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Influence of vegetation change on watershed hydrology: implications for paleoclimatic interpretation of lacustrine $\delta^{18}O$ records

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Abstract

Stratigraphic shifts in the oxygen isotopic ($\delta^{18}O$) and trace element (Mg and Sr) composition of biogenic carbonate from tropical lake sediment cores are often interpreted as a proxy record of the changing relation between evaporation and precipitation (E/P). Holocene $\delta^{18}O$ and Mg and Sr records from Lakes Salpetén and Petén Itzá, Guatemala were apparently affected by drainage basin vegetation changes that influenced watershed hydrology, thereby confounding paleoclimatic interpretations. Oxygen isotope values and trace element concentrations in the two lowland lakes were greatest between ~ 9000 and 6800 ^{14}C-yr BP, suggesting relatively high E/P, but pollen data indicate moist conditions and extensive forest cover in the early Holocene. The discrepancy between pollen- and geochemically-inferred climate conditions may be reconciled if the high early Holocene $\delta^{18}O$ and trace element values were controlled principally by low surface runoff and groundwater flow to the lake, rather than high E/P. Dense forest cover in the early Holocene would have increased evapotranspiration and soil moisture storage, thereby reducing delivery of meteoric water to the lakes. Carbonate $\delta^{18}O$ and Mg and Sr decreased between 7200 and 3500 ^{14}C-yr BP in Lake Salpetén and between 6800 and 5000 ^{14}C-yr BP in Lake Petén Itzá. This decline coincided with palynologically documented forest loss that may have led to increased surface and groundwater flow to the lakes. In Lake Salpetén, minimum $\delta^{18}O$ values (i.e., high lake levels) occurred between 3500 and 1800 ^{14}C-yr BP. Relatively high lake levels were confirmed by ^{14}C-dated aquatic gastropods from subaerial soil profiles ~1.0–7.5 m above present lake stage. High lake levels were a consequence of lower E/P and/or greater surface runoff and groundwater inflow caused by human-induced deforestation.

Introduction

Watershed hydrology and the related transfer of materials from terrestrial to aquatic systems are a function of several factors, including the lake/watershed ratio, lake morphometry, local climate variables (e.g., rainfall, temperature, and atmospheric humidity), and watershed characteristics such as bedrock geology, soil type, topography, and vegetation cover (Mason et al., 1994; Street-Perrott, 1995). Removal of vegetation as a result of climate change or clearance by humans, can reduce transpiration and soil moisture storage, thereby increasing water and material transfer to a lake or groundwater (Hibbert, 1967; Bosch & Hewlett, 1982; Bruijnzeel, 1990; Hornbeck & Swank, 1992; Stednick, 1996). Rates of water and material transfer to lacustrine systems, in turn, strongly influence in-lake biogeochemical processes. Historical changes in the delivery rates of dissolved and particulate materials can often be detected in accumulated lake sediments. Reliable paleoenvironmental reconstructions based on lacustrine sediment records therefore require an understanding of all factors that affect catchment processes, including human disturbance (Frey, 1969; Oldfield, 1978; Pennington, 1981; Binford et al., 1983; Deevey, 1984; Binford et al., 1987).

Previous paleoenvironmental studies in lowland Petén, Guatemala (Figure 1a) documented Pleistocene/ Holocene climate and vegetation changes and long-term human impacts on terrestrial and aquatic systems (Cowgill & Hutchinson, 1966; Deevey et al., 1979; Dahlin, 1983; Leyden, 1984, 1987; Rice et al., 1985; Vaughan et al., 1985; Binford et al., 1987; Leyden et al., 1993, 1994; Islebe et al., 1996; Curtis et al., 1998). In this study, Holocene environmental change in the Lake Salpetén watershed was inferred from multiple sediment cores. Stratigraphic variations in the oxygen isotopic ($\delta^{18}O$) and trace element (Mg and Sr) composition of biogenic carbonate from these cores were interpreted as reflecting altered watershed hydrologic balance that resulted from changes in climate, vegetation, and human disturbance in the catchment. Changes in material transfer to the lake bottom were inferred from sediment composition and accumulation, and vegetation changes within the catchment 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

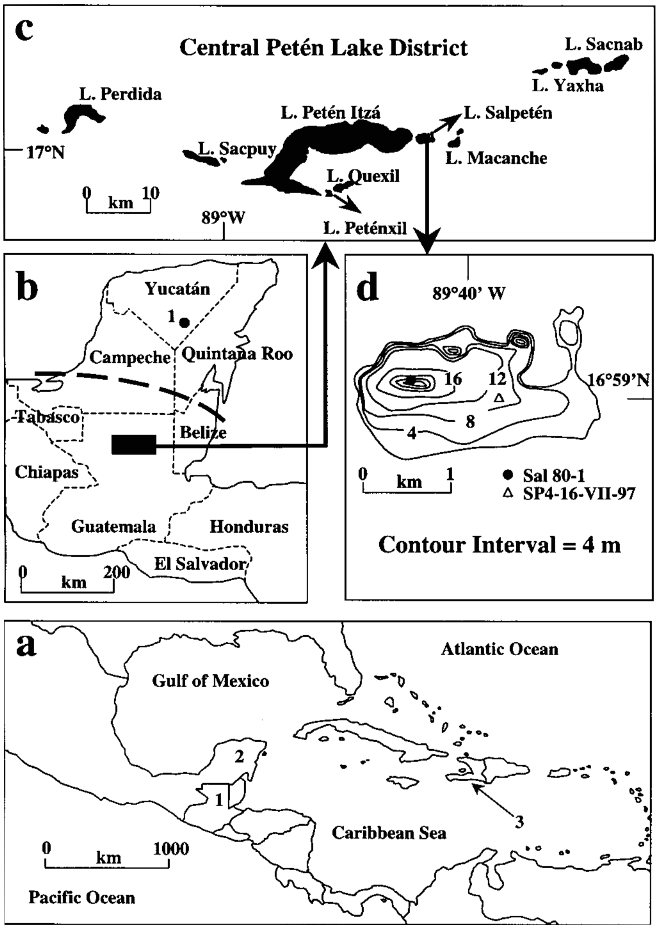

Figure 1. (a) Map of the Caribbean region showing lake study sites in northern Guatemala (1) and Lakes Chichancanab (2) and Miragoane (3). (b) Map of the Yucatán Peninsula showing lake study sites in northern Guatemala (black rectangle) and Lake Chichancanab (1). Heavy dashed line indicates the division between the northern and southern Maya lowlands. (c) Petén lake district in Guatemala. (d) Bathymetry and sediment core sites in Lake Salpetén.

records from Lake Chichancanab, Mexico (Hodell et al., 1995) and Lake Miragoane, Haiti (Hodell et al., 1991; Curtis & Hodell, 1993).

Proxy indicators of environmental change

Variations in the oxygen isotopic ratio ($^{18}O/^{16}O$) of lacustrine carbonate result from changes in the temperature of carbonate precipitation and shifts in the $^{18}O/^{16}O$ composition of the lake water from which the carbonate precipitates (Craig, 1965; Gonfiantini, 1965; Stuiver, 1968, 1970; Covich & Stuiver, 1974; Fritz et al., 1975). In tropical lakes, temperature fluctuations are assumed to be minor during the Holocene and variations in the $^{18}O/^{16}O$ of lacustrine carbonate are thought to have been dominated by changes in lake water residence time and the relative rates and isotopic composition of hydrologic inputs and outputs (Gasse et al., 1990; Talbot, 1990; Hodell et al., 1991; Lister et al., 1991; Curtis & Hodell, 1993; Curtis et al., 1996; Xia et al., 1997).

The hydrologic balance of a lake (dV_{LAKE}) is controlled by the transfer of water to and from the catchment according to the equation:

$$dV_{LAKE} = \sum I + P - \sum O - E$$

where $\sum I$ and $\sum O$ are the total surface and groundwater inflows (I) to, and outflows (O) from the lake, P is direct precipitation on the lake, and E is the evaporative loss from the lake (Pearson & Coplen, 1978; Gat, 1984). A similar expression can be written for the oxygen isotopic composition of the lake water:

$$dV_{LAKE}\,\delta_{LAKE} = \sum I\,\delta_I + P\,\delta_P - \sum O\,\delta_O - E\,\delta_E$$

where δ is the isotopic composition of the various inputs and outputs. In lakes with negligible surface outflow (O), the water balance and lake water isotopic composition is determined by the difference between evaporation and precipitation over the lake (E–P) and the water balance of the surrounding catchment ($\sum I$) (Craig, 1961; Gonfiantini, 1965; Fontes & Gonfiantini, 1967; Phillips et al., 1986). Lake water ^{18}O enrichment relative to the isotopic signature of direct precipitation and surface and groundwater inflow reflects preferential evaporative loss of $H_2{}^{16}O$ from the lake. Extended periods of enhanced evaporation or reduced precipitation and surface inflow (high E/P or E/I) result in higher $^{18}O/^{16}O$ ratios of lake water and precipitated carbonate. Conversely, increased surface inflow or precipitation (low E/P or E/I) results in lower $^{18}O/^{16}O$ ratios.

Trace element (especially Mg and Sr) chemistry of lacustrine ostracod valves also serves as an indicator of past changes in watershed hydrology (Chivas et al., 1993; Holmes, 1996). Decreased rainfall or surface flow results in enhanced precipitation of lacustrine carbonate and trace element enrichment of lake water (Hardie & Eugster, 1970). The magnesium and strontium concentration of biogenic carbonate in turn reflects the composition of the lake water and thereby the hydrologic balance of a lake (Chivas et al., 1985, 1986). Incorporation of magnesium in carbonate is also affected by water temperature (increased temperature results in increased Mg concentrations) but concentration in lake water is likely to be the major control in tropical lakes.

In addition to altering the hydrologic budget of a lake, increased or decreased precipitation and surface flow alter the transfer of dissolved and particulate materials to a lake. The flux of materials to the sediments is a function of the rate of material output from the catchment (Imboden & Lerman, 1978; Binford et al., 1983, 1987). Changes in the material input to a lake produce stratigraphic changes in both the sediment accumulation rate and the biogeochemistry of sediment (Deevey, 1984; Engstrom & Wright, 1984).

Changes in lacustrine sediment composition arise primarily from variations in material loss from the catchment, often the result of land-use changes. For example, soil changes resulting from deforestation and fires, produce variations in the flux of particulate matter to a lake (Dearing et al., 1987; 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 the pollen deposited in lake sediments. Deforestation enhances transport of soil nutrients and organic and inorganic matter to the lake, which is reflected in the sediment lithologic composition (Deevey, 1984; Binford et al., 1987).

Environmental setting

The Yucatán Peninsula includes the Department of Petén in northern Guatemala, Belize, and the Mexican states of Campeche, Yucatán, Quintana Roo, and portions of Tabasco and Chiapas (Wilson, 1980) (Figure 1b). The region is characterized principally by low-lying karsted limestones of Cretaceous and Tertiary age (Vinson, 1962). In the northern lowlands, karstic terrain is fully developed, surface drainage is virtually non-existent, and water falling on the land surface is either evaporated, transpired, or quickly delivered to the aquifer. Surface elevation is relatively low, however, and the water table is exposed in the many sinkholes where the limestone has collapsed. In the southern lowlands of Petén, surface elevation varies between 100 and 300 m above mean sea level and groundwater is fairly inaccessible. The karst landforms of Petén are less developed, however, and surface waters are perched, resulting in numerous lakes and seasonally inundated topographic depressions (Deevey et al., 1979). The lake district contains a series of terminal basins distributed along a series of east-west aligned faults at 17 °N latitude. Principal water bodies of the lake chain extend approximately 100 km from westernmost Lake Perdida eastward to the twin basins Yaxhá and Sacnab (Figure 1b).

Annual rainfall across the Yucatán Peninsula is highly variable, ranging from a low of 500 mm yr^{-1} along the northwest coasts of Yucatán and Campeche to over 2000 mm yr^{-1} in the southern lowlands (Wilson, 1980). Rainfall in Petén varies spatially and inter-annually from ca. 900–2500 mm, with a regional annual mean of 1601 mm (Deevey, 1978; Deevey et al., 1980). Heavy rains between June and October are associated with the northward migration of the inter-tropical convergence zone (ITCZ) and the Azores-Bermuda high-pressure system. This period is characterized by weak trade winds and warm sea surface temperatures in the Atlantic between about 10 ° and 20 °N. Conversely, dry conditions develop in November and December as the ITCZ and the Azores-Bermuda high move equatorward and strong trade winds become predominant (Hastenrath, 1976, 1984).

Seasonal aridity and the precipitation gradient in Yucatán are reflected by regional soil development and vegetation distribution. Soils in the northern lowlands are thin, and support dry-adapted scrub vegetation and some semi-evergreen forest of medium height (Wilson, 1980). Farther south, in Petén, the landscape is dominated by well-drained forest soils (Simmons et al., 1959) and tropical semi-deciduous and evergreen vegetation (Lundell, 1937). Forest soils of Petén are relatively fertile, but cultivation is restricted by steep slopes (Deevey et al., 1979; Rice et al., 1985). Under deforestation, the erosive potential of intense tropical rain contributes to the rapid transport of dissolved and particulate materials from the land to lakes.

Lake Salpetén

Lake Salpetén (16°58′N and 89°40′W) is a small ($A = 2.6$ km^2), terminal lake basin within the central Petén lake district (Figure 1c). The lake lies at 104 m a.s.l. and has a maximum depth of 32 m (Brezonik & Fox, 1974) (Figure 1d). Lake Salpetén is sulfate-rich (3000 mg l^{-1}) and relatively saline (4500 mg l^{-1} TDS), with a pH of 8.5 and a conductivity of ~ 3600 μS cm^{-1}. Surface water temperatures typically average between 27 ° and 30 °C throughout the year. Salpetén is an exceptional site for paleoenvironmental research. Evaporative loss from the lake is substantial, as demonstrated by the ~ 7‰ enrichment of lake water (+4.1‰, n = 4) relative to precipitation (–2.7‰, n = 11) and groundwater (–3.4‰, n = 2). Temporal changes in the $^{18}O/^{16}O$ ratio of the lake water are controlled primarily by changes in the ratio of evaporation to inputs, and past changes in the $^{18}O/^{16}O$ ratio of the lake water are recorded in the $\delta^{18}O$ record of carbonate in gastropod and ostracod shells preserved in the lake sediments. The basin lacks outflows, and sediments are therefore the ultimate sink of most dissolved and particulate matter that enters the lake.

Methods

In May 1980, fifteen meters of sediment were obtained from the deep basin of Lake Salpetén (Figure 1d) (Deevey et al., 1983). Sediments were collected with a 45.7-cm split-spoon sampler. The split-spoon sampler did not recover unconsolidated surface sediments, and uppermost retrieved deposits lie approximately 1.6 m below the sediment surface. In July 1997, additional sediment was recovered from Lake Salpetén in a water depth of 9.2 m (SP4-16-VII-97). Surface sediments were collected with a piston corer designed to retrieve undisturbed sediment-water interface profiles (Fisher et al., 1992) and deeper sections were taken in 1.0-m segments with a square-rod piston corer (Wright, 1984). The interface core was sectioned in the field at 1.0-cm intervals by upward extrusion into a sampling tray fitted to the top of the core barrel. Square-rod core sections were extruded in the field, stored in PVC pipe, transported to the laboratory, and sampled at 1.0-cm intervals. Archived core sections from the deep basin of Lake Salpetén, designated Sal 80-1, were reexamined in May 2000 and sampled at 5.0-cm intervals to a depth of ~ 1090 cm.

Sediment ages from Lake Salpetén were determined by accelerator mass spectrometry (AMS) of ^{14}C in terrestrial organic matter (wood, charcoal, and seed). Radiocarbon samples were measured at Lawrence Livermore National Laboratories and the National Ocean Science AMS Facility at Woods Hole Oceanographic Institution. Calibrated dates and calendar ages were calculated using the INTCAL98 program with a 100-yr moving average of the tree-ring calibration data set (Stuiver et al., 1998).

Oxygen and carbon isotopic ratios were measured on gastropod shells (*Cochliopina* sp.) and ostracod valves (*Heterocypris* sp., *Limnocythere* sp., and *Physocypria globula*.). Sediment samples were disaggregated in 3% H_2O_2 and washed through a 63-µm sieve. Coarse material (> 63 µm) was dried at 60 °C. Adult ostracod valves and gastropod shells were picked from the dried samples, soaked in 15% H_2O_2, cleaned ultrasonically in deionized water, and rinsed with methanol before drying. Aggregate samples of ~ 25 ostracod carapaces were measured from each 1.0-cm sediment sample. Gastropod shells of ~ 15 individuals were ground to a fine powder and a fraction of the ground carbonate was analyzed from each sample.

Carbonate samples for stable isotope analysis from Sal 80-1 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 on-line with a Finnigan MAT 252 mass spectrometer and compared to an internal gas standard. Carbonate samples from SP4-16-VII-97 were reacted in a common acid bath of 100% orthophosphoric acid at 90 °C using a VG/Micromass Isocarb preparation system. Stable isotope analysis of the resulting CO_2 gas was determined on-line with a triple-collector VG/Micromass Prism Series II mass spectrometer. All carbonate isotopic values are expressed in conventional delta (δ) notation as the per mil (‰) deviation from Vienna PeeDee Belemnite (VPDB). Precision for $\delta^{18}O$ and $\delta^{13}C$ of the samples analyzed with the Finnigan MAT 252 was ± 0.09‰ and ± 0.03‰, respectively. Precision of the samples analyzed with the VG/Micromass Prism Series II was ± 0.08‰ ($\delta^{18}O$) and ± 0.06‰ ($\delta^{13}C$).

Trace element (Mg and Sr) concentrations of the ostracods *Limnocythere* sp. and *Cytheridella ilosvayi* were determined by atomic absorption techniques with a Perkin-Elmer 3100 EDS spectrophotometer and coupled HGA-600 graphite furnace with autosampler. Multiple-valve samples (weighing ~ 50 µg) were measured after dissolution in 5.0-mL of 2% nitric acid. Elemental concentrations are expressed in µg g^{-1} by weight of calcium carbonate ($CaCO_3$). Precision of replicate sample analyses was ± 0.74% and ± 1.60% relative standard deviation for Mg and Sr, respectively.

Inorganic carbon (IC) in the sediments 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 estimated to be ± 0.6% based upon repeated analysis of reagent-grade calcium carbonate. Total carbon (TC) in the sediments 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

Nine AMS ^{14}C dates were obtained from core Sal 80-1 and yielded a maximum age of 8220 ^{14}C-yr BP (Table 1). Age-depth values were determined by linear interpolation between dated horizons, and linear sedimentation rates were extrapolated to the base of the sampled sediment profile (Figure 2). In core SP4-16-VII-97, eight AMS ^{14}C samples provided a basal age of 8780 ^{14}C-yr BP (Table 1). The age-depth profile in the core (Figure 3) demonstrates abrupt change in radiocarbon activity after 4130 ^{14}C-yr BP, suggesting erosion or non-deposition. Following the apparent hiatus in core SP4-16-VII-97, continuous sedimentation is evident since at least 1130 ^{14}C-yr BP.

Paleoenvironmental proxies from Sal 80-1 and SP4-16-VII-97 are plotted against depth (Figures 4 & 5) and results are discussed as a function of age (Figure 6). The $\delta^{18}O$ of biogenic carbonate in Lake Salpetén sediments was high prior to 7200 ^{14}C-yr BP and averaged 4.7‰ (Figure 6a). Between 7200 and 3500 ^{14}C-yr BP, mean values decreased to ~ 2.3‰, with the exception of a brief excursion toward higher values centered at 4700 ^{14}C-yr BP. Similarly, the trace element (Mg and Sr) concentration of biogenic carbonate was relatively high prior to 5240 ^{14}C-yr BP, but decreased by 4130 ^{14}C-yr BP (Figure 4b). Magnesium and strontium concentrations remained low after 4130 ^{14}C-yr BP. Minimum $\delta^{18}O$ values (averaging 1.7‰) occurred between 2400 and 1800 ^{14}C-yr BP (Figure 6a).

Basal sediments in Lake Salpetén had relatively high $CaCO_3$ concentrations that decreased to < 20% between 9500 and 9000 ^{14}C-yr BP (Figure 6b). Organic carbon

Table 1. AMS radiocarbon dates for samples from Lake Salpetén, Petén, Guatemala

Core	Sample type	Depth (cm)	Accession number	Target size (mg C)	Radiocarbon age (yr B.P.)	Calibrated age (A.D./B.C.)	Error (± yr BP)
Sal 80-1	Charcoal	157	CAMS 65387	0.03	1820	A.D. 199	190
	Charcoal	395	CAMS 65388	0.09	2200	219 B.C.	70
	Charcoal	503	CAMS 65389	0.11	2200	219 B.C.	60
	Charcoal	646	CAMS 64861	0.19	2430	459 B.C.	50
	Charcoal	707	CAMS 65390	–	2500	748 B.C.	60
	Charcoal	774	CAMS 64978	0.03	2990	1237 B.C.	190
	Charcoal	813	CAMS 65797	0.08	3160	1421 B.C.	80
	Charcoal	960	CAMS 65391	–	6990	5851 B.C.	50
	Charcoal	1012	CAMS 65798	–	8220	7235 B.C.	50
SP4-16-VII-97	Wood	20	OS 14151	–	480	A.D. 1436	65
	Wood	75	CAMS 55671	–	1130	A.D. 923	40
	Wood	97	OS 18641	–	4130	2646 B.C.	55
	Wood	103	OS 14153	–	5240	4026 B.C.	60
	Wood	151	OS 14152	–	8660	7610 B.C.	65
	Wood	162	OS 14154	–	8670	7623 B.C.	95
	Wood	165	OS 14156	–	8770	7890 B.C.	65
	Wood	172	OS 141551	–	8780	7900 B.C.	75

concentration was relatively low (< 5%) prior to 9000 ^{14}C-yr BP, but increased by 8200 ^{14}C-yr BP (Figure 6c). Organic carbon concentrations remained high (typically > 15%) between 8200 and 4300 ^{14}C-yr BP. Between 9000 and 4800 ^{14}C-yr BP, $CaCO_3$ content was highly variable, with maxima (30–40%) centered at 8100, 7100, and 6100 ^{14}C-yr BP. After 4800 ^{14}C-yr BP, $CaCO_3$ content increased steadily from < 30% to ~ 50%. Organic C concentration decreased from 20% to less than 5% between 4300 and 3200 ^{14}C-yr BP and remained low between 3200 and 1800 ^{14}C-yr BP.

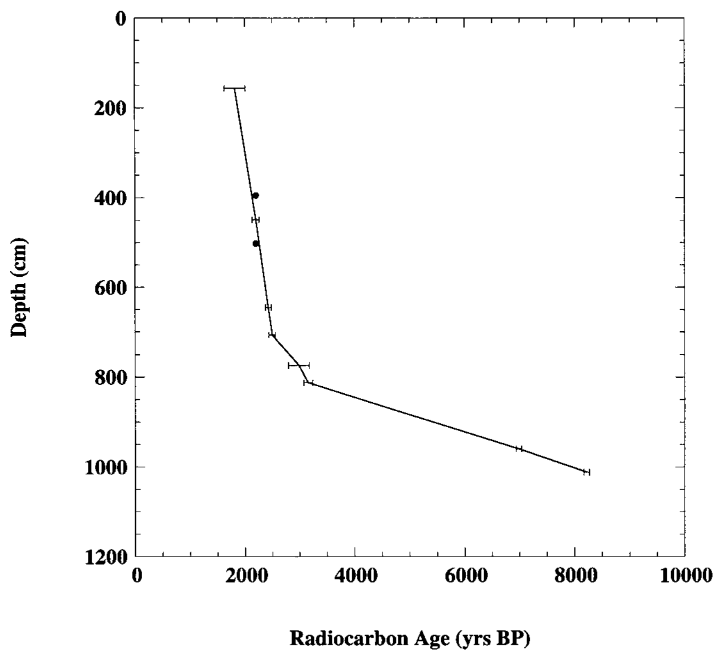

Figure 2. Radiocarbon age (^{14}C-yr BP) vs. depth for Lake Salpetén core Sal 80-1. Age-depth values were determined by linear interpolation between dated horizons.

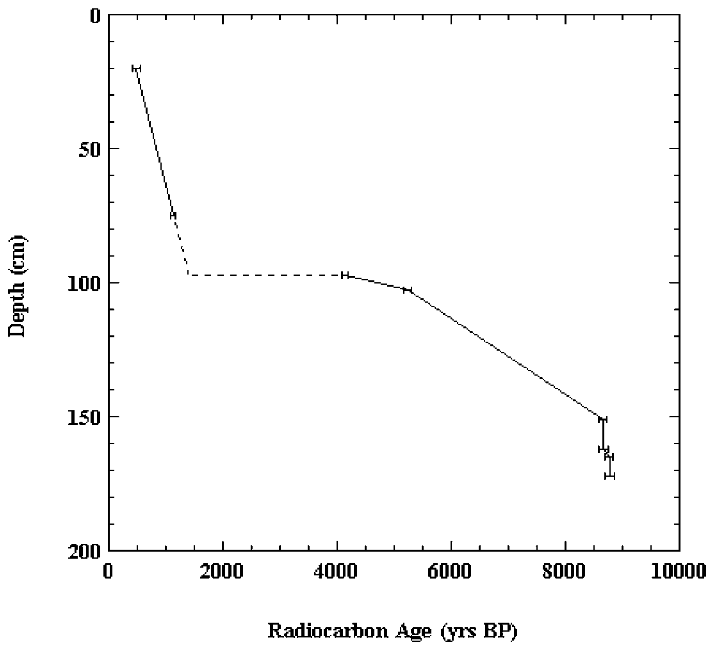

Figure 3. Radiocarbon age (^{14}C-yr BP) vs. depth for Lake Salpetén core SP4-16-VII-97. The age-depth profile in the core demonstrates abrupt change in radiocarbon activity after 4130 ^{14}C-yr BP.

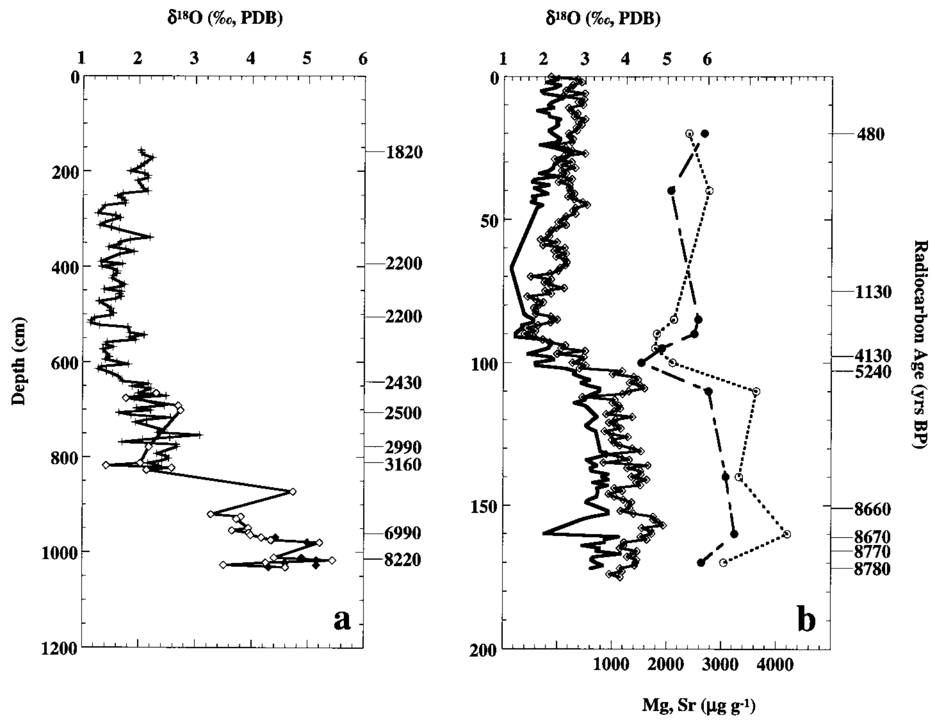

Figure 4. (a) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and two ostracods, *Physocypria globula* (crosses) and *Heterocypris* sp. (filled diamonds) vs. depth in core Sal 80-1. (b) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and the ostracod *Limnocythere* sp. (solid line) and Mg (open circles) and Sr (filled circles) concentrations in valves of the ostracod *Limnocythere* sp. vs. depth in core SP4-16-VII-97. Trace element analysis of core Sal 80-1 was not completed as a result of limited ostracod preservation.

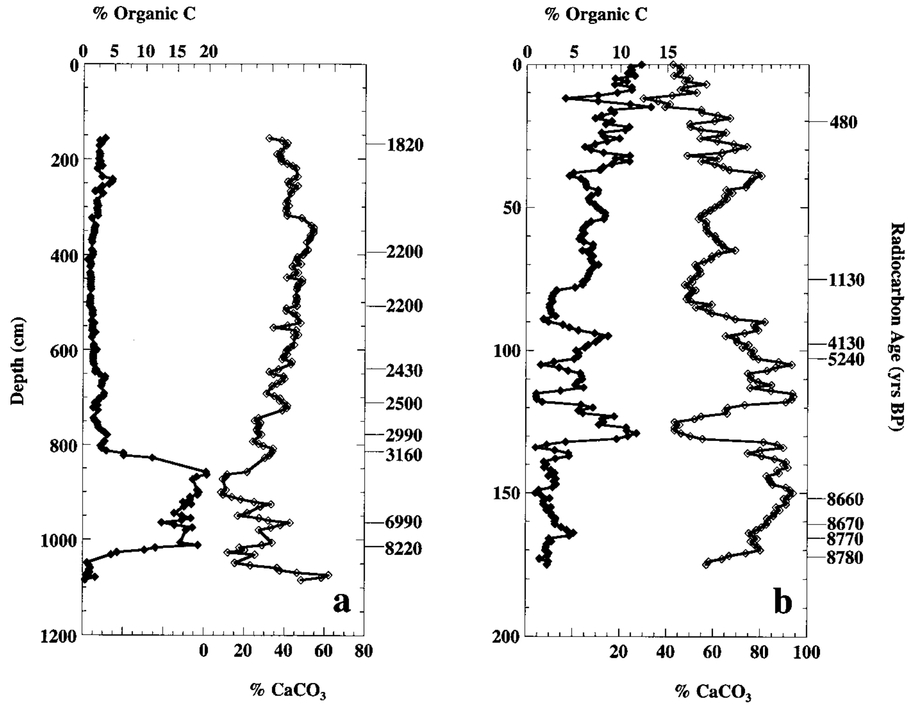

Figure 5. (a) Organic carbon content (filled diamonds) and $CaCO_3$ concentration (open diamonds) vs. depth in core Sal 80-1. (b) Organic carbon content (filled diamonds) and $CaCO_3$ concentration (open diamonds) vs. depth in core SP4-16-VII-97.

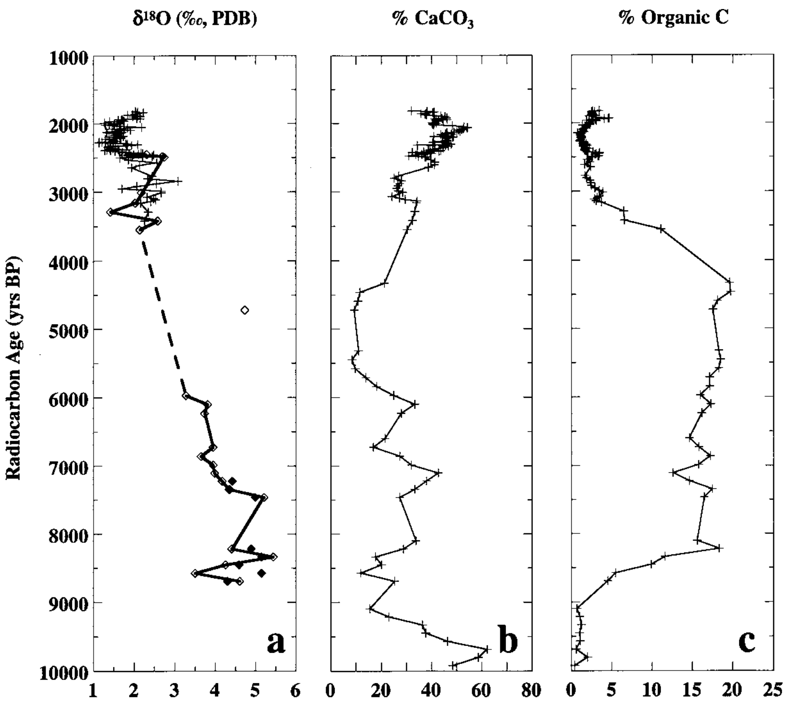

Figure 6. (a) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and two ostracods, *Physocypria globula* (crosses) and *Heterocypris* sp. (filled diamonds); (b) $CaCO_3$ concentration; and (c) organic carbon content vs. radiocarbon age (years BP) in Lake Salpetén core Sal 80-1.

Discussion

Early Holocene climate, vegetation, and watershed interactions (9000 to 7000 ^{14}C-yr BP)

The oldest date from Salpetén core SP4-16-VII-97 indicates that the lake filled above a depth of 9.2 m below present level by ~ 9000 ^{14}C-yr BP. This rise in water level coincided with the filling of Lake Petén Itzá ca. 9000 ^{14}C-yr BP (Curtis et al., 1998) and Lake Quexil prior to 8410 ^{14}C-yr BP (Vaughan et al., 1985). Early Holocene filling of Petén lakes is attributed to increased moisture availability following the arid late Pleistocene (Leyden, 1984; Leyden et al., 1993, 1994). Oxygen isotopic values in Lake Salpetén were highest in the early Holocene between ~ 9000 and 7200 ^{14}C-yr BP (Figure 7a). The isotope profile from nearby Lake Petén Itzá (Curtis et al., 1998) reveals a similar pattern of variation with the greatest δ^{18}O values between 9000 and 6800 ^{14}C-yr BP (Figure 7b). If the δ^{18}O signal is interpreted as a reflection of E/P, the records suggest that climate was relatively dry during the early Holocene, prior to ~ 6800 ^{14}C-yr BP. This interpretation is at odds with palynological evidence from many Petén lakes suggesting moist conditions and extensive lowland forests in the early Holocene from 9000 to 5600 ^{14}C-yr BP (Figures 7c & 7d) (Leyden, 1984, 1987; Vaughan et al., 1985; Islebe et al., 1996).

The discrepancy between early Holocene climatic inferences from pollen and δ^{18}O records may be reconciled if lake water δ^{18}O was controlled not only by E/P, but also by varying inputs of surface and groundwater flow to the lake as a result of changing vegetation density in the watershed. Input waters to Petén lakes display relatively light oxygen isotopic values, whether delivered by direct rainfall (–2.7‰) or surface and groundwater flow (–3.4‰). Reduced input of meteoric waters would decrease lake volume, increase the pro-

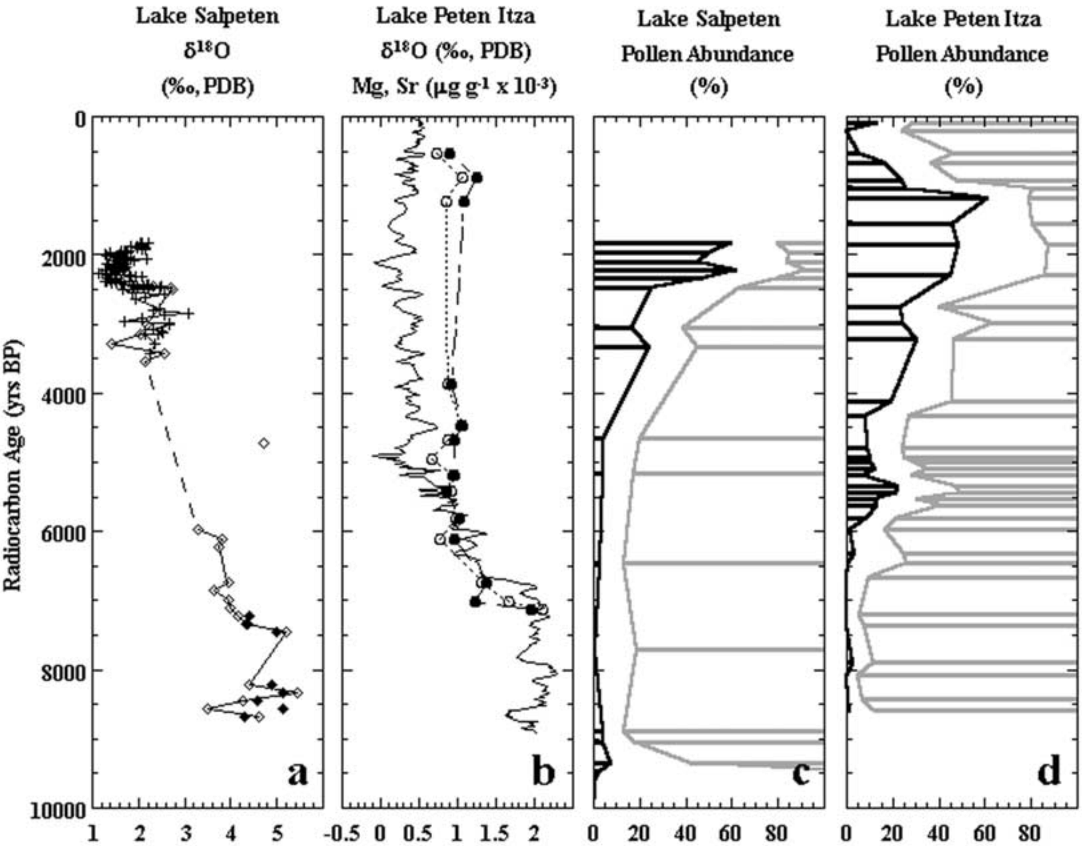

Figure 7. (a) Oxygen isotopic composition of gastropods and ostracods from Lake Salpetén core Sal 80-1, (b) δ^{18}O of gastropods from Lake Petén Itzá (solid line) and Mg and Sr concentrations (open and filled circles, respectively) in valves of the ostracod *Cytheridella ilosvayi*, and relative abundance of pollen types in the (c) Salpetén and (d) Petén Itzá sediment core vs. radiocarbon age. Black bars indicate the relative abundance of disturbance taxa and grey bars indicate the relative abundance of lowland forest taxa.

portion of the hydrologic budget lost to evaporation, and therefore increase lake water $\delta^{18}O$. Dense forest cover in the early Holocene may have increased evapotranspiration and soil moisture storage in the watershed, thereby reducing the catchment water yield and transport of isotopically light surface and ground waters to the lake. As a result, early Holocene $\delta^{18}O$ values of lake water and biogenic carbonate would have been high, explaining the contradictory climatic inferences from pollen and isotopes.

Oxygen isotopic records from Petén lakes may also have been influenced by changes in the isotopic composition of precipitation and/or reduced deepwater temperatures (and consequent high $\delta^{18}O$) during periods of elevated lake level. However, Mg and Sr values in ostracod shells from Lakes Salpetén and Petén Itzá suggest otherwise. Trace element concentrations in Lake Salpetén were highest in the early Holocene between ~ 9000 and 5200 ^{14}C-yr BP (Figure 4b). Magnesium and Sr concentrations in Petén Itzá were similarly high prior to ~ 6800 ^{14}C-yr BP (Figure 7b). This suggests that the $\delta^{18}O$ records, which parallel trace element concentrations, largely reflect basin hydrology rather than changes in the isotopic composition of precipitation. If the early to mid-Holocene decline in $\delta^{18}O$ reflected an increase in temperature, one would expect to see an increase in Mg values. Covariant decreases in both $\delta^{18}O$ and Mg eliminate temperature as a control on the lacustrine $\delta^{18}O$ record.

Middle Holocene deforestation and hydrologic change (7000 to 3500 ^{14}C-yr BP)

Middle Holocene pollen records indicate that lowland tropical forest taxa declined as open vegetation taxa increased (Leyden, 1984, 1987; Vaughan et al., 1985; Islebe et al., 1996). For example, lowland forest pollen of the *Moraceae-Urticaceae* group declined and pollen of the savanna indicator *Byrsonima* ('nance') increased in the Lake Petén Itzá record as early as ~ 5600 ^{14}C-yr BP. This suggests a transition to open forest as a result of climate change (increased E/P) or early human impact (Islebe et al., 1996; Curtis et al., 1998).

The $\delta^{18}O$ values and trace element concentrations of biogenic carbonate in cores from Lakes Salpetén and Petén Itzá decreased, coincident with the decline of lowland forest taxa. The discrepancy between the timing of the decrease in the two lacustrine records probably reflects minor chronological inaccuracies. The $\delta^{18}O$ decrease in Lake Salpetén occurred between ~ 7200 and 3500 ^{14}C-yr BP (Figure 7a), whereas $\delta^{18}O$ and Mg and Sr decreased between ~ 6800 ^{14}C-yr BP and 5000 ^{14}C-yr BP in Lake Petén Itzá (Figure 7b) (Curtis et al., 1998). Pollen and geochemical records in both lakes yield contradictory climate inferences if $\delta^{18}O$ and trace element data are interpreted as reflecting only changes in E/P. Forest decline may have reduced evapotranspiration and soil moisture storage, thereby increasing surface and groundwater flow to the lake basins, resulting in higher lake levels and lower lake water $\delta^{18}O$ values and trace element concentrations.

Late Holocene climate change and human disturbance (3500 to 1800 ^{14}C-yr BP)

Minimum $\delta^{18}O$ values in Lake Salpetén, suggesting high lake levels, were reached between 3500 and 1800 ^{14}C-yr BP (Figure 7a), corresponding to the period of Maya occupation. Independent evidence for high water level is provided by twelve AMS ^{14}C dates on aquatic gastropods from subaerial lacustrine deposits ~ 1.0–7.5 m above modern lake stage (Table 2). The oldest dates indicate that the lake filled to a depth of more than 7.0 m above modern lake stage between ~ 4600 and 3600 ^{14}C-yr BP, but most indicate episodes of high lake levels between 2700 and 1300 ^{14}C-yr BP. These ages coincide with reduced $\delta^{18}O$ in the core and support the inference for increased water input to Lake Salpetén. In contrast, mean $\delta^{18}O$ values in the Lake Petén Itzá core were nearly constant from 5000 ^{14}C-yr BP to present and fluctuated by only ~ 0.5‰ (Figure 7b) (Curtis et al., 1998). The 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 (Curtis et al., 1998).

High lake levels and $\delta^{18}O$ minima in Lake Salpetén may reflect increased precipitation and reduced evaporation beginning after 3500 ^{14}C-yr BP. Changes in the $\delta^{18}O$ record may also have been influenced by hydrologic changes in the watershed caused by Maya land-use practices. For example, minimum $\delta^{18}O$ values between 3500 and 1800 ^{14}C-yr BP coincide with Maya-induced deforestation around Lakes Salpetén and Petén Itzá (Figures 7c & 7d; Leyden, 1987; Islebe et al., 1996) and other Petén lakes (Tsukada, 1966; Deevey et al., 1979; Vaughan et al., 1985). During the same time interval, geochemical records from Lake Salpetén showed changes that reflect catchment soil erosion. Calcium carbonate values in core Sal 80-1 reached a maximum ~ 2000 ^{14}C-yr BP (Figure 6b). High carbonate values indicate agricultural forest clearance, accelerated ero-

Table 2. AMS radiocarbon dates for gastropod shells from soil samples (subaerial sediments) from the catchment of Lake Salpetén

Soil elevation (m)	Soil depth (cm)	Accession number	Carbonate $\delta^{18}O$ (‰, PDB)	Carbonate $\delta^{13}C$ (‰, PDB)	Radiocarbon age (yr BP)	Calibrated age (A.D./B.C.)	Age error (± yr)
~ 1.0	0	CAMS 56113	1.51	–2.39	1510	A.D. 573	50
	10	CAMS 56114	1.64	–2.29	1540	A.D. 542	40
	20	CAMS 56115	1.18	–3.18	2150	180 B.C.	40
	30	CAMS 56116	1.07	–2.55	2290	378 B.C.	40
~ 4.5	0	CAMS 60071	1.87	–1.85	1300	A.D. 694	40
	10	CAMS 60072	1.99	–1.89	1820	A.D. 199	40
	20	CAMS 60073	1.61	–1.83	2220	219 B.C.	40
~ 7.5	10	CAMS 60674	1.08	–3.20	1810	A.D. 208	40
	20	CAMS 60075	1.05	–3.07	2570	782 B.C.	40
	30	CAMS 60658	0.83	–4.88	2690	839 B.C.	40
	40	CAMS 60076	0.77	–2.49	3570	1910 B.C.	40
	50	CAMS 60659	1.17	–2.13	4570	3347 B.C.	40

sion of catchment soils, and transport of soil-derived carbonate to the lake. An increase in inorganic colluvium, termed the 'Maya clay', is documented in many Petén lakes during the period of Maya occupation (Cowgill & Hutchinson, 1966; Deevey et al., 1979; Brenner, 1983; Rice et al., 1985; Binford et al., 1987). As forests were removed by the rapidly expanding Maya population, evapotranspiration within the drainage basins may have been reduced, resulting in increased inflow to the lake.

Regional correlations and mechanisms of Holocene climate change

Most pollen records from circum-Caribbean lake sediment cores indicate that the wettest period throughout the lowland Northern Hemisphere tropics occurred between ~ 7000 and 5000 ^{14}C yr BP (Bradbury et al., 1981; Deevey et al., 1983; Leyden, 1985; Piperno et al., 1990; Islebe et al., 1996). Hodell et al. (1991) suggested that this period of high precipitation (low E/P) was related to an increase in the intensity of the annual cycle driven by the Earth's precessional cycle. This explanation is supported by meteorological studies of Caribbean climate that demonstrate a strong correlation between precipitation anomalies and intensity of the annual cycle (Hastenrath, 1984). Years with abundant rainfall coincide with an enhancement of the annual cycle when the ITCZ moves farther north during the summer rainy season and farther south during the winter dry season. Summer perihelion in the Northern Hemisphere, which occurred at ~ 8000 ^{14}C yrs BP, would have been associated with increased intensity of the annual cycle.

Early and middle Holocene $\delta^{18}O$ records from Petén lakes and Lakes Chichancanab (Mexico) and Miragoane (Haiti) exhibit relatively consistent patterns of millennial-scale change, suggesting that shifts in E/P were controlled (at least in part) by orbitally forced variations in seasonal insolation that modified the intensity of the annual cycle (Figure 8). Direct orbital forcing alone, however, does not fully explain the timing, magnitude, or rate of water level change recorded in the lakes. For example, inferred aridity during the earliest Holocene is more pronounced than predicted by insolation forcing. In Yucatán and Haiti, the early to middle Holocene moist period apparently began considerably earlier than in northern Guatemala. Discrepancies among sites may be a consequence of inaccuracies in age/depth relations. Alternatively, hydrologic processes that governed lake filling may have varied among sites, thereby confounding interpretation of the $\delta^{18}O$ records. Dense forest cover in the lowlands of northern Guatemala may have decreased surface runoff and groundwater flow to the lake basins more effectively than the scrub and open forest vegetation of the Chichancanab and Miragoane catchments (Leyden et al., 1998; Higuera-Gundy et al., 1999). Furthermore, rising sea level probably affected early Holocene filling of low-elevation Lakes Chichancanab and Miragoane, but was not a factor in the filling of inland, higher-elevation Lakes Salpetén and Petén Itzá (~ 110 m a.s.l.).

Oxygen isotopic values in Lakes Chichancanab and Miragoane increased abruptly after ~ 3200 ^{14}C-yr BP, suggesting late Holocene climatic drying (Figures 8c & 8d). During the same time interval, Lake Miragoane pollen records indicate the decline of mature forests

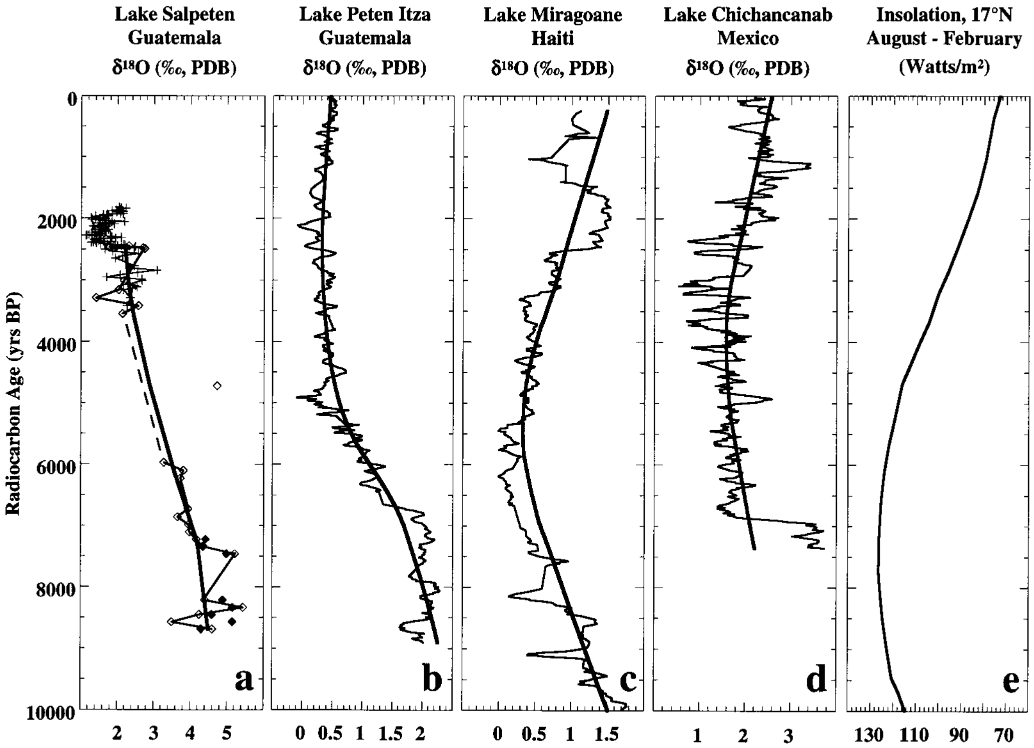

Figure 8. Oxygen isotope records from (a) Lake Salpetén core Sal 80-1, (b) Lake Petén Itzá, (c) Lake Miragoane, and (d) Lake Chichancanab vs. radiocarbon age, and (e) the difference in insolation (W/m^2) at 17 °N latitude between August and February, which is a measure of the intensity of the annual cycle (Hastenrath, 1984).

dominated by *Moraceae* and the prevalence of dry forest taxa (Higuera-Gundy et al., 1999). In contrast, $\delta^{18}O$ values in the Petén Itzá core remained relatively invariant after ~ 5000 ^{14}C-yr BP (Figure 8b) and minimum $\delta^{18}O$ values between 3500 and 1800 ^{14}C-yr BP in Lake Salpetén (Figure 8a) indicate high lake level, perhaps the result of altered basin hydrology and increased surface runoff and groundwater inflow. Moreover, late Holocene pollen-based climate reconstructions from Petén lakes have been confounded by the effects of Maya-induced deforestation (Deevey, 1978; Deevey et al., 1979; Vaughan et al., 1985; Leyden, 1987; Islebe et al., 1996). Neither Lake Chichancanab nor Lake Miragoane is believed to have suffered significant catchment disturbance by human populations prior to ~ 1400 ^{14}C-yr BP (Binford et al., 1987; Leyden et al., 1998). In the absence of human impact, the late Holocene lacustrine $\delta^{18}O$ records from northern Yucatán and Haiti probably reflect climatic variations more accurately. Removal of dry forest, scrub vegetation surrounding Lakes Chichancanab and Miragoane would only slightly increase water yield relative to the increase associated with clearing of lowland forest (Sahin et al., 1996). The best late Holocene paleoclimate records are thus likely to be found in lakes from minimally impacted drainage basins.

Conclusions

Discrepancies between Holocene climatic inferences based on pollen and geochemical records from Petén lakes can be reconciled if lake water $\delta^{18}O$ and trace element concentration was not controlled predominantly by E/P, but rather by the changing input of surface and groundwater to the lakes. Dense early Holocene forests caused high evapotranspiration and soil moisture storage in the watershed, thereby reducing the input of meteoric waters to the lakes. Forest decline after ~ 5600 ^{14}C-yr BP reduced evapotranspiration and soil moisture storage, thereby increasing run-off and groundwater flow to the lakes. High input of meteoric waters to the lakes is reflected by low $\delta^{18}O$ and Mg and Sr values in Lake Salpetén between 7200 and 3500 ^{14}C-yr BP and in Lake Petén Itzá between 6800 ^{14}C-yr BP and 5000 ^{14}C-yr BP. After 3500 ^{14}C-yr BP, $\delta^{18}O$ and trace element values in Lake Salpetén decreased, reaching minimum values between 3500 and 1800 ^{14}C-yr BP. High lake levels were confirmed by AMS ^{14}C dates of aquatic gastropods from subaerial soil (i.e., sediment) profiles that yielded ages ranging from 4570 to 1510 ^{14}C-yr BP. High lake levels may have resulted from increased precipitation (reduced E/P) and/or increased surface runoff and groundwater inflow as a consequence of human-induced deforestation. Results from Petén lakes suggest that natural or anthropogenic changes in vegetation cover within watersheds can alter lake hydrologic budgets, thereby confounding paleoclimatic inferences based on the δ^{18}O and trace element composition of biogenic carbonate.

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