

Climate change in lowland Central America during the late deglacial and early Holocene

MICHAEL B. HILLESHEIM,^{1*} DAVID A. HODELL,¹ BARBARA W. LEYDEN,² MARK BRENNER,¹ JASON H. CURTIS,¹ FLAVIO S. ANSELMETTI,³ DANIEL ARIZTEGUI,⁴ DAVID G. BUCK,⁵ THOMAS P. GUILDERSON,⁶ MICHAEL F. ROSENMEIER^{1†} and DOUGLAS W. SCHNURRENBERGER⁷

¹Department of Geological Sciences and Land Use and Environmental Change Institute (LUECI), University of Florida, Gainesville, FL 32611, USA

²Department of Geology, University of South Florida, Tampa, FL 33620, USA

³Geological Institute, Swiss Federal Institute of Technology ETH, Zurich, Switzerland

⁴Institut Forel and Department of Geology, University of Geneva, Geneva, Switzerland

⁵School of Natural Resources and Environment and Land Use and Environmental Change Institute (LUECI), University of Florida, Gainesville, FL 32611, USA

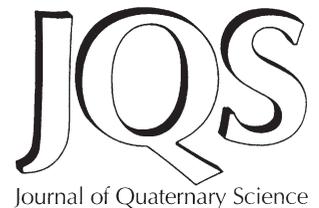
⁶Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, Livermore, CA 94551, USA

⁷National Lacustrine Core Repository and Limnological Research Center, University of Minnesota, Minneapolis, MN 55455, USA

Hillesheim, M. B., Hodell, D. A., Leyden, B. W., Brenner, M., Curtis, J. H., Anselmetti, F. S., Ariztegui, D., Buck, D. G., Guilderson, T. P., Rosenmeier, M. F. and Schnurrenberger, D. W. 2005. Climate change in lowland Central America during the late deglacial and early Holocene. *J. Quaternary Sci.*, **Vol. 20** pp. 363–376. ISSN 0267-8179.

Received 11 August 2004; Revised 8 February 2005; Accepted 10 February 2005

ABSTRACT: The transition from arid glacial to moist early Holocene conditions represented a profound change in northern lowland Neotropical climate. Here we report a detailed record of changes in moisture availability during the latter part of this transition (~11 250 to 7500 cal. yr BP) inferred from sediment cores retrieved in Lake Petén Itzá, northern Guatemala. Pollen assemblages demonstrate that a mesic forest had been largely established by ~11 250 cal. yr BP, but sediment properties indicate that lake level was more than 35 m below modern stage. From 11 250 to 10 350 cal. yr BP, during the Preboreal period, lithologic changes in sediments from deep-water cores (>50 m below modern water level) indicate several wet–dry cycles that suggest distinct changes in effective moisture. Four dry events (designated PBE1–4) occurred centred at 11 200, 10 900, 10 700 and 10 400 cal. yr BP and correlate with similar variability observed in the Cariaco Basin titanium record and glacial meltwater pulses into the Gulf of Mexico. After 10 350 cal. yr BP, multiple sediment proxies suggest a shift to a more persistently moist early Holocene climate. Comparison of results from Lake Petén Itzá with other records from the circum-Caribbean demonstrates a coherent climate response during the entire span of our record. Furthermore, lowland Neotropical climate during the late deglacial and early Holocene period appears to be tightly linked to climate change in the high-latitude North Atlantic. We speculate that the observed changes in lowland Neotropical precipitation were related to the intensity of the annual cycle and associated displacements in the mean latitudinal position of the Intertropical Convergence Zone and Azores–Bermuda high-pressure system. This mechanism operated on millennial-to-submillennial timescales and may have responded to changes in solar radiation, glacial meltwater, North Atlantic sea ice, and the Atlantic meridional overturning circulation (MOC). Copyright © 2005 John Wiley & Sons, Ltd.



KEYWORDS: Central America; Neotropics; palaeoclimate; lake sediments; geochemistry; stable isotopes; pollen; Holocene; Preboreal; deglaciation; Intertropical Convergence Zone; 8.2-kyr Event.

Introduction

The northern lowland Neotropics encompasses the area of the New World with elevations < 500 m and extends north from the equator to the Tropic of Cancer. It includes areas of south-

ern Mexico, Central America, and northern South America, as well as the West Indies (Fig. 1(A)). During the last deglaciation, this region experienced a pronounced increase in moisture availability. At the onset of deglaciation, all shallow lakes were dry and only deep basins contained water. The Petén Lake District of northern Guatemala (Fig. 1(B)) is one of the few regions with basins deep enough to have retained water during the arid late Pleistocene (Deevey *et al.*, 1980). Cores recovered from Lakes Quexil ($z_{\max} = 32$ m) and Salpetén ($z_{\max} = 32$ m) in 1980 provided some of the first insights into climatic and ecological change in the lowland Neotropics during the last

*Correspondence to: Michael B. Hillesheim, Department of Geological Sciences and Land Use and Environmental Change Institute (LUECI), University of Florida, Gainesville, FL 32611, USA. E-mail: mbhilles@ufl.edu

†Department of Geology and Planetary Science, University of Pittsburgh, Pittsburgh, PA 15260, USA (current address).

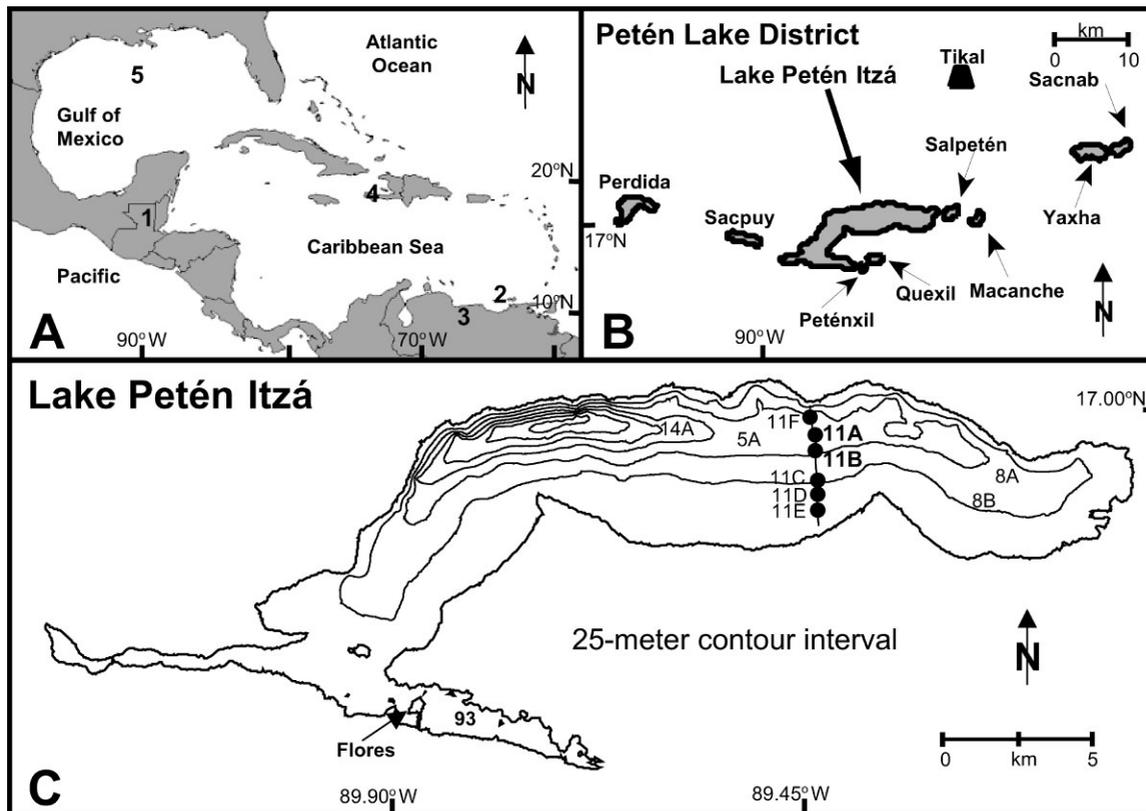


Figure 1 (A) Map of the northern lowland Neotropics and Gulf of Mexico showing site locations discussed in this study: 1, Petén Lake District, northern Guatemala; 2, Cariaco Basin, offshore northern Venezuela; 3, Lake Valencia, Venezuela; 4, Lake Miragoane, Haiti; 5, Louisiana slope, northern Gulf of Mexico. (B) Map of the Petén Lake District. (C) Bathymetric map of Lake Petén Itzá with the locations of cores collected along seismic line 11 in 2002. Black dots (●) indicate cores used in this study and the black line (—) is the location of seismic line 11. Cores along other seismic lines in the northern basin are also indicated (5A, 8A, 8B, 14A). 93 refers to the location of Core PI 6-VII-93 (Curtis *et al.*, 1998) taken from Petén Itzá's southern basin east of the island of Flores

Ice Age and the transition to the Holocene (Deevey *et al.*, 1983). Pollen and geochemical profiles demonstrated an abrupt shift from cool, arid conditions during the last glaciation to warmer, moister conditions during the early Holocene (Leyden *et al.*, 1993, 1994; Brenner, 1994). Because of incomplete recovery, drilling disturbance and poor dating resolution of the 1980 cores, details of the timing, rate and structure of this transition were uncertain.

In 1999, we completed a seismic survey (3.5 kHz) of six Petén lakes to identify potential targets for coring and drilling. We produced the first detailed bathymetric map of Lake Petén Itzá (Fig. 1(C)), the largest (100 km²) and deepest ($z_{\max} = 160$ m) lake in the region. Seismic profiles revealed a palaeoshoreline feature at ~58 m below modern lake level that possibly represents the lake lowstand during the last glacial period (Anselmetti *et al.*, in review). Because of the great depth of its northern basin, Lake Petén Itzá probably contains one of the few continuous sediment records in Central America of the last glaciation. Consequently, it is a prime target for future drilling operations to obtain long, undisturbed sediment cores for palaeoclimate studies.

In June 2002, we retrieved 13 Kullenberg-type piston cores from 10 sites in Petén Itzá's northern basin (Fig. 1(C)). Undisturbed glacial-age sediment was not recovered because of the thickness of cohesive late Holocene sediments, which limited penetration of the piston corer to ≤ 6 m. The oldest continuous sediment record recovered is dated to 11 250 cal. yr BP, and captures only the terminal part of the last deglaciation. Here we report on the palaeoclimatic history of the Petén lowlands during the late deglacial and early Holocene period (~11 250 to 7500 cal. yr BP). We compare our results from Lake Petén Itzá with proxy records from other sites under the

same climate regime, such as the marine Cariaco Basin off Venezuela (Haug *et al.*, 2001), as well as with the climate history of the high-latitude North Atlantic recorded in marine sediments and Greenland ice cores (Bond *et al.*, 2001; Stuiver *et al.*, 1995). We find a coherent response of climate throughout the circum-Caribbean region during the late deglaciation and early Holocene with tight linkages to climate changes observed in the high-latitude North Atlantic.

Study site

Lake Petén Itzá is located at ~17.00°N, 89.50°W in the Department of Petén, northern Guatemala (Fig. 1(B)). The lake is comprised of two connected basins. The deeper, northern basin ($z_{\max} = 160$ m) occupies a large half-graben formed by a series of east–west aligned *en echelon* faults (Vinson, 1962). The northern shore is marked by a steep karst ridge consisting of lower Tertiary limestone that follows the strike of the fault system, whereas the southern shore is gently sloping and, in places, rimmed by poorly drained seasonal swamps (*bajos*). The smaller southern basin is much shallower, averaging ~5 m water depth.

Lake Petén Itzá's water is dilute (11.22 meq l⁻¹) and dominated by calcium and bicarbonate, with magnesium and sulfate following closely in concentration (Table 1). Lakewater pH is high (~8.0) and is saturated for calcium carbonate, and there are abundant shells of carbonate microfossils in the lake sediments (Covich, 1976; Curtis *et al.*, 1998). Lake Petén Itzá is a terminal basin fed by precipitation, subsurface groundwater inflow, and a small input stream in the southeast. The

Table 1 Mean ion concentrations ($n=24$) and oxygen isotopic composition ($\delta^{18}\text{O}_{\text{lw}}$) V-SMOW of Petén Itzá lakewater ($n=34$) from samples collected at various sites and water depths in June and August 2002. Oxygen isotopic composition of precipitation ($\delta^{18}\text{O}_{\text{p}}$) interpolated from circum-Caribbean stations in the GNIP database (IAEA, 2001). $\delta^{18}\text{O}_{\text{lw}}$ was measured on a VG Prism II mass spectrometer using a modified method from Socki *et al.* (1992). Ion concentrations were measured on a Dionex Model DX 500 chromatograph. Bicarbonate was calculated as the sum of the cation charges minus the sum of chloride and sulfate charges

Ion	Concentration (meq l ⁻¹)	$\delta^{18}\text{O}_{\text{lw}}$ ($n=34$; ‰)	$\delta^{18}\text{O}_{\text{p}}$ (‰)
Ca ²⁺	3.15	+2.9‰	-4.0‰
Mg ²⁺	1.86		
Na ⁺	0.60		
Cl ⁻	0.26		
SO ₄ ²⁻	2.11		
HCO ₃ ⁻	3.24		
Total	11.22		

basin is effectively closed, lacking any surface outlets, although some downward leakage may occur. The oxygen isotopic composition of lakewater averages 2.9‰, which is enriched by ~7‰ relative to regional groundwater (Table 1). This reflects the importance of evaporation on the lake's water budget. Thermal stratification is persistent year round with hypolimnetic temperatures averaging ~25.4 °C, close to the mean annual air temperature (Fig. 2(A)). Hypolimnetic oxygen depletion has been observed near the deepest point in the lake (Fig. 2(B)).

Petén's climate is marked by uniformly high temperatures (averaging 25 °C), and variability is most notably expressed by temporal and spatial changes in rainfall. Precipitation in Petén varies from 900 to 2500 mm yr⁻¹, with a regional mean of ~1600 mm yr⁻¹ (Deevey *et al.*, 1980). Lake Petén Itzá is situated in a climatically sensitive region where the amount of rainfall is related to the seasonal migration of the Intertropical Convergence Zone (ITCZ) and Azores–Bermuda high-pressure system (Hastenrath, 1984). Approximately 90% of the precipitation falls during the summer months (June to October) driven by increased convection associated with the

northward migration of the ITCZ. The rainy season is often interrupted by a slight rainfall decrease in July and August, which is known as the canicula or 'little dry' (Magaña *et al.*, 1999). The rainy season usually ends by late October and is followed by a pronounced dry season during the winter (January to May) as the ITCZ moves equatorward (i.e. southward), allowing the Azores–Bermuda high to dominate the Gulf of Mexico and Caribbean Sea region. The region is also marked by subsidence during winter related to the descending limb of the Hadley Cell, centred at ~20°N (Waliser *et al.*, 1999).

Lake Petén Itzá's volume is sensitive to precipitation changes and has fluctuated markedly in the recent past. For example, mean annual rainfall during the period from 1934 to 1942 was relatively high (2055 mm yr⁻¹) and resulted in increased lake levels and flooding (Deevey *et al.*, 1980). In contrast, the early to mid-1970s were relatively dry (mean annual rainfall = 1415 mm yr⁻¹) resulting in lower lake levels. During the late 1970s lake level rose again in response to increased precipitation continuing until the early 1990s at which time the trend reversed. Petén Itzá's level varies seasonally, by as much as 80 cm, displaying a rise after the summer rainy season and a fall during the dry season with a 1–2 month lag (Deevey *et al.*, 1980). On seasonal to decadal timescales, lake volume and water chemistry are sensitive to changes in the balance between precipitation and evaporation.

Methods

In June 2002, we obtained six piston cores (Table 2) along a depth transect on seismic line 11 in water depths ranging from 9.7 to 63.2 m (Fig. 3). Cores were recovered using a Kullenberg-type piston corer triggered by a mud–water interface (MUCK) corer. In the laboratory, each core section was measured in its original polycarbonate liner for magnetic susceptibility and gamma-ray attenuation (GRA) bulk density at 0.5-cm increments using a GEOTEK Multi-Sensor Core Logger (MSCL). The instrument was calibrated at the start of each day using an aluminium standard. After logging, cores were split and imaged with a GEOTEK digital color linescan camera that was calibrated each day using a white ceramic calibration tile.

The primary focus of this study is on deep-water cores PI 8-VI-02 11A (water depth = 58.2 m), PI 5-VI-02 11B (water depth = 51.6 m), and shallow-water Core PI 6-VI-02 11D (water depth = 20.9 m), hereafter referred to as Cores 11A, 11B, and 11D. Core 11B was converted to the same sediment depth scale as Core 11A by correlating variations in colour (rgb) intensity (correlation coefficient, $r=0.84$) using Analyseries 1.1 (Paillard *et al.*, 1996). The working halves of Cores 11A and 11B were sampled at 1-cm and 5-cm intervals, respectively. Small subsamples (~5 cc) were taken for elemental

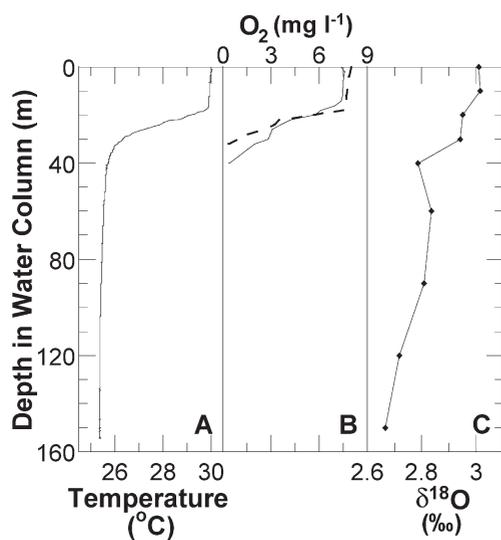


Figure 2 Water column profiles of (A) temperature (°C), (B) dissolved oxygen (mg l⁻¹), and (C) oxygen isotopic composition ($\delta^{18}\text{O}_{\text{lw}}$) V-SMOW of Petén Itzá lakewater. The temperature profile and samples for $\delta^{18}\text{O}$ analysis were collected on 13 August 2002 at 17.00°N, 89.85°W. Dissolved oxygen is plotted for May (solid line) and August (dashed line) 1980 (unpublished data)

Table 2 Length, water depth and location of cores retrieved along seismic line 11 in Lake Petén Itzá in June 2002. Cores are labelled by abbreviated lake name, day–month–year, and site

Core	Length (cm)	Depth (m)	Latitude (°N)	Longitude (°W)
PI 8-VI-02 11A	558	58.2	17.000	89.779
PI 5-VI-02 11B	515	51.6	16.998	89.779
PI 9-VI-02 11C	275	30.0	16.991	89.780
PI 6-VI-02 11D	210	20.9	16.988	89.779
PI 8-VI-02 11E	255	9.7	16.984	89.779
PI 9-VI-02 11F	585	63.2	17.004	89.780

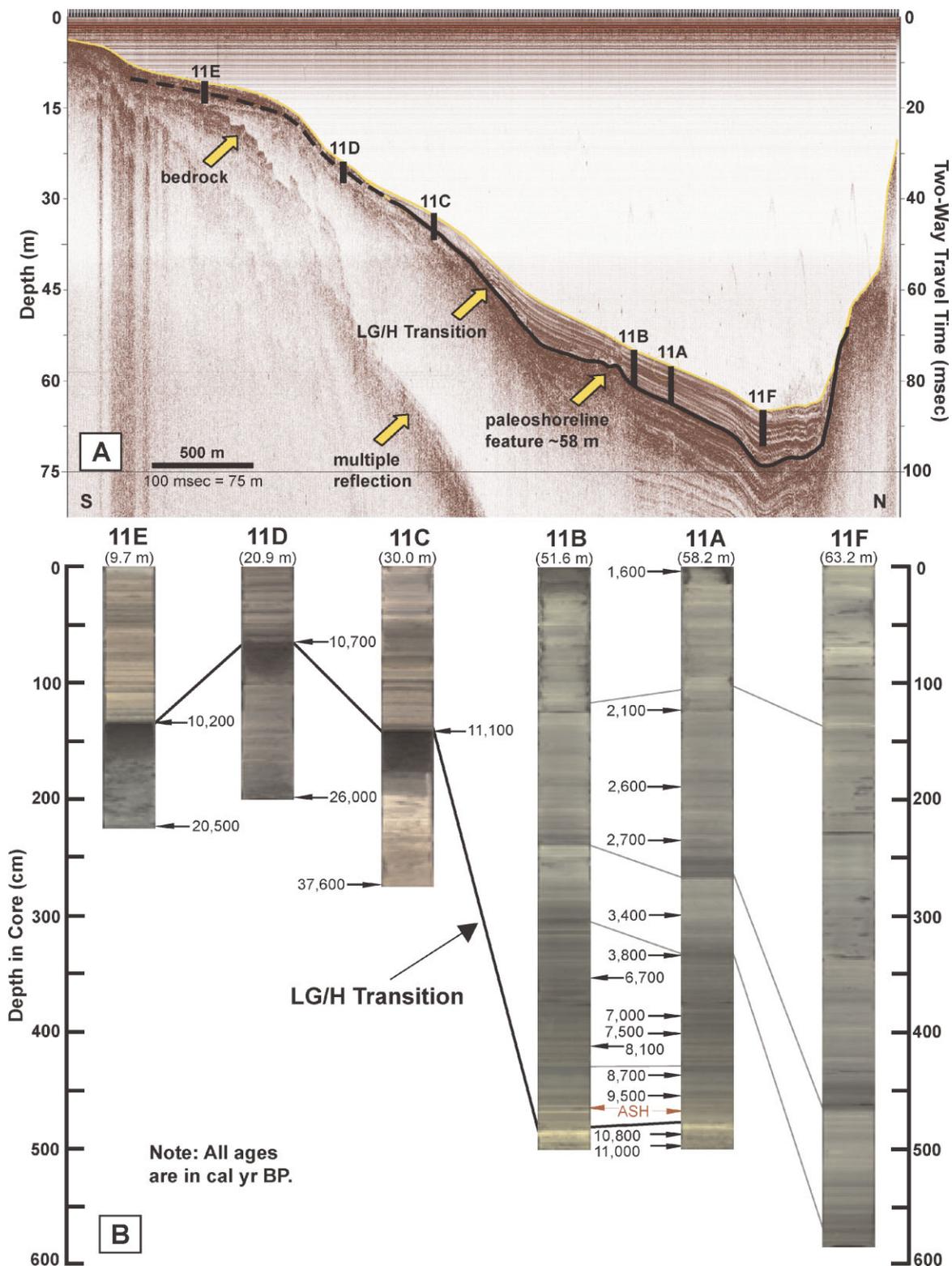


Figure 3 (A) High-resolution seismic image of line 11 showing the location of piston Cores 11A–11F (see also Table 2). Note the Late Glacial/Holocene (LG/H) sequence boundary (bold dashed–solid line) and the palaeoshoreline feature at ~58 m palaeo water depth. (B) Images of piston cores taken along a depth transect on seismic line 11. Water depths are relative to modern (2001) lake level. Dated horizons are indicated by arrows and inter-core correlations are marked by solid lines. The LG/H boundary (bold line) is marked by dark palaeosols in shallow-water Cores 11C, 11D, and 11E and gypsum precipitation in deeper-water Cores 11A and 11B. This figure is available in colour online at www.interscience.wiley.com/journal/jqs

analysis, oven-dried, and ground to a fine powder with a mortar and pestle. The remainder of the sample was wet-sieved at 63 μm to isolate microfossils and terrestrial organic material for isotopic and radiocarbon analysis. The three shallow-water cores are PI 9-VI-02 11C, PI 6-VI-02 11D, and PI 8-VI-02 11E (Table 2). Core 11D was selected as representative of the sediment stratigraphy in shallow-water cores (<35 m water depth)

and was sampled at 5-cm intervals for elemental composition and pollen analysis.

Radiocarbon ages were determined at the Center for Accelerator Mass Spectrometry (CAMS) at Lawrence Livermore National Laboratory. Eighteen samples of terrestrial organic matter (wood, leaf, or charcoal) and eight samples of aquatic gastropod (*Pyrgophorus* sp.) shell were dated. Samples of shell

material were used to date intervals where terrestrial organic matter was scarce. Samples of terrestrial organic matter used for radiocarbon dating were pretreated at the University of Florida using the acid–base–acid pretreatment method. Radiocarbon ages were determined after blank subtraction as determined on similarly pretreated and processed ^{14}C -free wood. Organic matter samples were also analysed for $\delta^{13}\text{C}$ content and corrected to -25‰ prior to calibration. Five samples from Core 11A were too small for $\delta^{13}\text{C}$ analysis and were assumed to have values of -25‰ . Calibrated ages, in calendar years before 1950 (cal. yr BP), were calculated using INTCAL98, with a 100-yr moving average of the tree ring calibration data set (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998).

Inorganic carbon (IC) was determined by coulometric titration using a UIC/Coulometrics 5011 coulometer coupled with a UIC 5240-TIC carbonate autosampler. Weight percent calcium carbonate (CaCO_3 , %) was calculated by multiplying IC by 8.33. Total weight percent carbon (TC) and sulphur (S, %) were measured using a Carlo Erba NA 1500 CNS elemental analyser with autosampler. Weight percent organic carbon (OC) was estimated by subtracting IC from TC. Weight percent organic matter (OM, %) was estimated by multiplying OC by 2.2.

Oxygen isotope ratios in Core 11A were measured on adult carapaces of the ostracod *Limnocythere* sp., which were isolated from the $>250\ \mu\text{m}$ sediment fraction. Ostracod valves were abundant from 558 to 500 cm, but gradually decreased above this level until they became absent at 484 cm. Ostracods were cleaned with 15% H_2O_2 to remove organic matter, and then washed with de-ionised water, rinsed in methanol, and dried. Ostracod aggregates of 15–20 valves were measured from each 1-cm sample. The samples were reacted in 100% orthophosphoric acid at 70°C using a ThermoFinnigan Kiel III

automated preparation system. Isotopic ratios of purified CO_2 gas were measured online with a ThermoFinnigan 252 mass spectrometer. All isotopic values are reported in standard delta (δ) notation relative to the VPDB standard. Analytical precision was estimated by measuring eight samples of a powdered carbonate standard (NBS-19) with each sample run. Precision for $\delta^{18}\text{O}$ was estimated to be $\pm 0.10\text{‰}$ (1σ ; $n = 24$ standards).

Samples for pollen analysis were processed using $1\ \text{cm}^3$ of sediment taken from Core 11B at 1-cm to 10-cm intervals and Core 11D at 5-cm intervals. One *Lycopodium* tablet (batch 124961; mean = 12 542; SD = ± 931) was added to each sample to calculate pollen concentration. Samples were processed using a standard sequence of HCL, KOH, HF and acetolysis. The residue was stained with basic fuchsin and suspended in tertiary butyl alcohol (TBA), and measured aliquots of the suspension were mounted in silicone oil on microscope slides. Pollen from Core 11D were counted at $400\times$ magnification to a minimum sum of 200, excluding pteridophytes, aquatic algae and macrophytes, and unidentifiable grains. For Core 11B, *Brosimum*-type pollen was also excluded from the sum to enhance the representation of the other taxa. Concentrations were converted to accumulation rates ($\text{grains cm}^{-1}\ \text{yr}^{-1}$) using the age model for Core 11B.

Results

Radiocarbon

All 13 radiocarbon samples from Core 11A yielded dates in stratigraphic order (Table 3). An age–depth model was derived

Table 3 Accelerator mass spectrometry (AMS) radiocarbon dates and calibrated ages of terrestrial organic and aquatic shell material from Lake Petén Itzá Cores 11A, 11B, 11C, 11D and 11E. All radiocarbon ages were corrected to $\delta^{13}\text{C}$ values of -25‰ , except leaf and wood samples (indicated with #), that were too small for $\delta^{13}\text{C}$ analysis and are assumed to be -25‰

Core	Sample type	Depth (cm)	CAMS number	$\delta^{13}\text{C}$ (‰)	Radiocarbon age (1σ , ^{14}C yr BP)	Calibrated age (2σ , cal. yr BP)	
PI 8-VI-02 11A	Leaf	75.5	99102	-23.31	1660 ± 40	1570 ± 60	
	Wood	190.5	99103	#	2095 ± 35	2070 ± 80	
	Wood	247.5	101228	#	2530 ± 60	2600 ± 150	
	Wood	297.5	99104	-22.27	2570 ± 45	2740 ± 35	
	Wood	353.5	92837	-30.30	3240 ± 30	3435 ± 55	
	Wood	369.5	99105	-31.68	3525 ± 25	3795 ± 75	
	Wood	446.5	101229	-23.63	6160 ± 70	7065 ± 180	
	Leaf	448.0	101230	#	6155 ± 50	7025 ± 135	
	Wood	463.5	99106	#	6585 ± 35	7470 ± 40	
	Wood	489.5	92836	-30.80	7925 ± 30	8720 ± 90	
	Wood	504.5	101231	-28.67	8425 ± 30	9475 ± 55	
	Wood	547.0	101232	#	9470 ± 110	10820 ± 340	
	Wood	552.0	101233	-22.70	9555 ± 30	11015 ± 75	
	PI 5-VI-02 11B	Shell	280.0	94597	-2.69	3970 ± 40	4440 ± 90^a
		Charcoal	365.0	94606	-27.12	5930 ± 40	6730 ± 70
Wood		433.0	92835	-28.30	7330 ± 40	8105 ± 85	
PI 9-VI-02 11C	Charcoal	140.0	94607	-15.67	9665 ± 45	11120 ± 60	
	Shell	140.0	94598	0.47	13810 ± 45	16580 ± 480^a	
	Shell	270.0	94599	-0.56	32420 ± 290	$37635 \pm 315^{a,b}$	
PI 6-VI-02 11D	Charcoal	75.0	94608	-17.92	9460 ± 45	10670 ± 110	
	Shell	75.0	94600	-1.50	11025 ± 40	13020 ± 150^a	
	Shell	140.0	94601	-1.85	19320 ± 70	22950 ± 730^a	
	Shell	205.0	94602	1.59	22110 ± 90	$26035 \pm 105^{a,b}$	
PI 8-VI-02 11E	Charcoal	165.0	94609	-12.71	9055 ± 50	10220 ± 70	
	Shell	165.0	94603	-3.68	10280 ± 40	12070 ± 310^a	
	Shell	250.0	94604	-3.50	17250 ± 60	20540 ± 650^a	

^a Ages are potentially subject to hard-water-lake error (Deevey and Stuiver, 1964).

^b Radiocarbon ages were too old for INTCAL98; therefore, ages were calibrated using the following equation from Bard *et al.* (1998): age (cal. yr BP) = $-3.0126 \times 10^{-6} (^{14}\text{C}$ age BP) $^2 + 1.2896 (^{14}\text{C}$ age BP) - 1005.

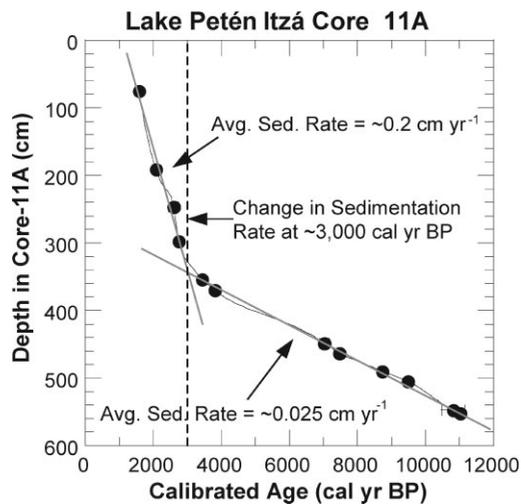


Figure 4 Calibrated age versus depth for Lake Petén Itzá Core 11A. Core chronology was established by linear interpolation between 13 calibrated radiocarbon dates measured on terrestrial wood and leaf samples. Average sedimentation rates vary between ~ 0.2 and ~ 0.025 cm yr^{-1} , with a transition in sedimentation rate occurring at ~ 3000 cal. yr BP. Dating error (2σ) on all samples is smaller than the plot symbols, except where indicated by error bar

by linear interpolation between dated horizons (Fig. 4). We estimated the basal age of Core 11A to be $\sim 11\,250$ cal. yr BP by extrapolating the linear sedimentation rate 6 cm beyond the lowermost radiocarbon date at 552 cm. Linear sedimentation rates (Fig. 4) average ~ 0.025 cm yr^{-1} from the base of the core at 558 cm to 335 cm (11 250 to 3000 cal. yr BP). Between 335 and 57 cm (3000 to ~ 1400 cal. yr BP), sedimentation rates increase to ~ 0.2 cm yr^{-1} associated with the onset of deposition of the Maya Clay unit. The temporal resolution averages ~ 40 yr for each 1-cm sample during the late deglacial and early Holocene part of the sequence (558 to 463 cm or 11 250 to 7500 cal. yr BP). Charcoal near the top of a soil horizon found in the three shallow-water cores (Table 3; Fig. 3(B)) was dated to constrain the onset of lacustrine deposition (i.e. flooding surface) associated with lake filling. Ages ranged from $11\,120 \pm 60$ cal. yr BP in Core 11C (31.8 m water depth) to $10\,220 \pm 70$ cal. yr BP in Core 11E (9.7 m water depth). In addition, three shell samples from the base of each shallow-water core were dated, yielding ages between 20 540 and 37 650 cal. yr BP (Table 3; Figure 3(B)).

Lithology and geochemistry

Shallow-water cores

Sediment properties of Core 11D (Fig. 5) represent the general pattern of lithologic and geochemical changes observed in the three shallow-water cores (< 35 m water depth). Pleistocene age sediments consist of dense, carbonate-rich silty clays overlain by a palaeosol that is marked by increased magnetic susceptibility (Fig. 5(D)), the presence of terrestrial gastropods and plant roots, and the absence of aquatic microfossils. The transition to lacustrine deposition, dated between 11 100 and 10 200 cal. yr BP (Fig. 3(B)), is marked by decreases in density and magnetic susceptibility concurrent with increases in percentage carbonate and organic matter content (Fig. 5), as well as the reappearance of lacustrine gastropods and ostracods.

Deep-water cores

The bottom 25 cm ($\sim 11\,250$ to 10 350 cal. yr BP) in both deep-water cores (11A, 11B) is composed of dense gypsum sands interbedded with silty clays. These lithologic changes are reflected in the calcium carbonate (CaCO_3), organic matter (OM), and sulphur (S) records (%; Fig. 6(A)–(C)). Gypsum-rich intervals are indicated by high sulphur content centred at $\sim 10\,400$, 10 700, 10 900 and 11 200 cal. yr BP. The intervening silty-clay layers contain higher percentages of carbonate and organic matter and are centred at $\sim 10\,500$, 10 800 and 11 000 cal. yr BP. At 10 350 cal. yr BP, gypsum precipitation ceased and sediments became dominated by calcium carbonate and organic matter. Between 10 350 and 7500 cal. yr BP, $\% \text{CaCO}_3$ decreased from 60% to 10% and %OM increased slightly, averaging 14%, with the remainder of the sediment being composed mostly of inorganic clay. Superimposed upon these long-term trends are small, short-term variations. At 9700 cal. yr BP, $\% \text{CaCO}_3$ decreased by 25% and %OM increased by $\sim 10\%$. From 9700 to 8800 cal. yr BP, $\% \text{CaCO}_3$ and %OM remained relatively constant averaging 25% and 14%, respectively. At 8800 cal. yr BP, $\% \text{CaCO}_3$ began to decrease more rapidly, coincident with a distinct increase in %OM. This trend is interrupted at 8100 cal. yr BP by a $\sim 10\%$ decrease in organic matter and a $\sim 5\%$ increase in carbonate, relative to the bulk sediment, that persisted until 7800 cal. yr BP when both proxies returned to pre-excursion values.

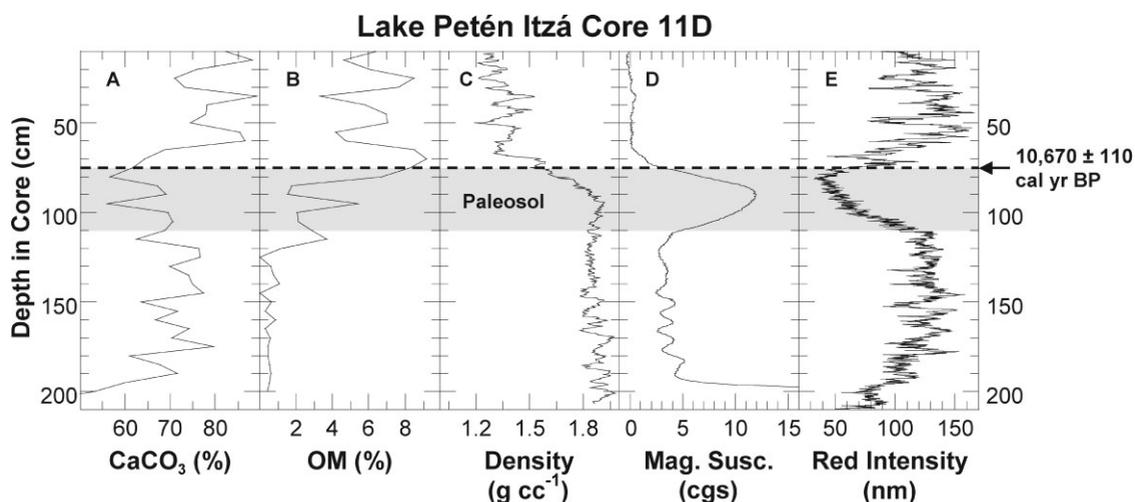


Figure 5 Sediment variables measured in Lake Petén Itzá Core 11D. (A) Weight percent calcium carbonate (CaCO_3 , %), (B) weight percent organic matter (OM, %), (C) GRA bulk density (g cc^{-1}), (D) magnetic susceptibility (cgs), (E) red colour intensity (nm). The dashed line denotes the dated lithologic transition from palaeosol to lacustrine deposition. The approximate thickness of the palaeosol is highlighted in grey

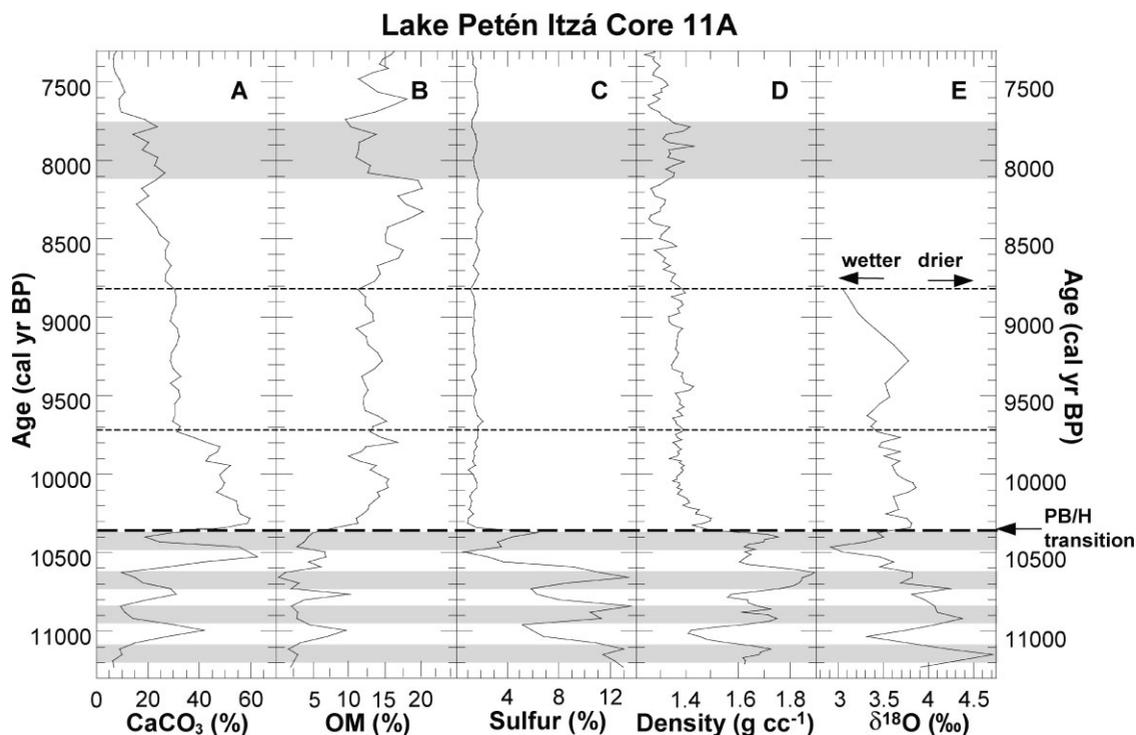


Figure 6 Sediment variables measured in Lake Petén Itzá Core 11A. (A) Weight percent calcium carbonate (CaCO_3 , %), (B) weight percent organic matter (OM, %), (C) weight percent sulphur (S, %), (D) GRA bulk density (g cc^{-1}), (E) oxygen isotopic composition ($\delta^{18}\text{O}$, ‰) of ostracods. Distinct changes in sediment composition are highlighted in grey

GRA bulk density follows lithologic changes between 11 250 and 10 350 cal. yr BP. High density is associated with gypsum beds and low density with clay-rich layers (Fig. 6(D)). At 10 350 cal. yr BP, density decreased abruptly followed by progressively lower values until the end of the record at 7500 cal. yr BP. This trend was interrupted, however, by a small (0.1 g cc^{-1}), short-lived (~ 100 yr) increase centred at 9400 cal. yr BP and a larger (0.2 g cc^{-1}), longer-lived (~ 300 yr) increase centred at ~ 8000 cal. yr BP.

Ostracods from the base of Core 11A record the greatest $\delta^{18}\text{O}$ values averaging 4.5‰ (Fig. 6(E)). Between $\sim 11\,250$ and $10\,400$ cal. yr BP, $\delta^{18}\text{O}$ generally decreased by $\sim 1.3\text{‰}$. Superimposed upon this trend are two distinct $\delta^{18}\text{O}$ minima centred at $11\,000$ and $10\,500$ cal. yr BP. At $10\,400$ cal. yr BP, $\delta^{18}\text{O}$ values increased slightly and then gradually decreased until ~ 9600 cal. yr BP, at which point values begin to increase again. At 9300 cal. yr BP, this trend reversed and $\delta^{18}\text{O}$ reached its lowest values of $\sim 3.1\text{‰}$ at 8800 cal. yr BP. Ostracods are absent in the sediment profile after this time.

Pollen

Pollen is poorly preserved prior to $\sim 11\,500$ cal. yr BP (86 cm) in Core 11D (Fig. 7(A)). Within the soil horizon (75 to 115 cm), *Pinus* (pine) predominates together with *Quercus* (oak), Asteraceae (composites) and Poaceae (grasses), but pollen concentrations are low. A small increase in aquatic taxa occurs as lacustrine deposition begins at $10\,670$ cal. yr BP (75 cm). After which *Brosimum*-type becomes a dominant taxon, along with *Gymnanthes*, pine, oak, *Bursera* and grasses.

The pollen percentage profile for Core 11B (Fig. 7(B)) can be divided into three zones that correlate with the sedimentological and isotopic records, but typically lag behind them by ~ 100 yr based on the high-resolution sampling of pollen (i.e. 1-cm intervals) over the late deglacial-to-early Holocene tran-

sition. Total pollen accumulation rates in Core 11B were low prior to $10\,250$ cal. yr BP, less than one-third of the values calculated for later deposits. Subsequent accumulation rates follow the percentage trends. *Brosimum*-type predominates in the basal zone from $11\,250$ to $10\,250$ cal. yr BP together with *Bursera*, *Ulmus*, *Gymnanthes*, pine and oak. Percentages for *Liquidambar* (sweet gum), *Typha* (cattail) and Asteraceae are notable. From $10\,250$ to 8250 cal. yr BP *Brosimum*-type, other Moraceae, *Celtis* and Melastomataceae/Combretaceae increase while *Liquidambar*, oak and, to a lesser extent, *Gymnanthes* decline. The aquatic alga *Botryococcus* increases sharply by 8800 cal. yr BP. After 8300 cal. yr BP *Brosimum*-type and *Gymnanthes* decrease and secondary forest taxa such as *Bursera* and *Cecropia* increase, followed later by increases in *Byrsonima*, grasses, and charcoal.

Interpretation of proxies

The oxygen isotopic ($\delta^{18}\text{O}$) composition of shell carbonate is influenced by temperature, $\delta^{18}\text{O}$ of lake water, and vital effects (i.e. taxon-specific fractionation) (Covich and Stuiver, 1974). Vital effects were minimised by measuring a single species of adult ostracod (*Limnocythere* sp.) from the $> 250 \mu\text{m}$ sediment fraction. Temperature change in the region during the Preboreal and early Holocene was probably small given the minor change in pollen assemblage relative to the last glacial, as well as the lack of cool indicator species such as *Juniperus* and *Liquidambar*. We therefore suggest that ostracod $\delta^{18}\text{O}$ was controlled principally by changes in the $\delta^{18}\text{O}$ of lakewater, which is related to the ratio of evaporation to precipitation (E/P) and the $\delta^{18}\text{O}$ of precipitation (Fontes and Gonfiantini, 1967; Gasse *et al.*, 1990). Between $11\,500$ and 8700 cal. yr BP, the $\delta^{18}\text{O}$ of western Caribbean seawater decreased by $\sim 0.4\text{‰}$ (Schmidt *et al.*, 2004), which should have decreased the $\delta^{18}\text{O}$

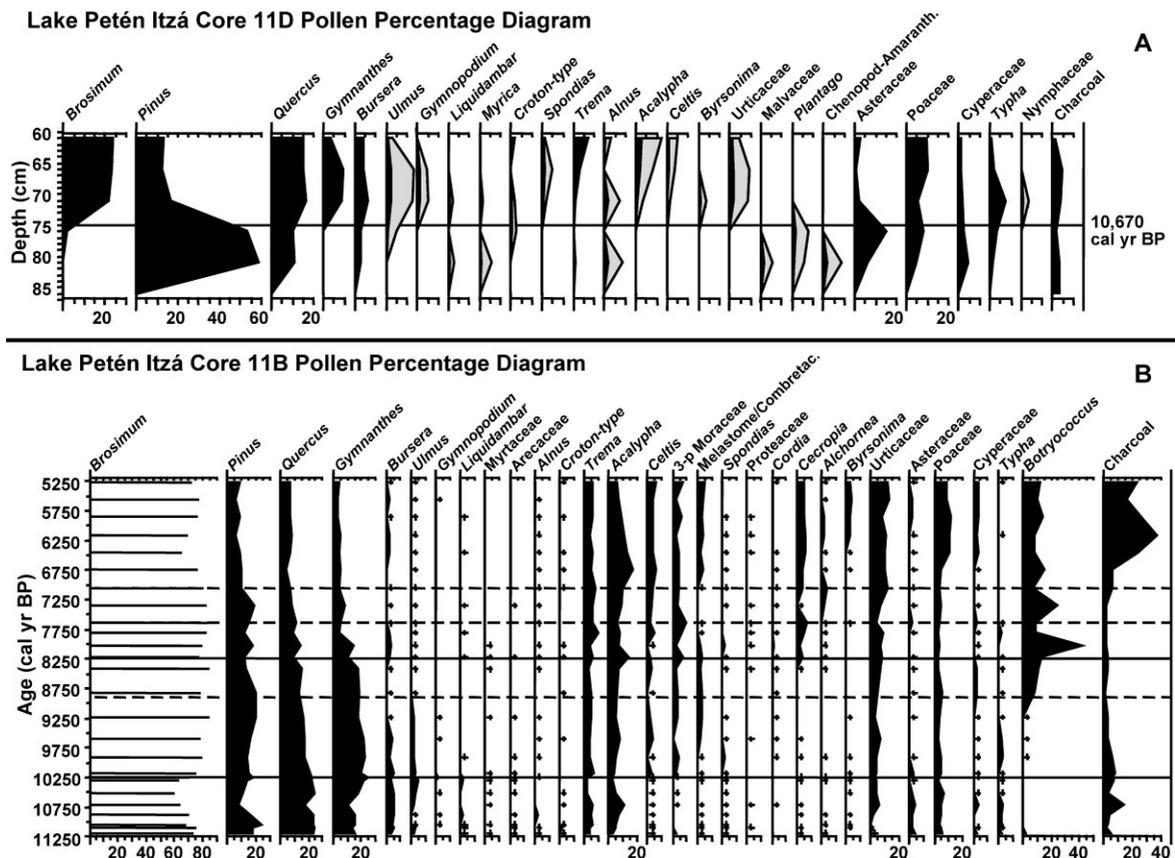


Figure 7 Pollen percentage diagrams from Lake Petén Itzá. (A) Shallow-water Core 11D and (B) deep-water Core 11B. For both cores, certain taxa (i.e. ferns, algae, aquatic vegetation, and unidentifiable pollen) were excluded from the pollen sum used to calculate percentages. Aquatic taxa were excluded so that the pollen sum reflects only terrestrial taxa. In addition, *Brosimum*-type was excluded from the pollen sum of Core 11B, in order to enhance the changes in the other taxa. Excluded taxa shown in the diagrams have percentages calculated using the restricted sum as their percentages have interpretative value. Grey curves in Core 11D represent a factor of five exaggeration of the original pollen percentages

of precipitation by the same amount because the Caribbean is the primary source of water vapour to the Yucatan Peninsula. The $\delta^{18}\text{O}$ of precipitation in the lowland tropics is also dependent on the amount of rainfall, with lower $\delta^{18}\text{O}$ values associated with greater precipitation (Rozanski *et al.*, 1993). Both the 'amount effect' and changes in E/P influence lakewater $\delta^{18}\text{O}$ in the same direction such that increased precipitation results in lower $\delta^{18}\text{O}$ values of both rainfall and lakewater. We conclude that variations in Lake Petén Itzá ostracod $\delta^{18}\text{O}$ represent relative changes in the ratio of precipitation to evaporation. Times of changing E/P are represented by changes in the slope of $\delta^{18}\text{O}$ and periods of unchanging $\delta^{18}\text{O}$ represent times of fairly constant E/P. We therefore compared the first derivative (i.e. changing slope) of the $\delta^{18}\text{O}$ signal with other proxy records (Fig. 8).

Changes in precipitation affect delivery rates of detrital material to the lake basin. Changing rainfall also alters the concentration of dissolved salts by enriching or diluting ions in lakewater. We used sediment lithology and chemical composition to infer past changes in detrital input and lakewater chemistry that are, in turn, controlled by climate. For example, greater gypsum content in sediments and associated increases in sediment density reflect lake volume reduction and higher lakewater salinity during dry periods, similar to the interpretation of the record from Lake Chichancanab, northern Yucatan Peninsula (Hodell *et al.*, 1995, 2001, 2005). Scanning electron micrographs (SEMs) of gypsum grains from Lake Petén Itzá show they are euhedral and have no evidence of rounding or abrasion that would indicate transport or redeposition. We suggest that gypsum formed authigenically in the littoral zone where the evaporation rate was higher than that in deeper, open water.

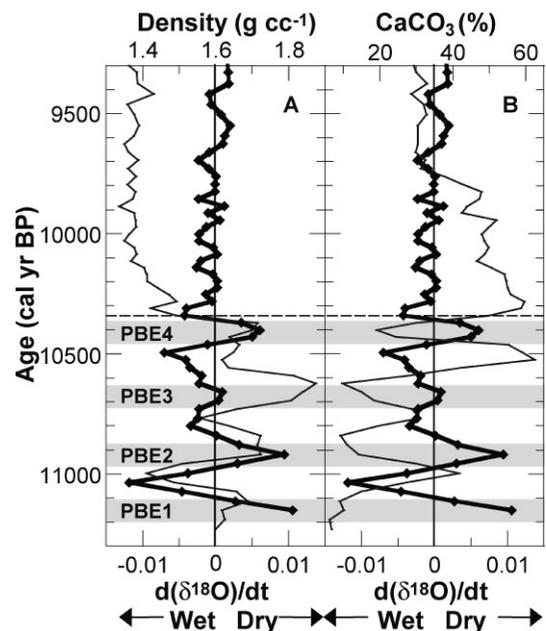


Figure 8 Comparison of the first-derivative of the detrended oxygen isotopic signal $d(\delta^{18}\text{O})/dt$ of ostracods from Petén Itzá Core 11A (\blacktriangle) with (A) density (g cc^{-1}) and (B) percentage carbonate (CaCO_3 , %). PBE1–4 designate dry events listed in Table 4 and are highlighted in grey

Discussion

Palaeoclimatic history of Lake Petén Itzá

Seismic profiles suggest that lake level was ~ 58 m below modern prior to 11 250 cal. yr BP (Fig. 3(A); Anselmetti *et al.*, in review), which required an $\sim 87\%$ (4.8 km^3) reduction in volume relative to today (Fig. 9). Shallow-water cores (< 35 m below modern lake level) contain a palaeosol indicating subaerial exposure. Cores from nearby, shallower Lakes Quexil and Salpetén also indicated lake-level lowering of at least 30–40 m during the last glacial period (Deevey *et al.*, 1983), which probably caused them to become ephemeral. Glacial-age vegetation consisted of sparse temperate thorn scrub and taxa such as *Juniperus*, indicating much cooler temperatures. One estimate suggests that temperatures were ~ 6.5 to 8°C lower than today (Leyden *et al.*, 1993, 1994).

During the last deglaciation, Petén climate was marked by an abrupt shift from arid to humid conditions, as inferred from multiple sediment proxies (Leyden *et al.*, 1993, 1994; Brenner, 1994). Pollen assemblages in shallow-water Core 11D (Fig. 7(A)) changed from *Pinus* to *Brosimum*-type dominance concomitant with the rise of other mesic taxa. This increase in mesic taxa correlates with a similar change in the pollen profile from nearby Lake Quexil that occurred after $\sim 12\,600$ cal. yr BP (Leyden *et al.*, 1993, 1994).

Deep-water cores from Lake Petén Itzá do not capture the entire deglacial sequence. Instead, they begin during the latter part of deglaciation, at $\sim 11\,250$ cal. yr BP. Predominance of *Brosimum* at the base of Core 11B indicates a mesic tropical forest had largely been established in Petén by this time (Fig. 7(B)). This is in agreement with $\delta^{18}\text{O}$ and pollen records from Lake Quexil that indicate warmer and wetter conditions following the glacial period (Leyden *et al.*, 1993, 1994). Dated palaeosols in shallow-water Petén Itzá cores, however, indicate lake level was at least 35 m lower than today (Fig. 3(B)). This conclusion is supported by the presence of *Typha* pollen in deep-water cores (> 50 m) that indicates a nearby littoral zone with emergent vegetation (Fig. 7(B)).

From 11 250 to 10 350 cal. yr BP, during the Preboreal period, the $\delta^{18}\text{O}$ of Lake Petén Itzá ostracods in Core 11A decreased, indicating greater effective moisture (Fig. 6(E)). The lithologic change in shallow-water cores (< 35 m water

depth), from subaerial exposure (palaeosol) to subaqueous deposition (lacustrine sediments), marks a lake-level rise at this time (Figs 3(B) and 5). In deep-water Core 11A, variations in oxygen isotopes, density and sediment composition records (Fig. 6) show a number of abrupt changes that suggest varying effective moisture and lake-level instability. These changes represent a series of wet–dry cycles of ~ 250 yr duration. During the Preboreal, four distinct dry events (designated PBE1–4) are identified by lithologic changes (i.e., increased gypsum content) centred at 11 200, 10 900, 10 700, and 10 400 cal. yr BP (Fig. 8). Changes in $\delta^{18}\text{O}$, as expressed by the first derivative of the signal, are consistent with the lithologic changes, typically leading them by ~ 50 yr (Fig. 8).

Low temporal sampling resolution (~ 50 – 200 yr) for pollen over this interval precludes detailed interpretation of vegetation changes during this period. Average climate conditions at the beginning of the Preboreal were somewhat cooler and drier. But after $\sim 10\,700$ cal. yr BP, pine, oak, *Liquidambar* and *Gymnanthes* begin to decline while *Acalypha*, *Trema* and charcoal increase (Fig. 7(B)). *Liquidambar* reflects cooler conditions and *Gymnanthes* prefers drier, open conditions. *Acalypha* and *Trema* are common tropical secondary or successional taxa. Although percentages for *Brosimum* do not increase, the accumulation rate is slightly higher. During this interval, pollen changes suggest moister, and perhaps warmer, climatic conditions, in general agreement with other proxy indicators.

An important change in Lake Petén Itzá's hydrology occurred at $\sim 10\,350$ cal. yr BP, coinciding with the Preboreal/Boreal boundary observed in northern European pollen records. At this time, an increase in effective moisture is inferred from sharp decreases in density and sulphur content that indicate cessation of gypsum precipitation (Fig. 6(C) and (D)) concomitant with increases in carbonate and organic matter content (Fig. 6(A) and (B)). A small increase in the abundance of *Brosimum*-type pollen lags this transition by ~ 100 yr (Fig. 7(B)). This represents the time at which Lake Petén Itzá's water chemistry fell below gypsum saturation and sediment lithology and density became less variable.

After 10 350 cal. yr BP, reduced variability of sediment composition and ostracod $\delta^{18}\text{O}$ values indicate a more persistent moist climate (Fig. 6). This is further supported by the resumption of lacustrine deposition at sites shallower than 8 m water depth (Fig. 3(B); Curtis *et al.*, 1998) indicating lake stage had risen to near present levels by 10 200 cal. yr BP (Fig. 9). Centred at 9400 cal. yr BP, small increases in density and $\delta^{18}\text{O}$ may suggest a brief (~ 100 yr) return to slightly drier conditions (Fig. 6(D) and (E)). At 8800 cal. yr BP, the abundance of aquatic algae (*Botryococcus*) increases (Fig. 7(B)) concomitant with an increase in organic matter content (Fig. 6(B)). This also marks the time when ostracods in deep-water cores disappear, indicating the onset of hypolimnetic anoxia, which may also account for better preservation of organic matter. Between 8100 and 7800 cal. yr BP, changes in sediment composition suggest a temporary return to drier conditions (Fig. 6(A)–(D)), after which humid climate conditions prevailed again.

From 10 250 to 8300 cal. yr BP, the pollen record also indicates a relatively moist, stable climate. The continued prominence of *Gymnanthes*, however, suggests that parts of the watershed were still edaphically dry or unstable, perhaps on some of the steep basin slopes. Given the limited sampling resolution (~ 200 yr) over this interval, the pollen record suggests drier conditions between 8100 and 7800 cal. yr BP, consistent with lithologic changes. This period also marks the beginning of fundamental changes in the vegetation, that include the introduction of other secondary taxa such as *Cecropia* that reflect a more diverse, open forest and increased seasonality (Fig. 7(B)).

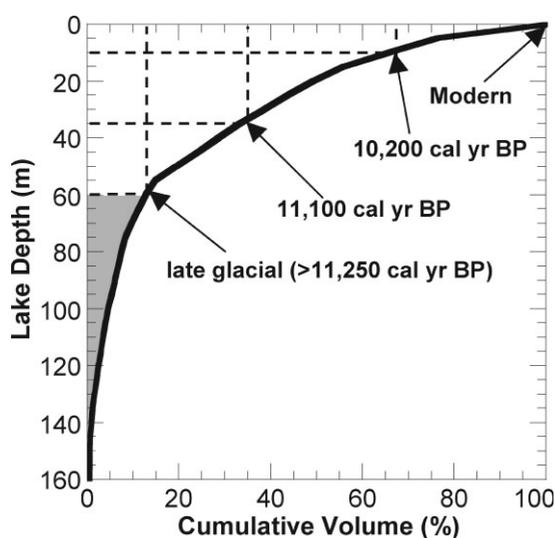


Figure 9 Hypsographic curve of cumulative lake volume versus water depth. Arrows indicate inferred lake levels based on flood surface ages

Comparisons with other circum-Caribbean and North Atlantic records

Here we compare our results with other marine and lake records from the circum-Caribbean and Gulf of Mexico, in order to assess the regional extent of palaeoclimatic changes observed in Lake Petén Itzá proxies. Previous studies of the last glacial-to-interglacial cycle have reported similarities between climate change in the Caribbean and the high-latitude North Atlantic on a variety of timescales (Hughen *et al.*, 1996; Peterson *et al.*, 2000). Consequently, we also compare the Petén Itzá record with proxy signals from the high latitudes of the northern hemisphere. When comparing Lake Petén Itzá proxies with other palaeoclimate records, each signal was plotted on its own independent timescale and no attempt was made to align features in the records.

There are relatively few lacustrine records of the last deglaciation in the circum-Caribbean because most shallow lakes remained dry well into the early Holocene. Lake Miragoane, Haiti, and Lake Valencia, Venezuela, are two lakes located on the northern and southern rim of the Caribbean Basin, respectively, that were deep enough to preserve records of the latter part of the last deglaciation (Bradbury *et al.*, 1981; Leyden, 1985; Hodell *et al.*, 1991; Curtis and Hodell, 1993; Brenner *et al.*, 1994; Higuera-Gundy *et al.*, 1999; Curtis *et al.*, 1999, 2001). Between 11 250 and 8800 cal. yr BP, oxygen isotopic records from Lakes Miragoane, Valencia and Petén Itzá are similar, showing large-scale fluctuations superimposed on a long-term trend of increasing moisture availability (Fig. 10). Several $\delta^{18}\text{O}$ maxima and minima can be correlated among the records within the uncertainty of the chronologies. The

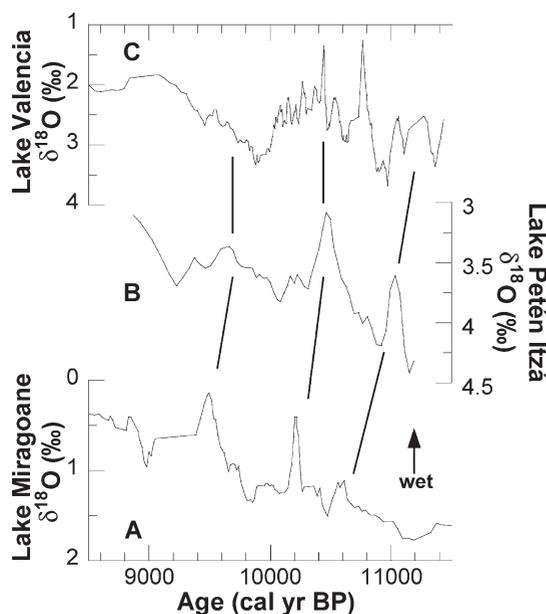


Figure 10 Oxygen isotopic ($\delta^{18}\text{O}$) records of ostracods from three circum-Caribbean lakes. (A) Lake Miragoane, Haiti, $\delta^{18}\text{O}$ (‰, 5-pt running average) of *Candona* sp. (Curtis and Hodell, 1993), (B), Lake Petén Itzá, $\delta^{18}\text{O}$ (‰, 3-pt running average) of *Limnocythere* sp., and (C) Lake Valencia, Venezuela, $\delta^{18}\text{O}$ (‰, 5-pt running average) of *Heterocypris* sp. (Curtis *et al.*, 1999). Radiocarbon dates (^{14}C yr BP) from Lakes Miragoane are based on carbonate microfossils (ostracods) that are susceptible to hard-water-lake error (Deevey and Stuiver, 1964), whereas the Lakes Valencia and Petén Itzá chronologies are based on terrestrial organic matter. Dates from the Lakes Miragoane and Valencia cores were calibrated to calendar years by the authors, in order to compare them with the Petén Itzá record. Each proxy is plotted on its own respective timescale

records from Lakes Valencia and Petén Itzá are the most reliably dated because the chronologies are based exclusively on radiocarbon analysis of discrete terrestrial macrofossils. In contrast, the Lake Miragoane chronology is based on radiocarbon analysis of carbonate microfossils with the assumption of a constant hard-water lake error. Despite the dating uncertainty in Lake Miragoane the pattern of oxygen isotope variation is similar to that from the other lakes, indicating several abrupt shifts in moisture availability on an otherwise long-term trend toward wetter conditions in the early Holocene.

The most detailed records of the last deglaciation and Holocene in the Caribbean come from studies of laminated sediments in the Cariaco Basin (11°N) off the coast of Venezuela (Hughen *et al.*, 1996, Peterson *et al.*, 2000, Haug *et al.*, 2001). In general, the density record from Lake Petén Itzá Core 11A and the titanium (Ti) record from Cariaco (Haug *et al.*, 2001) are remarkably similar (Fig. 11). Both sediment proxies are probably related indirectly to precipitation. In Cariaco, rainfall controls the input of terrigenous material to the basin. Low Ti values in Cariaco sediments have been interpreted to reflect drier climate and vice versa. In Lake Petén Itzá, detrital input and lakewater chemistry control sediment composition, which is related indirectly to precipitation.

The Petén Itzá record does not extend back to the time of the Bølling–Allerød or Younger Dryas events that are prominent in the Cariaco Ti signal, but rather begins in the Preboreal period when Cariaco Ti values were generally increasing but lower than those of the early Holocene. The late deglacial is marked by density variations in Core 11A that correspond to changes in Ti content of Cariaco Basin sediments. Within dating uncertainties, reductions in Cariaco Ti content correlate to intervals of increased density and gypsum content in Lake Petén Itzá (Fig. 11(B)), suggesting that the dry Preboreal events (PBE1–4) observed in the Petén Itzá record may have been pan-Caribbean in extent.

At the end of the Preboreal period ($\sim 10\,350$ cal. yr BP), both records show an abrupt shift that suggests increased precipitation (Fig. 11(A)). This transition is more pronounced in the Petén Itzá density record than in Cariaco Ti owing to the sudden cessation of gypsum deposition in the lake. During the early-to-middle Holocene, the two records track one another closely, and display two synchronous returns to inferred drier conditions (Fig. 11(A)). The first, centred at 9400 cal. yr BP was small and brief (~ 100 yr). The second, centred at 8000 cal. yr BP, was greater in magnitude and longer in duration (~ 300 yr). The latter event is also recorded in a speleothem from lowland Costa Rica, that suggests this was a time of protracted dry conditions associated with a weakening of the Central American monsoon (Lachniet *et al.*, 2004).

The Gulf of Mexico may have played an important role in regional climate during the last deglaciation because of meltwater input derived from the retreat of the Laurentide Ice Sheet. Many studies have focused on the early part of the last deglaciation (i.e. Bølling–Allerød/Younger Dryas transition) when meltwater was redirected from the Mississippi River to the St. Lawrence River into the North Atlantic (Clark *et al.*, 2001). A recent study, however, suggests that episodic meltwater floods to the Gulf of Mexico continued after the Younger Dryas period. Aharon (2003) identified four events during the Preboreal, when meltwater was rerouted south. These are centred at 9900, 9700, 9400 and 9100 ^{14}C yr BP. When converted to calendar years, these events are dated at 11 000, 10 700, 10 400 and 10 000 cal. yr BP (Table 4; Aharon, pers. comm.). Three of these discharge events correspond with the Preboreal events (PBE1–4) observed in the Petén Itzá density and $\delta^{18}\text{O}$ records (Fig. 8), as well as variations in Cariaco Ti (Fig. 11(B)). This may suggest that episodic meltwater discharge into the Gulf of Mexico

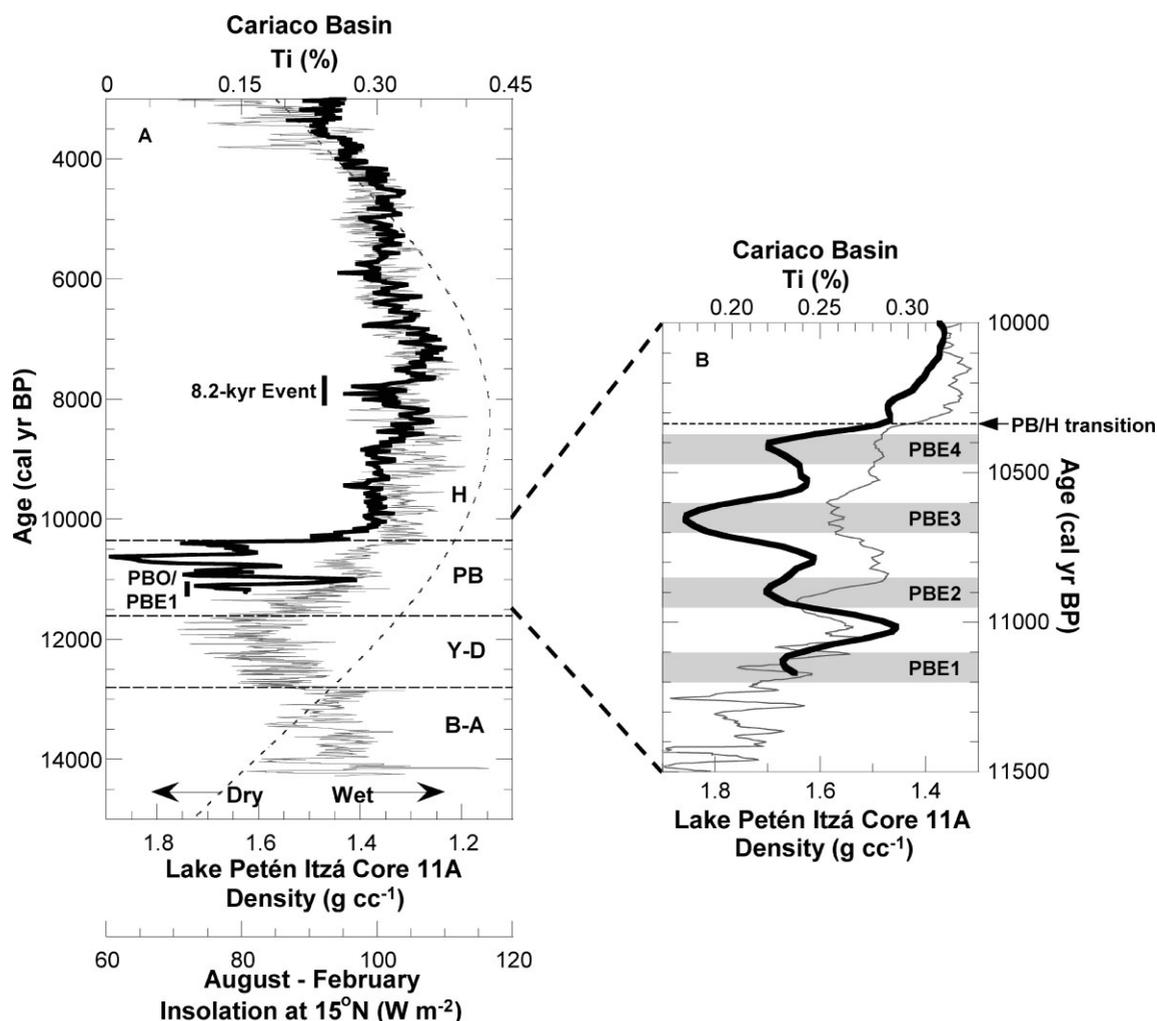


Figure 11 (A) Comparison of density in Petén Itzá Core 11A (black line; g cc⁻¹; inverted scale) and the 3-point running average of percentage titanium (grey line; Ti, %) from the Cariaco Basin (Haug *et al.*, 2001), with August minus February insolation (dashed line; W m⁻²) at 15°N latitude (Berger, 1978). Major cooling events observed in the North Atlantic such as the 8.2-kyr Event (Alley *et al.*, 1997) and Preboreal Oscillation (PBO) at ~11 300 cal. yr BP (Björck *et al.*, 1997) are highlighted. Major climate intervals are denoted as the Holocene (H), Preboreal (PB), Younger Dryas (Y-D), and Bølling-Allerød (B-A). Each record is plotted on its own respective timescale. Comparisons were not made younger than 3000 cal. yr BP because human disturbance of the Petén Itzá watershed began at this time. (B) Petén Itzá Core 11A density smoothed with a 5-pt running average and Cariaco Ti smoothed with a 15-pt running average to mute short-term variability. The Cariaco Ti record was shifted 100 yr younger by correlating the PB/H transition. This was done to better illustrate changes observed in the two records

during the Preboreal was sufficient to cause regional changes in E/P. Alternatively, the correlation between Caribbean drying and meltwater pulses to the Gulf of Mexico may represent

Table 4 Chronology of Preboreal events (PBE1–4) recorded in Lake Petén Itzá Core 11A compared to periods of increased meltwater delivery (MWF-5 a,c,e,g) to the Gulf of Mexico (Aharon, 2003) and the Preboreal Oscillation (PBO; Björck *et al.*, 1997)

Preboreal events (PBE)	Interval ^a (cal. kyr BP)	GOM and NA events	Interval ^b (cal. kyr BP)
4	10.31–10.46 (10.4)	MWF-5 g	9.75–10.07 (10.0)
3	10.66–10.80 (10.7)	MWF-5 e	10.23–10.43 (10.4)
2	10.88–11.01 (10.9)	MWF-5 c	10.69–10.79 (10.7)
1	11.15–> 11.25 (11.2)	MWF-5 a	10.97–11.07 (11.0)
		PBO	11.15–11.30

^a Numbers in parentheses indicate times of peak aridity in the Petén, inferred from $\delta^{18}\text{O}$ of ostracods.

^b Numbers in parentheses indicate maximum meltwater fluxes to the Gulf of Mexico.

responses to a common forcing, such as deglaciation in high northern latitudes.

Finally, we compared the Lake Petén Itzá $\delta^{18}\text{O}$ and density records with palaeoclimate changes in the high-latitude North Atlantic. Bond *et al.* (2001) used fluctuations in haematite-stained grains (HSG) to infer increases in drift ice associated with cool SST in the North Atlantic for the last glaciation and Holocene. From 11 500 to 8000 cal. yr BP, they recognised four distinct peaks in HSG abundance centred at 11 300, 10 300, 9400 and 8400 cal. yr BP (Fig. 12). The three older events correlate with inferred dry climate conditions observed in the Petén Itzá $\delta^{18}\text{O}$ record (Fig. 12). Furthermore, three dry events that are also observed as density increases in our record coincide with abrupt decreases in $\delta^{18}\text{O}$ values measured in the GISP2 ice core (Fig. 13; Stuiver *et al.*, 1995). The two most pronounced events, the Preboreal Oscillation (PBO) at ~11 300 cal. yr BP (Björck *et al.*, 1997), which corresponds to PBE1 (Table 4), and the 8.2-kyr Event (Alley *et al.*, 1997) have been interpreted as periods of significant cooling in the high-latitude North Atlantic. We, therefore, suggest that cool events in the high-latitude North Atlantic were associated with dry climate conditions in the northern lowland Neotropics.

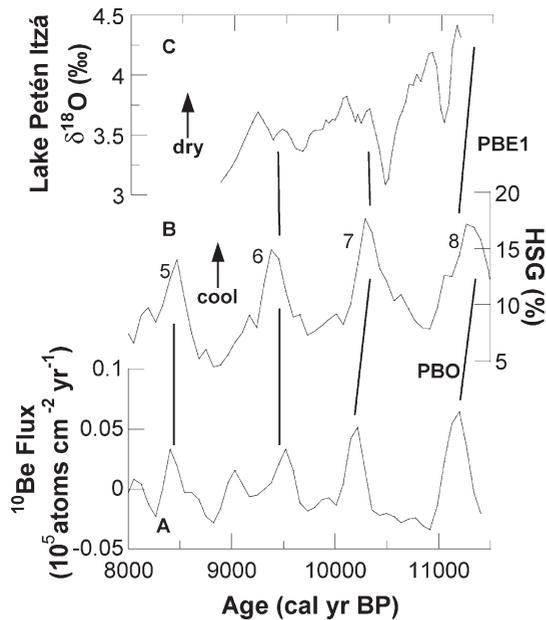


Figure 12 (A) ^{10}Be Flux (10^5 atoms $\text{cm}^{-2}\text{yr}^{-1}$, smoothed and detrended) from the GISP2 and GRIP ice core records, after Bond *et al.* (2001), (B) stacked plot of percentage haematite stained grains from North Atlantic cores (HSG, %; Bond *et al.*, 2001), and (C) oxygen isotopic composition ($\delta^{18}\text{O}$, ‰) of Petén Itzá ostracods, which was smoothed with a 3-pt running average to better illustrate long-term variability. Each record is plotted on its own respective timescale

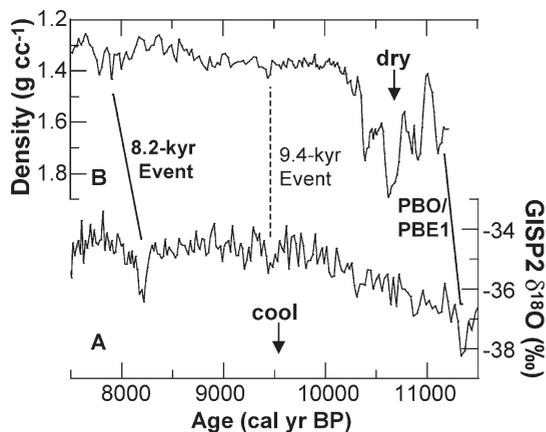


Figure 13 Comparison of (A) oxygen isotopic composition ($\delta^{18}\text{O}$, ‰) of ice from the GISP2 ice core (Stuiver *et al.*, 1995) with (B) density (g cc^{-1}) from Lake Petén Itzá Core 11A. Major cooling events observed in the North Atlantic such as the 8.2-kyr Event (Alley *et al.*, 1997) and Preboreal Oscillation (PBO) at $\sim 11\,300$ cal. yr BP (Björck *et al.*, 1997) are highlighted. Each proxy is plotted on its own respective timescale

Proposed mechanisms for climate change in the Preboreal and early Holocene

Precipitation proxies in the circum-Caribbean display a coherent response during the latter part of the last deglaciation and early Holocene. This is expected because the northern Neotropics are influenced by the same climate regime, where rainfall is related to the intensity of the annual cycle and associated displacement of the Atlantic ITCZ and Azores–Bermuda high-pressure system (Hodell *et al.*, 1991; Peterson *et al.*, 2000; Haug *et al.*, 2001). This mechanism has been shown to operate on short (annual) to long (millennial) timescales (Hodell *et al.*, 1991; Chang *et al.*, 1997; Peterson *et al.*, 2000; Haug *et al.*, 2001; Lea *et al.*, 2003; Black *et al.*, 2004;

Chiang *et al.*, 2003; Chiang, 2004). In turn, the mean latitudinal position of the ITCZ and Azores–Bermuda subtropical high may be influenced by numerous factors including: tropical Atlantic SST, seasonal insolation, glacial boundary conditions, thermohaline circulation (via cross-equatorial heat transport and sea-ice extent), solar irradiance and ocean–atmosphere interactions such as NAO and ENSO (Hodell *et al.*, 1991; Chang *et al.*, 1997; Peterson *et al.*, 2000; Haug *et al.*, 2001; Poore *et al.*, 2003; Chiang *et al.*, 2003; Chiang, 2004).

Long-term (glacial-to-interglacial) variations in Neotropical precipitation are affected by the seasonal distribution (August–February) of insolation related to variations in Earth's orbit, especially precession (Hodell *et al.*, 1991; Haug *et al.*, 2001). Precipitation in the northern Neotropics is enhanced when the seasonal insolation difference is high, and reduced when seasonality is low. Long-term changes in palaeoprecipitation proxies (density and $\delta^{18}\text{O}$) in Lake Petén Itzá cores generally follow the signal of seasonal insolation difference at 15°N (Fig. 11(A)), but orbital forcing cannot explain the abrupt changes that occurred during the latter part of the last deglaciation and early Holocene.

During the Preboreal, the general trend toward more humid conditions in the Petén was interrupted by four dry events (PBE1–4) centred at 11 200, 10 900, 10 700 and 10 400 cal. yr BP (Fig. 8). The oldest event (PBE-1) coincides with the Preboreal Oscillation (Björck *et al.*, 1997; van der Plicht *et al.*, 2004). The three younger events appear to correspond with episodic diversions of glacial meltwater to the Gulf of Mexico (Table 4; Aharon, 2003) that were related to changes in glacial boundary conditions (Clark *et al.*, 2001). A possible explanation for this phenomenon comes from modelling studies that impose a 6°C cooling of the Gulf of Mexico related to the delivery of cold, fresh glacial waters to the basin (Overpeck *et al.*, 1989; Oglesby *et al.*, 1989; Maasch and Oglesby, 1990). These model sensitivity tests imply that cooling Gulf of Mexico SST results in a westward shift of the Azores–Bermuda high to a position over the Intra-Americas Sea. This would, in turn, suppress convection and reduce precipitation in the region (Hastenrath, 1984). Meltwater discharge during the Preboreal, however, was significantly less than that used to infer a 6°C cooling of the entire Gulf of Mexico. Instead, meltwater fluxes were more comparable in magnitude to the largest of modern Mississippi River floods (Aharon, 2003), which probably only influenced part of the Gulf of Mexico. Alternatively, early Holocene deglacial events may have provided meltwater input to both the Gulf of Mexico and high-latitude North Atlantic, thereby causing cooling, an expansion of sea ice, and perhaps a reduction in the Atlantic meridional overturning circulation (MOC).

Two of the Preboreal events (PBE1 and PBE4) at 11 200 and 10 400 cal. yr BP, respectively, and an early Holocene event at 9400 cal. yr BP may correlate to increases in the $\delta^{18}\text{O}$ records of Lake Miragoane and Lake Valencia (Fig. 10; Curtis and Hodell, 1993; Curtis *et al.*, 1999). These dry events together with another early Holocene event centred at 8000 cal. yr BP, correspond within chronological uncertainty, to cooler temperatures inferred from low $\delta^{18}\text{O}$ values in the GISP2 ice core (Fig. 13; Stuiver *et al.*, 1995) and fluctuations in the abundance of haematite-stained grains (HSG) found in high-latitude North Atlantic cores (Fig. 12; Bond *et al.*, 2001). Bond *et al.* (2001) related fluctuations in HSG abundance with changes in the production rates of ^{10}Be and ^{14}C , such that increases in HSG abundance occurred during times of decreased solar irradiance and cold temperatures in the high-latitude North Atlantic. Increased aridity in the circum-Caribbean may have accompanied these cooling events, and may have been caused by a southward displacement of the ITCZ related to decreased solar

luminosity, increased sea ice or decreased thermohaline circulation. For example, the 8.2-kyr Event, which was the most prolonged and distinct cooling in the northern hemisphere since the Younger Dryas, may have been related to both reduced solar luminosity (Muscheler *et al.*, 2004) and a reduction in MOC from glacial meltwater discharge into the North Atlantic via Hudson Strait (Clark *et al.*, 2001). A similar mechanism has been proposed for the Preboreal Oscillation (PBO; Björck *et al.*, 1997; van der Plicht *et al.*, 2004).

Summary and conclusions

Seismic data indicate that prior to the last deglaciation, the level of Lake Petén Itzá was about ~58 m below modern stage. This observation is consistent with previous studies of Petén lakes that indicated arid climate conditions. The last deglaciation saw an abrupt shift from arid to humid conditions in the region. The age of our oldest deep-water core from Petén Itzá is 11 250 cal. yr BP, which captures only the terminal part of this arid-to-humid transition. Pollen assemblages from the base of the core demonstrate that a mesic forest had been largely established by ~11 250 cal. yr BP, but Petén Itzá's level was still > 35 m below modern. From 11 250 to 10 350 cal. yr BP during the Preboreal Period, cores from < 35 m below present water depth contain soil horizons, indicating subaerial exposure. During this period, deep-water cores from > 50 m below present water depth contain gypsum sands interbedded with carbonate-rich clays, indicating several dry-wet cycles. Oxygen isotopic ratios generally decrease during the Preboreal Period indicating a trend toward more mesic conditions. This trend was interrupted, however, by four Preboreal dry events (PBE1-4) centred at 11 200, 10 900, 10 700 and 10 400 cal. yr BP. The four events correlate to decreases in the Cariaco Ti record, suggesting they may have been widespread throughout the circum-Caribbean region. Three of the events coincided with episodic diversions of glacial meltwater to the Gulf of Mexico.

At 10 350 cal. yr BP, increased precipitation and lake-level rise is documented in deep-water cores by cessation of gypsum deposition and the onset of lacustrine sedimentation at shallow-water sites. Increased abundance of tropical mesic pollen taxa lagged this transition by ~100 yr. This event occurred at the Preboreal-to-Boreal transition and coincided with an inferred increase in precipitation recorded throughout the circum-Caribbean. The early Holocene was a time of generally reduced climate variability and steadily increasing moisture availability except for two returns to slightly drier conditions centred at 9400 and 8000 cal. yr BP. The prolonged dry event at 8000 cal. yr BP correlates with inferred reductions in circum-Caribbean precipitation as well as cooler temperatures in the high-latitude North Atlantic.

Comparison of the Lake Petén Itzá record with other lacustrine and marine records from the circum-Caribbean and Gulf of Mexico suggests that climate in this region responded coherently during the Preboreal and early Holocene periods. In addition, lowland Neotropical climate during this time was tightly linked to high-latitude climate change in the North Atlantic. Although the general trend in the northern lowland Neotropics was toward increasingly mesic conditions from ~11 250 to 7500 cal. yr BP, climate became drier at times of cooling in the high-latitude North Atlantic. We speculate that the observed changes in lowland Neotropical precipitation were related to the intensity of the annual cycle and associated displacements in the mean position of the ITCZ and Azores-Bermuda

subtropical high. This mechanism operated on a variety of timescales and responded sensitively to changes in solar radiation, glacial meltwater, North Atlantic sea ice extent and Atlantic MOC.

Acknowledgements We thank Liseth Perez, Laura Rodriguez and Jacobo Blijdenstein for field assistance, Gerard Bond for HSG and ¹⁰Be data, Paul Aharon for calibrating the LOUIS stacked-core age model, and Mike Binford for assisting with limnological data collection in 1980. We also thank Arturo Godoy and Roan McNab from the Wildlife Conservation Society (WCS), Margaret and Michael Dix from the Universidad del Valle, and the Consejo Nacional de Areas Protegidas (CONAP) for providing logistical assistance in the field. The Limnological Research Center (LRC) at the University of Minnesota supplied the coring system. Radiocarbon analyses were performed under the auspices of the U.S. Department of Energy by the University of California Lawrence Livermore National Laboratory (contract W-7405-ENG-48). Paul Baker and Larry Peterson provided thoughtful reviews that improved the manuscript. This work was supported by National Science Foundation (NSF) grant ATM-0117148 and is a publication of the University of Florida, Land Use and Environmental Change Institute (LUECI).

References

- Aharon P. 2003. Meltwater flooding events in the Gulf of Mexico revisited: implications for rapid climate changes during the last deglaciation. *Paleoceanography* **18**: DOI:10.1029/2002PA000840.
- Alley RB, Mayewski PA, Sowers T, Stuiver M, Taylor KC, Clark PU. 1997. Holocene climate instability: a prominent, widespread event 8,200 yr ago. *Geology* **25**: 483–486.
- Anselmetti FS, Ariztegui D, Hodell DA, Hillesheim MB, Brenner M, Gilli A, McKenzie JA. In review. Late Quaternary climate-induced lake level variations in Lake Petén Itzá, Guatemala, inferred from seismic stratigraphic analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology*.
- Bard E, Arnold M, Hamelin B, Tisnerat-Laborde N, Cabioch G. 1998. Radiocarbon calibration by means of mass spectrometric ²³⁰Th/²³⁴U and ¹⁴C ages of corals; an updated database including samples from Barbados, Mururoa, and Tahiti. *Radiocarbon* **40**: 1085–1092.
- Berger A. 1978. Long-term variations of daily insolation and Quaternary climate change. *Journal of the Atmospheric Sciences* **35**: 2362–2367.
- Björck S, Rundgren M, Ingólfsson O, Funder S. 1997. The Preboreal oscillation around the Nordic Seas: terrestrial and lacustrine responses. *Journal of Quaternary Science* **12**: 455–465.
- Black DE, Thunell RC, Kaplan A, Peterson LC, Tappa EJ. 2004. A 2000-year record of Caribbean and tropical North Atlantic hydrographic variability. *Paleoceanography* **19**: DOI:10.1029/2003PA000982.
- Bond G, Kromer B, Beer J, Muscheler R, Evans MN, Showers W, Hoffman S, Lotti-Bond R, Hajdas I, Bonani G. 2001. Persistent solar influence on North Atlantic climate during the Holocene. *Science* **294**: 2130–2136.
- Bradbury JP, Leyden BW, Salgado-Labouriau ML, Lewis WM, Schubert C, Binford MW, Frey DG, Whitehead DR, Weibezahn FH. 1981. Late Quaternary environmental history of Lake Valencia, Venezuela. *Science* **214**: 1299–1305.
- Brenner M. 1994. Lakes Salpetén and Quexil, Petén, Guatemala, Central America. In *Global Geological Record of Lake Basins*, Vol. 1, Kelts K, Gierlowski-Kordesch E (eds). Cambridge University Press: Cambridge, UK; 377–380.
- Brenner M, Curtis JH, Higuera-Gundy A, Hodell DA, Jones GA, Binford MW, Dorsey KT. 1994. Lake Miragoane, Haiti (Caribbean). In *Global Geological Record of Lake Basins*, Vol. 1, Kelts K, Gierlowski-Kordesch E (eds). Cambridge University Press: Cambridge, UK; 403–405.
- Chang P, Ji L, Li H. 1997. A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air-sea interactions. *Nature* **385**: 518–518.

- Chiang JCH. 2004. Present-day climate variability in the tropical Atlantic: a model for paleoclimate change? In *The Hadley Circulation: Past, Present, and Future*, Diaz HF, Bradley RS (eds). Cambridge University Press: Cambridge, UK; 465–488.
- Chiang JCH, Biasutti M, Battisti DS. 2003. Sensitivity of the Atlantic Intertropical Convergence Zone to Last Glacial Maximum boundary conditions. *Paleoceanography* **18**: DOI:10.1029/2003PA000916.
- Clark PU, Marshall SJ, Clarke GKC, Hostetler SW, Liccardi JM, Teller JT. 2001. Freshwater forcing of abrupt climate change during the last glaciation. *Science* **193**: 283–287.
- Covich AP, Stuiver M. 1974. Changes in oxygen 18 as a measure of long-term fluctuations in tropical lake levels and molluscan populations. *Limnology and Oceanography* **19**: 682–691.
- Covich AP. 1976. Recent changes in molluscan diversity of a large tropical lake (Lago de Petén, Guatemala). *Limnology and Oceanography* **21**: 51–59.
- Curtis JH, Hodell DA. 1993. An isotopic and trace element study of ostracods from Lake Miragoane, Haiti: a 10.5 kyr record of paleosalinity and paleotemperature changes in the Caribbean. In *Climate Change in Continental Isotopic Records, Geophysical Monograph 78*, Swart PK, Lohmann KC, McKenzie J, Savin S (eds). American Geophysical Union: Washington, DC; 135–152.
- Curtis JH, Brenner M, Hodell DA, Balser RA, Islebe GA, Hoogheemstra H. 1998. A multi-proxy study of Holocene environmental change in the Maya Lowlands of Petén, Guatemala. *Journal of Paleolimnology* **19**: 139–159.
- Curtis JH, Brenner M, Hodell DA. 1999. Climate change in Lake Valencia Basin, Venezuela, ~12 600 yr BP to present. *The Holocene* **9**: 609–619.
- Curtis JH, Brenner M, Hodell DA. 2001. Climate change in the circum-Caribbean (late Pleistocene-to-Present) and implications for regional biogeography. In *Biogeography of the West Indies*, Woods CA (ed.). CRC Press: Boca Raton, FL; 35–54.
- Deevey ES, Stuiver M. 1964. Distribution of natural isotopes of carbon in Linsley Pond and other New England lakes. *Limnology and Oceanography* **9**: 1–11.
- Deevey ES, Brenner M, Flannery MS, Yezdani GH. 1980. Lakes Yaxha and Sacnab, Petén, Guatemala: limnology and hydrology. *Archiv für Hydrobiologie* **57**: 419–460.
- Deevey ES, Brenner M, Binford MW. 1983. Paleolimnology of the Petén Lake District, Guatemala III: Late Pleistocene and Gamblian environments of the Maya area. *Hydrobiologia* **103**: 211–216.
- Fontes JC, Gonfiantini R. 1967. Component isotopique au cours de l'évaporation de deux bassins sahariens. *Earth and Planetary Science Letters* **3**: 258–266.
- Gasse F, Techet R, Durand A, Gilbert E, Fontes JC. 1990. The arid-humid transition in the Sahara and Sahel during the last deglaciation. *Nature* **346**: 141–146.
- Hastenrath S. 1984. Interannual variability and the annual cycle: mechanisms of circulation and climate in the tropical Atlantic sector. *Monthly Weather Review* **112**: 1097–1107.
- Haug GH, Hughen KA, Sigman DM, Peterson LC, Röhl U. 2001. Southward migration of the Intertropical Convergence Zone through the Holocene. *Science* **293**: 1304–1308.
- Higuera-Gundy A, Brenner M, Hodell DA, Curtis JH, Leyden BW, Binford MW. 1999. A 10 300 ¹⁴C-year record of climate and vegetation change from Haiti. *Quaternary Research* **52**: 159–170.
- Hodell DA, Curtis JH, Jones GA, Higuera-Gundy A, Brenner M, Binford MW, Dorsey KT. 1991. Reconstruction of Caribbean climate change over the past 10 500 years. *Nature* **352**: 790–793.
- Hodell DA, Curtis JH, Brenner M. 1995. Possible role of climate change in the collapse of the Maya civilization. *Nature* **375**: 391–394.
- Hodell DA, Brenner M, Curtis JH, Guilderson T. 2001. Solar forcing of the drought frequency in the Maya lowlands. *Science* **292**: 1367–1369.
- Hodell DA, Brenner M, Curtis JH. 2005. Terminal Classic drought in the Maya lowlands inferred from multiple sediment cores in Lake Chichancanab (Mexico). *Quaternary Science Reviews*. in press.
- Hughen KA, Overpeck JT, Peterson LC, Trumbore S. 1996. Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* **380**: 51–54.
- IAEA. 2001. IAEA/WMO global network of isotopes in precipitation: the GNIP Database. <http://isohis.iaea.org/> [1 July 2004].
- Lachniet MS, Asmerom Y, Burns SJ, Patterson WP, Polyak VJ, Seltzer GO. 2004. Tropical response to the 8200 yr B.P. cold event? Speleothem isotopes indicate a weakened early Holocene monsoon in Costa Rica. *Geology* **32**: 957–960.
- Lea DW, Pak DK, Peterson LC, Hughen KA. 2003. Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination. *Science* **302**: 1361–1364.
- Leyden BW. 1985. Late Quaternary aridity and Holocene moisture fluctuations in Lake Valencia Basin, Venezuela. *Ecology* **66**: 1279–1295.
- Leyden BW, Brenner M, Hodell DA, Curtis JH. 1993. Late Pleistocene climate in the Central American lowlands. In *Climate Change in Continental Isotopic Records, Geophysical Monograph 78*, Swart PK, Lohmann KC, McKenzie J, Savin S (eds). American Geophysical Union: Washington, DC; 165–178.
- Leyden BW, Brenner M, Hodell DA, Curtis JH. 1994. Orbital and internal forcing of climate on the Yucatan Peninsula for the past ca. 36 ka. *Palaeogeography, Palaeoclimatology, Palaeoecology* **109**: 193–210.
- Maasch KA, Oglesby RJ. 1990. Meltwater cooling of the Gulf of Mexico: a GCM simulation of climatic conditions at 12 ka. *Paleoceanography* **5**: 977–996.
- Magaña V, Amador JA, Medina S. 1999. The midsummer drought over Mexico and Central America. *Journal of Climate* **12**: 1577–1588.
- Muscheler R, Beer J, Vonmoss M. 2004. Causes and timing of the 8200 yr BP event inferred from comparison of the GRIP ¹⁰Be and the tree ring $\Delta^{14}C$ record. *Quaternary Science Reviews* **23**: 2101–2111.
- Oglesby RJ, Maasch KA, Saltzman B. 1989. Glacial meltwater cooling of the Gulf of Mexico: GCM implications for Holocene and present-day climates. *Climate Dynamics* **3**: 115–133.
- Overpeck JT, Peterson LC, Kipp N, Imbrie J, Rind D. 1989. Climate change in the circum-North Atlantic region during the last deglaciation. *Nature* **338**: 553–557.
- Paillard D, Labeyrie L, Yiou P. 1996. Macintosh program performs time-series analysis. *AGU Eos Transactions* **77**: 379.
- Peterson LC, Haug GH, Hughen KA, Röhl U. 2000. Rapid changes in the hydrologic cycle of the tropical Atlantic during the Last Glacial. *Science* **290**: 1947–1951.
- Poore RZ, Dowsett HJ, Verardo S, Quinn TM. 2003. Millennial-to century-scale variability in the Gulf of Mexico Holocene climate records. *Paleoceanography* **18**: DOI:10.1029/2002PA000868.
- Rozanski K, Araguá-Araguá L, Gonfiantini R. 1993. Isotopic patterns of modern global precipitation. In *Climate Change in Continental Isotopic Records, Geophysical Monograph 78*, Swart PK, Lohmann KC, McKenzie J, Savin S (eds). American Geophysical Union: Washington, DC; 1–36.
- Schmidt MW, Spero HJ, Lea DW. 2004. Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation. *Nature* **428**: 160–163.
- Socki RA, Karlsson HR, Gibson EK. 1992. Extraction technique for determination of oxygen-18 in water using preevacuated glass vials. *Analytical Chemistry* **64**: 829–831.
- Stuiver M, Reimer PJ. 1993. Extended ¹⁴C data base and revised Calib 3.0 ¹⁴C age calibration program. *Radiocarbon* **35**: 35–65.
- Stuiver M, Grootes PM, Braziunas TF. 1995. The GISP2 $\delta^{18}O$ climate record of the past 16 500 years and the role of the sun, ocean, and volcanoes. *Quaternary Research* **44**: 341–354.
- Stuiver M, Reimer PJ, Bard E, Beck JW, Burr GS, Hughen KA, Kromer B, McCormac G, van der Plicht J, Spurk M. 1998. INTCAL98 radiocarbon age calibration, 24 000–0 cal BP. *Radiocarbon* **40**: 1041–1083.
- van der Plicht J, van Geel B, Bohncke SJP, Bos JAA, Blaauw M, Speranza AOM, Muscheler R, Björck S. 2004. The Preboreal climate reversal and a subsequent solar-forced climate shift. *Journal of Quaternary Science* **19**: 263–269.
- Vinson GL. 1962. Upper Cretaceous and Tertiary stratigraphy of Guatemala. *Bulletin of the American Association of Petroleum Geologists* **46**: 425–456.
- Waliser DE, Shi Z, Lanzante JR, Oort AH. 1999. The Hadley Cell circulation: assessing NCEP/NCAR reanalysis and sparse in-situ estimates. *Climate Dynamics* **15**: 719–735.