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Chapter 4

Paleolimnological Approaches for Inferring Past Climate Change in the Maya Region: Recent Advances and Methodological Limitations

Mark Brenner
David A. Hodell
Jason H. Curtis
Michael F. Rosenmeier
Flavio S. Anselmetti
Daniel Ariztegui

INTRODUCTION

The past two decades have witnessed renewed interest in exploring the climatic context in which lowland Maya culture arose, developed, and collapsed. During the early part of the twentieth century, it was generally assumed that neotropical climate had remained relatively stable throughout the Holocene. That myth persists today, in part because isotope studies from Greenland ice cores indicated minimal temperature change at high latitudes over the last 11,500 years (Oldfield 2000). With respect to Mesoamerica, it was thought that recent instrumental measurements of temperature and rainfall on the Yucatán Peninsula were no different from climate conditions that prevailed during the period of ancient Maya florescence. As approaches to reconstructing past climate became more sophisticated and began to focus on low-latitude sites, it was apparent that tropical climate conditions following the Pleistocene were not static but varied considerably, especially with respect to moisture availability (e.g., Street and Grove 1976; Abbott et al. 1997; Bridgwater, Heaton, and O'Hara 1999; Verschuren, Laird, and Cumming 2000; deMenocal et al. 2000). Scientists working at sites around the globe inferred past climate changes using numerous sources of information, including marine and freshwater sediments, pollen analysis, tree-ring data, tree-line migration, shifting lake stages, and glacial advances and retreats. Soon, there was ample evidence of worldwide Holocene climate fluctuations on millennial and centennial timescales, and terms such as "hypsithermal," "Medieval Warm Epoch," and "Little Ice Age" became established in the paleoclimatology lexicon.

In the Maya lowlands, archaeologists were also beginning to uncover evidence of long-term climatic instability. At El Mirador, in Petén, Guatemala, Terminal Preclassic housemound groups were discovered in *bajos* that are presently subject to seasonal inundation. It is unlikely that considerable labor would have been invested for construction in areas prone to annual flooding (Dahlin 1983). At Lake Cobá, in Quintana Roo, Mexico, Maya-constructed walkways on the lakeshore are presently underwater, suggesting that they were built during a drier period when the lake stage was lower (Folan et al. 1983).

A reliable water source is essential for both human physiological requirements and agriculture. Given the high seasonality and low total annual rainfall across much of the Yucatán Peninsula, combined with the scarcity of accessible surface water and groundwater, the Maya took special steps to obtain sufficient water in some localities. Archaeologists have long noted the proximity of some Maya archaeological sites (e.g., Dzibilchaltún, Chichén-Itzá, Yaxhá, and Ceibal), to available water sources including wetlands, cenotes, lakes, and rivers (e.g., Wilson 1984; Adams 1980; Rice and Rice 1990). Where surface water and groundwater were not readily available, the Maya built water storage facilities such as *aguadas* and reservoirs (Matheny 1976; Scarborough and Gallopin 1991; Back 1995; Scarborough 1993). The importance of water to the Maya is clearly evident in their mythology, art, and architecture (Back 1981).

Because much of the Yucatán Peninsula lies in a zone that is climatically marginal for agriculture, small changes in precipitation income can have devastating effects on crop yields. The importance of the annual hydrologic cycle in the Maya lowlands has always been acknowledged. Once it became apparent that late Holocene climate was variable, archaeologists began to address how changes in moisture availability affected prehistoric Maya settlement distributions and cultural development (Folan et al. 1983; Gill 2000; Messenger 1990). It was suggested that long-term climate changes influenced the human population at Dzibilchaltún directly by limiting access to water in Cenote Xlacáh, and indirectly by affecting the local flora and fauna (Folan 1985).

Evidence for climate change during the Maya epoch was emerging from the archaeological record (Dahlin 1983; Folan 1985; Folan et al. 1983). Independently, researchers began to explore the paleoclimate of the Maya area using global circulation models (GCMs), and hindcasting climate on

the Yucatán Peninsula by correlation with long-term temperature records from distant sites. Gunn and Adams (1981) proposed that precipitation in southern Mexico was correlated with temperature in the northern hemisphere. They showed that cold northern hemisphere temperatures produce wetter weather on the Yucatán Peninsula, and warmer temperatures are associated with drier conditions due to a reduction in midwinter rainfall. By correlation with glacial advances and retreats in Alaska and Sweden, Gunn and Adams (1981) inferred shifts in available moisture for the Maya lowlands during the latter Holocene, and speculated that reduced precipitation around A.D. 900 (1150 ^{14}C yr B.P.) negatively impacted both Maya trade and agriculture.

More recently, Gunn, Folan, and Robichaux (1995) used an alternative approach to retrodict moisture availability in the Maya region. They first correlated mean global temperature with seasonal discharge from the Candelaria River, Mexico, for the years 1958 to 1990; this analysis convinced them to reverse their earlier ideas, and to conclude that global warm periods are associated with wetter climate in the Maya lowlands. They discovered that warm temperatures preclude a local dry season, interfering with *milpa* production by preventing burning of felled vegetation. Cold global temperatures postpone the onset of the seasonal rains and delay planting. Gunn, Folan, and Robichaux (1995) argued that intermediate temperatures promote the wet-dry seasonality that is most conducive to successful swidden agriculture. They inferred past mean global temperature using a combination of October orbital precession, an index of volcanism, and solar energy output. Next, they reconstructed discharge of the Candelaria River—that is, past climate conditions on the Yucatán Peninsula—for the period of Maya occupation (approximately the last 3,000 years), and related river runoff to several key developments in Maya prehistory. According to their model, the ninth century A.D. Maya Collapse occurred during a period of global cold temperatures and drought conditions in the Maya lowlands.

Paleoclimatological studies were also undertaken using variables found in lake and wetland sediment cores collected at sites throughout the Maya lowlands. Because direct measurements of past climate conditions cannot be made, "proxy" variables such as pollen, diatoms, and invertebrate remains are used to reconstruct past climate. Pollen analysis provided reliable information about the Pleistocene/Holocene climate transition in the southern Maya lowlands (Leyden 1984; Leyden et al. 1993; Leyden et al. 1994), but proved to be of little utility for inferring climate conditions in the later Holocene because watersheds in Petén, Guatemala, were densely populated (Rice and Puleston 1981; Rice, Rice, and Deevey 1985; Rice and Rice 1990), and regional, human-mediated deforestation over the last 3,000 to 4,000 years overwhelmed any climate signal (Bradbury et al. 1990; Deevey 1978; Deevey et al. 1979; Vaughan, Deevey, and Garrett-Jones 1985;

Brenner, Leyden, and Binford 1990; Leyden 1987; Hansen 1990; Brenner 1994; Islebe et al. 1996; Curtis et al. 1998).

Alternatively, paleoclimate inferences have been based on paleosalinity changes in water bodies, using assemblages of sedimented microfossils from groups of organisms with known salt tolerances. For instance, salinity changes over the past 7,000 years were inferred at Cobweb Swamp, northern Belize, based on subfossil communities of gastropod mollusks (snails) and small crustaceans called ostracods (Alcala-Herrera et al. 1994). Shifts in the salt content of the wetland were initially driven by sea-level rise. The last 3,500 years of the record is thought to have been influenced principally by climate change. Long-term lake level trends (i.e., moisture availability) were also inferred based on salinity preferences of diatom assemblages from lake sediment cores collected along a transect from Cenote San José Chulchacá, in the dry area west of Mérida, to Lake Cobá, in the wetter area of eastern Quintana Roo (Figure 4.1) (Whitmore et al. 1996). Each diatom taxon has specific ecological requirements and possesses a distinctive, identifiable siliceous "skeleton" or frustule that is preserved in lake sediments.

There are advantages and disadvantages to the various paleoenvironmental approaches that have been used to address late Holocene climate change in the Maya area. Modeling efforts did not require collection of new data, but relied on data from remote sites and assumed that modern correlations between climate variables in Yucatán (e.g., rainfall), and temperature in high-latitude sites, for example, could be projected back in time over millennia. Paleolimnological data from the Yucatán Peninsula have the advantage of coming from the region of interest, but there are limitations, nonetheless, to their use. As mentioned previously, human disturbance over the last three to four millennia had such a dramatic impact on regional vegetation that the pollen record could not be employed to decipher late Holocene climate change in the region. Inferences based on stratigraphic shifts in community assemblages of ostracods, gastropods, and diatoms require knowledge about the ecology of the enumerated taxa. There are, however, few ecological studies of the contemporary plankton and benthos in the Maya lowlands, and regional calibration data sets have not been developed that quantitatively relate water column chemistry to the relative abundance of various taxa. Furthermore, studies of stratigraphic shifts in subfossil communities require specialized sample preparations, identification of numerous taxa, and enumeration of large numbers of individuals.

As early as the 1970s, a new technology was being applied to the question of Maya paleoclimate—that is, reconstruction of past moisture conditions based on stratigraphic measurement of the oxygen isotope signal in the sedimented carbonate shells of aquatic invertebrates (Covich and Stuiver 1974). By the 1990s, advances in stable isotope mass spectrometry had facilitated the use of small sample sizes and permitted rapid analysis,

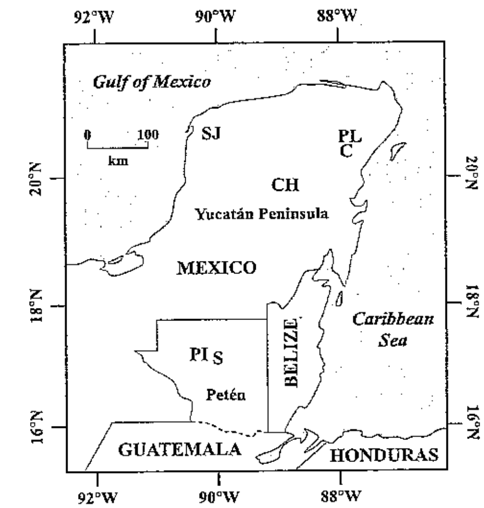


FIGURE 4.1. Map of the Maya lowlands. Coring sites referred to in the text include Cenote San José Chulchacá (SJ), Lake Punta Laguna (PL), Lake Cobá (C), and Lake Chichancanab (CH) in the northern Yucatán Peninsula, and Lake Petén-Itzá (PI) and Lake Salpetén (S) in the Central Petén Lake District, Guatemala.

making this technique ideal for generating high-resolution reconstructions of past changes in moisture availability. In this chapter, the theoretical basis for using stable isotope geochemistry to address Maya paleoclimate is explained. The selection of appropriate study lakes is discussed as are the advantages and limitations of applying isotope-based methods for climatic inferences. Results and interpretations from recent investigations in the Maya lowlands are also summarized.

THEORETICAL BASIS FOR ISOTOPE-BASED PALEOCLIMATE STUDIES

There are three naturally occurring stable isotopes of oxygen: ^{16}O, ^{17}O, and ^{18}O. The lightest isotope (^{16}O) is most common, representing 99.7630 percent of the isotope pool. ^{17}O is the rarest, representing only 0.0375 percent of the total, while ^{18}O accounts for 0.1995 percent. Because of their different masses, the isotopes behave differently when they enter into physical, chemical, and biological processes in the environment. It is the differential behavior of the heaviest (^{18}O) and lightest (^{16}O) isotopes, referred to as fractionation, and their changing relative abundance in the environment, that is exploited for paleoclimate reconstructions. The isotopic signature, or $\delta^{18}O$ of environmental samples, can be measured by isotope ratio mass spectrom-etry; results are presented in standard delta (δ) notation. This is an expres-sion of the $^{18}O/^{16}O$ ratio in a sample relative to the ratio in the Vienna Peedee Belemnite (VPDB) standard, expressed on a per mil (‰) basis:

$$\delta^{18}O = \frac{(^{18}O/^{16}O)sample - (^{18}O/^{16}O)VPDB}{(^{18}O/^{16}O)VPDB} \times 10^3 \ ‰$$

Talbot (1990), Chivas et al. (1993), Curtis and Hodell (1993), Holmes (1996), and others have presented the rationale for using the stable isotope ($\delta^{18}O$) signal in sedimented freshwater carbonate shells to reconstruct past climate conditions. The isotope ratio ($^{18}O/^{16}O$) in sedimented shell material is governed by the isotope ratio in the lakewater at the time the organism lived, the temperature at which carbonate precipitation occurred, and bio-logical fractionation by the organism (von Grafenstein, Erlernkeuser, and Trimborn 1999). Because there is little evidence for major excursions in the mean temperature of the tropics during the Holocene, the changing $\delta^{18}O$ of lake water over the last 10,000 years has been the major determinant of the $\delta^{18}O$ in shell carbonate. The $^{18}O/^{16}O$ ratio of lake water has, in turn, been controlled by hydrologic variables. Assuming that the $\delta^{18}O$ of regional rainfall has remained fairly constant during the Holocene, relative changes in hydrologic inputs (precipitation and runoff) and outputs (evaporation and outflow) govern the in-lake oxygen isotopic ratio (Figure 4.2).

In tropical, closed-basin lakes (i.e., those that lack significant overland bud-outflows), the factor that most influences the waterbody's hydrologic (Fontes get is the relationship between evaporation (E) and precipitation (P) and Gonfiantini 1967; Gasse et al. 1990; Lister et al. 1991). During dry peri-ods (high E/P), ^{18}O becomes relatively concentrated in the lake water be-cause $H_2{}^{16}O$, with its higher vapor pressure, is preferentially lost to evapo-ration. Conversely, during moist periods (low E/P), the water column is relatively depleted of ^{18}O (Figure 4.2). Measurements of the $\delta^{18}O$ signature of rainfall and lake water in the Maya lowlands demonstrate that ^{18}O is con-centrated in regional lakes due to evaporation (Figure 4.3).

Short-lived aquatic organisms that form calcium carbonate shells pre-serve a record of the E/P ratio that prevails during their lifetimes (von Grafenstein, Erlernkeuser, and Trimborn 1999). When they die, their re-mains are buried in sediments on the lake bottom, thereby preserving an ar-chive of past climate change (Figure 4.4). The stratigraphic paleoclimate re-cord can be deciphered by mass spectrometric measurement of the $\delta^{18}O$ of sedimented shells. More positive $\delta^{18}O$ values generally indicate higher E/P

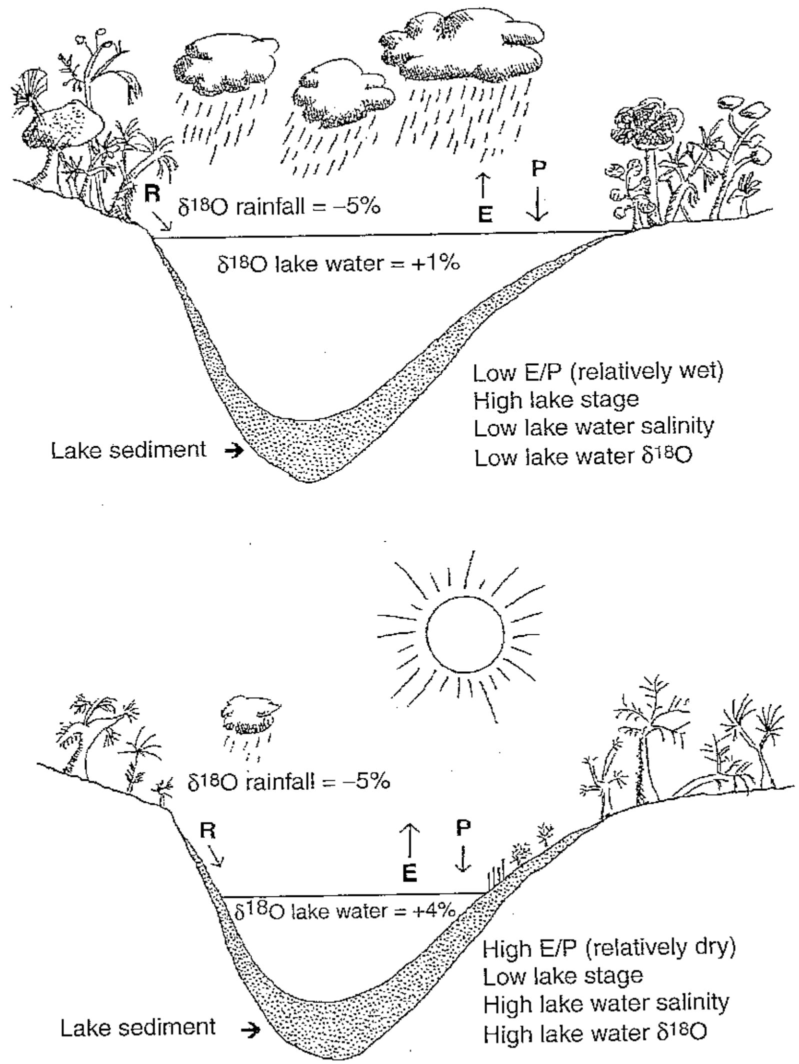


FIGURE 4.2. Cross section of a hypothetical closed-basin lake in the Maya low-lands, showing differences with respect to lake stage, lake water salinity, and lake water $\delta^{18}O$ during relatively wet (top) versus relatively dry (bottom) Holo-cene periods. Note that the $\delta^{18}O$ signal of input (rainwater) is assumed to have remained constant during wet and dry Holocene periods. In watersheds with negligible surface outflow, changes in lake volume are a function of the shifting relation between evaporation (E) and precipitation (P) over the lake and surface runoff (R).

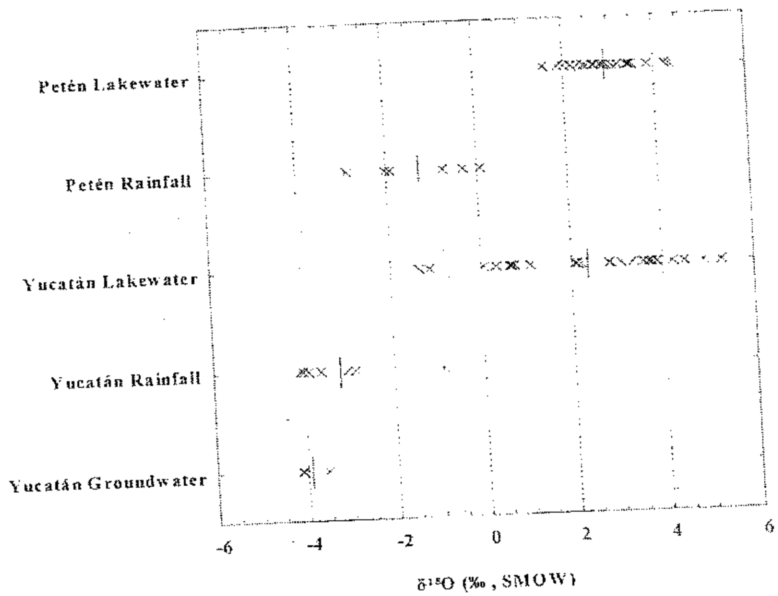
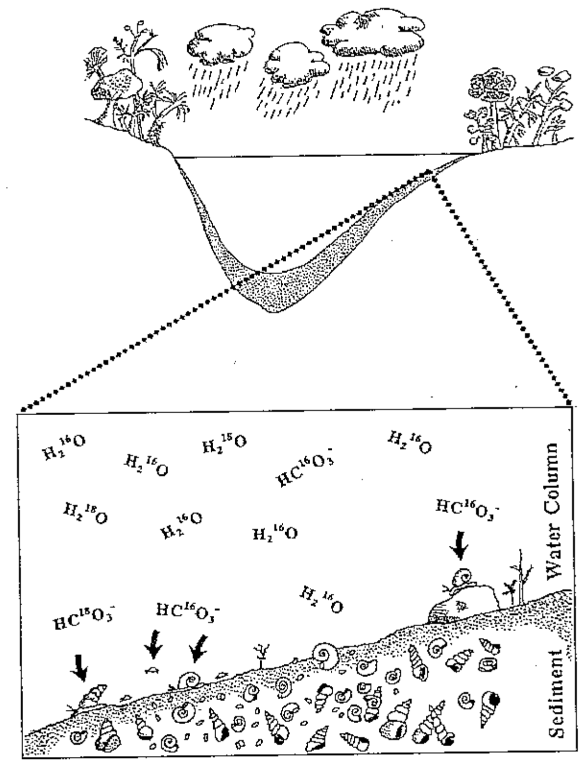


FIGURE 4.3. Oxygen isotope ($\delta^{18}O$) signatures of lake water, rainwater, and groundwater collected at sites in the Central Petén Lake District, Guatemala, and northern Yucatán Peninsula, Mexico. The $\delta^{18}O$ of samples is expressed relative to Standard Mean Ocean Water (SMOW). Individual sample results are indicated by the symbol *x*, and the mean value for samples is shown as a vertical line. For Petén, 22 lake water and 6 rainwater samples were analyzed. In the Yucatán Peninsula, the numbers of samples analyzed from lake water, rainwater, and groundwater were 32, 7, and 3, respectively. Differences between values for lake water and rainwater illustrate that ^{18}O is concentrated in lakes as a consequence of evaporation. In the long term, weighted mean $\delta^{18}O$ values for rainwater collected at other low-elevation, circum-Caribbean sites range from –5.65‰ to –4.01‰. (*Source:* Rozanski, Araguás-Araguás, and Gonfiantini 1993.)

(drier) conditions and/or decreased surface and groundwater inflow, whereas more negative values indicate relatively lower E/P (moister) conditions and/or increased inflow at the time the organism lived.

CHOOSING A STUDY SITE

The two taxonomic groups that are commonly employed for isotope-based climate reconstructions are the ostracods, which form bivalve shells of calcite, and the gastropods (snails), which are mollusks that make ar-



$$CO_2 + H_2{}^xO \leftrightarrow H_2C^xO_3$$
$$H_2C^xO_3 \leftrightarrow H^+ + HC^xO_3{}^-$$
$$HC^xO_3{}^- \leftrightarrow H^+ + C^xO_3{}^=$$
$$Ca^{++} + C^xO_3{}^= \leftrightarrow CaC^xO_3$$

FIGURE 4.4. Cross section of a hypothetical, closed-basin lake in the Maya lowlands (top), with a close-up of the sediment-water interface (bottom). The carbonate equilibrium equation is included below the diagram. Note that the isotopic signature of the water molecule ($H_2{}^xO$) in the equation, denoted by *x*, is reflected in the precipitated calcium carbonate (CaC^xO_3). Water and bicarbonate ions ($HC^xO_3{}^-$) contain $^{18}O/^{16}O$ ratios that reflect climate (E/P) conditions (see Figure 4.2). Ostracods and gastropods living above the sediment surface make calcium carbonate ($CaCO_3$) shells that reflect the $^{18}O/^{16}O$ ratio of the lake water in which they live. When they die, their remains are incorporated into the lake sediment, thereby preserving a record of past climate change (E/P) that can be interpreted through mass spectrometric measurement of the $\delta^{18}O$ of the sedimented shells.

agonite shells. The first step in undertaking a study of oxygen isotope changes through time is collection of a sediment core from an appropriate lake. Not all water bodies, however, yield satisfactory material for inferring past climate, and certain conditions must be met to increase the probability of generating an interpretable paleoclimate record (Table 4.1). First, the lake should lie within the geographic region of interest. Extrapolation of findings from distant sites is problematic because of spatial climate variability and differences in "boundary" conditions between the area of interest and the sediment study locality.

Second, the selected study lake should be relatively "closed" hydrologically, with most water lost to evaporation, rather than via overland outflow or seepage. Third, morphometrically, the lake must be sufficiently deep and have enough volume that it does not desiccate during periods of low rainfall. Drying at the core site can cause discontinuities (hiatuses) in sedimentation, and may lead to the erosion of sediments that were deposited previously. On the other hand, lakes that are too large and deep may display negligible isotope shifts in response to modest changes in E/P, because they lose or gain only a small fraction of their total volume when E/P changes occur. In general, tropical lakes with maximum depths of about 15 to 25 meters (m) yield acceptable records.

All factors that alter the hydrologic regime of a water body can potentially influence its volume, as well as the salinity and stable isotope signature of its water (see Figure 4.2). For lakes lacking outflows, shifts in precipitation and runoff (input) and evaporation (output) are probably of paramount importance to the hydrologic budget. Nevertheless, processes that change runoff (input) may also influence the lake hydrology. For instance, massive deforestation in a drainage basin can alter watershed transpiration and soil moisture storage (Bosch and Hewlett 1982; Bruijnzeel 1990; Stednick 1996), perhaps enhancing delivery of relatively ^{18}O-depleted rainfall and groundwater to the lake (Rosenmeier, Hodell, Brenner, Curtis, Martin, et al. 2002).

The chemistry of the lake water is crucial in that it must have sufficient dissolved inorganic carbon [e.g., carbonate (CO_3^{-2}) and bicarbonate (HCO_3^{-1})] to enable invertebrate shell formation. Furthermore, the lake sediments should be rich in shell remains (Figure 4.5).

Those remains must also be well preserved, because isotopic signatures can be altered by diagenesis (i.e., postdepositional changes such as dissolution and reprecipitation of shell carbonate). Sediment cores should possess remains of adult organisms from at least one taxon at all sediment depths throughout the entire record because some species may precipitate calcium carbonate ($CaCO_3$) out of isotopic equilibrium with the lake water. Species-specific isotopic fractionation during shell formation is referred to as "vital effects." If a single species is not encountered throughout the entire record,

TABLE 4.1. Criteria for lakes used in isotope-based ($\delta^{18}O$) reconstructions of Maya paleoclimate

Lake Variable	Requirements
Location	The lake should be located near the archaeological site or region of interest. The Maya lowlands are spatially variable with respect to moisture availability, making it difficult to extrapolate paleoclimate results over great distances.
Hydrology	The basin should be fairly closed with respect to hydrologic outputs, losing most of its water to evaporation (E).
Depth	The lake should be sufficiently deep that the core site will not desiccate during dry periods.
Volume	Excessively large, deep lakes may be unresponsive to climate changes as they gain or lose a small proportion of their total water volume due to changes in E/P. There must be a balance between sufficient depth and appropriate volume (V).
Water chemistry	Lake water must be sufficiently hard (i.e., have high enough bicarbonate [HCO_3^{-1}]) to permit shell formation and preservation.
Microfossil abundance	Well-preserved ostracod and snail shells should be abundant throughout the sediment core.
Microfossil taxonomy	Multiple individuals of a single taxon should be measured at each sampling level throughout the profile to control for "vital effects." Preferably, several taxa with differing ecologies should be analyzed to assess consistency of the long-term E/P signal.
Chronology	Chronology in karst areas should be based on AMS-^{14}C dating of terrestrial organic matter (e.g., seeds, stems, leaves, or pollen grains) to avoid hard-water-lake error. Lacking abundant terrestrial remains, steps can be taken to correct ages obtained from aquatic remains such as ostracods or snails.
Sedimentation rate	Sedimentation rates should be sufficiently high to permit paleoclimate reconstruction on a scale with high temporal resolution. Contiguous 1 cm sampling yields decadal resolution in lakes with mean sediment accumulation of 0.1 cm yr^{-1}. Excessively high sedimentation rates, especially those due to soil erosion, may dilute carbonate fossils.
Watershed	Factors that alter drainage basin hydrology (e.g., deforestation) may confound interpretation of $\delta^{18}O$ records.
Alternate proxies	Ideally, isotope-based E/P reconstructions should be supported by additional proxies such as sediment geochemistry (e.g., precipitated salts or trace metal ratios), or microfossils such as diatoms.

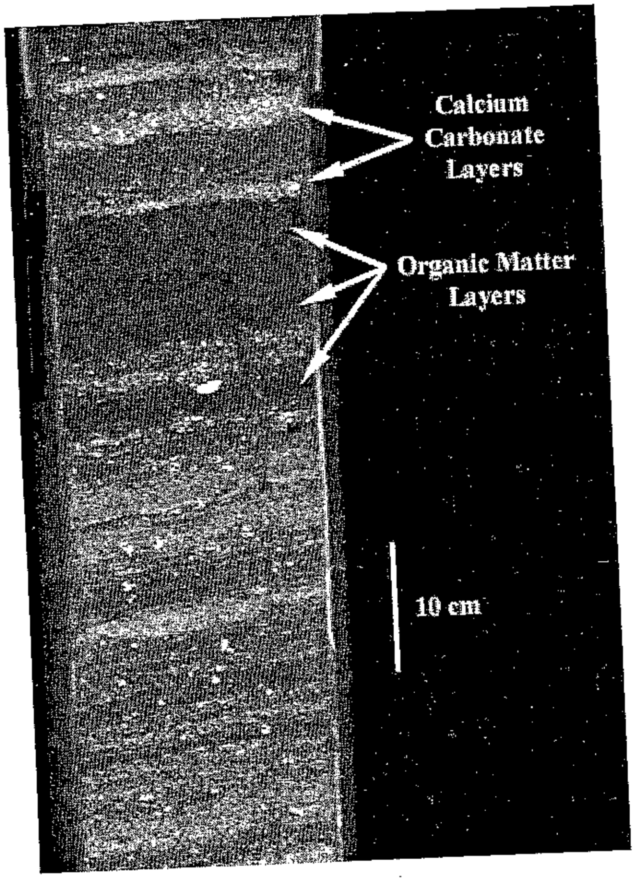


FIGURE 4.5. Section of a sediment core from Lake Punta Laguna in the Maya lowlands, showing finely laminated sediment consisting of organic matter (dark bands) and calcium carbonate (light bands). Note the high density of gastropods (white dots). Sampling intervals of 1 cm generally represent 5 to 20 years of sediment accumulation.

multiple species can be used, but their isotope signatures should be compared where they are found in the same stratigraphic levels to assess isotopic offset due to "vital effects." Ideally, several taxa found throughout the profile should be analyzed separately so that multiple isotope records can be generated using organisms that occupy different ecological niches. In each stratigraphic level, multiple individuals of a given taxon will constitute a sample, and sample measurement yields the mean $\delta^{18}O$ for the sedimented assemblage.

Prior to collecting and subsampling a core, it is impossible to determine whether all these criteria will be met. Nevertheless, the lake shoreline can be inspected for the presence of abundant small shells, and surface sediments throughout the lake can be sampled quickly with a dredge to establish the likelihood of collecting cores rich in carbonate remains.

The lake sediment core should have a continuous, datable record that covers the time period of interest. One analytical conundrum that has arisen in the Maya lowlands is the difficulty of generating reliable sediment chronologies. Lakes in karst country almost invariably have high bicarbonate (HCO_3^{-1}) concentrations in the water column that permit shell formation. It is, however, these same high bicarbonate concentrations that confound radiocarbon dates on organic matter and inorganic carbon (i.e., shells) of aquatic origin. This problem is called hard-water-lake error (HWLE) (Deevey and Stuiver 1964) and stems from the fact that some of the carbon in the carbon dioxide (CO_2) and bicarbonate (HCO_3^{-1}) used for photosynthesis by aquatic algae and other plants comes not from the atmosphere, but rather from the dissolution of ancient (e.g., Cretaceous, in the case of Petén, Guatemala), ^{14}C-depleted rocks in the watershed.

Similarly, snails and ostracods that form their shells from dissolved bicarbonate in such systems may incorporate ^{14}C-depleted carbon into their calcium carbonate shells. These processes lead to violation of a key assumption of radiocarbon dating—namely, that plants or shelled organisms fix ^{14}C from a carbon pool in equilibrium with the atmospheric ^{14}C concentration. This disequilibrium, which can make radiocarbon ages on aquatic organisms from karst lakes artificially old, can be transferred up the lacustrine food web, affecting dates on all materials of aquatic origin. The advent of accelerator mass spectrometry (AMS-^{14}C) dating has remedied this problem to some degree by enabling the analysis of small pieces of terrestrial organic matter identified in lake cores. The most reliable chronologies are thus based on AMS-^{14}C dating of terrestrial seeds, twigs, leaves, and even pollen grains. Paired dates (i.e., from the same core depths) on terrestrial (e.g., wood) and aquatic (e.g., shell) material from a core can provide an estimate of the magnitude of HWLE. These paired dates can sometimes be used to adjust, or correct, a chronology based largely on abundant shell remains.

The sediment accumulation rate at the coring site should be high enough to permit sufficient sampling resolution, but not so high that microfossils are diluted. Very high rates of sediment accumulation may permit high sampling resolution, but may also preclude recovery of a long record if the corer cannot penetrate the entire sediment lens. A rough rule of thumb is that average, long-term Holocene sedimentation rates in Maya-area lakes are on the order of 0.1 centimeters per year ($cm \cdot yr^{-1}$), so that contiguous 1 cm sampling intervals yield roughly decadal resolution on environmental changes.

Once again, it is difficult to evaluate the sedimentation rate at a core site prior to collection and dating of the profile.

High-resolution seismic surveying can help to assess sediment geometry and sediment distribution in the lake's subsurface (Figure 4.6). The reflections seen on seismic profiles usually image the bedding surfaces, and thus show stratigraphic horizons in the sediment record. By mapping geometric unconformities and units with characteristic seismic facies, one can establish a seismic stratigraphy, which partitions the sediment succession into seismic sequences. Mapping unconformities is particularly important, because it documents periods of nondeposition and/or erosion and can be used to reconstruct lake level changes. Thinning or thickening of seismic units also documents lateral changes in sedimentation rate. Seismic profiles can yield crucial information to identify appropriate coring sites for acquiring long, reliable sediment records of Holocene climate change that span the period of Maya occupation. Depending on the study objectives, seismic stratigraphies can be used to help choose the best coring sites to yield cores with high temporal resolution (high sedimentation rate, continuous sedimentation), long time sequences (low sedimentation rate, continuous sedimentation), or evidence of specific environmental changes (e.g., lake level fluctuations, or unconformities).

Recent studies in Lake Salpetén, Guatemala, illustrate the utility of using seismic surveys to choose coring locations. A north-south profile across the width of the lake (Figure 4.6) shows a steep, fault-controlled northern shore that leads to the deepest part of the basin (maximum depth = 32 m). In contrast, the slope of the southern shore is gentle. Intense sediment focusing has led to the deposition of a thick sediment package in the deepest part of the basin.

It is possible to correlate the seismic profile to Salpetén (SP) core 80-1, and identify lithologic changes associated with prominent reflectors, as well as determine the age of these reflectors using radiocarbon dating. Near the base of the sequence is a sharp reflector that coincides with the contact between gypsum and organic-rich, laminated gyttja, which marks the Pleistocene/Holocene boundary. The gyttja is overlain by a thick "Maya Clay" unit, which is an erosional deposit that has been equated with human-induced deforestation of the catchment. The base of the "Maya Clay" in SP core 80-1 was dated to 3160 ^{14}C yr B.P. (about 1400 B.C.), by AMS-^{14}C analysis of terrestrial material, suggesting that the Petén environment had already sustained Maya impact by the beginning of the middle Preclassic Period (about 1000 B.C.). The Maya clay is overlain by another organic-rich gyttja that was deposited following the Classic Collapse and subsequent reforestation of the watershed. Ongoing studies involve using the seismic profiles to map the three-dimensional distribution of the "Maya Clay" in the

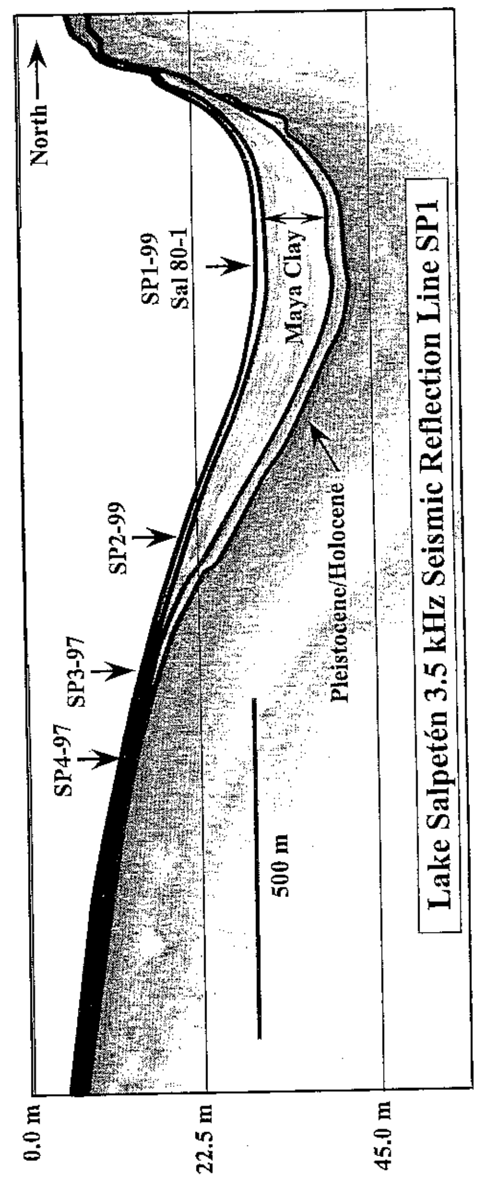


FIGURE 4.6. Single-channel, high-resolution seismic reflection profile from Lake Salpetén. Profile was recorded in digital format using a 3.5 kHz pinger source and GPS navigation. Note lateral variability in thickness of the sediment strata as well as several unconformities, in particular at shallow water depths. The profile for Lake Salpetén shows the sediment stratigraphy along the coring transects. Vertical arrows indicate approximate core locations. The seismic profile suggests the two cores from shallower water display discontinuities in sediment accumulation. Sub-bottom profiling, though not attainable in all lakes, can be enormously helpful for choosing core site.

basin and to estimate changing rates of erosion as a consequence of human impact on the environment.

COMPLEMENTARY SEDIMENT PROXIES

Study lakes should ideally contain additional sediment proxies that can be used to support isotope-based paleoclimate inferences. For instance, measurement of other geochemical variables in sedimented shell material, including their strontium/calcium ratio (Sr/Ca) and magnesium/calcium ratio (Mg/Ca), can be helpful for testing isotope-based paleoclimate interpretations (Chivas, De Dekker, and Shelley 1986; Chivas et al. 1993; Curtis and Hodell 1993; Holmes 1996). Diatoms have proven to be sensitive biological indicators of lake water salinity, which in turn reflects local changes in moisture availability (Fritz et al. 1999). In water bodies with high concentrations of dissolved ions, salts such as calcium sulfate ($CaSO_4$ = gypsum) or sodium chloride (NaCl) may precipitate during periods of low lake level and high salinity (Hodell, Curtis, and Brenner 1995). In addition to their utility as subjects for isotopic study, ostracod species assemblages reflect salt concentrations in a water body (Bridgwater, Heaton, and O'Hara 1999; Bridgwater, Holmes, and O'Hara 2000).

RESULTS AND DISCUSSION

Isotope Records from the Northern Yucatán Peninsula

The earliest effort to develop an isotope-based climatic reconstruction for the Yucatán Peninsula was undertaken by Covich and Stuiver (1974) at Lake Chichancanab. They retrieved a 12-m core of which approximately the topmost 9 m represented Holocene deposition. The oldest Holocene radiocarbon age in the core was 7380 ^{14}C yr B.P. (6230 B.C.), and $\delta^{18}O$ measurements were run on *Pyrgophorus* snail shells at 17 Holocene stratigraphic levels, yielding an average sampling interval of about 500 years. Today, improvements in mass spectrometry enable rapid $\delta^{18}O$ analysis of very small samples, permitting the generation of continuous, high-resolution isotope stratigraphies. Whereas Covich and Stuiver (1974) had to amass sample weights of about 50 milligrams (mg), modern instruments are capable of measuring isotope ratios in samples weighing as little as 20 μg (micrograms).

In the 1990s, lake sites on the northern Yucatán Peninsula were re-cored to obtain high-resolution paleoclimate records that span the period of Maya occupation. To date, the most reliable isotope-based reconstructions of past E/P have come from Lake Chichancanab (19°50' N, 88°45' W) and Lake

Punta Laguna (20°38' N, 87°37' W). A 4.9 m core was collected from 6.9 m of water in Lake Chichancanab (Figure 4.7) and 1 cm sectioning of the profile yielded a mean sampling resolution of about nineteen years. The section terminated on a paleosol that contained terrestrial gastropods (Hodell, Curtis, and Brenner 1995). The lake first filled with water about 8200 ^{14}C yr B.P. (7250 B.C.); the early lacustrine deposits are characterized by high concentrations of gypsum ($CaSO_4$), relatively high $\delta^{18}O$ values for gastropod and ostracod shells, and large numbers of *Ammonia beccarii,* a benthic foraminifer. *Ammonia beccarii* can tolerate a wide range of temperatures (10-35° C) and salinities [7-67 grams per liter ($g \cdot L^{-1}$)], but reproduce only at salt concentrations between 13 and 40 g L^{-1} (Bradshaw 1957). High E/P (i.e., dry climate) is inferred for this part of the early Holocene record based on biological and geochemical indicators that point to low lake level and high salinity.

Lake Chichancanab filled rapidly after about 7200 ^{14}C yr B.P. (6000 B.C.) and carbonate, rather than gypsum, began to precipitate. The stable isotope signatures of shells declined and *A. beccarii* was no longer present. These proxies all indicate relatively moister mid-Holocene conditions that persisted until about 3000 ^{14}C yr B.P. (1250 B.C.) (Figure 4.7). The moist conditions at Chichancanab reversed at about 3000 ^{14}C yr B.P. (1250 B.C.), and a drying trend is inferred from increased sulfur (gypsum) deposition and higher $\delta^{18}O$ values for gastropod and ostracod shells (Figure 4.7).

Late Holocene drying has also been documented in lake sediment cores from other sites around the Caribbean, including Haiti (Hodell et al. 1991) and northern Venezuela (Bradbury et al. 1981; Leyden 1985). Within chronologic uncertainty, the oxygen isotopic signal from Lake Miragoane, Haiti, agrees remarkably well with the geochemical records from Lake Chichancanab (Figure 4.8). Both indicate a trend toward drier climate in the late Holocene beginning at about 3000 ^{14}C yr B.P. (1250 B.C.) and intensifying at 2500 ^{14}C yr B.P. (750 B.C.) This climatic event coincides with the base of the "Maya Clay" unit in Lake Salpetén (Figure 4.6), raising the question to what extent vegetation and erosional changes at this time were due to climate and/or Maya deforestation.

According to the Lake Chichancanab and Lake Miragoane isotope records, swidden agriculture developed in the lowlands coincident with a climatic drying trend. The timing of cutting, burning, and planting was critical to successful agricultural harvests, especially under increasingly dry conditions in this region where rainfall is highly seasonal. The slash-and-burn cycle was no doubt administered by Maya leaders using calendars based on astronomic observations (Milbrath 1999). The importance of the timing and quantity of precipitation to Maya agriculture is evident in the widespread portrayal of the rain god Chac on architectural structures in the northern part of the peninsula.

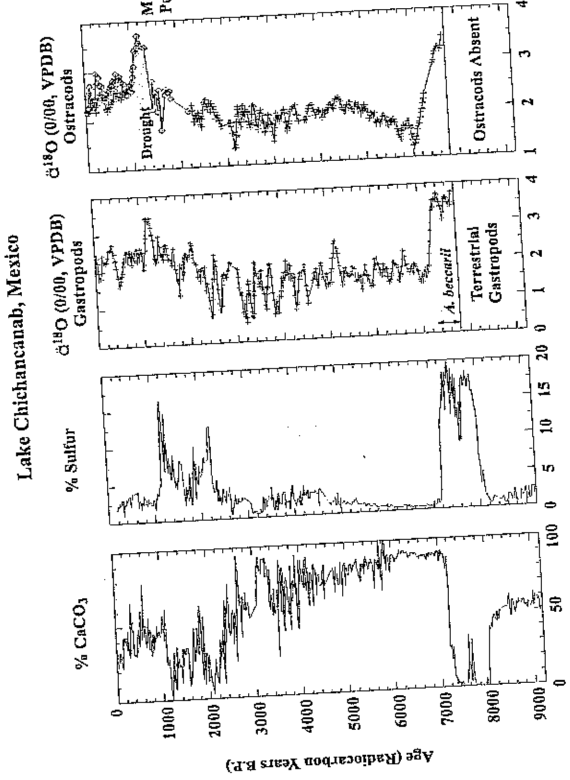


FIGURE 4.7. Paleoclimate record from Lake Chichancanab (Hodell, Curtis, and Brenner 1995) showing percent $CaCO_3$, percent sulfur (gypsum), $\delta^{18}O$ of gastropod shells *(Pyrgophorus coronatus)*, and $\delta^{18}O$ in ostracod shells *Physocypria* sp. (+) and *Cyprinotus* cf. *salinus* (✧), based on sampling at 1cm intervals. Values are plotted against time, with the major period of Maya occupation (between about 3000 and 400 ^{14}C yr B.P., or 1250 B.C. to A.D. 1450) indicated to the right of the plot. Oxygen isotopic results are three-point running means. The foraminifer *A. beccarii* was found between about 7800 and 7300 ^{14}C yr B.P. (6570 and 6150 B.C.). The gray line in the ostracod plot indicates the Late Classic drought dated at 1140 ± 35 ^{14}C yr B.P. (A.D. 920 ± 35 yr). (*Source:* Reprinted with permission from D.A. Hodell, J.H. Curtis, and M. Brenner, 1995, Possible role of climate in the collapse of Classic Maya civilization, *Nature*, 375, pp. 391-394. Copyright 1995 *Nature*.)

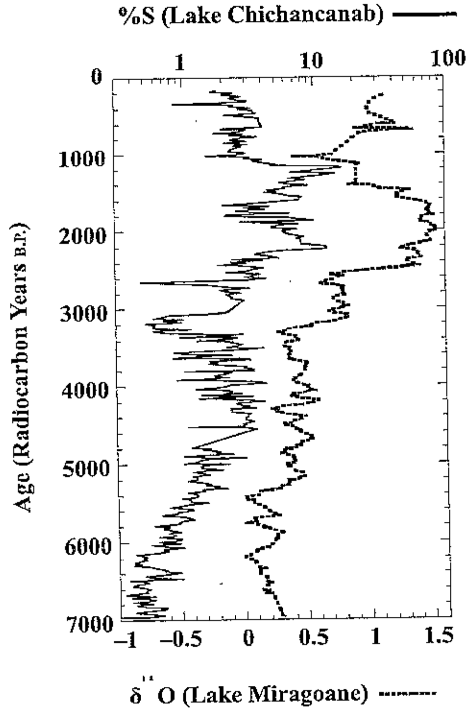


FIGURE 4.8. Weight percent sulfur (% S) record from Lake Chichancanab (plotted on a logarithmic scale) and a five-point running mean of the $\delta^{18}O$ record of the ostracod *Candona* sp. from Lake Miragoane, Haiti (Hodell et al. 1991; 1995). Note the general similarity of the records in the Holocene, especially the increase beginning at ~3000 ^{14}C yr B.P. (1250 B.C.), the abrupt increase at ~2500 ^{14}C yr B.P. (750 B.C.) and the decline at ~1000 ^{14}C yr B.P. (A.D. 1010). (Copyright 1995 *Nature*.)

Early Maya farmers dealt not only with interannual variations in the timing and amount of rainfall, as they do today, but also with pronounced decade-to-century shifts in available moisture between about 3000 ^{14}C yr B.P. (1250 B.C.) and about 1050 ^{14}C yr B.P. (A.D. 1000). In the Chichancanab core, samples from about 65 cm depth contain high concentrations of gypsum and high $\delta^{18}O$ values, indicating very dry (high E/P) climate conditions. The proxies suggest this was one of the driest periods of the Holocene since the initial filling of the lake (Figure 4.7). The culmination of this dry episode occurred at 1140 ^{14}C yr B.P. (A.D. 920), as determined by an AMS-^{14}C date on a terrestrial seed at 65 cm. It is probable that terrestrial vegetation grew close to the coring site during this period when the lake volume was shrinking. The timing of this very dry episode coincides closely with the Classic Maya Collapse (Hodell, Curtis, and Brenner 1995).

A 6.3 m core was collected from just over 6 m of water in Lake Punta Laguna (Figure 4.9). The basal age on the section is only 3500 ^{14}C yr B.P. (1820 B.C.), indicating that the mean sediment accumulation rate at the site was high (0.18·cm yr^{-1}). The core was sampled at 1 cm intervals, yielding an average sampling resolution of about six years. Long-term variability in E/P is evident in the high-resolution, isotope-based paleoclimate record. Although there are dramatic fluctuations in the 3.5-millennia record, mean conditions between ~1740 and 920 ^{14}C yr B.P. (A.D. 300-1100) were nevertheless drier than mean conditions before or after (Figure 4.9). The $\delta^{18}O$ peak at 1210 ^{14}C yr B.P. (A.D. 860) in the Punta Laguna section corresponds closely to the dated drought at Chichancanab (1140 ^{14}C yr B.P., or A.D. 920).

The isotope records from Lake Chichancanab and Lake Punta Laguna indicate that the Classic Collapse occurred during a drought episode. The two study lakes lie in the dry, northern part of the Maya lowlands, which was apparently least affected by the cultural decline (Lowe 1985). Depopulation during the ninth century A.D. was more characteristic of sites in the wetter, southern lowlands. If the drought were indeed widespread, one might expect the northern area to have been more impacted. Nevertheless, groundwater is more accessible in the north, because the shallow water table lies just below the land surface. Farther south, groundwater is largely inaccessible and lies at considerable depth.

Compared with settlements on the northern part of the peninsula, sites in the southern lowlands may have been more dependent on surface water in lakes, *bajos*, artificial reservoirs, and *aguadas*. As shallow depressions and reservoirs began to dry, increasing pressure would have been brought to bear on the lands and communities surrounding deeper lakes in central Petén. Archaeological surveys and test excavations in Petén show that many watersheds saw a dramatic population decline by the Terminal Classic (about A.D. 900). Whereas Late Classic Maya population (A.D. 550-800) densities in the drainage basins typically exceeded 200 persons per square kilometer (km^{-2}) (Rice and Rice 1990), densities dropped to less than 100 persons km^{-2} by the Terminal Classic. By the Late Postclassic (A.D. 1500), many watersheds were virtually abandoned, and there is evidence that local soils were stabilized as forests recolonized riparian areas. If drought played a role in the demographic decline in the southern lowlands, then paleoclimatic evidence should be found in Petén lake cores.

Isotope Records from Petén, Guatemala

The early work of Covich and Stuiver (1974) suggested that Lake Chichancanab would be a good study site. Covich's published and unpublished

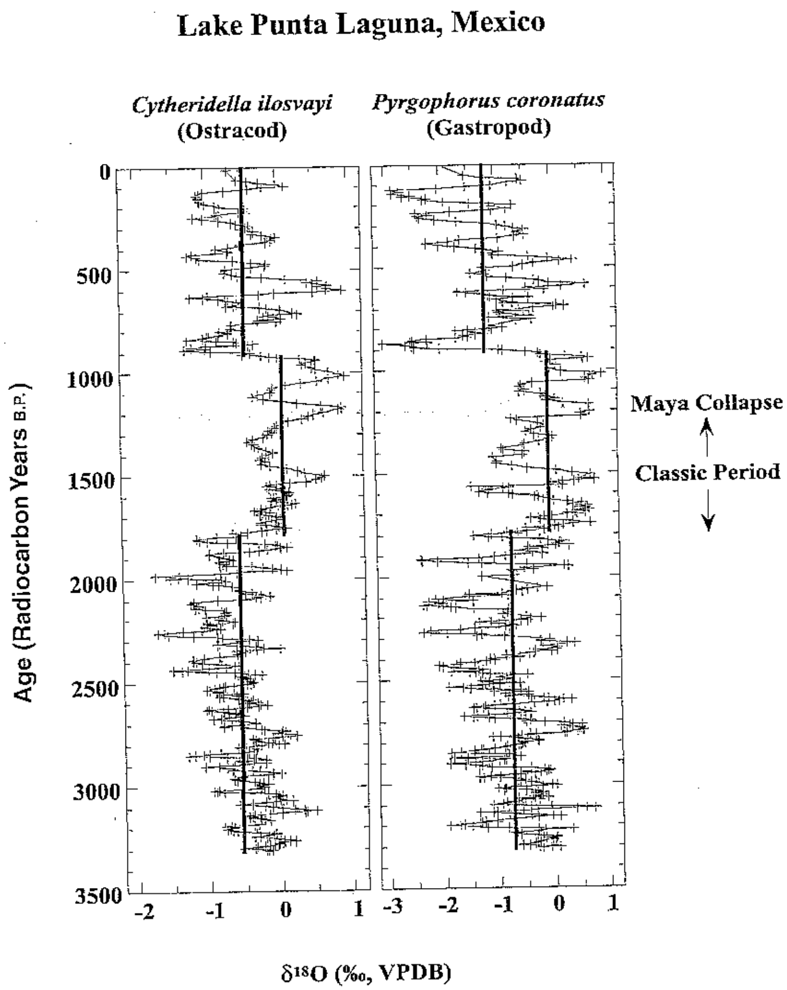


FIGURE 4.9. Oxygen isotope ($\delta^{18}O$) records based on the ostracod *Cytheridella ilosvayi* and gastropod *Pyrgophorus coronatus* from Lake Punta Laguna. Records represent five-point running means and span the last 3500 ^{14}C yr B.P. Note that the Classic Period is relatively dry compared to periods before or after. Also note the decadal to centennial variability in the record that presumably would have created difficulties for Maya agriculturalists. One of the driest episodes is dated to 1180 ^{14}C yr B.P. (A.D. 860), corresponding to the Maya collapse. (*Source:* Reprinted with permission from J.H. Curtis, D.A. Hodell, and M. Brenner 1996, Climate variability on the Yucatán Peninsula (Mexico) during the past 3500 years, and implications for Maya cultural evolution, *Quarternary Research,* 46, pp. 37-47. Copyright 1996 Academic Press.)

work was also used to prioritize study sites in Petén, Guatemala. His study of the subfossil mollusk communities in Lake Petén-Itzá sediments (Covich 1976) suggested that cores from this large water body (99 km^2) would yield plentiful material for isotope work. In 1993, a 5.45 m core was collected in 7.6 m of water from the southern basin of Lake Petén-Itzá, east of Flores (Curtis et al. 1998). A piece of wood at 527 cm depth had a radiocarbon age of 8840 ^{14}C yr B.P. (7930 B.C.), suggesting that each 1 cm sample represented, on average, about 19 years of sediment accumulation. Two isotope profiles were generated using the gastropods *Cochliopina* sp. and *Pyrgophorus* sp., and a composite ostracod record was completed using *Cytheridella ilosvayi* and *Candona* sp. (Figure 4.10). High isotope values in the early Holocene suggest persistent dry conditions following the arid late Pleistocene (Leyden 1984; Leyden et al. 1993; Leyden et al. 1994). The onset of moister conditions appears to have been postponed until about 6800 ^{14}C yr B.P. (5700 B.C.), after which there is a rather steady, 2,000-year decline in $\delta^{18}O$ values.

In contrast to the isotope-based inference for dry early Holocene conditions, the early Holocene pollen assemblage from Lake Petén-Itzá is characterized by high forest taxa which indicate that moist climatic conditions were established by 8500 ^{14}C yr B.P. (7550 B.C.) (Islebe et al. 1996; Curtis et al. 1998). The presence of aquatic microfossils in sediments deposited after about 9000 ^{14}C yr B.P. (8240 B.C.) provides evidence that conditions were certainly moister than in the arid late Pleistocene, when there was no lacustrine deposition at the site. High isotope signatures in the bottom 1.5 m of the core may reflect the fact that, during the early Holocene, a large proportion of lake water volume was lost each year to evaporation. Alternatively, dense vegetation cover in the early Holocene may have influenced watershed hydrology by increasing evaporation and soil water moisture storage (Rosenmeier, Hodell, Brenner, Curtis, Martin, et al. 2002).

After Lake Petén-Itzá attained a new isotopic equilibrium about 5000 ^{14}C yr B.P. (3770 B.C.), it demonstrated little fluctuation about the mean condition. Unlike the drying trend that characterized the last three thousand years of the records from the northern Yucatán Peninsula, Haiti, and northern Venezuela, there is no evidence for drying in the Lake Petén-Itzá isotope record. Although the Lake Petén-Itzá isotope results may reflect climatic stasis during the latter half of the Holocene, it is also possible that the lake was unresponsive to all but the most dramatic changes in E/P due to its extremely large volume. Seismic reflection studies, completed in the summer of 1999 revealed that Lake Petén-Itzá is a cryptodepression with a maximum depth of approximately 160 m (Anselmetti et al. 1999). The deepest part of the lake lies some 50 m below sea level. Even lake stage excursions of several meters represent a gain or loss of only a small fraction of the lake

Lake Petén-Itzá, Guatemala

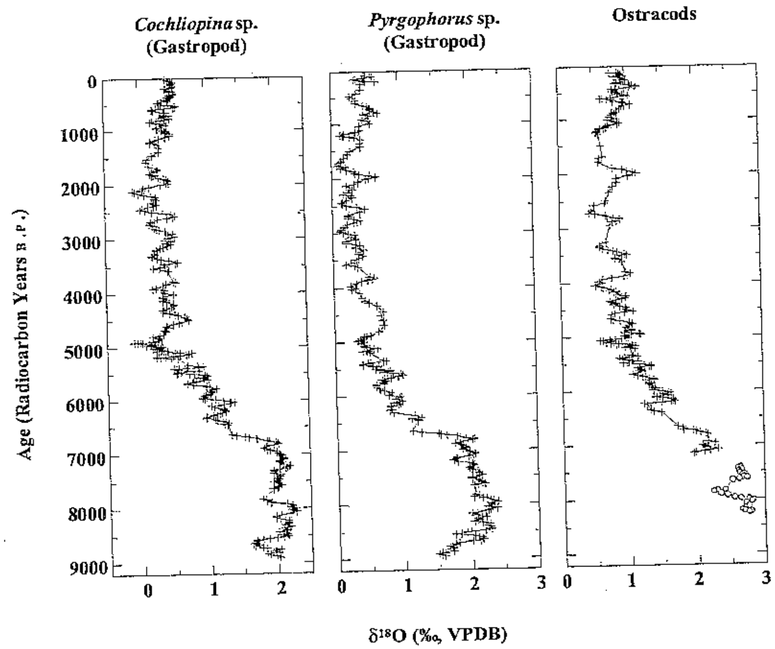


FIGURE 4.10. Oxygen isotope ($\delta^{18}O$) records from the Lake Petén-Itzá core, Petén, Guatemala (Curtis et al. 1998). Values are five-point running means plotted against ^{14}C yr B.P., and are based on gastropods (*Cochliopina* sp. and *Pyrgophorus* sp.) and ostracods (*Cytheridella ilosvayi* [solid line] and *Candona* sp. [open circles]). Note the fairly constant values after ~5000 ^{14}C yr B.P. (3770 B.C.), which probably reflect the unresponsive nature of this very deep (~160 m) and large (99 km^2) lake. (*Source:* Reprinted from J.H. Curtis, M. Brenner, D.A. Hodell, R.A. Balser, G. A. Islebe, and H. Hooghiemstra 1998, A multi-proxy study of Holocene environmental change in the Maya Lowlands of Petén, Guatemala, Figure 10, *Journal of Paleolimnology,* 19, pp. 139-159, with kind permission of Kluwer Academic Publishers.)

volume, and are unlikely to have significantly altered the $\delta^{18}O$ signature of the water column.

Recent research was done at Lake Salpetén (16°58' N and 89°40' W), a small (A = 2.6 km^2) water body that lies just east of Lake Petén-Itzá, near El Remate. The lake is 104 m a.s.l. (above sea level), has a maximum depth of 32 m (Brezonik and Fox 1974), and is similar in its chemistry to Lake Chichancanab. Lake Salpetén is saturated with respect to sulfate (3000 mg·L^{-1}), contains high concentrations of total dissolved solids (4500 mg·L^{-1} TDS), and has a conductivity of about 3600 $\mu S{\cdot}cm^{-1}$. The lake is prone to stage fluctuations; during the regression episode of the late 1990s, gypsum was precipitated on the stumps of drowned trees in the littoral zone. Cores were collected along a transect in Lake Salpetén in 1997 and 1999, at water depths of 9.2, 14.0, 16.3, and 23.2 m (see Figure 4.6). Abundant gastropod shells were evident in the profiles. Radiocarbon dates on cores from the two shallower sites suggested a sedimentation hiatus prior to 910 ^{14}C yr B.P. (A.D. 1150), but its duration could not be determined because some sediment deposited prior to that time may have been lost during the low stand. Discontinuities in the sediment record were confirmed by the seismic reflection profile (see Figure 4.6).

Oxygen isotope ($\delta^{18}O$) profiles from several cores have yielded consistent results (Figure 4.11; Rosenmeier, Hodell, Brenner, Curtis, and Guilderson 2002). Human alteration of watershed hydrology confounds interpretation of $\delta^{18}O$ as a proxy of changing E/P. Lower $\delta^{18}O$ values may represent decreased E/P (wetter climate) and/or greater surface runoff and groundwater inflow to the lake caused by human-induced deforestation. For example, declining $\delta^{18}O$ values between 1300 and 400 B.C. (3050-2340 ^{14}C yr B.P.) coincided with palynologically documented forest loss that may have led to increased inflow. Minimum $\delta^{18}O$ values occurred in the Middle and Late Preclassic Periods between 400 B.C. and A.D. 150 (2340-1860 ^{14}C yr B.P.) (Figure 4.11).

High lake stands are also documented at this time by radiocarbon dates on aquatic gastropods retrieved from pits dug into subaerial soils (i.e., lake sediments) located about 1 m to 7.5 m above the present (1999) lake stage (Rosenmeier, Hodell, Brenner, Curtis, Martin 2002). Following the period of minimum $\delta^{18}O$ values (400 B.C. to A.D. 150, or 2340-1860 ^{14}C yr B.P.), the signal increases in a series of steps at A.D. 150, 550, 850, and 1300 (1860, 1540, 1190, and 660 ^{14}C yr B.P.). These $\delta^{18}O$ increases may reflect a series of aridity increases and/or decreased hydrologic inputs as a consequence of forest recovery associated with population declines (Rosenmeier, Hodell, Brenner, Curtis, and Guilderson 2002). Within error associated with the sediment chronology, most of these steps in the $\delta^{18}O$ signal correspond with discontinuities in Maya cultural evolution: Preclassic abandonment (A.D. 150-200), Maya Hiatus (A.D. 500-550), and Terminal Classic Collapse (A.D. 800-950). Whether this coincidence represents a response of culture to climate change, or a response of environment to human disturbance is not known.

Oxygen isotope studies from lakes in Petén have thus far yielded ambiguous results. The $\delta^{18}O$ records from Lake Petén-Itzá indicate little change for the past five thousand years (see Figure 4.10), whereas those from Lake Salpetén show a series of stepped changes in the $\delta^{18}O$ equilibrium value of lake water. Contradictory results among Petén lakes may be due to anthropogenic vegetation changes that altered hydrologic budgets of individual

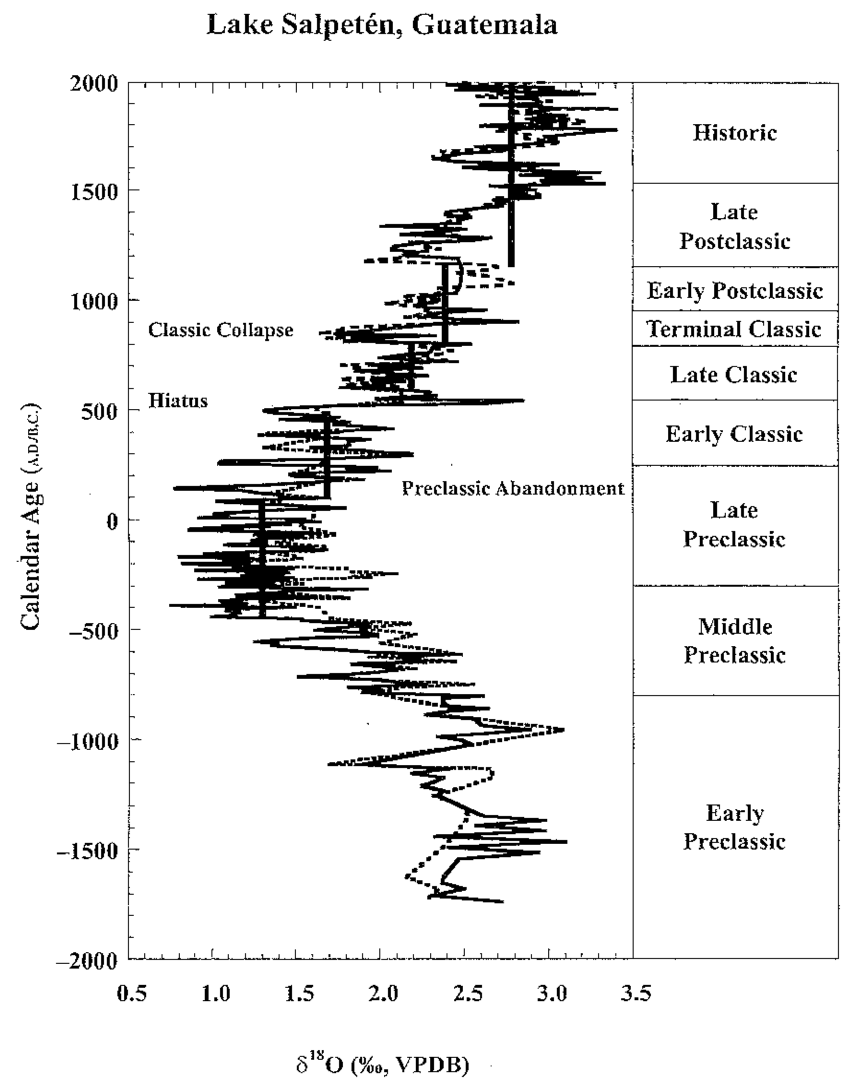


FIGURE 4.11. Oxygen isotope ($\delta^{18}O$) composition of ostracod valves *(Physocypria globula)* from cores taken in Lake Salpetén, Petén, Guatemala (Rosenmeier, Hodell, Brenner, Curtis, and Guilderson 2002). Note the shifts in the heavy black lines that delineate mean $\delta^{18}O$ values for each period. Isotope results are compared to major subdivisions of Maya cultural evolution.

lakes and produced different $\delta^{18}O$ patterns. In contrast, lakes in northern Yucatán yield consistent $\delta^{18}O$ patterns reflecting climatic variations (E/P). Hydrology in Petén watersheds may have been more subject to human impacts because the southern lowlands receive more rainfall, forest stature is greater, and local topography is steeper.

Objectives of Future Paleoclimate Research in the Maya Lowlands

Oxygen isotope records from Lake Chichancanab and Lake Punta Laguna have provided tantalizing evidence of a link between the Classic Maya Collapse and declining moisture availability in the ninth and tenth centuries A.D. Lake Petén-Itzá and Lake Salpetén in the Petén lowlands have yielded ambiguous results, most probably because of human disturbance of individual watershed hydrology. High-resolution paleoclimate records from other sites throughout the Maya lowlands are required to further address the hypothetical causal relationship between drought and cultural demise. The next challenge is to identify additional, appropriate lakes for isotopic study.

The best paleoclimate records from the southern lowlands will likely be found in lakes whose drainage basins were not densely settled, and those with low watershed/lake ratios. Archaeologists can help identify these optimal study lakes. Deciphering the paleoclimatic record from the Maya lowlands will involve collaboration between archaeologists and paleoenviromental scientists. A better understanding of the long-term relationship between climate and Maya culture will not only have bearing on the interpretation of the archaeological record, but should be informative about the future prospects for sustainable agriculture in a region that is once again becoming densely populated.

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