Climate, Tectonics or ...?: Speculations on the Recent Paleolimnology of Yellowstone Lake

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Abstract

The sediments of Yellowstone Lake may reveal the paleoecological history of this lake over the last few centuries. These sediments contain up to 60% biogenic silica derived from diatom frustules settling out from the overlying water. The sediment record reveals large variations in the diatom deposition over the last ~350 years. Some of these variations appear to correlate extremely well with independent climate records, particularly mean annual winter temperature and precipitation, derived from tree ring data extrapolations. A strong correlation occurs, for example, during an extended period of below-normal winter temperatures and above-normal precipitation seen during the late 1800s. Below-normal winter temperatures can significantly extend winter ice cover and shorten the ice-free, isothermal period during which the spring diatom bloom occurs. Rapid thermal stratification following prolonged ice cover may reduce annual diatom production and the subsequent silica deposition. Yet the forcing factors in this system may not be so straightforward. Sublacustrine hydrothermal springs found in the lake are a potential source of nutrients that may vary in strength and in time. These inputs may have the potential to alter the nutrient biogeochemistry of the lake. A pronounced chloride enrichment observed within the lake may be explained by input of a source of undiluted geothermal water equivalent to >0.06% of the riverine input. Calculations based upon silica removal indicate that such hydrothermal inputs would have to contribute up to ~10^7 moles of nitrogen annually, for example, in order to significantly shift the nutrient status of the lake. Observations to date, which are limited in space and time, do not appear to support such a flux for the whole lake. However, the existence of high-activity vents in past eras (as indicated by numerous relict features) or in unexplored regions of the lake cannot be ruled out.

Introduction

The integrity and value of the paleoecological history recorded in lake sediments is dependent upon a number of factors. For example, a good record contains: a relatively undisturbed sediment accumulation rate regime, a coherent and quantifiable sediment chronology, and minimal or quantifiable post-depositional alteration of biological or geochemical indicators as a consequence of diagenesis, mixing, migration, or other physical disturbances. The time scale of interest can range from years to millennia and is dependent upon the age of the lake and the existence of geochronological techniques to independently date various sediment horizons within the sediment column. Sediments within the depositional basins of Yellowstone Lake consist largely of a diatom ooze, up to 60% biogenic silica by weight, and have been examined in a number of studies for their potential in revealing the ecological history of this high-altitude system. Shero and Parker (1976) examined sediments from the South and Southeast arms of the lake and identified over 150 taxa of diatoms in cores with maximum ages estimated to be on the order of 1500 BP (years before the present). Many of these taxa were extremely rare, but the diversity of the flora indicates a potentially rich record of ecological changes and evolution over this period. In general, Shero and Parker (1976) observed a decrease in diatom abundance over the last 1500 years and hypothesized that lake productivity may have decreased over that period as a consequence of decreases in the annual nutrient supply, perhaps related to decreases in annual precipitation. More recently, Kilham et al. (1996) have provided an excellent review of the factors linking diatoms and climate change in the large lakes of the Yellowstone ecosystem.

The observations reported here are the result of work begun by our group some time ago (1983) in looking at the recent record (i.e., over the last 200 years) in these sediments, and in the sublacustrine hydrogeothermal activity within the Yellowstone Lake basin (Klump et al 1988).

Analytical Methods

Sediment cores were collected using a standard 7.5-cm-diameter Benthos gravity corer deployed from the U.S. Fish and Wildlife Service *R/V Cutthroat.* Intact cores in excess of 60 cm in length were retrieved. Upon returning to shore, cores were sectioned on a hydraulic extruder at 0.5- to 5-cm intervals. Sediment sections were placed in tared plastic 125-ml bottles, dried in an oven at 60˚C to a constant weight, and reweighed to determine the percent water content and porosity, assuming a dry sediment density of 2.3 g cm^{-3}. Sediments were pulverized in a mortar and pestle to a fine powder. Pb-210 activities were determined following a modification of the procedure of Robbins and Edgington (1975). An internal Po-208 standard was added to ~0.5 g of sediment to determine recovery efficiency, and the sediments were digested in 6N HCl at 95°C with sequential additions of 30% hydrogen peroxide. The solutions were cooled, filtered, pH-adjusted to 0.5 to 1.0, and amended with 100 mg of ascorbic acid. Po-210 and Po-208 were plated unto a polished copper disk in a boiling water bath and counted via low-level alpha spectrometry.

Pigments (chlorophyll and total carotenoids) were measured according to the spectrophotometric technique given by Strickland and Parsons (1972). Aliquots of whole, wet sediments were extracted in 90% acetone at approximately 30 ml per gram dry sediment for >20 hours in the dark, centrifuged, and the supernatant decanted into a 5-cm-path-length spectrophotometer cell. Values for carotenoids are reported as relative concentrations for time-series analysis and are roughly equivalent to ug g^{-1}dry sed. Daphnia winter-resting eggs or ephippia were counted in known wet-sediment aliquots under a dissecting microscope and are reported as number per gram dry sediment. Ephippia were easily recognized and counted, being the only large particles in these otherwise very fine-grained sediments. Biogenic silica was determined following the differential dissolution technique of DeMaster (1981) in which sediments are dissolved in a 85°C, 1% solution of Na_2CO_3. Sequential samples of the solution are analyzed for dissolved silica (Strickland and Parsons 1972) over a 5-hour period and the initial rapid rise in silica concentration taken as the dissolution of diatom frustules. Scanning electron microscopy micrographs made of both dissolved and untreated samples confirmed complete dissolution.

Results and Discussion

The cores examined here were collected from the deepest portion of the central basin of West Thumb in 1983 and 1985. The water depth here is ~310 feet and was, at the time, considered to be the deepest sounding in the lake. (Subsequently, more precise bathymetry with higher spatial resolution and remotely operated vehicle explorations of the bottom shifted the known deepest location to a small "canyon" southeast of Stevenson Island with soundings of nearly 400 feet.) Visually, the cores appeared to be relatively undisturbed, with a surface "flocculent layer" of a few millimeters. Below this surface floc, sediments were consolidated, highly porous (90% at 50 cm), fine-grained muds. X-radiographs revealed some apparent laminations, although not distinct or regular. In casual observations at the time of collection, benthic macroinvertebrate infauna were not observed and biogenic mixing is assumed to be minimal. The Pb-210 geochronology at this location for these cores (see Figure 1 for 1985) appears to bear this observation out. Excess Pb-210 (half-life 22.3 years) decreases exponentially from a value of ~23 pCi g^{-1} at the sediment–water interface to a supported value of <0.3 pCi g^{-1} at a depth of ~18 cm. Calculations from a curve fit of the excess Pb-210 data to the mass sediment accumulated with depth (g cm^{-2}) yield a net average mass sediment accumulation rate over this interval of ~22 mg cm^{-2} y^{-1} ($r^2 = 0.97$), or a linear sedimentation rate of ~0.16 cm y^{-1}. In general, Pb-210 dating may be extended to ~5 half-lives, or 100–120 BP. In this analysis we have extrapolated this rate to the length of core for chronological purposes, but add the caveat that dates prior to about 1870 are simple extrapolations and that these dates become increasingly sensitive to relatively small changes in sedimentation rate with increasing age. Indeed, one of the most interesting features in this core dates near the lower end for resolution by Pb-210 (1860–1900). This extrapolation, however, does fall within the range of sedimentation rates calculated by Shuey et al (1977) using paleomagnetic data of 0.100 to 0.213 cm y^{-1} (excluding the South Arm) for the period 1175 AD to present.

Diatom remains (intact frustules and fragments) make up the major portion of the sediment mass in the depositional basin of West Thumb. The biogenic silica content of these sediments reaches nearly 60% by weight at depth, but ranges from ~45% to 58% over the last 200 years (Figure 2). These changes in the biogenic silica content argue for significant changes in the production, burial, or

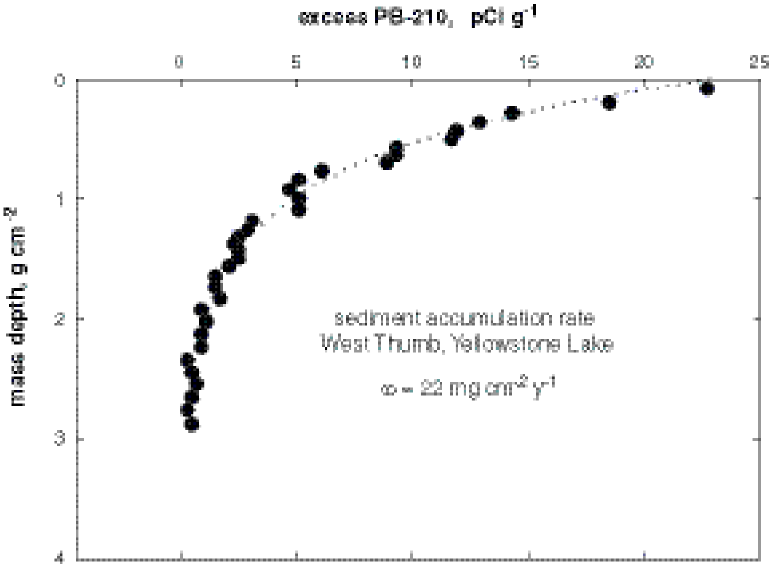

Figure 1. The Pb-210 geochronology for the West Thumb core (WT-85) shows a remarkably constant mass sedimentation rate averaging ~ 22 mg cm^{-2} y^{-1}.

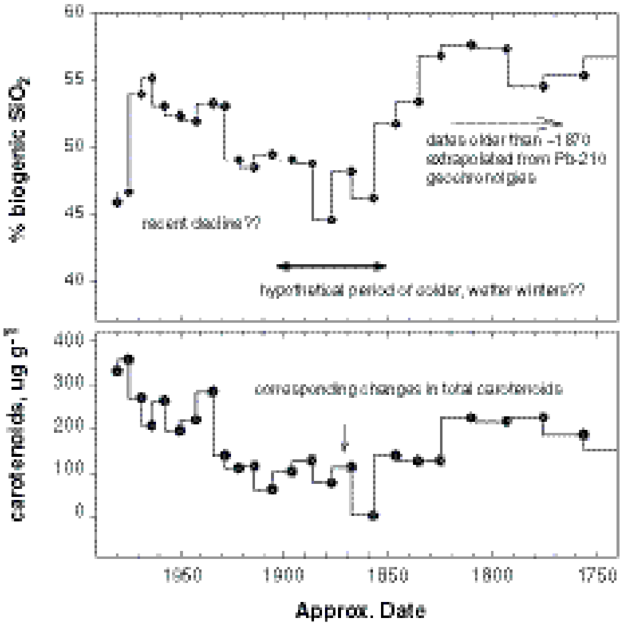

Figure 2. Percent biogenic silica (dry weight) and total carotenoids plotted as a function of time of deposition in WT-83. Both show significant drops in the late 1800s.

preservation of diatoms over time. Nothing within these cores indicates an alteration in the preservation of diatoms, although significant changes in speciation to species with more fragile, readily dissolved frustules is possible. Sedimentation rates, on the other hand, are remarkably constant. Our hypothesis is that the changes observed in the biogenic silica record are the result of changes in diatom production related to annual variations in whole lake productivity.

Of particular interest is the dramatic decrease in biogenic silica production (i.e., burial) in the late 1800s. At steady state, this decrease would translate to a drop in diatom production of 10–20%, depending on the reference period. The core used for this analysis was collected in 1983 and was sectioned at 1-cm intervals to a depth of 20 cm, 2-cm intervals to a depth of 30 cm, and then at 5-cm intervals to the bottom of the core. The lowest point in the biogenic silica stratigraphy (44.6%) occurs at 14–15 cm, an interval for which we place a date of ca. 1877. Quantification of simple algal pigments preserved in this record also shows a strong correlation with the diatom record, and carotenoids track biogenic silica content extremely well (Figure 2).

A principal goal of paleolimnology is, of course, to use such biotic tracers to decipher past conditions in the lake, in an attempt to determine how planktonic communities and the ecology of the system have responded to changes in climate, ecosystem structure, evolutionary pressure, and both naturally occurring (e.g., forest fires) and anthropogenic (e.g., watershed development) processes (e.g., Meyer et al. 1992; Kilham et al. 1996). In Yellowstone Lake, all of these types of processes are potential contributors to changing lake ecology.

Climate Changes

Temperate lakes, and perhaps high-altitude lakes in particular, are especially susceptible to changes in climate. One of the principal reasons for this is the annual physical cycle of most temperate lakes, which is driven by the annual temperature oscillation. A high-altitude lake in one of the coldest regions of the U.S., Yellowstone is ice covered for nearly six months of the year. Inter-annual changes in the temperature climate can vary the temporal extent of ice cover and of stratified and unstratified periods by several weeks or even longer. The ice free season begins with an isothermal, well mixed water column in the spring. As solar heating increases the lake shifts to a thermally stratified, stable water column in the summer, followed by overturn and mixing upon cooling again in the fall.

Determination of a climate signal in lake sediments is confounded by the variety of potential forcing functions. Correlations with other climate records, however, may be useful. Using the analysis of tree ring data, Douglas and Stockton (1975) reconstructed a long term seasonal temperature and precipitation record for the Yellowstone National Park region. This reconstruction dates back to 1750 with both seasonal and annual coverage. Of interest here is their reconstruction for winter temperatures and precipitation (Figure 3). A simple examination of this record shows what appears to be an anomalous period during the late 1800s. Prior to about 1860 and after about 1905, this record shows predicted winter tem-

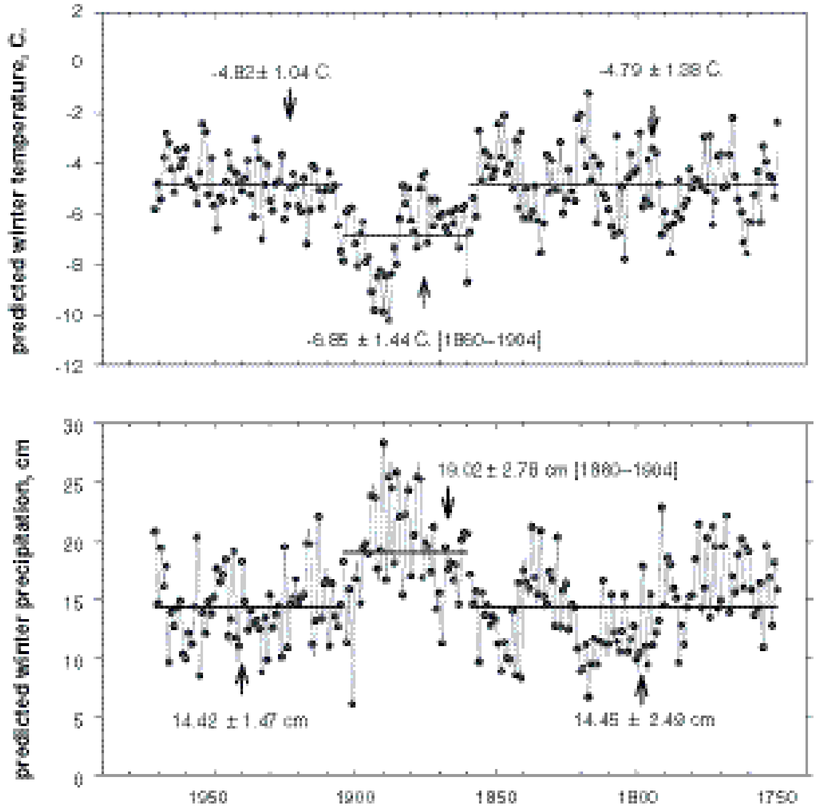

Figure 3. Paleoclimate predictions for winter (November-March) temperatures and precipitation in Yellowstone based upon tree ring climate reconstructions (from Douglas and Stockton 1974). Both below-average temperatures and above-average precipitation are predicted for the late 1800s, particularly during the period 1885–1894 when temperatures were nearly 4˚C lower and precipitation 8 cm higher than "normal" conditions prior to 1860 and after 1905.

peratures to vary about a remarkably constant mean value of ~4.8° ± 1.4°C. Predicted winter precipitation likewise varies about a relatively constant mean of ~14.4 ± 2.5 cm for these two long periods. During the late 1800s, however, there is a significant drop in predicted temperatures by at least 2°C, to a mean of –6.85° ± 1.4°C, concomitant with a significant increase in predicted winter precipitation by at least 4.5 cm. to 19.0 ± 2.8 cm. Assuming this is snow cover, this would be equivalent to ~45 cm of additional snowfall. Both of these would appear to be, in this region, significant climate variations. In fact, a closer examination of the data for 1860–1905 shows that for shorter periods of 10 to 20 years, the departures from the long-term means are even greater. For example, for the period 1885–1894, the average predicted winter temperature is –8.81° ± 0.89°C, 4°C below "normal," and the average predicted winter precipitation is 22.34 ± 3.88 cm, nearly 8 cm above "normal." The combined effect of colder-than-normal temperatures and above-average snowfall could easily extend the ice-cover period on Yellowstone Lake by weeks. Typically, ice-out occurs in late May or early June. Prolonging ice cover, even by two or three weeks, could have dramatic effects on lake ecology. Results from regional climate models for the impact CO_2 doubling on the thermal regime of Yellowstone Lake are indicative of the sensitivity of the lake to climatic-scale temperature changes. For example, the average annual surface temperature is increased by 1.6˚C for a $2xCO_2$ scenario (Hostetler and Giorgi 1995). This warming reduces the annual duration of ice cover by over six weeks, from 196 days to 152 days.

Primary production during the spring bloom is particularly important in deep lakes such as Yellowstone. During the spring transitional period, the lake is isothermal and well mixed. Because of mixing, algae throughout the water column may be exposed to light and have sufficient nutrients to sustain rapid growth. Once the lake warms, however, and begins to stratify, hypolimnetic nutrients are largely out of the reach of the photic zone and photosynthetic primary production is limited to the fairly shallow region of the epilimnion where nutrients, no longer being replenished from deeper waters, can be rapidly depleted. The spring bloom is triggered, in general, by light. Prolonged ice cover may have multiple effects. First, it insulates the water column from solar radiation, limiting algal growth, and secondly, it contracts the length of the isothermal spring bloom period. In the latter case, by the time ice-out occurs, solar heat inputs may be near their maximum and the lake can stratify very quickly, perhaps in a matter of days. The result: there is little time to extract stored hypolimnetic nutrients before they are “sealed off” by the rapidly forming thermocline, and the productivity of the spring bloom is significantly limited. Similar climatic-forcing effects have been observed as a consequence of El Niño events in Castle Lake, a temperate, subalpine lake in California. Year-to-year changes in the amount of snowfall from February through April, which determine the date of ice thawing (by up to more than one month later in the spring), coupled with early heating and stratification, resulted in significant interannual variations in heat stored within the lake (Strub et al. 1985). This ranged from early thaws with extended mixing and high productivity, to late thaws with incomplete mixing, a failure to renew photic zone nutrients, and consequent low productivity. During 1983, for example, when the lake remained ice covered until 6 July, primary production during the summer was only 25% of normal. In a similar situation, interannual variations in zooplankton abundance (principally the herbivorous *Daphnia hyaline)* in Lake Windermere, United Kingdom, strongly correlated with the timing of thermal stratification (George and Harris 1985). Interestingly, zooplankton biomass was higher in cool years coinciding with the period of maximum food availability, whereas in warm years the preferred algal food species tended to appear earlier and may have been in decline by the time Daphnia begin to reproduce.

The paleo-record of zooplankton is much less robust in lakes. Cladocerans, however, are a major component of the zooplankton in Yellowstone Lake and the sediments contain abundant ephippia, or winter-resting eggs. Recently, there is renewed interest in paleoecological studies using these resting eggs, including genetic and evolutionary histories (Hairston 1996). Here, however, we report only our observations on numbers deposited and preserved through time. Although this record is highly variable (Figure 4), it does appear that prior to about 1900 there are episodes of high and low ephippia production. One of the highest of these correlates to a date in the 1870s and 1880s, a period of low diatom production in the lake and cold, wetter winter weather. Three of the eight intervals in which numbers exceed 300 per g occur within this period. Coincidence, perhaps, but this observation may relate to spring bloom timing effects, low primary production, or changes in speciation. Daphnia stressed by low food resources, for example, may shunt more energy into egg production, resulting in greater ephippia abundance in years in which primary production is decreased. The clearest trend in the ephippia record, however, is the decline in numbers after about 1910. Speculations for the reason for this drop could include all of the above, as well changes in ecosystem dynamics, such as abundance shifts in planktivorous fish populations, an impact that can be triggered by the invasion of non-native species.

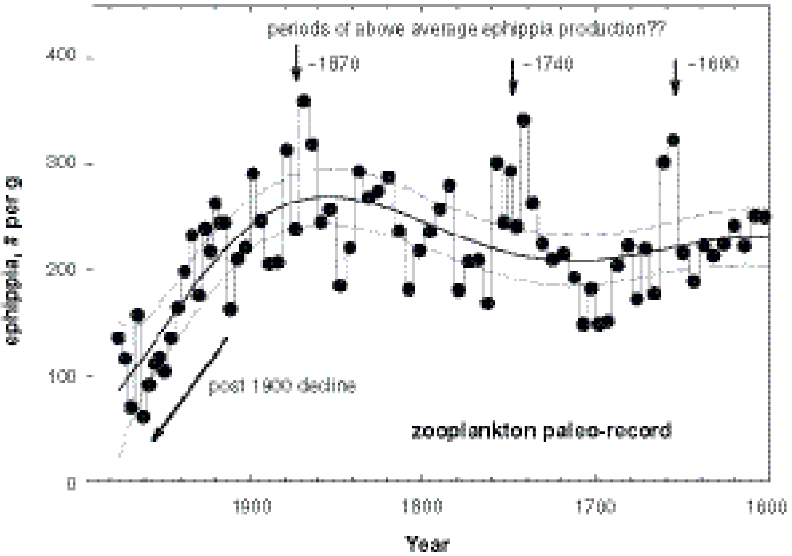

Figure 4. The abundance of ephippia (number per gram dry weight) in West Thumb sediments (WT-85) as a function of the year of deposition.

Geochemical Budgets and Geothermal Inputs

Its location within the Yellowstone caldera and over a geothermal hot spot makes Yellowstone Lake unique. Studies over the last 15 years have revealed a sublacustrine plumbing system made up of diverse underwater hot springs, fumaroles, and seeps (Klump et al. 1988). Could the geothermally enriched fluids emanating from these features have the potential to "fertilize" the lake to a degree sufficient to alter primary production? How much would be required? One approach is to simply calculate backwards from the known diatom accumulation rates. For example, if we assume that significant deposition only occurs at depths greater than 40 m, then slightly more than 50% of the area of the lake is depositional. Further, if we assume that the sediment accumulation rate averaged over this area is ~10 mg cm^{-2} y^{-1} (or roughly one-half of the 22 mg cm^{-2} y^{-1} measured in West Thumb), then the average biogenic silica removal rate (at 50% biogenic silica by weight in the sediment) for the whole lake is on the order of 2.5 mg (42 umol) Si cm^{-2} y^{-1}. The average depth of the lake is 42 m; hence, the average silica removal rate from lake water via burial is ~10 umol L^{-1} y^{-1}. Silica concentrations in Yellowstone Lake average over 150 umol L^{-1}. Hence, this rate of removal would hardly be detectable in lake water. Even if silica deposition were more widespread, e.g., at depths >20 m (i.e., 75% of the lake floor), and average sedimentation rates more rapid, e.g., 20 mg cm^{-2} y^{-1}, then annual silica depletion would rise to 30 umol L^{-1} y^{-1}. Although the data are infrequent and variable, this number approaches the depletion we have observed between inflow and outflow concentrations in the Yellowstone River.

Since the volume of the lake is ~1.4 x 10^{10} m^3, a 20% change in diatom production (i.e., in burial) would likewise require a mass of diatom silica of ~3 x 10^7 mol Si y^{-1}. Since silica is not limiting in this system, changes in the flux of another micronutrient would have to be responsible for any hypothesized "fertilization" effect, i.e., changes in new nutrient inputs over time. Take, for example, nitrogen or phosphorus. Average stoichiometries for diatom production vary, but an approximate Si:N ratio of ~1–2 and Si:P ratio of ~20–25 are reasonable (Redfield ratios are 16:16:1). To produce 3 x 10^7 mol Si y^{-1} would therefore require roughly 10^7 moles of nitrogen and 10^6 moles of phosphorus. To date the *highest* dissolved inorganic nitrogen and phosphorus concentrations we have measured in vent fluids are ~100 and 30 umol L^{-1}, respectively, measured in 1987–1989. Since that time the concentrations we have observed in vent fluids have been considerably more dilute. Thermal ponds on shore often have extremely high concentrations of inorganic nutrients. In Mary Bay, for example, we have measured dissolved ammonium concentrations in small ponds in excess of 600 umol L^{-1}. Assuming vent waters contain inorganic nutrients at the high end of our measurements in the lake, the hydrothermal flux required to effect a 20% diatom productivity shift from changes in nutrient supply is on the order of 0.3-1 x 10^8 m^3 y^{-1} or ~2%–7% of the riverine inflow. Silicon itself is a major constituent in hydrothermal vent waters, being readily leached from volcanic rocks rich in silicon at high temperatures. We have measured dissolved silica concentrations as high as 3 mmol L^{-1} in waters emanating from shallow vents in Sedge Bay (Remsen et al. 1990).

Chloride is a conservative element highly enriched in geothermal waters, and its flux into surface waters has been used as an indication of fluctuations in regional geothermal activity (Norton and Freidman 1985). Comparison of average chloride concentrations in the major inflow (<10 umol L^{-1}) and outflow

(~150 umol L^{-1}) shows that chloride is enriched by up to fifteenfold within the lake. Chloride concentrations in undiluted geothermal source waters in the park have been estimated to be as high as ~10–20 mmol L^{-1} (Truesdell et al. 1977; Fournier 1979). The highest we have measured in sublacustrine vents in the lake is ~3 mmol L^{-1} (Klump et al. 1992). At these levels it would require the addition of ~10^7 m^3 of vent water annually (~1% of river inflow) in order to raise the concentration in the Yellowstone River outflow by 140 umol L^{-1} (ignoring inputs from precipitation, which are assumed to be minor based upon low Cl^- levels in rainfall). If this flow were concentrated in 0.0001% of the lake bottom (i.e., 1,000 cm^2 of vents per 10 ha, which implies 3,500 such vents fields) the flow in these vents fields would need to average ~5 L min^{-1}. Our observations to date would seem to indicate that vents of this magnitude are not this numerous, but the task of accurately characterizing and quantifying activity at an areal frequency of only 1:100,000 is problematic.

The fact remains, however, that lake water is enriched in Cl^-, requiring a contribution of 1.7 x 10^7 mol Cl^- annually. This is equivalent to a ~0.06% contribution to the hydrologic budget from undiluted geothermal source waters (at 20 mmol L^{-1}). A variety of additional sources may be considered, e.g., surface runoff from contiguous geothermal areas in West Thumb, Mary Bay, Sedge Bay, Turbid Lake/Sedge Creek, and other areas; diffusion from geothermally enriched porewaters (see Aguilar et al., this volume); and wind-blown minerals or other dry deposition processes. Norton and Friedman (1985) estimate that 93% of the total chloride flux out of the park derives from hydrothermal sources, with the remainder divided among atmospheric inputs (2.7%), rock weathering (4%), and human contributions (0.2%).

It is apparent that geothermal activity varies over time. Whether this variability is sufficient to drive productivity shifts within Yellowstone Lake is still an open question, but our current observations, at least with respect to conventional nutrients, would seem to indicate that it is not. However, we still have explored only a very small fraction of the lake bottom. High-precision bathymetric charts being produced by the U.S. Geological Survey may help us answer these questions by assisting in pinpointing potentially active regions of the lake floor for further exploration. Furthermore, numerous sublacustrine vent field concretions, relict vent plumbing, and the meter-scale spires discovered in Bridge Bay (see Cuhel, Aguilar et al. this volume) are composed almost entirely (95%) of amorphous silica, indicating that very active, high-concentration vents have been common in the lake in the past. Hence, the potential for significant geothermally active episodes in the lake's history appears to be real.

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