Tectonics, global changes in sea level and their relationship to stratigraphical sequences at the US Atlantic continental margin

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The principal factors that control the extent of seas through geological time are vertical movements of the lithosphere and global changes in sea level. The relative height of the sea surface determines the facies and the thickness of sediments that can accumulate in a sedimentary basin. Backstripping studies show that the primary factors affecting the subsidence of rifted sedimentary basins are thermal contraction, following heating and thinning of the lithosphere at the time of rifting, and sedimentary loading. Factors such as compaction, palaeobathymetry, erosion and global sea level changes also contribute, but their combined affects are small compared to those of thermal contraction and sedimentary loading. Simple models have been constructed which combine the effects of sedimentary loading and thermal contraction with those of compaction, sub-aerial erosion and global changes in sea level. In the models it was assumed that the lithosphere was heated and thinned by stretching at the time of rifting, sedimentary loading occurs by flexure of a lithosphere that progressively increases its flexural rigidity with age following rifting and, that sediment compaction and bathymetry change across a basin but do not vary significantly with geological time. Furthermore, different assumptions were made on the magnitude of curves of global sea level changes and the relationship between denudation rate and regional elevation. The models show that tectonics, in the form of thermal contraction of the lithosphere and flexure and slowly varying global changes in sea level, can explain a number of the stratigraphic features of the US Atlantic continental margin. In this Paper some of the implications of these results are examined for studies of (a) sea level changes through geological time; and (b) the maturation history of continental margin basins.

Keywords: Sea level; Stratigraphy; Tectonics

The development of stratigraphical modelling techniques have provided a powerful new tool with which to quantitatively examine the various processes that have shaped the sedimentary record. Recent studies (Steckler, 1981; Watts *et al.*, 1982; Steckler & Watts, 1982; Beaumont *et al.*, 1982) have focussed on modelling the stratigraphy of rifted Atlantic-type continental margin basins. These margins, which are associated with some of the largest accumulations of sediments on the Earth's surface, have been of particular interest to the offshore oil industry.

An important outstanding problem in stratigraphy is the nature of the control on the sedimentary record. Pitman (1978) has argued that the principal control on the development of transgressive and regressive sequences are changes in the rate of sea level. In another study, Watts (1982) has suggested that tectonics rather than sea level plays a major role in controlling stratigraphical sequences. Currently there is considerable controversy regarding the relative importance of sea level and tectonics in controlling the stratigraphical record.

The acquisition of large quantities of seismic reflection and well data from the world's continental margin basins during the past decade has provided a high-quality data set with which to evaluate the nature of the control on the stratigraphical record. One of the most detailed studies carried out to date has been the one by Vail *et al.* (1977). They used seismic sequence analysis techniques to establish a curve of changes in coastal onlap and offlap which they attributed to variations in sea level. By these techniques Vail *et al.* (1977) estimated a global sea level curve for the entire Phanerozoic.

The Vail *et al.* (1977) sea level curve exhibits large amplitude (up to a few hundred metres) oscillatory sea level changes. It therefore differs from the curve deduced by Watts and Steckler (1979) from analysis of the tectonic subsidence at individual wells in a margin. This curve is significantly smoother than the Vail *et al.* (1977) curve.

We believe that the question of the magnitude and shape of the sea level curve can be addressed by stratigraphical modelling techniques. By comparing the

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predicted stratigraphy to observed basin sections it should be possible to test the validity of the various sea level curves.

The stratigraphical record is the sum, however, of a number of geological processes which interact with each other through time. Beaumont *et al.* (1982), for example, have constructed models which combine the effects of thermal contraction and sedimentary loading, with compaction and erosion. It is difficult, however, in their models to evaluate the relative importance of the various factors that control the stratigraphical record. Our modelling philosophy has been to identify a series of parameters which are of importance in stratigraphy and are of known magnitude. A 'sensitivity analysis' is then carried out in which each parameter is permitted to vary with respect to a common reference.

In this Paper stratigraphical modelling techniques have been applied to the US Atlantic continental

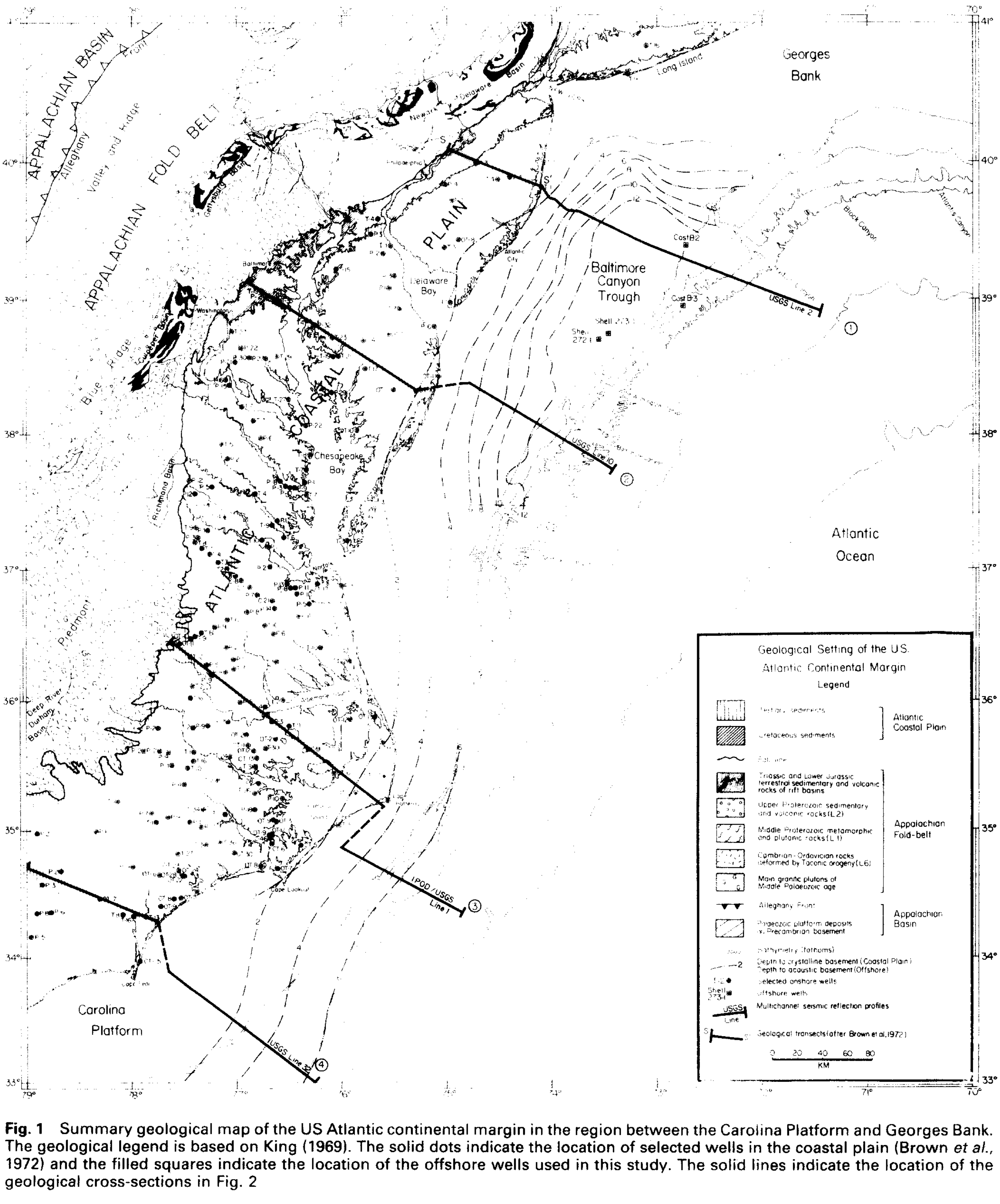

Fig. 1 Summary geological map of the US Atlantic continental margin in the region between the Carolina Platform and Georges Bank. The geological legend is based on King (1969). The solid dots indicate the location of selected wells in the coastal plain (Brown *et al.*, 1972) and the filled squares indicate the location of the offshore wells used in this study. The solid lines indicate the location of the geological cross-sections in Fig. 2

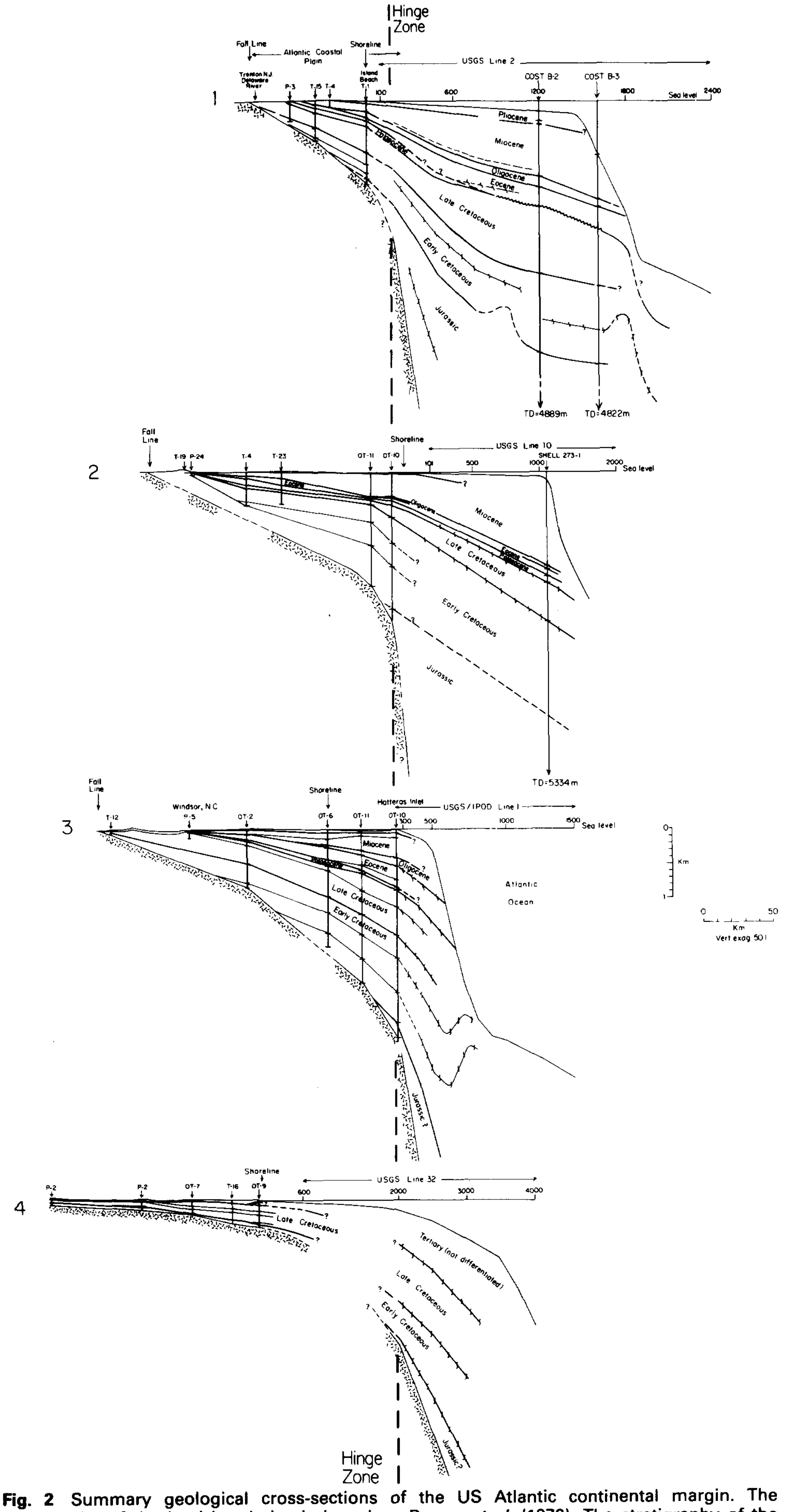

Fig. 2 Summary geological cross-sections of the US Atlantic continental margin. The stratigraphy of the land boreholes is based on Brown *et al.* (1972). The stratigraphy of the offshore wells is based on Scholle (1977a, 1977b), and Libby-French (1981). The solid lines in the offshore indicate prominent seismic reflectors identified on USGS multichannel seismic Lines 2, 10, 32 and USGS/IPOD Line 1

margin. This margin was selected for study because it is relatively old, well sedimented, and has been intensely investigated by the petroleum industry. Particular emphasis is given to the relative role of tectonics and changes of sea level in controlling margin stratigraphy. The role of other factors, such as sub-aerial erosion, sediment supply and compaction are also discussed. By systematically varying these factors in our models we hope to gain a better understanding of the development of continental margin basins.

Geological setting of the US Atlantic continental margin

The US Atlantic continental margin between the Carolina platform and Georges Bank (Fig. 1) consists of a thick sequence (up to 12 km) of gently dipping sediments which accumulated following rifting of the North American and African plates during Late Triassic/Early Jurassic. The actual age of onset of sea floor spreading is not well known. By dating the volcanic rocks on the conjugate sides of the Atlantic, Van Houten (1977) concluded that rifting began $\approx$ 196 Ma., at or near the Triassic/Liassic boundary. The oldest well identified magnetic anomaly, M-25, is known from Deep Sea Drilling Project (DSDP) sites 100 and 105 to be $\approx$ 146 m.y. The East Coast Magnetic Anomaly (ECMA), which is believed to represent the ocean/continent boundary, is separated from M-25 by 325–375 km of undated oceanic basement. DSDP Site 534 in the Blake-Bahama basin (e.g. Sheridan, 1983) suggests an age of $\approx$ 160 m.y. for the ECMA, which is significantly younger than Van Houten's estimated age of rifting.

The post-rifting history of the margin has been described in a number of previous publications (Schlee *et al.*, 1976; Poag, in press; and Schlee, 1981). Non-marine conditions in the Triassic gave way to a predominantly shallow marine environment in the Jurassic. During rapid subsidence of the margin, sandstones and shales accumulated in near-shore regions and limestones were deposited in the outer shelf, slope and rise regions. A period of reef development characterized this early phase of margin development. Similar environmental conditions continued into the Early Cretaceous. By Late Cretaceous times, however, a thinner near-shore sequence was deposited and a thicker deep-water sequence accumulated. Eventually, by the middle of the Late Cretaceous the prograding sediment wedge built over the reef ridge, distributing terrigineous material from the inner shelf across the outer shelf, to the continental slope. The wedge continued to upbuild and prograde during the Tertiary, but erosion during the Pliocene and Pleistocene considerably reduced the width of the shelf.

Fig. 2 shows the generalized stratigraphy of the margin between Cape Fear and Long Island. Each profile in the Figure is stacked on the 'hinge zone' which marks an abrupt increase in depth to continental basement at the margin (e.g. Watts & Steckler, 1979). The shelf break occurs in the vicinity of the hinge zone on profiles 3 and 4 (Fig. 1) and some distance seaward of the hinge on profiles 1 and 4. The stacked profiles show that both the fall-line and the shelf edge vary in distance from the hinge zone along strike of the margin. In general, stratigraphical horizons increase in dip at the hinge zone and are truncated at the present day continental slope and rise. Jurassic sediments are inferred to comprise a significant portion of the sedimentary wedge off shore. However, they have only been sampled at relatively few localities in the study area. Coarse non-marine sandstone, as sampled in the Esso-Hatteras Light borehole 1 (Fig. 1), is thought to be representative of much of the thickest Jurassic sequence in the region between Cape Hatteras and Long Island. The COST B-2, B-3 and Shell 273 wells penetrated similar rocks of Jurassic age, but their precise age is poorly known. The Jurassic appears to pinch out at the hinge zone (Fig. 2) and is known to be absent beneath much of the coastal plain. Overstepping the Jurassic sediments in a landward direction are beds of Lower Cretaceous age. Well data show that Lower Cretaceous beds wedge out up-dip and only a few isolated outcrops of these beds exist in the coastal plain. Fig. 2 shows that the Lower Cretaceous pinches out in the vicinity of the fall-line. In profile 1, however, the beds are clearly overlapped by beds of Late Cretaceous age. The Upper Cretaceous outcrops extensively in the coastal plain, where individual formations are youngest near the shore-line. Overlying the Upper Cretaceous beds are thick sequences of Tertiary sediments.

Biostratigraphical data from the COST B-2 and B-3 wells and seismic sequence analysis techniques indicate that several widespread unconformities occur in the stratigraphical record in the Baltimore Canyon Trough (e.g. Poag, 1980; in press; Schlee, 1980). Poag (in press) has recognized prominent unconformities (time gaps of 4 m.y. and longer) between the Hauterivian and Barremian, Turonian and Coniacian, Palaeocene and Maastrichtian, late Eocene and late Oligocene, and the middle Miocene and Pliocene. The most striking of these unconformities separate the Palaeocene/Maastrichtian and late Eocene/late Oligocene where 17–18 m.y. and 8 m.y., respectively, are missing. Both the Palaeocene and Oligocene pinch-out landward of the COST B-2 and B-3 wells, in the vicinity of the hinge zone (Olsson *et al.*, 1980; Grow *et al.*, 1979).

The deep structure of the margin, in contrast to the shallow structure, is poorly known. Gravity and geoid modelling (e.g. Watts & Steckler, 1979) have shown that the crust thins from 30 to 35 km in the vicinity of the New Jersey shoreline to $\approx$ 14 km beneath the outer shelf. Unfortunately, there have only been a few seismic refraction experiments carried out to verify the gravity and geoid modelling. Pakiser and Steinhard (1968), using a number of long-range seismic refraction profiles, showed that the thickness of the crust beneath the coastal plain is $\approx$ 35 km. At sea, Ewing and Ewing (1959) and Sheridan *et al.* (1979) estimated a crustal thickness of $\approx$ 17 km beneath the slope. Seismic refraction profiles obtained as part of the Large Aperture Seismic Experiment (LASE) experiment suggest that the crust thins by $\approx$ 15 km beneath the outer shelf, supporting the predictions of the gravity and geoid models.

Backstripping studies at the COST B-2 and B-3 wells (e.g. Steckler & Watts, 1978) suggest that $\approx$ 20 km of crustal thinning has occurred beneath the outer shelf, in accord with the gravity and seismic results.

Modelling the stratigraphy

The stratigraphy of rifted continental margins is the result of a number of factors which interact with each other through geological time. The most important are tectonic movements of the basement and sedimentary loading. Other factors include changes in relative sea level, compaction and erosion. These latter factors are small in terms of overall basin development, but they are important modifiers of the stratigraphy, particularly in the coastal plain. To quantify stratigraphy it is necessary to introduce a simple parameterization of the factors that contribute to basin development. The approach employed in this Paper is to use parameters that are empirically based wherever possible. However, some factors, such as sub-aerial erosion, can at best only be approximated.

Tectonic subsidence

Studies have now shown that the later stages of tectonic subsidence of continental margins are a result of thermal contraction of the lithosphere (e.g. Sleep, 1971; Watts & Ryan, 1976; Falvey, 1974; McKenzie,

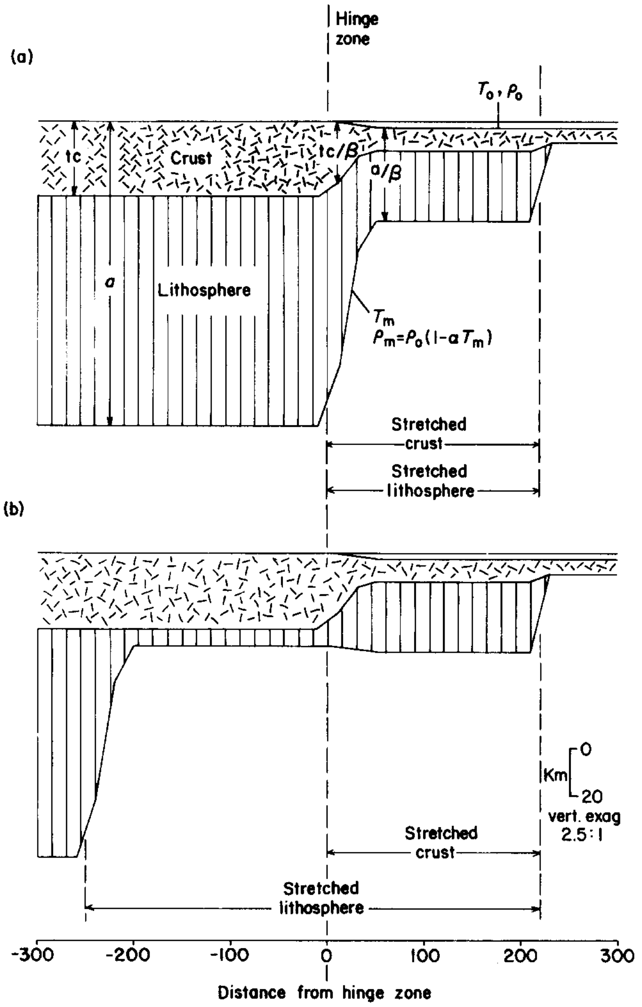

Fig. 3 Schematic diagram showing the principal features of the stretching model used in this Paper. (a) 1-layer stretching model based on McKenzie (1978); (b) 2-layer stretching model based on Royden and Keen (1980). In both models a 225 km wide zone of stretched crust and lithosphere is assumed between the hinge zone and oceanic lithosphere. In (a) stretched lithosphere is limited to this zone while in (b) it extends 250 km landward of the hinge zone

Table 1 Sea level changes through geological time determined at the COST B-2 well[a]

Age (Ma.)	Sea level (m, above present)
5.0	0.0
25.0	20.6
33.0	27.3
39.0	76.7
42.0	81.5
51.0	87.0
55.0	91.7
66.0	122.9
73.0	123.3
83.0	139.7
87.0	123.8
88.0	112.5
91.0	91.2
98.0	84.9
105.0	77.6
(122.0)	(61.1)
(137.0)	(46.6)
(144.0)	(39.8)
(150.0)	(33.9)
(161.0)	(23.3)

[a] Values in parenthesis were not determined from biostratigraphical data at the COST B-2 well but were used in the complete model

1978). The heating event responsible may be either 'passive' or 'active'. In the stretching model (McKenzie, 1978), the lithosphere is extended and the hot underlying material rises passively beneath the developing rift. The amount of extension at any point in the model is described by a parameter, β. Active heating, which is not necessarily associated with extension, may also be involved in the rifting process (e.g. Illies, 1978).

The tectonic subsidence of the COST B-2 and B-3 wells (Figs. 1 and 2) suggests that β is in the range 3–4 seaward of the hinge zone (e.g. Watts, 1981). A β value of 3.0–3.5 probably corresponds to the maximum thinning that is possible prior to hot underlying material penetrating the thinned continental crust (Le Pichon & Sibuet, 1981). In this study, a β value of 3.1 has been assumed for the continental crust seaward of the hinge zone.

Two types of stretching model are considered in this Paper: the first, a one-layer model in which the hinge zone marks the limit of crustal and lithospheric thinning, the second, a two-layer model in which the hinge zone is only a crustal thinning boundary and lithospheric thinning extends into the hinterland (Fig. 3). The one-layer model is particularly appealing because of its simplicity. It appears to be a common observation in extensional regimes, however, that the zone of anomalous lithosphere extends over a wider region than its overlying rift (e.g. Searle, 1970). Combined seismic refraction, attenuation, gravity, and electrical methods confirm this result for both the Rhine graben and East African rift system (Illies, 1978).

In our modelling a two-dimensional 'box' model (Steckler, 1981) has been used for the variations of crustal and thermal structure across a margin. A box width of 20 km was selected. Other parameters for the thermal model are given in Table 1.

Flexure

Most models consider sediment and water loading on a locally supported (Airy) crust or a flexurally supported rigid crust (e.g. Watts & Ryan, 1976). Airy compensation is considered as a special case of flexure where the rigidity vanishes.

Active and passive heating changes the flexural strength of the lithosphere by causing thermally activated creep at depth. A simple rule for the variation of flexural strength with temperature has been suggested by Watts (1978) and Bodine *et al.* (1981). They show, based on observations of flexure in oceanic regions as well as theoretical modelling, that the depth to the 450°C isotherm is a good approximation to the thickness of the oceanic lithosphere that is capable of elastically supporting loads over geologic time. Karner *et al.* (1983) have suggested that this approximation may also apply to continental lithosphere.

The equations governing flexure are described by Watts *et al.* (1982). It has been suggested by some authors (Walcott, 1970; Sleep & Snell, 1976; Beaumont, 1978) that the flexural strength of the lithosphere will ultimately vanish. The processes by which this may occur are as yet poorly known, however, so that this effect has not been included in the models presented here.

In our modelling, a numerical integration technique (Bodine, 1981) has been used to calculate the flexure at each box across the margin.

Sea level

Sea level variations occur over a wide variety of time scales (Fig. 4). Bond (1979) and Watts and Steckler (1979) attempted to determine the long-term sea level variations to a resolution of ≈ 50 m.y. These curves are 'empirical' since they used the stratigraphical record to determine sea level variations in areas where the tectonics is known. The curve of Pitman (1978), however, is 'mechanistic'. In this method changes in sea level were calculated from variations in the volume of mid-ocean ridge crests. Both types of sea level curves typically show a high during the early part of the Late Cretaceous and a fall since then.

Vail *et al.* (1977), however, attempted to determine the short-term changes in sea level to a resolution of ≈ 1 m.y. This curve consists of high-order cycles of sea level variations superimposed on a long-term curve. The long-term curve is similar to the form of the Pitman (1978) curve and is believed to be a result of mid-ocean ridge volume changes. The high-order cycles, however, imply high rates of sea level change which cannot be explained by any non-glacial mechanisms.

We believe that the pre-Quaternary high-order cycles in the Vail *et al.* (1977) curve probably do not represent sea level changes. Thus, in our modelling studies only long-term sea level curves, such as those deduced by Pitman (1978) and Watts and Steckler (1979) have been used. The purpose of using the long-term curves is to determine to what extent these curves can account for the overall stratigraphy of the margin.

However, since the publication of the long-term sea level curves, there have been improvements in the geomagnetic time-scale and down-hole biostratigraphy making revision necessary. For example, Kominz (in press) has re-evaluated the effect on sea level of changes in mid-ocean ridge crest volumes using more recent time-scales. She has shown that the maximum possible height of sea level above the present level at 80 Ma. was 365 m while the minimum was 45 m with a most probable height of 230 m above present day.

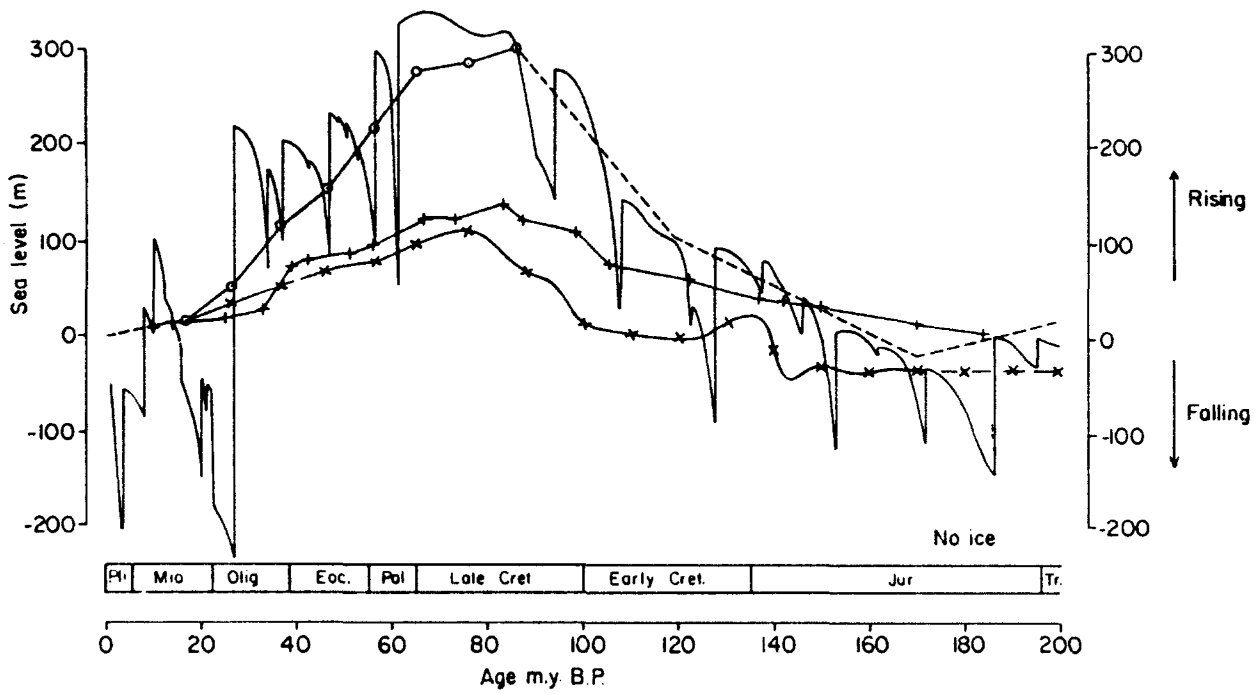

Fig. 4 Sea level changes through geological time. ○, Sea level curve determined by Pitman (1978) from estimates of changes in the volume of mid-ocean ridge crests; ---, continuation of Pitman's curve based on the first-order cycle of Vail *et al.*, 1977; +, determined in this paper from an analysis of the tectonic subsidence at the COST B-2 well; ×, curve determined by Watts and Steckler (1979) from an analysis of the tectonic subsidence at five wells off New Jersey and Nova Scotia; ——, the Vail *et al.* (1977) relative sea level curve, calibrated according to Vail and Hardenbohl (1979)

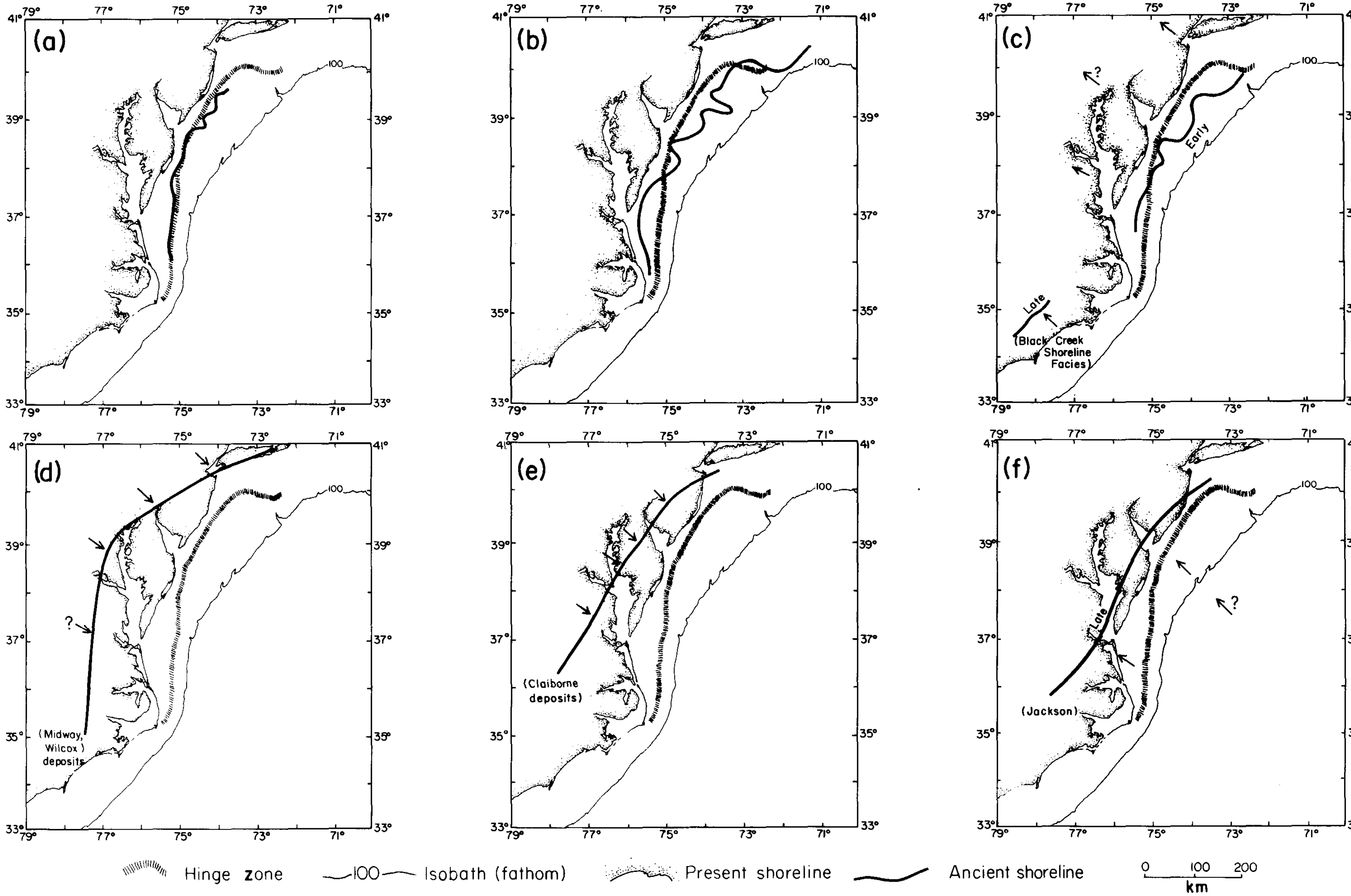

Fig. 5 Approximate positions of ancient shorelines at the US Atlantic continental margin. The Mesozoic shorelines (a–c) are based on seismic facies analysis techniques (Schlee, 1981) while the Tertiary shorelines (d–f) are based on biostratigraphical data (Richards, 1967; Colquhoun & Johnson, 1968). (a) Triassic; (b) Jurassic, (c) Cretaceous; (d) Palaeocene; (e) Eocene; (f) Oligocene

Recent revisions in biostratigraphy (Poag, in press) have been used to estimate a new sea level curve for the COST B-2 well in the US Atlantic continental margin. The method used involved constructing a model for the depth to stratigraphical horizons in a simply compacting, thermally subsiding, sedimentary basin where second-order effects such as sea level changes have not occurred. If t_0 is the time for the start of subsidence, the horizons 80 m.y. after t_0 and later are determined by four additional constants: Y_b, the asymptotic subsidence at infinite time; τ, the thermal time constant for the cooling plate; ϕ_0 the initial porosity of sediments at zero depth; and Y_p, the porosity depth constant (e.g. Angevine & Turcotte, 1982). Though reasonable constraints can be put on these constants at a particular drill site, they cannot usually be determined *independently* to the degree of accuracy needed to derive sea level. The five constants, however, can be varied simultaneously, within the constraints imposed, in such a way as to minimize the differences between predicted and observed stratigraphy at a well.

A minimization technique has been used to estimate the sea level curve from biostratigraphical data at the COST B-2 well (Fig. 4; Table 1). Horizons prior to 100 m.y., (within 80 m.y. of rifting) have not been predicted. The anomalously large depth to the base of the mid-Miocene (950 m) cannot be reasonably explained by any choice of the five constants and sea level changes. Neogene horizons, therefore, have not been used in the derivation of a sea level curve.

The overall form and amplitude of the new sea level curve is similar to the original Watts and Steckler (1979) curve. This is true even though (1) different age horizons and water depths have been used for the COST B-2 well; (2) the Watts and Steckler (1979) curve is a composite of five different wells; and (3) the form of the equations governing thermal subsidence and compaction used here is different. The principal difference in the new curve is the high rate of sea level fall in the Palaeocene and the Oligocene.

Sediment supply

Considerable morphological differences are known between rifted continental margins with high and low sediment supply, even in regions where the underlying tectonic subsidence is similar. Therefore, it is of importance in stratigraphical modelling to address the question of sediment supply and its role in determining the palaeobathymetry through time.

The concepts of 'graded shelf' and 'equilibrium profile' of a climax shelf (e.g. Johnson, 1919; Bruun, 1962) are useful in developing a model for coastal plain and continental shelf morphology and facies distribution. In these concepts the slope of the shelf surface and the distribution of facies reach a steady state, except during the periods of rapid variations in sea level. This reflects a *dynamic* equilibrium between the rate of sediment supply and the ability of wave energy to re-distribute sediment of differing grain sizes.

If this balance remains unchanged these concepts can

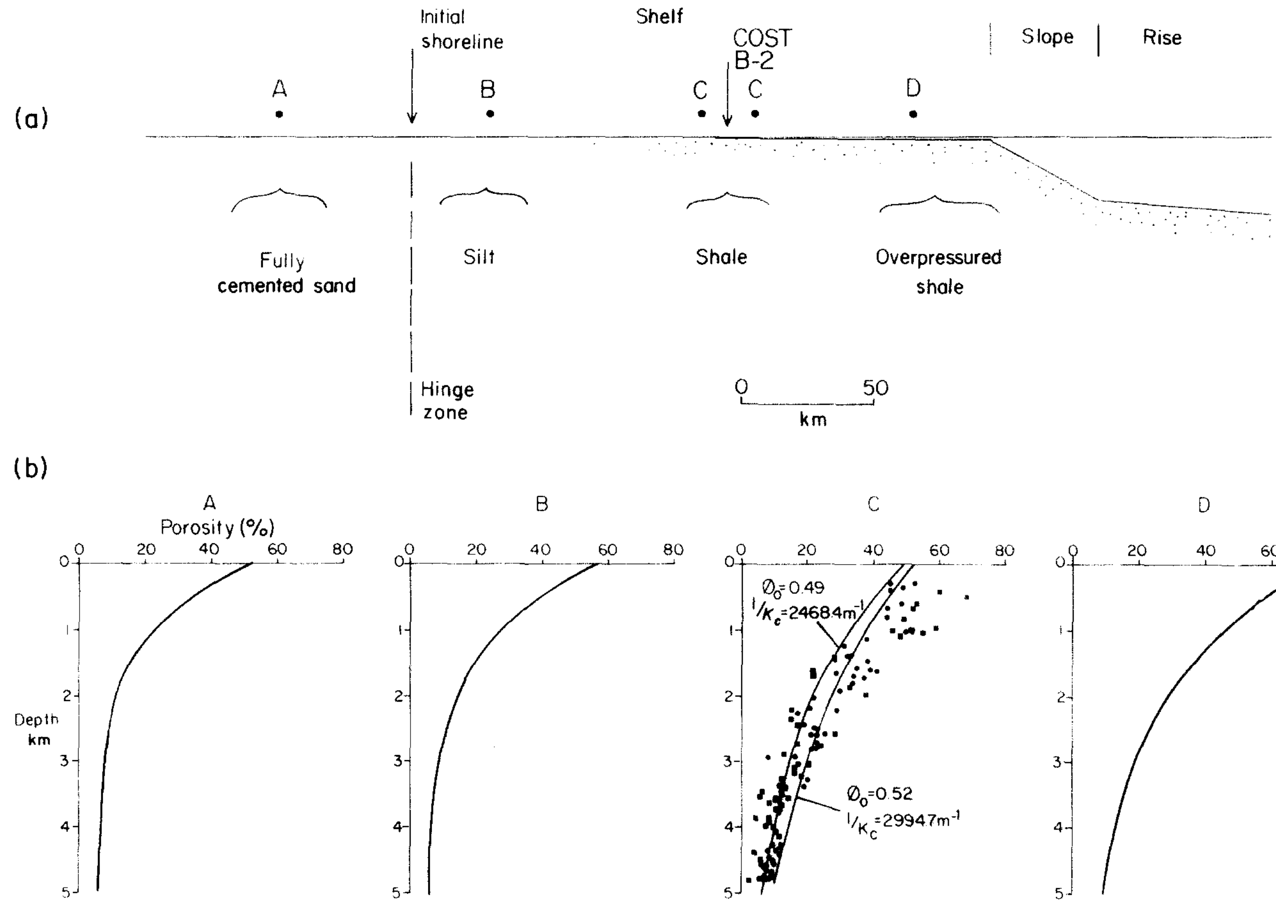

Fig. 6 Water depth and porosity *versus* depth curves across the modelled margin. (a) 'Equilibrium profile' of a shelf, slope and rise. Note that the initial shoreline coincides with the position of the hinge zone. (b) Porosity *versus* depth curves. Locality A is landward of the initial shoreline, B is in the inner shelf, C is in the middle shelf, and D is in the outer shelf

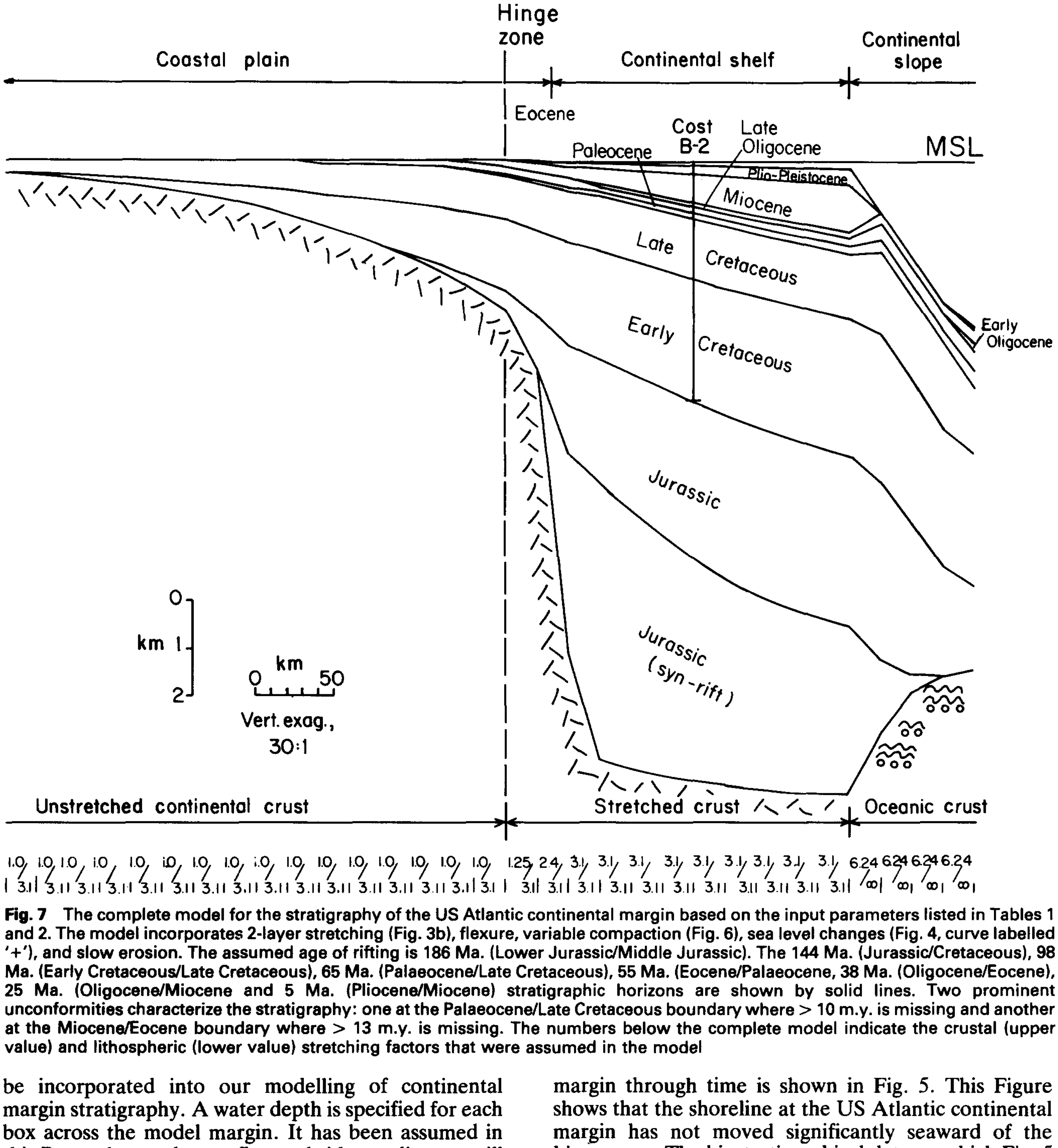

Fig. 7 The complete model for the stratigraphy of the US Atlantic continental margin based on the input parameters listed in Tables 1 and 2. The model incorporates 2-layer stretching (Fig. 3b), flexure, variable compaction (Fig. 6), sea level changes (Fig. 4, curve labelled '+'), and slow erosion. The assumed age of rifting is 186 Ma. (Lower Jurassic/Middle Jurassic). The 144 Ma. (Jurassic/Cretaceous), 98 Ma. (Early Cretaceous/Late Cretaceous), 65 Ma. (Palaeocene/Late Cretaceous), 55 Ma. (Eocene/Palaeocene, 38 Ma. (Oligocene/Eocene), 25 Ma. (Oligocene/Miocene and 5 Ma. (Pliocene/Miocene) stratigraphic horizons are shown by solid lines. Two prominent unconformities characterize the stratigraphy: one at the Palaeocene/Late Cretaceous boundary where > 10 m.y. is missing and another at the Miocene/Eocene boundary where > 13 m.y. is missing. The numbers below the complete model indicate the crustal (upper value) and lithospheric (lower value) stretching factors that were assumed in the model

be incorporated into our modelling of continental margin stratigraphy. A water depth is specified for each box across the model margin. It has been assumed in this Paper that as the sea floor subsides, sediments will infill to maintain a constant water depth and preserve the equilibrium profile of the margin.

If the balance between sediment supply and wave energy changes, however, complications are introduced which cannot be modelled simply yet. In the present study the simpler case is chosen since it appears that the US Atlantic margin between the Carolina platform and Georges Bank has remained in an approximately steady state climax margin for much of its evolution. For example, Swift and Heron (1969) have shown in a study of the Lower Cretaceous to lowermost Tertiary units off Carolina that during the transgressive and regressive stages of the Cretaceous, the margin assumed a 'climax' profile.

A further indicator of variation in morphology of the margin through time is shown in Fig. 5. This Figure shows that the shoreline at the US Atlantic continental margin has not moved significantly seaward of the hinge zone. The biostratigraphical data on which Fig. 5 is based do not have sufficient resolution, however, to accurately determine the position of the shoreline. Thus, the shoreline position represents averages over relatively long periods of geological time (up to a few tens of m.y.).

Compaction

The tectonic subsidence of a margin is amplified by sediment loading in direct proportion to the difference in density between the sediments and the underlying mantle material which they displace. The average density of any sedimentary column will vary, however, with lithology and the extent of compaction and cementation that has taken place.

The average density of a sediment column can be predicted from porosity–depth data discussed by Rieke and Chilingarian (1974) and Magara (1978) as well as other sources given in Bond and Kominz (1984). However, the application of these curves to actual stratigraphical sequences in basins is complicated by the significant heterogenities present in these sequences and the unknown percentages of water-expandable clays.

Athy (1930) was the first to suggest that sediment porosities exponentially decrease with depth. Though many other empirical forms have since been suggested the following exponential form has been used in our modelling:

$$\phi(z) = \phi_o e^{-z/C_z} \quad (1)$$

where z is the burial depth, C_z is the depth constant of compaction, ϕ_o is the initial zero-depth porosity and $\phi(z)$ is the porosity at depth z. Equation (1) appears to be satisfactory in 'normally compacting' sections as long as ϕ_o and C_z are allowed to vary with lithology. The processes that may interfere with normal compaction are overpressuring and early cementation. However, Sclater and Christie (1980) have shown, that when the overpressuring factor, λ (the ratio between excess and lithostatic pressure), is a constant, overpressuring can be accounted for by suitable choice of constant C_z in Equation (1). Early cementation is more difficult to model and its effects have not been included here.

Fig. 6 shows the variation of porosity with depth assumed across our idealized model of the US Atlantic continental margin. The compaction constants, ϕ_o and C_z, were selected to be broadly consistent with (1) the porosity derived from seismic interval velocities calculated along USGS line 25 (Grow *et al.*, 1979); (2) the measured sandstone porosities in the COST B-2 well (Rhodehamel, 1977); (3) the existence of overpressuring in the middle and outer shelf as evidenced by high mud weights required during drilling; (4) the possibility of early cementation in near-shore deposits; and (5) the generalized facies relationships for the Cretaceous and Tertiary formations in New Jersey described by Owens and Sohl (1969).

In our modelling each box was assigned a different porosity–depth curve (Fig. 6). The average density in the sediment column was calculated from the porosity–depth curve for each box assuming a grain density of 2.65 g cm^{-3}.

Erosion

In his 1971 study, Sleep suggested that erosion of uplifted areas marginal to a newly formed ocean basin could lead to significant amounts of crustal thinning and subsequent subsidence. Although it is now generally agreed that Sleep over-estimated the magnitude of this effect, it is clear that erosion may be a critical factor in understanding coastal plain stratigraphy.

The average erosion rate or denudation rate, dh/dt, for the Appalachian regions from the early Triassic to the present has been estimated by Gilluly (1968) on the basis of present day volumes of sediments in the US Atlantic margin. He found rates of at least 0.4 cm ky^{-1}. It is natural to assume, however, that this rate varied in the past as climatic conditions changed or as the average regional elevation, h_{av}, of the source terrain varied. For example, Ahnert (1970) has shown that the denudation rates of 20 modern day drainage basins in the US and Europe vary linearly as a function of mean relief estimated over a 20 km^2 area.

Pitman and Golovchenko (1983) suggest that mean elevation (with respect to base level) rather than mean relief is a better measure of the 'potential energy' of a landscape subject to erosion. They calculate that for the basins studied by Ahnert (1970) the regional elevation falls between 65 and 80% of the mean relief. Using this result they suggest:

$$\frac{dh}{dt} = h_{av}/K_e \quad (2)$$

where h is the regional elevation and K_e is an erosion time constant equal to 6.67 m.y.

They also pointed out that denudation rates today are 2–10 times those typical for pre-Recent times. Therefore, in this paper two cases have been considered: 'fast' erosion where K_e = 13.3 m.y. and 'slow' erosion where K_e = 66.6 m.y.

The mean elevation of a landscape will not follow a simple exponential, however, because of isostatic adjustment. In the model presented in this Paper, erosion is treated as an upward load on a continuous elastic plate.

Results

The parameters discussed in the previous section have been combined into a single model that we believe is applicable to the US Atlantic continental margin. In this section a model is presented for the stratigraphy of the margin, incorporating preferred values of the variable parameters described previously. We refer to the model that best fits observed sections of the margin as the 'complete model'. By referring to it as complete we do not imply that it explains all the details of observed stratigraphy at the margin. On the contrary, the complete model only accounts for the overall stratigraphic framework of the margin. We have found it useful to develop a complete model, however, because it has allowed us to evaluate the individual control of each model parameter by selectively altering each of the variables one at a time.

Fig. 7 shows a complete model for the stratigraphy of the margin based on the parameters summarized in Table 2. The model includes 2-layer stretching, flexure, variable compaction, slow erosion, and the COST B-2 well sea level curve.

The stratigraphy of the model margin is characterized by a seaward-dipping wedge of Mesozoic–Tertiary sediments. The Jurassic pinches out in the vicinity of the hinge zone and is absent beneath much of the coastal plain. The Jurassic is overstepped by Early Cretaceous which in turn progressively onlaps previously eroded basement rocks. In the model, Late Cretaceous sediments extend over the adjacent continental interior. Beginning in the Campanian with the fall in sea level (Fig. 4) sediments progressively offlap. In outcrop, sediments young in a seaward direction. As subsidence rates slow down during the later stages of margin development, the rate of sea level change

Table 2 Summary of parameters assumed in model calculations

ρ_o	= mean density of the lithosphere	= 3.33 g cm^{-3} at 0°C
ρ_a	= mean density of the asthenosphere	= 3.18 g cm^{-3} at 1333°C
ρ_g	= grain density	= 2.65 g cm^{-3}
ρ_w	= density of water	= 1.03 g cm^{-3}
ρ_c	= density of crust	= 2.80 g cm^{-3}
ρ_m	= density of mantle	= 3.33 g cm^{-3}
K_g	= thermal conductivity of grains	= 5.0 × 10^{-3} cal °C^{-1} cm^{-1} s^{-1}
K_w	= thermal conductivity of water	= 1.5 × 10^{-3} cal °C^{-1} cm^{-1} s^{-1}
α	= coefficient of volume expansion	= 3.4 × 10^{-5}
τ	= thermal time constant of lithosphere	= 64.2 m.y.
T_m	= solidus temperature	= 1333°C
t_c	= crustal stretching prior to stretching	= 31.2 km
a	= lithosphere thickness	= 125.0 km
T_e	= elastic thickness	= $Z_{450°C}$
E	= Young's modulus	= 1 × 10^{12} dyne cm^{-2}
σ	= Poisson's ratio	= 0.25
b	= radioactive heat generation depth constant	= 10.0 km

becomes increasingly important in controlling the model stratigraphy. This is well illustrated in the Oligocene where a drop in sea level from 75 to 25 m (Fig. 4) produces an interval of non-deposition on the shelf during the early part of this epoch. The thicknesses of the Neogene were not predicted by the model but are based on measured thicknesses in the vicinity of the COST B-2 well (Poag, 1980).

The model stratigraphy agrees well with the observed regionalized stratigraphy of the Atlantic continental margin between the Carolina Platform and the George's Bank (Fig. 2). The overall pattern of onlap and offlap is well reproduced though the width of the observed coastal plain varies along the margin (Fig. 2). The model coastal plain stratigraphy most closely resembles profile 4 (Fig. 2). The outcrop pattern on profile 1 and the subcrop pattern on profile 3 resembles the model patterns. A distinctive feature of the Tertiary stratigraphy, reproduced by the model, is an Oligocene pinch-out near the hinge zone. Fig. 8 compares the observed and predicted sedimentary accumulation in the biostratigraphically well-defined section of the well. The difference between the two curves can be explained by water depth variations at the COST B-2 well. These variations are approximately proportional to the water depths used and correspond well to the inferred palaeobathymetry at the well. Thus, the long-term sea level curve derived from data at the COST B-2 well appears to be consistent with the overall stratigraphical and palaeogeographical evolution of the margin.

The model stratigraphy can be used to predict those periods during the development of the margin when conditions may have been favourable for net landscape erosion and the formation of ancient denudational surfaces or peneplains. Prevalence of erosion is deduced during the Jurassic since thermal uplift kept most of the region landward of the hinge zone above sea level. This period of net erosion was terminated during the Early Cretaceous by extensive subsidence and a rise in sea level. Other periods of erosion are inferred during the Palaeocene and Oligocene when a drop in sea level exposed much of the region landward of the hinge zone. These periods of erosion inferred from the model correspond well to times of formation of the Fall Zone, Schooley and Harrisburg erosion surfaces of the Appalachians (e.g. Melhorn & Edgar, 1980).

There is an inherent non-uniqueness, however, in the modelling of stratigraphical sequences in sedimentary basins. Different combinations of the initial parameters, for example, can lead to similar stratigraphy. Therefore, a limited attempt has been made in this study to demonstrate the uniqueness of the proposed model by changing one set of parameters at a time (Fig. 9).

1-Layer stretching

In contrast to the complete model in Fig 9a, post-rift Jurassic sediments created by 1-layer stretching (Fig. 9b) extend beneath the coastal plain. The Cretaceous and Tertiary sequences are similar. In the complete model the thermal bulge created by extended lithospheric heating is sufficient to restrict Jurassic sediments, while in 1-layer stretching, flexure causes subsidence of the entire coastal plain sequence below sea level.

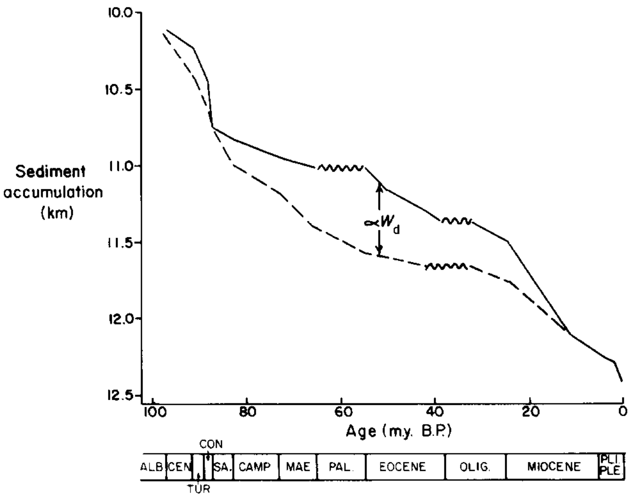

Fig. 8 Comparison of observed and predicted sediment accumulation, for the well dated part of the COST B-2 well. ——, Observed sediment accumulation from Poag (in press); ---, predicted accumulation based on the complete model in Fig. 7. The difference between the predicted and observed stratigraphy which is greatest (up to 500 m) for the period from the Maastrichtian to the Oligocene, is attributed to an increase in water depth. The wavy lines indicate unconformities where significant amounts of time (> 8 m.y.) are missing

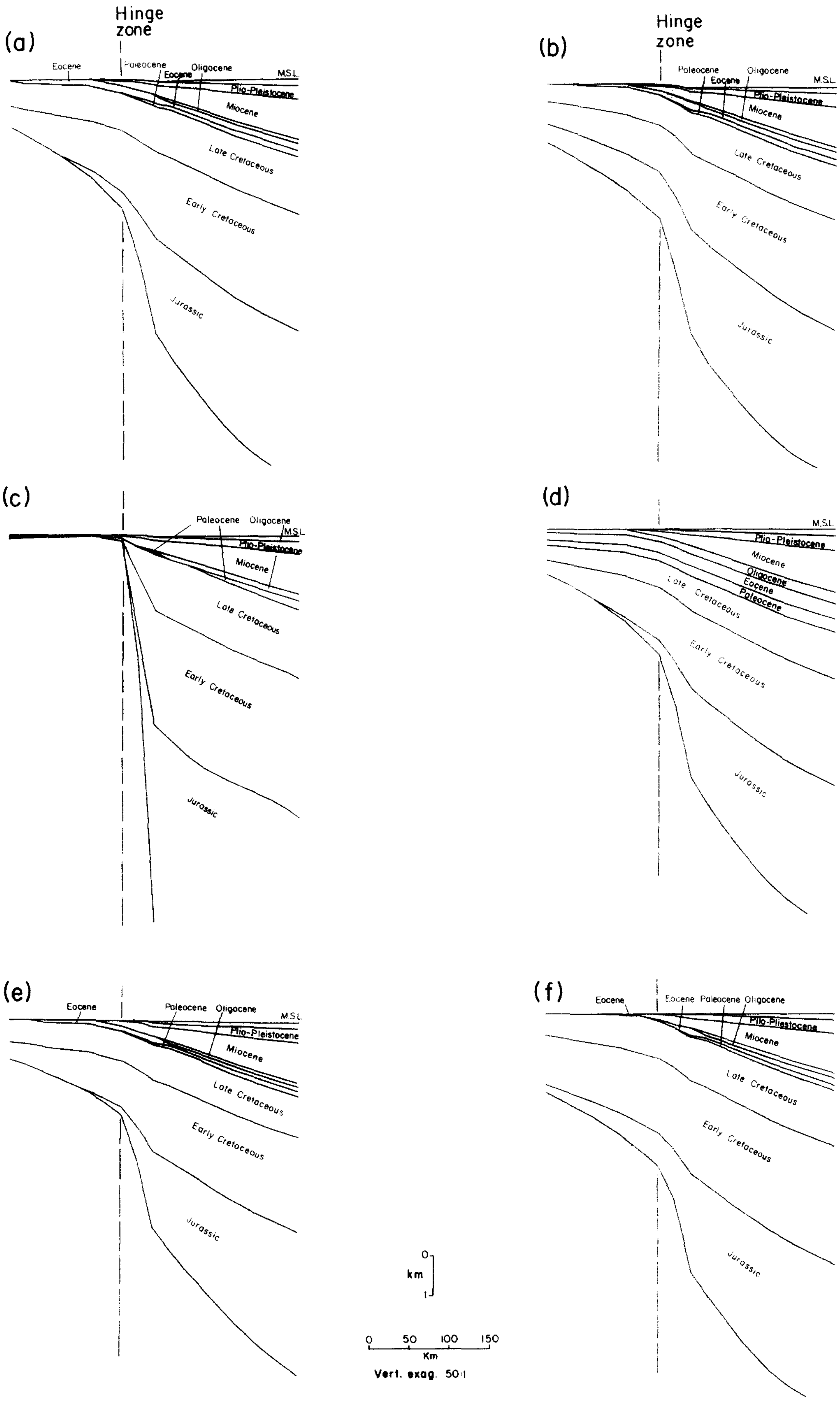

Fig. 9 Detailed model stratigraphy between the coastal plain and outer shelf. (a) Complete model; (b)–(e) complete model with different input parameters removed; (b) 1-layer stretching (Fig. 3a with stretching factor equal to 1 landward of the hinge zone); (c) no flexure (Airy-type crust); (d) constant sea level (no changes in sea level with respect to present day); (e) constant compaction (no variation in porosity–depth curve across the margin); (f) Fast erosion (erosion time constant = 13.3 m.y.)

No flexure

The role of lithospheric flexure is illustrated by comparing Fig. 9a and 9c. Fig. 9c is based on an Airy-type model for the response of the crust to sediment loading. This model is characterized by the absence of a coastal plain, an abrupt hinge zone, and an increased subsidence rate in the shelf region. The subsidence of the coastal plain in the complete model is therefore the consequence of the lateral strength of the lithosphere which in effect 'couples' the thinned crust to the continental interior across the hinge zone.

Constant sea level

Fig. 9d shows the model stratigraphy predicted for constant sea level. While the Jurassic and Early Cretaceous stratigraphy are similar for the complete and constant sea level models, the Late Cretaceous and Tertiary vary significantly. The Late Cretaceous and Tertiary units do not pinch-out but extend across the hinge zone into the continental interior due to a small, continuous, subsidence in this area. The Late Cretaceous is thicker in the complete model because of a net rise of sea level between 100 and 65 m.y., whereas the Tertiary is thinner because of the overall fall in sea level.

Fig. 10 shows a comparison of the stratigraphy predicted using a 'high' sea level curve taken from Pitman (1978) (maximum sea level of ≈ 300 m) and a 'low' sea level based on Watts and Steckler (1979) (maximum of ≈ 107 m). The high sea level (Fig. 10a) curve shows a significantly thicker Late Cretaceous and a thinner Tertiary than the low sea level curve (Fig. 10b).

The thickness of post-Palaeocene sediments associated with the two sea level curves can be estimated by considering the amount of sea level fall in comparison to the tectonic subsidence. For example, for the post-Palaeocene, thermal contraction models predict ≈ 350 m of post-rift tectonic subsidence in outer shelf regions. The amount of sea level change for the high and low curves is 265 and 107 m which subtracts from the thermal subsidence to give a net tectonic subsidence of ≈ 95 and 240 m, respectively. Multiplying the tectonic subsidence by an isostatic factor of $(\rho_m - \rho_w)/(\rho_m - \rho_s)$ gives a total sediment accumulation of ≈ 250 and 625 m. The measured thicknesses of the computed post-Palaeocene is 150 and 800 m for the high and low sea level curves, respectively (Fig. 10), agreeing to better than 200 m with the isostatic 'rule of thumb'.

These results suggest that the observed stratigraphy of the US Atlantic margin can be explained better by a low amplitude rather than a high amplitude sea level curve. However, further studies on other old margins around the world are needed to better constrain the amplitude of the long-term sea level curve.

Constant compaction

The absence of compaction in Fig. 9e leads to a subtle but systematic difference in the thickness of individual sedimentary units in the modelled margin. The upper units appear thinner because they have 20% porosity, which is lower than comparable units in the complete model. The lower unit is similar in thickness because

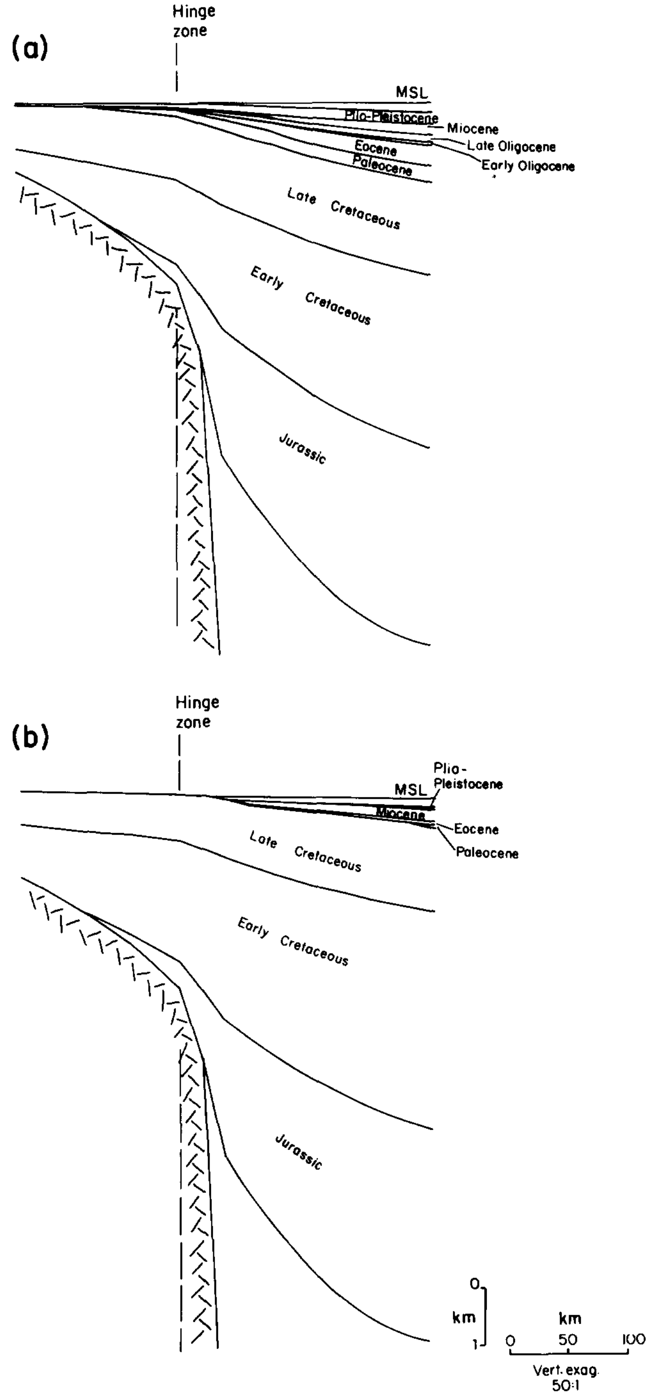

Fig. 10 Detailed model stratigraphy between the coastal plain and outer shelf for different long-term sea-level curves. (a) Low sea level curve after Watts and Steckler (1979); (b) high sea level curve after Pitman (1978). The Miocene wedge, assumed in the complete model, has been omitted in both models. The low sea level curve produces a relatively thick Tertiary wedge while the high sea level curve produces a relatively thin wedge. The presence of a reasonably thick Tertiary wedge off the US East Coast (e.g. Fig. 2) is support for a low rather than a high sea level curve

the porosity approaches that used in the complete model.

A model in which the porosity depth curve varies across a margin (Fig. 6) is in accord with seismic reflection profiles of the margin. For example, Fig. 11 shows a comparison of the complete model 40 km seaward of the hinge zone to USGS multichannel seismic Line 10 (Grow *et al.*, 1979). The depths to individual stratigraphical horizons in the complete

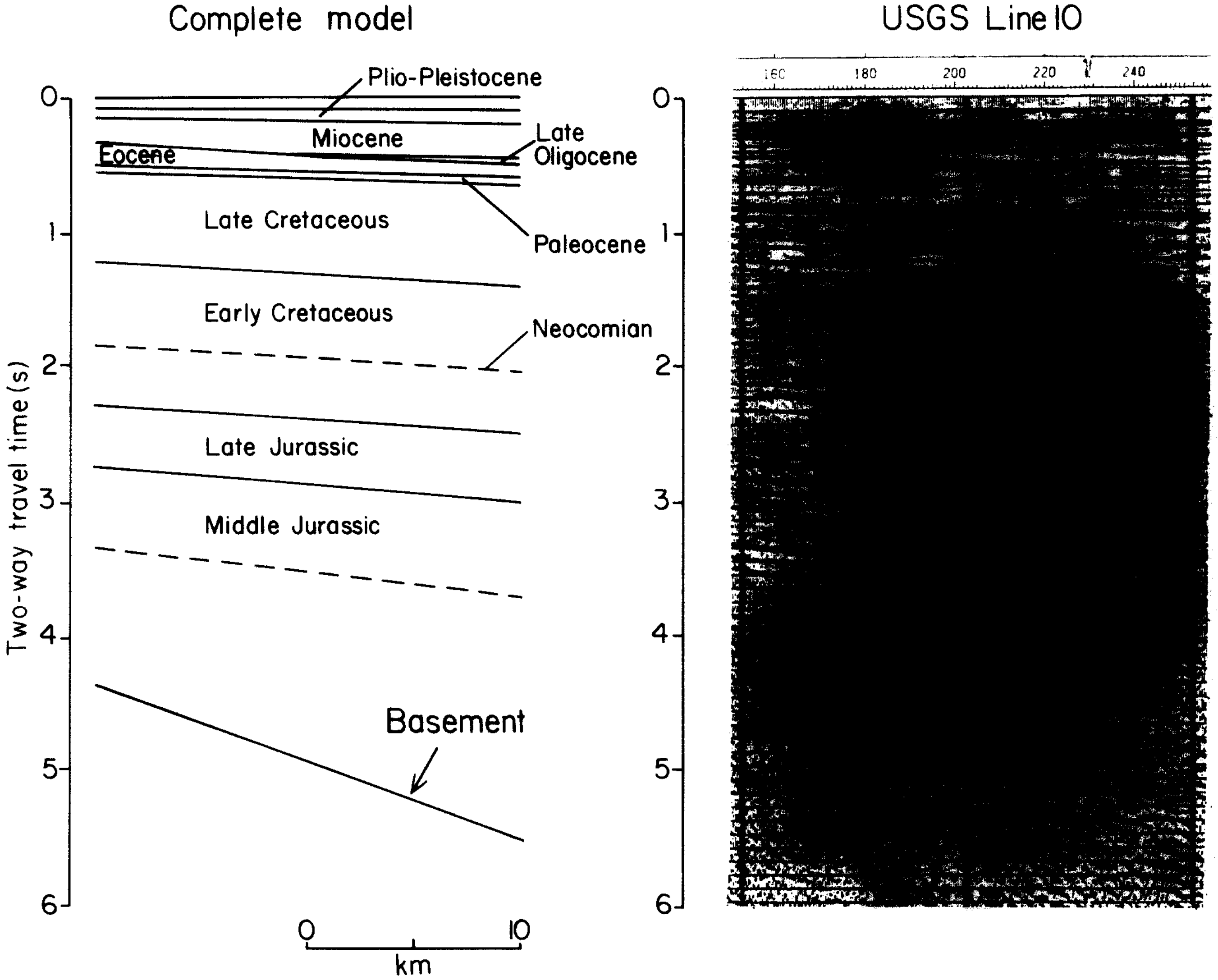

Fig. 11 Comparison of the complete model stratigraphy and USGS Line 10 off Delaware (Fig. 1) 50 km seaward of the hinge zone. The travel time for the complete model was computed from the porosity of each stratigraphic unit assuming a travel time of 619.9 $\mu s\ m^{-1}$ and 229.6 $\mu s\ m^{-1}$ in the pore fluid and matrix, respectively

model were converted to two-way travel time using a linear relationship between porosity and transit time. Fig. 11 shows that there is good agreement between the slopes of individual stratigraphic horizons in the complete model and dips of prominent seismic reflectors on Line 10.

Fast erosion

Fig. 9f shows the model stratigraphy predicted for fast erosion. The model is based on an erosion time constant one-fifth that used in the complete model. The principal differences between the two models occur in the coastal plain where there is sub-aerial erosion. In the Jurassic, the region landward of the hinge zone remained emergent due to thermal uplift. A combination of rising seas and fast erosion reduced the emergent region below sea level before the end of the Jurassic. As a result, the upper Jurassic extends beneath the entire coastal plain. The outcrop pattern differs from slow erosion mainly in that fast erosion removes Eocene deposits, thereby exposing Late Cretaceous deposits.

Discussion

This study has shown that the major features of the pre-Miocene development of the US Atlantic continental margin can be explained by the simple stretching model of McKenzie (1978) provided that the effects of flexure, compaction, shelf sediment dynamics, sub-aerial erosion and sea level changes are included. The modelling philosophy presented follows that of Steckler (1981), Steckler and Watts (1981), and Watts (1982). A simple model of the tectonic subsidence is used to predict the observed thicknesses of the stratigraphical units. This approach is unlike that adopted by Beaumont *et al.* (1982), for example, where the sediment thickness at each model time is determined by 'decompaction of the appropriate part of the sediment section as presently observed.'

An important aspect of the model presented in this Paper is that it includes the effects of changes in the relative rate of sea level and tectonic subsidence on the stratigraphy. The sensitivity of the landward extent of coastal marine deposits to the rate of change of sea level has been pointed out previously by Pitman (1978).

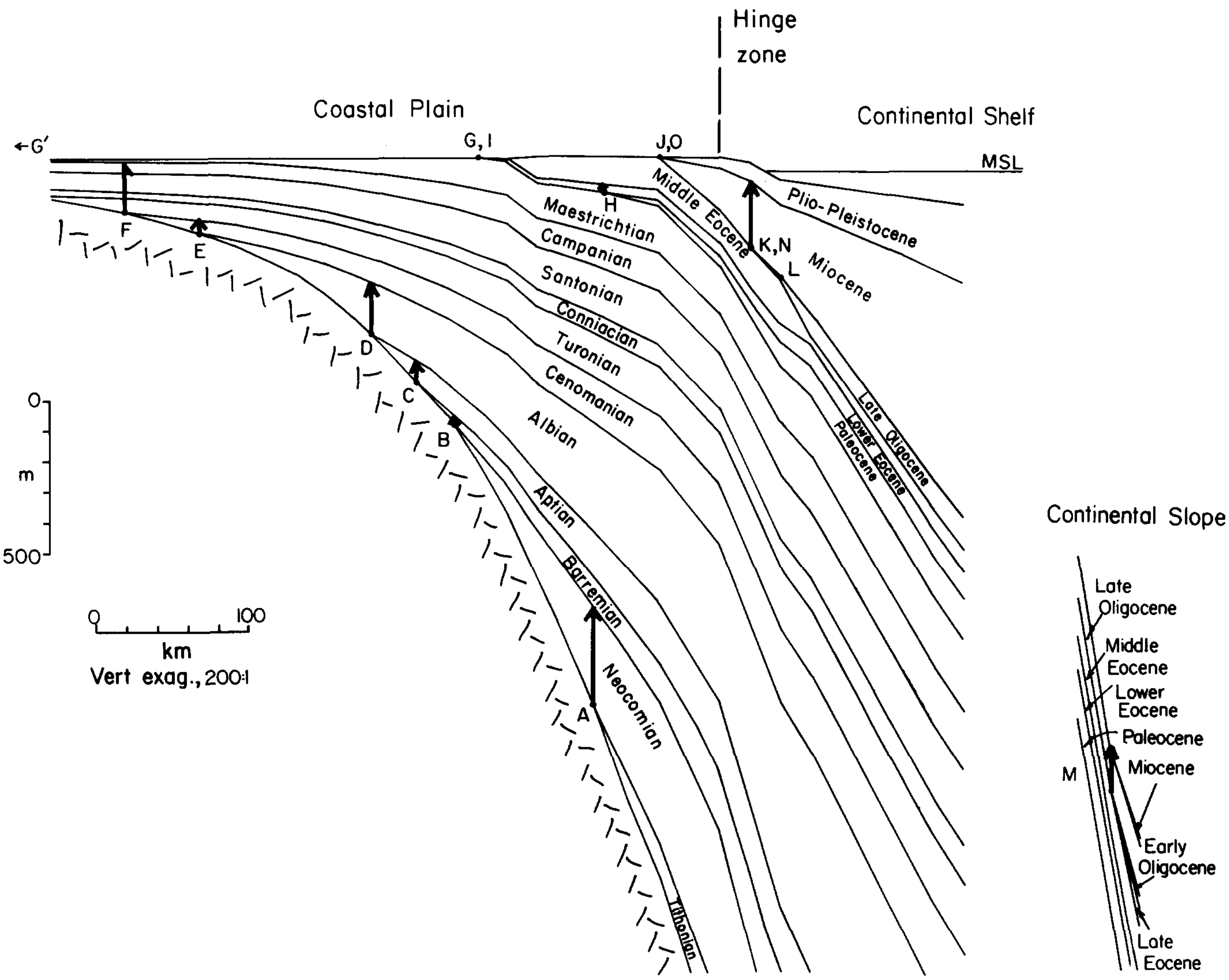

Fig. 12 Detailed model stratigraphy of the coastal plain and inner shelf region. The model is similar to the complete model in Fig. 7 except that more stratigraphical horizons are included and the vertical exaggeration has been increased. A–F, I and N and O indicate points of onlap of an overlying stratigraphical unit on an underlying unit. G and H and J–M indicate points of offlap. The heavy arrows show estimates of the vertical component of coastal onlap used to construct the cycle chart in Fig. 13

The synthetic stratigraphy created by Pitman (1978), however, differs in several ways from the work presented here. For example, consider the implications of the two models for the Oligocene. Biostratigraphical studies indicate that the Early Oligocene is generally missing across the US Atlantic shelf (e.g. Poag, in press). In Pitman's model this observation would require that there was a time transgressive movement of the beach across the shelf prior to the Oligocene. Pitman and Golovchenko (1983) have pointed out that such a movement would require a large fall of sea level *below* the shelf break. In the model presented here, however, the non-depositional surface is not time-transgressive, the sediments directly below the surface are not necessarily shallow-water, and sea level may not fall below the shelf break. Consequently, a sea level fall of only 50 m can produce the observed unconformity.

In this section the implications of the model are examined for (1) sea level changes; and (2) the maturation history of the US Atlantic margin.

Sea level changes

This study has shown that long-term sea level changes in combination with a simple tectonic model can explain a significant portion of the stratigraphy of the US Atlantic margin. The sea level curve that was used does not oscillate, but exhibits a gentle rise to 83 Ma. followed by a fall to the present day. This curve differs, however, from that used by Vail and colleagues to predict stratigraphy (e.g. Fig. 4).

To evaluate this discrepancy we have analysed our model stratigraphy using a similar technique to that used by Vail *et al.* (1977) to derive coastal onlap from observed basin sections. This technique is illustrated in Figs. 12 and 13. In the modelled stratigraphy the fine lines are equivalent to chronostratigraphical horizons which separate depositional sequences with lower and upper boundaries. Starting at the lowest point of coastal onlap (A, Fig. 12) the vertical component of coastal aggradation is estimated by tracing the lower boundary to successive points of onlap (B–G′). From these points the upper boundary is followed down-dip to a point vertically above the previous point of onlap. In this way, coastal onlap can be measured. Coastal offlap is estimated by tracing from the highest point of coastal onlap in an underlying sequence to the lowest point of onlap in an overlying sequence. For example,

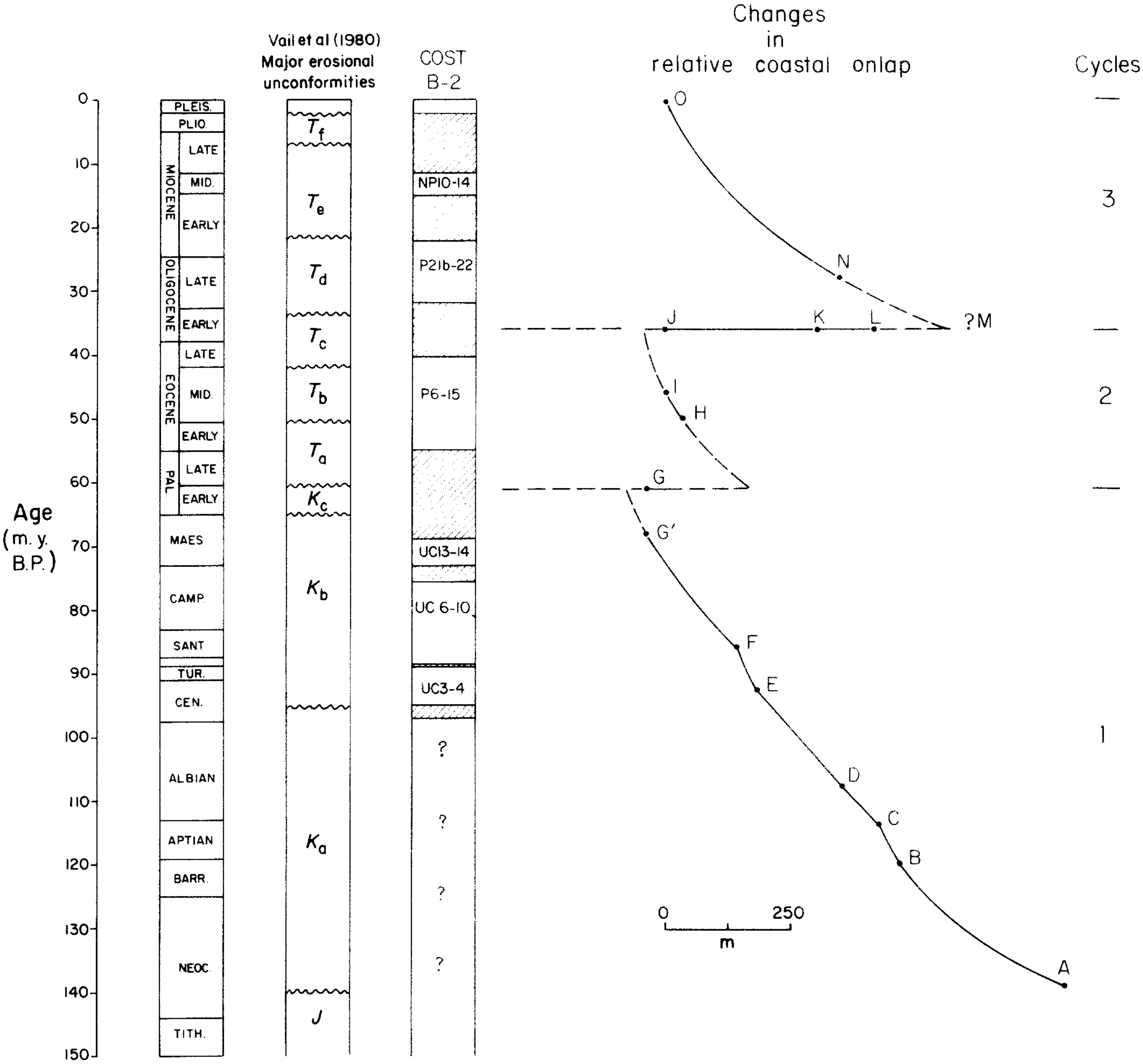

Fig. 13 A cycle chart derived from the model stratigraphy in Fig. 12 using the techniques of Vail *et al.* (1977). A–O correspond to onlap and offlap points in Fig. 12. Biostratigraphical data for the COST B-2 well is based on Poag (in press). The chronological data is based on Harland *et al.* (1982) and major erosional unconformities are based on Vail *et al.* (1980)

from point G the vertical distance to H is measured. The form of the resulting changes in the relative coastal onlap curve (Fig. 13) shows three major cycles of onlap. The first cycle occurs between 45 and 125 m.y. after rifting during a gradual flexing of the margin. As a result a progressive onlap of sediments onto the basement is produced, largely because of the increase in flexural strength of the basement following rifting (Watts, 1982). The second cycle occurs between 36 and 61 Ma. (Late Palaeocene to Early Oligocene). Towards the end of the first cycle, between F and G′, the relative rate of sea level change and tectonic subsidence at the point of onlap gradually becomes equal. At the beginning of the second cycle the rate of fall of sea level at the onlap point exceeds the rate of tectonic subsidence causing a downward shift in the pattern of onlap from G to H. The upper part of the second cycle exhibits a slower rate of fall of sea level, thereby giving rise to a pattern of onlap during the Eocene. The third cycle begins in a similar way to the second cycle with a rapid rate of sea level fall causing a downward shift in coastal onlap from J to M. A straightforward application of Vail's sequence analysis technique to point M, which is located near the base of the slope, gives a total change of apparent sea level fall between L and M of 2750 m. Clearly, the technique described by Vail *et al.* (1977) is not applicable to the case where large water depth changes occur across a basin. The upper part of the third cycle comprises of coastal onlap from points M to O. Since it is not possible to tie point L to M it is impossible to connect the second and third cycles. Rather, it was assumed that the highest point of coastal onlap, O, was on the same *datum* as the highest onlap point in the second cycle, J.

Each cycle shows a gradual increase in onlap terminated by a sudden downshift. The magnitude of cycles 1 and 3 is large, ≈ 750 m. This is of similar magnitude to the total coastal aggradation measured by

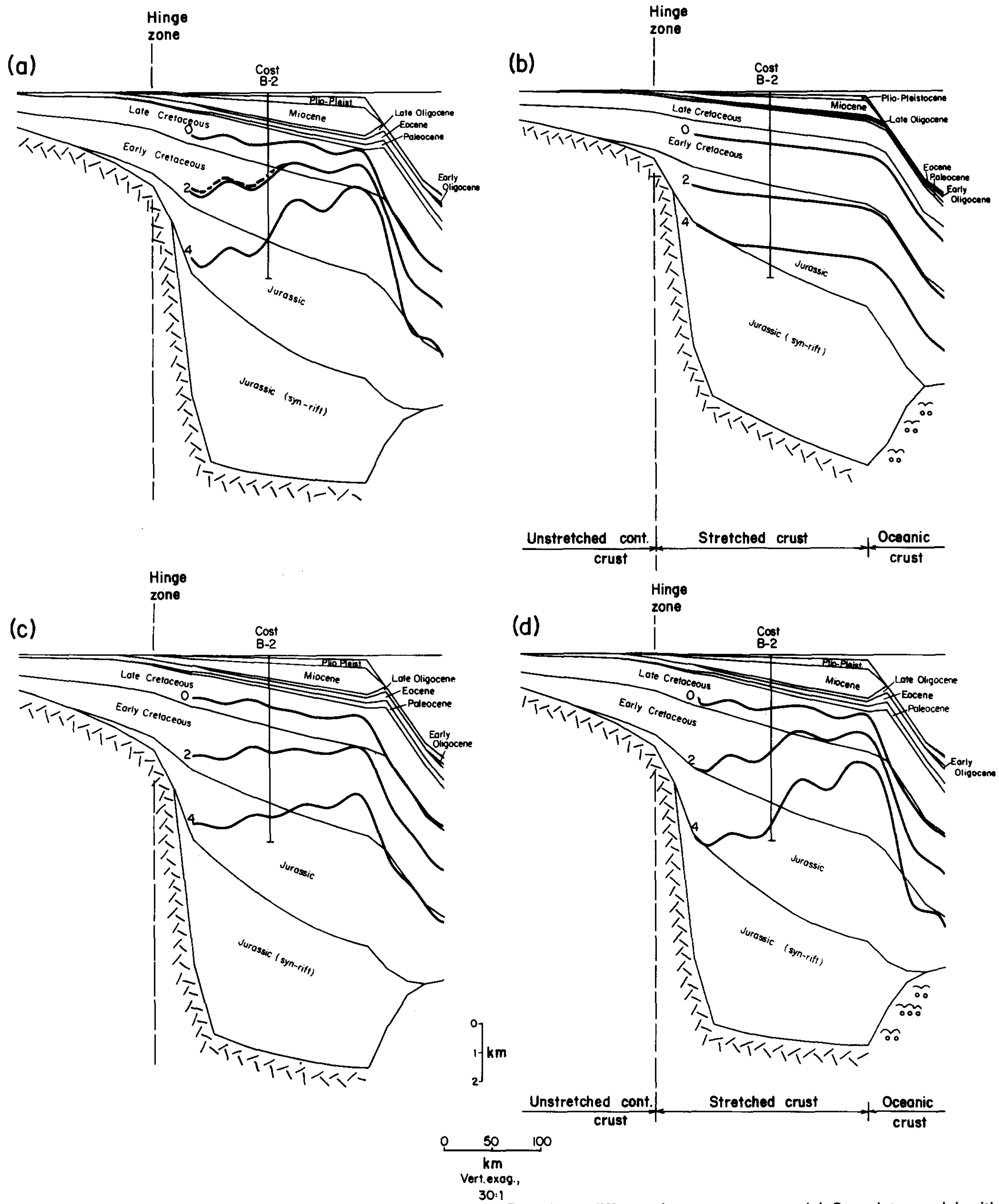

Fig. 14 Maturation indices across the complete model margin (Fig. 7) for different input parameters. (a) Complete model with parameters listed in Table 2 and variable compaction (Fig. 6) (crustal heat flow = 0.8 HFU; variable initial porosity = 0.5–0.7); (b) same as (a) but with a constant porosity of 20% for all stratigraphic units (no compaction); (c) same as (a) except that ϕ_0 was chosen to vary from 50 to 60% across the margin in contrast to the complete model (50–70%); (d) same as (a) except a limited crustal heat flow of 0.4 HFU. The dashed line in (a) shows maturation indices for a one-dimensional model which ignores the effects of lateral heat flow and flexure. Contours of maturation index are in units of 2 (2 = peak oil generation)

Vail *et al.* (1977, p.79) from observed seismic sections off northwestern Africa. Thus, cycles were obtained of similar shape and magnitude to those used by Vail *et al.* (1977) to derive their oscillatory sea level curve.

Comparison of the cycle chart in Fig. 13 with the Vail charts (e.g. Vail & Todd, 1981) shows significantly fewer cycles. Although we do not necessarily expect to see all major and minor cycles during the Jurassic and Cretaceous because of rapid tectonic subsidence at this time, we fully expect to see all Cenozoic cycles represented at this margin. The terminations of cycles 1 and 2 during the Palaeocene and Oligocene correspond well to both the major cycles (K_c/T_a and T_c/T_d) of Vail *et al.* (1980) and biostratigraphical data from the COST B-2 and B-3 wells (Poag, in press). The absence of sequence boundaries, T_c/T_b and T_b/T_a, raises several interesting questions which will be considered elsewhere (Thorne & Watts, submitted).

We conclude that a long-term sea level curve derived from a single stratigraphic well can explain many of the fine details of the stratigraphy of the US Atlantic margin. Furthermore, oscillatory changes in sea level, such as those proposed by Vail *et al.* (1977), are not required to explain the stratigraphy at the margin.

Maturation history

A number of authors have attempted to show that once the stratigraphy and changes in basal heat flow are reasonably well known, the thermal maturation history of a basin can be predicted (e.g. Keen, 1979; Royden *et al.*, 1980; Sclater & Christie, 1980). The maturation of hydrocarbons, however, is exponentially dependent on temperature. Thus, variations in the model parameters controlling the stratigraphy of a basin could lead to large uncertainties in predicted maturation indices.

Fig. 14 shows maturation indices across the complete model margin which have been computed for a variety of different input parameters. Following the procedure of Sclater and Christie (1980), the time–temperature history of a given sedimentary unit can be predicted from the basal heat flow and the thermal conductivity *versus* depth profile through time. The basal heat flow is determined from the lithospheric stretching with an additional component provided by crustal radioactivity. It is assumed that the thermal conductivity of the sediments is given by:

$$K_s = \phi \, . \, K_w + (1 - \phi) \; K_g \tag{3}$$

where ϕ is porosity, K_w is the thermal conductivity of water and K_g is the thermal conductivity of grains. The maturation index, TTI, (Wapples, 1980) has been calculated from:

$$\log_{10} \text{TTI} = \log_{10} \int_0^t 2^{(T(t)-105)/10} \, . \, dt \tag{4}$$

Contours of the maturation index for various model runs (isomats) are shown in Fig. 14. A value of $\log_{10}$ TTI of 2 corresponds to peak oil generation. Fig. 14a shows isomats for the complete model. The depth to peak oil generation is $\approx$ 3.5 km in the area between the hinge zone and the COST B-2 well, quickly shallowing to $\approx$ 2.6 km beneath the outer shelf. In the area of the slope, the depth to peak oil generation increases to $\approx$ 4.0 km below the sea floor.

While the stratigraphy shown in Figs. 14a–d is similar, the depth to peak oil generation varies significantly. The factor most influencing this depth is the porosity–depth function. In Fig. 14b a constant porosity of 20% is assumed for all units. The resulting isomats are smooth, parallel, and dip gently seaward. The region in the vicinity of the hinge zone is more mature than areas further seaward because of lateral heat flow from an area of relatively high heat flow landward of the hinge (where the lithospheric stretching factor, γ, is 3.1 and crustal heat flow is 0.8/3.1 HFU). In general, chronostratigraphical horizons dip more steeply seaward than the isomats particularly in the lower parts of the section. As a result, potential source beds in the Upper Jurassic, for example, would be more thermally mature beneath the shelf break than in the region of the hinge. However, the actual depth to the isomat corresponding to peak petroleum generation is shallower in the region of the hinge. In Fig. 14c the initial porosity was chosen to vary from 50 to 60% across the margin, in contrast to the complete model where outer shelf sediments reach a porosity of 70%. In this example, there is little variation in the depth to peak oil generation across the shelf. However, the variation in maturity of source rocks across the shelf is large. For example, the Upper Jurassic has a maturation index > 4 below the shelf break and an index of 2 near the hinge.

As shown in Fig. 14 estimates of thermal maturation are strongly model dependent. A further variation is shown in Fig. 14a. In this example, an approach following that of Sclater and Christie (1980), Keen and Barrett (1981) and Sawyer *et al.* (1982) is used. In this approach, the time variation in heat flow is estimated by best fitting a one-dimensional thermal subsidence model (neglecting lateral heat flow) to an Airy backstripped sediment accumulation curve. Fig. 14a shows that in this case this method gives a virtually equivalent result to the complete model. This is true even though the stretching factors derived from backstripping the model margin are significantly lower in all cases than those used in the complete model (β of 2.6 *versus* 3.1 at COST B-2). Lateral heat flow in the complete model leads to additional cooling of the shelf. The near equivalence of the dashed and solid lines in Fig. 14a arises from a trade-off of two-dimensional factors; horizontal heat flow and flexure. Thus, the one-dimensional approach, which is apparently widely used in the oil industry (C. Tapscott, personal communication) may be equivalent in certain cases to a fully two-dimensional model. It is suggested, however, that an investigation of other margin geometries be carried out before a one-dimensional model is used to predict the thermal regime of frontier basins.

A comparison of predicted and observed maturation indices in the COST B-2 well is complicated by uncertainties in both model parameters and accurate calibration of the various geochemical maturation indicators. For example, the depth to predicted peak oil generation ranges from 2.5 to 3.2 km within 20 km of the projected position of the COST B-2 well for the various models in Fig. 14. Uncertainties in the observed depth to peak generation in the COST B-2 well have also been reported by Claypool *et al.* (1977). Estimates range from 2.8 km to the base of the well (4.8 km), depending on which method of maturity analysis is used. One possibility that will help better define the prediction of thermal maturation in continental margins is a better understanding of thermal conductivity–porosity changes in shelf sediments. For example, in the model margin, it is the thermal conductivity in the upper Tertiary part of the section that has controlled the maturation of deeper source beds.

Quantitative stratigraphy is now possible in many basins of the world where subsidence is controlled by well known tectonic processes such as thermal contraction and flexure. In these basins, model studies such as the one described here should enable a better understanding of the various factors that control the deposition and preservation of the stratigraphical record. One of the more difficult tasks will be incorporating the facies concept into stratigraphical modelling. This concept has not yet been included in the models presented in this Paper. Future work, however, should enable a better understanding of the changing patterns of facies in sedimentary basins through time. The development of more refined tectonic models that include the facies concept offers the most promise of determining the nature of the control on the stratigraphical record during the next decade.

Acknowledgements

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Appendix

The principal steps of the FORTRAN Program SUBSID2DC, used to construct the modelled stratigraphy in Figs. 7, 9, 10, 11, 12 and 14 are summarized below:

BEGIN

- I. Input model parameters
  - a) Table 2
  - b) Box size, box number, stretching factors, compaction factors
  - c) Equilibrium profile
  - d) Table 1
- II. Compute initial subsidence and heat flow for each box

DO FOR EACH TIME STEP:

- III. Compute thermal subsidence flexure and heat flow
  See Fig. A1 and Bodine (1981)
- IV. Output model parameters
  - a) Depth to basement
  - b) Depth to top of sediments
  - c) Maturation indices

END.

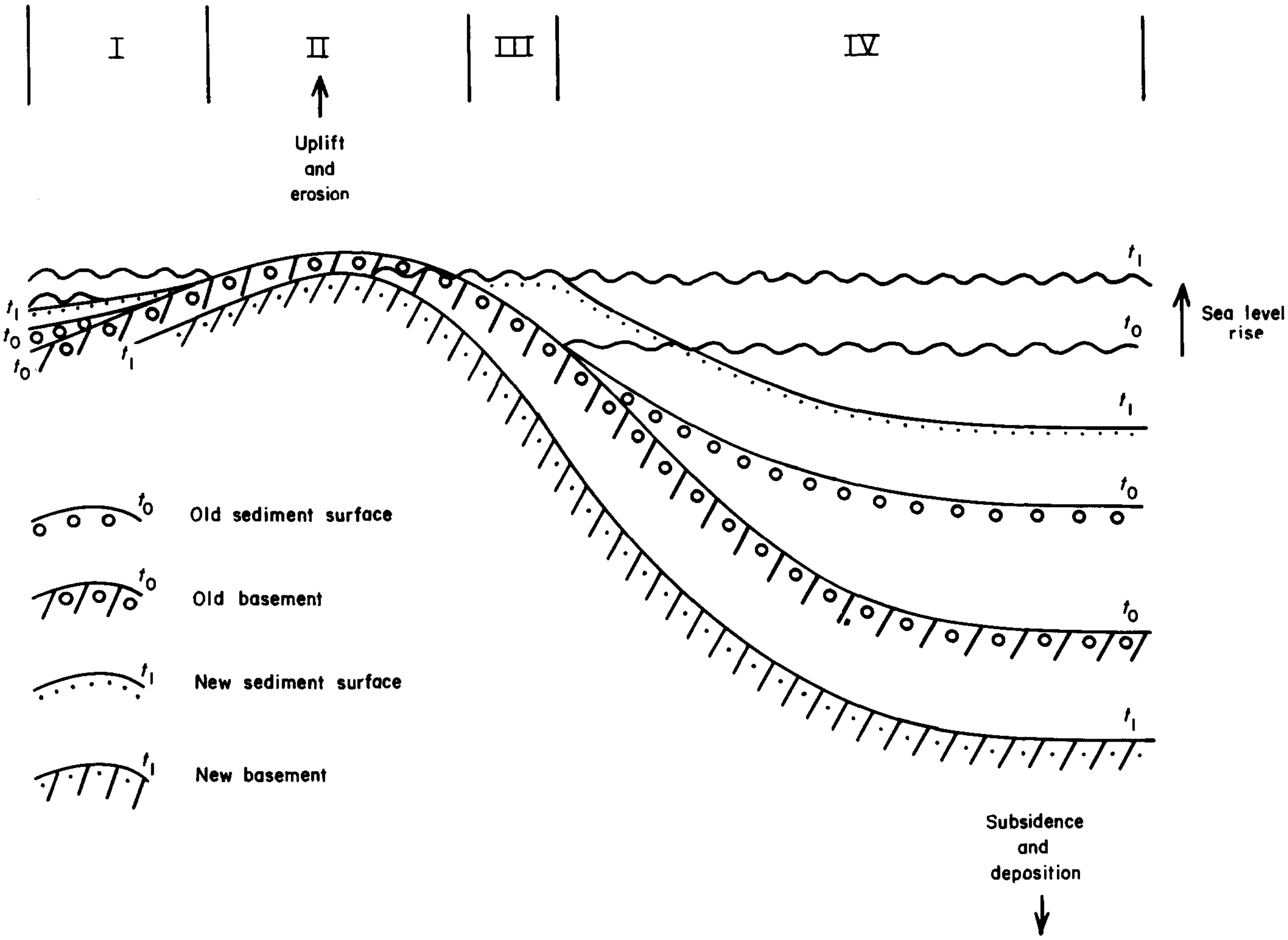

Fig. A1 Schematic diagram illustrating how program SUBSID2DC considers erosion and deposition for the case of a sea level rise. A sea level rise from time t_0 to t_1 results in the following regions of erosion and deposition. I, Infilled by sediments of variable thickness; II, not infilled by sediments; III, same as I; and IV, infilled by sediments of constant thickness. In the flexure calculation the region of sediment infill between t_0 and t_1 constitutes the 'load' while the region of the basement depression between t_0 and t_1 is the 'depression'. In the erosion calculation the height of sea level at t_0 defines the denudation 'load' at t_1 Erosion is calculated by treating this load as an upward force on an elastic plate. A uniform density, ρ_s, is assumed for the material of the load and the infill