Environmental change in northern Belize since the latest Pleistocene

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ABSTRACT: Cores from the New River Lagoon in Belize have been analysed using a range of palaeolimnological proxies to reconstruct change since the latest Pleistocene. The combination of diatom and stable isotope analyses (supported by mineralogical and major element data) indicate that the New River Lagoon has been a freshwater system throughout the period of record. For most of this time the lagoon has been hydrologically open, but there are indications that it may have become closed in the latest Pleistocene or very early Holocene. This may have been associated with a drier climate and lower sea level. Mid-Holocene records are consistent with those from elsewhere in the region, indicating a stable and moist climate. The late Holocene was marked by greater variability, but there is no clear evidence of any drying in the late Classic coincident with the Maya ‘collapse’ ca. AD 900. Our results indicate that the continuity of occupation at the Maya site of Lamanai on the shores of the New River Lagoon is consistent with continued access to freshwater during periods of regional climatic variability. The importance of conditions at individual sites when considering both system response and human vulnerability to climate change is emphasised. Copyright © 2009 John Wiley & Sons, Ltd.

KEYWORDS: climate; Holocene; Maya ‘collapse’; diatoms; stable isotopes.

Introduction

The Yucatán peninsula has become a major focus for research because of interest in the apparent association between changes in the Maya culture and climatic shifts. Papers on the wider region (Hodell et al., 1995, 2005a, 2007; Curtis et al., 1996; Haug et al., 2003; Neff et al., 2006) have drawn attention to the correlation in time between the ‘collapse’ of the Maya civilisation ca. AD 750–900 (late or terminal Classic) and the occurrence of a severe dry period, or periods, between ca. AD 800 and 1000. The high-resolution record from Punta Laguna (Curtis et al., 1996; Hodell et al., 2007) also indicated that the Maya hiatus coincided with a first major dry phase at AD 585. The theme of cultural response to climate change has also been taken up by Furley (1998), Gill (2000) and Webster (2002) and in a wider context by deMenocal (2001).

Belize lies between the better-studied areas of the northern and south-western Yucatán peninsula (Brenner et al., 2002); limited palaeoenvironmental research within Belize so far has been based largely on palynology. The aim of this paper is to determine the environmental changes that have affected the New River Lagoon in northern Belize (Fig. 1) over the Holocene, encompassing the Maya occupation of the region, based on results from two sites: Hillbank in the south and at Lamanai in the mid section of the lagoon.

Environmental change and Maya chronology in northern Belize

Belize forms part of the south-eastern Yucatán peninsula between approximately 88° 45' and 89° 15' W and 15° 45' and 18° 30' N. It is bordered by Mexico, Guatemala and the Caribbean Sea (Fig. 1). The climate of Belize is typical of the Northern Hemisphere tropical Americas, with a summer wet season driven by the northward movement of the Inter-Tropical Convergence Zone (ITCZ). The climate in the north of the country is both drier and more seasonal than in the south (Esselman and Botes, 2001). Annual precipitation in our study area in northern Belize is in the range 1524–2032 mm (Walker, 1973).

Northern Belize is underlain by Cretaceous and Tertiary carbonates (mainly limestone), with a discontinuous cover of Pliocene well-drained sands and gravels and Quaternary alluvium. Three fault lines trending NNE to SSW influence
the direction of the major river valleys of the Booth River, the Río Bravo and the New River. The New River (Fig. 1) is one of the longest rivers in this area and enters the Bahía Chetumal near Corozal. The New River Lagoon is a freshwater body lying approximately 60 km from the estuary. The vegetation of the well-drained, acidic sand ridges is pine savanna, while more calcareous sediments support lowland, moist evergreen forest (Bridgewater et al., 2002). Herbaceous swamps, seasonally flooded savanna and marshland occupy the freshwater lowlands. There are mangrove forests along the coast and remnant mangroves are found well inland, including New River Lagoon.

In common with adjacent areas of lowland Mexico and Guatemala, Belize has evidence for extensive Maya occupation by the late Preclassic (ca. 300 BC to AD 250). Most Maya sites in Belize seem to have reached their peak in the early Classic (ca. AD 250–600). While a number of these sites do show evidence for abandonment in the late Classic, there appear to have been key sites that did not collapse particularly along parts of the coast and the site of Lamanai (see below) (Graham, 2001).

Two wetland sites in northern Belize have been studied previously: Cobweb Swamp, adjacent to the Mayan site of Colha (Alcala-Herrera et al., 1994; Jacob and Hallmark, 1996). Zea mays was found in Cobweb Swamp and Laguna de Cocos, Albion Island (Fig. 1). An 8000 a record was obtained from Cobweb Swamp, adjacent to the Mayan site of Colha. One water sample from the lagoon yielded isotopic values of $\delta^{18}O$ and $\delta^D$. This $\delta^{18}O$ is close to that recorded in precipitation at Petén Itza in Guatemala (Curtis et al., 1998) and lies between values for lake water at Coba and San José Chulchaca in the Mexican part of the Yucatán (Whitmore et al., 1996). Marfia et al. (2004) report spot samples of surface and groundwaters from Belize, but unfortunately none were from the New River Lagoon. Samples from the Río Hondo (see Fig. 1) had values of $-3.3\%$ $\delta^{18}O$ and $-1.0\%$ $\delta^D$ and $-3.4\%$ $\delta^{18}O$ and $-16.0\%$ $\delta^D$, respectively. Other reported $\delta^{18}O$ values for lakes in both the Mexican and Guatemalan part of the

The New River Lagoon

The New River Lagoon is about 35 km long, with an average width of 1 km and a maximum depth of about 10 m. The lagoon probably represents a widened valley section tracing a fault line, lying between the limestone Yaxbac Hills to the west and sandy Pliocene ridges to the east. Water samples indicated a pH of 7.3–8.1, electrical conductivity (EC) in the range 0.9–1.3 mS cm$^{-1}$, and Mg sulphate and carbonate chemistry. One water sample from the lagoon yielded isotopic values of $-1.0\%$ $\delta^{18}O$ and $-5.2\%$ $\delta^D$. This $\delta^{18}O$ is close to that recorded in precipitation at Petén Itza in Guatemala (Curtis et al., 1998) and lies between values for lake water at Coba and San José Chulchaca in the Mexican part of the Yucatán (Whitmore et al., 1996). Marfia et al. (2004) report spot samples of surface and groundwaters from Belize, but unfortunately none were from the New River Lagoon. Samples from the Río Hondo (see Fig. 1) had values of $-3.3\%$ $\delta^{18}O$ and $-1.0\%$ $\delta^D$ and $-3.4\%$ $\delta^{18}O$ and $-16.0\%$ $\delta^D$, respectively. Other reported $\delta^{18}O$ values for lakes in both the Mexican and Guatemalan part of the

Figure 1  Map of Belize showing the location of the New River Lagoon and other sites referred to in the text (A and B) and sites across the Yucatán Peninsula with (triangles) and without (circles) evidence for drought in the Late Classic (C)
Yucatán are significantly higher than those from the New River Lagoon, reflecting the fact that the lagoon is an open system.

Although freshwater today, there are isolated patches of mangrove along the shores of the New River Lagoon. Their origin is subject to debate and it has been suggested that they may be relict features from a period of higher sea level (Romney, 1959; Furley and Ratker, 1992).

Two areas of the lagoon – Hillbank (in the south) and Lamanai (in the mid section) (Fig. 1), were the focus of this study. The settlements of Hillbank and Lamanai are both on the west side of the lagoon, lying on significantly higher ground than the swampy east side, and hence more attractive for occupation. There are extensive ancient raised field complexes on the broad flood plain.

Hillbank has been extensively modified in the recent past, largely due to logging operations in the 19th and early 20th centuries. Most of the archaeological record has been destroyed, but the available data indicate that the site may have been occupied during the terminal Postclassic or early Spanish period. It is also believed to have been the site of Colcotz, one of the 15 Spanish colonial towns in Belize (Pendergast et al., 1993). Lamanai was a major Mayan site, possibly established as early as 1500 BC, with a substantial population by 300 BC (Pendergast, 1975). Major phases of construction seem to have occurred in the Preclassic (up to 300 BC) and in the Classic (ca. AD 250–600). Information on population levels is limited, but the peak was probably in the Preclassic and into the early Classic, when it may have reached 10,000 (Lambert et al., 1984). Ridged field systems seem to indicate intensive agriculture and maize seems to have been the agricultural staple (Lambert and Amason, 1978). According to Pendergast (1981), there is no evidence for any sort of collapse or abandonment of Lamanai in the late (or terminal) Classic Maya ‘collapse’. The Postclassic period shows a transformation of social organisation and technology, but no change in local occupation. There are extensive ancient raised field complexes on the broad flood plain.

Methods

Nine cores were collected from the New River Lagoon over two field seasons in 1999 and 2000. A Livingstone corer was used to collect longer cores and a mini-Kullenberg corer to retrieve shorter cores with the sediment–water interface intact. All cores were X-rayed at the British Geological Survey (Edinburgh), cut in half lengthways, photographed and the stratigraphy described. All core samples were kept in cold store. Three of the cores were selected for detailed analysis. In 1998, the Natural History Museum (NHM) London took a core from the southern part of the New River Lagoon for pollen analysis. We were given access to this core material (here called Hillbank 1998).

A chronological framework for the cores has been provided by radiocarbon (14C) and radionuclide measurements. Radiocarbon dating in areas of limestone bedrock is problematic. The difficulties of obtaining reliable radiocarbon dates were illustrated by many of the early lake sediment studies in the Petén region of Guatemala (e.g. Deevey et al., 1979; Leyden et al., 1994). The ideal solution would be to date plant macrofossils that have derived their CO2 from the atmosphere, but the cores from Belize yielded few plant macrofossils. Paired dates (plant macrofossils and carbonates from the same level in the core) can yield estimates of the hardwater error in a system, which can then be applied to correct carbonate dates from other levels in the same sequence. Paired dates were obtained from the Hillbank 1998 and Lamanai 1999 cores. These two pairs indicated a hardwater error of 1527 and 1660 a., respectively. The latter estimate was used to correct gastropod dates on the Lamanai 1999 core where only these were available. All dates reported here are accelerator mass spectrometry (AMS) dates (Table 1). For the purposes of comparison with the archaeological record, the 14C dates have been calibrated using CALIB4.4 (Stuiver and Reimer, 1993) to yield dates in calibrated years BP and years BC or AD.

210Pb dating was attempted to try to provide a chronology for events over the last 150 a. As well as calculating the level of unsupported 210Pb in the sediment, the anthropogenic radionuclide 137Cs was also measured. Radionuclide activities were measured by direct gamma spectroscopy using a Canberra low-background LGe detector housed in a graded Pb–Cd–Cu shield. Contiguous 1 cm samples were taken from the top of the cores, sealed in plastic Petri dishes and stored for a minimum of 3 weeks prior to analysis to ensure radioactive equilibrium between 226Ra and 222Rn. Total 210Pb was determined from its photopeak at 46.5 keV and the supported component was evaluated using the gamma emissions of 214Pb and 214Bi.

Unsupported 210Pb activity was calculated by subtracting the supported activity from the total activity. 137Cs was analysed via its photopeak at 661.7 keV. These dating methods were applied to three cores, but only Hillbank 2000 (see below) had sufficient levels of 210Pb for counting. The 210Pb fallout rate in northern Belize was found to be very low (0.09 pCi cm−2 a−1) compared with the global average (0.5 pCi cm−2 a−1). The reasons for this are not clear (cf. Kim and Rejmánková, 2002).

Table 1 AMS radiocarbon dates from the New River lagoon cores

<table>
<thead>
<tr>
<th>Site</th>
<th>Code</th>
<th>Depth (cm)</th>
<th>Material</th>
<th>14C a BP</th>
<th>δ13C VPDB±0.1</th>
<th>Cal. a BP (2σ)</th>
<th>Cal. a BC/AD (2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hillbank</td>
<td>CAMS-77198</td>
<td>113.5</td>
<td>Gastropod (G)</td>
<td>3990±40</td>
<td>−7.9</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>AA-42417</td>
<td>113.5</td>
<td>Organic matter (OM)</td>
<td>2463±48</td>
<td>−27.6</td>
<td>2359–2621</td>
<td>789–402 BC</td>
</tr>
<tr>
<td></td>
<td>AA-39721</td>
<td>153</td>
<td>OM</td>
<td>4752±66</td>
<td>−30.3</td>
<td>5440–5600</td>
<td>3650–3370 BC</td>
</tr>
<tr>
<td></td>
<td>AA-39722</td>
<td>400</td>
<td>OM</td>
<td>130±55</td>
<td>−28.0</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>CAMS-45870</td>
<td>990</td>
<td>OM</td>
<td>6020±50</td>
<td>—</td>
<td>6730–6989</td>
<td>5046–4744 BC</td>
</tr>
<tr>
<td></td>
<td>CAMS-77197</td>
<td>1242</td>
<td>OM</td>
<td>9840±60</td>
<td>−27.6</td>
<td>11 160–11 340</td>
<td>9140–8600 BC</td>
</tr>
<tr>
<td>Lamanai</td>
<td>AA-35787</td>
<td>38.5</td>
<td>OM</td>
<td>810±40</td>
<td>−32.4</td>
<td>AD 1224, 1231, 1239</td>
<td>AD 1161–283</td>
</tr>
<tr>
<td></td>
<td>CAMS-77196</td>
<td>38.5</td>
<td>G</td>
<td>2470±40</td>
<td>−6.9</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>SUERC-4104</td>
<td>72–75</td>
<td>G</td>
<td>3260±50</td>
<td>−6.7</td>
<td>[ca. AD 400]</td>
<td>[ca. AD 400]</td>
</tr>
<tr>
<td></td>
<td>SUERC-4108</td>
<td>178–180</td>
<td>G</td>
<td>4445±35</td>
<td>−5.6</td>
<td>—</td>
<td>[ca. 900 BC]</td>
</tr>
<tr>
<td></td>
<td>AA-35786</td>
<td>259</td>
<td>OM</td>
<td>3070±50</td>
<td>−27.1</td>
<td>3159–3383</td>
<td>1433–1113 BC</td>
</tr>
<tr>
<td></td>
<td>CAMS-77195</td>
<td>312</td>
<td>OM</td>
<td>3440±40</td>
<td>−28.6</td>
<td>3629–3780</td>
<td>1880–1616 BC</td>
</tr>
<tr>
<td>Outpost</td>
<td>CAMS-77200</td>
<td>13</td>
<td>OM</td>
<td>360±40</td>
<td>−29.9</td>
<td>314–411</td>
<td>AD 1442–1642</td>
</tr>
</tbody>
</table>

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Table 2  Basic data for cores from the New River lagoon described in the text

<table>
<thead>
<tr>
<th>Core</th>
<th>Core length (cm)</th>
<th>No of $^{14}$C dates</th>
<th>Proxies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hillbank 1998</td>
<td>1400</td>
<td>7 (1 paired)</td>
<td>$\delta^{18}$O, $\delta^{13}$C, diatoms</td>
</tr>
<tr>
<td>Hillbank 2000</td>
<td>70</td>
<td>0</td>
<td>$\chi$, XRD, %CaCO$_3$, $\delta^{18}$O, $\delta^{13}$C, diatoms</td>
</tr>
<tr>
<td>Lamanai 1999</td>
<td>319</td>
<td>6 (1 paired)</td>
<td>$\chi$, LOI, XRD, XRF, %CaCO$_3$, $C/N$, $\delta^{18}$O, $\delta^{13}$C, av. P, diatoms</td>
</tr>
<tr>
<td>Outpost 2000</td>
<td>74</td>
<td>1</td>
<td>$\chi$, %CaCO$_3$, $\delta^{18}$O, $\delta^{13}$C</td>
</tr>
</tbody>
</table>

The four cores used for this study (Hillbank 1998, Hillbank 2000, Lamanai 1999 and Outpost 2000) were analysed using a range of proxies (summarised in Table 2). It was not possible to apply all proxies to all cores. The analysis of fossil diatom assemblages was the main focus of the study.

Magnetic susceptibility ($\chi$) was measured on all cores except Hillbank 1998 using a loop sensor (Bartington Instruments). Where sufficient core material was available, mass specific measurements were also carried out. In carbonate-rich sediments susceptibility is usually very low, but influxes of clay or iron-rich soils from catchment slopes may result in higher values. Loss-on-ignition (LOI) was carried out at 550°C and 900°C to obtain estimates of percentage organic matter plus bound water and carbonate content, respectively (Dean, 1964). A separate estimate of inorganic carbon (carbonate) content was obtained using a carbonate bomb (Müller and Gastner, 1971).

A limited number of samples were analysed using X-ray diffraction (XRD) to estimate bulk mineralogy and X-ray fluorescence (XRF) for elemental composition. XRD analyses were carried out on powdered samples using a Philips PW1800 automated diffractory system. Mineralogical information can help to identify different sediment sources, with different clay minerals forming under different combinations of parent material and weathering conditions. XRF analysis was carried out using a Philips Panalytical PW2404 wavelength dispersive XRF spectrometer. The relative abundances of the major oxides can also help to identify sediment sources.

C/N ratios (organic carbon/total nitrogen) were determined using a Carlo-Elba NA 1500 Elemental Analyser. Larger than normal samples (40 mg) were needed to yield reproducible nitrogen values (Meyers and Lallier-Vergès, 1999). The separation of aquatic from terrestrial sources is assisted by the use of C/N where values $<$10 are taken as typical of aquatic vegetation (Brenner et al., 1999).

For stable isotope analysis ($\delta^{18}$O, $\delta^{13}$C) of bulk carbonate, CO$_2$ was liberated from 1 mg of sample using a conventional method (McCrea, 1980; Wachtler and Hayes, 1985). The resulting gas was analysed in an AP2003 triple-collector mass spectrometer. Precision and accuracy are better than ±0.2% at 1σ. Data are reported relative to VPDB. Some analyses were also carried out on gastropods of the genera Cochliopina and Pyrgophorous. These genera had been used for previous isotopic studies in the Yucatan peninsula by Covich and Stuiver (1974) and by Curtis et al. (1996, 1998). Both are typical of freshwater, relatively deep lakes in this area (Covich, 1983). Individual taxa were not abundant in the cores, but it was felt useful to compare the results from individual organisms with those obtained from bulk carbonate. Measurements were carried out on thoroughly cleaned, adult specimens using a VG isogas PRISM II mass spectrometer. This has a higher sensitivity than the AP2003.

Oxygen isotope records of lacustrine carbonates have been used as the main palaeoclimatic proxy for other lake studies across the Yucatan peninsula (Covich and Stuiver, 1974; Hodell et al., 1991; Curtis et al., 1996; Rosenmeier et al., 2002a; Hodell et al., 2005a,b). In closed basin lakes, evaporative concentration causing enrichment in $^{18}$O and more positive $\delta^{18}$O values is often the dominant signal. In hydrologically open systems, however, a range of other factors, such as precipitation source region or temperature, may be more important (Leng and Marshall, 2004). The analysis of either bulk carbonate or of individual carbonate microfossils (in this case gastropods) can result in problems of interpretation (Ito, 2001). The calcite in bulk sediment samples may come from different sources, or reflect seasonal rather than annual average conditions. Based on the one $\delta^{18}$O water value obtained from the New River Lagoon and using the annual average temperature for this area of Belize (26°C) calcite precipitating from the lagoon today would have a value of $\sim3.2\%e$ $\delta^{18}$O VPDB (based on Hays and Grossman, 1991). This value is similar to the mean value of the bulk carbonates in the cores presented here (see below). Taken together with the variation between calcite and aragonite in different sections of the cores, we believe that we can assume that the bulk carbonate is authigenic in origin. Gastropods will record isotope composition over a period of time and may have values affected by habitat-specific, localised conditions. In this case, gastropods were not found throughout the cores and using bulk samples allowed complete records to be obtained. In addition, carbon isotopes ($\delta^{13}$C) reflect provenance and changes in lake total dissolved inorganic carbon (TDIC). This is further influenced by within-lake processes such as productivity of the lake (e.g. Curtis et al., 1998).

Samples were prepared for diatom analysis using a standard method (Battarbee, 1986). A minimum of 400 valves were counted for each level using a 1000× oil immersion lens with bright field illumination as standard. Diatoms were identified using the standard published florae (e.g. Patrick and Reimer, 1966; Kramer and Lange-Bertalot, 1986, 1988, 1991a,b). Floras from other tropical areas were particularly valuable (e.g. Foged, 1984; Gasse, 1986). Results (as percentages) were plotted using the software C2 and zoned using CONISS. The diatom data from the cores were also analysed using detrended correspondence analysis (DCA).

Modern diatom and water samples were collected from a range of Belizean lakes to assist in the interpretation of the fossil diatom assemblages. Collections were made initially from 28 sites (Breen, 2002), with a further 17 being sampled in 2003 (Breen and Furley, 2003). Electrical conductivity (EC) and pH were recorded for all sites. Modern diatom and water chemistry results were analysed using canonical correspondence analysis (CCA). The theft of some samples meant that the water chemistry dataset was insufficient to develop transfer functions for major environmental parameters. Instead, we have applied the transfer function for EC developed by Reed (1998). Reed's study was used in preference to that of Whitmore et al. (1996) as it included many more of the species found in the New River Lagoon cores.

Results: Hillbank

The Hillbank 1998 core was 1381 cm long. No detailed stratigraphic description of the core was made by the NHM and...
the core was not photographed. The sediments were largely silts and clays, with some sand-size material in the upper 900 cm. Two of the core segments (1339–1290 and 1198–1128 cm) were lost and were not available for sampling for this project. In spite of this, it remains the longest record from the New River Lagoon, although discontinuous below 1127 cm. The core has a chronology based on $^{14}$C dates from five levels (see Table 1). Additional dating samples were collected by the NHM, but were not available for this study. The date from 400 cm is anomalously young and has been excluded from discussion.

The rate of sedimentation at Hillbank is problematic. Plotting an age–depth curve (not shown) indicates that the top of the sequence may be missing. Coring in the lagoon showed that the top sediments are highly unconsolidated and it is quite possible that the top is missing. A linear regression of the dates believed to be reliable yields an $r^2$ of 0.8, but with an intercept of 2780 $^{14}$C a BP. Calculating sedimentation rates between pairs of dates indicates major changes in accumulation rate believed to be reliable yields an age–depth curve (not shown) indicates that the top of the core corresponds to diatom zone 2. The samples from 1381 and 1349 cm are also offset from the bulk of the core. These samples show strong positive covariance, which may indicate that the lagoon was hydrologically closed at this stage.

Above 1198 cm, $\delta^{18}$O values are more stable at around $-3.5\%$. In the bulk carbonate there are low values ($-4.1\%$) at 630 cm. Values then remain fairly stable until 193 cm, above which bulk $\delta^{18}$O values oscillate between $-4.3\%$ and $-3.3\%$ to the top of the core. The $\delta^{18}$O$_{\text{gastropod}}$ record is broadly similar to the bulk carbonate in terms of both absolute values and trends. An unidentified species of gastropod, present only in the upper part of the core, showed even stronger variability than the bulk samples in the upper 193 cm.

Bulk carbonate $\delta^{13}$C values through the core are generally in the range $-2\%$ to $-4\%$ mean $(-3.2\%)$. There are three major, low-value departures from this at 1230 to 1210 cm ($-4.8\%$ to $-8.1\%$), 109 cm ($-5.3\%$) and in the top 20 cm ($-4.6\%$) ($<2359–2621$ cal. a BP). These are all more than 1 SD less than the mean, with the results from 1220 and 1210 cm being more than 2 SD less. $\delta^{13}$C is only $>1$ SD higher than the mean at $80\%$ ($-1.87\%)$.

The $\delta^{13}$C$_{\text{gastropod}}$ record from Cochliopina sp. is most like that from the bulk carbonate (as with the oxygen isotopes), although the values are generally more negative. The results from the unidentified gastropod are the most extreme (range $-9.5$ to $+0.7\%$), highlighting species-specific responses.

The 70 cm mini-Kullenberg core Hillbank 2000 (Fig. 4) was taken about 500 m from the site of Hillbank 1998 to ensure that the surface sediments were captured (see above). The core’s most distinctive feature is a lens of dark grey-brown clay between 30 and 23.5 cm. The results of $^{210}$Pb dating of this core added little further information. Low and relatively constant specific activities of $^{137}$Cs from the surface to 12 cm (Fig. 4) suggest that the sediment was subject to mixing. As a result, it has not been possible to derive a chronology from the radionuclide profiles. Input of $^{137}$Cs occurred mainly due to atmospheric testing of nuclear weapons through the 1950s and 1960s and its presence in the upper part of this core confirms that these sediments are recent.

The Hillbank 2000 core was analysed for magnetic susceptibility, percentage carbonate and by XRD, with the focus on the top 30 cm. The lower part of the core (70–30 cm) was dominated by carbonate (generally in the range 79–84%). This was confirmed by the XRD results, which indicated 95–100% calcite. In the top 30 cm, carbonate values were between 40% and 50% except in the clay lens, where they dropped to 33%. XRD data show that the lens was low in calcite (35–40%), with smectite (15–25%) and more quartz (15–25%). Away from the clay lens, calcite dominated.

Loop $\chi$ values peak in the clay unit (Fig. 4). Mass-specific values are only available for part of the core (the uppermost
Figure 2 Percentage diatom diagram Hillbank 1998. Dashed lines denote sections of core with no diatoms preserved.
Figure 3  Stable isotope data from Hillbank 1998 based on bulk carbonate and gastropods
Figure 4
Selected results from Hillbank 2000
Interpretation and discussion: Hillbank

Taken together, the two Hillbank cores provide a record of change from the latest Pleistocene to the recent past, although the resolution and dating of this sequence has been restricted by the missing samples. The diatoms present are mainly those common in modern samples from Belize, with the exception of those in diatom zone 2. Oxygen isotopes generally indicate a hydrologically open system (as today), except below 1248 cm. The data from the latest Pleistocene/early Holocene clearly reflect very different conditions from those prevailing through the rest of the Holocene and up to the present day.

The base of the Hillbank 1998 core, dating to the latest Pleistocene (diatom zone 1) indicates an open, freshwater system. δ¹⁸O values are close to the average for the whole core. Above a missing core segment, the section of core up to about 1200 cm, records the most significant changes in the whole sequence. High δ¹⁸O values, particularly between 1290 and 1245 cm, indicate evaporative concentration. Diatom valves are not preserved, possibly reflecting dissolution under high alkalinitities. When diatoms are preserved, the flora is dominated by *A. granulata*, a species found only in swampy conditions in modern diatom samples from Belize. The existence of swampy conditions with oxidation of organic matter may be confirmed by the occurrence of low δ¹³C values (−8 to −7‰). The date of 11 160–11 340 cal. a BP from 1160–11 340 cal. a BP from 1242 cm indicates that dry conditions and low water levels occurred in the very early Holocene, possibly starting in the latest Pleistocene and up to the present day. The last 270 cm of the core (covering about the last 5600 cal. a BP) indicates greater environmental variability in the later Holocene. In this part of the core (diatom zone 6), there are three periods with no diatom preservation. In this case, this cannot easily be attributed to increased evaporation, as there is a general shift to lower (more negative) δ¹⁸O values.

Results: Lamanai

The Lamanai 1999 core (319 cm long) has been the most intensively studied of those presented here (Table 2). Selected data are presented (Figs 5 and 6). The sediments were mainly greyish-brown silts and clays, with accumulations of gastropods, particularly between 180–179, 154–153, 75–72 and 30–20 cm. Dating control for Lamanai 1999 is provided by six ¹³C dates on five levels (see Table 1). Two of the dates were on gastropods only (180–178 and 75–72 cm) and the hard-water correction of 1660 a has been applied in this discussion. The age–depth curve for this core (with and without the corrected dates) is approximately linear (r² = 0.99 for OM dates only, 0.94 including corrected dates). As with the Hillbank 1998 core, the surface sediments are assumed to be missing and a date of AD 1500 has been assigned to the top of the core; the average rate of sedimentation is ~0.1 cm a⁻¹. For ease of comparison with the archaeological record, the results from this core are plotted against age (BC/AD).

Results from the geochemical analysis of this core are shown in Fig. 5. XRF measurements were made every 10 cm; LOI, percentage carbonates, available P and C/N every 5 cm and stable isotopes every 5 cm. XRD measurements were carried out on 17 samples scattered through the core. As might be expected, carbonates dominate the sequence (62–84.5%). The XRF results show that SiO₂ is next in abundance to carbonates, ranging from 27.4% to 1.3%. SiO₂ is generally most abundant before about 900 BC and has a secondary peak between ca. 170 BC and AD 270. Values since about AD 1240 are the lowest in the whole sequence. Very similar overall patterns are shown by Al₂O₃, Fe₂O₃ and TiO₂ values (not illustrated) and magnetic susceptibility (χ), indicating that similar processes are controlling the concentrations of these non-carbonate sediments. Larger inputs of allochthonous material in the lower part of the core are the most likely explanation and could be derived from Pliocene sands and gravels to the east. Linear interpolation of the available dates indicates that this occurred prior to ca. 1300 BC. The XRD results show that the core is dominated by calcite, with some aragonite, quartz and smectite. Smectite was found only in the lower part of the core (below 260 cm, >1433 to 1132 BC) and in a sample from 146.5 cm (ca. 350 BC). Aragonite was present in low concentrations in many of
Figure 5  Age–depth plot of selected geochemical and isotope results from Lamanai 1999
Figure 6  Age–depth percentage diatom diagram, Lamanai 1999. Dashed lines denote sections of core with no diatoms preserved.
the samples. The highest percentage of aragonite (40%) was recorded in a sample from 177.5 cm (ca. 350 BC), but concentrations of >5% were present since about AD 500. Quartz also peaked at 177.5 cm (20%) and was again present in small amounts in most samples.

The results of the diatom analysis are shown in Fig. 6. CONISS identified seven major zones, although some of these are barren of diatoms either completely (zones 3 and 6), or in part (zone 1). Diatoms are poorly preserved through zone 1 (ca. 1800–800 BC). D. elegans, D. tenuis and Nitzschia amphibia var. rostrata are the dominant species. D. tenuis is not common in modern samples from the lagoon or other sites across Belize. In modern samples, N. amphibia spp. formed significant proportions (27% and 15%) of assemblages in two sites sampled for modern diatoms (pH 7.72 and 7.45, EC 0.44 and 2.22 mS cm⁻¹, respectively). Denticula tenuis is hardly present in the rest of the sequence. Zone 2 (221–181.5 cm, ca. 800–390 BC) has increased percentages of M. smithii var. lacustris. M. smithii and Cyclotella distinguenda are also present in lower percentages. C. distinguenda was not found in modern samples from the lagoon, but dominated (54%) the modern sample with 15% N. amphibia. M. smithii was very rare in modern samples. Zone 3 (181.5–172.5 cm, ca. 390–300 BC) was barren of diatoms and corresponded to the maximum of aragonite and a peak in quartz. Diatoms were preserved in zone 4 (172.5–74 cm, ca. 300 BC to AD 700) and species are more diverse. C. distinguenda, M. smithii var. lacustris and Achnanthes exigua reach their highest percentages in this zone. A. exigua is largely confined to this part of the core. Zone 5 (74–40 cm, ca. AD 700 to 1070) shows an abrupt change in flora. An unknown (Species 15) dominates at the base of the zone, but the remainder is mainly D. elegans accompanied by Cyclotella plitivensis. The ecology of C. plitivensis is not well known and it was not found in the modern samples collected in Belize. Above this, zone 6 (40–9 cm, ca. AD 1070–1400) is again barren. Two samples from the top of the core make up zone 7 dominated by D. elegans and a variety of forms of Brachysira neoexilis.

The results of oxygen and carbon isotope analyses of bulk carbonates and C/N are shown in Fig. 5. Oxygen isotopic values lie in the range −5.3 to −3.2‰ (mean −4.0‰). These are similar to those in the top of Hillbank 1998 and in Hillbank 2000. Values are slightly higher since about 600 BC (mean −3.9‰), than in the bottom of the core (mean −4.2‰), but only one sample falls more than 1 SD from the mean. Carbon isotopic values show much greater variability (−1.9 to −5.8‰, mean −3.5‰), but there are two broad zones above and below ca. AD 1000. The more negative values after this (mean −4.9‰, mainly more the 1 SD below the mean) are more akin to those in Hillbank 2000. The mean for the lower part of the core is −3.0‰. Between ca. 320 BC and AD 1000 there are three excursions to higher values (in the range −1.2 to −2.8‰). C/N values in the core are all above 10, indicating that higher plant material is a significant source of organic material. Apart from the top two samples, values range from 21.6 to 44.6. There is no obvious relationship between the higher values of δ¹³C and changes in C/N. At the top of the core, however, the most negative δ¹³C do correspond to the lowest C/N ratios, but the shift to more negative δ¹³C occurs before the C/N changes.

A mini-Kullenberg core (Outpost 2000) was again taken near the longer core to capture the surface sediments. The sediments are similar to those in the Lamanai core. The core only has one date from 13 cm, ca. AD 1440–1640, and a limited range of analyses was carried out. Once again, the sediments are dominated by carbonates (74–83%). Magnetic susceptibility measurements made using the loop detector gave negative values throughout. Bulk carbonate and gastropod stable isotope results are shown in Fig. 7. The δ¹⁸O record (bulk sediment) has very similar values to those in Lamanai 1999. The results from Pyrgoophorus sp. show enriched δ¹⁸O above 45 cm and particularly between 35 and 10 cm. Cochliopina sp. shows a similar, but less pronounced, trend. Unfortunately an isotope record from gastropods is not available from Lamanai 1999 to compare with the differences between the bulk and gastropod data from the Outpost core.
Bulk $\delta^{13}$C values are similar to those in the top 50 cm of Lamanai 1999. From the base to 40 cm in the Outpost core, the results are very stable, with values averaging near $-4\%o$. Above 40 cm values are more variable, but with a negative trend, reaching $-8.1\%o$ at 5 cm. The gastropod values are generally more negative (as was the case in Hillbank 1998) and do not show the same trends as the results from the bulk sediment.

**Interpretation and discussion: Lamanai**

The cores taken close to the archaeological site of Lamanai provide a more complete and detailed record of change over the last 4000 a than the Hillbank cores. Again, diatom taxa in the sediments are generally the same as those found today, here with the exception of *D. tenuis* and *C. pitivensis*. $\delta^{18}$O values are similar to those from the Holocene sediments from Hillbank 1998, with only *Pyrgophorus* from recent sediments in Outpost 2000 showing significant enrichment. Even this, however, was not of the same magnitude as the enrichment at the base of the Hillbank 1998 core.

The Lamanai 1999 core covers the period since about 1800 BC, with a much less variable accumulation rate than Hillbank 1998. Only the top 270 cm of Hillbank 1998 overlaps with the Lamanai 1999 core. The lower part of the Lamanai 1999 core is marked by the highest inputs of allochthonous materials such as aluminium, iron and silicon. Diatoms are poorly preserved in zone 1. Here, as throughout the whole core apart from the very top, the C/N ratio suggests significant inputs of terrestrial organic material. Around 350 BC there is a marked change. Diatoms are not preserved, the highest level of aragonite (40%) is recorded and there is a peak in quartz. It seems likely that these indicate increased evaporation, but this is not reflected by higher $\delta^{18}$O, except through the broad trend of slightly increased values after ca. 600 BC. Between ca. 170 BC and AD 270 there is a secondary peak in silica, iron and aluminium. Magnetic susceptibility increases and then declines. Taken together, these indicate increased erosion from the catchment. The timing of this is consistent with building at Lamanai. It has been suggested that the terminal Preclassic to early Classic was marked by drying, as well as significant anthropogenic disturbance of the environment (Hodell et al., 2007). Our data are consistent with this.

At AD 700 there is an abrupt change in the diatom flora and $\chi$ shows a general decline. This may reflect the cessation of the major Classic phase of construction at Lamanai. The three peaks in $\delta^{13}$C between ca. 320 BC and AD 1000 in this core may indicate increasing productivity in response to higher nutrient inputs from the growing Maya population.

A further change is indicated after ca. AD 1070. The loss of diatoms may indicate higher alkalinites associated with more evaporation, but this is not apparent in the $\delta^{18}$O record. Magnetic susceptibility is extremely low and the rate of sediment accumulation is slower here than in other parts of the core. There is a marked shift in $\delta^{18}$C to lower values and the top few C/N ratios are the lowest in the core, indicating a greater contribution from aquatic organic matter. Taken together, these data may reflect reduced activity at the Lamanai site, less erosion and the recovery of forest in the Postclassic and Hispanic periods.

Although the record from Lamanai 1999 does seem to indicate periods of drier climate, there is no clear evidence for this in the $\delta^{18}$O record from bulk sediment. None of the values here are as high (enriched) as those in the lower part of Hillbank 1998. The widely reported drought of the late Classic seems to have left little trace in the Lamanai core, which is consistent with the archaeological interpretation of continuous settlement. The gastropod isotope data from the upper half of Outpost 2000, however, do show markedly higher values. Based on the one radiocarbon date available for this core, this period of isotopic enrichment started ca. AD 970, which does correspond with the late Classic. Rosenmeier et al. (2002b) have suggested that vegetation clearance, with resulting increased runoff, can influence $\delta^{18}$O records, but in the absence of pollen data we cannot test this hypothesis.

**Conclusions**

The results from this study need to be compared with those already produced for the circum-Caribbean. The record from Hillbank indicates dry conditions before 11 160–11 340 cal. a BP, in the earliest Holocene (or latest Pleistocene). The lagoon may not have had an outflow. Conditions significantly drier than present are also recorded in Quexlabel and Petén-Itzá in Guatemala (Leyden et al., 1993; Hillesheim et al., 2005) and Lake Miragoane, in Haiti (Hodell et al., 1991). Some records from the Basin of Mexico also show dry conditions at this time.

The transition to relatively moist early Holocene conditions was completed at Hillbank by 9500–10 000 cal. a BP. In general terms, there is a period of lake filling in Guatemala and the Yucatán Peninsula ranging from before 8000 cal. a BP (San José Chulchaca; Leyden et al., 1996) to 10 200 cal. a BP (Lake Petén-Itzá; Curtis et al., 1998). This change has been attributed to increased moisture availability as the mean latitude of the ITCZ moved northwards (Haug et al., 2001) and to sea level rise, which pushed up freshwater aquifers (Fairbanks, 1989; Watts and Hansen, 1994). The Hillbank 1998 record shows that the lagoon was freshwater for the whole of the record (inferred from the diatoms and gastropods). There is no evidence for seawater incursions as recorded at nearby Cobweb swamp. The discontinuous patches of mangrove have been considered as relic formations whose present-day distribution results from waterborne transport in flood conditions.

Moist and quite stable conditions are recorded in the New River Lagoon from the early Holocene to 5600 cal. a BP. The mid Holocene in Petén-Itzá (7700–5400 cal. a BP) and Lake Miragoane (7900–5900 cal. a BP) is also characterised by wetter conditions. The wettest period of the Holocene in San José Chulchaca centred on ca. 5800 cal. a BP (Leyden et al., 1996; Whitmore et al., 1996). After the mid Holocene, conditions in the New River Lagoon seem to have become more variable.

The New River Lagoon cores give some indication of disturbance in the late Preclassic/early Classic, but we see no evidence for drought at the time of the Maya hiatus (ca. AD 530–590). There are very limited signs of the late Classic drought associated with the Maya ‘collapse’. Although the late Classic dry period is thought to be wide in its extent (Hodell et al., 1991; Horn and Sanford, 1992; Metcalfe et al., 1994; Metcalfe, 1995; Hodell et al., 1993; Curtis et al., 1996; Haug et al., 2003), there is increasing recognition of its complexity, both temporally and spatially. Higher resolution studies of sites where late Classic drought is evident have shown that the period was characterised by multiple dry events (Hodell et al., 2005a, 2007). Spatial variability is also important; there is only one record in Guatemala from Lake Salpetén (Rosenmeier et al., 2002a) that has preliminary evidence for drying at this time. This episode is not detected at all in Lake Petén-Itzá (Curtis et al., 1998) (Fig. 1(C)). Curtis et al. believe that this is due to the large size of the lagoon, which would render the
sequence much less sensitive to fairly short-lived shifts in climate. The suggestion that the size of the New River Lagoon cushioned any impact of dry conditions in the late Classic seems to be confirmed by the δ18O record (authors’ unpublished data) from the nearby, much smaller, Honey Camp Lagoon (Fig. 1A) and (C), which shows a strong enrichment at this time.

Pendergast (1987) has given three possible reasons for the continuity of occupation at Lamanai: its location at the edge of a large lake providing a rich and varied diet; that the lagoon provided the inhabitants with an easy means of communication with the northern Maya area and other parts of Mesoamerica facilitating trade; and thirdly, undetectable factors such as the strength and personality of the community’s leaders, which would have provided stability at a time of crisis. Our new results indicate a fourth reason: that the New River Lagoon was such a large body of water that it was relatively insensitive to climatic variations over the Holocene. Hence, the population living along its shores would have benefited from a reliable source of water.

The New River Lagoon appears to have been a permanent water body for much longer than many of the lakes in the Yucatán and therefore has the potential to produce a record extending back towards the last glacial maximum. The latest glacial/earliest Holocene appears to have been a period of marked variability in Belize, when conditions may have been drier than at any other time in the Holocene. The influence of lower sea levels must be considered, however. The relative magnitude of environmental changes in the transition from glacial to interglacial conditions compared with the more widely discussed impacts of the droughts of the late Classic period warrants further investigation across the circum-Caribbean region.

In spite of some limitations with the data, the cores from the New River Lagoon represent the longest and most detailed palaeolimnological record from Belize yet produced. Future work could resolve some of the outstanding issues raised by this study.

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