Global and Planetary Change 66 (2009) 93–99

Contents lists available at ScienceDirect

Global and Planetary Change

journal homepage: www.elsevier.com/locate/gloplacha

Late Pleistocene Sea level on the New Jersey Margin: Implications to eustasy and deep-sea temperature

James D. Wright [a,*], Robert E. Sheridan [a], Kenneth G. Miller [a], Jane Uptegrove [b], Benjamin S. Cramer [c], James V. Browning [a]

[a] Department Geological Sciences, 610 Taylor Road, Rutgers University, Piscataway, NJ 08854, United States
[b] New Jersey Geological Survey, Trenton, NJ 08625, United States
[c] Department of Geological Sciences, 1272 University of Oregon, Eugene, OR 97403, United States

ARTICLE INFO

Article history:
Accepted 14 March 2008
Available online 30 November 2008

Keywords:
sea level
climate change

ABSTRACT

We assembled and dated a late Pleistocene sea-level record based on sequence stratigraphy from the New Jersey margin and compared it with published records from fossil uplifted coral reefs in New Guinea, Barbados, and Araki Island, as well as a composite sea-level estimate from scaling of Red Sea isotopic values. Radiocarbon dates, amino acid racemization data, and superposition constrain the ages of large (20–80 m) sea-level falls from New Jersey that correlate with Marine Isotope Chrons (MIC) 2, 3b, 4, 5b, and 6 (the past 130 kyr). The sea-level records for MIC 1, 2, 4, 5e, and 6 are similar to those reported from New Guinea, Barbados, Araki, and the Red Sea; some differences exist among records for MIC 3. Our record consistently provides the shallowest sea level estimates for MIC3 (~25–60 m below present); it agrees most closely with the New Guinea record of Chappell (2002; ~35–70 m), but contrasts with deeper estimates provided by Araki (~85–95 m) and the Red Sea (50–90 m). Comparison of eustatic estimates with benthic foraminiferal $\delta^{18}O$ records shows that the deep sea cooled ~2.5 °C between MIC 5e and 5d (~120–110 ka) and that near freezing conditions persisted until Termination 1a (14–15 ka). Sea-level variations between MIC 5b and 2 (ca. 90–20 ka) follow a well-accepted 0.1‰/10 m linear variation predicted by ice-growth effects on foraminiferal $\delta^{18}O$ values. The pattern of deep-sea cooling follows a previously established hysteresis loop between two stable modes of operation. Cold, near freezing deep-water conditions characterize most of the past 130 kyr punctuated only by two warm intervals (the Holocene/MIC 1 and MIC 5e). We link these variations to changes in Northern Component Water (NCW).

1. Introduction

Large changes in northern hemisphere ice sheets (NHIS, the "ice ages") during the past 2.6 myr have caused large and rapid changes in sea level (Shackleton and Opdyke, 1973), particularly during the Brunhes chron (past 780 kyr) when the 100-kyr Milankovitch-scale terminations resulted in over 100 m sea-level rises in less than 10 kyr (Broecker et al., 1968; Fairbanks, 1989). Two such terminations (1 and 2 of Broecker et al., 1968; ca. 10–18 and 130–135 ka, respectively) occurred during the past 130 kyr (the late Pleistocene), in addition to several smaller precessional- (19/23 kyr) and obliquity- (~41 kyr) scale changes (e.g., Shackleton, 2000; Waelbroeck et al., 2002).

Oxygen isotopic records of marine carbonates reflect ice-growth and decay events as increases (e.g., MIC[1] 2, 3b, 4, 5b, 5d, and 6) and decreases (e.g., MIC 1, 3a, 3c, 5a, 5c, and 5e) but also reflect changes in seawater temperature and local $\delta^{18}O_{seawater}$ variations in surface waters. Untangling the effects of global compositional changes in $\delta^{18}O$ (global $\delta^{18}O_{seawater}$, the ice-volume effect) from deep-sea or upper ocean water temperature changes has been the holy grail of Pleistocene paleoceanography. Mg/Ca provides one independent means of evaluating temperature effects, though this method has its limitations also (e.g., Rosenthal et al., 1997). Sea level varies as a function of ice-volume changes; thus, if global sea-level changes are known, this places constraints on the temperature and ice-volume components of the $\delta^{18}O$ record (e.g., Chappell and Shackleton, 1986; Cutler et al., 2003; Waelbroeck et al., 2002).

One complicating factor to understanding $\delta^{18}O_{seawater}$ variations has been that the history of sea-level change has been poorly known. For example, prior to the work of Fairbanks (1989), the position of sea-level at the last glacial maximum (LGM, ~20 ka) was not known within a factor of two and could have been anywhere between 80 and 150 m below present (Milliman and Emery, 1968). Fairbanks (1989) established that sea level at the LGM was 120 m below present by drilling and radiometrically dating corals (*Acropora palmata*) offshore of Barbados. The Barbados estimates for the past 20 kyr have been

* Corresponding author.
E-mail addresses: jdwright@rci.rutgers.edu (J.D. Wright), kgm@rci.rutgers.edu (K.G. Miller).

[1] Although dubbed oxygen isotopic stages by paleoceanographers for decades, the term stage is a stratigraphic term reserved for characterizing time-rock units (Hedberg, 1976). The proper term for isotopic variations are zones in depth and chron in time and to maintain consistency with prior terminology we use the term Marine Isotopic Chron (MIC).

0921-8181/$ – see front matter © 2009 Elsevier B.V. All rights reserved.
doi:10.1016/j.gloplacha.2008.03.013

confirmed in the uplifted Huon terraces of New Guinea (Cutler et al., 2003), in Tahiti (Bard et al., 1996), the Sunda shelf (southeast Asia; Hanebuth et al., 2000), and Bonepart Gulf (northwest Australia; Yokoyama et al., 2001)(see summary in Lambeck and Chappell, 2001); sea-level changes in these regions were similar, though details of the Termination 1 rise, in particular the exact timing and amplitude of the Termination 1a rise (14–15 ka), still require verification.

Studies of the uplifted fossil reefs in Barbados (Mesollela et al., 1969; Matthews, 1973; Fairbanks and Matthews, 1978), Haiti (Dodge et al., 1983), and New Guinea (Bloom et al., 1974; Chappell and Shackleton, 1986; Chappell et al., 1996; Chappell, 2002; Cutler et al., 2003) have provided calibrations for late Pleistocene sea-level changes from MIC 2 to 6 (150–20 ka). Studies of uplifted fossil reefs from Araki atoll (Urmos, 1985) and Barbados (Peltier and Fairbanks, 2006) have provided estimates for MIC 3 sea level. Detailed sea-level records have been recently derived and used to deconvolve the effects of temperature from the δ^{18}O record (Cutler et al., 2003; Waelbroeck et al., 2002). However, all of these regions used in sea-level reconstructions are experiencing uplift and the assumption of constant uplift rates over a 100+ kyr introduces a source of error of up to ±15 m. For example, Peltier (1998) showed that the late Pleistocene uplift rate assumed for the Huon peninsula (1.9 mm/yr) is too high for the Holocene (0.65 mm/yr). Siddall et al. (2003) provided a different estimate of eustatic changes by using Red Sea δ^{18}O values as a proxy for salinity changes caused by sea-level changes; this record is dependent on the models for salinity variations and needs verification. Sea-level records from tectonically quiescent regions such as the New Jersey margin spanning the last 150 kyr are valuable to constrain eustatic history.

Studies in New Jersey of the Holocene (Miller et al., this volume) and upper Pleistocene (Ashley et al., 1991) have provided a record of sea-level change back to the LGM. The Pleistocene section on this passive continental margin is thin and spotty due to low sediment accommodation space during an interval of rapid eustatic change. Therefore, a continuous late Pleistocene record cannot be obtained at one location. By using seismic stratigraphy to map upper Pleistocene sequences across the New Jersey continental shelf, Sheridan et al. (2000) pieced together a history of late Pleistocene sea-level change. Although the physical stratigraphy is well constrained (Fig. 1), their age control was limited. Here, we take advantage of new dates (Table 1) for seismic sequences mapped by Sheridan et al. (2000) and illustrated on Fig. 1 and use them to derive a sea-level curve for the late Pleistocene of New Jersey. We compare this record with sea-level estimates from New Guinea, Barbados, and Red Sea and with the deep-sea δ^{18}O record (Fig. 2).

2. High-resolution seismic data, New Jersey continental shelf

High-resolution seismic reflection profiles (~1–1.5 m resolution) across the New Jersey continental shelf provide a record of upper Pleistocene unconformity-bounded units (sequences) (Ashley et al., 1991; Carey et al., 1998; Sheridan et al., 2000, and references therein). Sheridan et al. (2000) compiled a composite seismic profile from Barnegat Inlet, NJ to the continental slope that showed seven seismic units (1/2, 3a, 3b, 3c, 4a, 4c, and 5) above a prominent reflector assigned to MIC 6 or older (Fig. 1). Dates for the seven units (Table 1) are derived from: 1) core samples (vibracore and coreholes) from the New Jersey shelf for lower stands of sea level that are preserved from near the modern beach (e.g., major interstadials) to the outer continental shelf; and 2) subaerial outcrops in New Jersey for MIC5e for the highest stand of sea level (Knebel et al., 1979; Wellner et al., 1993; Carey et al., 1998; Sheridan et al., 2000; McHugh and Olson, 2002; Uptegrove, 2003). Only superposition or broad amino acid racemization ages were available for dating MIC 3b, 3c, and 4. The Holocene sea-level (MIC1) record for New Jersey is from Miller et al. (this volume). All of our dates have been calibrated to calendar ages (Table 1) using CALIB 5.0.1 (Stuiver et al., 2005).

Unit 1/2 consists of lower transgressive and overlying highstand deposits. The transgression has been ^{14}C dated near the modern shoreline as ~4–8 ka and younger (Table 1; Fig. 1; Psuty, 1986; Miller et al., this volume). Unit 1/2 overlies a major subaerial erosional surface (R2 of Ashley et al., 1991) associated with the LGM. Lowstand deposits immediately above the LGM unconformity have been ^{14}C dated in shelf cores as 8–18 ka (Table 1); ^{14}C dates below the unconformity are 21 to 35 ka (Table 1; Knebel et al., 1979; Uptegrove, 2003).

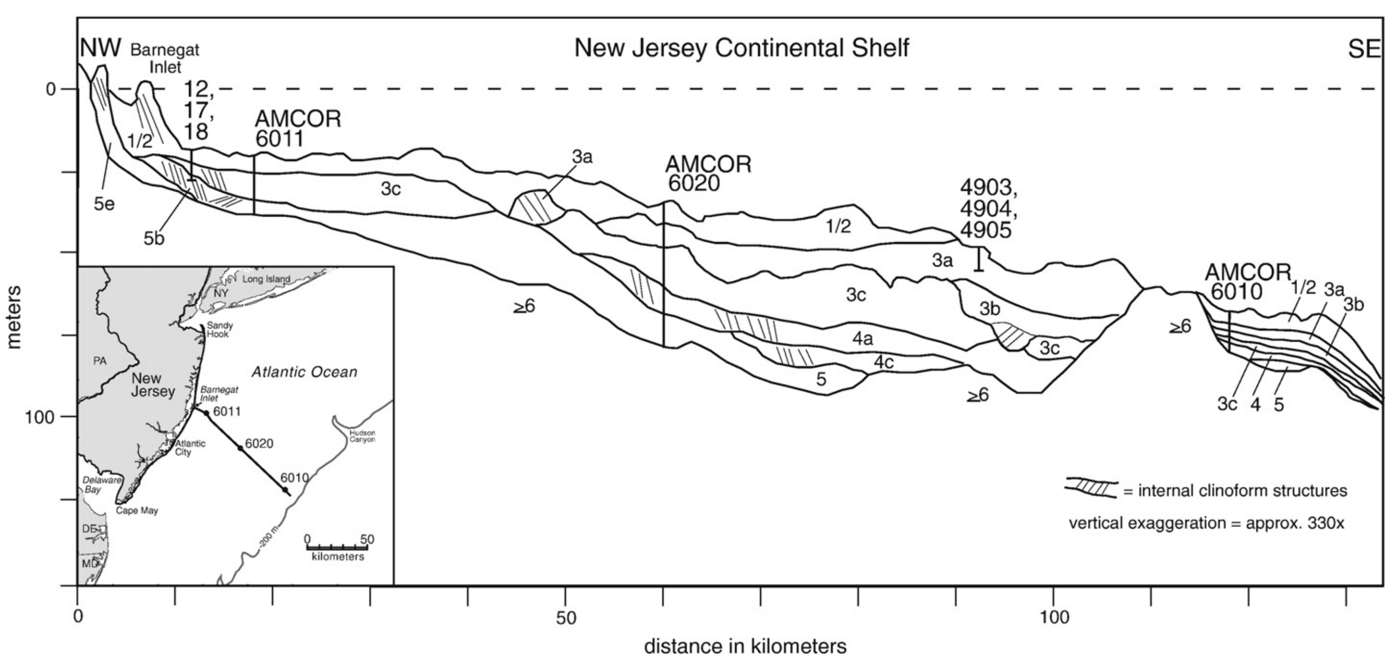

Fig. 1. Seismic stratigraphic cross section of the Late Pleistocene off Barnegat Inlet, New Jersey (modified after Sheridan et. al., 2000). AMCOR sites 6011,6020,6010 are located, as are vibracores 12,17,18 from Uptegrove (2003), and vibracores 4903,4904,4905 from Knebel et al. (1979), all core data referred to in Table 1. Vibracores are projected to the cross section using seismic correlations. Numbers 1/2, 3a, 3b, 3c, 4, 5, 5e, 5b, and 6 refer to correlations to marine isotope chrons. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 1

Location	Time interval	Reference	Age control	Age, ka	Elevation, m	Age error	Vertical error
NJ Holocene data from Miller et al. (this volume)							
NJ shelf	termination 1	Duncan et al. (2001)	14C age of 12300 in core Core 26/27 of 12,300 +/-450 recalbrated to calendar years before present	14.4	-78.0	0.5	3.0
NJ shelf	LGM	Dillon and Oldale (1978)	14C dates of ~21 ka on various cores calibrated to calendar years before present	21.0	-120.0	?	5.0
NJ shelf	3a	Knebel et al. (1979)	14C dates (28-36 ka; 35 ka best estimate) in vibracores 4902, -03, -04, -05	35.0	-30.0	7.0	5.0
NJ shelf	3b	Uptegrove (2003)	Superposition below 3a, above 3c	45.0	-60.0	10.0	5.0
NJ AMCOR 6020	3c	Sheridan et al. (2000)	Amino acid racemization (AAR) zone; 14Cdates (>40 ka) in vibracores 12a, 17	55.0	-20.0	10.0	5.0
NJ AMCOR 6020	4	Sheridan et al. (2000)	Superposition below 3c, above 5 in AMCOR 6020	70.0	-75.0	10.0	5.0
NJ Barnegat Inlet	5b	Uptegrove (2003)	Superposition below 3c shoreline, above stage 6 unconformity	90.0	-20.0	10.0	5.0
NJ outcrops	5e	Sheridan et al. (2000); Uptegrove (2003)	AAR zone in AMCOR 6020; AAR zone in vibracores 12b, 13, 18; 14C dates >40 ka	125.0	6.0	5.0	5.0
NJ shelf	6	Sheridan et al. (2000)	Below MIS 5 in AMCOR 6020; correlation of R reflector to nannofossil zones in ODP 903, 1073, and AMCOR 6010	130.0	-120.0	20.0	5.0

We date two sea-level highstands during MIC 3 using ^{14}C and amino acid racemization. High-resolution seismic stratigraphy from the incised valley of the ancestral Hudson River preserves two highstand systems tracts at 30 m and 20 m below modern sea level. Radiocarbon ages of 21–35 ka on the former (Knebel et al., 1979) suggest correlation with MIC 3a. Superposition, amino acid racemization, and radiocarbon

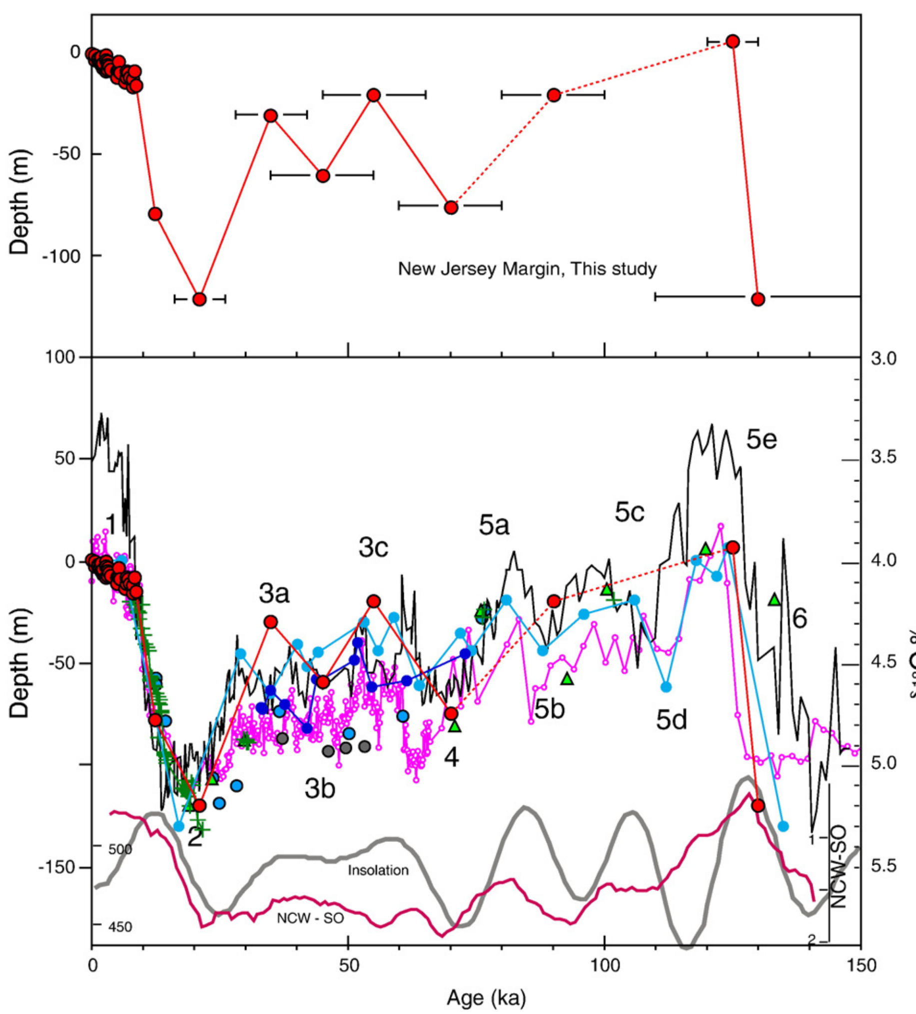

Fig. 2. Comparison of our sea-level record from mid-Atlantic U.S. margin (solid red circles) with Huon New Guinea terraces (blue filled circles, Cutler et al., 2003),light blue filled circles connected by lines (Chappell and Shackleton, 1986) dark blue circles connected by lines in MIC3, Chappell, 2002), Barbados (Green +, Fairbanks, 1989; Bard et al., 1990; green triangles, Cutler et al., 2003), the Araki record of Urmos (1985; gray circles), the Red Sea record of Siddall et al. (2003; pink circles), and benthic foraminiferal δ^{18}O record from Pacific (Carnegie Ridge) core V19-30 (Shackleton and Pisias, 1985; black line). Gray curve at bottom shows variations in insolation for June at 65°N latitude. 0 is modern sea level. NCW–SC is the difference between *Cibicidoides* benthic foraminiferal δ^{13}C records from northern North Atlantic Site 981 (Raymo et al., 2004) and Cape Basin 1089 (Hodell et al., 2003). The two sites were put onto a common age model by correlating δ^{18}O to V19-30 using the age model of Shackleton (2000). The difference curve was obtained by interpolating each δ^{13}C dataset to 1-kyr intervals, differencing, and then smoothing the difference curve with a 7-point (6 kyr) boxcar filter. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

dates on vibracores that yield ages older than 40 ka (Uptegrove, 2003) suggest correlation of the −20 m highstand with MIC 3c (Table 1). An intervening lowstand baymouth shoal is correlated with MIC 3b based on superposition with MIC 3a above and 3c below.

The age of deposits correlated to MIC 4 is moderately well constrained by superposition (Table 1) (Sheridan et al., 2000; Uptegrove, 2003), whereas ages of MIC 5b are constrained by amino acid racemization, and MIC 5e deposits are constrained by U–Th and amino acid racemization ages. Lowstand systems tracts are correlated with MIC 4 based on superposition in the AMCOR 6020 corehole between dated sediments of MIC 3c and 5 age (Sheridan et al., 2000; Table 1). Highstand deposits recovered in vibracores on the inner shelf have been dated by amino acid racemization as ~100 ka (Uptegrove, 2003), suggesting correlation to MIC 5c, though these are not used in our sea level curve because their paleodepths are not well constrained. Unit 5e can be traced on seismic profiles to the outcropping Cape May Formation in New Jersey where it is dated as equivalent to MIC 5e based on amino acid racemization (Sheridan et al., 2000) and U/Th dates from elsewhere on the Atlantic margin (Szabo, 1985) (Table 1). MIC 6 is observed in seismic profiles as a erosional surface truncating sloping beds, interpreted as a subaerial unconformity; it is poorly dated by superposition, with MIC 5 sediments above and correlation to nannofossil biostratigraphy at Sites 903, 1073, and AMCOR 6010 indicative of younger than MIC 8 (Sheridan et al., 2000).

We combine the age and depth estimates given above (Table 1) to form a composite sea-level estimate for New Jersey (Fig. 1). In general, sea-level studies of New Jersey have focused on documenting changes on the myr scale (e.g., Miller et al., 1998) because sedimentation and accommodation rates are generally too low to record Milankovitch-scale (10^4–10^5 yr) changes. The Pleistocene on the continental shelf is thin and spotty, though slope sites record very thick Pleistocene sequences (e.g., McHugh and Olson, 2002). The trick has been to get sufficient seismic coverage to link these disjunct Pleistocene shelf sequences, many of which are restricted to incised valleys, with the chronologic control on the outer shelf to slope. Sheridan et al. (2000) and this paper (Fig. 1) provide seismic control necessary to link together these sea-level variations. We show that the Milankovitch-scale sea-level cycles dated in carbonate terraces are also present on the New Jersey shelf. Six cycles of varying amplitude span MIC 6 to 2 having a ~20 kyr cyclicity, with a seventh cycle beginning at the base of Termination 1 (Fig. 1). Limitations in stratigraphic and seismic resolution prevent identification of higher-frequency events.

The New Jersey sea-level curve reflects eustasy, geoidal, and local tectonic effects. Long-term subsidence on this margin is thermoflexural with a maximum of rate ~10 m/myr at the shelf edge (Steckler, 1981) and can be largely neglected in studying Pleistocene sequences. The main other subsidence effect is flexural response to Pleistocene sediment loading on the continental slope and rise and Glacial Isostatic Adjustment (GIA) to the loading and unloading of the Laurentide ice sheet (Peltier, 1998). Near-field effects of the ice loading and unloading are experienced north of Barnegat Inlet, NJ; the sections examined here experienced the same far-field effects that are the similar from Barnegat Inlet through the Carolinas (Peltier, 1998). It is more difficult to evaluate the far-field effects on the 10 to 100 kyr scale which can be as high as 1.4 mm/yr in the modern (Peltier, 1998). We estimate flexural loading subsidence for the New Jersey shelf as ~5 m at MIC 4. By assuming that the ~15 m of sediments between MIC2 and MIC6 subaerial erosion surfaces at the edge of the shelf (Fig. 1) is a measure of the subsidence of the MIC6 land surface, then the younger MIC4 land surface in the middle of the shelf has subsided less than ~5 m. Older MIC 5 elevations are near the basement hinge zone and show little or no subsidence. MIC 3 shorelines are also close to the basement hinge, with minimal subsidence. Thus, the corrections for subsidence, including loading effects, included in our peak and trough estimates are minimal.

Potter and Lambeck (2003) argued that the U.S. margin was strongly imprinted by far field effects (glacio-hydro-isostasy or GIA). They modeled a gradient of up to 20 m from the Barbados to the U.S. margin for the difference between the position of sea level during MIC5e and 5a. This difference is attributed to differential GIA to the loading of the Laurentide ice sheet in MIC6 (Potter and Lambeck, 2003).

To evaluate the global component of sea level in our New Jersey curve, we compare our composite sea-level estimate with sea-level records from New Guinea, Barbados, Araki, and the Red Sea (Fig. 2). The sea-level records for MIC 1, 2, 4, 5e, and 6 are reasonably similar amongst the five records (Fig. 2). Age uncertainties preclude a definitive comparison with what we have tentatively identified as MIC 5b. However, large differences exist among records for MIC 3. This lack of agreement could be attributed to one of the following uncertainties: 1) isostatic response to far field ice loading and unloading effects in New Jersey; 2) uplift histories of the New Guinea terraces and Araki atoll; and 3) models for scaling δ^{18}O-salinity variations in the Red Sea to sea level.

Large differences amongst records in MIC 3 cannot be reconciled. In MIC 3, the New Jersey margin agrees well with the New Guinea record of Chappell (2002), with differences attributable to age errors in the New Jersey margin record. However, the Red Sea, Araki, and New Guinea records of Cutler et al. (2003) are 10–30 m deeper during MIC 3. It is difficult to determine a eustatic estimate from these curves. Whether differences among curves are real is problematic because of errors in each. Sea levels determined from the uplifted areas of New Guinea and Araki are calculated assuming constant linear uplift rates since MIC 5e, which have been shown to be incorrect (Peltier, 1998) as noted above. The Red Sea record of Siddall et al. (2003) is based not on direct measurement of sea level but on an oxygen isotopic proxy for salinity changes caused by sea-level changes; it is thus difficult to evaluate the sources of error which is estimated as ±12 m. The aforementioned geoidal effects could have caused differences among the sea-level records on the order of 20–30 m. Thus, errors of up to ±15 m occur in the various data sets older that MIC2, so differences of 20–30 m are difficult to evaluate. It is remarkable given these potential uncertainties that the records between the New Jersey margin and New Guinea (Chappell, 2002) agree well (Fig. 2).

3. Sea level and oxygen isotope comparison

We compare sea-level fluctuations from the New Jersey margin, Barbados, New Guinea, Araki, and the Red Sea with the piston core V19-30 (Carnegie Ridge, eastern Pacific) benthic foraminiferal δ^{18}O record [*Uvigerina peregrina*] (Shackleton and Pisias, 1985; Fig. 2). The δ^{18}O curve (Fig. 2) is scaled to elevation using the δ^{18}O-sea level relationship of 0.1‰/10 m (Fairbanks and Matthews, 1978) to show the expected glacioeustatic component of the δ^{18}O record. The δ^{18}O and sea-level records were aligned at the 20 kyr (LGM) level. The amplitudes of variation in the New Jersey margin and New Guinea (Chappell, 2002) sea-level record agree with the scaled benthic foraminiferal δ^{18}O record from MIC 2 through 5b. The δ^{18}O and sea-level records depart in MIC 5e and 1 (Fig. 2), arguing for a large temperature component in MIC 1 and 5 and minimal in 2–5b. However, comparison with the Araki, Red Sea, and the New Guinea record of Cutler et al. (2003) suggest that there may have been a significant temperature component in MIC 3.

We use the New Jersey margin and New Guinea sea-level estimates (Chappell, 2002) to evaluate the contribution of temperature versus ice volume on the deep-sea benthic foraminiferal δ^{18}O record. The total glacial–interglacial range measured in the benthic foraminiferal δ^{18}O record from MIC 6 to 5e and MIC 2 to 1 is 1.8‰; this is 0.6–0.8‰ higher than the 1.2‰ (Broecker, 1986) or 1.0 ‰ (Schrag et al., 1996; Shackleton, 2000) range predicted for ice-volume change. This difference between the measured and predicted benthic foraminiferal

^{18}O ranges indicates that there was a ~2.5 °C temperature change in addition to the ice volume component over these glacial–interglacial cycles. This warming in deep-water temperature suggests a switch to a new deep-water mode at these times.

Various studies have plotted late Pleistocene benthic foraminiferal $\delta^{18}O$ values versus sea level (Chappell and Shackleton, 1986; Chappell et al., 1996, Chappell, 2002; Waelbroeck et al., 2002; Cutler et al., 2003) as we have here (Fig. 3). Previous studies have argued that the deep sea alternated between cold versus warm modes. Cutler et al. (2003) and Waelbroeck et al. (2002) recognized that warm (similar to today) deep-sea conditions were associated with MIC 5 and Holocene MIC 1 versus near freezing conditions throughout most of the last 130 kyr. However, these previous studies did not recognize that these warm intervals represented a distinctly different mode of climate operation.

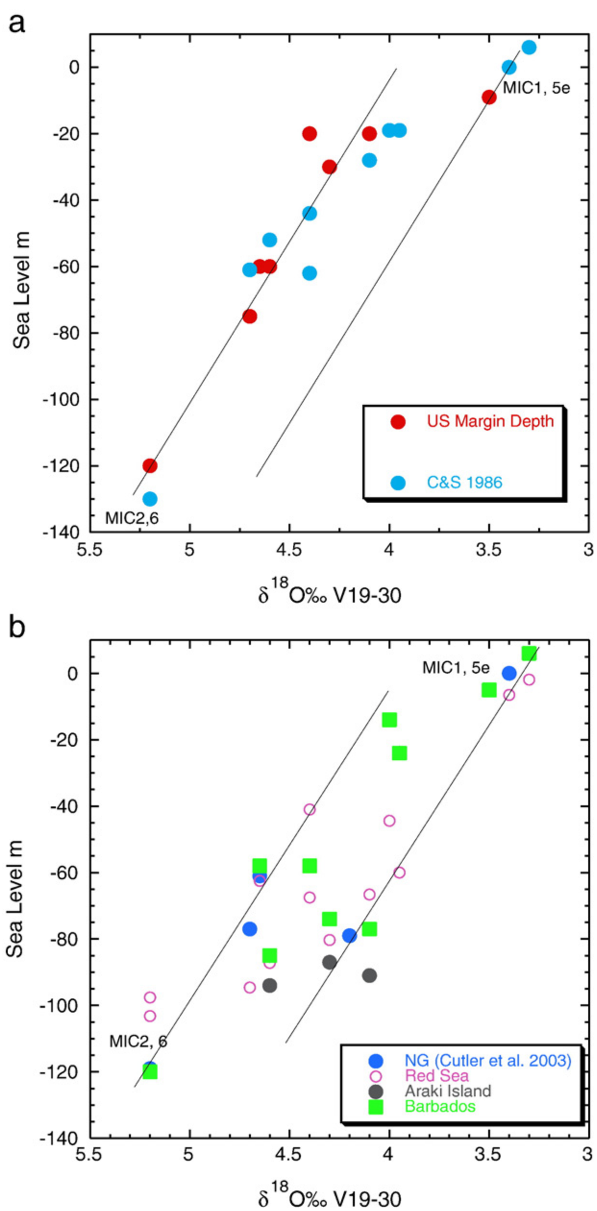

Fig. 3. New Jersey Margin (this study) and New Guinea (after Chappell, 2002) sea-level estimates (depth m) derived from Fig. 1 versus $\delta^{18}O$ values from V19-30 (a). Barbados, Araki, Red Sea and Cutler et al. (2003) sea level estimates for New Guinea versus $\delta^{18}O$ values from V19-30 (b). The lines in each graph have a slope of 100 m/1‰ (inverse of 0.11‰/10 m). The lines are anchored on the MIC 1/5e and MIC 2/6 points to show end member climate states. MIC = Marine Isotopic Chron.

We illustrate that deep-sea temperatures switched between two stable states of operations, one a cold mode that dominated late Pleistocene climate and the other a warm mode that only occurred during peak periods in northern hemisphere insolation during MIC 5 and 1. For each of the marine isotope chrons (Table 1; Fig. 2), sea level was plotted against the corresponding $\delta^{18}O$ value from V19-30 to illustrate the warm and cold modes. The New Jersey Margin and New Guinea data (Fig. 3a) were separated from the Cutler et al. (2003) New Guinea, Araki, and Red Sea data in another (Fig. 3b). The NJ margin and New Guinea data indicate that the warm modes were short-lived (~10 kyr) and were stable only during peaks in northern hemisphere insolation (Figs. 2, 3a). We note that the pattern of the glacial–interglacial climate change follows a hysteresis loop. The shift from the warm to cold mode followed MIC 5e and was associated with the largest decrease in northern hemisphere insolation of the past 150 kyr. Once this shift occurred, the climate system remained in the cold mode until the next termination.

In contrast to our preferred interpretation, if deeper MIC 3 sea levels from Barbados, Araki Island and the Red Sea are correct (Fig. 3b), then a significant deep-water warming (~1 to 2 °C) is required at this time (Fig. 3b). Waelbroeck et al. (2002) also argued that there was little to no warming in the deep Pacific during MIC3, though they similarly used New Guinea sea levels to constrain ice volume (e.g., Fig. 3a). Our NCW record (Fig. 2) independently suggests little change in deep water during MIC 3, which is consistent with minimal temperature change. If true, then this would argue that the shallower MIC 3 sea level estimates from New Guinea and New Jersey are more accurate. However, independent estimates of temperature change in the deep Pacific and/or additional sea-level data are needed to resolve the differences within MIC 3.

4. Discussion

The greatest mystery in middle–late Pleistocene $\delta^{18}O$ records (i.e., the Brunhes Chron, past 780 kyr) is the strong eccentricity signal because it is not predicted by direct insolation effects (e.g., Imbrie et al., 1993; Shackleton, 2000). It is clear from Shackleton (2000), Cutler et al. (2003), and this study (Fig. 1) that much of the $\delta^{18}O$ changes on the 100 kyr scale are due to the large warming during terminations that causes the sea-level and $\delta^{18}O$ records to diverge during the peak warmth of the MIC 1/Holocene and MIC 5e. The MIC 5e departure was short-lived (~10 kyr, which is <1/2 precessional cycle). Imbrie et al. (1992, 1993) noted that climate is linearly related to insolation forcing on the precessional and obliquity scales, but suggested that climate responds non-linearly to forcing on the 100-kyr eccentricity scale due to feedbacks either in the cryospheric or carbon cycles. Shackleton (2000) attributed the strong eccentricity component of the 100 kyr cycle to atmospheric pCO_2 changes. Imbrie et al. (1993), Shackleton (2000) and Cramer et al. (2003) suggested that the carbon cycle is the primary factor influencing temperatures in deep-water source regions and that eccentricity cyclicity evident in carbon-cycle proxy records imply that precessional forcing is the primary orbital component influencing atmospheric pCO_2. As Shackleton (2000) first suggested, Cutler et al. (2003) and this study show that the amplitude of deep-sea temperature change during precessional cycles is generally very small to non-existent. However, during eccentricity maxima, when precessional effects are at maximum, the system is propelled into a warm state. Our hysteresis curve implies that the climate system as reflected in deep ocean temperatures is not only jolted out of glacial states, but appears to be jolted back into them in concert with the largest changes in insolation (Fig. 2).

Deep-water changes are implicated in the abrupt warming and cooling associated with peak interglacials (MIC 5e and 1). Northern Component Water (NCW) originating in the Nordic seas is a source of heat to the deep sea (see summary in Imbrie et al., 1993). The difference between NCW and Southern Ocean $\delta^{13}C$ records provides a proxy for NCW intensity (Oppo and Fairbanks, 1997). The most significant features in the difference record (Fig. 2) are the low values recorded during MIC 5e and 1. This is interpreted to reflect a strong North Atlantic contribution to the Southern Ocean. In MIC 3a, 3c, 5a, and 5c, which represent interglacials with respect to ice volume, the conveyor contribution to the Antarctic was much lower relative to MIC 1 and 5e. Thus, the geochemical data indicates that the conveyor is responsible for the temperature anomalies in the deep-sea $\delta^{18}O$ record during MIC 5e and the Holocene.

5. Conclusions

Though fossil sunshine of coral reefs (including uplifted terraces) is considered the gold standard for reconstructing late Pleistocene sea-level changes, we show that a siliciclastic margin such as New Jersey can provide reasonable limits on sea level over the past 130 kyr. Our New Jersey sea level record agrees well with coral records for most of the past 130 kyr with the exception of MIC3 when all records show considerable differences. Though our record for this interval is the shallowest, it does compare well with the New Guinea coral record of Chappell (2002), but differs most for those in Araki Island and limited data from Barbados. Following previous studies, we show that the deep sea alternated between a warm and cold mode on 100 kyr short eccentricity scale, with likely little change on precessional or obliquity scales. We link these two modes to changes in NCW.

Acknowledgments

We thank G.M. Ashley, M.E. Katz, and D.V. Kent for comments and G.M. Ashley for discussions and J.D. Milliman, G. Camoin, and three anonymous reviewers for comments. Supported by National Science Foundation grants OCE0084032, EAR97-08664, EAR99-09179, and EAR0307112 (Miller).

References

Ashley, G.N., Wellner, R.W., Esker, D., Sheridan, R.E., 1991. Clastic sequences developed during late Quaternary glacio-eustatic sea-level fluctuations on a passive margin: example from the inner continental shelf near Barnegat Inlet, New Jersey. Geological Society of America Bulletin 103, 1607–1621.

Bard, E., Hamelin, B., Fairbanks, R.G., Zindler, A., 1990. Comparison of ^{14}C and Th ages obtained by mass spectrometry in corals from Barbados. Implications for the sea level during the last glacial cycle and for the production of ^{14}C by cosmic rays during the last 30,000 years. Nature 345, 405–410.

Bard, E., Hamelin, B., Arnold, M., Montaggioni, L.F., Cabioch, G., Faure, G., Rougerie, F., 1996. Deglacial sea level record from Tahiti corals and the timing of global meltwater discharge. Nature 382, 241–244.

Bloom, A.L., Broecker, W.A., Chappell, J.M.A., Matthews, R.K., Mesolella, K.J., 1974. Quaternary sea level fluctuations on a tectonic coast: new $^{230}Th/^{234}U$ dates from the Huon Peninsula, New Guinea. Quaternary Research 4, 185–205.

Broecker, W.S., 1986. Oxygen isotope constraints on surface ocean temperatures. Quaternary Research 26, 121–134.

Broecker, W.S., Thurber, D.L., Goddard, J., Ku, T.L., Matthews, R.K., Mesollela, K.J., 1968. Milankovitch hypothesis supported by precise dating of coral reefs and deep sea sediments. Science 159, 297–300.

Carey, J.S., Sheridan, R.S., Ashley, G.M., 1998. Late Quaternary sequence stratigraphy of a slowly subsiding passive margin, New Jersey continental shelf. American Association of Petroleum Geologists Bulletin 82, 773–791.

Chappell, J., 2002. Sea-level changes forced ice breakouts in the last glacial cycle: new results from coral terraces. Quaternary Science Reviews 21, 1229–1240.

Chappell, J., Shackleton, N.J., 1986. Oxygen isotopes and sea level. Nature 324, 137–140.

Chappell, J., Omura, A., Esat, T., McMulloch, M., Pandelfi, J., Ota, Y., Pillans, B., 1996. Reconciliation of Late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. Earth and Planetary Science Letters 141, 227–236.

Cramer, B.S., Wright, J.D., Kent, D.V., Aubry, M.-P., 2003. Orbital climate forcing of $\delta^{13}C$ excursions in the late Paleocene–early Eocene (Chrons C24n–C25n). Paleoceanography 18, 1097, doi:10.1029/2003PA000909.

Cutler, K.B., Edwards, R.L., Taylor, F.W., Cheng, H., Adkins, J., Gallup, C.D., Cutler, P.M., Burr, G.S., Bloom, A.L., 2003. Rapid sea-level fall and deep-ocean temperature change since the last interglacial period. Earth and Planetary Science Letters 206, 253–271.

Dodge, R.E., Fairbanks, R.G., Benninger, L.K., Maurasse, F., 1983. Pleistocene sea levels from raised coral reefs of Haiti. Science 219, 1423–1425.

Fairbanks, R., Matthews, R.K., 1978. The marine oxygen isotope record in Pleistocene coral, Barbados, West Indies. Quaternary Research 10, 181–196.

Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea level record influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature 342, 637–642.

Hanebuth, T., Stattegger, K., Grootes, P.M., 2000. Rapid flooding of the Sunda Shelf: a late-glacial sea-level record. Science 288, 1033–1035.

International stratigraphic guide: a guide to stratigraphic classification, terminology and procedure. In: Hedberg, H.D. (Ed.), International Subcommission on Stratigraphic Classification of IUGS Commission on Stratigraphy. John Wiley and Sons, New York, New York. 200 pp.

Hodell, D.A., Charles, C.D., Curtis, J.H., Mortyn, P.G., Ninnemann, U.S., Venz, K.A., 2003. Data report: oxygen isotope stratigraphy of ODP Leg 177 sites 1088, 1089, 1090, 1093, and 1094. In: Gersonde, R., Hodell, D.A., Blum, P. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results, vol. 177. [Online]. Available from World Wide Web: <http://www-odp.tamu.edu/publications/177_SR/chap_09/chap_09.htm>. [Cited 2004-08-10].

Imbrie, J., Boyle, E., Clemens, S., Duffy, A., Howard, W., Kukla, G., Kutzback, J., Martinson, D., McIntyre, A., Mix, A., Molfino, B., Morley, J., Peterson, L., Pisias, N., Prell, W., Raymo, M., Shackleton, N., Toggweiler, J., 1992. On the structure and origin of major glaciation cycles, 1. Linear responses to Milankovitch forcing. Paleoceanography 7, 701–738.

Imbrie, J., Berger, A., Boyle, E.A., Clemens, S.C., Duffy, A., Howard, W.R., Kukla, G., Kutzback, J., Martinson, D.G., McIntyre, A., Mix, A.C., Molfino, B., Morley, J.J., Peterson, L.C., Pisias, N.G., Prell, W.L., Raymo, M.E., Shackleton, N.J., Toggweiler, J.R., 1993. On the structure and origin of major glaciation cycles, 2. The 100,000-year cycle. Paleoceanography 8, 699–735.

Knebel, H.J., Wood, S.A., Spiker, E.C., 1979. Hudson River: evidence for extensive migration on the exposed continental shelf during Pleistocene time. Geology 17, 254–258.

Lambeck, K., Chappell, J., 2001. Sea-level change through the last glacial cycle. Science 292, 679–686.

Matthews, R.K., 1973. Relative elevation of late Pleistocene high sea level stands: Barbados uplift rates and their implication. Quaternary Research 3, 147–153.

McHugh, C.M.G., Olson, H.C., 2002. Pleistocene chronology of continental margin sedimentation: new insights into traditional models, New Jersey. Marine Geology 86, 389–411.

Mesollela, K.J., Matthews, R.K., Broecker, W.S., Thurber, D.L., 1969. The astronomical theory of climate change: Barbados data. Journal of Geology 77, 250–274.

Miller, K.G., Mountain, G.S., Browning, J.V., Kominz, M., Sugarman, P.J., Christie-Blick, N., Katz, M.E., Wright, J.D., 1998. Cenozoic global sea-level, sequences, and the New Jersey Transect: results from coastal plain and slope drilling. Reviews of Geophysics 36, 569–601.

Milliman, J.D., Emery, K.O., 1968. Sea levels during the past 35,000 years. Science 162, 1121–1123.

Oppo, D., Fairbanks, R.G., 1997. Variability in the deep and intermediate water circulation of the Atlantic Ocean during the past 25,000 years: Northern hemisphere modulation of the Southern Ocean. Earth and Planetary Science Letters 86, 1–15.

Peltier, W.R., 1998. Postglacial variations in the level of the sea: implications for climate dynamics and solid-earth geophysics. Reviews of Geophysics 36, 603–689.

Peltier, W.R., Fairbanks, R.G., 2006. Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. Quaternary Science Reviews 25, 3322–3337.

Potter, E., Lambeck, K., 2003. Reconciliation of sea-level observations in the Western North Atlantic during the last glacial cycle. Earth and Planetary Science Letters 217, 171–181.

Psuty, N.P., 1986. Holocene sea level in New Jersey. Physical Geography 7, 156–167.

Raymo, M.E., Oppo, D.W., Flower, B.P., Hodell, D.A., McManus, J.F., Venz, K.A., Kleiven, K.F., McIntyre, K., 2004. Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene. Paleoceanography 19, PA2008–PA2021.

Rosenthal, Y., Boyle, E.A., Slowey, N., 1997. Temperature control on the incorporation of Mg, Sr, F and Cd into benthic foraminiferal shells from Little Bahama Bank: prospects for thermocline paleoceanography. Geochim. Cosmochim. Acta 61, 3633–3643.

Schrag, D.P., Hampt, G., Murray, D.W., 1996. Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean. Science 272, 1930–1932.

Shackleton, N.J., 2000. The 100,000-year ice-age cycle identified and fund to lag temperature, carbon dioxide and orbital eccentricity. Science 289, 1897–1902.

Shackleton, N.J., Opdyke, N.D., 1973. Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28–238: oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale. Quaternary Research 3, 39–55.

Shackleton, N.J., Pisias, N.G., 1985. Atmospheric carbon dioxide, orbital forcing, and climate. In: Sunquist, E.T., Broecker, W.S. (Eds.), The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present. American Geophysical Union, Washington, D.C., pp. 303–317.

Sheridan, R.E., Ashley, G.M., Miller, K.G., Waldner, J.S., Hall, D.W., Uptegrove, J., 2000. Onshore–offshore correlation of upper Pleistocene strata, New Jersey coastal plain to continental shelf and slope. Sedimentary Geology 134, 197–207.

Siddall, M., Rohling, E.J., Almogi-Labin, A., Hemleben, Ch., Meischner, D., Schmelzer, I., Smeed, D.A., 2003. Sea-level fluctuations during the last glacial cycle. Nature 423, 853–858.

Szabo, B.J., 1985. Uranium-series dating of fossil corals from marine sediments of southeastern United States Atlantic Coastal Plain. Geological Society of America Bulletin 96, 398–406.

Steckler, M.S., 1981, Thermal and mechanical evolution of Atlantic-type margins (Ph.D. thesis): New York, Columbia University, 261p.

Stuiver, M., Reimer, P.J., Reimer, R., 2005. Radiocarbon Calibration Program Revision 5.0.1. Copyright 1986–2005.
Uptegrove, J., 2003, Late Pleistocene facies distribution, sea level changes and paleogeography of the inner continental shelf off Long Beach Island, New Jersey [M.S. thesis]: New Brunswick, NJ, Rutgers University, 238 p.
Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J.-C., McManus, J., Lambeck, K., Balbon, E., Labracherie, M., 2002. Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. Quaternary Science Review 21, 295–305.
Wellner, R.W., Ashley, G.M., Sheridan, R.E., 1993. Seismic stratigraphic evidence for a submerged middle Wisconsin barrier: implications for sea-level history. Geology 21, 109–112.
Urmos, J.P., Oxygen isotopes, sea levels, and uplift of reef terraces, Araki Island, Vanuatu. 1985, M.S. thesis, Cornell University, 123pp.
Yokoyama, Y., Esat, T.M., Lambeck, K., 2001. Coupled climate and sea-level changes deduced from Huon Peninsula coral terraces of the last ice age. Earth and Planetary Science Letters 193, 579–587.