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A 4000-Year Lacustrine Record of Environmental Change in the Southern Maya Lowlands, Petén, Guatemala

Michael F. Rosenmeier,[1] David A. Hodell, Mark Brenner, and Jason H. Curtis

Department of Geological Sciences, University of Florida, P.O. Box 112120, Gainesville, Florida 32611

and

Thomas P. Guilderson

Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory 7000 East Avenue, Livermore, California 94551

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A 4000-yr sediment core record from Lake Salpetén, Guatemala, provides evidence for Maya-induced forest clearance and consequent soil erosion between ∼1700 cal yr B.C. and 850 cal yr A.D. Radiocarbon ages of wood, seeds, and charcoal support an age-depth model with average errors of ±110 cal yr. Relatively low carbonate $\delta^{18}O$ values between 1300 and 400 cal yr B.C. coincide with pollen evidence for forest loss, consistent with increased surface and groundwater flow to the lake. Minimum $\delta^{18}O$ values between 400 cal yr B.C. and 150 cal yr A.D. suggest a high lake level, as do ^{14}C-dated aquatic gastropods as much as 7.5 m above the present lake stage. High lake levels resulted from reduced evaporation-to-precipitation ratios, increased hydrologic input caused by anthropogenic deforestation, or both. The Preclassic abandonment (150 A.D.) and Early Classic/Late Classic boundary (550 A.D.) are marked by relatively high $\delta^{18}O$ values indicating reduced lake levels. Oxygen isotope composition increased further coincident with the Terminal Classic Maya demographic decline between 800 and 900 A.D. This period of high $\delta^{18}O$ may have been caused by the greater aridity that has been documented in northern Yucatán lakes or by decreased hydrologic input to the lake as a consequence of forest recovery. Reduced soil erosion after 850 cal yr A.D. coincided with the Terminal Classic Maya demographic decline and permitted forest recovery and resumption of organic sedimentation.

Key Words: **climate change; deforestation; Guatemala; Holocene; human disturbance; lake sediments; Maya; oxygen isotopes; sediment chemistry.**

INTRODUCTION

Pollen records from lowland Petén, Guatemala, document forest clearance by Maya populations beginning in the first millennium B.C. (Tsukada, 1966; Deevey, 1978; Vaughan *et al.*, 1985; Leyden, 1987; Islebe *et al.*, 1996). During the later Preclassic and Classic periods (∼250 B.C.–850 A.D.), as Maya settlement and agriculture expanded, much of Petén was deforested. Widespread cultivation accelerated soil erosion, which resulted in the accumulation of thick clay-rich deposits in many Petén lakes (Cowgill and Hutchinson, 1966; Deevey *et al.*, 1979; Brenner, 1983; Rice *et al.*, 1985; Binford *et al.*, 1987). This inorganic sediment accumulation, termed the *Maya clay*, was deposited over a period of ∼2600 yr and is equated with occupation of Petén watersheds (Binford *et al.*, 1987). Forest recovery and soil stabilization have been attributed to population decline following the Classic Maya period and further depopulation after European contact (Brenner *et al.*, 1990).

The chronological framework of Maya impact on terrestrial and lacustrine environments remains imprecise. The hard-water lake effect (Deevey and Stuiver, 1964) and redeposition of carbonate-rich basin soils of unknown isotopic composition (Vaughan *et al.*, 1985) invalidate radiocarbon dates on bulk sediments from Petén lakes. Chronologies have therefore relied largely on correlation between pollen assemblages and the archaeological prehistory of the region (Vaughan *et al.*, 1985; Brenner, 1994). Moreover, late Holocene climate reconstructions from pollen analyses are confounded by human-mediated deforestation in Petén (Deevey, 1978; Deevey *et al.*, 1979; Vaughan *et al.*, 1985; Leyden, 1987; Islebe *et al.*, 1996).

In this study, a composite sediment profile from Lake Salpetén provides the first high-resolution, accelerator mass spectrometry (AMS), ^{14}C-dated record of late Holocene environmental change in lowland Petén, Guatemala. Stratigraphic geochemical variations from this profile reflect changing hydrologic balance and material transfer to the lake resulting from both human disturbance and climate change. Shifts in the water balance of the lake were inferred from the oxygen isotopic composition ($\delta^{18}O$) of biogenic carbonate. Material transfer to the lake was inferred from sediment composition and accumulation, and watershed vegetation changes were reconstructed from fossil pollen assemblages. The Lake Salpetén record was compared with a profile from nearby Lake Petén Itzá (Curtis *et al.*, 1998) and sediment

[1] To whom correspondence should be addressed.

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records from the northern Yucatán Peninsula (Hodell *et al.*, 1995; Curtis *et al.*, 1996).

STUDY SITE

The Department of Petén, which occupies the northern third of Guatemala (Fig. 1a), is characterized principally by low-lying karsted limestone of Cretaceous and Tertiary age (Vinson, 1962). Most of the landscape has well-drained forest soils and tropical semideciduous and evergreen vegetation (Lundell, 1937). Terrain varies between 100 and 300 m above sea level. Groundwater lies well below the land surface. Surface waters, however, are perched, resulting in numerous lakes and seasonally inundated topographic depressions (Deevey *et al.*, 1979). The lake district contains closed basins along east–west *en echelon* faults at 17°N latitude (Fig. 1b). Lake Salpetén is a small, closed-basin lake that lies 104 m above sea level and has a maximum depth of 32 m (Brezonik and Fox, 1974) (Fig. 1c). Lake Salpetén is sulfate-rich (3000 mg L^{-1}) and brackish (4500 mg L^{-1} TDS), with a pH of 8.5 and a conductivity of $\sim$3600 μS cm^{-1}.

Rainfall in Petén varies spatially and interannually from 900 to 2500 mm, with a regional annual mean of 1600 mm (Deevey, 1978). Intense convection associated with the northward migration of the intertropical convergence zone (ITCZ) and the Azores–Bermuda high-pressure system produces heavy rains between June and October when the trade winds weaken and sea-surface temperatures warm in the Atlantic between 10° and 20°N (Hastenrath, 1984). Conversely, dry conditions develop in November and December as the ITCZ and the Azores–Bermuda high move equatorward and strong trade winds become predominant. A pronounced dry season prevails from January to May.

MATERIALS AND METHODS

In May 1980, a 15-m sediment core was obtained from the deep basin of Lake Salpetén (Deevey *et al.*, 1983) with a 45.7-cm split-spoon sampler. This core was designated Sal 80-1. Topmost sediments, above 1.6 m, were not recovered. In August 1999, two additional sediment cores were recovered from Lake Salpetén in water depths of 16.3 (SP2-19-VIII-99) and 23.2 m (SP1-17-VIII-99) (Fig. 1c). Surface sediments were collected with a sediment–water interface corer (Fisher *et al.*, 1992). Deeper sections were taken in 1.0-m segments with a modified square-rod piston corer (Wright *et al.*, 1984). Interface cores were sectioned in the field at 1.0-cm intervals. Square-rod core sections were extruded and sampled at 1.0-cm intervals in the laboratory. Archived core

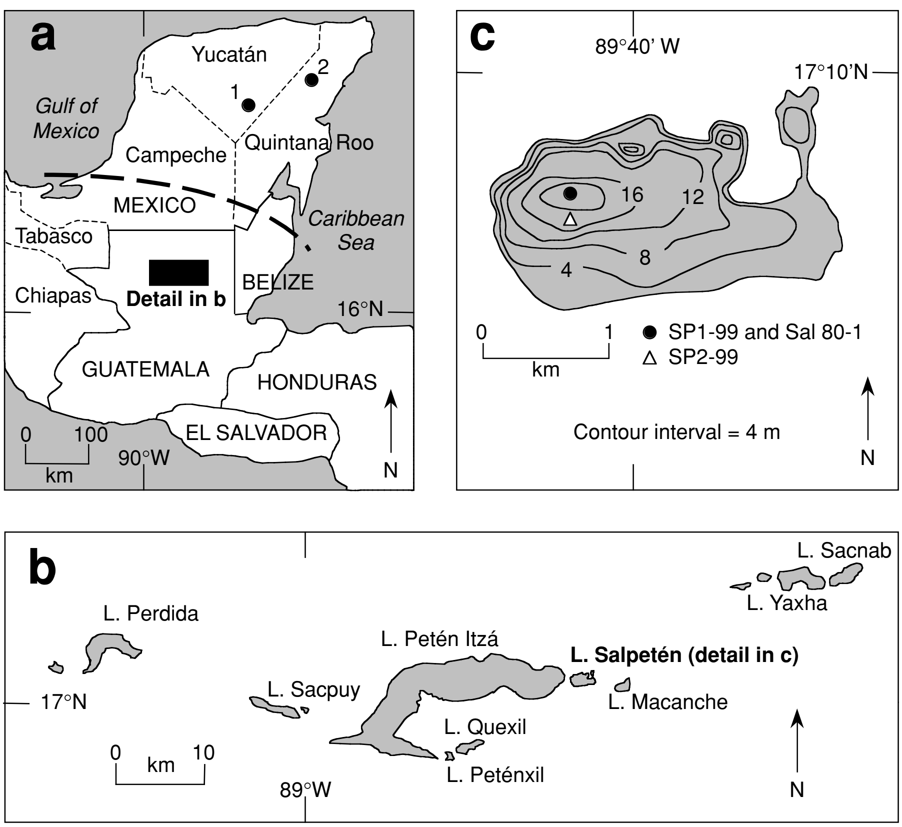

FIG. 1. (a) Map of the Yucatán Peninsula showing lake study sites in northern Guatemala (black rectangle) and Lakes Chichancanab (1) and Punta Laguna (2). Heavy dashed line indicates the division between the northern and southern Maya lowlands. (b) Petén lake district in Guatemala. (c) Bathymetry and sediment core sites in Lake Salpetén.

TABLE 1
Accelerator Mass Spectrometry Radiocarbon Dates for Samples from Lake Salpetén, Petén, Guatemala

Core	Material	Depth (cm)	Laboratory number	Age (^{14}C yr B.P.)	Intercept age[a] (cal yr A.D. or B.C.)	Age range at 2 SD (cal yr A.D. or B.C.)
SP1-99	Wood	24	CAMS 62943	180 ± 50	1670 A.D.	1640–1950 A.D.
SP2-99	Wood	59	CAMS 62939	400 ± 40	1460 A.D.	1430–1610 A.D.
SP2-99	Wood	108	CAMS 64858	1370 ± 50	660 A.D.	600–730 A.D.
SP2-99	Charcoal	114	CAMS 64859	1380 ± 140[b]	660 A.D.	400–960 A.D.
SP1-99	Wood	123	CAMS 60079	1320 ± 40	690 A.D.	650–770 A.D.
SP2-99	Charcoal	125	CAMS 64860	1690 ± 140[b]	370 A.D.	40–640 A.D.
SP2-99	Seed	153	CAMS 62940	1850 ± 50	170 A.D.	60–290 A.D.
Sal 80-1	Charcoal	174	CAMS 65388	2200 ± 70[b]	220 B.C.	400–70 B.C.
Sal 80-1	Charcoal	221	CAMS 65389	2200 ± 60[b]	220 B.C.	390–90 B.C.
SP2-99	Wood	231	CAMS 60749	2090 ± 40	100 B.C.	190–10 B.C.
SP2-99	Seed	236	CAMS 62941	2200 ± 50	220 B.C.	380–130 B.C.
Sal 80-1	Charcoal	271	CAMS 64861	2430 ± 50	460 B.C.	760–400 B.C.
Sal 80-1	Charcoal	296	CAMS 65390	2500 ± 60	750 B.C.	800–420 B.C.
SP2-99	Wood	309	CAMS 60078	2820 ± 40	960 B.C.	1070–870 B.C.
Sal 80-1	Charcoal	323	CAMS 64978	2990 ± 190[b]	1240 B.C.	1660–800 B.C.
SP2-99	Wood	339	CAMS 62942	3240 ± 40	1510 B.C.	1590–1420 B.C.
Sal 80-1	Charcoal	344	CAMS 65797	3160 ± 80[b]	1420 B.C.	1590–1260 B.C.

[a] Intercept age fit by the fourth-order polynomial in Figure 2.
[b] Target size less than 0.1 mg C.

sections from Sal 80-1 were sampled at 5.0-cm intervals and reexamined in May 2000.

Radiocarbon ages from Lake Salpetén were determined by AMS at Lawrence Livermore National Laboratories. All samples consisted of terrestrial organic matter (wood, seeds, and charcoal). Calibrated ages were calculated with the INTCAL98 calibration with a 100-yr moving average of the tree-ring calibration data set (Stuiver and Reimer, 1998; Stuiver *et al.*, 1998).

Oxygen isotopic ratios on valves of the ostracod *Physocypria globula* were measured. Adult ostracod valves were soaked in 15% H_2O_2, cleaned ultrasonically in deionized water, and rinsed with methanol before drying. Aggregate samples of ∼40 ostracod carapaces were measured from each sediment sample. Samples were reacted in 100% orthophosphoric acid at 70°C using a Finnigan MAT Kiel III automated preparation system. Isotopic ratios of purified CO_2 gas were measured online with a Finnigan MAT 252 mass spectrometer and compared to an internal gas standard. Isotopic values are expressed in conventional delta (δ) notation as the per mil (‰) deviation from Vienna PeeDee Belemnite. Precision for $\delta^{18}O$ samples was ±0.09‰.

Inorganic carbon (IC) was measured by coulometric titration (Engleman *et al.*, 1985) with a UIC/Coulometrics Model 5011 coulometer and coupled UIC 5240-TIC carbonate autosampler. Analytical precision is about ±0.6% based on repeated analysis of reagent-grade calcium carbonate. Total carbon (TC) was measured with a Carlo Erba NA 1500 CNS elemental analyzer with autosampler. Organic carbon (OC) was estimated by subtraction of IC from TC.

RESULTS

The high degree of stratigraphic correlation between Lake Salpetén sediment cores allowed construction of a composite depth series comprising 17 AMS ^{14}C dates (Table 1). Age–depth values for the last 4000 calibrated radiocarbon years were fit by a fourth-order polynomial (Fig. 2). Residual errors associated with the construction of the composite depth series average ±110 cal yr.

Paleoenvironmental proxies from Lake Salpetén are plotted and discussed relative to calendar age (cal yr A.D. or B.C.). Calcium carbonate concentrations in Lake Salpetén sediments increased gradually from 2000 cal yr B.C. to maximum values (>40%) between 700 cal yr B.C. and 850 cal yr A.D. (Fig. 3d). Organic carbon concentration was relatively high, typically >10%, before 1700 cal yr B.C. but decreased by 900 cal yr B.C. (Fig. 3c). Organic carbon concentrations remained low (<5%) between 900 cal yr B.C. and 850 cal yr A.D. Organic carbon concentration increased abruptly after 850 cal yr A.D. and again after 1400 cal yr A.D. but declined over the last 300 yr. After 850 cal yr A.D., $CaCO_3$ content was highly variable.

The $\delta^{18}O$ of biogenic carbonate decreased between 1300 and 400 cal yr B.C., from ∼2.7‰ to less than 1.5‰ (Fig. 3e). Minimum $\delta^{18}O$ values, averaging 1.3‰, occurred between 400 cal yr B.C. and 150 cal yr A.D. Oxygen isotopic values increased abruptly between 150 and 200 cal yr A.D. and averaged ∼1.7‰ between 200 and 500 cal yr A.D. Between 500 and 550 cal yr A.D., $\delta^{18}O$ increased by as much as 1.5‰ and averaged 2.2‰ between 550 and 850 cal yr A.D. Average $\delta^{18}O$ increased to

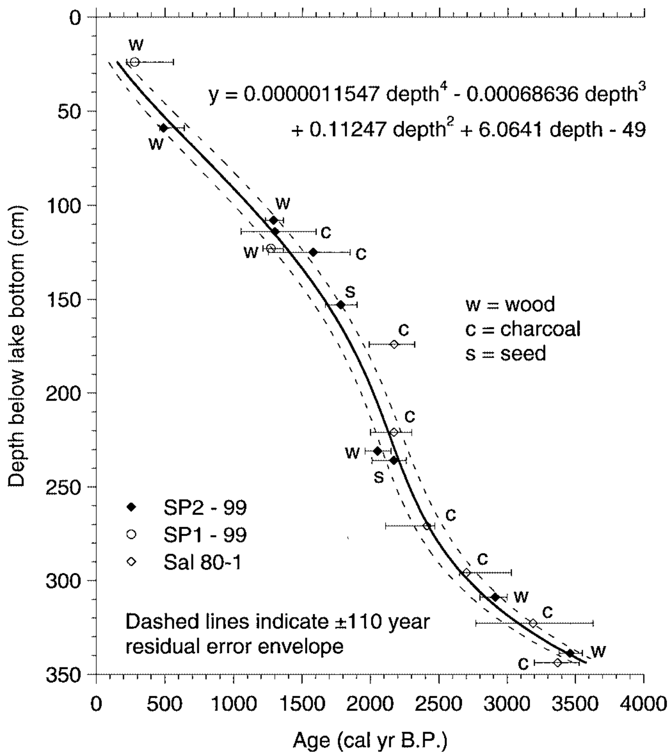

FIG. 2. Ages of plant fragments versus composite depth in cores from Lake Salpetén. Age–depth values were fit by a fourth-order polynomial (black line). Error bars delineate calibrated age range at two standard deviations; diamonds and circles are intercept ages used in polynomial.

2.4‰ after 900 cal yr A.D. and increased to 2.8‰ between 1300 cal yr A.D. and the present.

INTERPRETATION OF PROXY RECORDS

Variations in the oxygen isotopic ratio ($^{18}O/^{16}O$) of lacustrine carbonate are caused by changes in the temperature of carbonate precipitation and shifts in the $^{18}O/^{16}O$ of lake water from which the carbonate precipitates (Craig, 1965). In tropical lakes with negligible surface outflow, variations in the $^{18}O/^{16}O$ of lake water are dominated by changes in the rate of evaporation relative to the combined inputs of precipitation and surface and groundwater inflow (Fontes and Gonfiantini, 1967; Talbot, 1990; Curtis *et al.*, 1996; Rosenmeier *et al.*, in press). Extended periods of enhanced evaporation or reduced precipitation and surface and groundwater inflow raise $^{18}O/^{16}O$ ratios of lake water and precipitated carbonate. Conversely, increased surface and groundwater inflow or precipitation produce lower $^{18}O/^{16}O$ ratios.

In addition to altering lake hydrology, changes in surface flow and precipitation alter the transfer of dissolved and particulate material to a water body. Changes in material flux to the sediments result primarily from variations in material output from the catchment, sometimes due to land-use changes. For example, soil changes resulting from deforestation and fires produce variations in the flux of particulate matter to a lake (Walling, 1988; Dearing, 1991). Forest clearance and high rainfall accelerate alluviation and colluviation, and increases in sediment accumulation may reflect intensified surface flow and erosion within a watershed. Catchment deforestation, agricultural production, field abandonment, and subsequent plant succession are reflected by pollen deposited in lake sediments. Deforestation alters the transport of soil nutrients and organic and inorganic matter to the lake, which is reflected in the sediment lithologic composition (Binford *et al.*, 1987).

DISCUSSION

Human deforestation of the Lake Salpetén watershed is implied in the pollen record by reduction of high forest taxa, particularly pollen of the arboreal family Moraceae, and by expansion of disturbance taxa beginning ~1700 cal yr B.C. (Figs. 3a, 3b). For the same time interval, geochemical records from Lake Salpetén imply catchment soil erosion. Organic carbon content decreased after 1700 cal yr B.C. and inorganic materials dominated subsequent sediment accumulation (Figs. 3c, 3d). Forest clearance by the rapidly expanding Maya population destabilized water-shed soils, thereby accelerating their erosion, transport, and redeposition. Accelerated erosion and oxidation of exposed catchment soils rapidly depleted organic matter (forest litter and surface soil horizons) and enhanced delivery of soil-derived carbonate to the lake, resulting in the deposition of sediments with low organic content.

Archaeologically documented colonization of the Salpetén watershed dates to Middle Preclassic times, beginning ~1000 B.C. (Rice and Rice, 1990). Rapid accumulation of inorganic sediments in Lake Salpetén began before dense occupation of the basin in the Late Classic (600 to 800 A.D.). Catchment response to vegetation removal may not have been linearly related to population expansion. For example, initial forest clearance and soil destabilization may be attributed to pioneer settlement that is archaeologically unrecognized. Intensified basin occupation and deposition of the Maya clay correlates with increased disturbance taxa and the presence of maize (*Zea*) after 700 cal yr B.C. (Fig. 3b). Gradual decline of lowland forest and subsequent soil destabilization might also reflect climatic drying prior to human disturbance. However, forest loss as a result of increased aridity is incompatible with the coincident decrease in the $\delta^{18}O$ of biogenic carbonate in Lake Salpetén, beginning ~1300 cal yr B.C. (Fig. 3e).

The $\delta^{18}O$ values in Lake Salpetén decreased gradually from 1300 to 400 cal yr B.C. (Fig. 3e). Minimum $\delta^{18}O$ values, suggesting high water levels, occurred between 400 cal yr B.C. and 150 cal yr A.D., corresponding to the Middle and Late Preclassic Maya Periods. Evidence for high lake levels is also provided by 12 AMS ^{14}C dates on aquatic gastropods from soil pits (i.e., subaerial lacustrine deposits) ~1.0 to 7.5 m above modern lake stage (Rosenmeier *et al.*, in press). Several dates precede human disturbance of the watershed but most indicate high lake levels

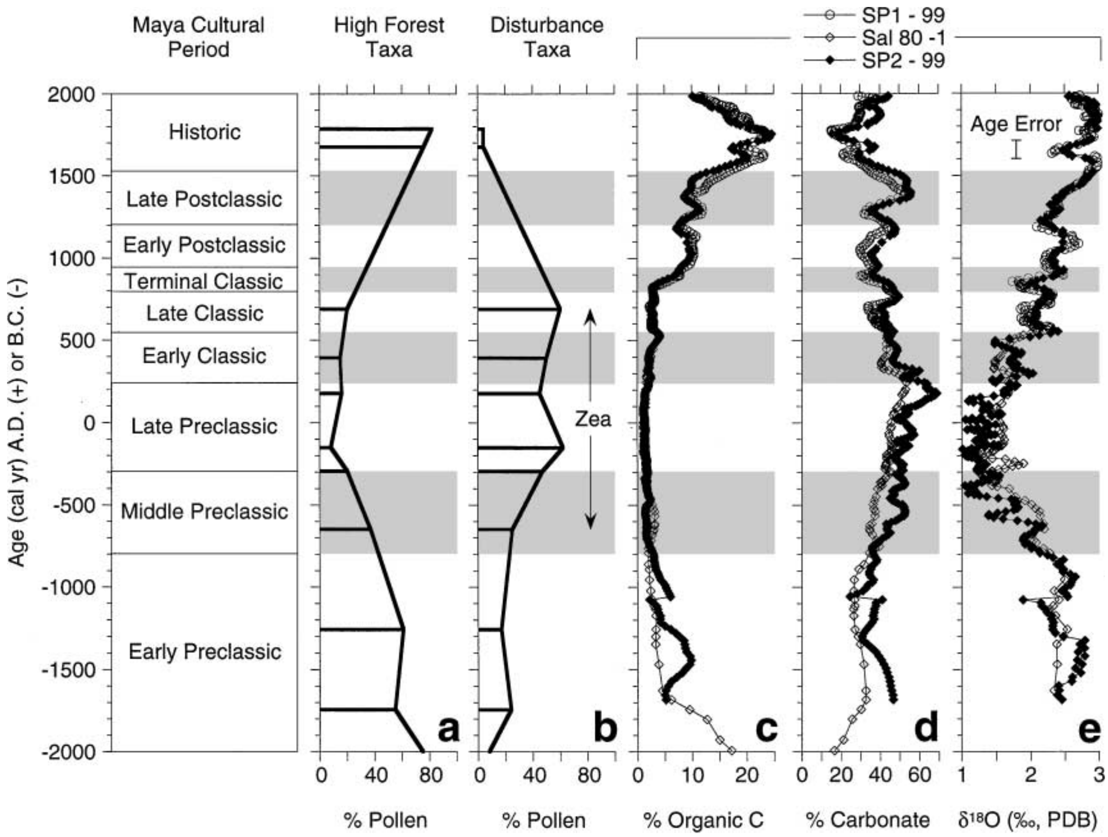

FIG. 3. (a) Relative abundance of high forest pollen taxa and (b) pollen taxa typical of disturbed lands from Lake Salpetén core Sal 80-1 (Leyden, 1987). High forest taxa include species of the families Moraceae (e.g., *Brosimum*, *Cecropia*, *Chlorophora*, and *Ficus*) and Urticaceae. Disturbance taxa include Amaranthaceae, Ambrosia, Compositae, Cyperaceae, and Gramineae. (c) Organic carbon content, (d) $CaCO_3$ concentration, and (e) oxygen isotopic composition versus age and Maya cultural periods (Rice and Rice, 1990). Geochemical data smoothed with a 5-point running mean to illustrate long-term trends. Error bar (plotted in e) shows the average residual error in the age–depth model (Fig. 2).

during Maya occupation, from the Middle Preclassic through the late Classic. These ages coincide roughly with δ^{18}O minima in the core and support the inference of high lake stage. Forest removal may have decreased evapotranspiration and soil moisture storage in the watershed, thereby increasing catchment water yield (Bosch and Hewlett, 1982; Stednick, 1996) and transport of isotopically light surface and ground waters to the lake. Increased delivery of meteoric waters would increase lake volume, decrease the proportion of the hydrologic budget lost to evaporation, and decrease lake water δ^{18}O. Alternatively, high water levels and δ^{18}O minima may reflect increased precipitation and reduced evaporation beginning after 1300 cal yr B.C.

In contrast, mean δ^{18}O values from nearby Lake Petén Itzá have been nearly constant over the past ∼4000 yr, fluctuating by only ∼0.5‰ (Fig. 4b). Low variability in the late Holocene Petén Itzá record may simply reflect the large volume and long residence time of the lake that make it relatively insensitive to either climatic or land-use changes. Moreover, the Maya clay is not documented as a distinct stratigraphic unit in the Petén Itzá core (Curtis *et al.*, 1998) although accelerated forest clearance is palynologically documented by 1000 cal yr B.C. (Islebe *et al.*, 1996). Lake Petén Itzá is substantially larger than Salpetén and the basin may be effectively buffered from watershed disturbances. Furthermore, erosional contributions to the coring site were probably low, as the site lies nearly 1 km from the steep north shore (Curtis *et al.*, 1998).

Oxygen isotopic records from Petén lakes may also have been influenced by changes in the isotopic composition of precipitation and by changes in air temperature. Trace element (Mg and Sr) concentrations in ostracod shells from Lakes Salpetén and Petén Itzá parallel the δ^{18}O records, suggesting that δ^{18}O records reflect basin hydrology rather than changes in the isotopic composition of rainfall (Rosenmeier *et al.*, in press). Furthermore, covariance in both δ^{18}O and Mg eliminate temperature as a control on the oxygen isotope record.

Reduced soil erosion in the Salpetén watershed after 850 cal yr A.D. is inferred from increased organic carbon deposition that coincides with palynological evidence of reforestation (Figs. 3a–3c). This environmental change coincides with the archaeologically documented decline of Maya population between 800 and 900 A.D. (Lowe, 1985). Reduced population in Petén watersheds following the Terminal Classic would

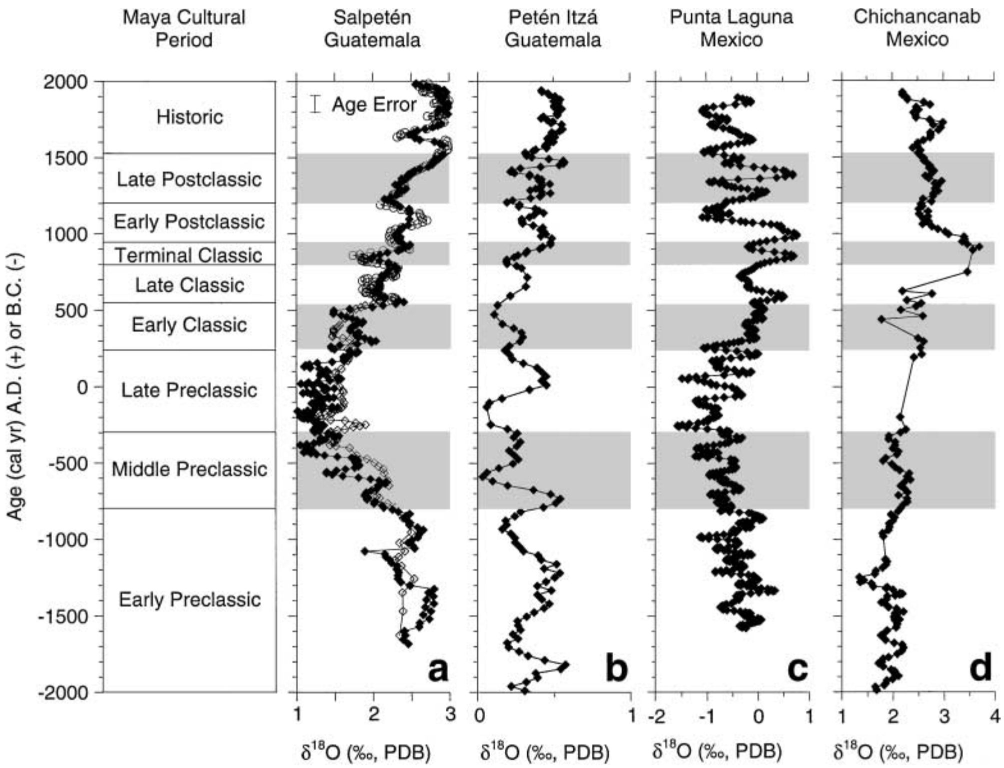

FIG. 4. Oxygen isotope records from (a) Lake Salpetén, (b) Lake Petén Itzá (Curtis *et al.*, 1998), (c) Punta Laguna (Curtis *et al.*, 1996), and (d) Lake Chichancanab (Hodell *et al.*, 1995) versus age and Maya cultural periods (Rice and Rice, 1990). Error bar (plotted in a) shows the average residual error in the age–depth model (Fig. 2).

have permitted recovery of some lowland forest vegetation and soil stabilization (Deevey *et al.*, 1979; Rice *et al.*, 1985; Binford *et al.*, 1987). Postclassic population densities in Petén, however, were probably sufficient to keep the region partially deforested (Brenner, 1994). The increase in organic carbon at the Late Postclassic/Historic boundary (Fig. 3c) delineates the upper temporal boundary of the Maya clay and is attributed to further decline of Maya populations following European contact.

Forest regeneration after 850 cal yr A.D. may have altered the hydrologic budget of Lake Salpetén by reducing surface and groundwater inflow, thereby causing an increase in δ^{18}O during the Postclassic and Historic Periods (Fig. 3e). This interpretation is at odds with evidence for increased δ^{18}O values after 150 cal yr A.D., nearly 700 yr prior to Terminal Classic depopulation of the catchment and forest recovery. Late Holocene climate changes may have disrupted regional evaporation and precipitation patterns, altering the hydrologic budget of Lake Salpetén. Strict climatic interpretation of the δ^{18}O record indicates relatively moist conditions during Middle and Late Preclassic settlement expansion, from 400 cal yr B.C. to 150 cal yr A.D. (Fig. 3e). Abrupt δ^{18}O increases centered at 150, 550, and 850 cal yr A.D. document stepwise climatic drying throughout the Classic and Early Postclassic. Maximum δ^{18}O values after 1300 cal yr A.D. reflect the driest conditions of the last ~4000 yr, during the Late Postclassic and Historic Period.

Temporal Correlations between Environmental and Cultural Changes

The paleoclimate history inferred from the Lake Salpetén δ^{18}O record indicates that shifts in moisture availability coincided with several discontinuities in Maya culture. The termination of Preclassic culture at the site of El Mirador, northern Guatemala (Dahlin, 1983), correlates temporally with inferred relative aridity between 150 and 200 cal yr A.D. (Fig. 3e). Moreover, the δ^{18}O increase between 500 and 550 cal yr A.D. corresponds to the boundary between the Early Classic and Late Classic periods (Fig. 3e). This "Maya Hiatus" represents a period of social upheaval and localized population declines (Gill, 2000) that may have been associated with climatic drying. Inferred aridity between 850 and 900 cal yr A.D. (Fig. 3e) occurred concomitant with the Terminal Classic Maya population decline between 800 and 900 A.D. (Lowe, 1985). Climate remained relatively dry throughout the Early Postclassic and δ^{18}O increased further during the Late Postclassic period.

Rather than reflecting climate changes, shifts in the Salpetén δ^{18}O record may simply reflect altered basin hydrology (varied surface runoff and groundwater inflow) related to forest clearance. Stepwise δ^{18}O increases after 150 cal yr A.D. may reflect periodic relaxation of catchment disturbance, temporary forest regrowth and soil stabilization, and establishment of new lake level and δ^{18}O steady-state conditions. Increased δ^{18}O at the Early Classic/Late Classic boundary coincides with a distinct organic carbon peak that may indicate reduced catchment disturbance and consequent soil stabilization.

Comparison with Northern Yucatán Sediment Records

The changing water level record from Lake Salpetén both contrasts with and complements late Holocene paleoenvironmental histories inferred from northern Yucatán lake cores. Lake Salpetén levels were relatively high during the Middle and Late Preclassic periods (Fig. 4a). Oxygen isotopic values increased abruptly at the boundary between the Preclassic and Classic and continued to increase stepwise throughout the Classic period. Similarly, the record from Lake Punta Laguna (Curtis *et al.*, 1996) displayed relatively low, but variable, δ^{18}O throughout the Preclassic period (Fig. 4c). Oxygen isotopic values in Punta Laguna increased abruptly after 250 A.D. and remained high throughout the Classic Period, suggesting dry conditions. In Lake Chichancanab, δ^{18}O increased abruptly at 800 A.D., marking the beginning of a 200-yr drought in the Terminal Classic and earliest Postclassic (Fig. 4d) (Hodell *et al.*, 1995, 2001). Oxygen isotope values in Salpetén increased throughout the Postclassic and Historic Periods, whereas δ^{18}O values in Punta Laguna and Chichancanab decreased after 1100 A.D.

Neither the Chichancanab nor Punta Laguna watersheds is believed to have been densely settled by the Maya (Castillo and Peraza, 1991; Leyden *et al.*, 1998). Given this minimal human impact, the δ^{18}O records from the northern Yucatán lakes probably reflect climatic variations (evaporation and precipitation) accurately. Furthermore, differences between the northern and southern lowlands with respect to vegetation stature and annual precipitation also influenced the relative importance of vegetation in controlling lacustrine hydrologic budgets. For example, clearing of the scrub vegetation that predominates in the Chichancanab catchment would increase water yield only about 5% relative to the increase associated with clearing lowland forest (Sahin and Hall, 1996). Even if the drying events recorded in northern Lakes Chichancanab and Punta Laguna were regional in extent, human-mediated changes in catchment hydrology in Petén may have obscured the climate signal. The effects of Maya deforestation therefore confound paleoclimatic interpretation of late Holocene δ^{18}O records from Petén lakes.

CONCLUSIONS

Palynological and geochemical records from Lake Salpetén indicate Maya-induced forest clearance and consequent soil erosion beginning ~1700 cal yr B.C. Reduced soil erosion after 850 cal yr A.D. coincided with the Terminal Classic Maya demographic decline. Forest recovery and increased organic carbon sedimentation after 1400 cal yr A.D. correlate with further depopulation of the watershed. Decreased δ^{18}O of biogenic carbonate between 1300 and 400 cal yr B.C. coincided with palynological evidence of forest loss. Low δ^{18}O and inferred high lake levels may have resulted from human disturbance, natural climate change, or both. A strictly climatic interpretation suggests higher precipitation during the expansion of Middle and Late Preclassic Maya settlement. Alternatively, minimum δ^{18}O values between 400 cal yr B.C. and 150 cal yr A.D. may have been a consequence of increased surface runoff and groundwater inflow to the lake related to Maya deforestation of the watershed. During the period of Preclassic abandonment (150 A.D.) and the Early Classic/Late Classic boundary (550 A.D.) δ^{18}O values increased as a consequence of decreased precipitation or temporary forest recovery. Similarly, δ^{18}O values increased during the Terminal Classic Maya demographic decline. Anthropogenic deforestation that alters lake hydrologic budgets may confound paleoclimatic inferences based on the δ^{18}O of biogenic carbonate.

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