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Late Cretaceous and Cenozoic sea-level estimates: backstripping analysis of borehole data, onshore New Jersey

William A. Van Sickel,*[1] Michelle A. Kominz,* Kenneth G. Miller† and James V. Browning†

*Department of Geosciences, Western Michigan University, Kalamazoo, MI, USA
†Department of Geological Sciences, Rutgers University, Piscataway, NJ, USA

ABSTRACT

Backstripping analysis of the Bass River and Ancora boreholes from the New Jersey coastal plain (Ocean Drilling Project Leg 174AX) provides new Late Cretaceous sea-level estimates and corroborates previously published Cenozoic sea-level estimates. Compaction histories of all coastal plain boreholes were updated using porosity–depth relationships estimated from New Jersey coastal plain electric logs. The new porosity estimates are considerably lower than those previously calculated at the offshore Cost B-2 well. Amplitudes and durations of sea-level variations are comparable in sequences that are represented at multiple boreholes, suggesting that the resultant curves are an approximation of regional sea level. Both the amplitudes and durations of third-order (0.5–5 Myr) cycles tend to decrease from the Late Cretaceous to the late Miocene. Third-order sea-level amplitudes in excess of 60 m are not observed. Long-term (10^8–10^7 years) sea level was approximately constant at 30–80 m in the Late Cretaceous, rose to a maximum early Eocene value of approximately 100–140 m, and then fell through the Eocene and Oligocene.

INTRODUCTION

One of the more complex problems involved in unravelling Earth history is determination of the record of global sea-level change (eustasy). Any estimate derived from continental or continental margin data is necessarily a combination of both eustatic and tectonic changes, and is also impacted by local or regional sedimentary processes and their climatic controls (e.g., Sloss, 1963; Bond, 1979; Watts & Steckler, 1979; Posamentier *et al*., 1988). Proxies, such as oxygen isotopes, can be used to estimate the volume of water sequestered as glacial ice (Miller *et al*., 1987), but this signal is overprinted by variations in temperature and, to a lesser extent, salinity (Miller, 2002). Estimates of eustasy from variations in the changing volume of the oceans, including ridge volumes, sediment volumes, large igneous provinces, continent–continent collision and continental extension (e.g., Harrison, 1990; Larson, 1991) are complicated by the fact that subduction removes increasingly large portions of the data as estimates range towards earlier times (e.g., Kominz, 1984).

Sea-level change occurs in a cyclic, but generally not periodic, manner on various time scales (e.g., Vail *et al*., 1977). The magnitude of long-term (50–200-Myr duration) variations has recently become controversial. Estimates for the Late Cretaceous to the Miocene based on changing ocean volumes (e.g., Harrison, 1990) are somewhat higher than estimates based on continental hypsometry (Bond, 1979) and on backstripping (Watts & Steckler, 1979). However, recent analysis of the age distribution of ocean floor suggests that there may have been no change in spreading rates for the last 180 Myr, and thus, no related sea-level change (Rowley, 2002). Although this does not eliminate other contributions to ocean volume change, estimates by Harrison (1990) suggest that the primary control on long-term eustasy was mid-ocean ridge volumes. Backstripping of the Cenozoic section from New Jersey (Kominz *et al*., 1998) indicates a relatively small, but significant, long-term Cenozoic sea-level fall. Because this record did not include the entire Cretaceous sea-level rise, Kominz *et al*. (1998) suggested that the magnitude of long-term sea-level fall was underestimated. The two additional boreholes analysed here were drilled into the Late Cretaceous in part to determine whether backstripping would support the larger- or smaller-magnitude estimates of long-term sea-level change.

Backstripping of the New Jersey coastal plain sediments also allows us to estimate the magnitude of the equally controversial third-order (about a half to three million year duration) sea-level changes. Sea-level estimates based on sequence stratigraphic data (e.g., Haq *et al*., 1987) are controversial both because of the proprietary nature of the data on which the estimates were based and because of limitations of the method (e.g., Miall, 1986, 1992; Christie-Blick & Driscoll, 1995). Finally, sequence stratigraphic

[1] Present address: Polaris Energy Inc., 180 W. Michigan Avenue, Suite 508, Jackson, MI 49201, USA.
Correspondence: Michelle A. Kominz, Deptartment of Geosciences, Western Michigan University, Kalamazoo, MI 49008, USA. E-mail: michelle.kominz@wmich.edu

data suggest that some very large magnitude third-order sea-level changes of 100 m or more occurred (Haq *et al*., 1987). The only known mechanism for such large, rapid eustatic variations is continental glaciation (e.g., Pitman & Golovchenko, 1991). As such, these events are particularly problematic when they occur during greenhouse climates, such as the Late Cretaceous. Thus, this is an important time period in which to determine the magnitude of sea-level change. Backstripping of Cenozoic New Jersey coastal plain strata (Kominz *et al*., 1998) previously indicated that the magnitude of sea-level change was roughly half that estimated by the Haq *et al*. (1987) method. Testing for a eustatic origin requires that estimates of the sea-level magnitude be obtained at multiple locations. The New Jersey boreholes provide overlapping middle Eocene (~45 Ma) through late Miocene (~7 Ma) data. Two new boreholes drilled at Ancora and Bass River, New Jersey provide overlapping data and extend from the Late Cretaceous through middle Miocene. If third-order sea-level variations are consistent within this data set, local tectonism can be eliminated as a mechanism for relative sea-level change.

REGIONAL SETTING

Two new boreholes, Bass River and Ancora (Fig. 1), were drilled by the Ocean Drilling Program (ODP) as Leg 174AX (Miller *et al*., 1998a, 1999a). These sites targeted Cretaceous sections beneath the New Jersey coastal plain.

The New Jersey coastal plain forms the western margin of the Baltimore Canyon Trough, a large sedimentary basin that underlies the continental shelf along the middle Atlantic, east coast of the United States (Grow & Sheridan, 1988). This basin formed as a result of Late Triassic–Early Jurassic rifting between North America and Africa (Grow & Sheridan, 1988). By Cretaceous time, the crust attained sufficient flexural rigidity for offshore sediments that filled the thermally subsiding margin to generate accommodation landward of the actively stretched crust (Watts, 1981; Steckler *et al*., 1999). The resulting coastal plain consists of Lower Cretaceous to Holocene strata that dip gently seaward and thicken down-dip (Olsson *et al*., 1988). Lowstand systems tracts are generally absent on the coastal plain so that sequences represent stacked transgressive and highstand systems tracts (Miller *et al*., 1998b).

BACKSTRIPPING METHOD

Backstripping is a quantitative method of estimating tectonic subsidence, which is defined as the vertical movement of basement in the absence of both sediment loading and sea-level change (Watts & Ryan, 1976). We apply a one-dimensional analysis and assume an Airy isostatic response to loads (e.g., Bond *et al*., 1989). The calculation of basement subsidence in water, R1 (also called first reduction or accommodation), is a modification of the general backstripping equation of Steckler & Watts (1978):

$$R1 = TS + \Delta SL\left(\frac{\rho_a}{\rho_a - \rho_w}\right) = S^*\left(\frac{\rho_a - \rho_{S^*}}{\rho_a - \rho_w}\right) + WD \quad (1)$$

where TS is tectonic subsidence, ΔSL is eustatic sea-level change, S^* is decompacted sediment thickness, WD is paleodepth and ρ_a and ρ_w are the density of the asthenosphere (3.18 g cm^{-3} at 1300 °C) and sea water (1.03 g cm^{-3}), respectively. The first reduction ($R1$) removes the effects of sediment loading and compaction, producing curves that are an approximation of the subsidence that would have occurred if the basin had subsided in water without sediment deposition (Bond *et al*., 1989). $R1$ includes both tectonic subsidence and eustatic sea-level change.

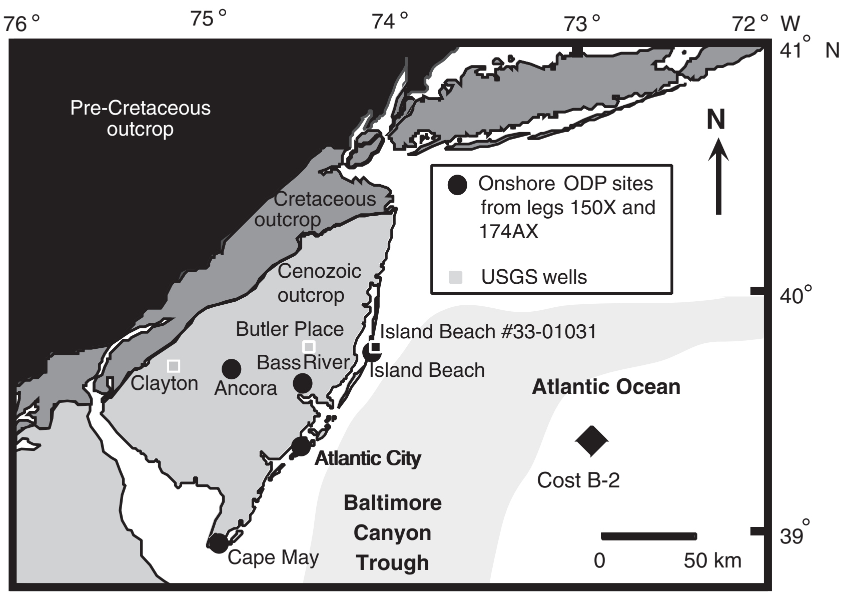

Fig. 1. Location map of New Jersey coastal plain boreholes and wells used in this study, including Ocean Drilling Program Leg 150X (Island Beach, Atlantic City, and Cape May) and Leg 174AX (Bass River and Ancora) sites. The USGS wells (Butler Place and Island Beach #33-01031) were used in conjunction with the Island Beach borehole and the offshore Cost B-2 well to estimate compaction of sand and mud. The USGS Clayton and Island Beach wells were used to estimate thicknesses and lithologies beneath the Bass River and Ancora boreholes. The Baltimore Canyon Trough (grey, off-shore shading) is a Mesozoic–Cenozoic sedimentary basin, which continues to the northeast and to the south beyond the boundaries of this map.

Estimation of S^* requires the assumption of compaction rates. Ideally, compaction is based on porosity estimates from the wells or boreholes in the region of the study, as is the case here.

Seaward of the coastal plain, thermal subsidence and sediment loading dominated accommodation (Steckler *et al.*, 1988). The flexural response of the adjacent, unstretched crust generated sufficient accommodation onshore for sediment accumulation, thus forming the coastal plain. It is this subsidence, a response to offshore loading that forms the 'tectonic' subsidence of the coastal plain. Because the offshore sediment load decreases with time, as the offshore thermal cooling decreases, the coastal plain subsidence, is likewise, exponentially decaying in form. That is, the form of subsidence beneath the New Jersey coastal plain was also thermal (e.g., Steckler & Watts, 1978; Kominz *et al.*, 1998). The best-fit thermal curves (McKenzie, 1978), subtracted from $R1$ curves, produce a second set of curves that are designated as $R2$ (second reduction; Bond *et al.*, 1989).

$$R2_{SL} = (R1 - TS)\left(\frac{\rho_a - \rho_w}{\rho_a}\right). \qquad (2)$$

That is, subsidence because of offshore sediment loading (TS) was removed from $R1$, the subsidence in water. The remaining subsidence, $R2$, includes variations in sea level as well as any non-thermal tectonic or out-of-plane sediment-loading component of $R1$. We use the subscript, SL, to indicate that, in the absence of non-thermal tectonics, this is sea-level change. Similar magnitudes and durations of $R2_{SL}$ among different borehole sites indicate isolation of regional sea level. Large-scale, regional tectonics could be present in this signal.

Thermal subsidence is calibrated to an ocean floor with a thermal decay constant of 36 Myr and an equilibrium plate thickness of 95 km (Stein & Stein, 1992). Best-fit thermal subsidence is calculated by first fitting an exponential curve, with a decay constant of 36 Myr, by linear regression to the $R1$ curve. The best-fit value of the exponential fit at 0 Myr is then used to constrain a best-fit thermal plate model (McKenzie, 1978; Kominz *et al.*, 1998). The thermal subsidence of a stretched plate is a better simulation of the form of 'tectonic' subsidence than is a simple exponential curve. Initial thermal subsidence was assumed to start at 150 Ma. This is younger than the age of formation of the passive margin (Olsson *et al.*, 1988) because the flexural response of the coastal plain was increasingly delayed landward (e.g., Steckler *et al.*, 1988).

POROSITY ESTIMATES FROM SONIC LOGS

The previous study of sea level from boreholes along the New Jersey margin by Kominz *et al.* (1998) adapted sand and mud compaction curves from the outer shelf Cost B-2 well porosities (Rhodehamel, 1977). However, samples of sand and mud porosities in the upper kilometre were limited (Fig. 2). Because most of the boreholes analysed in this study only penetrated ~1 km, and sedimentation rates were much slower on the coastal plain than on the shelf, we have developed new porosity–depth relationships. The new porosity curves are constrained at depth by the offshore data (Fig. 2), but the portions of the curves used to decompact the New Jersey coastal plain boreholes are only constrained by the new coastal plain data.

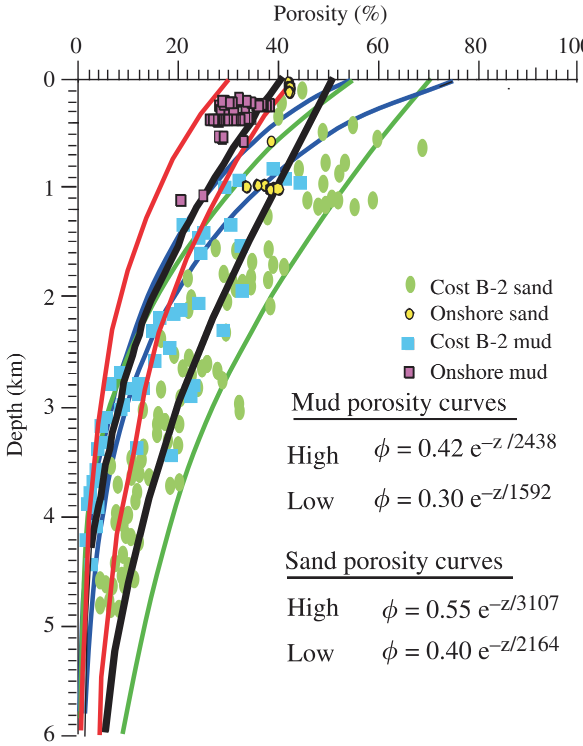

Fig. 2. Cost B-2 sand and mud porosities are compared with estimates from onshore sites (Island Beach borehole, Island Beach #33-01031, and Butler Place wells; see Fig. 1 for well/borehole locations). The thick solid black (sand) and red (mud) lines represent the new high and low, end-member porosity curves used in this study. The blue and green lines are mud and sand curves from the COST B2 well. In the equations, the depths, z, are in metres.

Three wells, the Island Beach borehole (ODP Leg 150X), the Island Beach well #33-01031, and Butler Place well (Fig. 1) were analyzed. Gamma logs were used to define sand and mud intervals using the Schlumberger (1974) method for estimating clay volume (Van Sickel, 2000). Sand was defined as having less than 5% clay by volume and mud as having more than 50% clay by volume. Porosity was estimated from sonic logs for intervals that met these criteria.

Because New Jersey coastal plain sediments are relatively shallowly buried and unconsolidated, we utilized the Raymer–Hunt equation (Raymer *et al.*, 1980) to estimate porosity. This formula, based on empirical data, is applicable to shallow, unconsolidated formations:

$$\text{Sonic porosity} = 63 \times \left[1 - \frac{\Delta t_{ma}}{\Delta t_{log}}\right] \qquad (3)$$

where Δt_{ma} is the interval transit time of the matrix and Δt_{log} the interval transit time of the formation. We assumed a travel time of 168 $\mu s\,m^{-1}$ (Schlumberger, 1972) for sand and 233 $\mu s\,m^{-1}$ (Erickson & Jarrard, 1998) for mud. The method was applied to sonic logs for the three onshore sites (Van Sickel, 2000).

Although the data are insufficient to define the form of the curves (Fig. 2), they were fit to exponential decay curves rather than linear curves because the physics requires decreasing rates of compaction with depth (e.g., Athy, 1930; Baldwin & Butler, 1985). Additionally, the exponential is the form of compaction curves used by the backstripping programme that we employed (Kominz *et al.*, 1998).

Both the lower and upper porosity limits for sand and mud are considerably lower than those used by Kominz *et al.* (1998, Fig. 2). The new curves bring the surface sands to near closest packing, whereas surface porosities are considerably higher than this at the B2 well. The lower and upper limits for surface mud are reduced even more than sand. Previous studies have suggested that one factor affecting compaction is the rate of sedimentation (e.g., Audet & McConnell, 1992). Porosities tend to be higher when sedimentation rates are high (e.g., Schmoker & Gautier, 1989). Sedimentation rates at offshore sites beneath the modern continental shelf are much higher than beneath the coastal plain because of higher accommodation rates. This is consistent with the lower porosities found in the coastal plain data, as compared with the offshore COST B-2 data. Perhaps more crucial in backstripping estimates, the porosities of both sandstones and shales decrease much less rapidly with depth over the top kilometre according to the new curves. This suggests that earlier backstripping results overestimated the amount of accommodation made available by compaction. As a result, tectonic subsidence and/or eustasy may have been underestimated.

BOREHOLE DATABASE

Data for the Cape May, Atlantic City and Island Beach boreholes are from Kominz *et al.* (1998) and updated for the Oligocene following Kominz & Pekar (2001). The more recent Bass River and Ancora boreholes (Fig. 1; Table 1) were drilled as part of ODP Leg 174AX (Miller *et al.*, 1998a, 1999a). These boreholes were continuously cored through Miocene to Upper Cretaceous (Cenomanian) strata. Pleistocene and younger strata are absent at both boreholes with the exception of a thin Holocene layer.

Lithologic descriptions include onsite and subsequent laboratory physical characterization of core samples including relative percentages of quartz sand, silt-clay, glauconite, mica and carbonate components (Miller *et al.*, 1998a, 1999a). Grain densities were calculated from the relative proportions of different lithologies. Sequence bounding unconformities were identified by Miller *et al.* (1998a, 1999a) as paraconformities inferred from biostratigraphic and Sr isotopic breaks and from physical

Table 1. (a) Bass River borehole data (ODP Leg 174AX), (b) Ancora borehole data (ODP Leg 174AX)

Unit	Depth (m)	Depth (ft)	Age (Ma)	Water depth (m)
(a)				
Bass River	596.35	1956.5	93.6	60
Bass River	594.37	1950.0		50
Bass River	588.27	1930.0		40
Bass River	574.55	1885.0		30
Bass River	567.24	1861.0		20
Bass River	560.56	1839.1		20
Bass River	556.27	1825.0		40
Bass River	553.25	1815.1		20
Bass River	550.60	1806.4	92.2	20
Unconformity	550.60	1806.4	91.5	0
Magothy1	542.55	1780.0		0
Magothy1	534.02	1752.0	90.0	0
Unconformity	534.02	1752.0	88.0	0
Magothy2	527.31	1730.0		0
Magothy2	520.97	1709.2	86.8	0
Unconformity	520.97	1709.2	85.2	20
Cheesequake	519.42	1704.1		30
Cheesequake	513.05	1683.2	84.3	10
Unconformity	513.05	1683.2	84.2	100
Merchantville	509.02	1670.0		125
Merchantville	504.45	1655.0		100
Woodbury	479.76	1574.0		50
Woodbury	467.87	1535.0		20
Woodbury	453.55	1488.0		20
LowerEnglish	448.85	1472.6	76.8	20
Unconformity	448.85	1472.6	76.7	20
UpperEnglish	439.07	1440.5	75.9	20
Unconformity	439.07	1440.5	75.8	60
Marshalltown	435.56	1429.0		75
Wenona	423.98	1391.0		65
MtLaurel1	411.18	1349.0		60
MtLaurel2	394.57	1294.5	71.0	40
Unconformity	394.57	1294.5	69.0	80
Navesink	389.39	1277.5		60
NewEgypt	383.66	1258.7	64.5	25
Unconformity	383.66	1258.7	61.7	105
Hornerstown1	382.99	1256.5	61.2	105
Unconformity	382.99	1256.5	58.5	105
Hornerstown2	380.67	1248.9		130
Vincetown1	378.20	1240.8	58.0	105
Unconformity	378.20	1240.8	56.5	105
Vincetown2	372.77	1223.0		125
Vincetown2	355.10	1165.0		140
Vincetown2	347.05	1138.6	55.0	105
Unconformity	347.05	1138.6	54.2	135
Manasquan1	345.83	1134.6	54.0	135
Unconformity	345.83	1134.6	53.5	185
Manasquan2	337.87	1108.5	52.8	155
Unconformity	337.87	1108.5	52.2	155
Manasquan3	312.42	1025.0		125
Manasquan3	309.83	1016.5	50.7	100
Unconformity	309.83	1016.5	50.4	135
Manasquan4	299.10	981.3	49.8	125
Manasquan4	292.58	959.9	49.7	100
Unconformity	292.58	959.9	48.6	135
LowerSharkR1	290.90	954.4	48.4	135

Table 1. (Continued)

Unit	Depth (m)	Depth (ft)	Age (Ma)	Water depth (m)
Unconformity	290.90	954.4	47.3	135
LowerSharkR2	284.08	932.0	47.2	135
Unconformity	284.08	932.0	44.4	135
LowerSharkR3	270.00	885.8	43.4	135
Unconformity	270.00	885.8	41.0	75
UpperSharkR1	263.23	863.6	40.0	75
Unconformity	263.23	863.6	37.0	75
UupperSharkR2	258.05	846.6	36.5	75
Unconformity	258.05	846.6	35.7	125
AbseconInlet1	256.07	840.1	35.6	100
Unconformity	256.07	840.1	35.5	125
AbseconInlet2	234.91	770.7	35.2	100
Unconformity	234.91	770.7	35.1	125
AbseconInlet4	208.76	684.9	34.8	75
Unconformity	208.76	684.9	34.2	100
AbseconInlet3	205.86	675.4	33.8	100
Unconformity	205.86	675.4	32.7	75
AtlanticCity	169.26	555.3	32.1	75
Unconformity	169.26	555.3	21.0	20
Kirkwood	166.73	547.0		30
Kirkwood	157.28	516.0		40
Kirkwood	148.29	486.5		20
Kirkwood	116.74	383.0	20.2	20
Unconformity	116.74	383.0	20.1	20
Kirkwood	100.28	329.0	19.8	0
Unconformity	100.28	329.0	17.5	20
Kirkwood	88.06	288.9		40
Kirkwood	70.84	232.4		20
Kirkwood	62.48	205.0	16.5	20
Unconformity	62.48	205.0	16.1	20
Kirkwood	50.90	167.0		20
Kirkwood	45.54	149.4		20
Kirkwood	42.98	141.0		20
Kirkwood	40.51	132.9	15.6	20
Unconformity	40.51	132.9	12.5	0
Cohansey	37.80	124.0	12.4	0
Unconformity	37.80	124.0	12.3	0
Cohansey	31.06	101.9	11.9	0
Unconformity	31.06	101.9	9.4	0
Cohansey	12.44	40.8	8.8	0
Unconformity	12.44	40.8	8.0	0
Cohansey	6.00	19.7	7.5	0
Unconformity	6.00	19.7	0.011	− 6
CapeMay	0.91	3.0	0.000	− 8.53
(b)				
Potomac	356.62	1170.0	98.5	− 5
Potomac	353.02	1158.2	97.6	− 5
Unconformity	353.02	1158.2	97.5	20
Potomac2	351.90	1154.5	97.3	20
Unconformity	351.90	1154.5	97.2	− 5
Potomac3	349.95	1148.1	97.0	− 5
Unconformity	349.95	1148.1	96.4	50
Bass River	342.90	1125.0		30
Bass River	340.16	1116.0		20
Bass River	338.61	1110.9	95.0	10
Unconformity	338.61	1110.9	94.9	40
Bass River2	329.95	1082.5	93.8	10

Table 1. (Continued)

Unit	Depth (m)	Depth (ft)	Age (Ma)	Water depth (m)
Unconformity	329.95	1082.5	93.7	30
Bass River3	323.85	1062.5	92.8	10
Unconformity	323.85	1062.5	90.0	0
Magothy1	300.84	987.0	88.3	0
Magothy1	295.66	970.0		0
Magothy1	291.82	957.4	86.8	0
Unconformity	291.82	957.4	85.0	25
Cheesequake	288.13	945.3	84.7	20
Unconformity	287.89	944.5	84.5	50
Merchantville	275.24	903.0		70
Woodbury	242.99	797.2		40
LowerEnglish	241.50	792.3	77.5	20
Unconformity	241.50	792.3	77.5	40
UpperEnglish	234.70	770.0		20
UpperEnglish	230.80	757.2	76.0	10
Unconformity	230.80	757.2	75.9	30
Marshalltown	224.03	735.0		50
Wenonah	218.85	718.0		30
MtLaurel	198.52	651.3	71.8	20
Unconformity	198.52	651.3	69.0	60
Navesink+	186.69	612.5	64.5	50
Unconformity	186.69	612.5	63.0	60
Hornerstown1	184.86	606.5	61.0	60
Unconformity	184.86	606.5	60.0	60
Vincetown1	182.58	599.0	57.8	60
Unconformity	182.58	599.0	57.1	60
Vincetown2	171.33	562.1	55.5	45
Unconformity	171.33	562.1	55.5	45
Vincetown3	159.17	522.2	55.0	45
Unconformity	159.17	522.2	54.2	70
Manasquan1	158.25	519.2	54.0	70
Unconformity	158.25	519.2	52.9	185
Manasquan2	140.58	461.2	52.3	155
Unconformity	140.58	461.2	52.2	155
Manasquan3	136.77	448.7	52.0	125
Unconformity	136.77	448.7	49.7	125
LowerSharkR4	136.16	446.7	49.5	125
Unconformity	136.16	446.7	48.6	125
LowerSharkR5	135.03	443.0	48.4	125
Unconformity	135.03	443.0	46.2	100
LowerSharkR6	130.33	427.6	46.1	100
Unconformity	130.33	427.6	44.4	140
LowerSharkR7	118.45	388.6	43.7	110
Unconformity	118.45	388.6	41.0	75
UpperSharkR	109.67	359.8	40.5	50
Unconformity	109.67	359.8	37.2	50
UpperSharkR	80.38	263.7	36.7	50
Unconformity	80.38	263.7	21.2	35
Kirkwood	69.25	227.2	20.2	20
Unconformity	69.25	227.2	20.0	20
Kirkwood	51.18	167.9	19.5	10
Unconformity	51.18	167.9	11.9	5
Kirkwood	8.92	29.3	11.2	0
Unconformity	8.92	29.3	8.8	0
Cohansey	2.74	9.0	8.3	0
Unconformity	2.65	8.7	0.5	0
Surface	0.00	0.0	0.0	0

stratigraphy, including irregular contacts, reworking, bioturbation, major facies changes and γ-ray peaks.

Paleodepth estimates are generally constrained by lithofacies, and/or by benthic foraminiferal biofacies (assemblages). Foraminiferal depth zones are assigned as inner neritic (0–30-m paleodepth), middle neritic (30–100 m) and outer neritic (100–200 m). Higher paleobathymetric resolution is available for the Oligocene sections from two-dimensional paleoslope modelling leading to paleodepth range errors of at least $\pm$ 10 m (inner neritic), $\pm$ 20 m (middle neritic) and $\pm$ 30 m (outer neritic) (Pekar & Kominz, 2001). Relative water depth changes throughout the section, especially within sequences, are also better constrained. For example, paleodepths clearly shoal upsection in sequences above maximum flooding surfaces.

Decompacted sediment thickness is calculated using porosity–depth curves of sand, mud, calcarenite and siltstone. Glauconite is assumed to compact like sand. Carbonates and siltstones are decompacted using the generalized curves of Bond & Kominz (1984).

Biostratigraphic data are tied to the Gradstein *et al.* (1995) Cretaceous time scale and to the Berggren *et al.* (1995) Cenozoic time scale. Ages are assigned to the top and base of each sequence and constrained by integrating magnetostratigraphic, biostratigraphic and Sr isotopic data (Hernandez *et al.*, 2000). Age uncertainties are typically $\pm$ 0.5 Myr for the Cenozoic (Miller *et al.*, 1988) and $\pm$ 1.0 Myr for the Cretaceous (Miller *et al.*, 2003, 2004). Although sedimentation rates were probably variable (Miller, 1997), sedimentation rates between sequence boundaries are assumed to be constant when more precise estimate of ages within sequences is beyond the accuracy of our age control.

Backstripping requires stratigraphic information to basement to account for accommodation because of compaction and to evaluate the amplitude of long-term sea-level change (Kominz *et al.*, 1998). Thicknesses of strata beneath the Bass River and Ancora boreholes are interpolated between those at the Clayton-1 well (Owens *et al.*, 1998) and Island Beach well #33-01031 (Gill *et al.*, 1963). Generalized lithologies of older sediments are taken from the Ancora borehole (Miller *et al.*, 1999a; Olsson *et al.*, 1988).

RESULTS

Effect of porosity ranges on $R1$ and $R2_{SL}$ estimates

The wide range of porosity (Fig. 2) suggests that considerable error could be introduced by inaccuracies in decompaction. To test this, we applied both low and high end-member porosity–depth curves to the Bass River and Ancora boreholes (Fig. 3). The $R1$ curves generated with high end-member porosity–depth curves require a maximum of approximately 50 m more subsidence in both boreholes (Fig. 3a). The variation of $R2_{SL}$ estimates, because of high vs. low end-member porosity is 20 m or less at Bass River and 5 m or less at Ancora (Fig. 3b). $R2_{SL}$ results are less sensitive to porosity assumptions because each curve is fit individually to a theoretical thermal curve and because adjustment for water loading results in a decrease in the magnitudes by a factor of about 0.67 ($= [\rho_a - \rho_w]/\rho_a$; see Eqn (2)). Because the difference is slight, and for simplicity, only the $R2_{SL}$ results obtained from low end-member porosity–depth relations are discussed below.

$R1$ results, Bass River and Ancora boreholes

The $R1$ subsidence curves for Bass River and Ancora are largely consistent with the assumption of thermally controlled tectonic subsidence (Fig. 3a). $R1$ curves younger than 98 Myr are derived from borehole data whereas older portions represent estimated stratigraphy to basement. The gaps in the $R1$ curves represent non-deposition between sequence boundaries. In general, subsidence at Bass River was greater than at Ancora, which is consistent with its location down-dip from the Ancora borehole. As discussed earlier, the thermal form of subsidence was most likely because of the flexural response to sediment loading of the offshore, thermally subsiding, passive margin.

$R2_{SL}$ results, all boreholes

The best-estimate $R2_{SL}$ results of all boreholes are plotted relative to present sea level (Fig. 4). This is possible because the upper portion of the boreholes included Holocene sediments, so that the $\approx$0-Ma $R2_{SL}$ result is set according to the current height of the borehole. Overall, the timing and magnitude of $R2_{SL}$ estimates from Bass River and Ancora are consistent with each other and with estimates from the ODP Leg 150X boreholes (Fig. 4). This suggests that at least a regional sea-level signal has been isolated. There is particularly good agreement for Cretaceous through early Eocene estimates with more variability in the latest Cretaceous. Interestingly, both the duration of sequences and the magnitudes of sea-level falls required by the $R2$ results decrease from the Late Cretaceous to the Neogene. There are several possible explanations for this observation. It is possible that amalgamated sequences in older strata are more difficult to resolve than in the younger strata, resulting in the decrease in cycle duration. The decreasing magnitudes may be a function of the progradational nature of this margin. More of the transgressive and regressive portions of section are preserved in the older sequences. In the Cretaceous the boreholes were located on the middle to inner shelf. Recently the coastal plain has accumulated marine strata only during extreme highstand events. As the sediments accumulated, these coastal plain boreholes came to record less of the sea-level cycle. Thus, the magnitude of sea-level change recorded by $R2_{SL}$ decreases with time, whereas the actual magnitude of sea level may not.

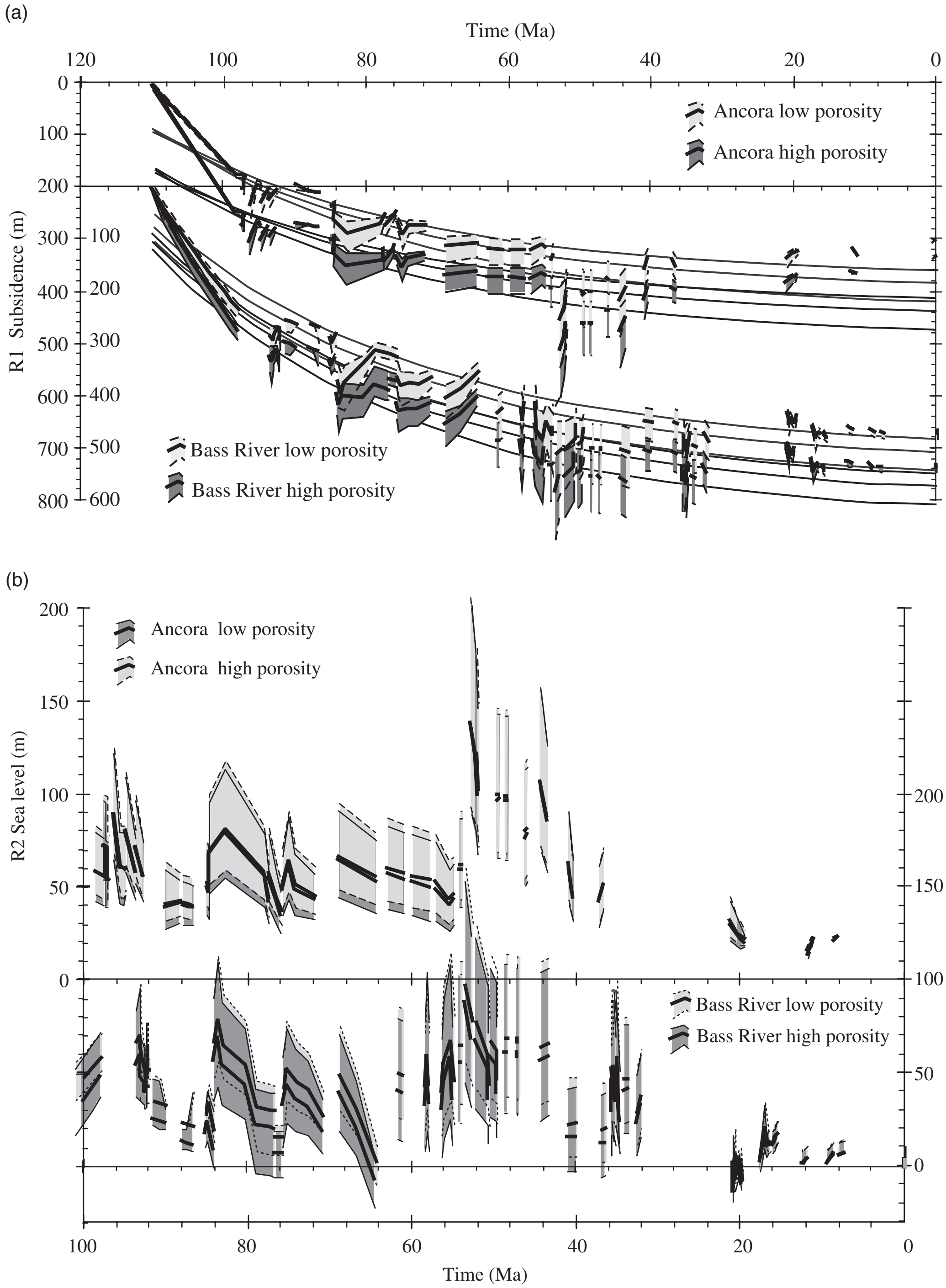

Fig. 3. (a) $R1$ results for Bass River and Ancora boreholes using both high and low end-member porosity–depth curves. Each $R1$ curve is fit to a thermal, tectonic subsidence curve (solid lines). $R1$ curves older than 98 Myr represent estimated stratigraphic thickness, ages and lithologies beneath the borehole. Bass River results are offset by 200 m from Ancora results in order to distinguish the two data sets. Thus, the left side of the vertical axis refers to Ancora curves whereas the right side refers to Bass River results. (b) $R2_{SL}$ plotted in reference to present-day sea level ($R2_{SL} = 0$). These curves represent the difference between $R1$ results and the best-fit thermal curves and are corrected for water loading. $R2_{SL}$ estimates older than 95 Myr are not plotted because they do not represent borehole data. Bass River results are offset by 100 m from Ancora results in order to distinguish the two data sets. Thus, the left side of the vertical axis refers to Ancora curves whereas the right side refers to Bass River results.

The long-term $R2_{SL}$ results indicate roughly constant sea level through the Late Cretaceous (Fig. 4). Sea level appears to have risen from the Paleocene to a maximum in the early-to-middle Eocene. The subsequent long-term sea-level fall was completed prior to the Miocene.

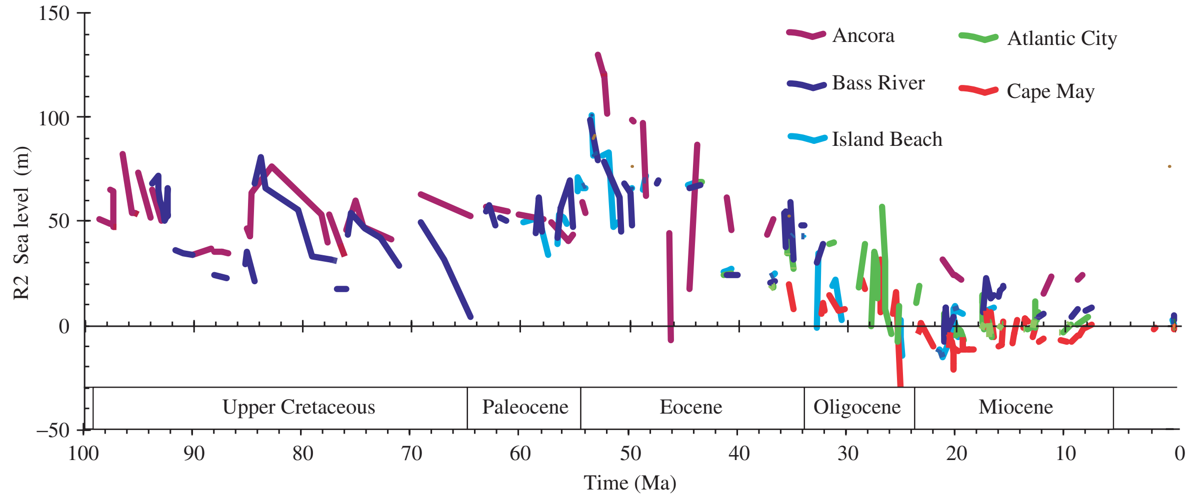

Fig. 4. $R2_{SL}$ results for the five New Jersey coastal plain boreholes. Only the best estimate values of water depth are used in order to compare the results from all five boreholes. Although the estimates of $R2_{SL}$ are not identical, there is considerable similarity in timing and magnitude short-term sea level.

DISCUSSION

Comparison of $R2_{SL}$ to published, long-term sea-level estimates

The long-term trends in sea level are generally inconsistent with previously published sea-level curves (Fig. 5). Long-term sea-level change was estimated based on changes in oceanic ridge volumes by Kominz (1984). The sea-level estimates of Kominz (1984) are similar to our maximum $R2_{SL}$ results from the Eocene maximum to the present (Fig. 5). However, the $R2_{SL}$ curves from older sequences are lower and diverge from the Kominz (1984) sea-level curve. It is important to recognize that although our results are far below the best estimate sea-level curve derived by Kominz (1984) from ridge volume data, her error range increased from about 10 m at present to about 200 m at 80 Ma owing to increasing uncertainty in the age of magnetic reversals and in the paleogeographic reconstructions. Thus, the $R2_{SL}$ results from the New Jersey coastal plain are actually within the uncertainty of the ridge volume results (Fig. 5).

The most recent analysis of sea-floor age data suggests that there may have been no change in ridge volumes with time (Rowley, 2002). This does not eliminate other contributions to ocean volume change. Harrison (1990) estimated sea-level change because of ridge volumes, Cretaceous volcanism, extension of continents resulting from break-up, compression of continents during continent–continent collision, and changing flux of sediment to the ocean. His estimate encompassing all of these mechanisms was roughly equivalent to that of the ridge volume curve alone, suggesting that, without variations in mid-ocean ridge volumes, ocean volume changed little over the past 180 Myr. An additional contribution to the long-term fall of as much as 20 m may be accounted for if the average ocean temperature has decreased by 18 °C (Sahagian, 1988). This change accounts for only a fraction of the higher Cretaceous $R2_{SL}$ values (Figs 4 and 5). We observe 45–180 m of long-term lowering that cannot be explained by these other mechanisms, and again imply a long-term drop because of decreasing rates of sea-floor spreading. Thus, the Rowley (2002) conclusions are not compatible with the long-term sea-level trends obtained from the New Jersey coastal plain.

The early to middle Eocene eustatic maximum is not seen in the ridge volume-derived curve. This suggests that factors other than sea-floor spreading generated the sea-level maximum. The sea-level high could have been influenced by tectonic controls on the size of the ocean basins. The earliest Eocene generation of the Norway/Greenland passive margin (Talwani & Eldholm, 1977) might have decreased ocean volumes through extension of passive margins and emplacement of large volumes of basalt, causing rapid sea-level rise. The subsequent sea-level fall might have been enhanced by continent–continent collision of India and Asia, which increased the ocean volume by doubling continental crustal thickness. The late early Eocene was a climatic optimum and may represent a continental ice volume minimum and, thus, a sea-level high. Subsequent increases in small glaciers (Browning *et al.*, 1996; Abreu *et al.*, 1998) and in ocean-water cooling (Sahagian, 1988) both caused sea level to fall through the middle-to-late Eocene.

Our $R2_{SL}$ estimates are substantially lower than the Haq *et al.* (1987) long-term sea-level curve derived from global sequence stratigraphic data and recalibrated to the Gradstein *et al.* (1995) and Berggren *et al.* (1995) biostratigraphic time scales following Abreu *et al.* (1998). Closer agreement between the Haq *et al.* (1987) long-term curve and the maxima in third-order $R2_{SL}$ results is obtained by shifting the long-term curve downward by 110 m, causing an overlap with $R2_{SL}$ best-estimates at 10, 24, 27 and 53 Ma. The long-term trends of the shifted Haq *et al.* (1987) and the New Jersey records are similar (Fig. 5). Both show little long-term change in the Late Cretaceous, although the Haq *et al.* (1987) curve does show a slight falling trend.

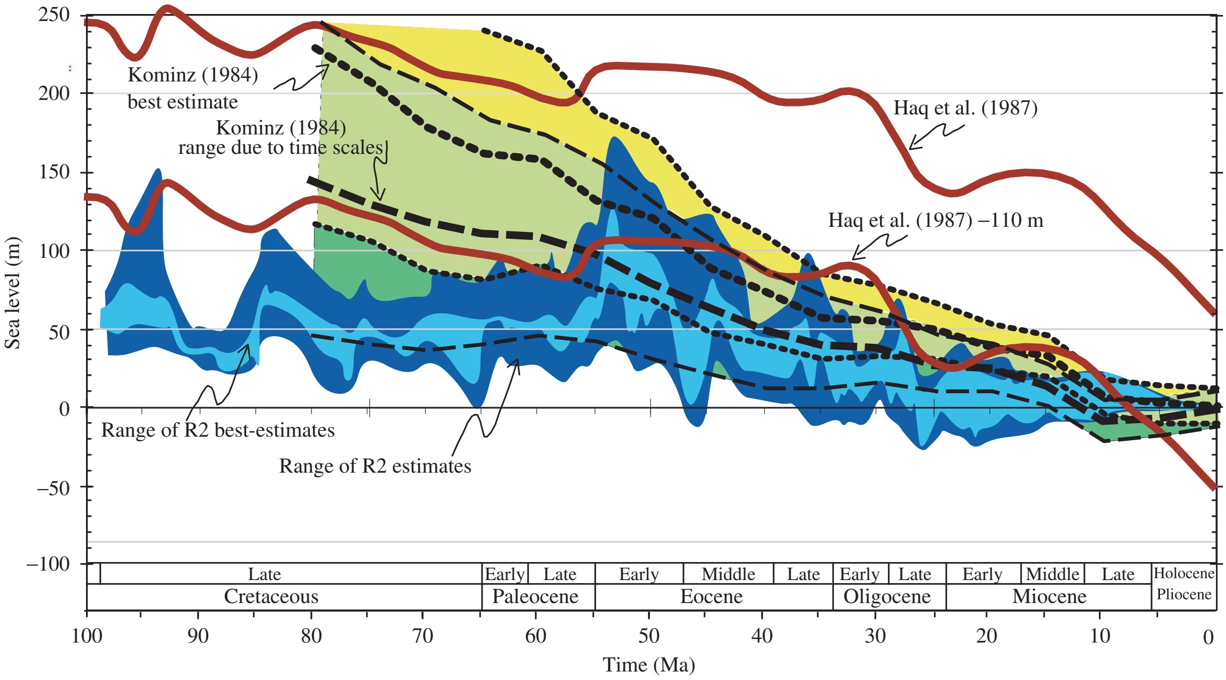

Fig. 5. Smoothed sea-level ($R2_{SL}$) results from the five boreholes. The dark blue is the result of connecting the maximum and minimum possible $R2_{SL}$ results, whereas the light blue indicates the range of best-fit estimates. No effort is made here to incorporate sea-level falls represented by sequence boundaries. The $R2_{SL}$ results are compared with the Haq *et al.* (1987) long-term sea-level curve (higher thick red curve) corrected to the Gradstein *et al.* (1995) and the Berggren *et al.* (1995), time scales following Abreau *et al.* (1998). Agreement is facilitated by a downward shift of 110 m (lower thick red curve). Comparison is also made with the Kominz (1984) long-term sea-level curve. The thick black dotted line represents her best estimate magnetic anomaly time scale with ranges because of uncertainties in paleogeographic reconstructions indicated by the thinner dotted lines. Full ranges are indicated by the yellow and light green coloured regions. Minimum ridge volumes were obtained by calibrating paleomagnetic time scales to biostratigraphic time scales that maximized the duration of the Cretaceous quiet zone. The thick dashed and thin lines indicate the mean and the ranges of sea-level change with this time scale. The green region represents this result whereas light green represents the overlap in sea-level estimates between the Kominz (1984) best-estimate and biostratigraphic time scale-constrained sea-level estimates.

A rise in earliest Eocene time to a maximum with subsequent sea-level fall through the Oligocene is also indicated by both curves (Fig. 5). The final third-order sea-level fall indicated by Haq *et al.* (1987), which began in middle Miocene and continues to the present, is much lower in magnitude or absent in $R2_{SL}$ results.

The 110-m difference between the long-term Haq *et al.* (1987) eustatic estimates and the $R2_{SL}$ curves is most likely a function of uncertainties in ridge volume as discussed above. The long-term sea-level estimate of Haq *et al.* (1987) was derived from backstripping a well from the west-African passive margin by Hardenbol *et al.* (1981), coupled with a Turonian estimate based on ocean volume changes as estimated by Kominz (1984). Thus, if the lower-most ridge volume sea-level curves are most appropriate, as discussed above, the Haq *et al.* (1987) long-term magnitudes are also too high by the same amount.

Comparison of $R2_{SL}$ to published, short-term sea-level estimates

$R2_{SL}$ from boreholes

Defining the magnitude of third-order sea-level change from backstripping is complicated by the presence of hiatuses and uncertainties in paleowater depth. Sea-level falls between sequences are not recorded in the coastal plain boreholes. We can only infer the actual magnitude of the sea-level change from the difference between the best-estimate $R2_{SL}$ values where sediments are present. Dotted lines drawn connecting sequences on Figs 6 and 7 serve to indicate the presence of lower sea-level values during sequence boundaries, but their magnitudes are unconstrained. Even where sediments are present (i.e., the hiatuses are minimal or deposition was apparently continuous), the absolute magnitudes are uncertain. However, the best-estimate paleowater depths generally reflect real deepening and shallowing of facies on the order of 10's of metres.

Upper Cretaceous

Both the Ancora and the Bass River boreholes sampled strata of Turonian and younger age (Fig. 6a). As noted above, the overall durations and estimated magnitudes of $R2_{SL}$ are similar at both locations. A few sequences are present only in one borehole. This may be the result of either dating inaccuracies (e.g., the recognition of three vs. two Magothy sequences (Fig. 6a) relies entirely on biostratigraphy; Miller *et al.*, 2004), or absence of

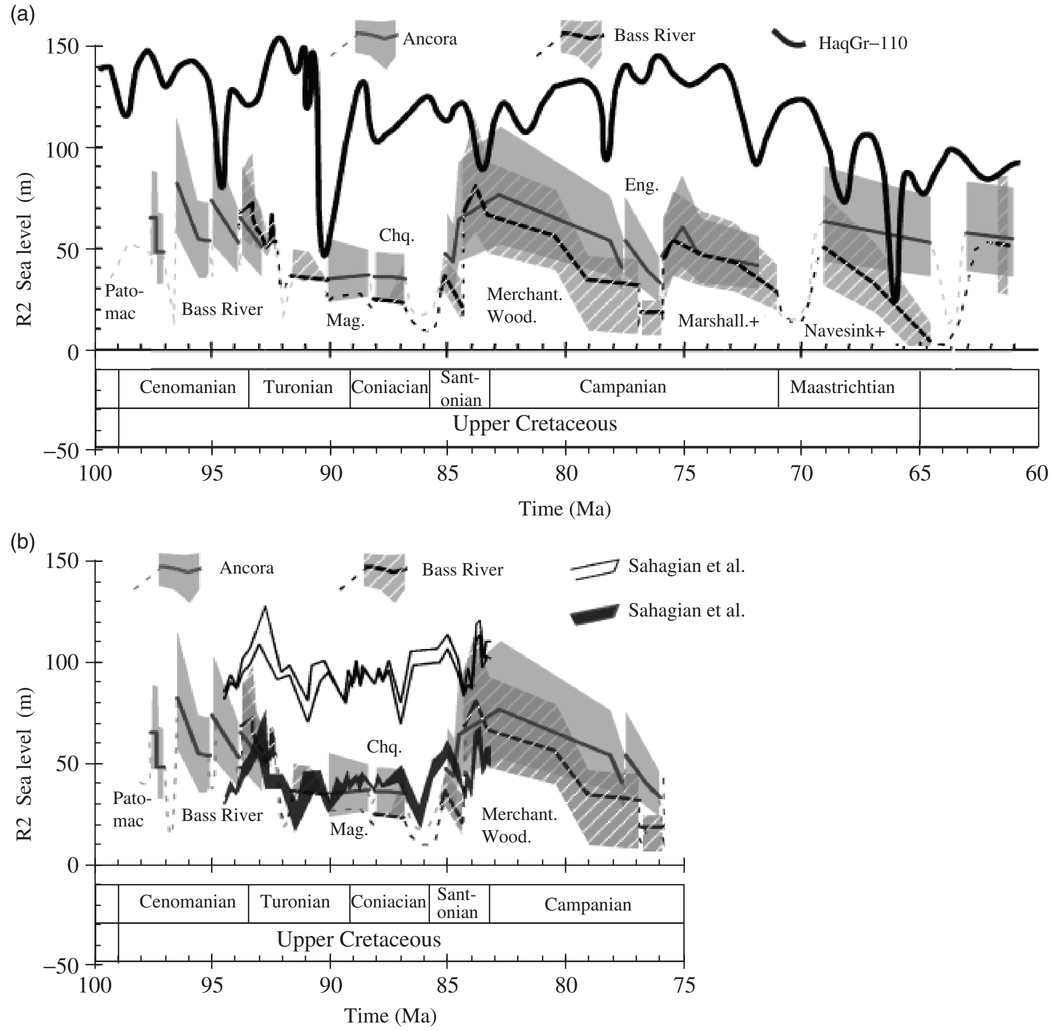

Fig. 6. (a) Cretaceous $R2_{SL}$ results for the Ancora and Bass River boreholes compared with the Haq *et al.* (1987) third-order sea-level curve recalibrated for biostratigraphic time scales (Abreau *et al.*, 1998) and shifted down by 110 m. Sequences are named according to the dominant formation of which they are composed. Potomac, Potomac Formation; Bass River, Bass River Formation; Mag., Magothy Formation; Chq., Cheesequake Formation; Merchant. Wood., Merchantville Formation and Woodbury Formation; Eng., Englishtown Formation; Marshall.+, Marshalltown, Wenonah and Mt Laurel Formations; Navesink+, Navesink Formation and the base of the Hornerstown Formation. (b) Early Late Cretaceous $R2_{SL}$ results for the Ancora and Bass River boreholes compared with sea-level estimates from backstripping the Russian Platform (Sahagian *et al.*, 1996). The black curve shows the Russian Platform data shifted down by 50 m. Ages of sea-level maxima and minima have also been shifted by less than 1 Myr to correspond to the $R2_{SL}$ results from New Jersey.

deposition at one of the locations during a particular sea-level rise. The Campanian Englishtown Formation is the only Late Cretaceous sequence for which the calculated Ancora and Bass River sea-level magnitudes do not overlap within the modelled error ranges (Eng., Fig. 6a). The latest Maastrichtian Navesink Formation also exhibits non-overlapping $R2_{SL}$ estimates, although it begins with compatible estimates. There are numerous ways to introduce error into the $R2_{SL}$ results that could result in this situation. If compaction does not follow the theoretical porosity–depth curves, because of undercompaction, or secondary dissolution, the accommodation generated in one location may be different from that at another, resulting in an apparent difference in $R2_{SL}$. Alternatively, nearby sediment loading may deflect the crust, generating accommodation that is unrelated to tectonics or eustasy. However, we attribute the two greatest mismatches (the Englishtown and Navesink sequences) to uncertainties in water-depth estimates: (1) the paleodepth estimates for the Navesink sequence were probably too high in the deep neritic setting of Bass River and (2) the paleoenvironments for the Englishtown sequence are poorly constrained. To reduce the errors, one option is to develop a two-dimensional backstripping approach as was done by Kominz & Pekar (2001) for the Oligocene section. This requires a considerably more extensive database than that provided by the two ODP boreholes.

As was the case with the long-term curve, the recalibrated Haq *et al.* (1987) third-order sea-level curve has been shifted down 110 m for comparison with the $R2_{SL}$ results (Fig. 6). In most cases, the timing of the sequence boundaries differ considerably between the $R2_{SL}$

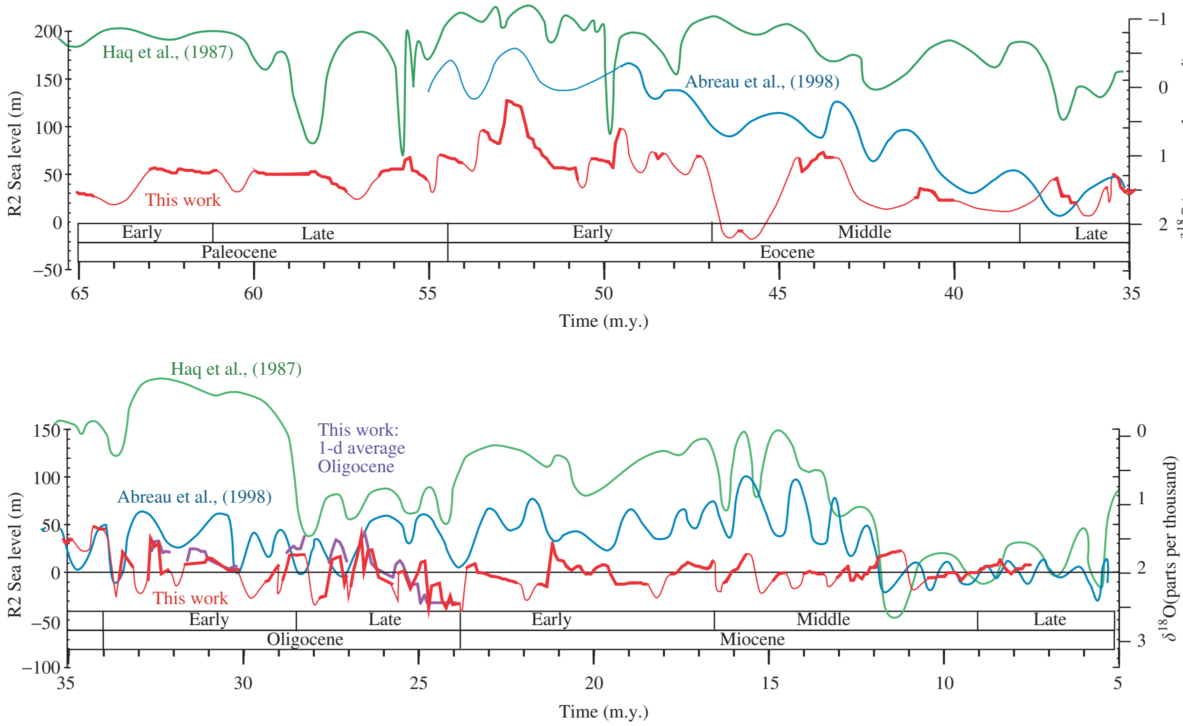

Fig. 7. Cenozoic $R2_{SL}$ results averaged from the five New Jersey coastal plain boreholes compared with other third-order sea-level estimates (thick red lines). Thin red line indicates that sea level was lower during sequence boundaries. The purple RZ_{SL} results do not include 2-dimensional Oligocene estimates. The global eustatic curve from sequence data is the Haq *et al.* (1987) sea-level curve recalibrated for more recent biostratigraphic time scales (Abreau *et al.*, 1998). The averaged $\delta^{18}O$ curve represents Abreu & Anderson's (1998) compilation. Light $\delta^{18}O$ values are plotted towards the top of the graph and indicate minimum continental ice and/or maximum temperatures.

and Haq *et al.* (1987) curves. Major sea-level falls of the Haq *et al.* (1987) curve often correspond to depositional sequences (i.e., highstands) in the $R2_{SL}$ curve (e.g., during the Navesink+ and Merchant. Wood. sequences; Fig. 6a). In other cases the sequence boundaries of the New Jersey coastal plain appear to correspond with the sea-level maxima of the Haq *et al.* (1987) curve (e.g., 70, 76 and 92 Ma). In some cases uncertainties in dating allow the two data sets to be reconciled. The errors associated with the Haq *et al.* (1987) chronology are complex (Haq *et al.*, 1988), whereas the uncertainties are about ± 0.5 Myr in biostratigraphic ages alone (Gradstein *et al.*, 1995) and ± 1.0 Myr for most of the Cretaceous Sr isotope dated $R2_{SL}$ curves (Miller *et al.*, 2004). Thus, the rather large 95-, 90-, 83.5-, 78.5-, 72- and 66-Ma sequence boundaries of the Haq *et al.* (1987) curve appear to correspond to $R2_{SL}$ hiatuses at about 95, 92, 84.5, 76, 70 and 64 Ma (as discussed in Miller *et al.*, 2003). The large, 100–110-m sea-level falls at 90 and 66 Ma in the Haq *et al.* (1987) curve require formation of significant glacial ice because of their magnitude and duration (e.g., Pitman & Golovchenko, 1991); in fact, such amplitudes would imply storage of ice in excess of the modern. Such large magnitude, rapid events are not corroborated by the $R2_{SL}$ results, which instead indicate variations in sea level of 10–25 m with occasional excursions up to 60 m (Santonian–Campanian sequences, Fig. 6a). In general, the magnitude of sea-level change required by the $R2_{SL}$ results might still reflect, in part, an ice-volume change. These sea-level changes would not require continent-scale ice sheets, but rather, more modest climatic changes (e.g., 25% of East Antarctica; Miller *et al.*, 1999c, 2003; DeConto & Pollard, 2003).

Comparison of the Late Cretaceous $R2_{SL}$ results with the sea-level estimates of Sahagian *et al.* (1996) reveals some striking similarities (Fig. 6b). Their Late Cretaceous sea-level curve was based on backstripping of strata from the Russian platform and Siberia. Although the relationship between eustatic variations and sedimentation are different on passive margins (Sahagian *et al.*, 1996), the amplitude estimates from this cratonic setting are comparable with those from the New Jersey coastal plain (Fig. 6b). A downward shift of the Russian platform curve by ~50 m and shifting of sequence boundaries by less than 1 Myr results in remarkable similarity with the late Turonian and Coniacian $R2_{SL}$ results if one ignores the high-frequency (less than 1 Myr) variability reported by Sahagian *et al.* (1996). The similarity in these results is the first indication that we may be converging on actual eustatic amplitudes for this time interval. As such, it suggests that the New Jersey margin was not impacted by non-thermal or local tectonics on the scale of these sequences (1–5 Myr) and that it is an excellent location to study eustatic sea-level change. However, there is a long-term, ~50-m offset between the two data sets. This could be due either to uplift of the Russian Platform, or to large scale, long-term subsidence of the entire New Jersey region.

Cenozoic

The $R2_{SL}$ results from the Ancora and Bass River boreholes tend to corroborate previous Cenozoic $R2_{SL}$ estimates (Fig. 4). Therefore, we discuss these results briefly. All of the best-estimate results from the New Jersey coastal plain sites were averaged to generate the $R2_{SL}$ curves of Fig. 7. Where only hiatuses occur, thin red lines are drawn to indicate lowered sea level; however, the magnitudes are not constrained (Fig. 7). Oligocene sea level was taken from the two-dimensional backstripping results of Kominz & Pekar (2001), which included results from the Bass River borehole. The two-dimensional sea-level curve was hung on the latest Eocene one-dimensional backstripping estimates. The difference between averaging and two-dimensional backstripping is compared in this figure. In general, the shapes of the two curves are quite similar, although, they are not identical. This is a result of using only the best-constrained water depths to determine sea level as well as to the two-dimensional loading effects (Kominz & Pekar, 2001). There are also sequences present in (and/or extended by) the two-dimensional result that are not seen in the one-dimensional data, because several wells were used by Kominz & Pekar (2001) to fill in gaps in the ODP borehole data set.

Large amplitude (100 m or more) eustatic variations predicted by Haq *et al.* (1987) at 58, 56, 50, 28.5 and 12 Ma) are not seen in the $R2_{SL}$ results. The Paleocene and Eocene examples occur in the middle of sequences and correspond to minor sea-level falls of 10–20 m and rises of 10–40 m in the $R2_{SL}$ curve. This suggests that these large falls were over-estimated by the Haq *et al.* (1987) approach, and/or that they are indicative of regional tectonics elsewhere at that time. The Oligocene fall occurs within a sequence in New Jersey. Age uncertainties of $\pm$ 0.5 Ma for the $R2_{SL}$ results could shift that sequence towards compatibility in timing with the Haq *et al.* (1987) curve. However, the change in Haq *et al.* (1987) sea-level highstand values from ~160 m above the $R2_{SL}$ estimates before the fall to ~60 m above the $R2_{SL}$ estimates after the fall is incompatible with $R2_{SL}$ results. This $R2_{SL}$ result is corroborated by a qualitative sequence interpretation of Oligocene marine strata in Kazakhstan (Pinous *et al.*, 1999). Two large Miocene sea-level falls at about 16 and 15 Ma correspond with unconformities on the New Jersey coastal plain. The long-term middle Miocene fall of over 150 m (Haq *et al.*, 1987) is not represented in the $R2_{SL}$ results. On the Haq *et al.* (1987) curve three consecutive highstands are separated by falls of 40, 40 and 50 m, whereas the $R2_{SL}$ results show no change. Only a regional, 130 m, 5 Myr, middle Miocene subsidence event in New Jersey could reconcile the two sea-level curves.

Oxygen isotopes are a measure of continental ice volume but are also dependent on temperature and salinity. The timing of many of the sequence boundaries in the New Jersey margin have been tied to rapid rises in $\delta^{18}O$ (Miller *et al.*, 1996). However, comparison of $R2_{SL}$ estimates with the smoothed $\delta^{18}O$ curve of Abreu & Anderson (1998) does not consistently support this relationship (Fig. 7). This may, in part, reflect uncertainties in dating. The Abreu & Anderson (1998) curve indicates a long-term sea-level maximum in the Early Eocene and subsequent Eocene fall that is consistent with our results, and with earlier global $\delta^{18}O$ curves (e.g., Shackleton & Kennett, 1975; Miller *et al.*, 1987; Prentice & Matthews, 1988). Our results do not show the middle Miocene sea-level fall of Abreu & Anderson (1998) and other global $\delta^{18}O$ curves (Shackleton & Kennett, 1975; Miller *et al.*, 1987; Prentice & Matthews, 1988; John *et al.*, 2004). However, a recent recalibration of the effects of ice volume and temperature on $\delta^{18}O$ curves (Miller *et al.*, 2003) suggests that the long-term Miocene sea level may have been much more consistent with the $R2_{SL}$ results than previously thought.

CONCLUSIONS

Late Cretaceous to Miocene sea-level estimates from backstripping are remarkably consistent among five ODP boreholes obtained on the New Jersey coastal plain, suggesting that we have isolated regional sea-level magnitudes. Late Cretaceous sea level was about 50–120 m above present sea level. Maximum sea level (70–170 m above present) occurred in the early Eocene with a subsequent long-term fall to the earliest Miocene. Evidence of long-term Miocene sea-level fall is lacking in the New Jersey coastal plain strata. The overall long-term changes are within the uncertainty suggested by changing volumes of mid-ocean ridges.

Magnitudes of third-order sea-level change decreased with time from relatively high amplitudes (as much as 60 m) in the Late Cretaceous to relatively low amplitudes in the Miocene. This is most likely because of the progressive shallowing from shelf to nearshore environments, and the resultant loss of more of the sea-level cycle in the younger sequences. Comparison of the $R2_{SL}$ results with the Haq *et al.* (1987) global sea-level curve is generally problematic. In particular, the large-scale ~100-m rapid sea-level falls suggested by sequence modelling (Haq *et al.*, 1987) are, at best, much larger than or, at worst, incompatible with $R2_{SL}$ results. A eustatic origin for Late Cretaceous third-order sequences is suggested by the remarkable similarity in timing and amplitude trends between the Siberian Platform (Sahagian *et al.*, 1996) and the New Jersey Margin. Glacioeustatic control of late middle Eocene to Holocene sequences is suggested by the link between sequence boundaries and oxygen isotopic increases (Miller *et al.*, 1998).

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