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Citation for published version:
Henderson, GM, Robinson, LF, Cox, K & Thomas, AL 2006, 'Recognition of non-Milankovitch sea-level highstands at 185 and 343 thousand years ago from U-Th dating of Bahamas sediment' Quaternary Science Reviews, vol 25, no. 23-24, pp. 3346-3358., 10.1016/j.quascirev.2006.03.003

Digital Object Identifier (DOI):
10.1016/j.quascirev.2006.03.003

Link:
Link to publication record in Edinburgh Research Explorer

Document Version:
Author final version (often known as postprint)

Published In:
Quaternary Science Reviews

DOI: 10.1016/j.quascirev.2006.03.003

Recognition of non-Milankovitch sea-level highstands at 185 and 343 thousand years ago from U-Th dating of Bahamas sediment

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Abstract

Thirty-one new bulk-sediment U-Th dates are presented, together with an improved $\delta^{18}O$ stratigraphy, for ODP Site 1008A on the slopes of the Bahamas Banks. These ages supplement and extend those from previous studies and provide constraints on the timing of sealevel highstands associated with marine isotope stages (MIS) 7 and 9. Ages are screened for reliability based on their initial U and Th isotope ratios, and on the aragonite fraction of the sediment. Twelve 'reliable' dates for MIS 7 suggest that its start is concordant with that predicted if climate is forced by northern-hemisphere summer insolation following the theory of Milankovitch. But U-Th and $\delta^{18}O$ data indicate the presence of an additional highstand which post-dates the expected end of MIS 7 by up to 10 kyr. This event is also seen in coral reconstructions of sealevel. It suggests that sealevel is not responding in any simple way to northern-hemisphere summer insolation, and that tuned chronologies which make such an assumption are in error by ≈10 kyr at this time. U-Th dates for MIS 9 also suggest a potential mismatch between the actual timing of sealevel and that predicted by simple mid-latitude northern-hemisphere forcing. Four dates are earlier than that predicted for the start of MIS 9. Although the most extreme of these dates may not be reliable (based on the low-aragonite content of the sediment) the other three appear robust and suggest that full MIS 9 interglacial conditions were established at 343 ka. This is ≈8 kyr prior to the date expected if this warm period were driven by northern-hemisphere summer insolation.

1.0 Absolute dating of Pleistocene sealevel changes

Sealevel represents a first-order feature of the surface earth environment with significant implications for global climate (Shackleton, 1987). It both responds to climate, through processes such as temperature, precipitation and seasonality, and influences it, through processes including ocean circulation and moisture supply (Lambeck et al., 2002). The global nature of sealevel change also makes it a powerful stratigraphic tool, allowing correlation between distant records if they contain an imprint of past sealevel.

For several decades, the assumption has been that sealevel during the Pleistocene is closely related to summer insolation at 65^{o}N, following the theory of (Milankovitch, 1930). To test this assumption, and to assess the mechanisms driving climate change, requires development of a history of sealevel changes with an absolute chronology independent of any climate model. Prior to about 40 ka, when the ^{14}C chronometer no longer provides information, such independent chronology comes exclusively from U/Th dating. Other indicators have the potential to tell us about the amplitude of change, including sedimentology (Murray-Wallace 2002) and oxygen isotopes in the Red Sea (Siddall et al., 2003) but only material that can be reliably U/Th dated provides information about the timing of these changes.

Corals and cave carbonates (speleothems) have provided the best U/Th constraints on Pleistocene sealevel. Corals grow in seawater and so constrain the minimum sealevel. Some species are known to have growth restricted to the upper 5 to 10 meters, thus providing a sealevel datum with some precision. Corals are dateable with U/Th techniques, but older samples become progressively prone to alteration, particularly when subject to sub-aerial exposure. The difficulty of finding pristine examples of appropriate coral species has limited the application of U/Th dating, but corals have, nevertheless, formed the major tool on which Pleistocene sealevel has been established (e.g. Bard et al., 1996; Gallup et al., 1994; Mesolella et al., 1969; Stirling et al., 1998; Stirling et al., 2001). The recent development of an alpha-recoil model to correct for some diagenetic alteration somewhat alleviates this limitation (Thompson and Goldstein, 2005). But coral data can still be difficult to interpret because of the nature of reef development. Corals do not grow continuously upward, as sediment cores do, but respond in a complex way to changes in sealevel. Careful field work is required to put the corals in context, and to assure that they are found *in situ*, rather than above their formation location in a storm deposit, or below in a fall. Corals are also prone to dissolution and erosion. Powerful examples of this have been seen recently in the Pacific after El Niño warming bleached coral reefs. Huge volumes of these reefs subsequently eroded within only a few years (Eakin, 1996; Reake-Kudla et al., 1996). The corals left for us to date therefore represent a preservationally biased sample of those that have escaped erosion and dissolution. Corals provide powerful sealevel constraints, but they are not perfect.

Most forms of speleothem grow in air so their presence indicates that sealevel was below them when they formed. Investigating diagenesis of speleothem carbonate is more difficult than for corals because of the lack of knowledge about the expected ($^{234}U/^{238}U$) value at the time of their formation. But speleothems frequently consist of dense calcite with large crystals and are expected to be diagenetically robust. The presence of dated speleothem therefore provides an upper bound on sealevel. Absence of speleothem, on the other hand, does not provide sealevel information because their formation can be prevented by other variables (e.g. moisture levels or CO_2 degassing).

The application of both coral and speleothem records to constrain past sea level also requires detailed knowledge of the uplift/subsidence history of the region in which the samples are found. This is especially important in tectonically active regions.

The $\delta^{18}O$ record captured in deep-sea cores contains a significant sealevel signal. Chronology of this record could therefore provide information about sealevel change, particularly if other controls on $\delta^{18}O$ (temperature and local salinity changes) could be deconvolved. Such a chronology would have the significant advantage that $\delta^{18}O$ records in the deep ocean form continuously, without the sporadic formation and preservation seen for coral and speleothems. Sadly, techniques to date most marine cores with sufficient precision have not yet been established. Early attempts to use $^{231}Pa/^{230}Th$ (Broecker and Van Donk, 1970) provided ages, but with uncertainties that were large and difficult to assess. More recent application of the ^{230}Th excess method (Francois et al., 2004) provides information about sediment accumulation rate, but cannot directly date past events. In general, it is this excess of ^{230}Th in marine sediments which prevents their dating with U/Th techniques. Sediments are, in effect, the exact opposite of corals and speleothems. The insoluble nature of Th means that, while carbonates precipitating from water have very low ^{230}Th, sediment underlying seawaters have high ^{230}Th. Variations in the amount of excess ^{230}Th with time prevent the precise use of this tool to assess past ages.

Carbonate platform settings, such as the Bahamas, provide a compromise between the advantages of corals and deep-sea cores. A large fraction of the sediment formed in these areas is carbonate, directly precipitated from seawater and therefore with very low initial Th concentrations. This sediment accumulates at least semi-continuously, providing more stratigraphy than either corals or speleothems. The first work on this material demonstrated that it had sufficiently low initial ^{230}Th, particularly during highstand periods, to allow dating of the bulk sediment from the last interglacial (Slowey et al., 1996). Further work explored the use of isochrons to date the material, and established some limits on the application of the technique to Bahamas sediments (Henderson and Slowey, 2000; Henderson et al., 2001). That work indicated that the penultimate deglaciation (Termination 2) occurred earlier than suggested by traditional Milankovitch models of climate change, thereby challenging our understanding of the mechanisms driving glacial-interglacial cycles. Recovery of longer cores during ODP Leg 166 allowed the first application of these dating techniques to earlier highstands with age constraints on Marine Isotope Stage (MIS) 7 (Robinson et al., 2002). This paper continues that work, providing additional constraints on the timing of MIS 7 and new information about the timing of highstands associated with MIS 9.

2.0 Sedimentary setting and samples

Samples for this study were taken from ODP Leg 166, Site 1008A. This Site was drilled on the leeward (SW) slope of the Great Bahama Bank (GBB) at 23°36.64' N, 79°5.01'W in 437 m of water. Mineralogical data indicate clear glacial-interglacial cycles in the aragonite fraction of the sediment (Malone, 2000). These are caused by flooding of the GBB during interglacial periods providing a large area for growth of aragonite-precipitating organisms such as *Halimeda* algae. During glacial periods, the banks are exposed, aragonite formation rates decrease, and the sedimentation rate of cores also reduce dramatically. Sediment aragonite percentages during highstands are typically 70-80%, with the remainder consisting of high-Mg and low-Mg calcite (and typically <1% detrital material). At depths in the core greater than 50 metres below sea floor (mbsf), corresponding to MIS 11, dolomite also starts to be seen (Malone, 2000), indicating the presence of diagenetic alteration of carbonate minerals at this depth.

Sedimentalogical changes associated with glacial-interglacial cycles are more extreme at this site than those seen on the leeward slopes of the Little Bahama Bank (LBB). Initial U/Th dating of Bahamas sediment was conducted on those LBB cores (Henderson et al., 2001; Slowey et al., 1996) which contain a smooth record of glacial to interglacial change. By contrast, Site 1008A demonstrates abrupt changes from glacials to interglacials, with glacial portions typically having sedimentation rates >10 times slower than those in the interglacials. This makes the GBB cores less appropriate for the dating of glacial periods or intermediate climate states. Sedimentation rates during interglacial periods, however, can be extremely rapid (up to 60 cm/kyr) so these GBB cores are well suited to assessing the duration and timing of sealevel highstands.

Foraminiferal oxygen isotope records recovered from ODP Leg 166 sediments show glacial-interglacial amplitudes of up to 2.5‰ (e.g. Robinson et al., 2002). This amplitude of change reflects the effects of sealevel, temperature, and the diagenesis of glacial-age sediment. Sealevel change causes a global change of ≈1‰ in $\delta^{18}O$ (Schrag et al., 2002). Glacial-interglacial temperature changes in the Caribbean are thought to be ≈2.5°C (Schmidt et al., 2004), and therefore could be responsible for an additional 0.6‰ of the observed $\delta^{18}O$ change.

The timing of this temperature change in the Caribbean relative to sealevel change is not completely clear, but probably does not differ by more than a couple of thousand years. Early work as part of the CLIMAP project (CLIMAP, 1984) suggested that temperature lagged sealevel in the North Atlantic, and led it in the

South, with the Caribbean at the hinge and not showing a significant phase difference. More recent Mg/Ca work (Schmidt et al., 2004) has again indicated a slight temperature lead for Terminations 1 and 2, 10^{o} south of the Bahamas in the Columbian basin. Paleothermometry on the Bahamas cores themselves is confounded by the presence of calcite overgrowths on foraminifera leading to high Mg/Ca during glacial periods (Rosenthal et al., 1997), and by very low alkenone concentrations. In the absence of such direct assessment, we assume that changes in sealevel and temperature at this site are within 2 kyr of one another.

The remainder of the $\delta^{18}O$ signal (≈0.9‰) is caused by addition of isotopically heavy oxygen to glacial sediments by seafloor diagenesis (Malone et al., 2001). This process is directly related to the rate of sedimentation, and therefore to sealevel and climate. When sealevel is high and the climate warm, greater areas of the Bahamas banks are exposed and aragonite productivity is high, leading to high sedimentation rates and minimal cementation. When sealevel is low, particularly during glacial conditions, sedimentation rates fall and cementation occurs leading to a higher $\delta^{18}O$ value. In summary, although control of $\delta^{18}O$ in these GBB sediments is complex, the major changes are synchronous with sealevel change.

Existing *G. sacculifer* and bulk-carbonate $\delta^{18}O$ stratigraphies for Hole 1008A (Robinson et al., 2002) were augmented by 28 new measurements on *G. sacculifer* (300-355 micron fraction) to increase resolution, particularly during the MIS 9 interval. For U/Th chronology, a total of 31 bulk sediment samples were selected spanning MIS 7 and MIS 9, with emphasis on the latter. These are in addition to 22 previously dated samples from 1008A which focused on the MIS 7 interval (Robinson et al., 2002).

3.0 Analytical techniques

Bulk sediment samples (≈1 g) were processed following techniques of Henderson et al. (2001) and analysed by multi-collector ICP mass spectrometry following Robinson et al. (2002). Briefly, samples were dissolved in 7.5N HNO_3, U and Th separated using a single anion-exchange column. Typical blanks for this procedure were $3x10^{-10}$ g ^{238}U, and $3x10^{-11}$ g ^{232}Th and are small compared to sample U and Th masses.

U and Th isotopes were analysed on a Nu Instrument MC-ICP-MS using a standard bracketing technique against the international U standard, CRM-145, and an in-house Th standard (ABC-2). In each case, small beams (^{234}U, ^{230}Th, ^{229}Th) were analysed in an ion-counter, while other beams (^{238}U, ^{235}U, ^{236}U, ^{232}Th) were analysed synchronously in Faraday collectors. This analytical technique yields precision in the critical $^{230}Th/^{229}Th$ isotope ratio typically better than 1‰, and final ^{230}Th concentrations (incorporating weighing errors and uncertainty in the spike concentration) of ≈2‰.

Calculated ages are corrected for the presence of initial ^{230}Th using the measured ^{232}Th concentrations. An initial $^{232}Th/^{230}Th$ atom ratio of 20,000 is assumed (equivalent to a $^{230}Th/^{232}Th$ activity ratio of 9.3). The initial Th is therefore assumed to be richer in ^{230}Th than typical crustal Th by approximately a factor of ten. This reflects the fact that most of the initial ^{230}Th in these sediments is derived from seawater, which gains ^{230}Th from decay of U. This leads to a $^{232}Th/^{230}Th$ atom ratio of seawater in the Bahamas at ≈400m water depth of ≈15,000 (Robinson et al., 2004). An uncertainty to the correction for initial ^{230}Th of ±50% is assumed and combined with analytical error to give the final uncertainty on the corrected ages.

4.0 Results

4.1 Stratigraphy

Interglacial portions of the $\delta^{18}O$ record are significantly expanded relative to glacial portions (Fig. 1). Interglacial portions are characterized by *G. sacculifer* values of ≈-1‰, and bulk sediment values of ≈0‰, while glacial portions are ≈2‰ heavier. The upper 12 m of the record is at relatively low resolution but clearly shows the Holocene (0-6 mbsf) and MIS 5 (7-10 mbsf).

MIS 7 falls in the interval 12 to 20 mbsf where the $\delta^{18}O$ curve features four pronounced lows in both the *G. sacculifer* and bulk carbonate records (seen as highs in the inverted scale of Figs 1 and 2). The shallowest of these (≈12.5 mbsf) features less extreme $\delta^{18}O$ values (particularly in the bulk carbonate record) and is probably equivalent to MIS 6.5 (Martinson et al., 1987). The other three lows have approximately equal $\delta^{18}O$, at values typical of full interglacial conditions for this site. These are assumed to represent the highstand associated with MIS 7 which therefore covers the depth interval 12.8 to 20.0 mbsf. Termination 3, at the start of MIS 7, is difficult to identify on the basis of the *G. sacculifer* record alone, but the first change to low foraminiferal $\delta^{18}O$ values corresponds to the clear change in bulk sediment values which makes selection of this event at 20.0 m a reasonable choice (Fig. 2).

A similar pattern is seen for the MIS 9 section of the core. Low $\delta^{18}O$ values from 21-23 mbsf (seen only in the *G. sacculifer* record) are probably equivalent to MIS 8.5 (Martinson et al., 1987). Low $\delta^{18}O$ values seen in both records from 22.6 to 27.2 mbsf are assumed to represent the highstand corresponding to MIS 9. And Termination 4 is defined by an abrupt change in both the *G. sacculifer* and bulk-sediment records at ≈27.2 mbsf.

4.2 Chronology

Compared to average continental crust, analysed samples have high U concentrations (4.1-14.3 ppm) and low Th concentrations (0.06 – 0.28 ppm) (Table 1). They all exhibit significant ^{230}Th-^{234}U-^{238}U disequilibrium with calculated raw ages (i.e. not corrected for initial ^{230}Th) ranging from 147 to 406 ka (Table 1). Corrections for initial ^{230}Th range from 1 to 25 kyr, with a median value of 4 kyr, reflecting the generally low Th concentrations of these samples. Five samples have age corrections >10 kyr (corresponding to ^{232}Th concentrations of >200 ppb). Ages for these samples are not useful to constrain the timing of climate change due to the large correction and are discarded. This follows the approach used in previous dating studies of Bahamas sediments (Henderson et al., 2001; Robinson et al., 2002).

δ^{234}U(T) (calculated with corrected ages) range from 105 to 209, with an average identical to the modern seawater value (146; Robinson et al., 2004). This average value suggests overall closure of the sediment mass to exchange of U with other reservoirs, but the range of values observed indicates the presence of some movement of U within the sediment. Samples with values outside the range 135 to 155 are assumed to have suffered sufficient alteration to their U/Th system that resulting ages are unreliable. This range is identical to that used in Robinson et al. (2002) and allows for uncertainty in the seawater value of δ^{234}U through time (Henderson 2002). No attempt to correct ages for recoil mobility of nuclides using measured δ^{234}U has been conducted (e.g. Thompson et al. 2003). Bulk sediment δ^{234}U(T) which differ significantly from seawater indicate transport of ^{234}U over sufficient distances that insoluble ^{230}Th is unlikely to accompany it, so such open-system age models are unlikely to be robust in this setting.

Based on their ^{232}Th concentrations and δ^{234}U(T) values, 11 of the 31 ages are assumed to provide reliable ages. These range from 190 - 363 ka and are illustrated, together with 12 reliable ages from Robinson et al. (2002), in Fig. 1. Two of these ages are significantly out of age sequence and cannot be correct (22.30 mbsf and 28.49 mbsf). These two samples are from sections of the core with the lowest measured aragonite content of any dated samples (≈50% and ≈40% respectively) (Malone, 2000). Aragonite is the mineral in the sediment with a high U/Th ratio which provides most of the age information (Henderson et al., 2001). It appears that when the fraction of such aragonite in the core is as low as 50%, accurate age information cannot be derived from bulk sediments. These ages are not considered further, and care must be taken in interpreting other ages where the aragonite content may be low. The remaining ages define the timing of the MIS 7 and MIS 9 portions of the δ^{18}O record from Site 1008A.

5.0 Discussion

Most models to explain glacial-interglacial sealevel cycles assume that they are linked to changes in summer insolation at mid-latitudes in the Northern hemisphere. This assumption, following Milankovitch (1930), is inherent to tuned timescales of climate change, including the widely used SPECMAP timescale (Imbrie et al., 1984; Martinson et al., 1987). Several of these tuned climate records are shown in Fig. 3, including a new tuned benthic δ^{18}O stack (Lisiecki and Raymo, 2005). Direct U-Th chronology for sealevel changes in this study, and from corals (Thompson and Goldstein, 2005), can be compared to these tuned timescales to test the assumption that glacial-interglacial cycles are linked to mid-latitude summer insolation. We make such comparison here, but first consider the quality of the age information derived from Hole 1008A.

5.1 Age reversals and the quality of the U/Th data

Ages are expected to increase down-core. In general, ages that pass the isotopic rejection criteria follow this pattern (Fig. 1). There are, however, several age reversals, particularly within MIS 7 (Fig. 2). These age reversals must be caused either by incomplete closure of the U-Th system, or by stratigraphic disturbance.

The failure of U-Th chronology in those samples that do not pass the rejection criteria indicates the possibility for disturbance of the U-Th system in these Bahamas sediments. $\delta^{234}U(T)$ values close to seawater are, however, not easy to explain if samples have been disturbed. Movement of U within the core is expected to perturb both the measured $\delta^{234}U$ and the age, both of which significantly alter the $\delta^{234}U(T)$. It is also noteworthy that, where replicate samples are both reliable based on their $\delta^{234}U(T)$ values, their ages are within error of one another (e.g. 2 samples at 19.48m, and 3 at 25.9m).

The stratigraphy in this off-shelf setting is also not perfect. Sediment transport occurs rapidly from the Bahamas banks to the slopes at present and is driven by tidal processes. Storage of sediment on the banks, or higher up the slopes, cannot be ruled out, however, and these sediments might be transported down-slope to place older sediment above newer in slope cores. Such down-slope transport would obviously perturb the stratigraphy. It would not, however, alter the geochemistry so that $\delta^{18}O$ and U/Th ages would remain intact and coupled together. U/Th ages on sediment with a $\delta^{18}O$ lower than zero therefore still imply highstand conditions at that time even in the presence of stratigraphic disturbance.

Overall, the presence of age-reversals does point to an imperfect system, either geochemical or stratigraphic, but this record is the best independently dated marine stratigraphy available. It is interesting to contrast these Bahamas ages with those from coral terraces. Dating of corals is generally performed on individual samples distributed laterally so that no relative stratigraphic information is provided. Coral ages are therefore not normally subject to the stratigraphic test that these Bahamas sediments have undergone. It is also notable that, of the many MIS7 corals that have been dated, only ≈5 give ages with a $\delta^{234}U(T)$ close to modern seawater (Robinson et al. 2003). This contrast with the 12 ages with good $\delta^{234}U(T)$ that are presented here on Bahamas sediments. Although the Bahamas sediments are not perfect, the fact that they come with some stratigraphic control, and have experienced submarine preservation, do give them certain advantages over subaerially exposed corals. Corals, of course, have their own advantages, such as their provision of a more direct assessment of sealevel. These two archives therefore provide complementary information in the quest to define a radiometric chronology for Pleistocene sealevel.

5.2 An additional sealevel highstand post-dating MIS 7

The early portions of MIS 7 have been discussed previously (Robinson et al., 2002) and new ages reported here do not change the interpretation. The start of MIS 7 at Site 1008A agrees well with that expected from records tuned to 65°N summer insolation. The coral record also agrees with SPECMAP (Thompson and Goldstein, 2005) and suggests that Termination 3 occurred at ≈245 ka (Fig. 4).

The more interesting feature occurs at the end of MIS 7. Sediment with low $\delta^{18}O$ (in both *G. sacculifer* and bulk records) extends from 18 to 14 mbsf (Figs 1 and 2). Based on typical highstand sedimentation rates for this core, and on the duration of this interval from U/Th ages, this is very likely to represent MIS 7.3 to MIS 7.1. A new U/Th age of 190 ka from the $\delta^{18}O$ peak immediately following this interval (13.80m; Table 1) supports this interpretation and is in agreement with previous U/Th chronology for this period based on both corals (Thompson and Goldstein, 2005) and speleothems (Bard et al., 2002). Together, these archives provide convincing evidence that MIS 7.1 ends at 190 ka, again in good agreement with SPECMAP. Immediately above this MIS 7.1 section at Site 1008A, however, lies an additional $\delta^{18}O$ low (13.7 to 13.0 mbsf). A date from this low $\delta^{18}O$ interval, and two ages from slightly greater depth, give ages that are significantly younger than 190 ka. An age reversal at this depth (Fig. 2) make interpretation somewhat uncertain, but these ages suggest the presence of highstand conditions lasting some 10 ka after the expected SPECMAP age for the end of MIS 7.

A similar sealevel event was identified in the coral record of Thompson and Goldstein (2005) which shows a discrete sealevel highstand from ≈187 to 180 ka (Fig. 4). The chronology of both the Site 1008A and coral records are based on few U/Th ages, but the close agreement in the duration and age of this sealevel event in two completely independent records suggests that there is a sealevel highstand which post-dates the normally recognized end of MIS 7. This event is not MIS 6.5, which is clearly seen as an additional feature in both the Bahamas sediment and Barbados coral records (Fig. 4).

Some constraints can be placed on the duration and amplitude of this sealevel event. Typical highstand sedimentation rates for the Site 1008A location would suggest a duration of 5-10 kyr, and the coral record suggests a duration of ≈6 kyr. The duration is therefore long compared to typical millennial climate events as seen, for instance, in the Greenland ice cores during the last glacial (Grootes et al.,

1993) and is closer to the duration of highstands associated with MIS 5.5 or MIS 5.3. In the Barbados coral record, sealevel reaches –13 m during this event, but this is in conflict with speleothems from Italy which continue to grow during this period at an altitude of –18 m. This suggests that the tectonic reconstruction at either Barbados and/or Italy is not completely accurate, perhaps due to glacio-eustatic effects (e.g. Potter and Lambeck 2003) but overall suggests that sealevel must have been close to –15 m during this period.

The presence of a sealevel highstand at ≈180 ka represents a challenge to the idea that Pleistocene climate is driven by summer insolation at 65°N. Sealevel is increasing (and therefore ice is melting) when 65°N summer insolation is at one of its lowest points of the last 400 kyr. The presence of sub-orbital oscillations of sealevel has been recognized before (Esat et al., 1999; Thompson and Goldstein, 2005). But this highstand, occurring in the period normally thought of as a portion of the MIS 6 glacial, joins Termination 2 (Gallup et al., 2002; Henderson and Slowey, 2000) in providing a challenge to orbital models of climate change, and in providing clues about the mechanisms for Pleistocene climate change.

5.3 Implications of the additional highstand for orbital tuning of δ^{18}O records

The additional highstand at 185 ka is not seen in orbitally-tuned δ^{18}O records (Imbrie et al., 1984; Lisiecki and Raymo, 2005) despite the significant sealevel component expected in these records. The most recent tuned δ^{18}O stack (Lisiecki and Raymo, 2005) collects more than 50 benthic records from around the globe and stacks them to provide a combined record of sealevel and deep-ocean temperature for the last 5.3 Myr (Fig. 3, final curve). This is the most comprehensive benthic stack yet published, and indicates δ^{18}O of 4.4‰ at ≈185 ka, 1.2‰ heavier than the modern value in the same stack, and therefore requiring sealevel significantly lower than today at 185 ka. There are only two possible explanations for the high $\delta^{18}O$ values at 185 ka in this and other tuned records. Either deep-ocean temperatures were very cold at this time, thus cancelling the $\delta^{18}O$ signal from high sealevel, or the timescales of these tuned records are incorrect.

It is very unlikely that deep-ocean temperatures can be cold enough to cancel the $\delta^{18}O$ signal from a –15m highstand at ≈185 ka. The Holocene to Last Glacial Maximum (LGM) change in $\delta^{18}O$ is 1.8‰ in the Lisieki and Raymo stack. This is thought to represent 1‰ (Schrag et al., 2002) due to a sealevel change of 125 m, plus about 3°C of deep-ocean cooling at the LGM. This cooling takes bottom waters throughout the ocean to very close to their freezing point (Adkins et al., 2002). By comparison, $\delta^{18}O$ at ≈185 ka is 1.2‰ heavier than modern values. If sealevel is -15 m at this time (as constrained by the coral, speleothem, and Bahamas records) only ≈0.1‰ of this difference is explained by sealevel. The remaining 1.1‰ requires deep-ocean temperatures to be more than 4°C colder than today, and therefore colder than during the close-to-freezing LGM. Deep-ocean cooling therefore cannot explain the mismatch between tuned benthic $\delta^{18}O$ stacks and the sealevel record. This conclusion is reinforced by the more qualitative but intuitive argument that it is difficult to imagine ice-sheets shrinking to cause a sealevel highstand at a time of extreme cold.

It therefore appears that the timescale of tuned records is incorrect at the end of MIS 7. The large change in $\delta^{18}O$, normally tuned to the decrease in northern hemisphere summer insolation at 190 ka, actually occurs at ≈180 ka. Tuned records are therefore in error by about 10 kyr at this point. This is significantly larger than the normally quoted errors on such timescales (e.g. 5 ka, Martinson et al., 1987) and reflects a failure of the basic assumption on which such tuning relies. Having accepted that the tuned timescale is not robust, benthic $\delta^{18}O$ records do, in fact, contain evidence for this end-MIS 7 highstand. In the Lisieki and Raymo stack, for instance, there is a pronounced shoulder at the end of MIS 7 (arrow “a”, Figure 2), with $\delta^{18}O$ values only ≈0.2‰ heavier than at MIS 7.1 (Lisiecki and Raymo, 2005). In individual records, a discrete peak is sometimes seen at this point. The well-known V19-30 record, for instance, contains an event of similar duration and slightly lower amplitude than those of MIS 7.3 and MIS 7.1 (Shackleton and Pisias, 1985).

Adjusting timescales for marine cores so that the end of MIS 7 is at 180 ka rather than 190 ka does not generate unrealistic changes in sedimentation rate. If such a timescale adjustment implied large temporal changes in sedimentation rate it might argue against its validity. Recalculating sedimentation rates in the Lisiecki and Raymo stack, assuming that the record at 190 ka is 10 kyr too old, leads to global sedimentation rates during MIS 6 that are 23% higher than the average for the last million years. This compares to sedimentation rates during the last glacial (MIS2-4) that are ≈32% higher than average and is therefore not unreasonable.

5.4 Possible mechanisms for this additional sealevel highstand

The additional late-MIS 7 sealevel highstand cannot be explained by ice-sheet melting in response to northern-hemisphere summer insolation. Instead, it must be due to a climate response to insolation at another season/latitude, or to a natural oscillation in the climate system. Sub-orbital sealevel changes during the last glacial (Siddall et al., 2003; Yokoyama et al., 2001) correlate with changes in North Atlantic and Greenland temperatures and are thought to be caused by coupling between ice sheets and ocean circulation via fresh-water supply and heat transport. Such coupling is less likely at the end of MIS 7 when northern-hemisphere ice sheets are small (as is the case, in general, for sub-orbital sealevel changes during highstands (Thompson and Goldstein, 2005)).

Two insolation mechanisms can be suggested, but neither is without problems. The first is to suggest that the very low northern-hemisphere summer insolation at this time (Fig. 3; top curve) keeps conditions in the North Atlantic region cold enough that moisture transport is suppressed (by lower rates of evaporation, or by the presence of sea-ice) and ice cannot grow. Problems with this mechanism are that the ice-sheets actually seem to shrink at 187 ka, requiring active melting rather than just a cessation in growth, and there are other periods with equally low northern-hemisphere summer insolation when ice sheets do not decrease (e.g. 230 ka). The second possible insolation mechanism is that summer insolation in the southern hemisphere reaches a peak at 187 ka, just as the additional sealevel rise is observed. This timing coincidence is appealing and might suggest that the sealevel rise is caused by a melting of southern-hemisphere ice. Southern hemisphere ice is generally thought to contribute only ≈25m to glacial-interglacial sealevel changes, but recent evidence is suggesting that it contributes a significant fraction (≈half) to sealevel changes associated with millennial events (Rohling et al., 2004). A larger role for southern hemisphere ice might therefore be invoked for this late MIS 7 event. Again, though, a problem with this mechanism is that other, equally large peaks in southern-hemisphere insolation are not accompanied by observable sealevel highstands.

Atmospheric CO_2 levels may also play a role in this sealevel event. The Vostok CO_2 record at the end of MIS 7 shows a series of peaks which are not easy to relate directly to events in the marine $\delta^{18}O$ record. Two peaks of ≈ 240 ppmV appear to correspond to MIS 7.3 and MIS 7.1 and are followed by a peak of 231 ppmV (Petit et al. 1997) (arrow “b”, Figure 2). This additional peak occurs at 191 ka according to the Vostok age scale but is 3 kyr younger in the Shackleton timescale tuned using atmospheric $\delta^{18}O$. There is latitude to move this event to still younger ages because of the breadth of $\delta^{18}O_{atm}$ peaks in this interval. The additional sealevel peak might, therefore, be caused by greenhouse warming induced by a change in the carbon cycle. This explanation alters the question rather than answering it. The cause of a change in atmospheric pCO_2 at this time is not clear.

5.5 The timing of highstands associated with MIS 9

Termination 4, at the start of MIS 9, is reasonably clear in the $\delta^{18}O$ records and occurs at ≈26.2 mbsf. Four dates immediately above the Termination are all older than the SPECMAP age for this event of 338 ka (Figure 4). The oldest of these dates (363 ka) predate the SPECMAP age very significantly but are from an interval where the aragonite content may be low (measurements above and below these ages are 78% and 24% aragonite respectively). The other three ages are replicates at a single depth where sediment has a measured aragonite fraction of 78%, and the $\delta^{18}O$ of both *G. sacculifer* and bulk sediment indicate full interglacial conditions. These three replicates all pass U-Th criteria and are within error of one another, averaging 343 ka (±4 ka, 2σ). Although the very old date can be called into question based on low aragonite fraction, it is more difficult to question the well-constrained dates at ≈343 ka. These ages therefore suggest that full interglacial conditions were achieved ≈5 kyr (or more) before the age for Termination 4 in the SPECMAP chronology (338 ka). Deglaciation typically takes ≈6 kyr (e.g. Broecker and Henderson 1999) so the expectation is that full-interglacial conditions should lag the midpoint of deglaciation by ≈3 kyr. A 5 kyr lead therefore represents an 8 kyr mismatch from that expected if climate if forced by the mid-latitude northern hemisphere. This is significantly greater than the combined error on the three ages for this event so there is some confidence in suggesting that Termination 4 is earlier than expected from traditional Milankovitch forcing.

There is very little previous data for MIS 9 with which to compare this new Bahamas data. A single coral date from Huon Peninsula (Galewsky et al., 1996) is also earlier than the SPECMAP age for Termination 4, but is within error (Fig. 5). High-precision dates from Henderson Island (Stirling et al., 2001) do not provide any evidence for an early start to MIS 9, but coral preservation is generally biased to the end of highstand periods because of the possibility of marine erosion of early highstand material.

In the absence of supporting evidence from other records it is premature to be confident about the age for the start of MIS 9 based on the limited data presented here. Nevertheless, this data does suggest that MIS 9 may start significantly earlier than predicted by tuned records such as SPECMAP. This discrepancy would require, as for the late end of MIS 7, revision of tuned chronologies, and would provide new information about the mechanisms driving glacial-interglacial climate (e.g. Alley et al. 2001). For instance, a role for the southern hemisphere might be indicated by the correlation of the early start to MIS 9 with a peak in southern-hemisphere insolation at 345 ka.

6. Conclusions

Bulk-sediment U-Th dates have been used to date MIS 7 and 9 as identified in a $\delta^{18}O$ stratigraphy from ODP Site 1008A on the leeward slope of the Great Bahamas Bank. After screening dates based on their initial U and Th isotope ratios, 'reliable' dates allow the testing of tuned chronologies based on the assumption that climate is driven by changes in northern-hemisphere summer insolation. Based on both $\delta^{18}O$ and U-Th data, we identify a sealevel highstand at the end of MIS 7, previously seen in coral data, which lasts from ≈187 to 180 ka and therefore post-dates the expected end of MIS 7 by some 10 kyr. U-Th dates at the beginning of MIS 9 also suggest that tuned chronologies may be wrong. Based on three replicate ages at a single sediment depth, full interglacial conditions of MIS 9 appear to have been established by 343 ka, ≈8 kyr earlier than expected. These results indicate that the phasing between climate change and northern hemisphere insolation is not constant, so chronologies based on this assumption (e.g. SPECMAP) may have systematic errors. These results also provide new information about the mechanisms linking insolation changes with changes in climate during the Pleistocene.

Acknowledgements

Nick Shackleton first put one of us (GMH) in contact with Niall Slowey, thus starting the use of U-Th techniques to date Bahamas sediment. The research presented in this paper grew directly from that contact and we thank Nick for initiating this research, and for valuable discussion as it has developed. GMH also acknowledges, with great gratitude, the help that Nick provided in keeping an otherwise thoroughly geochemical Ph.D. thesis on the climate straight-and-narrow, and for his support and inspiration since then. He will be greatly missed.

We thank Mark Siddall, Bill Thompson and an anonymous referee for their constructive comments on an earlier draft of the manuscript, and the Gary Comer Abrupt Climate Change Fellowship for funding that enabled completion of this research.

Figure 1: $\delta^{18}O$ and aragonite records for ODP Hole 1008A with bulk sediment U/Th ages. The upper curve (blue) is $\delta^{18}O$ for *G. sacculifer* (300-355micron fraction); the middle curve (green) is $\delta^{18}O$ for bulk carbonate; and the lower curve (orange) is aragonite percentage (from Malone, 2000). U/Th ages are corrected ages shown in bold in Table 1, together with reliable ages from Robinson et al. (2002). Two of these ages are plotted in grey and are clearly out of sequence. These are discussed in the text, and not considered as part of the age model for this core. Symbol size for the U/Th ages approximates the final 2σ uncertainty (Table 1). Vertical grey bands define the low $\delta^{18}O$ values corresponding to MIS 9 and MIS 7.

Figure 2: A blow-up of the MIS 7 section of Figure 1. Bulk-sediment U/Th ages are shown with their 2σ uncertainties. Note the general increase of age with depth, but the presence of two age reversals. The upper curve (blue) is $\delta^{18}O$ for *G. sacculifer* (300-355micron fraction) and the lower curve (green) is $\delta^{18}O$ for bulk carbonate. Marine isotope substages are selected based on these curves and shown by labelled vertical grey bands. Note the presence of an additional highstand, labelled as "??".

Figure 3: A comparison of age constraints for the timing of MIS9 and MIS 7 from Bahamas sediments with previous sealevel reconstructions and some other relevant data. The vertical grey bands represent the timing for MIS 9 and MIS 7 based on U/Th dating of bulk Bahamas sediment (see other Figures). These are compared with, from the top: 65°N insolation; open-system coral ages from Barbados (Thompson and Goldstein, 2005); Vostok CO_2 (Petit et al., 1999); sealevel curve based on Red-Sea $\delta^{18}O$ (Siddall et al., 2003); a "sealevel" curve based on correcting $\delta^{18}O$ for temperature using Mg/Ca (Lea et al., 2000); a "sealevel" curve based on comparison

of Vostok and ocean $\delta^{18}O$ records (Shackleton, 2000); and a stack of 57 benthic $\delta^{18}O$ records (Lisiecki and Raymo, 2005). Note that, while the grey bands and the upper two records are plotted on an absolute timescale. Other records are plotted on their published timescales which are based on some form of tuning. In the case of the Vostok CO_2 record, this tuning is to only two age control points. Other records are tuned to some form of mid-latitude northern-hemisphere summer insolation curve. Arrow "a" marks the shoulder at the end of MIS 7, and "b" the extra peak in the CO_2 curve, as discussed in the text.

Figure 4: All available absolute constraints on sealevel during MIS 7. Blue lines show intervals of speleothem growth at the Bahamas (Smart and Richards, 1992) and Italy (Bard et al., 2002) and are an upper bound for sealevel (assuming that there are no unrecognized hiatuses in the record). The red line is a reconstruction of sealevel based on open-system coral ages from Barbados which are shown as red points (Thompson and Goldstein, 2005). All coral data has been shifted –7 m compared to the published record. This brings coral and speleothem sealevel into closer agreement and reflects probable small errors in the reconstruction of tectonic movement at the various sites. The new Bahamas $\delta^{18}O$ data are plotted below these literature data in blue (*G. sacculifer*) and green (bulk sediment). They are placed on an age scale assuming a constant sedimentation rate between Termination 3 (at 243 ka) and the end of MIS 7 (at 180 ka). Ages for these tied points are based on bulk sediment ages as shown in Figure 2 and discussed in the text. The lowermost curve shows the SPECMAP curve for comparison (Imbrie et al. 1984). Note the presence of an additional highstand in both the coral and Bahamas data – highlighted by the vertical grey band.

Figure 5: All available absolute constraints on sealevel during MIS 9. A single blue triangle is a speleothem age (Lundberg and Ford, 1994). Coral ages are shown in red from Henderson Island (Stirling et al., 2001); Huon (Galewsky et al., 1996); and Muruoa (Camoin et al., 2001). The new Bahamas $\delta^{18}O$ data are plotted below these literature data in blue (*G. sacculifer*) and green (bulk sediment). They are placed on an age scale assuming a constant sedimentation rate between the start of full interglacial conditions (at 343 ka – based on bulk sediment ages as discussed in the text) and the end of MIS 9 (at 204 ka – based on the SPECMAP age scale). The lowermost curve shows the SPECMAP curve for comparison (Imbrie et al. 1984). Note that the new Bahamas chronology suggests that MIS 9 started earlier than expected from northern hemisphere mid-latitude orbital forcing (as highlighted by the vertical grey band).

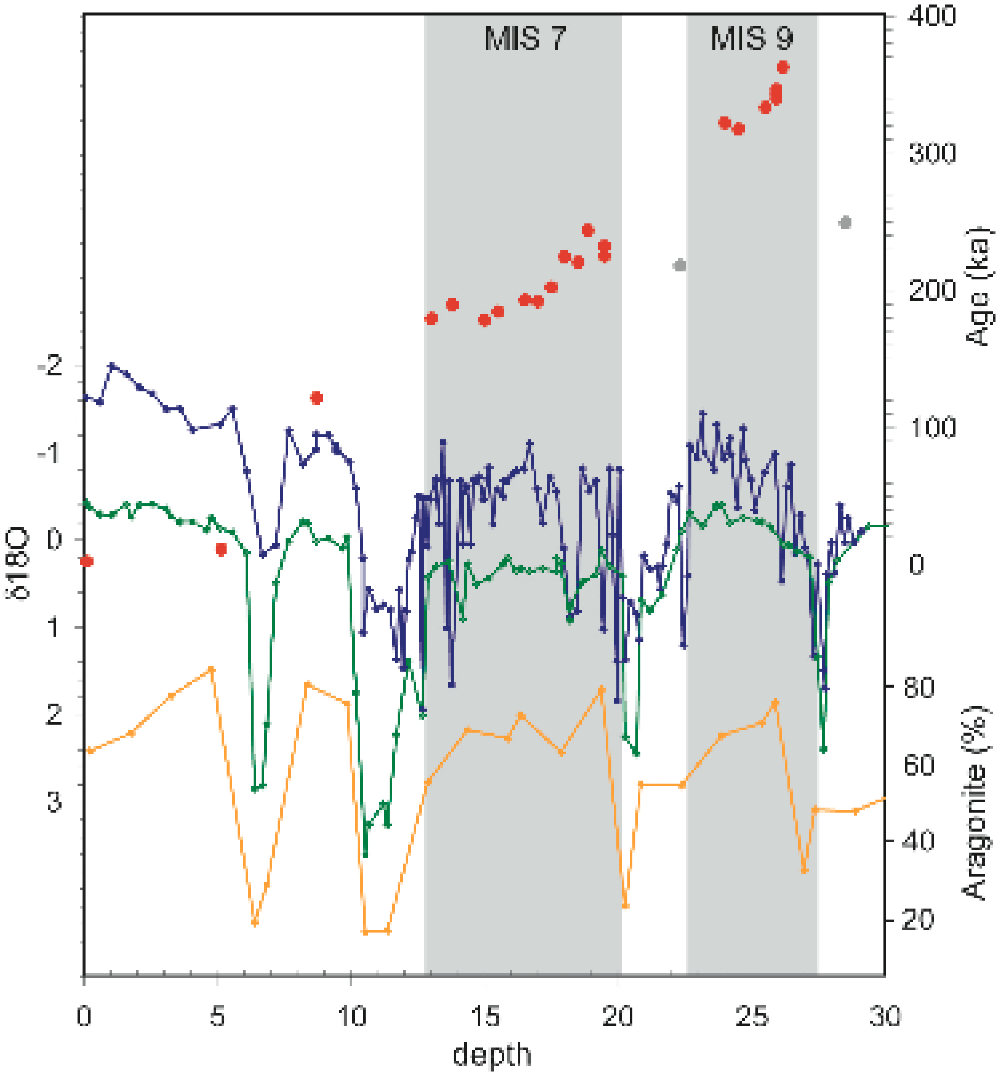
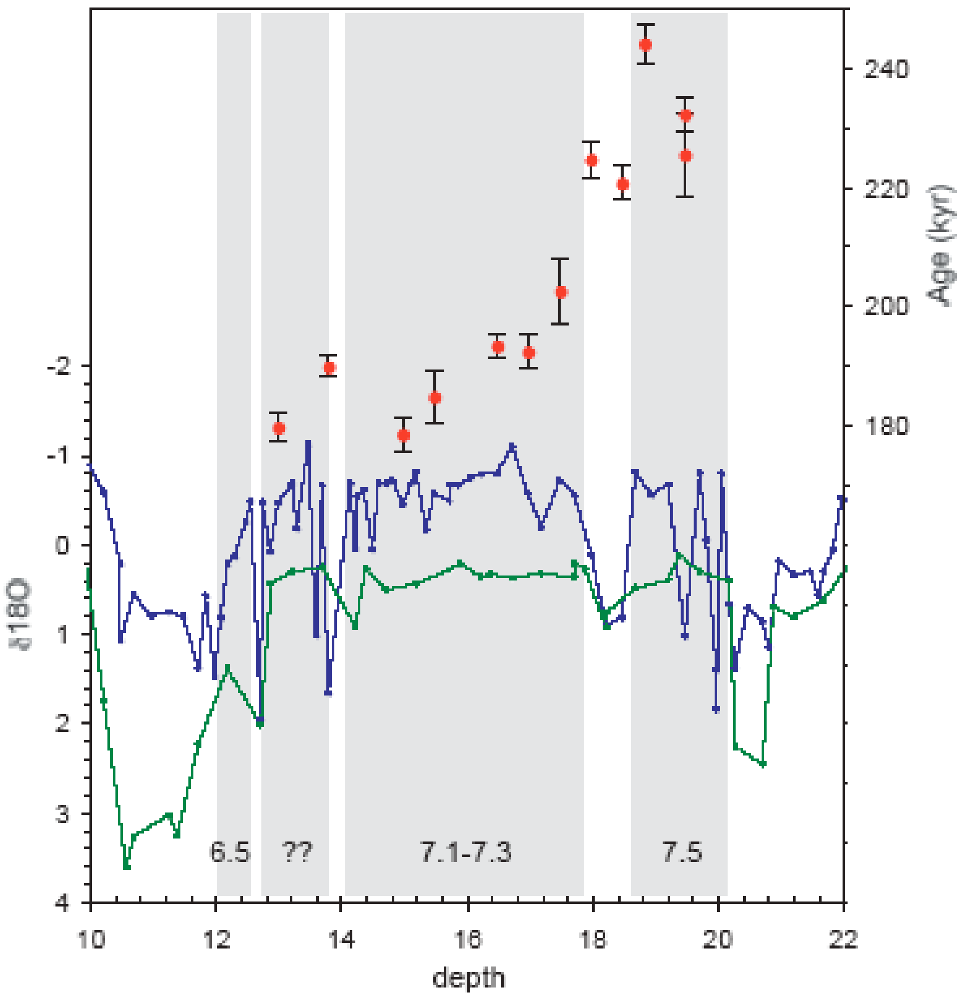

MIS 7
MIS 9
δ18O
-2
-1
0
1
2
3
Age (ka)
400
300
200
100
0
Aragonite (%)
80
60
40
20
0
5
10
15
20
25
30
depth

Figure 1



240
220
200
180
Age (kyr)
-2
-1
0
1
2
3
4
δ18O
6.5
??
7.1-7.3
7.5
10
12
14
16
18
20
22
depth

Figure 2

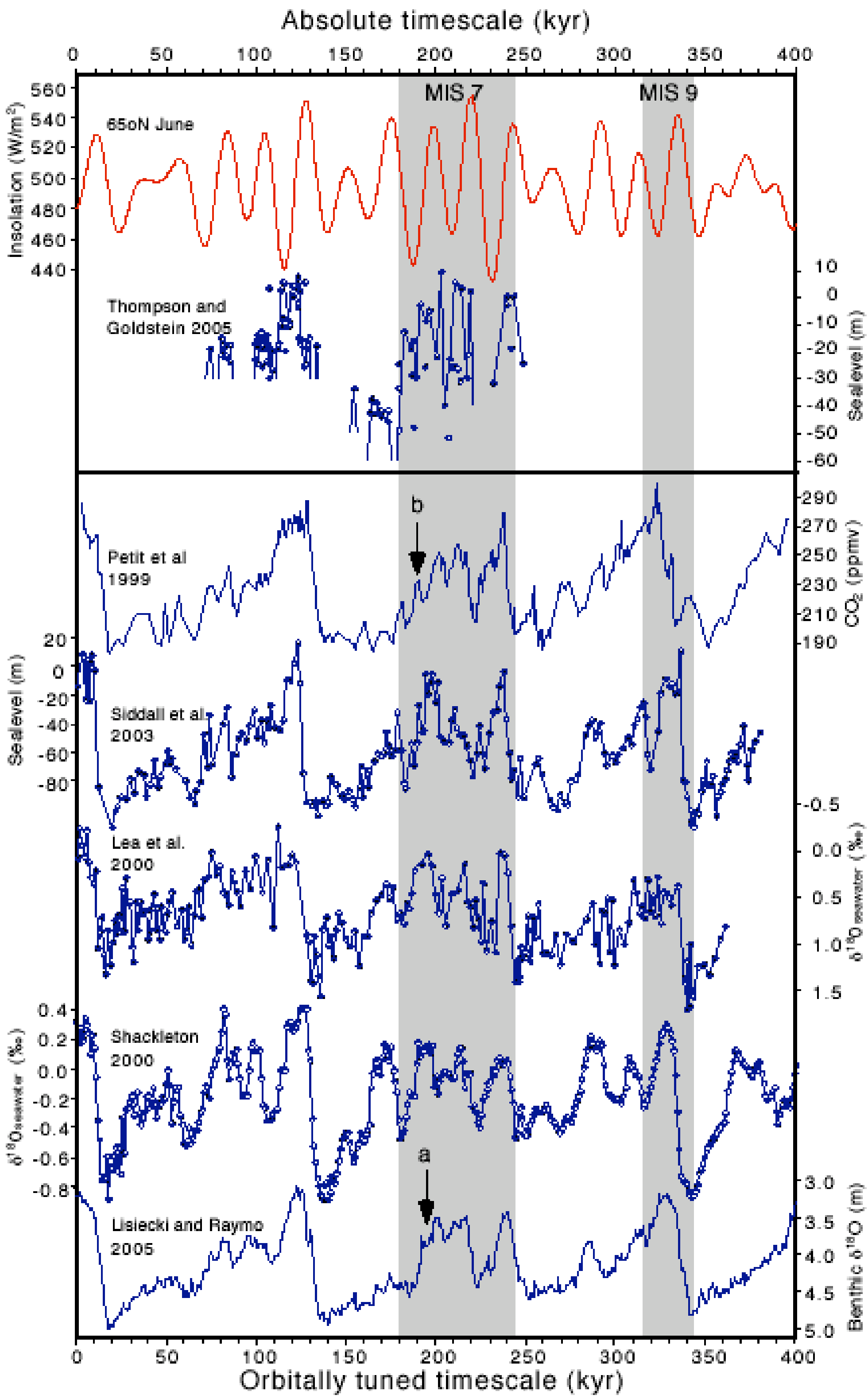
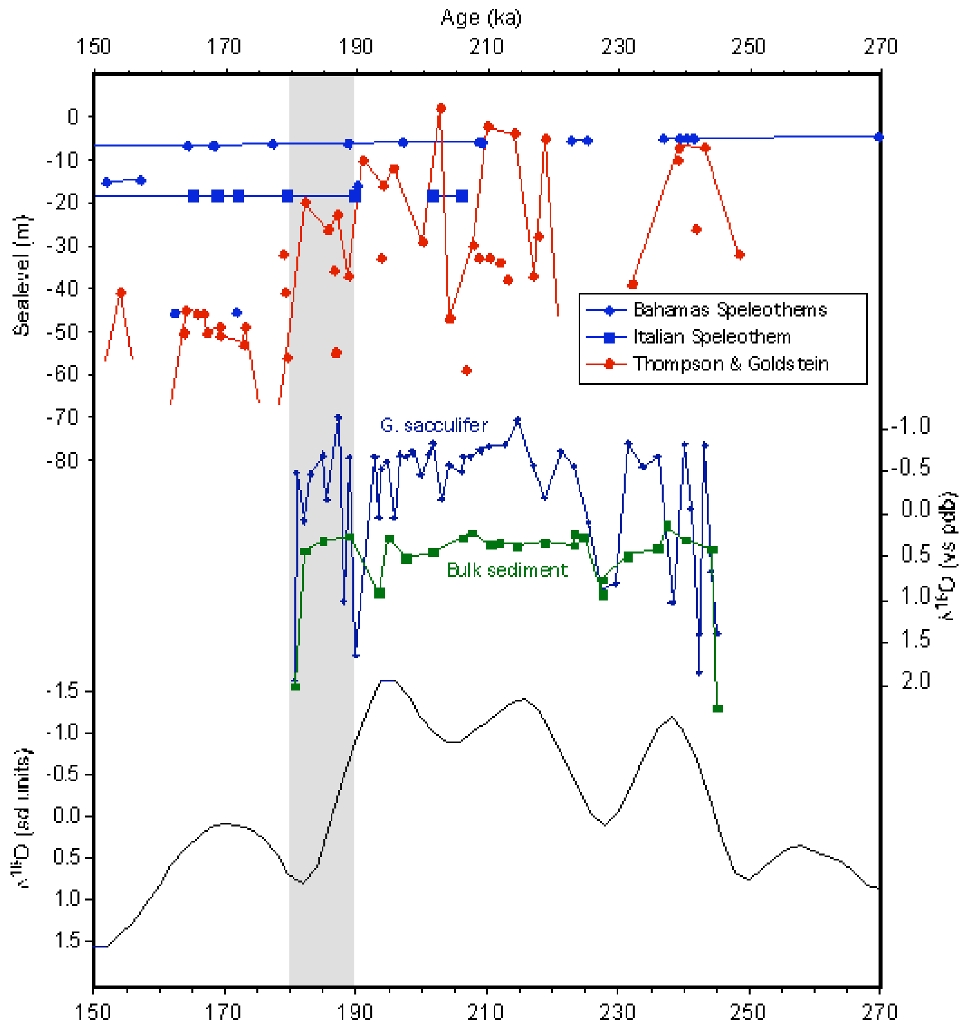
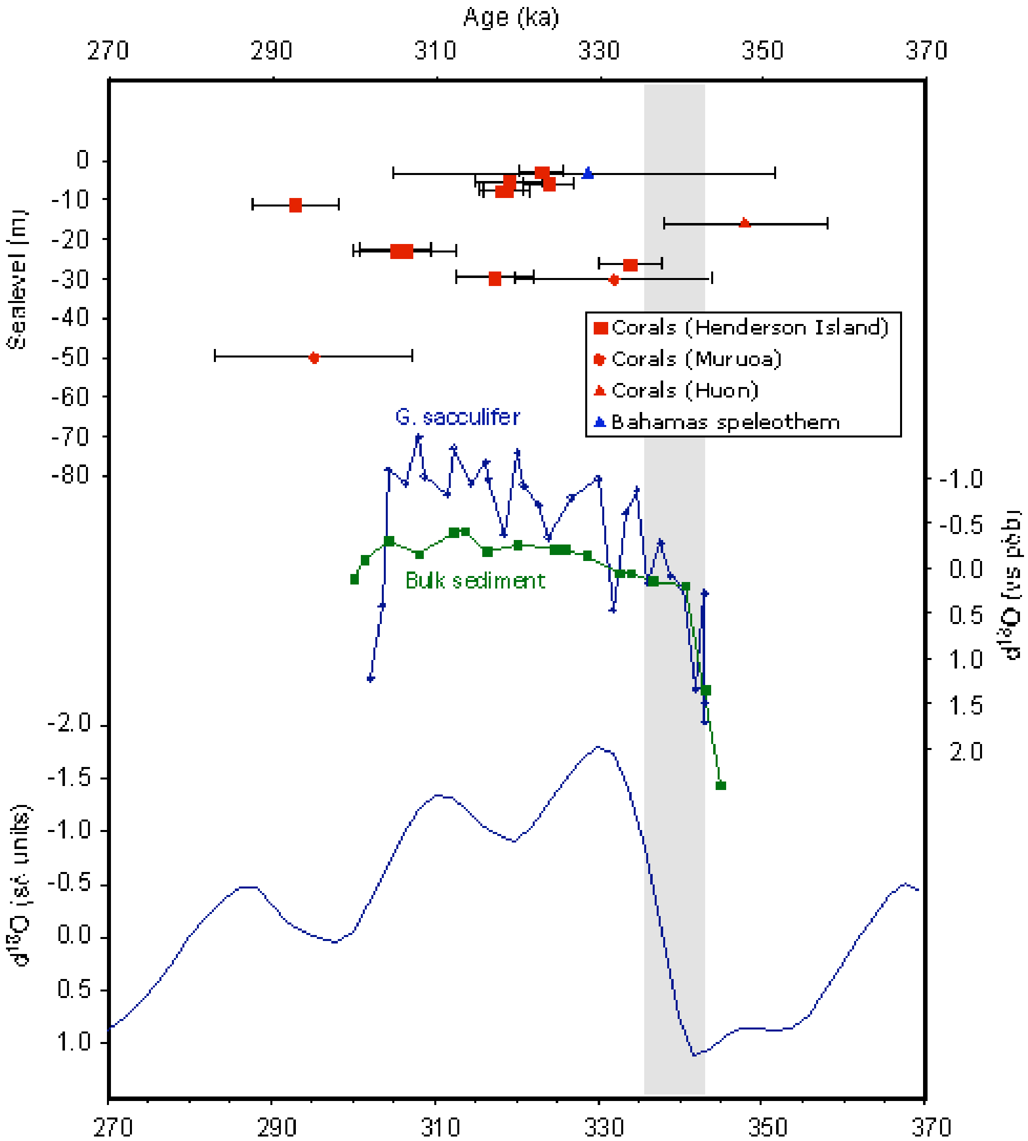

Absolute timescale (kyr)
MIS 7
MIS 9
65oN June
Insolation (W/m2)
Thompson and Goldstein 2005
Sealevel (m)
b
Petit et al 1999
CO2 (ppmv)
Siddall et al 2003
Sealevel (m)
Lea et al. 2000
δ18Oseawater (‰)
Shackleton 2000
δ18Oseawater (‰)
a
Lisiecki and Raymo 2005
Benthic δ18O (m)
Orbitally tuned timescale (kyr)

Figure 3



Age (ka)
Sealevel (m)
Bahamas Speleothems
Italian Speleothem
Thompson & Goldstein
G. sacculifer
Bulk sediment

Figure 4



Age (ka)
Sealevel (m)
Corals (Henderson Island)
Corals (Muruoa)
Corals (Huon)
Bahamas speleothem
G. sacculifer
Bulk sediment
d13O (vs pdb)
d18O (std units)

Figure 5

Depth (mbsf)	Core	Sect	top (cm)	^{238}U (ppm)	^{232}Th (ppb)	^{230}Th (ppt)	$\delta^{234}U(0)$ ‰	($^{230}Th/^{238}U$)	Raw age (ka)	Corr. age (ka)	$\delta^{234}U(T)$ ‰
12.30	2	4	120	7.453 ±0.006	272.3 ±0.8	100.2 ±0.3	92 ±1	0.821 ±0.003	147 ±1	*137 ±5*	*136 ±2*
12.56	2	4	146	4.981 ±0.004	280.3 ±0.8	70.1 ±0.2	100 ±1	0.860 ±0.003	159 ±1	*144 ±7*	*151 ±3*
12.75	2	5	15	7.171 ±0.006	206.9 ±0.6	116.4 ±0.4	88 ±1	0.992 ±0.003	242 ±3	*235 ±5*	*171 ±2*
13.80	2	5	120	8.192 ±0.007	77.1 ±0.2	122.2 ±0.4	80 ±1	0.912 ±0.003	193 ±2	**190 ±2**	**138 ±2**
18.84	3	2	124	5.464 ±0.005	73.4 ±0.2	87.8 ±0.3	74 ±1	0.982 ±0.003	248 ±3	**244 ±3**	**149 ±1**
20.07	3	3	97	3.499 ±0.003	222.9 ±0.7	59.3 ±0.2	63 ±2	1.035 ±0.003	337 ±8	*320 ±8*	*163 ±1*
20.98	3	4	38	8.584 ±0.010	166.9 ±0.5	127.7 ±0.2	65 ±4	0.909 ±0.002	201 ±3	*196 ±4*	*112 ±6*
21.61	3	4	101	14.316 ±0.017	181.9 ±0.5	195.0 ±0.3	88 ±4	0.832 ±0.002	152 ±1	*149 ±2*	*134 ±6*
22.30	3	5	20	7.243 ±0.006	160.9 ±0.5	113.3 ±0.4	75 ±1	0.956 ±0.003	224 ±2	**218 ±4**	**140 ±2**
22.48	3	5	38	7.042 ±0.006	142.3 ±0.4	120.8 ±0.4	68 ±1	1.048 ±0.003	352 ±9	*346 ±9*	*183 ±1*
22.63	3	5	53	6.144 ±0.005	164.2 ±0.5	102.6 ±0.3	68 ±1	1.020 ±0.003	300 ±5	*293 ±6*	*157 ±2*
22.63	3	5	54	6.254 ±0.008	158.2 ±0.2	106.5 ±0.1	70 ±1	1.041 ±0.001	330 ±4	*325 ±5*	*175 ±3*
23.27	3	5	117	6.954 ±0.008	123.3 ±0.2	121.3 ±0.2	67 ±1	1.066 ±0.002	406 ±9	*403 ±9*	*209 ±2*
23.63	3	6	3	7.173 ±0.009	111.3 ±0.2	124.5 ±0.1	64 ±1	1.060 ±0.001	402 ±8	*400 ±8*	*197 ±4*
23.98	3	6	38	6.888 ±0.006	66.3 ±0.2	111.1 ±0.3	60 ±1	0.986 ±0.003	268 ±4	*265 ±4*	*128 ±1*
23.98	3	6	38	7.632 ±0.009	69.9 ±0.1	127.5 ±0.1	57 ±1	1.021 ±0.001	324 ±4	**322 ±4**	**142 ±3**
24.24	3	6	64	9.364 ±0.011	75.8 ±0.1	153.2 ±0.2	51 ±1	0.999 ±0.002	300 ±3	*299 ±3*	*119 ±1*
24.49	3	6	89	8.544 ±0.007	172.0 ±0.5	133.3 ±0.4	54 ±1	0.953 ±0.003	242 ±3	*236 ±5*	*105 ±1*
24.49	3	6	89	8.253 ±0.010	67.7 ±0.1	137.2 ±0.1	55 ±1	1.016 ±0.001	319 ±4	**317 ±4**	**135 ±3**
24.98	3	6	138	7.382 ±0.009	75.1 ±0.1	124.4 ±0.2	66 ±1	1.029 ±0.002	319 ±4	*318 ±4*	*161 ±2*
25.48	3	7	38	8.238 ±0.010	72.1 ±0.1	138.2 ±0.2	56 ±1	1.025 ±0.002	335 ±4	**333 ±5**	**144 ±2**
25.90	3	7	80	7.531 ±0.006	73.2 ±0.2	126.0 ±0.4	51 ±1	1.022 ±0.003	343 ±8	**340 ±8**	**135 ±2**
25.90	3	7	80	7.457 ±0.009	75.2 ±0.1	125.1 ±0.2	51 ±1	1.025 ±0.002	348 ±5	**346 ±5**	**137 ±2**
25.90	3	7	80	7.261 ±0.009	76.9 ±0.1	122.0 ±0.1	53 ±1	1.026 ±0.001	345 ±5	**344 ±5**	**140 ±3**
26.13	4	1	3	5.543 ±0.007	101.3 ±0.1	93.9 ±0.1	54 ±1	1.036 ±0.002	366 ±6	**363 ±7**	**149 ±2**
26.32	4	1	22	5.230 ±0.006	106.0 ±0.1	89.0 ±0.1	59 ±1	1.040 ±0.002	360 ±6	*355 ±6*	*160 ±2*
26.48	4	1	38	5.896 ±0.007	117.4 ±0.3	98.3 ±0.2	46 ±4	1.019 ±0.002	350 ±12	*346 ±12*	*122 ±10*
27.15	4	1	105	4.121 ±0.005	447.3 ±1.3	71.4 ±0.1	58 ±4	1.059 ±0.002	428 ±23	*403 ±27*	*181 ±12*
27.76	4	2	16	4.480 ±0.005	388.7 ±1.1	73.0 ±0.1	49 ±4	0.995 ±0.002	298 ±7	*276 ±14*	*106 ±8*
28.49	4	2	89	8.114 ±0.007	121.9 ±0.4	130.6 ±0.4	69 ±1	0.983 ±0.003	254 ±3	**249 ±4**	**141 ±2**
29.13	4	3	3	7.994 ±0.010	144.6 ±0.4	128.6 ±0.2	61 ±4	0.983 ±0.002	263 ±5	*259 ±6*	*128 ±8*

Table 1: U and Th concentrations and U-Th ages for bulk carbonate samples from ODP hole 1008A. Errors are 2σ. Corr. Age is the age corrected for initial ^{230}Th assuming an initial $^{232}Th/^{230}Th$ atom ratio of 20,000. Ages marked in bold are considered reliable based on their $\delta^{234}U(T)$ and ^{232}Th concentrations, those in italics are not considered reliable.

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