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ORIGINAL PAPER

Palaeoenvironmental evolution of Cenote Aktun Ha (Carwash) on the Yucatan Peninsula, Mexico and its response to Holocene sea-level rise

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Abstract A 61-cm core was obtained from 4 m below the water table in Cenote Aktun Ha, on the Yucatan Peninsula, Mexico. The cenote is 8.6 km from the Caribbean coast and its formation and evolution have been largely affected by sea-level change. The base of the core dates to 6,940–6,740 cal year BP and overlying sediments were deposited rapidly over the subsequent ~200 years. The pollen record shows that the cenote evolved from a marsh dominated by red mangrove (*Rhizophora mangle*) and fern (Polypodiaceae) to an open-water system. These vegetation changes were controlled by water level and salinity and are thus useful indicators of past sea level. At the base, the $\delta^{13}C_{org}$ isotopic ratios reveal the influence of terrestrial vegetation (−29‰ VPDB), but shift to more negative values up-core (−33‰), indicating an influence from particulate matter in the flooded cenote pool. Although microfossil populations were nearly absent through most of the core, the microfossil assemblage in the upper 6 cm of the core is dominated by the juvenile foraminifer *Ammonia tepida* and the thecamoebian genus *Centropyxis*. These populations indicate open-water conditions in the cenote and a major environmental shift around 6,600 cal year BP, which is related to sea-level rise in the Caribbean basin. These data fit well with previously established sea-level curves for the Caribbean Sea. Our reconstruction of the environmental history of Cenote Aktun Ha helps elucidate the floral and hydrological history of the region, and highlights the utility of cenote sediments for studying the Holocene sea-level history of the Caribbean Sea.

Keywords Holocene sea level · Thecamoebians · Foraminifera · Cenote · Cave · Mangrove

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Introduction

The Yucatan Peninsula of Mexico is a carbonate platform comprised of Eocene-age limestone in the interior, with off-lapping sequences to a Quaternary coastline (Weidie 1985). The peninsula has been tectonically stable since the late Pleistocene (Szabo et al. 1978). It contains a coastal fresh–saline, density-stratified aquifer with a freshwater lens floating on encroaching seawater. Mixing at the fresh–saline interface generates a zone of undersaturated waters with respect to $CaCO_3$ (Back et al. 1986; Smart et al. 1988) resulting in the formation of extensive, anastomosing, phreatic cave systems (Smart et al. 2006) that are connected hydrologically with the Caribbean Sea. In the Mexican state of Quintana Roo, there are currently 165 explored submerged systems, including Sistema Ox Bel Ha and Sistema Sac Actun, which are the two longest underwater cave systems in the world, measuring 170 and 155 km, respectively (QRSS 2008). The cave systems have multiple sinkholes or cenotes that formed due to the collapse of subterranean caverns and which open the cave to the surface. Sediment accumulations in the cenotes have been studied rarely (e.g. Alvarez Zarikian et al. 2005) and there is little that is known about their environmental evolution through Holocene sea-level rise.

The objectives of this study are to test the applicability of the cenote sediment record as a palaeoenvironmental archive and sea-level indicator. Specifically, we use a multiproxy approach of gross lithology (i.e. organic matter content), microfossils (foraminifera, thecamoebians, ostracods), organic ($\delta^{13}C$) and carbonate isotope ($\delta^{13}C$ and $\delta^{18}O$) geochemistry, and palynology to examine sediments from Cenote Aktun Ha, commonly known as Carwash, in the state of Quintana Roo. The results presented here show that cenote sediment records: (1) provide valuable information about past vegetation and hydrological conditions in and around the cenote; and (2) have great potential for testing and refining previous western Caribbean Basin sea-level curves (e.g. Lighty et al. 1982; Fairbanks 1989; Blanchon and Shaw 1995; Toscano and Macintyre 2003; Gischler and Hudson 2004).

Study area

Cenote Aktun Ha (UTM coordinates: 16Q 449224 2241909) is located on the Yucatan Peninsula 8 km northwest of the town of Tulum, in the Mexican state of Quintana Roo (Fig. 1a). Modern vegetation on the eastern coast of the Yucatan Peninsula is a tropical arid, medium forest. The cenote is oval-shaped, with a large surface area ($\sim$1,000 m^2) and a maximum depth of 4.5 m. The cenote is 8.6 km from the coast and the water level is approximately equivalent to

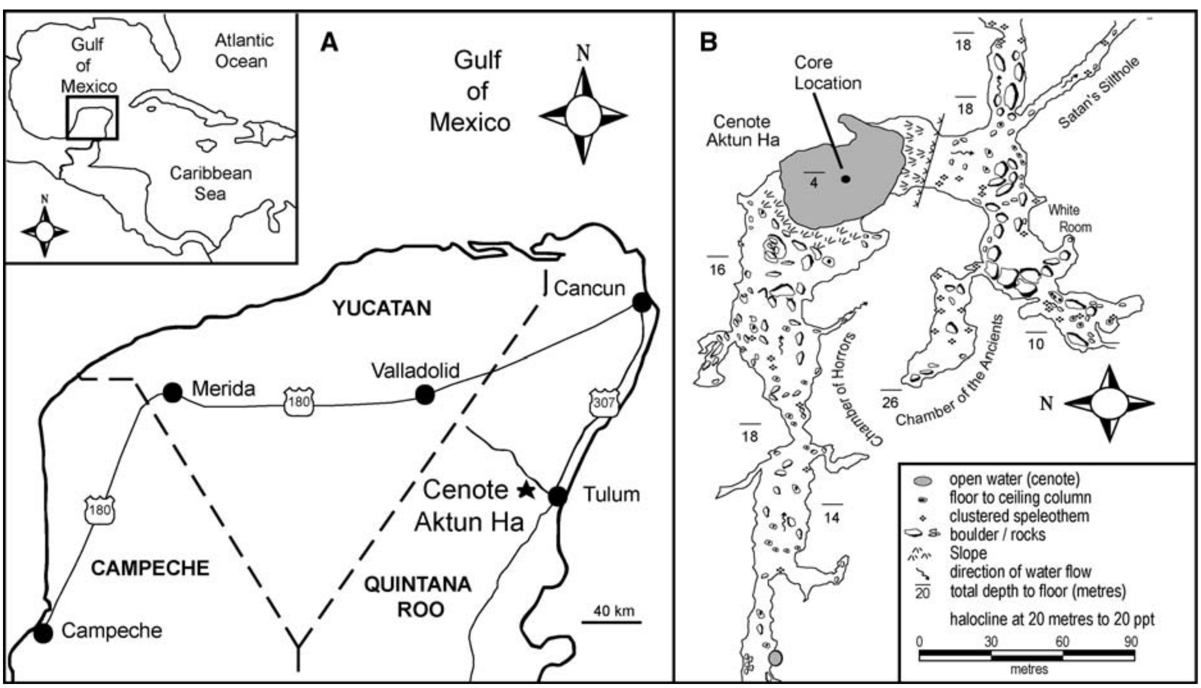

Fig. 1 (**a**) Location map of the Caribbean region (inset) and the Yucatan Peninsula, and (**b**) a portion of the exploration survey of Sistema Aktun Ha showing Cenote Aktun Ha, the core location, and the Chamber of the Ancients. Adapted from Coke and Young (1990)

mean sea level (msl) as the hydraulic gradient across the peninsula is on the order of 0.5–1.0 cm km^{-1} (Marin and Perry 1994; Beddows 2004). The water has an average temperature of 25°C and a salinity of ~1.5 ppt. The basin is connected to a cave system, Sistema Aktun Ha, which has a total explored extent of 2.8 km, with passages running both coastward and inland (Fig. 1b). Physico-chemical conditions of the cenote and adjacent cave water are seasonally and inter-annually consistent, based on periodic monitoring from 2000 to 2007 (Beddows 2004; Beddows et al. 2007) and measured semi-diurnal water-level fluctuations are on the order of 5 cm or less (Beddows 2004).

Background research

Post-glacial sea-level rise

There have been a number of investigations dealing with the post-glacial sea-level history of the Caribbean basin. Lighty et al. (1982) dated reef-crest corals (*Acropora palmata*—which grow within 5 m of sea level) from the tropical western Atlantic and produced a minimum sea-level curve that spanned the past 10,000 years. Additional research by Fairbanks (1989), using *A. palmata* from Barbados, extended the sea-level curve back to 17,000 year BP. On the basis of hiatuses in *A. palmata* growth, two periods of extreme rise in sea level were identified: melt-water pulse IA (mwp-IA) (12,000 year BP) and melt-water pulse IB (mwp-IB) (9,500 year BP). A third period of extreme sea-level rise at ~7,600 year BP was also identified from several locations in the Caribbean, where the rate exceeded 45 mm $year^{-1}$ (Blanchon and Shaw 1995; Blanchon et al. 2002). Obtaining a reliable coral sample that has not been subjected to taphonomic processes such as transport or erosion, however, can be difficult. Intertidal mangrove peats have also been used as indicators of sea-level rise (e.g. Ellison 1993), although others have argued that these data may also be problematic due to continuous decomposition and compaction of the organic matter (e.g. Gischler 2006). Recently, Toscano and Macintyre (2003) presented a calibrated sea-level curve that incorporated new and previously published coral and mangrove peat dates, to constrain sea level to both an upper and lower limit. Although this curve is similar to those presented by Lighty et al. (1982) and Fairbanks (1989), the precision of this method may not be sufficient to record minor sea-level oscillations <2 m (Scott et al. 1995). Using mangrove peat in conjunction with marsh foraminifera assemblages, Blum et al. (2001) presented Holocene sea-level data from the Texas coast of the Gulf of Mexico, which indicated a mid-Holocene (~6,800–4,800 year BP) sea-level highstand of 1.95 m above modern sea level. Blum et al. (2001) reference similar highstands in other low-latitude coastal locations such as western Africa, eastern South America, and Australia. There have yet to be any studies from Caribbean locations to document this highstand.

Holocene climate change

Another major focus of palaeoenvironmental research in the region has been Holocene climate history and its effect on the ancient Maya (e.g. Curtis et al. 1996; Leyden et al. 1996; Whitmore et al. 1996; Hodell et al. 2001; Leyden 2002). These studies were conducted using the sediment records from isolated inland lakes with no known associated cave systems (i.e. closed basins), making their primary hydrologic input/output precipitation and evaporation. Ratios of evaporation to precipitation were assessed through isotopic ($\delta^{18}O$) analysis of ostracod and gastropod shells (Curtis et al. 1996; Hodell et al. 2001). Hodell et al. (2001) found a dominant 208-year drought cycle in the last 2,600 years of the record, which is tied closely to solar activity. These closed-basin lakes are generally more than 40 km from the Caribbean coast, making them ideal for studying regional climate change, but there is no similar research devoted to the evolution of cenotes from the cave systems closer to the coast.

Palynology

The few palynological studies from the southeastern Yucatan Peninsula document climatic, sea-level, and anthropogenic processes that have shaped the flora of the region. In the longest record from the area, Leyden et al. (1998) generated a ~8,400-year pollen sequence from Lake Coba, within the Coba archaeological site. The initial flora was dominated by semi-deciduous and swamp forest. Indicators of deforestation, such as Poaceae and Chenopodiaceae pollen, are apparent beginning ~3,600 cal year BP, and maize pollen, which is indicative of local cultivation, is present by ~2,800 cal year BP. By AD 1240, the population of Coba had declined, and the forest began to recover to its pre-disturbance state.

Farther south, in the Rio Hondo region near the Belize border, Torrescano and Islebe (2006) investigated a core that revealed a ~5,000-year record of mangrove and tropical forest change. The area initially consisted of forest dominated by trees of the Moraceae and Fabaceae families. Between 4,600 and 4,000 cal year BP, the mangroves *Rhizophora mangle* and *Conocarpus erectus* became more abundant, likely in response to the flooding of the Rio Hondo as sea level rose. After that time, *R. mangle* decreased in importance and *C. erectus* dominated the record, perhaps reflecting a shift toward a higher-salinity, seasonally dry environment. Due to a hiatus in sedimentation, the record is missing the last 2,500 years.

Finally, Islebe and Sanchez (2002) studied a core from a mangrove system near the coastline about 30 km south of Cancun. They infer humid conditions from 2,500 to 1,500 ^{14}C year BP, when the pollen of *R. mangle* and elements of the nearby semi-evergreen tropical forest dominated the record. From 1,500 to 1,200 ^{14}C year BP, *C. erectus* was dominant, suggesting drier and more open conditions, which Islebe and Sanchez (2002) suggest may be linked to the period of Maya cultural decline. After that time, *R. mangle* again dominated for about 200 years, until *C. erectus* returned and persisted from ~1,000 ^{14}C year BP until the present.

Palaeoenvironmental indicators

Thecamoebians and foraminifera are testate rhizopods that are very useful in reconstructing past conditions in coastal systems because they are known to inhabit almost every aquatic environment, and are very sensitive to salinity, temperature and eutrophication (Scott et al. 2001). The tests have excellent preservation potential and because they typically occur in large numbers (>10^3/cc), detailed environmental reconstructions have been conducted in several studies from both temperate and tropical zones (e.g. Miller et al. 1982; Honig and Scott 1987; Hodell et al. 2005; Reinhardt et al. 2005; Peros et al. 2007a). Thecamoebians have never been reported from cenotes or caves in Yucatan and there is only one foraminifer taxon that has been reported from Aguada X'caamal (Hodell et al. 2005) and Lake Chichancanab (Hodell et al. 1995), two closed-basin lakes on the peninsula. On the island of Barbados, however, ponds have been found to contain thecamoebian assemblages (Roe and Patterson 2006). They generally display low abundances and diversities, and are typically dominated by centropyxids, which are indicative of brackish environmental conditions (Riveiros et al. 2007; van Hengstum et al. 2008a). Other than these few studies, thecamoebian and foraminiferal distributions are virtually unknown in cenote and cave systems (van Hengstum et al. 2008b).

Pollen analysis was also used to document the vegetation history of Cenote Aktun Ha and infer changes in the hydrology of the basin. Mangrove communities are common in quiet-water settings around the circum-Caribbean, and recent palynological investigations of these systems provided evidence of sea-level and climate changes in the region (e.g. Torrescano and Islebe 2006; Peros et al. 2007a, b). The main mangrove species in the Caribbean is *Rhizophora mangle* (red mangrove), which typically grows in brackish water in sheltered systems (Tomlinson 1986). Its pollen is produced in large quantities and is dispersed by both wind and water, although a recent investigation of modern coastal sediments has shown that when *R. mangle* pollen exceeds ~30% of the pollen sum, the source plant(s) were likely within a 5 m radius of the coring site (Davidson 2007). Thus, *R. mangle* pollen is a good indicator of local vegetation dynamics, and because of its affinity for brackish settings, can be used as a palaeosalinity indicator.

Analysis of δ^{13}C has the potential to identify the dominant sources of organic and inorganic carbon to the cenote and cave system. Pohlman et al. (1997) presented $\delta^{13}C_{org}$ data from in and around Cenote Maya Blue, which is part of the phreatic Sistema Naranjal, as well as total organic carbon (TOC) values from within the cenote and cave passages. Cenote Maya Blue is about 5.5 km from the coast and is a very similar system to the Aktun Ha study site, which is about 8.6 km away. Their results show that the terrestrial vegetation has a δ^{13}C of −29‰ while the pool particulate organic matter (PPOM), which describes all organic matter in the water column in the cenote, has a value of −35‰ (Fig. 2). The TOC

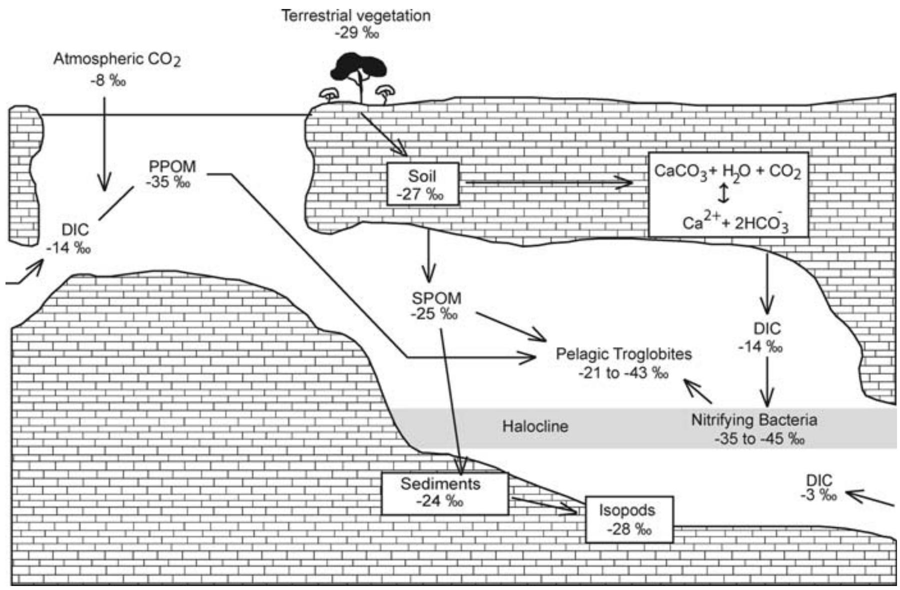

Fig. 2 δ^{13}C cycling model in a Yucatan cave system. Adapted from Pohlman et al. (1997)

values of sediments presented by Pohlman et al. (1997) show that organic carbon ranged from as high as 24% at the entrance to a flooded cave, to as low as 0.1% more than 60 m into the cave's interior.

Methods

A 5-cm-diameter, sediment push core was obtained by SCUBA divers in August, 2006 from the relatively flat top of the breakdown pile of the cenote at a depth of 4 m below modern water level. A high, central location away from the sides of the breakdown pile was selected to avoid disturbed sediment. Several attempts were required to find a cavity amongst the breakdown pile of limestone blocks. Initial penetration of the sediment was 1.3 m though the final sediment core length was only 61 cm. This discrepancy is likely due to compaction of the sediment as it has a high organic content (average 55%) and porosity. Although over penetration of the core due to plugging at the penetration front cannot be ruled out, the loose consistency of the sediment suggests that this was not a factor.

Fossil pollen was concentrated by subjecting 18 roughly evenly spaced subsamples from the core to treatment with dilute HCl, KOH, acetolysis solution, and HF, and sieving with both 150-µm and 7-µm nylon mesh (Faegri and Iversen 1989; Cwynar et al. 1979). A known number of exotic spores of the moss *Lycopodium clavatum* were added to each subsample so pollen concentration (expressed in grains/cc) could be estimated. Identifications were made using published atlases (Palacios Chávez et al. 1991; Roubik and Moreno 1991) as well as the Neotropical Pollen Database (Bush and Weng 2007). In the majority of cases at least 300 pollen and spores were enumerated per level; in a few cases, this number was not reached because of very low pollen concentrations.

Microfossil samples were wet-sieved with a 45-µm screen to remove silt- and clay-sized particles, and examined wet using a binocular dissecting microscope (max. 80X). Samples were split randomly using a wet-splitting apparatus and standard methods (Scott and Hermelin 1993). The assemblages were identified to the species level using published taxonomic information (Moore 1961; Medioli and Scott 1983; Pointier 2001; Scott et al. 2001). The populations are presented as down-core variations in specimens cm^{-3} (Table 1).

Ten bulk sediment samples were analysed for isotopic ratios of $\delta^{13}C_{org}$. Each sample was first rinsed in 30 ml of 10% HCl solution for 12 h to oxidize carbonates such as shell fragments or calcite grains. The samples were then rinsed twice with distilled water and dried for 24 h at 67°C in an oven. The remaining organic matter was ground into a fine homogenous powder using a mortar and pestle. Subsamples weighing 40–50 µg were placed into a Costech elemental analyzer where they were combusted and analysed with a Delta Plus XP mass

Table 1 Down-core changes in foraminifera/thecamoebians, gastropods, and ostracods (cm^{-3})

Depth (cm)	2–3	3–4	4–5	7–8	18–19	24–25	36–37	48–49
Total number of specimens	425	893	336	130	168	376	205	351
Total number of specimens (cm^{-3})	340	714.4	268.8	13	16.8	37.6	20.5	35.1
Foraminifera/thecamoebians								
Ammonia tepida "juvenile"	218.4	536	152.8	0	0	0	0	0
Arcella vulgaris	0.8	0.8	0.8	0	0	0	0	0
Centropyxis aculeata "aculeata"	20	22.4	4.8	0	0	0	0	0
C. aculeata "discoides"	9.6	0.8	0	0	0	0	0	0
C. constricta "aerophile"	0	4.8	0	0	0	0	0	0
C. constricta "constricta"	3.2	6.4	0	0	0	0	0	0
Tintinnid favella	0	6.4	1.6	0	0	0	0	0
Trochammina macrescens	0	0.8	0	0	0	0	0	0
Gastropods								
Pyrgophorus coronatus	32	24	7.2	3.2	9.0	21.1	3.3	13.2
P. marmorata	0	0	0	0	0.6	0.3	0.3	0
Ostracods								
Darwinula stevensoni	56	112	101.6	8.8	5.7	7.3	7.7	8.2
Cytheridella ilosvayi	0	0	0	0	1.5	8.9	9.2	13.7

spectrometer. The results were compared against internal, IAEA and NIST standards (NBS21, ANU-Sucrose). Isotopic ratios were expressed in standard delta notation (δ) in per mil (‰) against Vienna PeeDee Belemnite (VPDB).

Adult specimens of the ostracod *Cytheridella ilosvayi* were picked from the 250-μm fraction at 9 intervals through the core. The carapaces were cleaned and sonicated in distilled water to remove any bulk organic matter. The samples were analysed for isotopic fractions of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ using a VG Optima mass spectrometer at the Laboratory for Stable Isotope Science at the University of Western Ontario in London, Ontario, Canada.

Total Organic Carbon (TOC) was analysed in 29 bulk samples using the loss on ignition (LOI) technique presented by Heiri et al. (2001).

Radiocarbon (AMS) analyses were performed at the Beta Analytic Radiocarbon Dating Laboratory in Miami, Florida on three twigs and one shell picked from four intervals in the core. Twig samples were pretreated using an acid/alkali/acid procedure to remove carbonates and any secondary organic matter, while the shell was pretreated using an acid etch method. The samples were then analysed for radiocarbon age and calibrated using the IntCal04 database (Reimer et al. 2004). The dates presented here are in calendar years before present (cal year BP).

Results

Lithology and LOI

Nearly the whole core, from 4 to 61 cm, was comprised of a dark, organic-rich peat with several woody fragments as well as whole and fragmented gastropod shells. From 0 to 4 cm was comprised of a coarse filamentous algae with whole and fragmented gastropod shells and plant fragments. LOI values varied throughout the core (Fig. 3). Values varied about a mean of 55.8% (by weight) though values as high as 71% and as low as 32% were found, but no clear trend was documented in the core.

Pollen

On the basis of the pollen taxa (Fig. 3; Table 2), three pollen zones were identified. Zone 3 comprises the bottommost samples (51–61 cm) and is dominated by the red mangrove species *Rhizophora mangle*. A small quantity of fern pollen (Polypodiaceae), possibly that of *Acrostichum* (mangrove fern), and a number of trees (e.g. Sapotaceae, Solanaceae, and *Spondias mombin*) also characterize this zone. On the basis of their morphologies, most of the Solanaceae in the core appear to be *Solanum*, whereas the

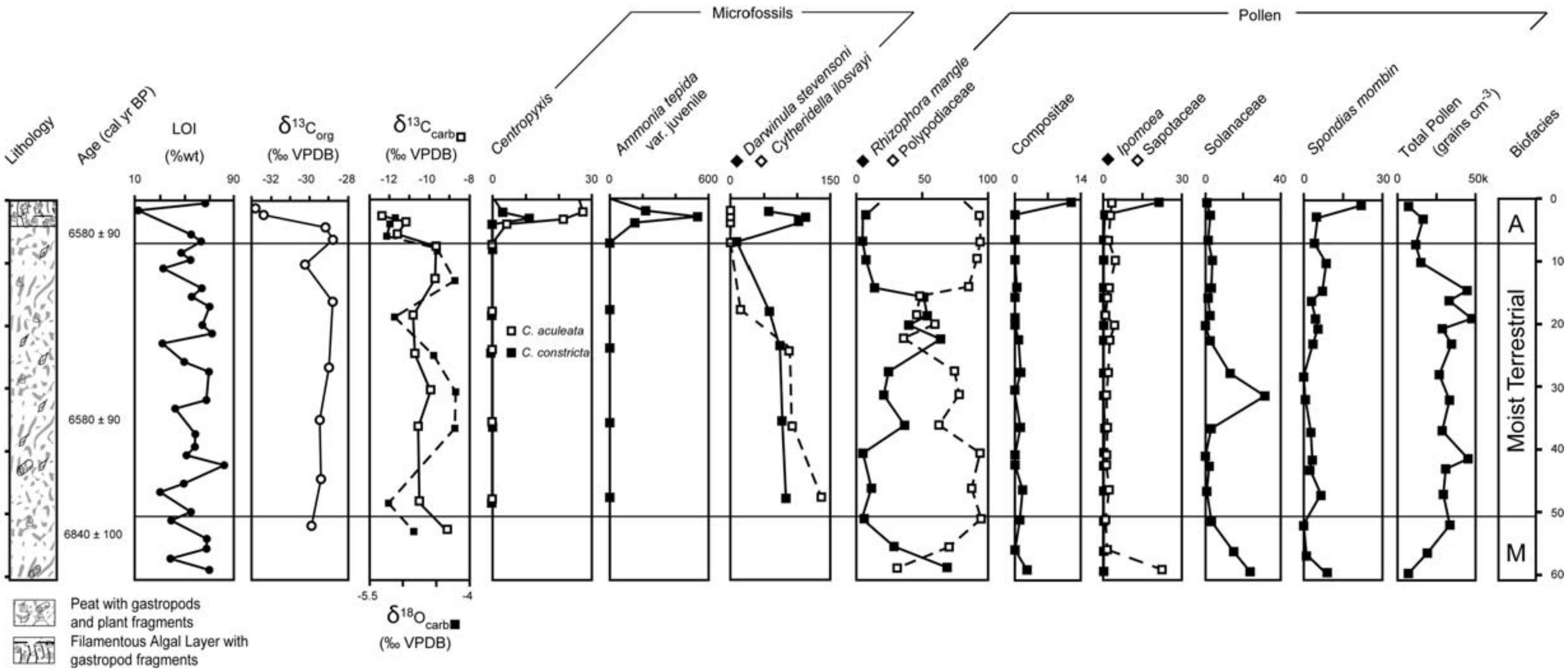

Fig. 3 The down-core results of stable isotope (‰ VPDB), microfossil (specimens cm^{-3}), pollen analysis (relative abundance), and the associated chronology. The combined interpretation of the proxies established three facies: Mangrove (M), Moist terrestrial, Aquatic (A)

Table 2 List of all pollen taxonomic groups identified in the core

Dominant local wetland plants	Herbs, vines/lianas, mosses and other wetland plants	Trees and shrubs
Polypodiaceae	Amaranthaceae	*Aegiphila*
Rhizophora	Compositae	*Alibertia*
	Convolvulaceae (non-echinate)	*Alnus*
	Cyperaceae	Anacardiaceae
	Fabaceae-Mimosoideae	Apocynaceae
	Fabaceae-Caesalpinioideae	Apiaceae
	Poaceae	Araliaceae
	Ipomoea-type	Arecaceae
	Lamiaceae	Bignoniaceae
	Lycopodium	*Brosimum alicastrum*
	Passiflora	*Bursera*
	Polypodium (smooth, monolete)	Crassulaceae
	Polypodium (verrucate, monolete)	*Croton*-type
	Rubiaceae	Cucurbitaceae
	Selaginella	Euphorbiaceae
	Sparganium-type	Malphigiaceae
	Styzolobium	Moraceae
	Typha	*Myrica*
	Trilete-other	*Pinus*
		Sapindaceae
		Sapotaceae
		Solancaceae
		Spondias mombin

This list does not include unknowns, which in some cases were classified morphologically

Sapotaceae is likely *Pouteria campechiana*. Pollen concentration is at its lowest in Zone 3.

Pollen Zone 2 (7–51 cm) was characterized by significant increases in Polypodiaceae and pollen concentration, and decreases in *Rhizophora* and a number of the trees and shrubs that are present in Zone 3. At 46 cm, Malphigiaceae increases to ~10% of the tree and shrub sum. The pollen concentration of Zone 2 is much higher than the other zones, reaching almost 48,000 grains/cc at some depths.

Zone 1 comprised the uppermost sample, from 0 to 7 cm. Despite an abundance of Polypodiaceae and a drastic decrease in pollen concentration (similar to Zone 3), the palynological richness (as defined by Birks and Line 1992; Peros and Gajewski 2008) of Zone 1 is high, containing a variety of trees, shrubs, and other herbaceous plants. *Spondias mombin* is at its most abundant in this zone, and other taxa, such as *Ipomoea* and Compositae, record very high concentrations compared to deeper levels.

Microfossils

The upper interval (0–6 cm) of the core had an assemblage that was dominated by what we have identified as juveniles of the foraminifer *Ammonia tepida* (Fig. 3). All specimens were underdeveloped, with at most 1–2 chambers in addition to the proloculus. No mature adult specimens were found in the assemblage. The specimens are thought to be juvenile *A. tepida* as they were found with the thecamoebians *Centropyxis constricta* and *Centropyxis aculeata,* indicating salinity of ~3.3 ppt, consistent with the modern water salinity. Based on this close association, we suspect they are juvenile *A. tepida* rather than a new species; *Ammonia tepida* and *Centropyxis* spp. are typically found in the upper reaches of estuaries (e.g. Scott et al. 2001). There are no published data for these juvenile specimens (D.B. Scott pers. comm. 2007), but they may be undocumented because of their small size (~50 µm). Many microfossil studies use a 63-µm screen compared to the 45-µm screen that was used in this study. Thus, many of the small specimens enumerated in this study may have been lost in previous investigations. If this interpretation is correct, then the abundance of juveniles suggests short-term periods of higher salinities, allowing them to bloom, but does not allow them to mature into adults. Limnetic tintinnids, which are generally pelagic protozoans, are also present in this interval. The remainder of the core, from 6 to 61 cm, is dominated by ostracods, *Cytheridella ilosvayi* and *Darwinula stevensoni*, and the gastropod *Pyrgophorus coronatus*. The ostracods have been documented in other studies from the Caribbean (e.g. Holmes 1998), inhabiting fresh to slightly saline lakes, in the shallow waters between the reed swamp and the open water.

Isotopic geochemistry

The $\delta^{13}C_{org}$ values show a distinct shift to more positive values from −32.8‰ to −28.8‰ at the top of the core (0–6 cm), and then a subtle shift back to more negative values (∼ −29.5‰) from 6 to 61 cm (Fig. 3; Table 3). A similar trend can be seen in the $\delta^{13}C_{carb}$ values, which have the most negative values (−12.4‰) at the top of the core (2–3 cm), then increase to −9.7‰ at 7–8 cm, and then exhibit minor fluctuations downcore, with a mean of −10.2‰. Finally, the $\delta^{18}O_{carb}$ values fluctuate throughout the core from −4.21‰ to −5.25‰, but shift to more negative values at around 6 cm. The shift in $\delta^{13}C$ and $\delta^{18}O$ of both the carbonate and organic fractions at 6 cm marks a large change in both water chemistry and organic matter production/deposition within the cenote.

Chronology

The three wood fragments and shell selected for ^{14}C-dating showed a slight downcore increase in age. The bottom of the core (54–55 cm) was dated to 6,840 ± 100 cal year BP and was similar to the shell date at 55–57 cm (7,340 ± 90 cal year BP); the offset is probably due to the hard-water effect from inorganic carbon in the ground water. The upper samples (34–36 cm and 5–6 cm) have a similar age of 6,580 ± 90 cal year BP (Table 4). The dates indicate the rate of accumulation through most of the core, from 5 to 57 cm, was approximately 0.65 cm year^{-1}. The date of 6,580 ± 90 cal year BP at 5–6 cm indicates a nearly 1,000-fold decrease in sediment accumulation rate to 1.9×10^{-3} cm year^{-1} for the past 6,500 years, i.e. a virtual depositional hiatus, as reported by Torrescano and Islebe (2006).

A fifth ^{14}C date was obtained from Coke et al. (1991), who dated a piece of charcoal from a hearth within the Aktun Ha cave system at a depth of 27 m. The charcoal was within a section of the cave known as the Chamber of the Ancients and was dated to 8,250 ± 80 year BP, 9,230 ± 200 cal year BP (recalibrated with IntCal 04; Reimer et al. 2004) indicating that the water level in the cave was below this elevation at that time.

Table 3 Stable isotope ($\delta^{13}C$ and $\delta^{18}O$, ‰VPDB) results

Depth (cm)	Organic	Carbonate	
	$\delta^{13}C$ (‰)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)
0–1	−32.8	–	–
1–2	−32.4	–	–
2–3	–	−5.12	−12.37
3–4	−29.2	−5.20	−11.17
5–6	−28.8	−5.25	−11.59
7–8	–	−4.49	−9.66
9–10	−30.2	–	–
12–13	–	−4.22	−9.71
15–16	−28.8	–	–
18–19	–	−5.13	−10.81
24–25	–	−4.54	−10.72
25–27	−29.0	–	–
30–31	–	−4.21	−9.93
34–36	−29.5	–	–
36–37	–	−4.22	−10.56
43–45	−29.4	–	–
48–49	–	−5.22	−10.49
51–52	−29.9	–	–
52–54	–	−4.84	−9.10

The organic ratio is from bulk peat samples and the carbonate ratio is from ostracod (*C. ilosvayi*) valves

Discussion

Environmental reconstruction of Cenote Aktun Ha

The pollen record shows three distinct palynofacies through time as water level within the cenote increased. The bottom zone is characterized by

Table 4 AMS ^{14}C dates from organic material (twigs) and carbonate material (ostracods)

Lab ID	Location	Core depth (cm)	Material	$\delta^{13}C$ (‰)	Conventional age (year BP)	2Σ calibration range(cal year BP)
Beta-226967	Cenote	5–6	Twig	−29.7	5,790 ± 40	6,670–6,490
Beta-226968	Cenote	34–36	Wood Frag	−29.3	5,780 ± 40	6,670–6,480
Beta-226969	Cenote	54–55	Wood Frag	−29.3	5,990 ± 40	6,940–6,740
Beta-235036	Cenote	55–57	Shell	−9.9	6,390 ± 40	7,421–7,256

Dates were calibrated using the IntCal04 database (Reimer et al. 2004). The offset between organic and carbonate matter is probably a result of the hard-water effect

generally high but variable amounts of *R. mangle*. This plant provides a good indicator of contemporary sea level as red mangrove grows within the intertidal zone in brackish environments. The fern *Acrostichum*, which is often found in mangrove environments, becomes abundant in Zone 2. This, along with the abundant mangrove pollen, indicates that conditions were wet and terrestrial, and water level was close to the top of the breakdown pile 6,500–7,000 cal year BP (Fig. 4a). The high abundance of these pollen grains within the core indicates a localized input from plants and trees living on the breakdown pile in the cenote depression, rather than from regional inputs. A similar environment can be seen in some modern flooded cenotes such as Cenote Eden (16Q 473097 2265716), where portions of the breakdown pile form small islands of ferns, shrubs and grasses in the middle of the cenote pool (Fig. 5). The sediments accumulating in crevices of the breakdown pile are organic-rich and have $\delta^{13}C_{org}$ values of ~ -29‰, indicative of terrestrial vegetation (Fig. 2).

The decline in pollen concentration at 9 cm and the reduction of *R. mangle* through Zones 1 and 2 indicates the flooding of the breakdown pile (4 m below MSL) at 6,500 cal year BP (Fig. 4b). The continued rise in sea level eventually formed the flooded cenote, as documented in the top 5 cm of the core (Fig. 4c). At approx. 6,500 cal year BP, this rising water table (and sea level) caused the die-off of all the vegetation on the breakdown pile and cessation of direct input of pollen grains from this vegetation. The top 5 cm of the core shows an increase in palynological richness and the appearance of grass (Poaceae) and possibly some weeds (Compositae), suggesting that there was a disturbance in the local vegetation over the past 6,500 years. Nevertheless, because the sedimentation rate was so low, it is impossible to determine the exact timing of the vegetation change. The shift in $\delta^{13}C_{org}$ values from −29‰ to −32‰ in the upper 2–3 cm is indicative of flooding. This rise in water level increased algal

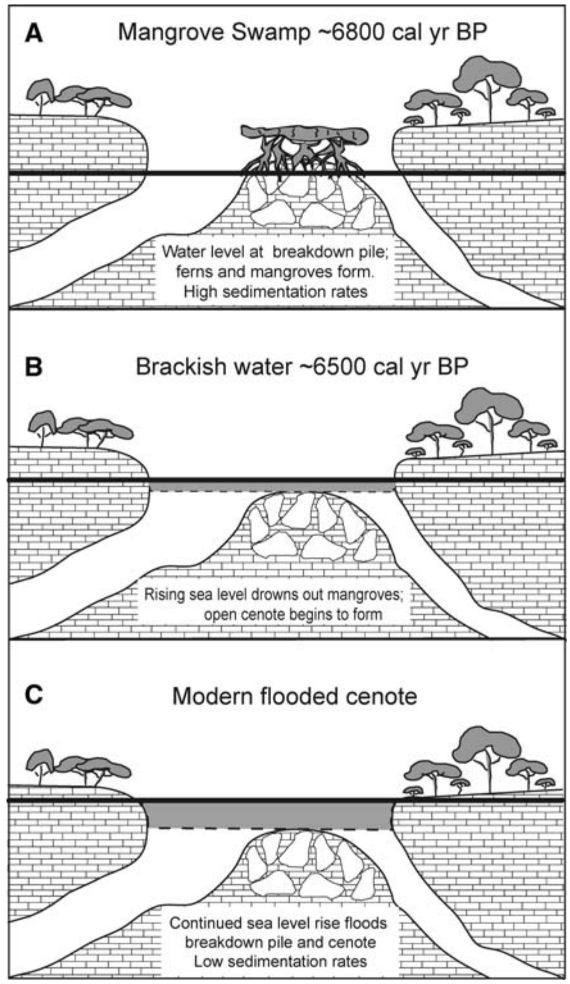

Fig. 4 An interpretive diagram showing flooding sequence in Cenote Aktun Ha. Shaded region represents flooding on the breakdown pile due to sea level rise

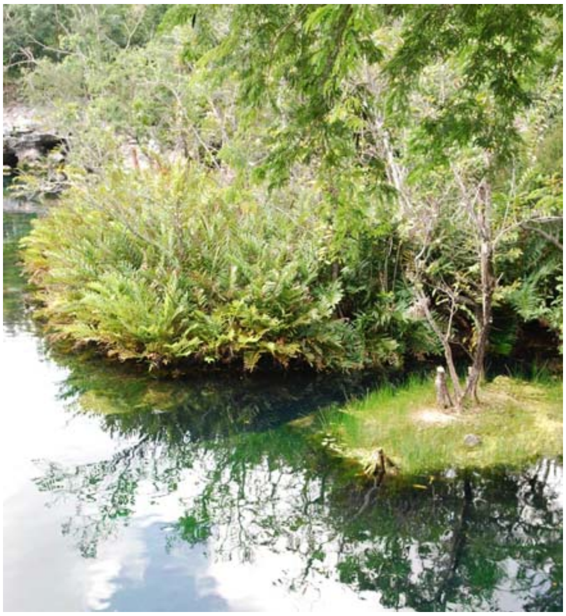

Fig. 5 A picture of the modern conditions in Cenote Eden. The fern islands are growing in the middle of the cenote pool on top of collapsed boulders

productivity, causing higher accumulations of PPOM, and a shift to more negative $\delta^{13}C_{org}$ values.

The microfossil populations within the core followed similar trends as the pollen, with two distinct biofacies (Fig. 3) showing a similar flooding history. The biofacies 1 microfossil populations at the top of the core (<6 cm) indicate a shift to a flooded cenote with slightly brackish ($\sim$3 ppt) conditions, as indicated by *A. vulgaris*, *Centropyxis* spp., juvenile *A. tepida*, and pelagic tintinnids (van Hengstum et al. 2008a). The basal section (>6 cm) of the core was defined by a lack of thecamoebians and foraminifera, but was dominated by the ostracods *C. ilosvayi* and *D. stevensoni* and the freshwater gastropod *P. coronatus*. The ostracods were likely living in the shallow waters near the collapse pile at the margin of the mangroves. These species have been found in other lakes on the Yucatan (e.g. Curtis et al. 1996; Whitmore et al. 1996; Hodell et al. 1995, 2005).

The reason for the lack of foraminifera and thecamoebians below 6 cm in the core is unclear, although it may be due to harsh environmental conditions (i.e. low O_2) in the marsh/mangrove. Marsh and mangrove sediments routinely have well-preserved foraminifera and thecamoebians with agglutinated tests (Scott et al. 2001) and the preservation of pollen and ostracods would suggest that foraminifera and thecamoebians should also be preserved. It could be that the concentrations of thecamoebians and foraminifera tests were low due to the relatively high sedimentation rate of this interval (0.65 cm year^{-1}). This, coupled with possible anoxic conditions due to organic matter decay, may have further limited population densities and the few specimens present may simply have gone undetected. In contrast, the upper, condensed horizon had relatively high concentrations of microfossils because the high OM inputs were lost with the water level change and open cenote conditions.

The rapid upcore change in $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ starting around 6 cm to more negative values (−11.59‰ and −5.25‰, respectively), might also indicate a transition to fresher conditions. The $\delta^{13}C_{carb}$ of the DIC of the modern open cenote is −14‰ while marine waters below the halocline have values of −3‰ (Pohlman et al. 1997). Therefore, the shift to negative values could reflect a slight salinity shift towards fresher conditions, which can also be seen in the $\delta^{18}O_{carb}$ values. This is consistent with the pollen record, which shows mangrove dominating at the base of the core, consistent with more brackish conditions. However, the reason for the salinity shift is unclear from these data. It may be that salinity was higher due to increased evaporation from the shallow water body during the mangrove phase. Subsequent flooding due to rising sea level and changing hydrological conditions may have caused a decrease in salinity. Alternatively, the shift in $\delta^{13}C_{carb}$ values could be related to changes in productivity in the cenote that accompanied rising water levels.

Implications

The dates obtained from the core were compared with the calibrated sea-level curve of Toscano and Macintyre (2003) (Fig. 6) and show good agreement with the mangrove peat data points from other Caribbean locations. This result should be tested in other cenotes. There are two main advantages to using cenote records: (1) the breakdown pile has elevational stability, and (2) provided the transition from marshy to open flooded cenote conditions can be identified, an accurate and precise sea-level point can be determined. A high-resolution timescale of Holocene sea-level rise in the Caribbean could be constructed by sampling other cenotes with collapse piles of different elevations.

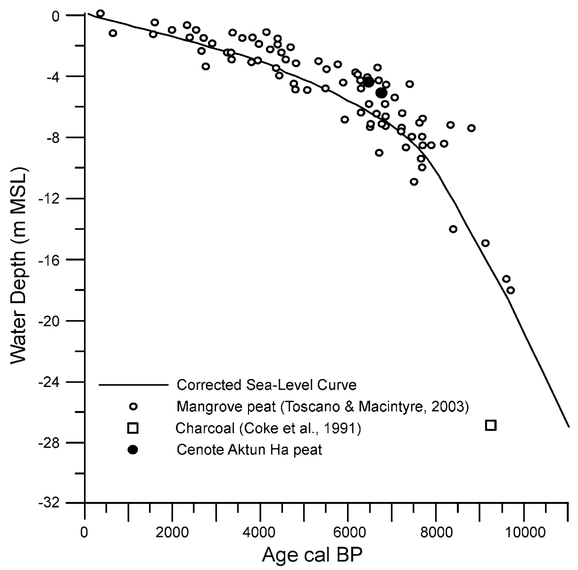

Fig. 6 Local Caribbean sea-level curve from Toscano and Macintyre (2003) with their mangrove peat data, cave charcoal (Coke et al. 1991), and this study. The symbols for the data from this study encompass the error range on the date

Current sea-level points in the Caribbean have methodological issues related to elevation (accuracy and precision), so the cenotes could provide an additional proxy to test these arguments. Mangrove peat is susceptible to compaction and older basal peat is often contaminated with younger peat (Gischler 2006), while reef-crest corals (*A. palmata*) have an error range of up to 10 m (Blanchon 2005), making it difficult to determine minor sea-level fluctuations on the order of 1–2 m. In addition, obtaining older samples for sea level points requires coring in the deep ocean (>100 m; Fairbanks 1989). Results presented here suggest a multi-proxy approach, and dating the transition in a cenote from semi-terrestrial marsh to open water conditions overcomes these problems. Thus, older sea level information can be obtained by sampling a cenote with a deeper collapse pile. For example, Cenote Angelita (16Q 439555 2226701), which is located 11.7 km from the coast, has a depth of 30 m at the collapse pile, giving the potential to track sea level change much further back in time.

The data from Cenote Aktun Ha are also consistent with a growing body of literature that is documenting coastal environmental changes during the middle Holocene across the Caribbean region. In a recent palaeoecological study from Laguna de la Leche, a 67-km^2 wetland located on the north coast of central Cuba, Peros et al. (2007a, b) used palynological and micropalaeontological evidence to show that the wetland formed ~6,500 cal year BP as a result of sea level rise. In Belize, Yang et al. (2004) obtained sediment cores from Chetumal Bay that showed thick accumulations of *R. mangle* peat. The peat was deposited on top of the Pleistocene limestone at a depth of 5.2 m below modern sea level and the basal age was 6,330 ± 70 year BP (uncalibrated). In a similar environment on the island of Jamaica, Digerfeldt and Hendry (1984) obtained sea-level points of ~4.5 m below modern sea level from *R. mangle* peat that dated to 5,790–6,080 cal year BP (uncalibrated). Synchronous with the Jamaican site, Florida palynological evidence documents a middle Holocene transition from oak- to pine-dominated flora, indicating a shift from dry to wet conditions across the peninsula (Grimm et al. 1993) and enrichment in δ^{18}O from ostracod shells, indicating a shift from dry to wet conditions and rising sea level (Alvarez Zarikian et al. 2005). The changes seen throughout Florida, northern Cuba, Belize, and at Cenote Aktun Ha, which all occurred in tectonically stable areas, suggest that coastal Caribbean vegetation and hydrological systems underwent a widespread shift ~6,500 cal year BP, either as a direct response to the marine transgression, or indirectly, through the effects of rising sea level on the phreatic aquifer. We hypothesize that other, similar studies from tectonically stable areas throughout coastal areas of the Caribbean will show changes similar to those identified at Cenote Aktun Ha.

Conclusions

Considerable debate exists concerning the Holocene sea-level history of the western Caribbean basin. Palaeoenvironmental data collected from Cenote Aktun Ha, near Tulum in the eastern Yucatan Peninsula, provide evidence that assists in interpreting both the relative sea-level history of the region, as well as the impacts of sea-level changes on the hydrology and vegetation of the region. The multi-proxy approach used on this core allowed for the reconstruction of the environmental changes within the cenote from ~6,800 cal year BP to the present, which can be directly linked to sea-level rise in the Caribbean. Our results show that relative sea level reached 4 m below present level by ~6,800 cal year BP, which is consistent with established sea-level curves for the Caribbean. Further research is required, but these preliminary results show that cenote sediment records may be useful sea level indicators for the Caribbean.

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