Geological Society, London, Memoirs

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Geological Society, London, Memoirs 2014, v.41; p21-34.
doi: 10.1144/M41.3

Notes

Chapter 3

History of continental shelf and slope sedimentation on the US middle Atlantic margin

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Abstract: We describe sedimentation on the storm-dominated, microtidal, continental shelf and slope of the eastern US passive continental margin between the Hudson and Wilmington canyons. Sediments here recorded sea-level changes over the past 100 myr and provide a classic example of the interplay among eustasy, tectonism and sedimentation. Long-term margin evolution reflects changes in morphology from a Late Cretaceous–Eocene ramp to Oligocene and younger prograding clinothem geometries, a transition found on several other margins. Deltaic systems influenced Cretaceous and Miocene sedimentation, but, in general, the Maastrichtian–Palaeogene shelf was starved of sediment. Pre-Pleistocene sequences follow a repetitive model, with fining- and coarsening-upward successions associated with transgressions and regressions, respectively. Pleistocene–Holocene sequences are generally quite thin (<20 m per sequence) and discontinuous beneath the modern shelf, reflecting starved sedimentation under high rates of eustatic change and low rates of subsidence. However, Pleistocene sequences can attain great thickness (hundreds of metres) beneath the outermost shelf and continental slope. Holocene sedimentation on the inner shelf reflects transgression, decelerating from rates of approximately 3–4 to around 2 mm a^{-1} from 5 to 2 ka. Modern shelf sedimentation primarily reflects palimpsest sand sheets plastered and reworked into geostrophically controlled nearshore and shelf shore-oblique sand ridges, and does not provide a good analogue for pre-Pleistocene deposition.

Supplementary material: References used in the comparison of all dates for New Jersey localities in Figure 3.8 are available at http://www.geolsoc.org.uk/SUP18749.

The US middle Atlantic margin

The US middle Atlantic region (offshore New Jersey–Delaware: Figs 3.1 & 3.2) is a classic passive continental margin dominated by modern siliciclastic sedimentation, generally low sediment input during the last 10 myr and a long (180–200 myr) history of sedimentation that resulted in over 16 km of sediment buried beneath the modern shelf (Fig. 3.3) (e.g. Grow & Sheridan 1988). The margin was the subject of early studies because of its proximity to researchers using emerging technologies during the early twentieth century, and its physiography was used in defining the terms shelf, slope and rise (Heezen *et al.* 1959). Oil exploration generated great interest in the late 1970s and early 1980s, with attendant seismic profiling (e.g. Poag 1985), stratigraphic test wells (e.g. Scholle 1980), and 32 exploration wells primarily on the outer continental shelf and slope. Although oil and gas prospects were deemed uneconomical, focus on the stratigraphy of continental margins and sea-level change led to this region becoming a target of academic drilling by the Deep Sea Drilling Project (DSDP) Leg 95 (summary in Poag 1985), the Ocean Drilling Program (ODP) legs 150 (Mountain *et al.* 1994), 174A (Christie-Blick *et al.* 2003) and onshore ODP legs 150X and 174AX (summary in Browning *et al.* 2008), and the Integrated Ocean Drilling Program (IODP) Expedition 313, New Jersey Shallow Shelf (Mountain *et al.* 2010). These academic efforts have returned a wealth of information that provides unparalleled documentation of Cretaceous–Miocene sedimentation from the onshore coastal plain to the continental rise. However, Plio-Pleistocene sediments on this margin are very thin and discontinuous and, despite great efforts to core and date (e.g. Carey *et al.* 2005; Goff *et al.* 2005), are still poorly known. In this contribution, we: (1) review the physiographical and oceanographical settings and modern sedimentation on this margin; (2) discuss the geological structure and Cretaceous–Miocene sedimentation; (3) review Plio-Pleistocene sedimentation including the last glacial cycle (approximately the last 20 kyr), contrasting it with older periods; and (4) provide a brief overview of Holocene, instrumental and future sea-level changes on this margin.

Physiographical and oceanographical setting

The shelf, slope and rise physiographical provinces on the US middle Atlantic margin (offshore New Jersey and the Delmarva Peninsula: Fig. 3.1) were defined on the basis of the seafloor gradient: the shelf dips seawards with a gradient of less than 1:1000 (<0.06°), the continental slope with a gradient greater than 1:40 (>1.4°) and the continental rise with a gradient of about 1:100 (*c.* 0.6°) (Heezen *et al.* 1959; Emery 1968). The continental shelf is wide (>150 km) in this region and the water depth at the shelf–slope break averages about 135 m (Table 3.1) (Heezen *et al.* 1959). The shelf can be roughly divided into inner (0–40 m), middle (*c.* 40–100 m) and outer shelf (>100 m), marked by scarps (mid Shelf and Franklin: Fig. 3.2) that are both cut by the modern and palaeo-Hudson River valleys (Fig. 3.2) (Goff *et al.* 1999). The inner–middle shelf is flatter than the outer shelf (Fig. 3.2). The continental slope is incised by numerous submarine canyons, the largest of which indent the shelf–slope break. We focus here on the continental shelf and slope sedimentation between the Hudson and Wilmington canyons (Fig. 3.1).

Three water masses are found in the middle Atlantic shelf and slope (the following summary is largely derived from Beardsley

From: Chiocci, F. L. & Chivas, A. R. (eds) 2014. *Continental Shelves of the World: Their Evolution During the Last Glacio-Eustatic Cycle*. Geological Society, London, Memoirs, **41**, 21–34. http://dx.doi.org/10.1144/M41.3

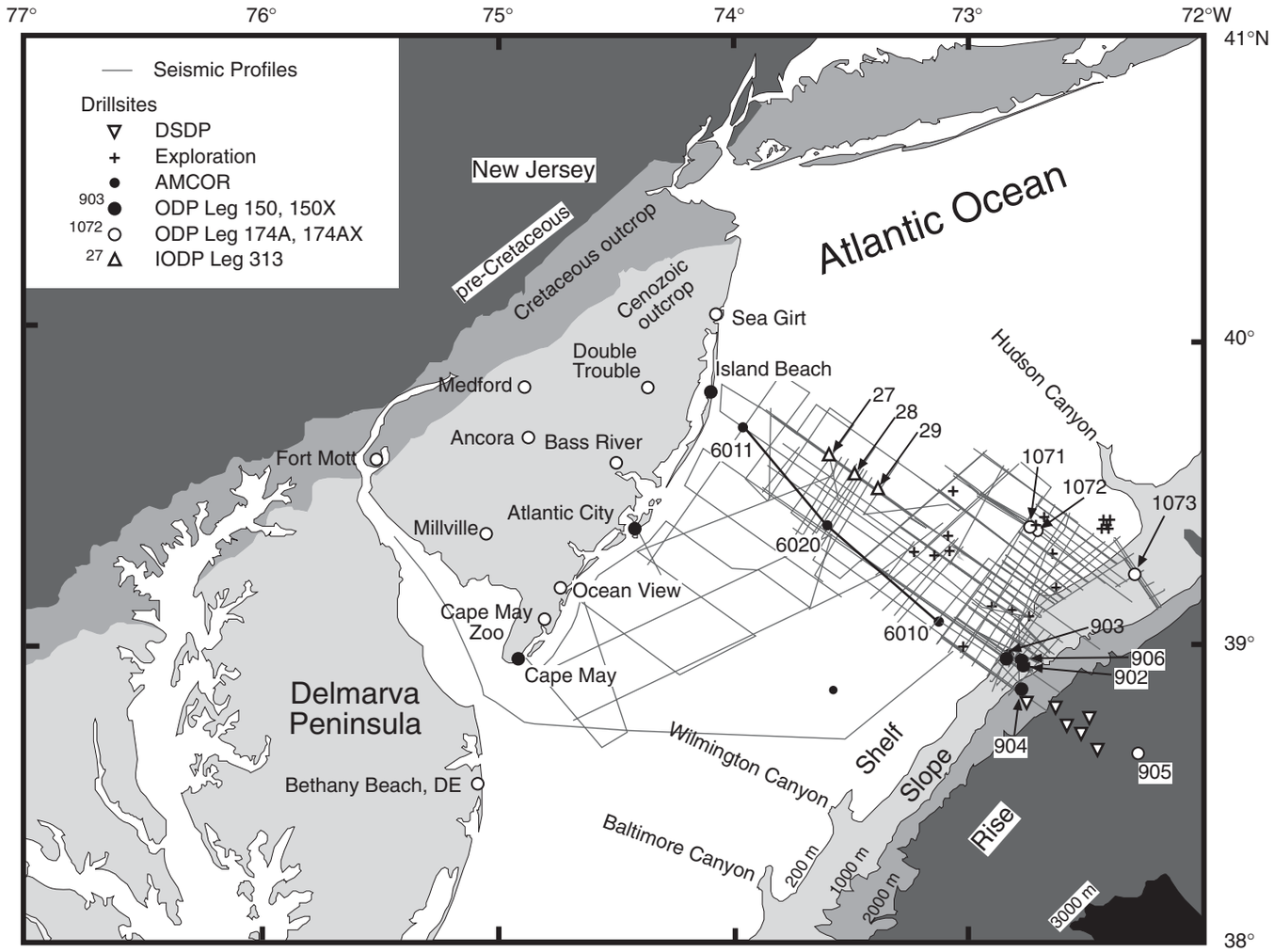

Fig. 3.1. Generalized location map showing coastal plain outcrop ages, generalized bathymetry, shelf edge, onshore and offshore coreholes, and multichannel seismic profiles collected in the 1990s that target Cenozoic strata. The black line connecting AMCOR 6011 and 6010 is shown in Figure 3.6.

& Boicourt 1981): (1) relatively fresh Shelf Water (salinity of <35‰); (2) more saline Slope Water (35–36‰); and (3) warm (>18 °C), salty (>36‰) Gulf Stream water. Slope Waters generally dominate on the continental slope, although the front between Shelf and Slope Waters is not rigidly fixed to the shelf–slope break. In addition, Gulf Stream eddies (warm core rings) advect warm, salty water to the slope and occasionally to the shelf. The source of Slope Water has been attributed to local winter cooling but it probably has a source from the Labrador Current, as indicated by the $\delta^{18}O_{seawater}$ tracer (Fairbanks 1982). A cyclonic gyre of NE-flowing Slope Water and SW-flowing Shelf Water is geostrophically balanced by regional wind forcing, although the processes controlling the position and movement of the boundary between Shelf and Slope Waters are poorly understood. The main thermocline and oxygen minimum zones in this region are shallow and seasonably stable (<400 m), with relatively high O_2 values in the minimum zone (>3 ml l^{-1}; *c.* 94 μm kg^{-1}: Miller & Lohmann 1982). The continental rise falls within the influence of the Western Boundary Undercurrent (WBUC), a strong SW-flowing bottom current composed primarily of North Atlantic Deep Water with an admixture of Antarctic Bottom Water (Heezen *et al.* 1966). The strongest flow of this deep geostrophic current is between water depths of 3000 and 4900 m on the continental rise, although the current has migrated up and down the rise through time, and may have impinged on the lower slope at times (Heezen *et al.* 1966).

The physical oceanography of the US middle Atlantic continental shelf can be characterized by timescales that correspond to annual, seasonal, storm, tidal and wave components (Beardsley & Boicourt 1981). Historical analyses indicate that the long-term annual flow on the New Jersey shelf is alongshore to the SW at average velocities of less than about 20 cm s^{-1}, with values during winter storms and hurricanes exceeding 30–60 cm s^{-1} (Beardsley & Boicourt 1981; Lyne *et al.* 1990; Gong *et al.* 2010). A recent overview of annual and seasonal changes in surface water circulation on the shelf (Gong *et al.* 2010) confirms this earlier result. Mean surface flow over the New Jersey shelf for the period 2002–2007 was 2–12 cm s^{-1} along the shelf and offshore to the south (Gong *et al.* 2010). Both the Hudson Shelf Valley (Fig. 3.2) and the shelf edge act as dynamic barriers that define the continental shelf circulation. Topography, seasonal stratification and wind forcing control surface flow, which is in the approximate direction of the wind during the winter season when the water column is unstratified and more to the right of the wind during the summer season when the water column is stratified.

New views of the seasonal circulation have emerged (Gong *et al.* 2010). During the summer, SW winds drive cross-shelf offshore flows and favour upwelling. During the winter, cross-shelf

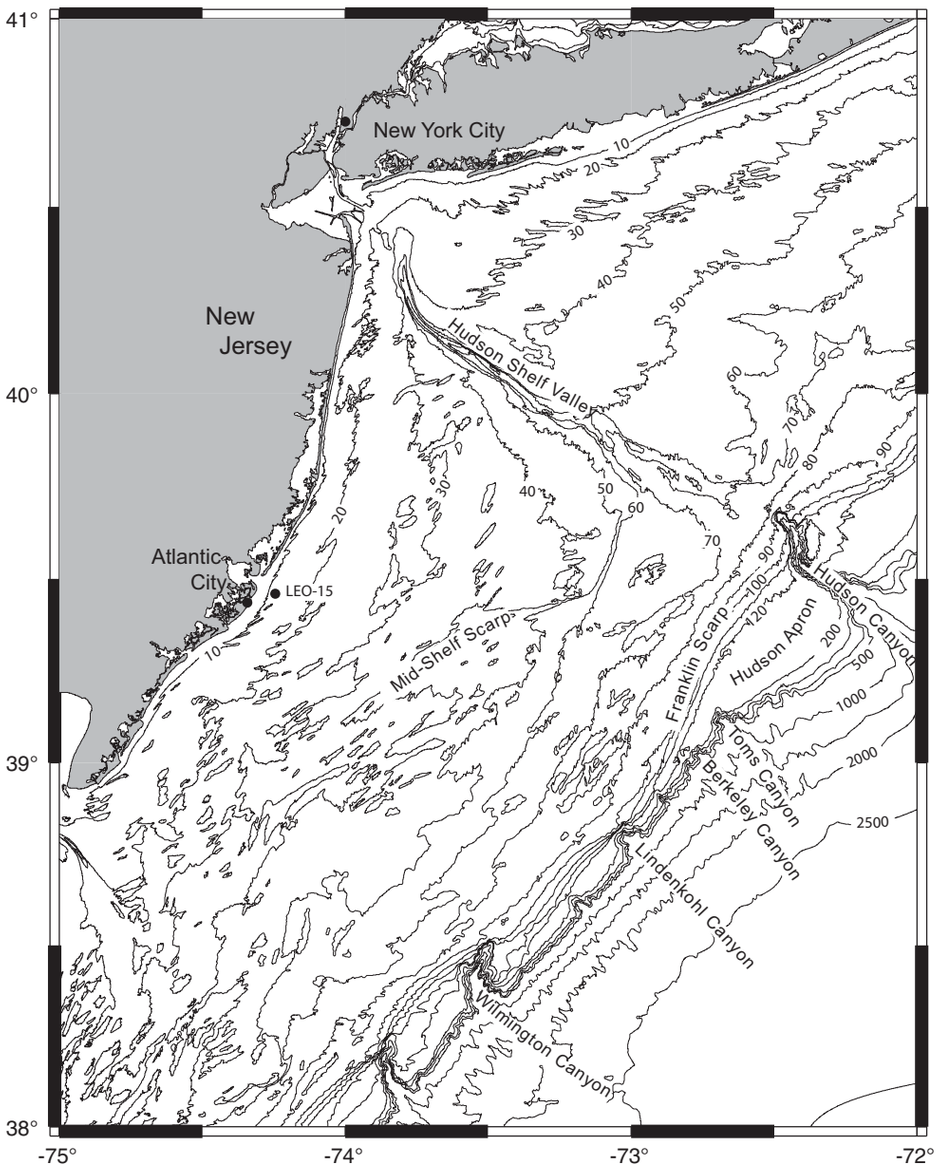

Fig. 3.2. Bathymetry of the US Atlantic margin between the Hudson and Wilmington canyons showing the 10 m contour interval.

offshore flows are driven by NW winds. Longshore NE winds associated with storm events drive energetic along-shelf flows during the autumn and spring. Thus, surface transport is cross-shelf during summer and winter, and along-shelf during the spring and autumn. Measurements indicate that the currents strong enough to entrain and transport sediment were confined to high-energy events such as storms. At higher frequencies, less energetic semi-diurnal astronomical tides with an average tidal range of 1.2 m (i.e. microtidal) drive flows that are generally orientated in the shore-normal direction. The maximum velocity of the near-surface (5 m below) tidal currents averages about 20 cm s^{-1}, and that of the near-bottom (2 m above) tidal currents averages about 10 cm s^{-1}. Although the velocities of the tidal currents are lower than other currents, most non-tidal currents flow alongshore and tidal currents often dominate the onshore/offshore signal. At very high frequencies, the energy spectrum is dominated by surface waves that are usually less than 1 m, travelling shorewards with the larger waves generated by offshore events and propagating in from offshore (Glenn *et al.* 2008).

Recent underwater measurements on the shelf by autonomous underwater vehicles ('gliders') show that the combined action and timing of surface waves, tides and storm-driven currents can explain the observed temporal variation in storm resuspension (Glenn *et al.* 2008). Key to sediment dispersal is the role of the pycnocline, which determines how far up in the water column the sediment is resuspended and made available for transport: (1) on the inner shelf, the presence or absence of a seasonal pycnocline is a function of upwelling/downwelling; (2) on the middle shelf, a persistent seasonal stratification limits direct linkage between the upper wind-driven boundary layer and the lower combined wave and current boundary layer until mixing occurs in the full water column, usually in the autumn; and (3) on the outer shelf, the effects of surface waves are limited to only the most severe storms (Glenn *et al.* 2008).

Modern sedimentation

The modern US middle Atlantic continental shelf is generally starved because sediment input is trapped in estuaries. Surface sediments across the shelf are almost exclusively sand (i.e. >63 mm: Hollister 1973). Modern sediments fail to follow the classic graded shelf model of progressive fining offshore (Swift 1969). Instead, the complete dominance of sands on the modern shelf prompted Emery (1968) to classify modern shelf

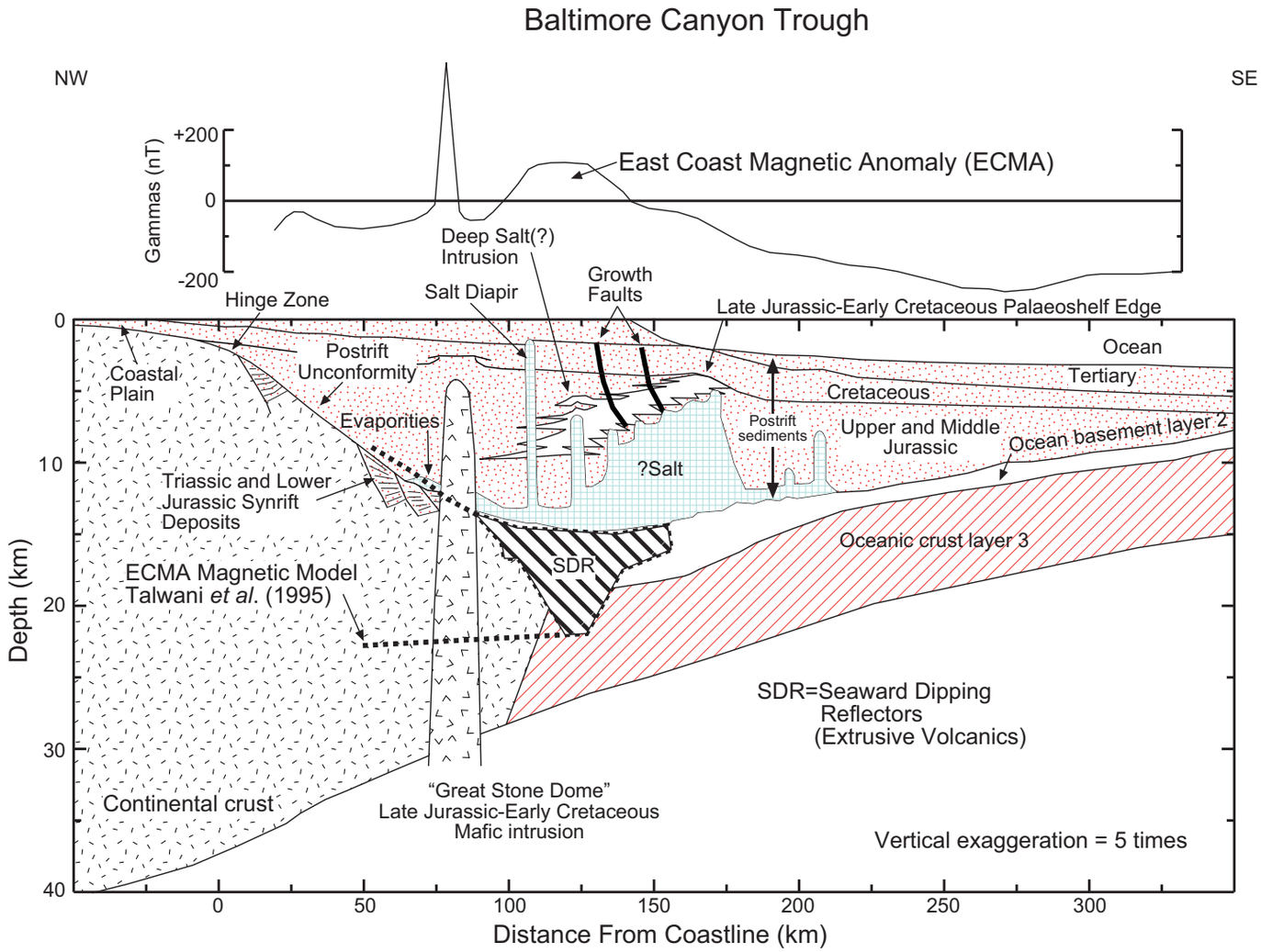

Fig. 3.3. Updated cross-section of the Baltimore Canyon Trough, modified after Grow & Sheridan (1988). The modifications are based on all publicly available seismic data, deep drilling, and most recent interpretations and magnetic modelling of the EDGE experiment (Talwani *et al.* 1995). The most significant differences from previous work are the interpretation of thick salt and clarification of the transition from ocean to continental crust.

sedimentation in this region and many others throughout the world as 'relict,' a product of the Holocene transgression and not in equilibrium with modern shelf processes. Swift *et al.* (1971) recognized that the sands were relict, in the sense that they were initially deposited in a different environment (e.g. shoreface sediments now residing on the middle shelf) but subsequently have been redeposited in a modern hydrodynamic equilibrium, and applied the term 'palimpsest'. Thus, the terms 'relict' and 'palimpsest' used in this Memoir were derived first from this well-studied margin.

Table 3.1. *Characteristics of the US middle Atlantic margin (Hudson Canyon to Wilmington Canyon)*

Length of the shelf	*c.* 200 km
Average width	150 km
Mean tide range	1.2 m
Waves	Generally <1 m
Currents	<60 cm s^{-1}
Dominating process (wave/current/tide)	Storm
Average depth of the shelf break	135 m
Siliciclastic/carbonate/authigenic/glacial sedimentation	Siliciclastic
Modern/relict/palimpsest (if possible in approximate %)	Palimpsest
Tectonic trend over the last glacial cycle (stable/uplifting/subsiding)	Stable

The sands on the modern shelf are arrayed into a series of ridges and swales that reflect multiple processes which mold shelf sedimentation today (Fig. 3.2) (e.g. Ashley *et al.* 1991; Goff *et al.* 1999, 2005). Although the source of modern quartz sand in this region is ultimately from the Appalachian Mountains via major river systems (Hudson, Delaware and Susquehanna), the maturity of the sands argues for the recycling of coastal plain sediments in the nearshore zone (Pazzaglia & Gardner 1994), with the likely source being primarily from the onshore Miocene Cohansey and Kirkwood formations and younger surficial units. These sands were reworked in nearshore/shoreface environments during the Holocene transgression across the shelf.

Today, the nearshore zone has both shore-attached and -detached sand ridges (1–12 m thickness, 2–20 km in length, 1–5 km spacing: Goff *et al.* 2005) that often run at oblique angles to the shoreline (Fig. 3.2) (Ashley *et al.* 1991). Similar ridges are found on the middle shelf off New Jersey in approximately 40 m of water (e.g. Stubblefield *et al.* 1983) and on the outer shelf (Goff *et al.* 1999). The origin of these ridges has been controversial, with early studies favouring an abandoned barrier origin (see the summary in Swift *et al.* 1973). However, numerous studies have shown that these ridges are reworked in

hydrodynamic equilibrium with shelf currents (Swift *et al.* 1973; Stubblefield *et al.* 1983; Rine *et al.* 1991; Goff *et al.* 1999, 2005). Although the sands of inner and middle shelf ridges were emplaced during transgression, they are strongly modified by modern currents (Stubblefield *et al.* 1983; Goff *et al.* 2005). The role of modern currents on outer shelf ridges is still debated. Goff *et al.* (1999) suggested that the outer shelf ridges were largely erosional, whereas other studies have suggested strong modification by currents (Swift *et al.* 1973; Stubblefield *et al.* 1983; Rine *et al.* 1991). Studies in the nearshore zone at the LEO15 site (15 m present depth offshore of Tuckerton, NJ) (Fig. 3.2) show that longshore currents (particularly alongshore geostrophic storm-generated flows) are sufficiently energetic to entrain and transport sands (Styles & Glenn 2005). We suggest that reworking of shelf sand to form ridges by geostrophic currents is analogous to build-up of deep-sea drift deposits (Heezen *et al.* 1966), albeit despite differences in grain size and velocities. This explains the oblique orientation of many of the ridges and the fact that many may be erosional remnants.

Recent Chirp seismic reflection data have identified several interesting features on the outer continental shelf (Goff *et al.* 2005). In addition to sand ridges that are largely erosional remnants, NE–SW striations identified as 'sand ribbons' occur in swales and appear to be deposited by currents in water depths of 50–100 m. Elongated pits approximately 0.5–1.5 km long, 1–3 km wide and up to 10 m deep are erosional remnants. Buried Pleistocene erosional channels display both a bathymetric and backscatter expression in this region. Finally, striations interpreted as iceberg grooves have been mapped on the outer continental shelf (Goff *et al.* 1999; Goff & Austin 2009) that may have been filled by transparent and/or chaotic fill interpreted as a catastrophic flooding event (Fulthorpe & Austin 2004).

Slope sediments reflect a mixture of downslope transport and pelagic sedimentation. The modern continental shelf break at approximately 135 m in this region is related primarily to the position of sea level during the last glacial maximum that resulted in the shedding of sands directly into the deep sea, v. finer-grained muds deposited on to the slope during the Holocene. Surface sediments consist of upper slope silts that grade down to lower slope clays (Hollister 1973). Because most modern riverine sediment input is trapped in estuaries, relatively little coarse terrigenous sediment reaches the slope today, although there is evidence that more coarse material was deposited on the slope during glacials (Christensen *et al.* 1996). Thus, modern sedimentation on the slope is primarily hemipelagic and sediments are derived from muds carried as suspended material from river discharge or from resuspended shelf sediments carried off the shelf (Doyle *et al.* 1979). Slope sediments have high (>1%) but variable organic carbon values and are generally carbonate poor (<20%), with carbonate content increasing downslope as a result of increased input of pelagic carbonates and decreased input of terrigenous material (Miller & Lohmann 1982). Considerable speculation has centred on whether Holocene sedimentation on the slope is dominated by sediments transported downslope (slide, slumps, debris flows, turbidites) or pelagic/hemipelagic rain. Examination of surface samples shows a pattern of largely *in situ* benthic biofacies (Miller & Lohmann 1982). Submersible observations of the lower slope in the immediate region show that pelagic sediments drape the bottom; outcrops are restricted to occasional near-vertical walls. Visual and core evidence for large- and small-scale transport is largely limited to blocks found at the foot of the slope (2200 m in this region) and sporadic turbidity-current activity in some canyon thalwegs (McHugh *et al.* 1993; Pratson *et al.* 2007).

Geological structure and early margin history

The US middle Atlantic continental shelf and slope is a classic passive margin, and in this review we focus on the region between the Hudson and Wilmington submarine canyons that contains a thick sedimentary record of up to 16 km. Rifting and subsequent separation from NW Africa occurred during the Late Triassic–earliest Jurassic (*c.* 230–190 Ma), forming a series of rift basins that extend from the onshore today to beneath the modern shelf (Fig. 3.3: e.g. Grow & Sheridan 1988). Seafloor spreading began prior to the Bajocian (*c.* 175 Ma: Middle Jurassic), with the likely opening beginning off Georgia by around. 200 Ma and progressing northwards to the US middle Atlantic margin (Withjack *et al.* 1998). This south–north 'zipper' rifting is associated with a diachronous post-rift unconformity that separates active 'rift-stage' (synrift) deposits (strongly influenced by syndepositional horst and half-graben structures) from passive margin 'drift-stage' deposits which accumulated in a progressively widening and deepening basin. Prior to the Plio-Pleistocene, post-rift history of the middle Atlantic region is dominated by simple thermal subsidence, sediment loading, lithospheric flexure, compaction and sea-level changes (Watts & Steckler 1979; Reynolds *et al.* 1991; Miller *et al.* 2005; Kominz *et al.* 2008), though mantle dynamics impacted the longer-term (more than 1 myr) record (e.g. Rowley 2013). Local normal faulting (minor, except for several large growth faults beneath the modern outer continental shelf), rare salt diapirism and a single Early Cretaceous igneous intrusion (the Great Stone Dome: Fig. 3.3) locally complicate the otherwise simple passive margin post-rift tectonic history (Fig. 3.3) (Poag 1985). Glacial Isostatic Adjustments (GIA) played a major role following the development of large northern hemisphere ice during the past 2.7 myr (Peltier 1998). Most of these are far-field effects but, during major Pleistocene glacials (Marine Isotope Chrons (MIC) 2, 6 and, perhaps, others – note that most authors use the term Marine Isotope Stage (MIS) but this is a stratigraphically incorrect usage of the term 'stage', the proper term is 'chron') (Stanford *et al.* 2001), continental ice sheets reached northern New Jersey and influenced shelf–slope sedimentation through near-field GIA effects.

Up to 16 km of post-rift sediments accumulated along the US middle Atlantic region in an offshore basin termed the 'Baltimore Canyon Trough' (BCT: Fig. 3.3). The Jurassic section is composed of shallow-water limestones and shales (typically 8–12 km) that are restricted to the offshore BCT (Fig. 3.3). In the BCT, salt of probable Jurassic age has progressively migrated as a diapiric ridge seawards over the initial ocean crust under the East Coast Magnetic Anomaly (Fig. 3.3). This salt migration is similar to what has been documented in the Gulf of Mexico and the Nova Scotian margin.

Long-term global sea-level rise plus thermal subsidence and flexural bending of the crust beneath the coastal plain led to progressive widening of the BCT during the Early Cretaceous (*c.* 120–140 Ma: Fig. 3.3) (Watts & Steckler 1979; Olsson *et al.* 1987). A fringing great barrier reef marked the edge of the shelf–slope break during the Jurassic–Early Cretaceous, and the margin experienced mixed siliciclastic and carbonate deposition (Jansa 1981; Poag 1985) (Fig. 3.3). The reef prograded seawards to a position no more than a few tens of kilometres seawards of the modern shelf edge during the Early Cretaceous and then disappeared. It is not clear what caused the demise of the barrier reef. Deltaic sediments subsequently overstepped the reef in the Early Cretaceous and, during a long interval of Late Cretaceous transgression, the shelf–slope break moved landwards once again. River input to the shelf may have overwhelmed the carbonate platform and reef, although climatic change and northward latitudinal drift may also be implicated in the demise of this great barrier reef of eastern North America. Siliciclastic input resulted in moderately thick (*c.* 2–3 km) offshore Cretaceous strata containing several major sandbodies and onshore deposits that are generally deltaically influenced (e.g. Sugarman *et al.* 1995; Kulpecz *et al.* 2008; Browning *et al.* 2008). After these Cretaceous pulses of sand input into the BCT, accumulation rates were generally low–moderate during the latest Cretaceous–Palaeogene

when siliciclastic and carbonate fine-grained sediment accumulated (Poag 1985). Deposition during this latter interval occurred on a carbonate ramp with a gradient of around 1:500 (Steckler *et al.* 1999; see discussion below).

Cretaceous–Miocene sedimentation

The passive US middle Atlantic margin is a natural laboratory for unravelling sea-level history and the response of sedimentation to sea-level changes. Drilling onshore and offshore of New Jersey and Delaware has yielded a more than 100 myr record of sea-level changes (Fig. 3.4) (Miller *et al.* 2005; Browning *et al.* 2008; Kominz *et al.* 2008). Fundamental to reconstructing sea level is the realization that relative sea-level falls (i.e. the combination of global sea level (eustasy) and tectonism) cause erosional unconformities and that these unconformities can be used to divide the stratigraphic record into sequences (Vail *et al.* 1977).

Early studies of the US middle Atlantic region recognized that the sea encroached the onshore coastal plain during transgressions and retreated during regressions numerous times during the Late Cretaceous–Miocene (*c.* 100–5 Ma: Owens & Sohl 1969; Olsson 1975). It is clear that these transgressions and regressions moulded the stratigraphic record buried beneath the modern shelf (Poag 1985). More recent studies placed these transgressive

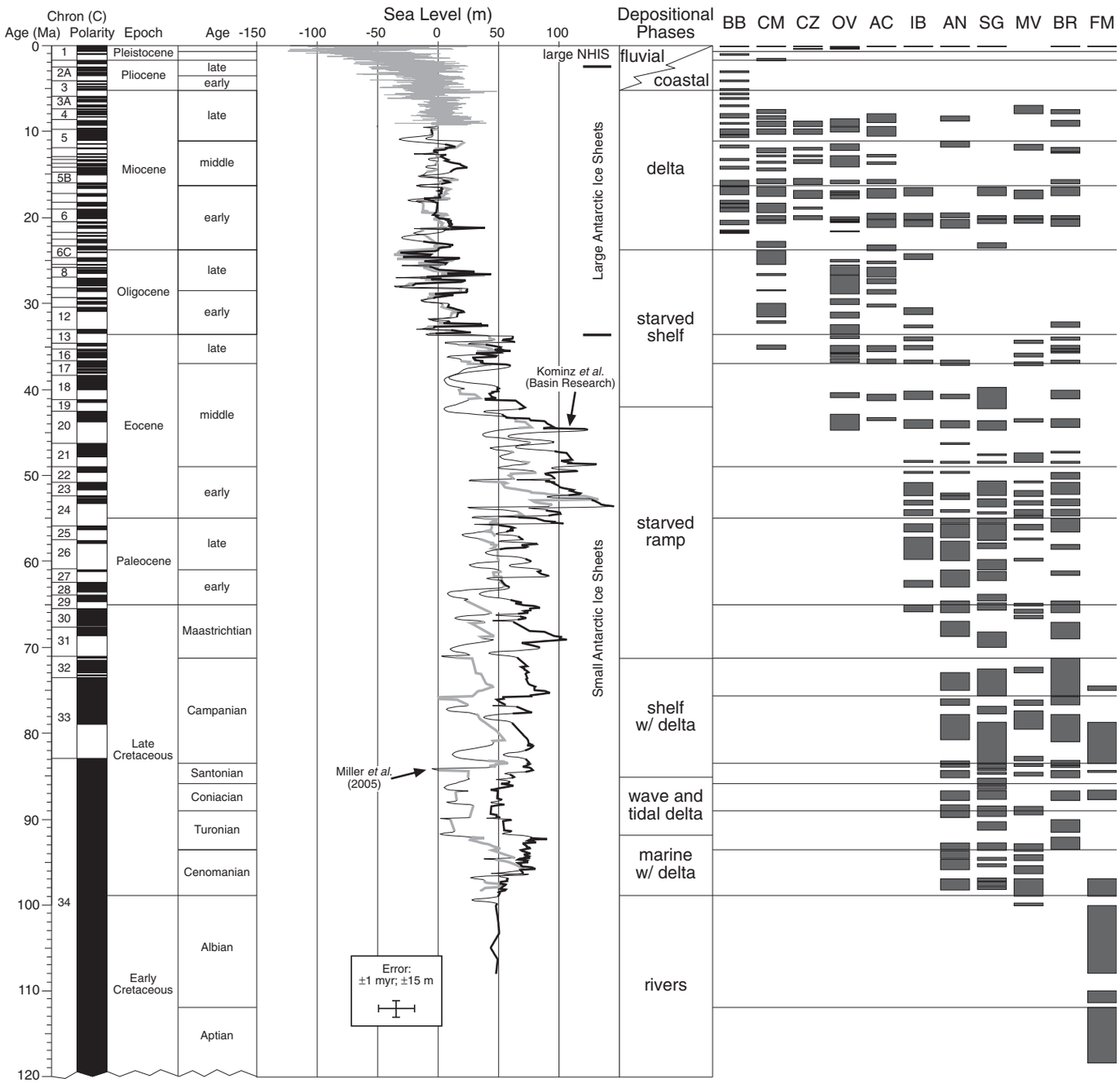

Fig. 3.4. Sea-level curves of Miller *et al.* (2005) and Kominz *et al.* (2008), depositional regimes of the onshore coreholes (Fig. 3.1), and distribution of sediments in sequences in the onshore coreholes as a function of time. BB, Bethany Beach core; CM, Cape May core; CZ, Cape May Zoo core; OV, Ocean View core; AC, Atlantic city core; IB, Island Beach core; AN, Ancora core; SG, Sea Girt core; MV, Millville core; BR, Bass River core; FM, Fort Mott core. Modified after Browning *et al.* (2008).

and regressive facies into a sequence stratigraphic framework. Continuous coring by the ODP legs 150X and 174AX onshore, and 150 and 174A offshore (see the summary in Miller *et al.* 2005; Browning *et al.* 2008), has provided one of the best-dated records of sequences (unconformity-bounded units) and a well-defined history of sea-level changes for the past 100 myr. To differentiate the effects of eustasy from other influences, a modelling technique termed 'backstripping' (Watts & Steckler 1979) was applied to this well-dated record of water-depth changes, progressively removing the effects of compaction, loading and thermal subsidence (see the summary in Miller *et al.* 2005). In the absence of regional or local tectonics, backstripping provides a global sea-level estimate, with the greatest uncertainty resulting from errors in estimates of water depth at the time of deposition. Backstripped records from 11 onshore ODP coreholes (Fig. 3.4) generally yielded similar sea-level estimates (summary in Kominz *et al.* 2008) that compare well to those from other passive margins and epicontinental seas (e.g. US Gulf Coast, NW Europe, the Russian Platform) and the oxygen isotope proxy for glacioeustasy (Miller *et al.* 2005), suggesting that global sea-level changes were a dominant process controlling million-year-scale sequences on this margin.

Although sea level appears to be a dominant control on million-year-scale sequences (Miller *et al.* 2005), facies changes within sequences reflect changes in accommodation (including effects of sea level and subsidence), sediment supply and provenance (Posamentier *et al.* 1988). In contrast to modern sedimentation, Cretaceous–Miocene nearshore and shelf sequences follow a classic pattern of a graded shelf and Walther's law, where the horizontal pattern is repeated vertically (Browning *et al.* 2008). In contrast, such patterns are not often observed in Plio-Pleistocene sequences due to low accommodation rates, low sediment supply to the shelf and high rates of global sea-level change. In Cretaceous–Miocene sections, a sequence consists of a basal unconformity overlain by transgressive sands (transgressive systems tract, TST) dominated by glauconite in the Late Cretaceous–Palaeogene and quartz in the Miocene. The TST is overlain by a coarsening-upward regressive highstand systems tract (HST). Two different facies models were derived for the region (Browning *et al.* 2008): one for riverine/deltaic influence sequences, and one for classic storm-dominated shoreface or neritic environments. Close to riverine influence, the HSTs consist of lower prodelta silts and upper delta front sands (Sugarman *et al.* 1993; Browning *et al.* 2008). Away from riverine influence, the HST consist of offshore muds, lower shoreface, shelly, heavily bioturbated, heterolithic silty fine and very fine sands, distal upper shoreface fine–medium sands with admixed silts, and upper shoreface/foreshore fine–coarse, well-sorted sands, with opaque heavy mineral laminae (see Browning *et al.* 2008). The changing water-depth patterns inferred from lithofacies changes are mirrored in the distribution of benthic foraminifera, giving confidence that the transgressive–regressive packages reflect shifting water depths on a graded shelf.

Backstripping applied to these transgressive–regressive sequences in the coastal plain wells reveal numerous Late Cretaceous–Miocene sea-level cycles: 15–17 Late Cretaceous, 6 Paleocene, 12 Eocene, 7 Oligocene and 14 Miocene sequences are shown (Fig. 3.5). The amplitudes of the eustatic estimates are best constrained (better than ± 10 m resolution) in the Oligocene by two-dimensional (2D) backstripping. They are less constrained in the Late Cretaceous–Eocene where most lowstands are missing, and are poorly constrained in the Neogene because much of each cycle is missing (Miller *et al.* 2005; Kominz *et al.* 2008). Despite this limitation, 'Icehouse' sequences of the last 33 myr show high amplitudes (up to 60 m), consistent with control by growth and decay of large (near modern-sized) Antarctic ice sheets (Miller *et al.* 2005). 'Greenhouse' Late Cretaceous–Eocene sequences show lower amplitudes (typically 25 m, although a few are 40 m); nevertheless, a 25 m global sea-level change in less than 1 myr can be explained only by the growth and decay of a significant ice sheet (i.e. about one-third of the modern ice volume). This apparent conflict with the well-documented history of warm global temperatures at this same time (e.g. Huber *et al.* 2002) has been explained by invoking small, ephemeral ice sheets in the Greenhouse world (see the summary in Miller *et al.* 2005).

The 100–1000 myr history of the US middle Atlantic margin reflects the changing influences of two large river systems during the Cretaceous and latest Oligocene–Miocene. Sediment thickness measured in onshore wells and mapped in offshore seismic profiles indicates a north source inferred to be the ancestral Hudson River draining the northern Appalachian Mountains, and a central source inferred to be the ancestral Delaware or Susquehanna rivers draining the central Appalachians (Kulpecz *et al.* 2008; Monteverde *et al.* 2008).

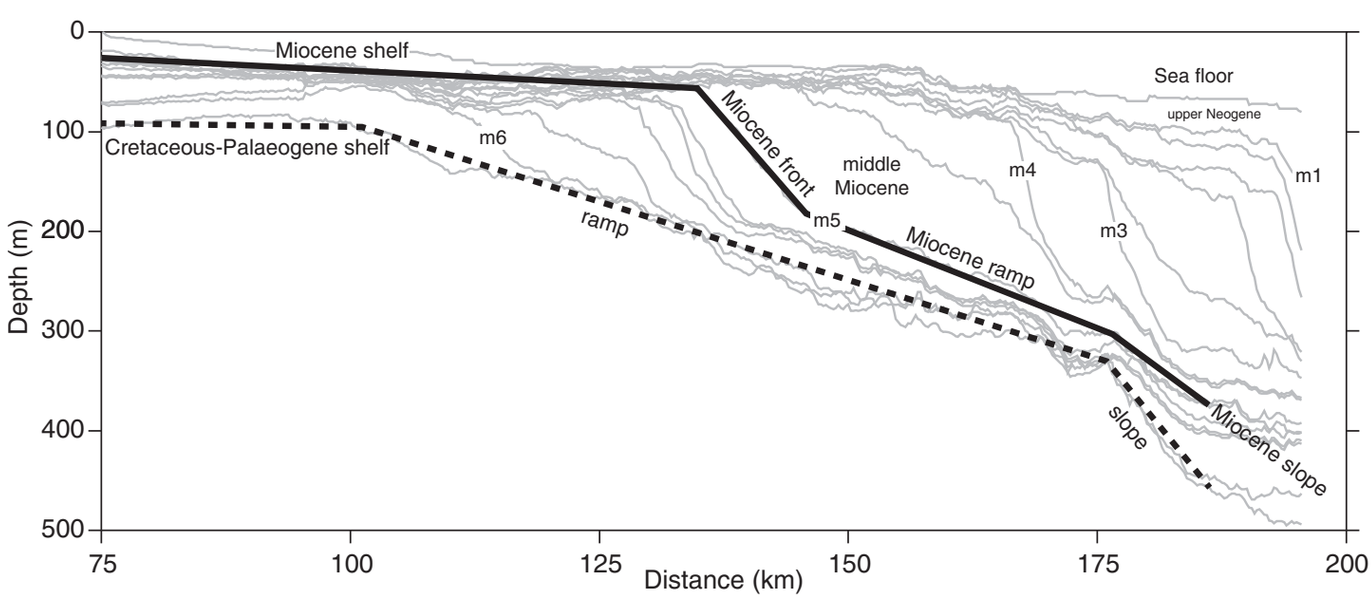

Fig. 3.5. Two-dimensional backstripped cross-sections across the New Jersey shelf. Generalized profiles for the Late Cretaceous (thick dashed line) and Miocene (thick solid line) are labelled as shelf, ramp, front (Miocene only) and slope. Modified after Steckler *et al.* (1999). Origin is the modern shoreline.

Browning *et al.* (2008) noted that the sediment facies evolved through eight depositional regimes controlled by changes in accommodation, long-term sea level and sediment supply (Fig. 3.4): (1) the Early Cretaceous consisted of anastomosing riverine environments deposited during a time of warm climates, high sediment supply and high accommodation; (2) the Cenomanian–early Turonian was dominated by marine sediments with a minor deltaic influence associated with long-term sea-level rise; (3) the late Turonian–Coniacian was dominated by non-marine fluvial wave and tidal delta systems associated with long-term sea-level fall; (4) the Santonian–Campanian consisted of marine deposition under the influence of a wave-dominated delta associated with a long-term sea-level rise and increased sediment supply; (5) Maastrichtian–middle Eocene deposition consisted primarily of starved, carbonate ramp–shelf environments associated with very high long-term sea level and low sediment supply; (6) the late Eocene–Oligocene was a starved siliciclastic shelf associated with high long-term sea level and low sediment supply; (7) early–middle Miocene sediments were deposited on a prograding shelf under a strong wave-dominated deltaic influence associated with a major increase in sediment supply and accommodation; and (8) over the past 10 myr, low accommodation and eroded coastal systems were associated with a long-term sea-level fall and low rates of sediment supply due to bypassing.

The evolution of the US middle Atlantic margin also reflects a long-term change from a carbonate ramp to a prograding siliciclastic margin. A major switch from carbonate ramp deposition to starved siliciclastic sedimentation (the 'carbonate switch') occurred progressively from the middle Eocene onshore to earliest Oligocene on the continental slope in response to global and regional cooling (Miller *et al.* 1997). Sedimentation rates increased dramatically in the late Oligocene–Miocene (Poag 1985) due to increased input from the hinterland (Poag & Sevon 1989; Pazzaglia & Gardner 1994). These thick sandbodies, arrayed as prograding units beneath nearly the entire modern shelf, have recently been continuously cored in the inner shelf region by IODP Expedition 313 (Mountain *et al.* 2010). IODP drilling recovered about 16 early–middle Miocene sequences at three sites spanning topset, inflection, foreset and bottomset deposits that provide an unprecedented coring of facies across seismically imaged sequences.

A change in margin morphology occurred during the carbonate switch, as revealed by reconstruction of past depositional surfaces using 2D backstripping (Fig. 3.5) (Steckler *et al.* 1999). Three Cretaceous–Eocene physiographical provinces can be recognized (Fig. 3.5): shelf (1:1000; 0–100 + m water depths), ramp (1:300; 100–325 m water depths) and slope (<1:100; 325–2000 m water depths). Four Miocene physiographical provinces can be recognized (Fig. 3.5): shelf (1:1000; 0–50 m water depths); front (<1:40; 50–200 m water depths); ramp (1:300; 200–350 m water depths); and slope (>1:40; 350–2000 m water depths). A flatter shelf with one sharp shelf edge developed in the Pleistocene.

The modern continental shelf break occurs in a zone of thickened oceanic crust that extends from thinned continental crust beginning approximately 10 km eastwards of the Great Stone Dome (*c.* 80 km in Fig. 3.3; *c.* 100 km in Fig. 3.5) to typical oceanic crust seawards of the dome. A change in declivity analogous to the modern shelf–slope break has existed and varied in position within this zone since rifting. At the time of separation from NW Africa (*c.* 180 Ma), the shelf–slope break was approximately 25–30 km landwards of its present location (Fig. 3.3). Two-dimensional backstripping places a break between a more steeply dipping ramp and the slope seawards of the Great Stone Dome (*c.* 175 km in Fig. 3.5; 105 km in Fig. 3.3) during the Late Cretaceous–Palaeogene. Terrestrial sediment supply decreased markedly in the Maastrichtian–Palaeogene, resulting in carbonate ramp–shelf deposition to a break in slope landward of the Great Stone Dome (Fig. 3.3). Siliciclastic input increased once again in the late Oligocene and spiked in the middle Miocene (Miller *et al.* 1997; Steckler *et al.* 1999) in what appears to be a global pattern (Bartek *et al.* 1991; Lavier *et al.* 2001) that may have been a result of global climatic cooling (Steckler *et al.* 1999). This led to two regions of distinct change in seafloor declivity for the Neogene: (1) a shelf–front break whose position was controlled by the advance and retreat of shallow-water (<100 m) clinothems; and (2) the shelf–slope break that remained close to the previous reef edge (175 km: Fig. 3.5). By the Pleistocene, the shelf–front break, controlled by clinothem location, prograded close to its present position (Fig. 3.5). In this way, the structural shelf break has existed within a 75 km-wide zone just seawards of continental crust since the time of first seafloor spreading; water depth at the shelf break has varied from around 100 m to slightly more than 300 m (Fig. 3.5), as observed on other modern margins.

Plio-Pleistocene sedimentation

The Pliocene onshore in New Jersey consists of fluvial gravels of the Pennsauken Formation that are poorly dated (Stanford *et al.* 2001). The Pliocene section in Delaware is non-marine and also poorly dated, but is fully marine in the Yorktown Formation in Virginia. This implies about 20 m of differential subsidence/uplift between Virginia and New Jersey/Delaware. The Pliocene is surprisingly poorly represented beneath the shelf and upper slope (Mountain *et al.* 2007). The reasons for this are unclear, although low sediment input and accommodation and/or mantle dynamic effects (Rowley 2013), may have been exacerbated by GIA effects of the development of large northern hemisphere ice sheets at around 2.7 Ma (Shackleton & Opdyke 1973).

Mountain *et al.* (2007) provided a detailed summary of Pleistocene sedimentation on this margin. The Pleistocene is characterized by a low-relief hinterland that provided minimal sediment input, extensive reworking on a wide shelf and little accommodation from thermal subsidence. As a result, the Pleistocene section beneath the inner–middle shelf is thin and spotty, and stacked in complex patterns owing to low sediment input and low accommodation during an interval of rapid eustatic change. For example, short-lived increases in accommodation space were fed by downcutting of channels occurring in concert with short-term sea-level rise at suborbital frequencies (e.g. meltwater pulse 1a) (Nordfjord *et al.* 2006, 2009; Christensen *et al.* 2013). Therefore, a continuous late Pleistocene record cannot be obtained at one location beneath most of the shelf. The exception is a thick accumulation beneath the outermost continental shelf and upper slope, particularly in the region of the Hudson Apron to the west of the Hudson Shelf Canyon (Mountain *et al.* 2007), as discussed below.

Two groups have worked extensively on Pleistocene sequences beneath the New Jersey shelf. The first emphasized the inner–middle shelf (Ashley *et al.* 1991; Carey *et al.* 1998, 2005; Sheridan *et al.* 2000; Wright *et al.* 2009). The second emphasized the middle–outer shelf wedge and upper slope (Duncan *et al.* 2000; Goff *et al.* 2005; Gulick *et al.* 2005; Nordfjord *et al.* 2006, 2009; Goff & Austin 2009). The latter focused on the last glacial cycle (the interval from the Last Glacial Maximum (LGM) to present) as discussed in the next section.

By using seismic stratigraphy to map upper Pleistocene sequences across the New Jersey continental shelf (Fig. 3.6), Sheridan *et al.* (2000) pieced together a history of late Pleistocene (i.e. since 130 ka) sea-level change. High-resolution seismic reflection profiles (*c.* 1–1.5 m resolution) across the New Jersey continental shelf provide a record of Pleistocene unconformity-bounded sequences (Ashley *et al.* 1991; Carey *et al.* 1998; Sheridan *et al.* 2000 and references therein). Sheridan *et al.* (2000) compiled a

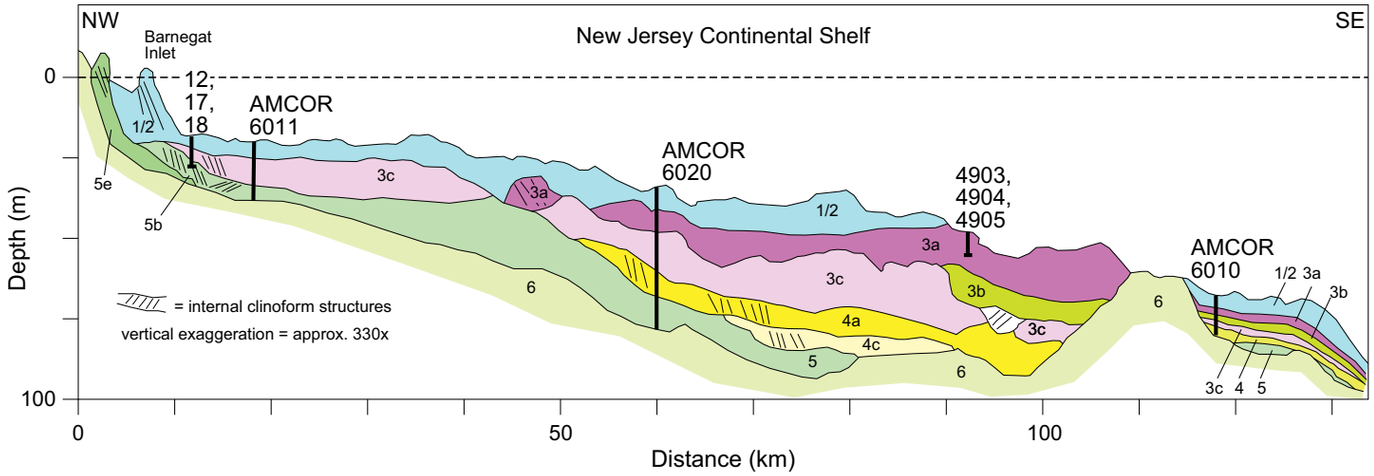

Fig. 3.6. Seismic stratigraphic cross-section of the late Pleistocene off Barnegat Inlet, New Jersey. Numbers 1/2, 3a, 3b, 3c, 4, 5, 5e, 5b and 6 refer to correlations to Marine Isotope Chrons. Shown are AMCOR sites 6011, 6020 and 6010, vibracores 12, 17 and 18 from Uptegrove (2003), and vibracores 4903, 4904 and 4905 from Knebel *et al.* (1979). Modified after Sheridan *et al.* (2000) by Wright *et al.* (2009). The profile is located in Figure 3.1.

composite seismic profile from Barnegat Inlet, New Jersey 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. 3.6). Incision by the palaeo-Hudson River preserved thick MIC 3 and 4 deposits (up to 30 m) in the shelf valley that crossed the shelf at various places, whereas increased sediment discharge by the Hudson River is inferred for MIC 1 and 5 (Sheridan *et al.* 2000).

Although seismic profiles constrain the Pleistocene physical stratigraphy on the continental shelf, age control is difficult. Wright *et al.* (2009) reviewed age control on the Pleistocene sequence delineated previously (Carey *et al.* 1998; Sheridan *et al.* 2000) using radiocarbon dates, amino acid racemization data and superposition; they constrained the ages of large (20–80 m) sea-level falls, and correlated them with MIC 2, 3b, 4, 5b and 6 (the past 130 kyr: Fig. 3.7). They noted that, despite the proximity of New Jersey to the Laurentide ice sheet, sea-level records for MIC 1, 2, 4, 5e and 6 are similar to those reported from New Guinea, Barbados and the Red Sea (Fig. 3.7), with some differences among records for MIC 3. The New Jersey record consistently provides the shallowest sea-level estimates for MIC 3 (*c.* 25–60 m below present: Wright *et al.* 2009), with a barrier system migrating to within about 1 km of the modern barrier (Ashley *et al.* 1991). This difference may be due to a GIA (Potter & Lambeck 2003) effect. Otherwise, the New Jersey record approximates the global record (Fig. 3.7). This can be explained by the fact that the portion of New Jersey directly influenced by the peripheral bulge of the ice sheet was north of Island Beach. A GIA correction must be applied to correct the New Jersey Pleistocene estimate for its far-field response (e.g. Peltier 1998) but this correction of 5–10 m is within the errors of the Pleistocene relative sea-level estimates.

Large volumes of sediment were deposited on the outer shelf and slope during the mid–late Pleistocene (post-750 ka),

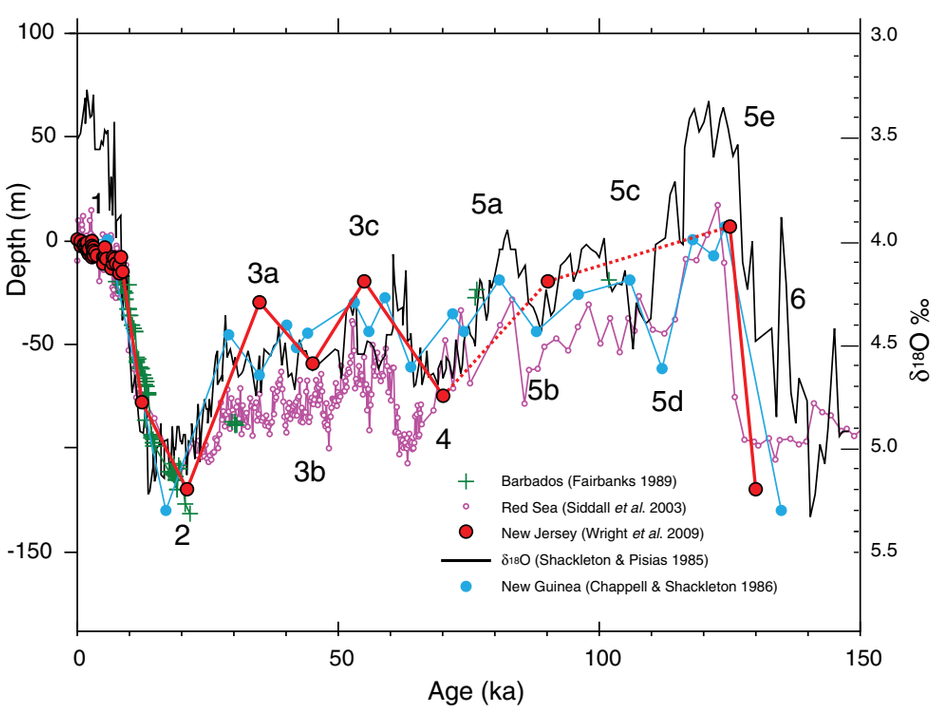

Fig. 3.7. Comparison of the sea-level record from the US middle Atlantic margin with the Huon New Guinea terraces (Chappell & Shackleton 1986), Barbados (Fairbanks 1989; Bard *et al.* 1996), the Red Sea record of Siddall *et al.* (2003) and the benthic foraminiferal $\delta^{18}O$ record from Pacific (Carnegie Ridge) core V19–30 (Shackleton & Pisias 1985: black line). After Wright *et al.* (2009).

extending the continental margin several tens of kilometres past the pre-Quaternary shelf edge (Mountain *et al.* 2007). Here we focus on erosion by submarine canyons and the rapid sedimentation that often occurs in their interfluves.

Submarine canyons off New Jersey are impressive features (e.g. Hudson Canyon is one of the largest in the world) that have a long history of study in this region (Shepard 1934; Daly 1936) and controversy on their origin. Several classes of submarine canyons cut the slope: (1) V-shaped large submarine canyons (Hudson and Wilmington: Fig. 3.2) cut deeply (*c.* 1000 m) into the slope, and much of the way across the shelf and across the continental rise; (2) smaller submarine canyons (Toms and Lindenkohl: Fig. 3.2) cut less deeply into the slope (*c.* 400 m) and breach the shelf break but cannot be traced across it; (3) U-shaped canyons are restricted to the lower slope; and (4) submarine rills and gullies are smaller features on the slope. Pratson *et al.* (2007) reviewed the processes that form these largely erosional features, caused by processes including turbidity currents and intraslope failure due to spring sapping, slides, slumps, and structural control. V-shaped canyons are cut largely by turbidity currents; U-shaped canyons are formed by other submarine processes (Mountain *et al.* 2007). Proximity to modern river systems suggests a link to sediment that most probably supplied these Pleistocene turbidity currents. Pratson *et al.* (2007) noted that submarine canyons may form during times of sea-level lowering, but also noted that intraslope processes unrelated to sea-level, including canyon piracy and intraslope failures, are potentially important in forming slope canyons.

Despite often being assumed to be Pleistocene in age, submarine canyons have a very long history. The modern Hudson River, for example, is apparently structurally controlled by the earliest Jurassic Palisades Sill, and an ancestral Hudson River has delivered large amounts of sediment to the rise near its current location since the Early Jurassic (Poag & Sevon 1989). However, the palaeo-Hudson River migrated south of its position during the Pleistocene as discussed above (Carey *et al.* 2005), and the lower slope portion of the Hudson Canyon can be dated to a seismic reflector thought to mark the approximate beginning of the growth and decay of large northern hemisphere ice sheets (*c.* 2.7 Ma: Mountain & Tucholke 1985). Buried Miocene canyons occur in the region and are likely antecedents for the Plio-Pleistocene Hudson Canyon (Miller *et al.* 1987); the evolution of a buried Miocene canyon is discussed by Mountain *et al.* (2007).

Mountain *et al.* (2007) provided an overview of Pleistocene outer shelf and slope sedimentation. Pleistocene sections at ODP sites 903 (444 m water depth near Berkeley Canyon) and 1073 (639 m water depth, Hudson Apron) recovered over 350 and 500 m, respectively, of sediment assigned to the Bruhnes Chron (last 780 kyr), with sedimentation rates greater than 60 m Ma^{-1} (Christensen *et al.* 1996; Austin *et al.* 1998; McHugh & Olson 2002). Physical properties data (e.g. magnetic susceptibility) integrated with biostratigraphy provide an astronomical chronology for these sites; this chronology shows that deposition of primarily silty clay was largely continuous, although it was punctuated by a few short hiatuses (Christensen *et al.* 1996; McHugh & Olson 2002). Mountain *et al.* (2007) mapped four Pleistocene seismic sequences on the outer shelf and upper slope on the Hudson Apron, and dated them on the slope: (1) sequence p4/yellow is early Bruhnes (MIC 19–12); (2) sequence p3/green is mid Bruhnes (MIC 12–9); (3) sequence p2/blue correlates with MIC 8; and (4) sequence p1/purple is the last interglacial to Holocene (MIC 5e-1). It appears that the sequence boundaries on the slope are generally associated with major glacials (MIC 6, 8, 12 and 20), although sedimentation was continuous across several glacial cycles, and, where there appears to be a cause–effect association, the phase relationships and the link to glacioeustasy are not simple. These sites bear testimony to the dominance of hemipelagic sedimentation between canyon thalwegs on the slope. Within canyon thalwegs and adjacent to them (particularly on the right-hand sides looking down-canyon), downslope deposition dominates.

The last deglaciation

Studies of the upper Pleistocene–Holocene on the inner continental shelf of New Jersey (Ashley *et al.* 1991; Miller *et al.* 2009) and Delaware (Ramsey & Baxter, 1996) have provided a record of sea-level change since the Last Glacial Maximum (LGM; *c.* 20–26 ka). A major erosional surface has been mapped on the inner shelf (R1 of Ashley *et al.* 1991); the erosional event occurred prior to the LGM, probably in MIC 4. During the LGM, a widespread unconformity was eroded beneath the shelf (R2 of Ashley *et al.* 1991). The relationship of the R reflector (Gulick *et al.* 2005) on the outer continental shelf is controversial and there are two interpretations: (1) an approximately 40 ka lowstand surface (Gulick *et al.* 2005; Goff & Austin 2009); and (2) correlation with MIC 6 (Sheridan *et al.* 2000) (Fig. 3.6). It is clear that this reflector is not the LGM as originally interpreted (Milliman *et al.* 1990) but is, instead, an older erosional surface.

Recent studies of the middle–outer shelf Pleistocene document that the surface stratigraphy of the last glacial cycle is more complicated than previously thought (see the summary in Mountain *et al.* 2007). The stratigraphy above the R reflector (=MIC 6 of Sheridan *et al.* 2000; =40 ka and Heinrich event 4 of Goff & Austin 2009) (Fig. 3.6) consists of the following upsection (Duncan *et al.* 2000; Nordfjord *et al.* 2006, 2009): (1) an outer shelf wedge; (2) a dendritic 'channels' reflector interpreted as fluvial channels formed during the LGM, (3) marine fill of the channels reflector during the early Holocene; and (4) a reflector formed as a transgressive ravinement surface. Detailed mapping of the shelf wedge suggest that it consists of subaqueous delta deposits, whereas the sediments above the ravinement reflector consist of lagoonal/back barrier and tidal channel deposits (Nordfjord *et al.* 2009).

The deglaciation is poorly sampled on the continental shelf. Dillon & Oldale (1978) reported dates of approximately 21 ka and a depth of 120 m for the LGM in cores from the outer shelf of New Jersey (Fig. 3.7), very similar to the estimate of 120 m from Barbados (Fairbanks 1989) that has been modelled as approximately 127 m of eustatic lowering (Peltier & Fairbanks 2006). During the MIC 2 sea-level fall, incision and reworking dominated (Goff *et al.* 2005; Christensen *et al.* 2013). Sea level rose rapidly during the last deglaciation and the timing of infilling of the channels (*c.* 16–14 ka) is consistent with the timing of meltwater pulse 1a (Christensen *et al.* 2013). Infilling occurred rapidly (0.5–1 cm a^{-1}) but in agreement with modern estuarine rates of sedimentation (Christensen *et al.* 2013). By 8.8 ka, sea level rose to a point 12 m below modern sea level, creating a thin (3–4 m) barrier system on top of an erosional transgressive ravinement surface (R3 of Ashley *et al.* 1991). The shore-attached ridges were deposited during the sea-level rise younger than approximately 9 ka, and are now reshaped and redeposited by modern currents. The present-day barrier and lagoon complex postdates 8.8 ka (Ashley *et al.* 1991). Holocene middle–outer shelf sediments are reworked, and dates on middle–outer shelf shell material are not in stratigraphic succession, suggesting the reworking of glacial sediments (Alexander *et al.* 2003; Christensen *et al.* 2013).

The record of late Holocene sea-level rise has been complicated by the fact that the rise of approximately 9 m over the past 5 kyr is small relative to uncertainties in the method. Psuty (1986) interpreted a slowing of the rate of rise at approximately 2 ka in New Jersey. Detailed evaluation of sea-level rise in New Jersey during the Holocene was performed by evaluating the 'indicative meaning', evaluating numerous uncertainties and identifying the most reliable points (circled points in Fig. 3.8). During the Holocene, sea level rose moderately rapidly from approximately 8 to

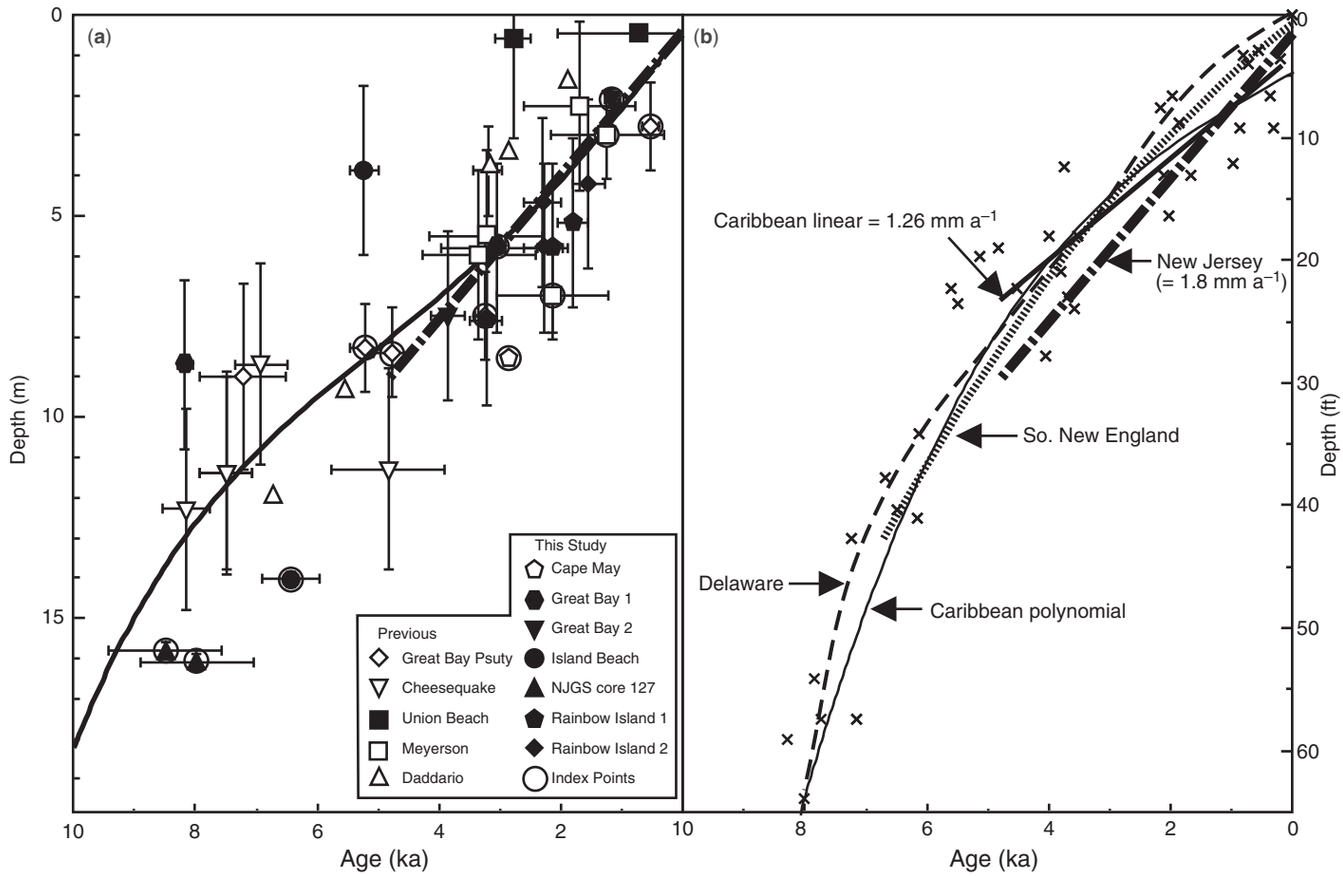

Fig. 3.8. (**a**) Comparison of all dates for New Jersey localities with error bars; data are compiled in Miller *et al.* (2009). The linear regression of 1.8 mm a^{-1} (thick dashed line) is fit to all of the data as two segments: 8–0 and 5–0 ka BP. The thick black line is a polynomial fit to the data. Note that the different regressions yield essentially the same record from 5 to 0 ka BP. Data points with no apparent depth errors are from marsh deposits with 30 cm vertical errors. (**b**) Comparison of our sea-level data and regression (red) with the sea-level record of Fairbanks (1989: crosses), which is based on the western Atlantic reef data of Lighty *et al.* (1982) for ages of less than about 6.4 ka BP. Two regressions through the reef data are shown: the first is a third-order polynomial; the other is a linear regression for all Lighty *et al.* (1982) data younger than 5.5 ka BP. The Delaware (dashed: Ramsey & Baxter 1996) and Southern New England (stippled: Donnelly *et al.* 2005) sea-level records are shown for comparison. Also shown is the record of the past 2 ka from Kemp *et al.* (2011). Modified after Miller *et al.* (2009).

5 ka (Fig. 3.8). The New Jersey record only requires a rise of about 8 m in this period (2.7 m ka^{-1} = mm a^{-1}) but the more complete record from Delaware (Ramsey & Baxter 1996) suggests upwards of 11–12 m (4 mm a^{-1}: Fig. 3.8). The rate of rise slowed around 5 ka, and has averaged 1.8 mm a^{-1} in both New Jersey and Delaware over the past 5 kyr (Miller *et al.* 2009). Much of this rise is due to GIA subsidence. Miller *et al.* (2009) assumed that the current GIA subsidence of 1 mm a^{-1} could be applied to a linear rise in sea level over the past 5 kyr, and concluded that the global rise in sea level was 0.75 ± 0.5 mm a^{-1}. However, modelling of Pacific island records requires a minimal global rise in sea level over the past 2–3 kyr (Peltier *et al.* 2002). Kemp *et al.* (2011) noted minimal eustatic change over the past 2 kyr from detailed studies in North Carolina compared to other less constrained records from throughout the world (Fig. 3.8). Miller *et al.* (2013) revisited the New Jersey sea-level record and documented a 1.4–1.6 mm a^{-1} rise from 2 ka to 1800 Common Era.

Instrument scale, present and future sea-level changes

Twentieth-century tide gauge data from the US middle Atlantic margin reveal a regional sea-level rise of approximately 3 mm a^{-1}, with Atlantic City and Sandy Hook yielding higher rates (*c.* 4 mm a^{-1}) due to compaction (Psuty 1986; Miller *et al.* 2009). This is consistent with the global increase of 1.8 mm a^{-1} in the twentieth century derived from tide gauge data (Church & White 2006) after the GIA subsidence effects of 1.3 mm a^{-1} of subsidence are accounted for (Miller *et al.* 2013). It is clear that the rates of sea-level rise are accelerating in the twenty-first century; global sea level is rising at a rate of 3.3 mm a^{-1} in 2010 CE (Cazenave & Llovel 2010) and accelerating at a rate that intersects 80 cm of rise by 2100 CE (Rahmstorf 2007). Semi-empirical predictions of future sea-level rise (e.g. Vermeer & Rahmstorf 2009) are even higher (1.2 ± 0.6 m by 2100 CE). The US middle Atlantic margin will continue to subside due to GIA effects at 1 mm a^{-1} plus local compaction effects, bringing a minimum rise of about 1 m to this region by 2100 CE. The modern coastal environment is generally starved of sediments in this region, exacerbating the effects of a modern sea-level rise of 3 mm a^{-1} together with 1–2 mm a^{-1} of regional and local subsidence. With a 1 m rise in relative sea level by 2100, the '100 year flood' mark of 2.9 m for much of the New Jersey shoreline will be breached annually by storm surges flooding major airports and highways in the region. Beach erosion will be increasingly common and severe; following the Bruun rule, a 1 m rise in sea level would erode the shoreline by 50–100 m (Kyper & Sorensen 1985), with marsh rollback of 1 km given a 1:1000 gradient. We conclude that this region will be severely impacted by sea-level changes in the twenty-first century, affecting a coastal population of over 40 million people and an economic engine of over $50 billion annually.

We thank F. Pazzaglia, G. Ashley and an anonymous reviewer for their reviews, R. Peltier for discussions of GIA effects, and J. Goff and C. Fulthorpe for input on outer shelf sedimentation and Pleistocene stratigraphy. This overview is the product of a long history of study of this margin, and we thank R.K. Olsson, C.W. Poag and the late J. Owens for their pioneering efforts and inspiration, and NSF and ODP/IODP for funding seismic profiling and drilling.

References

Alexander, C., Sommerfield, C. *et al.* 2003. Sedimentology and age control of late Quaternary New Jersey shelf deposits. *In*: *Transactions of the American Geophysical Union, Fall Meeting 2003*. American Geophysical Union, Washington, DC, Abstract OS52B-0909.

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

Austin, J. A., Jr., Christie-Blick, N., Malone, M., Mountain, G. S. the Leg 174A Shipboard Party. 1998. Miocene to Pleistocene sand-rich sequences and sea-level changes at the New Jersey outer shelf: results from Leg 174A. *JOIDES Journal*, **24**, 4–6, 19.

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.

Bartek, L. R., Vail, P. R., Anderson, J. B., Emmet, P. A. & Wu, S. 1991. Effect of Cenozoic ice sheet fluctuations in Antarctica on the stratigraphic signature of the Neogene. *Journal of Geophysical Research*, **96**, 6753–6778.

Beardsley, R. C. & Boicourt, W. C. 1981. On estuarine and continental shelf circulation in the Middle Atlantic Bight. *In*: Warren, B. A. & Wunch, C. (eds) *Evolution of Physical Oceanography*. MIT Press, Cambridge, MA, 198–233.

Browning, J. V., Miller, K. G., Sugarman, P. J., Kominz, M. A., McLaughlin, P. P. & Kulpecz, A. A. 2008. 100 Myr record of sequences, sedimentary facies and sea-level change from Ocean Drilling Program onshore coreholes, U.S. Mid-Atlantic coastal plain. *Basin Research*, **20**, 227–248.

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

Carey, J. S., Sheridan, R. E., Ashley, G. M. & Uptegrove, J. 2005. Glacially-influenced late Pleistocene stratigraphy of a passive margin: New Jersey's record of the North American ice sheet. *Marine Geology*, **218**, 155–173.

Cazenave, A. & Llovel, W. 2010. Contemporary sea level rise. *Annual Review of Marine Sciences*, **2**, 145–173.

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

Christensen, B. A., Hoppie, B. W., Thunell, R. C., Miller, K. G. & Burckle, L. 1996. Pleistocene age models. *In*: Mountain, G. S., Miller, K. G. *et al.* (eds) *Proceedings of the ODP, Scientific Results*, **150**. Ocean Drilling Program, College Station, TX.

Christensen, B. A., Alexander, C., Goff, J. A., Turner, R. J. & Austin, J. A. 2013. The Last Glacial: Insights from continuous coring on the New Jersey continental shelf. *Marine Geology*, **335**, 78–99.

Christie-Blick, N., Austin, J. A., Jr & Malone, M. J. (eds) 2003. *Proceedings of the ODP, Scientific Results*, **174A**. Ocean Drilling Program, College Station, TX.

Church, J. A. & White, N. J. 2006. A 20th century acceleration in global sea-level rise. *Geophysical Research Letters*, **33**, L1602.

Daly, R. A. 1936. Origin of submarine canyons. *American Journal of Science*, **31**, 401–420.

Dillon, W. P. & Oldale, R. N. 1978. Late Quaternary sea-level curve: reinterpretation based on glacio-tectonic influence. *Geology*, **6**, 56–60.

Donnelly, J. P., Cleary, P., Newby, P. & Ettinger, R. 2005. Coupling instrumental and geological records of sea-level: evidence from southern New England of an increase in sea-level rise in the late 19th century. *Geophysical Research Letters*, **31**, L05203.

Doyle, L. J., Pilkey, O. H. & Woo, C. C. 1979. Sedimentation on the eastern United States continental slope. *In*: Doyle, L. J. & Pilkey, O. H. (eds) *Geology of Continental Slopes*. SEPM Special Publications, Tulsa, **27**, 119–130.

Duncan, C. S., Goff, J. A., Austin, J. A. & Fulthorpe, C. S. 2000. Tracking the last sea level cycle: seafloor morphology and shallow stratigraphy of the latest Quaternary New Jersey middle continental shelf. *Marine Geology*, **170**, 395–421.

Emery, K. O. 1968. Positions of empty pelecypod valves on the continental shelf. *Journal of Sedimentary Petrology*, **38**, 1264–1269.

Fairbanks, R. G. 1982. The origin of continental shelf and slope water in the New York Bight and Gulf of Maine. *Journal of Geophysical Research*, **87**, 5796–5808.

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.

Fulthorpe, C. S. & Austin, J. A., Jr. 2004, Shallowly buried, enigmatic seismic stratigraphy on the New Jersey mid-outer shelf: evidence for latest Pleistocene catastrophic erosion? *Geology*, **32**, 1013–1016.

Glenn, S. M., Jones, C., Twardowski, M., Bowers, L., Kerfoot, J., Webb, D. & Schofield, O. 2008. Glider observations of sediment resuspension in a Middle Atlantic Bight fall transition storm. *Limnology and Oceanography*, **53**, 2180–2196.

Goff, J. A. & Austin, J. A. 2009. Seismic and bathymetric evidence for four different episodes of iceberg scouring on the New Jersey outer shelf: possible correlation to Heinrich events. *Marine Geology*, **266**, 244–254.

Goff, J. A., Swift, D. J. P., Duncan, S. C., Mayer, L. A. & Hughes-Clarke, J. 1999. High-resolution swath sonar investigations of sand ridge, dune and ribbon morphology in the offshore environment of the New Jersey margin. *Marine Geology*, **161**, 307–337.

Goff, J. A., Austin, J. A., Jr *et al.* 2005. Recent and modern marine erosion on the New Jersey outer shelf. *Marine Geology*, **216**, 275–296.

Gong, D., Kohut, J. T. & Glenn, S. M. 2010. Seasonal climatology of wind-driven circulation on the New Jersey Shelf. *Journal of Geophysical Research*, **115**, C04006.

Grow, J. A. & Sheridan, R. E. 1988. U.S. Atlantic continental margin: a typical Atlantic-type or passive continental margin. *In*: Sheridan, R. E. & Grow, J. A. (eds) *The Atlantic Continental Margin*. The Geology of North America, 1–2. Geological Society of America, Boulder, CO, 1–7.

Gulick, S. P. S., Goff, J. A., Austin, J. A., Jr, Alexander, C. R., Jr, Nordfjord, S. & Fulthorpe, C. S. 2005. Basal inflection-controlled shelf-edge wedges off New Jersey track sea-level fall. *Geology*, **33**, 429–432.

Heezen, B. C., Tharp, M. & Ewing, M. 1959. *The Floors of the Oceans*. Geological Society of America, Special Papers, Boulder, **65**.

Heezen, B. C., Hollister, C. D. & Ruddiman, W. F. 1966. Shaping of the continental rise by deep geostrophic contour currents. *Science*, **152**, 502–508.

Hollister, C. D. 1973. *Atlantic Continental Shelf and Slope of the United States – Texture of Surface Sediments from New Jersey to Southern Florida*. United States Geological Survey, Professional Papers, **529-M**.

Huber, B. T., Norris, R. D. & MacLeod, K. G. 2002. Deep-sea paleotemperature record of extreme warmth during the Cretaceous. *Geology*, **30**, 123–126.

Jansa, L. F. 1981. Mesozoic carbonate platforms and banks of the eastern North American margin. *Marine Geology*, **44**, 97–117.

Kemp, A. C., Horton, B. P., Donnelly, J. P., Mann, M. E., Vermeer, M. & Rahmstorf, S. 2011. Climate related sea-level variations over the past two millennia. *Proceedings of the National Academy of Sciences*, **108**, 11 017–11 022.

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.

Kominz, M. A., Browning, J. V., Miller, K. G., Sugarman, P. J., Misintseva, S. & Scotese, C. R. 2008. Late Cretaceous to Miocene

sea-level estimates from the New Jersey and Delaware coastal plain coreholes: an error analysis. *Basin Research*, **20**, 211–226.

Kulpecz, A. A., Miller, K. G., Sugarman, P. J. & Browning, J. V. 2008. Subsurface distribution of Upper Ccretaceous sequences and facies, New Jersey Coastal Plain. *Journal of Sedimentary Research*, **78**, 112–129.

Kyper, T. & Sorensen, R. 1985. Potential impacts of selected sea level rise scenarios on the Beach and Coastal Works at Sea Bright, New Jersey. *In*: Magoon, O. T., Converse, H., Miner, D., Clark, D. & Tobia, L. T. (eds) *Coastal Zone '85*. American Society of Civil Engineers, New York, 2645–2655.

Lavier, L. L., Steckler, M. S. & Brigaud, F. 2001, Climatic and tectonic control on the Cenozoic evolution of the West African margin. *Marine Geology*, **178**, 63–80.

Lighty, R. G., MacIntyre, I. G. & Stuckenrath, R. 1982. *Acropora palmata* reef framework: a reliable indicator of sea level in the western Atlantic for the past 10,000 years. *Coral Reefs*, **1**, 125–130.

Lyne, V. D., Butman, B. & Grant, W. D. 1990. Sediment movement along the U.S. east coast continental shelf: I. Estimates of bottom stress using the Grant-Madsen model and near-bottom wave and current measurements. *Continental Shelf Research*, **13**, 397–428.

McHugh, C. M. G. & Olson, H. C. 2002. Pleistocene chronology of continental margin sedimentation: new insights into traditional models, New Jersey. *Marine Geology*, **185**, 389–411.

McHugh, C. M., Ryan, W. B. F. & Schreiber, B. C. 1993. The role of diagenesis in exfoliation of submarine canyons. *American Association of Petroleum Geologists Bulletin*, **77**, 145–172.

Miller, K. G. & Lohmann, G. P. 1982. Environmental distribution of Recent benthic foraminifera on the northeast United States continental slope. *Geological Society of America Bulletin*, **93**, 200–206.

Miller, K. G., Melillo, A. J., Mountain, G. S., Farre, J. A. & Poag, C. W. 1987. Middle to late Miocene canyon cutting on the New Jersey continental slope: biostratigraphic and seismic stratigraphic evidence. *Geology*, **15**, 509–512.

Miller, K. G., Browning, J. V., Pekar, S. F. & Sugarman, P. J. 1997. Cenozoic evolution of the New Jersey coastal plain: changes in sea level, tectonics, and sediment supply. *In*: Miller, K. G. & Snyder, S. W. (eds) *Proceedings of the ODP, Scientific Results*, **150X**. Ocean Drilling Program, College Station, TX, 361–373.

Miller, K. G., Kominz, M. A. *et al.* 2005. The Phanerozoic record of global sea-level change. *Science*, **310**, 1293–1298.

Miller, K. G., Sugarman, P. J. *et al.* 2009. Sea-level rise in New Jersey over the past 5000 years: implications to anthropogenic changes. *Global and Planetary Change*, **66**, 10–18.

Miller, K. G., Kopp, R. E., Horton, B. P., Browning, J. V. & Kemp, A. C. 2013. A geological perspective on sea-level rise and impacts along the U.S. mid-Atlantic coast. *Earth's Future*, **1**, 3–18.

Milliman, J. D., Jiezao, Z., Anchun, L. & Ewing, J. I. 1990. Late Quaternary sedimentation on the outer and middle New Jersey continental shelf: results of local deglaciation. *Journal of Geology*, **98**, 966–976.

Monteverde, D. H., Mountain, G. S. & Miller, K. G. 2008. Early Miocene sequence development across the New Jersey margin. *Basin Research*, **20**, 249–267.

Mountain, G. S. & Tucholke, B. E. 1985. Mesozoic and Cenozoic geology of the U.S. Atlantic continental slope and rise. *In*: Poag, C. W. (ed.) *Geologic Evolution of the United States Atlantic Margin*. Van Nostrand Reinhold, New York, 293–341.

Mountain, G. S., Miller, K. G. & Blum, P. 1994. *Proceedings of the Ocean Drilling Program, Initial reports*, **150**. Ocean Drilling Program, College Station, TX.

Mountain, G. S., Burger, R. L. *et al.* 2007. The long-term stratigraphic record on continental margins. *In*: Nittrouer, C. A., Austin, J. A., Jr, Field, M. E., Kravitz, J. H., Syvitski, J. P. M. & Wiberg, P. L. (eds) *Continental-Margin Sedimentation: From Sediment Transport To Sequence Stratigraphy*. International Association of Sedimentologists, Special Publications, **37**, 381–458.

Mountain, G., Proust, J.-N., McInroy, D., Cotterill, C. the Expedition 313 Scientists 2010. *Proceedings of the International Ocean Drilling Program, Expedition 313, Initial Report*. Ocean Drilling Program, College Station, TX.

Nordfjord, S., Goff, J. A., Austin, J. A., Jr. & Gulick, S. P. S. 2006. Seismic facies of incised valley-fills, New Jersey continental shelf: implications for erosion and preservation processes acting during late Pleistocene/Holocene transgression. *Journal of Sedimentary Research*, **76**, 1284–1303.

Nordfjord, S., Goff, J. A., Duncan, L. S. & Austin, J. A., Jr. 2009. Shallow stratigraphy and transgressive ravinement on the New Jersey shelf: implications for sedimentary lobe deposition and latest Pleistocene-Holocene sea level history. *Marine Geology*, **266**, 232–243.

Olsson, R. K. 1975. *Upper Cretaceous and Lower Tertiary Stratigraphy, New Jersey Coastal Plain. Second Annual Field Trip Guidebook*. Petroleum Exploration Society, New York.

Olsson, R. K., Melillo, A. J. & Schreiber, B. L. 1987. Miocene sea level events in the Maryland coastal plain and the offshore Baltimore Canyon trough. *In*: Ross, C. & Haman, D. (eds) *Timing and Depositional History of Eustatic Sequences: Constraints on Seismic Stratigraphy*. Cushman Foundation for Foraminiferal Research, Special Publications, **24**, 85–97.

Owens, J. P. & Sohl, N. F. 1969. Shelf and deltaic paleoenvironments in the Cretaceous-Tertiary formations of the New Jersey Coastal Plain. *In*: Subitsky, S. (ed.) *Geology of Selected Areas in New Jersey and Eastern Pennsylvania and Guidebook of Excursions*. Rutgers University Press, New Brunswick, NJ, 390–408.

Pazzaglia, F. J. & Gardner, T. W. 1994. Late Cenozoic flexural deformation of the middle U.S. Atlantic passive margin. *Journal of Geophysical Research*, **99**, 12 143–12 157.

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.

Peltier, W. R., Shennan, I., Drummond, R. & Horton, B. 2002. On the postglacial isostatic adjustment of the British Isles and the shallow viscoelastic structure of the Earth. *Geophysical Journal International*, **148**, 443–475.

Poag, C. W. 1985. Depositional history and stratigraphic reference section for central Baltimore Canyon trough. *In*: Poag, C. W. (ed.) *Geologic Evolution of the United States Atlantic Margin*. Van Nostrand Reinhold, New York, 217–263.

Poag, C. W. & Sevon, W. D. 1989. A record of Appalachian denudation in postrift Mesozoic and Cenozoic sedimentary deposits of the U.S. middle Atlantic margin. *Geomorphology*, **2**, 119–157.

Posamentier, H. W., Jervey, M. T. & Vail, P. R. 1988. Eustatic controls on clastic deposition I – Conceptual framework. *In*: Wilgus, C. K., Hastings, B. S., Kendall, C. G. St. C., Posamentier, H. W., Ross, C. A. & Van Wagoner, J. C. (eds) *Sea-Level Changes: An Integrated Approach*. SEPM Special Publications, Tulsa, **42**, 109–124.

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.

Pratson, L. F., Hutton, E. W. H., Kettner, A. J., Syvitski, J. P. M., Hill, P. S., George, D. A. & Milligan, T. G. 2007. The impact of floods and storms on the acoustic reflectivity of the inner continental shelf: a modeling assessment. *Continental Shelf Research*, **27**, 542–559.

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

Rahmstorf, S. 2007. A semi-empirical approach to projecting future sea-level rise. *Science*, **315**, 368–370.

Ramsey, K. W. & Baxter, S. J. 1996. *Radiocarbon Dates from Delaware: A Compilation*. Delaware Geological Survey, Report of Investigations, **54**.

Reynolds, D. J., Steckler, M. S. & Coakley, B. J. 1991. The role of the sediment load in sequence stratigraphy: the influence of flexural isostasy and compaction. *Journal of Geophysical Research*, **96**, 6931–6949.

Rine, J. M., Tillman, R. W., Culver, S. J. & Swift, D. J. P. 1991. Generation of Late Holocene ridges on the middle continental shelf of New Jersey, USA – evidence for formation in a mid-shelf setting based upon comparison with a nearshore ridge. *In*: Swift, D. J. P., Oertel, G. F., Tillman, R. W. & Thorne, J. A. (eds) *Shelf Sand*

and Sandstone Bodies; Geometry, Facies and Sequence Stratigraphy. International Association of Sedimentologists, Special Publications, **14**, 395–426.

Rowley, D. B. 2013. Sea level: Earth's dominant elevation – Implications for duration and magnitudes of sea level variation. *Journal of Geology*, **121**, 445–454.

Scholle, P. A. 1980. Geological studies of the COST No. B2 well, United States mid-Atlantic continental slope area. *United States Geological Circular*, **833**, 1–132.

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, DC, 303–317.

Shepard, F. P. 1934. Canyons off the New England coast. *American Journal of Science*, **27**, 24–36.

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, C., Meischner, D., Schmelzer, I. & Smeed, D. A. 2003. Sea-level fluctuations during the last glacial cycle. *Nature*, **423**, 853–858.

Stanford, S. D., Ashley, G. M. & Brenner, G. J. 2001. Late Cenozoic fluvial stratigraphy of the New Jersey Piedmont: a record of glacioeustasy, planation, and incision on a low-relief passive margin. *Journal of Geology*, **109**, 265–276.

Steckler, M . S., Mountain, G. S., Miller, K. G. & Christie-Blick, N. 1999. Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping. *Marine Geology*, **154**, 399–420.

Stubblefield, W. L., Kersey, D. G. & McGrail, D. W. 1983. Development of middle continental shelf sand ridges: New Jersey. *American Association of Petroleum Geologists Bulletin*, **67**, 817–830.

Styles, R. & Glenn, S. M. 2005. Long-term sediment mobilization at LEO-15. *Journal of Geophysical Research*, **110**, C04S90.

Sugarman, P. J., Miller, K. G., Owens, J. P. & Feigenson, M. D. 1993. Strontium-isotope and sequence stratigraphy of the Miocene Kirkwood Formation, southern New Jersey. *Geological Society of America Bulletin*, **105**, 423–436.

Sugarman, P. J., Miller, K. G., Bukry, D. & Feigenson, M. D. 1995. Uppermost Campanian–Maestrichtian strontium isotopic, biostratigraphic, and sequence stratigraphic framework of the New Jersey Coastal Plain. *Geological Society of America Bulletin*, **107**, 19–37.

Swift, D. J. P. 1969. Inner shelf sedimentation: process and products. *In*: Stanley, D. J. (ed.) *The New Concepts of Continental Margin Sedimentation: Application to the Geological Record*. American Geological Institute, Washington, DC, DS-5-1–DS-5-26.

Swift, D. J. P., Stanley, D. J. & Curray, J. C. 1971. Relict sediments, a reconsideration. *Journal of Geology*, **79**, 322–346.

Swift, D. J. P., Duane, D. B. & McKinney, T. F. 1973. Ridge and swale topography of the middle Atlantic Bight, North America: secular response to the Holocene hydraulic regime. *Marine Geology*, **15**, 227–247.

Talwani, M., Ewing, J., Sheridan, R. E., Holbrook, W. S. & Glover, L., III . 1995. The EDGE experiment and the U.S. East Coast Magnetic Anomaly. *In*: Banda, E., Talwani, M. & Torne, M. (eds) *Rifted Ocean–Continent Boundary, Volcanic Margin Concepts*. NATO ASI Series, **5E**, 1–26.

Uptegrove, J. 2003. *Late Pleistocene facies distribution, sea level changes and paleogeography of the inner continental shelf off Long Beach Island, New Jersey*. MS thesis, Rutgers University, USA.

Vail, P. R., Mitchum, R. M., Jr & Thompson, S., III. 1977. Seismic stratigraphy and global changes of sea level, part 4: global cycles of relative changes of sea level. *In*: Payton, C. E. (ed.) *Seismic Stratigraphy – Applications to Hydrocarbon Exploration*. American Association of Petroleum Geologists, Memoirs, **26**, 83–9.

Vermeer, M. & Rahmstorf, S. 2009. Global sea level linked to global temperature. *Proceedings of the National Academy of Science of the United States of America*, **106**, 21 527–21 532.

Watts, A. B. & Steckler, M. S. 1979. Subsidence and eustasy at the continental margin of eastern North America. *In*: Talwani, M., Hay, W. W. & Ryan, W. B. F. (eds) *Deep Drilling Results in the Atlantic Ocean: Continental Margins and Paleoenvironments*. American Geophysical Union, Maurice Ewing Symposium, **3**, 218–234.

Withjack, M. O., Schlische, R. W. & Olsen, P. E. 1998. Diachronous rifting, drifting, and inversion on the passive margin of central eastern North America: an analog for other passive margins. *American Association of Petroleum Geologists Bulletin*, **82**, 817–835.

Wright, J. D., Sheridan, R. E., Miller, K. G., Uptegrove, J., Cramer, B. S. & Browning, J. V. 2009. Late Pleistocene sea level on the New Jersey Margin: implications to eustasy and deep-sea temperature. *Global and Planetary Change*, **66**, 93–99.