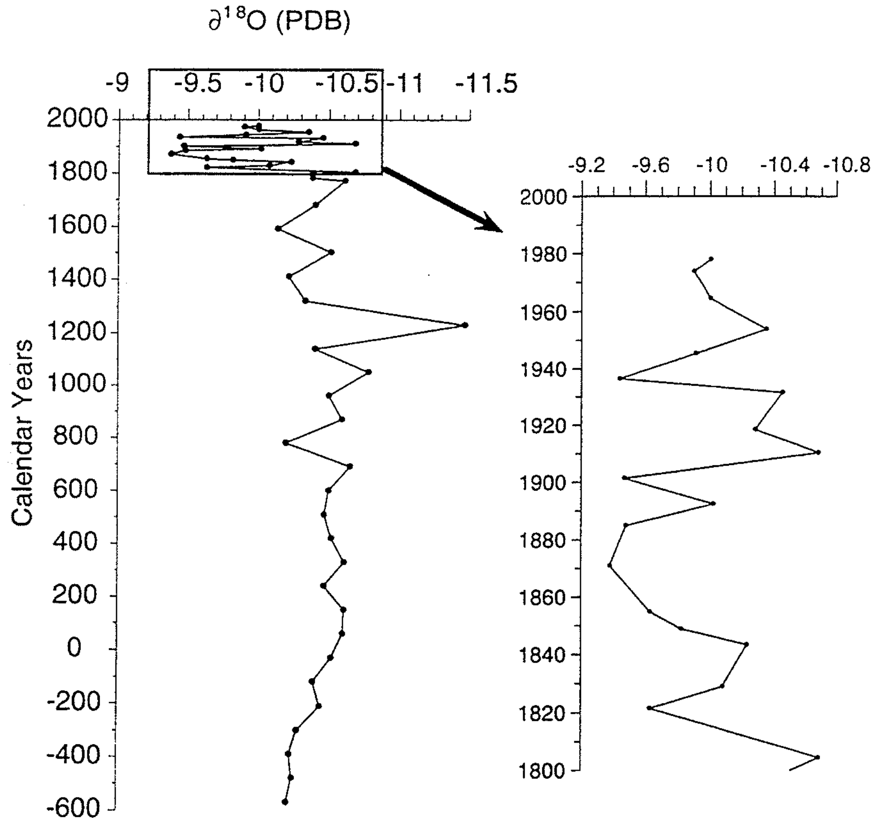

Fig. 33: Distribution of $\delta^{18}O$ values of fine-grained calcite (< μ) from Green Lake core GL-6. See text for additional discussion.

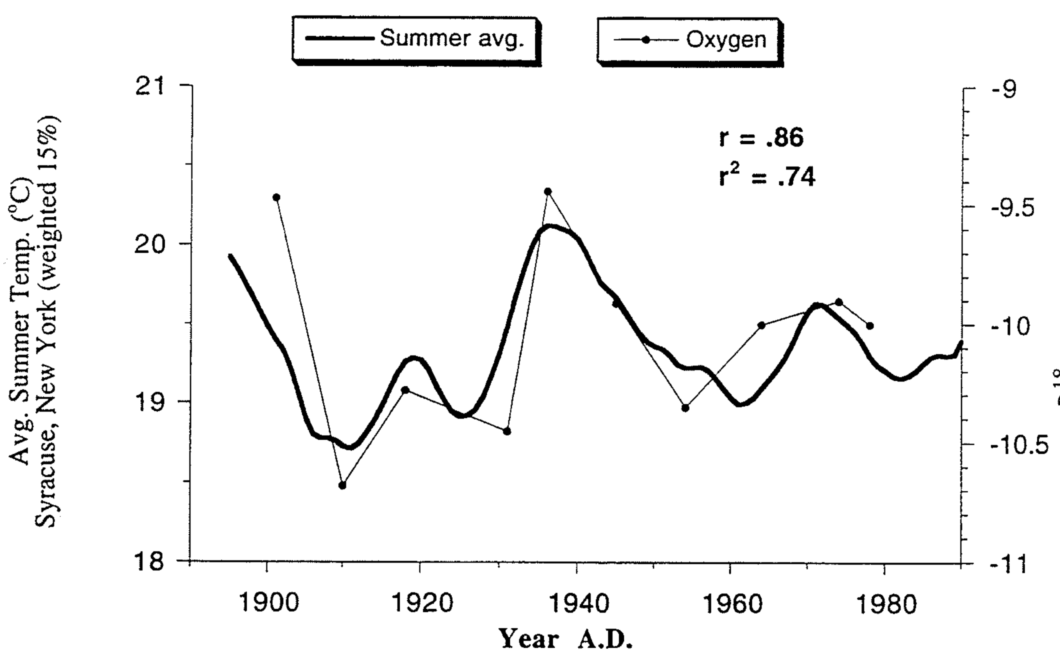

Fig. 34: Correlation of Green Lake core GL-6 $\delta^{18}C$ values over the last century with yearly average summer temperatures for Syracuse, N.Y. (smoothed 15%). The strong correlation between decadel trends in temperature and $\delta^{18}O$ (r^2=.74) is thought to be a function of the type of airmass advecting into the area. See text for additional discussion.

This isotopic evidence for a modest cooling trend between 500 B.C.. and 700 A.D. is consistent with ice core data sets which indicate global cooling during the past 4-5 ka following the mid-Holocene Hypsithermal (Larsen et al., 1995; Thompson et al., 1995). Both glacier reconstructions (Nesje and Kvamme, 1991) and general circulation models (Liao et al., 1994) indicate that Hypsithermal summer temperatures were 2-3 °C warmer than pre-industrial modern values, due to precessional changes.

Following an extreme $\delta^{18}O$ depletion event (one data point) of ~1 ‰ at ~1250 A.D. $\delta^{18}O$ values remain relatively constant until the early to middle 1800's and "colder" than modern values. Since then, $\delta^{18}O$ values become enriched (heavier) by as much as 1 ‰, although there is significant variability (Fig. 33). The increased variability is due in part to a lower sampling interval afforded by the higher accumulation rates after 1830 A.D. Whereas the pre-1830 A.D. average sampling interval is 1 sample per 90 years, the interval increases to about 1 sample every 14 years after 1830 A.D. allowing resolution of decadel scale variability.

It is tempting to speculate that the enrichment (warming) of $\delta^{18}O$ values from Green Lake after the early 1800's is related to the anthropogenic rise of atmospheric carbon dioxide and global warming since the early to middle 1800's (Keeling et al., 1989). Absolute $\delta^{18}O$ values over the past ~150 years are as heavy as -9.3 ‰ which is unprecedented over the 2500 year record in core 6 (Fig. 33). However, a 1 ‰ increase in $\delta^{18}O$ would indicate an ~3 °C summer temperature rise, assuming that the entire isotopic shift is temperature dependent. Mean global surface temperatures have risen by only ~0.5 °C since the mid-1800's. However, part of this observed increase in $\delta^{18}O$ values from Green Lake over the past ~150 years could also be due to an increase of moisture from warm, isotopically heavy Gulf of Mexico air masses, and/or evaporative enrichment.

To elucidate the potential causes of enriched and depleted $\delta^{18}O$ values in Green Lake throughout the last century, a calibration using regional climate data was performed. Figure 34 shows the graphical comparison of $\delta^{18}O$ values from ~1900 to 1980 A.D. with yearly average summer temperatures for Syracuse, N.Y. When the temperature data is smoothed (a 15% running average function was utilized to remove interannual variability) decadel trends are evident that display a remarkable correlation (r= .86) with $\delta^{18}O$ values (Fig. 34). Periods of cooler temperatures (1910 to 1930 and 1950 to 1970) correlate with isotopically depleted values while warmer periods (1900's and the late 1930's to early 1940's) correlate with isotopically enriched values. However, if the $\delta^{18}O$ values were solely recording surface water temperatures in Green Lake during these periods, the exact opposite relationship would be observed (i.e. cool periods represented by isotopically enriched values). Therefore, the dominant control on the $\delta^{18}O$ carbonate signature must be a function of the water composition in Green Lake, which is largely controlled by the type of airmass bringing precipitation into the area. As stated above, enriched values could be due to an increase of moisture from warmer temperature, isotopically heavy Gulf of Mexico source air masses while depleted values may result from cooler temperature, isotopically depleted Canadian or Pacific source airmasses. Because Green Lake lies directly below the modern day position of the jet-stream, and the position of the jet-stream controls the type of airmass coming into the region, the Green Lake $\delta^{18}O$ record provides a very sensitive proxy for jet-stream fluctuations and climate change.

Final Comments

From Stop #5 on the northwest shore of Green Lake we will proceed along the path and return to our starting point at the Park's boathouse. Here we can pause for any final comments, discussion or suggestions anyone would like to make regarding the geology, limnology and paleoclimatology of Green Lakes State Park.

Overall we hope you found this field trip educational as well as enjoyable, and that you have gained a better appreciation of the value of Green Lakes State Park as a natural laboratory for the study of fundamental natural processes such as deglaciation, meromixis and whitings as well as the enormous potential of these lakes as archives of a high-resolution record of natural climate variability over the past 10,000 years. Thanks for joining us.

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