Earth and Planetary Science Letters, 41 (1978) 1–13

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SUBSIDENCE OF THE ATLANTIC-TYPE CONTINENTAL MARGIN OFF NEW YORK

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Received January 18, 1978
Revised version received May 26, 1978

The continental margin off New York consists of a thick sequence of at least 4.8 km of mainly shallow-water sediments. Such large thicknesses of shallow-water sediments cannot be produced by the effects of sediment loading alone. We have used biostratigraphic data from the COST B-2 well to examine the origin of the subsidence of this margin. The contribution of sediment loading to the subsidence has been evaluated and removed. Corrections for effects of compaction, water depth and changes of sea-level were also included. The remaining subsidence has been interpreted in terms of a simple thermal model for the cooling lithosphere. Based on this model a thickness of the thermal lithosphere of 113–139 km is estimated. The total subsidence and crustal thinning beneath the COST B-2 well are also estimated and used to place constraints on models for the origin of the margin.

1. Introduction

The Atlantic-type continental margin of the East Coast, U.S., is characterized by a substantial thickness of seaward-dipping Mesozoic and Tertiary sediments. Early seismic refraction measurements [1,2] revealed a high-velocity refractor (>5.0 km/s) at dephts of 5–6 km beneath the outer continental shelf. This refractor had been variously interpreted as either crystalline basement or high-density sedimentary rocks [1–3]. Recent multi-channel seismic reflection profiles, however, have shown prominent reflections at greater depths, and Schlee et al. [4] estimate that at least 12 km of sediments underlie the continental shelf off New York.

In March 1976 the first deep stratigraphic well, known as the COST B-2 well, was drilled off the East Coast, U.S., 150 km east of Atlantic City, New Jersey ([5,6]; Fig. 1). The well penetrated a sequence of mostly Mesozoic and Tertiary sands and shales to a depth of 4.8 km. Biostratigraphic data [5] have shown that most of these sediments were deposited in shallow-water environments.

Such large thicknesses of shallow-water sediments cannot be produced by the effects of sediment loading. In order to accumulate large thicknesses of sediment loading alone, large water depths are required. In the local isostatic or Airy model [7], for every 1 km of sediment deposited, the basement subsides about 600 m. Thus the total sediment accumulation is limited to about $2\frac{1}{2}$ times the water depth. Even in a flexural model [8] where the lateral strength of the crust induces subsidence, large water depths are still required near the shelf edge.

In order to explain the large thicknesses of shallow-water sediments there must be other processes causing contemporaneous subsidence. The origin of the subsidence at Atlantic-type margins has been discussed by a number of authors. Although most hypotheses appeal to one or more factors they may be divided into three main groups:

(1) Thermal cooling of the lithosphere following uplift and subaerial erosion [9–11], or subcrustal thinning [12] at the time of initial rifting.

(2) Crustal stretching or "necking" due either to regional extension [13], or to differential loading at a margin [14,15].

Lamont-Doherty Geological Observatory Contribution No. 2685.

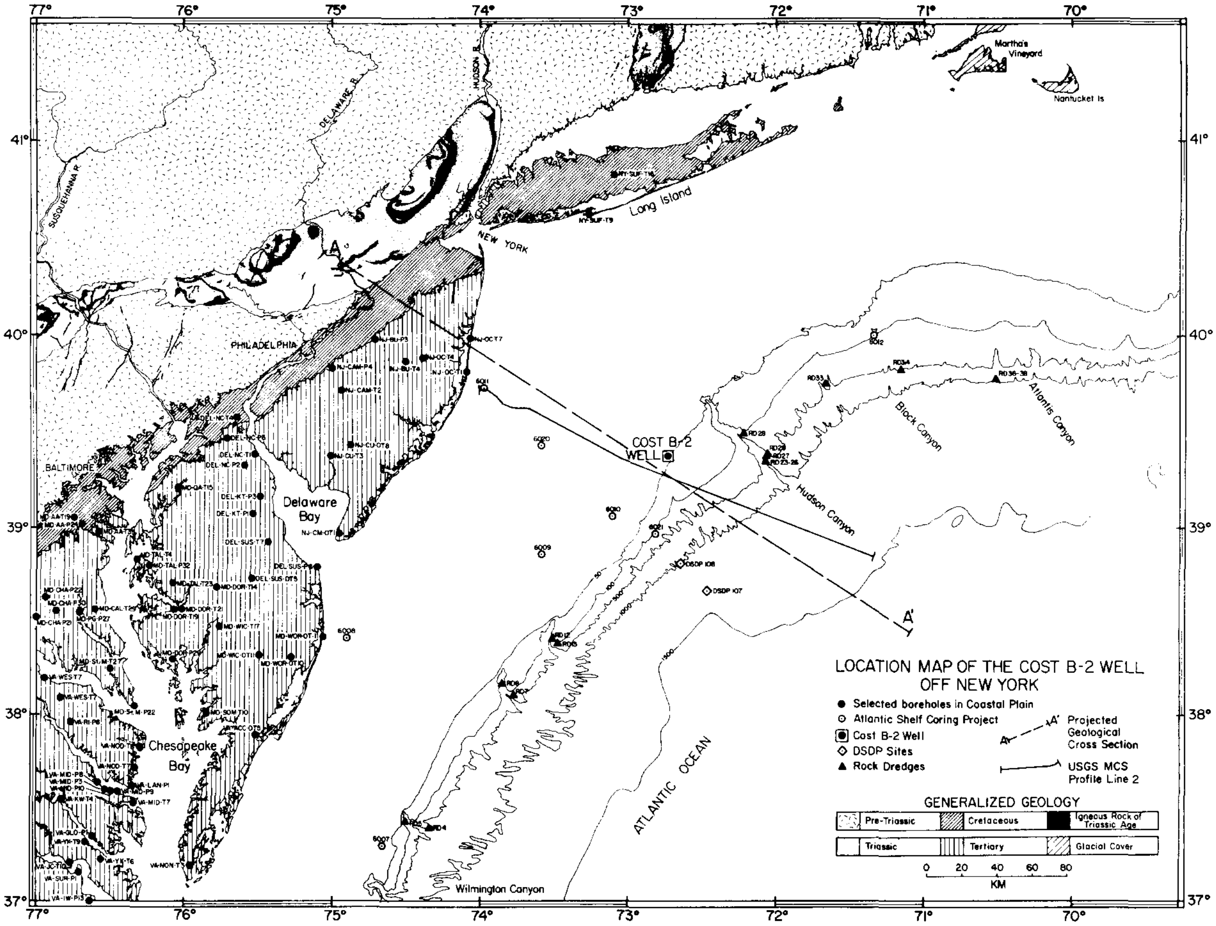


Fig. 1. Location map of the COST B-2 well [6] on the outer continental shelf off New York. The location of other boreholes is based on Brown et al. [33] for the coastal plain and Hathaway et al. [34] for the Atlantic shelf and slope. The location and identification of the rock dredges are from a compilation by Gibson et al. [35].

(3) Deep crustal metamorphism leading to an increase in overall density of the crust [16,17].

Although it is presently not clear which of these factors is the main cause of the subsidence, a useful approach to the problem is to account quantitatively for the effect of sediment loading. In this way, that part of the subsidence which is not caused by the weight of the sediments can be isolated. Sleep [9] corrected for the effect of sediment loading using stratigraphic data from wells in the coastal plain of the U.S. and showed that the subsidence was exponential in form. Watts and Ryan [18] used biostratigraphic data from the western Mediterranean and showed that the subsidence not caused by sediment loading generally followed the form of the empirical ocean ridge curve.

The purpose of this paper is to present the results of a quantitative analysis of biostratigraphic data from the COST B-2 well and to use these results to understand the origin of the subsidence of the continental margin off New York. The COST B-2 well provides an important new data base with which to carry out these studies: (1) it is the first deep stratigraphic well on the outer continental shelf off the East Coast, U.S.; (2) the stratigraphic data are supplemented by detailed geophysical logging and paleontologic analysis; and (3) the well was drilled on a relatively old continental margin near the ocean/continent bound-

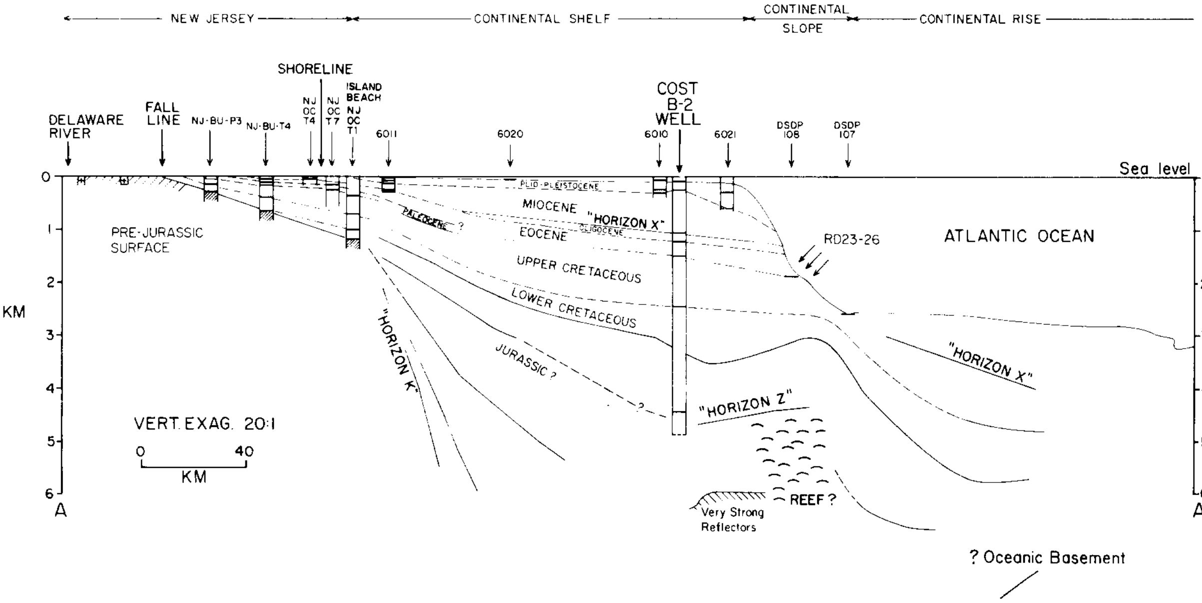


Fig. 2. Schematic cross-section of the continental shelf and margin off New York along profile A-A' (Fig. 1). The identification of sedimentary horizons is based mainly on the COST B-2 well and other boreholes in the shelf and margin. The continuity of sedimentary horizons between boreholes is based on the nearby multichannel Line 2 ([4]; Fig. 1). The solid lines in the cross-section are based on prominent seismic horizons summarized in fig. 11 of Schlee et al. [4].

ary and therefore contains a valuable record of the subsidence following rifting of the margin.

2. Analysis of data

To remove the effects of the sedimentary loading and isolate the origin of the subsidence two procedures were followed. First, biostratigraphic data from the COST B-2 well were used to reconstruct the sedimentary section as it occurred during the development of the margin. Second, using simple models of sediment unloading the margin is "backstripped" and the depth to basement without the sediment load calculated.

The information required to reconstruct the sedimentary section is shown schematically in Fig. 3. The column consists of a sediment layer of thickness S^* overlain by a water layer of thickness W_d. The column lies below the former sea-level which differs from present day by Δ_{SL}. In this study the long-term eustatic component of sea-level determined by Pitman [20] and Vail et al. [19] is used for Δ_{SL} (Fig. 3).

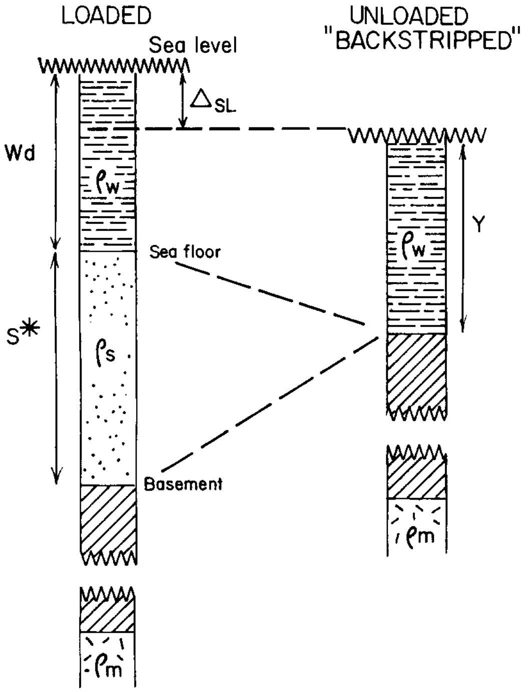


Fig. 3. Schematic diagram of a reconstructed (loaded) sedimentary section and a "backstripped" (unloaded) sedimentary section. The parameters are defined in the text.

TABLE 1

Summary of total sediment thicknesses, porosity, water depth and sea-level change through time for the COST B-2 well off New York

	Age (m.y. B.P.)	Horizon depth (ft)	Total sediment thickness (ft)	Porosity (%)	Corrected sediment thickness (ft)	Water depth (ft)	Sea-level change (ft)	Sea-level correction (ft)	Computed basement depth					
									local loading (ft)		flexural loading			
											present-day shelf (ft)		eroded shelf (ft)	
Plio-Pleistocene	0	0	42,000	54.0	42,000	0–90	302	433	16,866	16,956	20,047	20,137	same as present-day shelf	
U. Miocene	5	522	41,478	52.5	41,734	0–50	302	433	16,769	16,819	19,895	19,945		
M. Miocene	11.4	784	41,216	46.0	41,597	0–300	272	390	16,762	17,062	19,848	20,148		
L. Miocene	15.7	3169	38,831	41.0	40,148	3–600	272	390	16,547	16,847	19,413	19,713		
U. Oligocene	22.5	3208	38,792	39.5	40,159	3–600	374	522	16,406	16,706	19,249	19,549		
L. Oligocene	32.5	3623	38,377	38.5	39,885	3–600	496	715	16,113	16,413	18,883	19,183		
U. Eocene	38	3694	38,306	38.0	39,837	3–600	618	889	15,922	16,222	18,669	18,969	17,921	18,221
M. Eocene	44	3816	38,184	37.5	39,755	3–1500	692	993	15,788	16,988	18,535	19,735	17,777	18,977
L. Eocene	49	4241	37,759	36.5	39,465	3–1500	804	1153	15,522	16,722	18,239	19,439	17,480	18,680
Paleocene	55	4576	37,424	35.5	39,233	3–1500	866	1242	15,349	16,549	18,044	19,244	17,278	18,478
Maastrichtian	65	4612	37,388	35.0	39,208	3–1500	1089	1562	15,020	16,220	17,733	18,933	16,949	18,149
Campanian	72.5	4712	37,288	34.0	39,137	3–1500	1144	1642	14,914	16,114	17,662	18,862	16,876	18,076
Santonian	80	5348	36,652	32.0	38,680	3–1500	1144	1642	14,748	15,948	17,473	18,673	16,693	17,893
Sant.-Con.	83	5675	36,325	29.5	38.438	0–50	1133	1628	14,373	14,423	17,050	17,100	16,276	16,326
Turonian	88	6609	35,391	26.0	37,678	3–600	1059	1534	14,490	14,790	17,112	17,412	16,354	16,654
Cenomanian	94	7222	34,778	24.0	37,235	3–600	988	1402	14,461	14,761	17,049	17,349	16,310	16,610
Albian	100	7742	34,258	22.0	36,807	0–600	833	1195	14,212	14,812	16,771	17,371	16,023	16,623
Aptian	106	8512	33,488	19.0	36,158	0–300	642	922	14,249	14,549	16,762	17,062	15,997	16,297
Barremian	112	9612	32,388	16.0	35,196	0–300	491	706	14,114	14,414	16,528	16,828	15,755	16,055
Haut.-Val.	118	10,612	31,388	13.0	34,290	0–300	409	588	13,902	14,202	16,210	16,510	15,445	15,745
Berriasian	130	11,612	30,388	10.0	33,353	0–50	376	541	13,608	13,658	15,879	15,929	15,063	15,113
Jurassic (?)	135	14,262	27,738	7.5	30,786	0–50	376	541	12,673	12,723	14,575	14,625	13,806	13,856
Base of B-2	134–141	15,655	26,345	7.0	29,400	0–50	219–376	325–541	12,168	12,326	13,892	14,050	13,243	13,401
Basement	195 (?)	42,000	0	–	0	0	0–75	0–108	0	108	0	108	0	108

This sea-level curve which has been calibrated by Pitman [20] based on the volume of mid-oceanic ridges reaches a maximum amplitude of about 350 m in the Late Cretaceous (Table 1). The depth of the water layer W_d, at the COST B-2 well is obtained from palaeoenvironmental analysis by Smith et al. [5]. This analysis was based on benthonic foraminifera and the presence or absence of dinoflagellates.

The sediment layer S^* (Fig. 3) represents the sediment layer as it existed at some former time. Due to the effect of compaction it does not correspond to the thickness observed in the well. As sediments are progressively buried, pore water is expelled and the layers decrease in overall thickness. In order to evaluate the size of this effect in the well, we utilize porosity values determined from down-hole density and sonic logs [21]. A smooth curve representing the variation of porosity with depth is obtained from these values (Fig. 4). We assume that independent of their lithology, individual sediment layers follow this porosity-depth curve as the overburden pressure increases. We also assume that sediment layers beneath the base of the well have been fully compacted. The procedure then used is to remove the overburden of

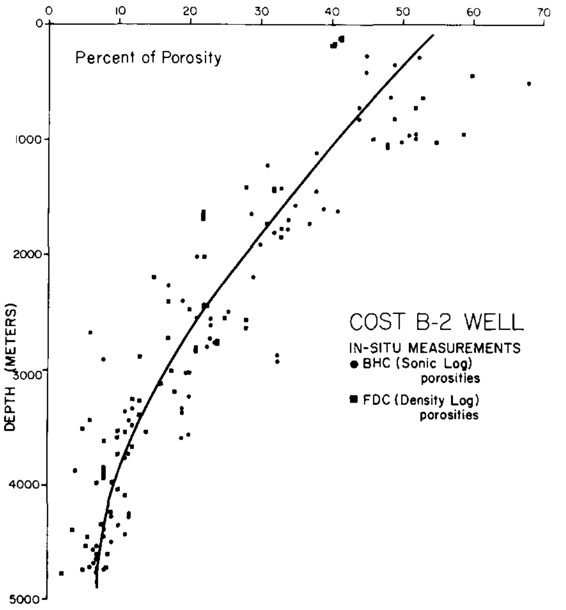


Fig. 4. Porosity-depth curve for the COST B-2 well based on data tabulated by Rhodehamel [21]. The heavy line indicates a "smooth" curve assumed in the computation of the reconstructed sedimentary section.

younger sediments and allow the underlying sediments to slide up the curve to their former position. The total sediment thickness obtained is S^*, which is the thickness the sediments possessed at the former time. For these calculations the total present sediment thickness is assumed to be 12.8 km (42,000 ft) based on the multi-channel seismic profile Line 2 [22]. The values used for the data reduction are listed in Table 1.

"Backstripping" (Fig. 3) involves the removal of sediment loads and also of water loads due to changes in sea-level. The result of "backstripping" is the calculation of Y (Fig. 3), the depth to basement in the absence of surface loads. We use simple models in which the lithosphere responds to sediment loads either by local loading of the Airy-type or flexural loading of a thin elastic plate.

In the local loading model the lithosphere responds only to the load immediately above it and "backstripping" corresponds to a simple balancing of mass columns. Following Fig. 3, Y is then given by:

$$Y = S^*\left[\frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)}\right] + W_d - \Delta_{SL}\left[\frac{\rho_m}{(\rho_m - \rho_w)}\right] \qquad (1)$$

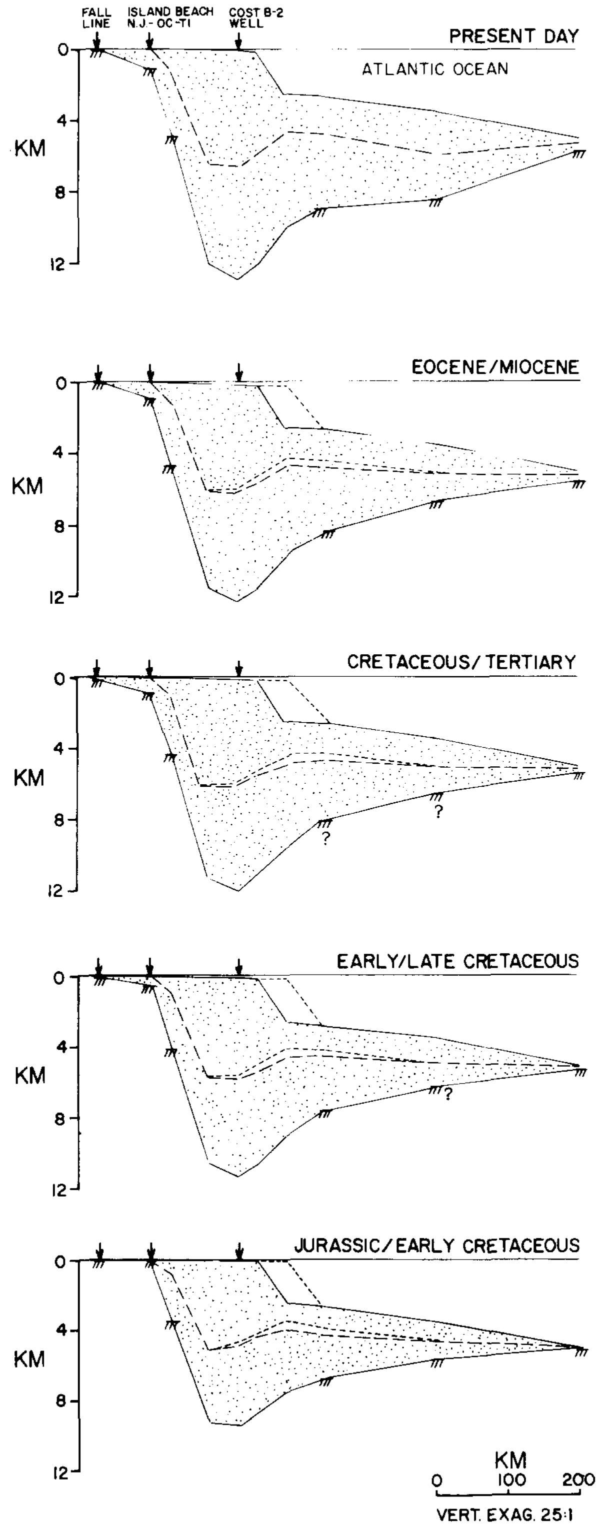
where ρ_m is the mean mantle density, ρ_w is the mean water density and ρ_s is the mean saturated sediment density. The mean sediment density is given by:

$$\rho_s = \frac{\sum_i [\phi_i \rho_w + (1 - \phi_i)\rho_g]\, T_i}{S^*} \qquad (2)$$

where T_i is the interval thickness, ϕ_i is the interval porosity and ρ_g is the grain density. We assume a constant grain density of 2.65 g/cm^3 throughout the section. The average grain density of the stratigraphic intervals does not vary by more than a few percent in the COST B-2 well [5, table 1].

In the flexural model the lithosphere is assumed to have lateral strength and the depression of the basement surface depends not only on the load above it but also on the surrounding load. "Backstripping" in the flexural model therefore requires knowledge of the sedimentary section on either side of the COST B-2 well as well as the flexural rigidity of the underlying basement.

We then construct sedimentary sections of the margin (Fig. 5) using the COST B-2 well, the Island Beach well [23] and DSDP sites 105, 107 and 108 [24]. The stratigraphy of the section between the



drill sites is interpolated using data from the nearby multi-channel Line 2 ([4]; Figs. 1, 2). Each point used in the computation is corrected for comparison and for density variations based on the porosity-depth curve for the COST B-2 well (Fig. 4).

Schlee et al. [4] have suggested that the shelf edge was at one time further seaward as is evidenced by the truncation of prominent seismic horizons at the continental slope and upper part of the rise (Fig. 2). To allow for this possibility we use two flexural models: one in which the shelf edge maintained its present position through time and one in which a once more extensive shelf was eroded in the Late Eocene to its present position. In both these models the load is comprised not only of sediments but also of water due to changes in sea-level. Except near the shoreline, the load due to sea-level fluctuations is constant and therefore a local loading model is used for the water load.

Studies of the deformation of the lithosphere caused by long-term surface loads show that there may be a difference between the flexural rigidities of the continental and oceanic lithosphere [8,25]. At the present time, however, uncertainties as to the nature of the ocean/continent transition preclude the use of more complex models with lateral variations in elastic properties. We therefore use a simple continuous elastic plate model with a flexural rigidity of 1×10^{30} dyne-cm, a value generally intermediate to those obtained from continents and oceans.

We summarize the results of "backstripping" sediments for the local and flexural loading models at the COST B-2 well in Fig. 6. Several aspects of the "backstripping" method are apparent in this figure. The most striking result is that although sediment loading contributes in a major way to the observed subsidence there is a large residual subsidence due to other causes. We refer to this residual subsidence as the "tectonic driving force" associated with the margin.

Fig. 5. Reconstructed sedimentary sections of the continental shelf and margin for the present day and earlier times. The question marks indicate uncertain depths and the dotted area indicates the total sediment thickness at a particular time. The long-dashed line is computed depth to basement without the sediment load assuming the shelf was constant through time. The short-dashed line indicates the depth assuming the shelf extended further seaward prior to the Late Eocene.

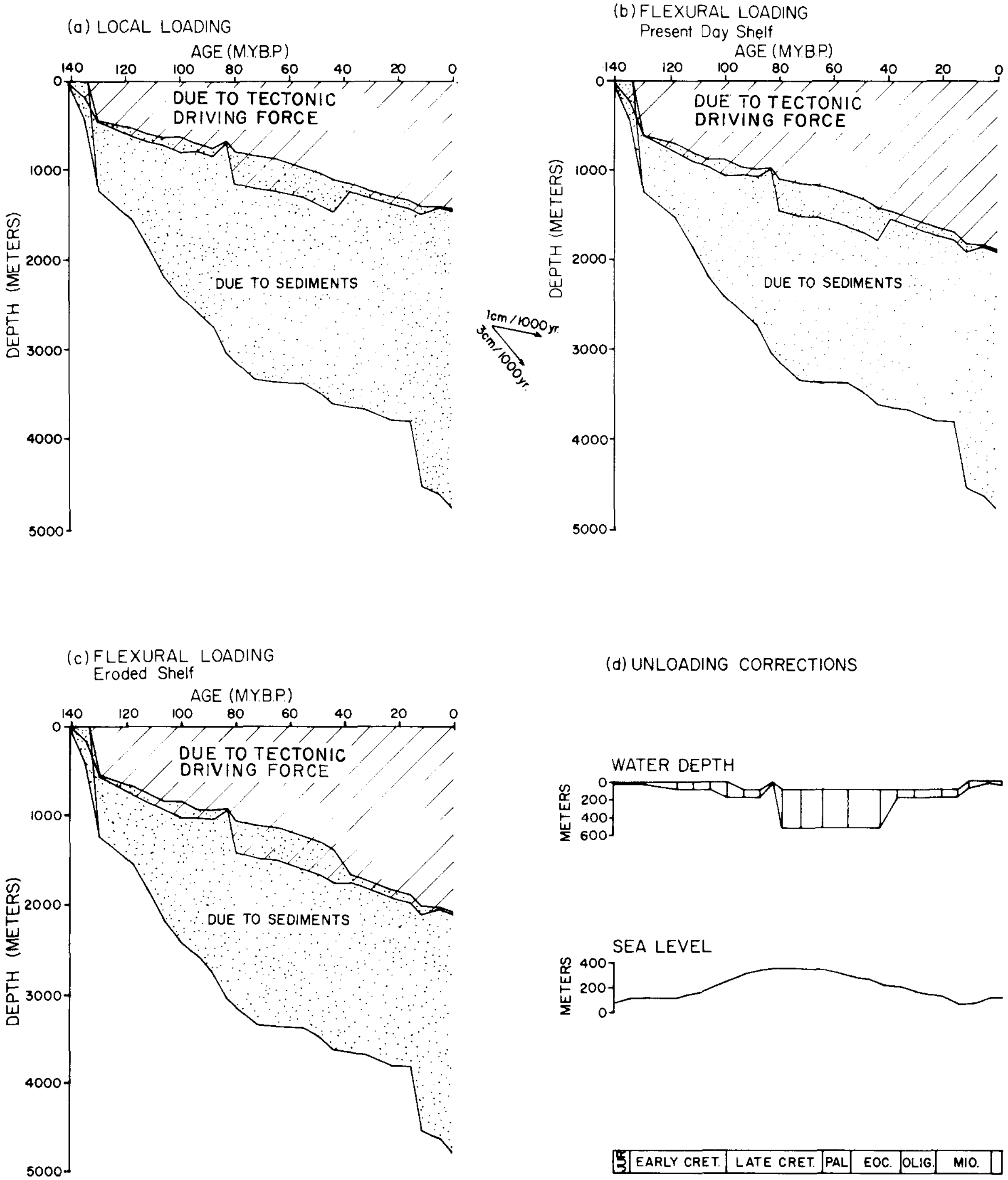


Fig. 6. Summary of subsidence curves for the COST B-2 well for three models of "backstripping" sediments from the margin. In each model the dotted area indicates that part of the observed subsidence which can be attributed to sediment loading and the cross-hatched area indicates that part which can be attributed to other "tectonic driving forces". Fig. 6d is a plot of the water depth and sea-level corrections used to "backstrip" the sediments from the margin.

The magnitude of the "tectonic driving force" is larger for the flexural models than for the local loading model. This is because the response of the lithosphere to surface loads in the flexural models extends over a broad area of the margin. A remarkable feature of the "tectonic driving force" for each model is the smoothness of the curve as compared to the observed subsidence. The rapid subsidence observed during the Miocene is nearly eliminated from the "tectonic driving force". This is due to the unconsolidated nature of the overlying sediments and the decrease in water depth due to a rapid influx of sediments. The sharp change in the rate of the observed subsidence at about 80 m.y. B.P. is also nearly eliminated from the "tectonic driving force". This is the result of changes in the sedimentation rate caused by the peaking of sea-level in the Late Cretaceous and its subsequent fall.

3. Interpretation and results

Although a number of hypotheses have been proposed for the origin of Atlantic-type margins only models in which the subsidence is thermally driven [9–11,13] are easily testable. It is difficult, for example, to evaluate how the subsidence would change through time in the models proposed by Bott [14,15] and Falvey [16]. Phase change models [17] cannot yet incorporate changes in sedimentation rates.

In thermal models the subsidence of the margin is the result of contraction associated with the cooling of thinned lithosphere and crust. Subaerial erosion [9,11] and crustal stretching or "necking" [12,13] have been proposed as the causes of the thinning. In this section we compare the "tectonic driving force" isolated from biostratigraphic data in the COST B-2 well with predictions based on simple thermal models.

We use the cooling plate model which Parsons and Sclater [26] have recently shown explains the subsidence of the ocean ridges to ages of at least 160 m.y. The advantage of this model is that it provides a number of simple tests of the thermal origin of the subsidence. The ocean ridge model [26] does not strictly apply to a continental margin. At a margin the entire lithosphere has probably not been heated to the solidus temperature. In this case the difference in the shape of the subsidence curve occurs primarily in the early history of the cooling and should not significantly affect the subsidence pattern for the sedimentary record in the COST B-2 well.

As pointed out by Parsons and Sclater [26] the subsidence curve for the cooling plate model follows a relatively simple pattern. During the early cooling history the subsidence is similar to a cooling half-space and the depth of the ocean floor is proportional to $t^{1/2}$. Later during the cooling history, the subsidence is arrested and the ocean floor depth exponentially decays to a steady state value. This suggests the use of two simple plots to examine the subsidence: depth against $t^{1/2}$ and log depth against t.

In order to apply the cooling plate model to subsidence of the margin at the COST B-2 well we need to make the following assumptions: (1) the depth to basement beneath the well is 12.8 km, based on seismic profile Line 2 [22]; (2) the sediments beneath the well were deposited at or near sea-level; and (3) the subsidence began 195 m.y. B.P., corresponding to the time of extensive basaltic igneous activity along the margins of the North Atlantic ocean at the Triassic-Liassic boundary [27].

Fig. 7 shows the plots of depth of basement against the square root of age since subsidence began for the COST B-2 well. The linear part of the depth against $(\text{age})^{1/2}$ plot is poorly constrained. Data from the COST B-2 well penetrates only a small part of the linear portion of the plot and both the age of initial subsidence and the total sediment thickness are uncertain. However, even allowing reasonable variations in these parameters does not alter the observation that the "tectonic driving force" deviates from a straight line. We can, therefore, estimate values for the slope of the straight line, C_1, and the time of departure from this straight line, C_2.

Fig. 8 shows the plots of log depth against age. The parameter D_0 (Fig. 8) is equivalent to C_3 of Parsons and Sclater [26] and represents the asymptotic value of the subsidence. The y-axis of this plot is the log of normalized elevation above this value. The amplitude of the exponential portion of the subsidence, C_3' of Parsons and Sclater [26], has a value 8 D_0/π^2, giving the value of the intercept on the y-axis. The range of D_0 we consider acceptable are those in which this value fell within 5 m.y. of the assumed

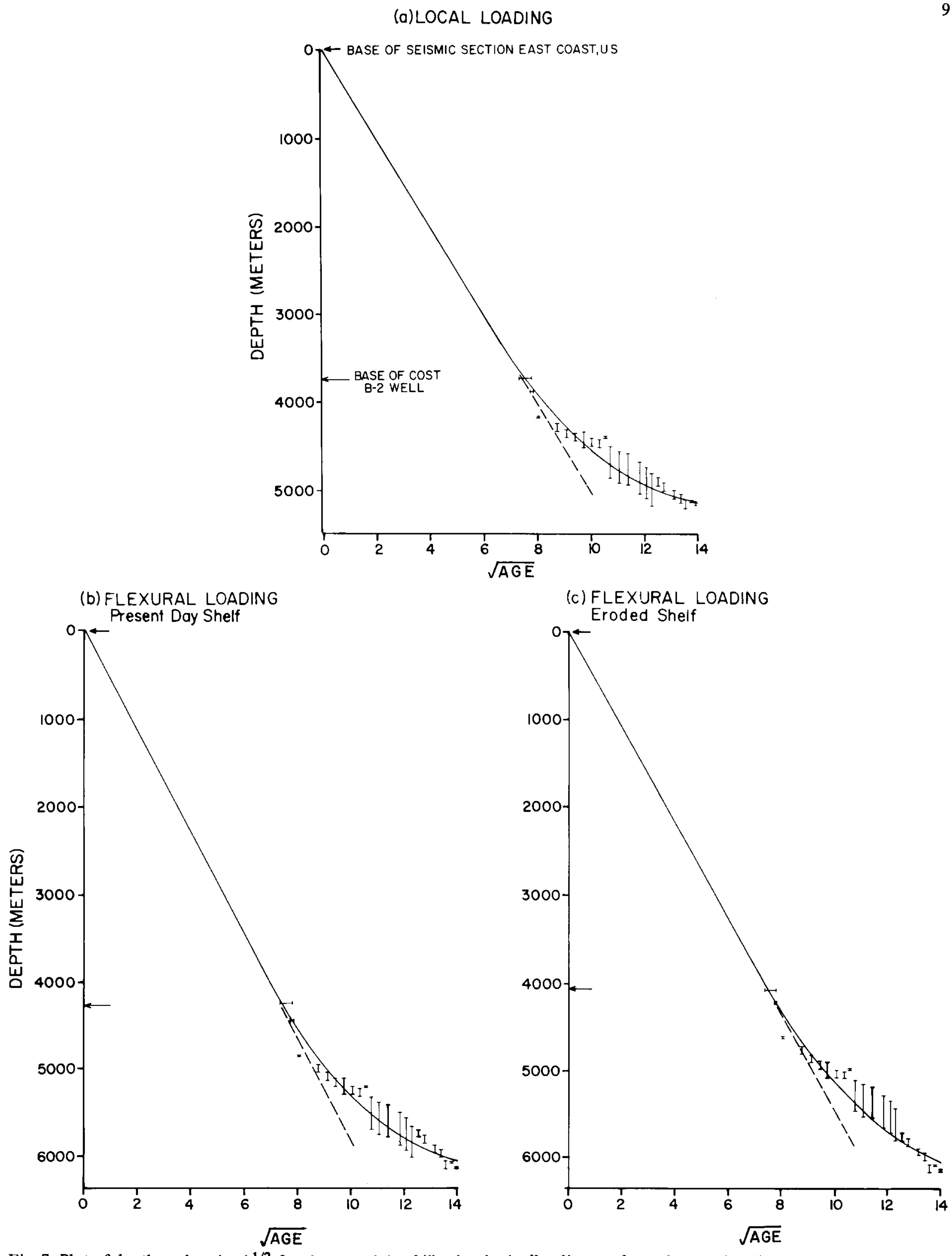


Fig. 7. Plot of depth against $(\text{age})^{1/2}$ for three models of "backstripping" sediments from the margin. The age assumed for the start of the subsidence is 195 m.y. B.P. [27]. The vertical bars reflect the range of water depth.

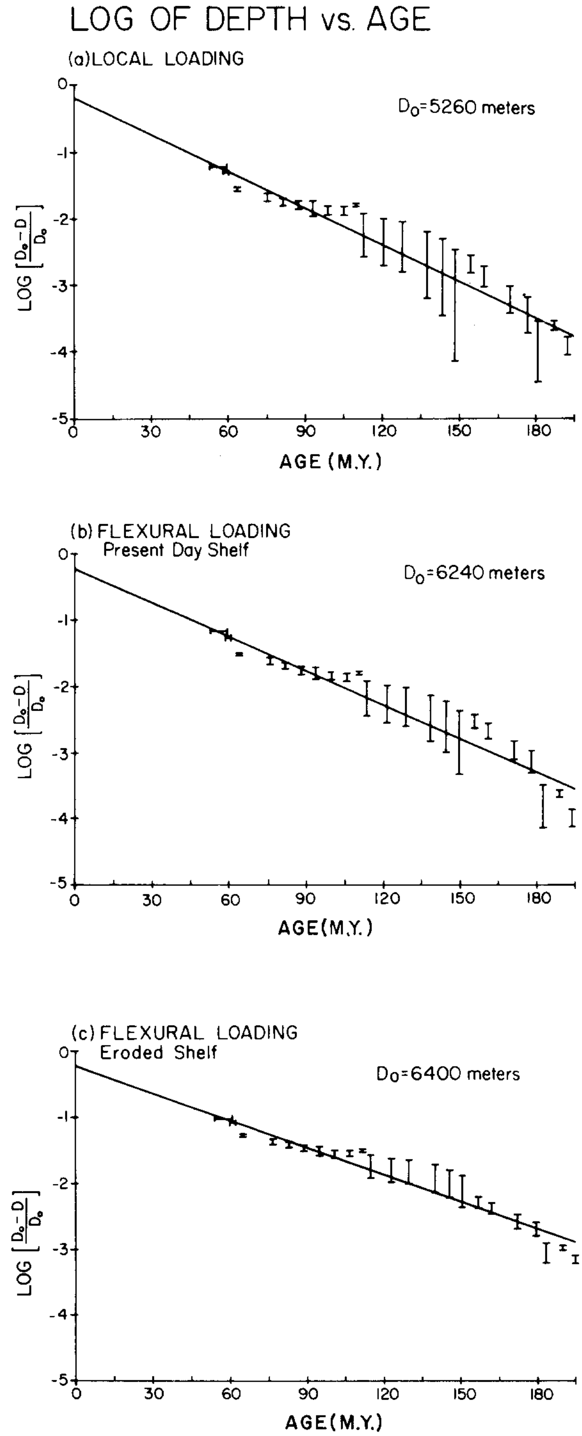


Fig. 8. Plot of $\log(D_0 - D)/D_0$ against age for three models of "backstripping" sediments from the margin. D is the reconstructed depth and D_0 is the asymptotic value of D. The vertical bars reflect the range of water depth.

age of 195 m.y. B.P. The straight line in these plots is determined by a least squares fit to the data for each value of D_0. The slope of this line gives an estimate of the time constant of the exponential term, C_2'.

The estimates of the parameters from the plots in Figs. 7 and 8 are summarized in Table 2. These parameters can be used to obtain independent estimates of the thermal thickness of the lithosphere. The thickness of the lithosphere can be determined from the ratio of the asymptotic value to the rate of subsidence, the age at which the subsidence departs from the cooling half-space model, and the time constant of the exponential portion of the curve. There is good agreement between these different estimates (Table 2). The range of estimates of the thickness is 113–123 (local loading), 115–125 (flexural loading with present-day shelf) and 125–139 km (flexural loading with eroded shelf). The subsidence predicted by the mean estimate for each model of sediment loading is shown as a solid line in Fig. 7.

These estimates of the thermal thickness of the lithosphere depend on the simple cooling plate model and the previously stated assumptions. Reasonable variations (about 2 km) in the assumed total sediment thickness will cause variations in D_0 which are less than those obtained between the sediment loading models. The presence of deep-water sediments beneath the well would correspond to a greater initial depth to basement and a smaller estimate of D_0. The maximum initial depth for the basement would be similar to the depths of mid-ocean ridge crests (about 2.5 km). We have, however, constructed models in which the subsidence is similar to normal mid-ocean ridges and found that we could not satisfactorily fit the well data. The age for the initial subsidence may be less than the 195 m.y. B.P. we are assuming. Younger ages would not significantly effect the estimate of D_0 but would decrease the estimates of the thermal thickness.

The calculated "tectonic driving force" generally agrees well with the observed data from the COST B-2 well (Figs. 7, 8). The local loading model and the flexural loading model with the present-day shelf give similar estimates for the thermal thickness (Table 2). The main difference is the large "tectonic driving force" for the flexural model. This observation places important constraints on the origin of the subsidence. The addition of a large shelf prior to erosion in the

TABLE 2

Thermal parameters deduced in the computations

Constant	Estimates			Thickness estimate a (km)		
Local loading						
C_1 m/(m.y.)$^{1/2}$	483	497	516	114	119	123
C_3 meters	5243	5260	5289			
C_2' m.y.	52	55	60	113	116	121
C_2 m.y.	(55)	(64)	(75)	(112)	(121)	(131)
Flexural loading (present-day shelf)						
C_1 m/(m.y.)$^{1/2}$	563	577	603	116	122	125
C_3 meters	6203	6240	6279			
C_2' m.y.	54	59	64	115	120	125
C_2 m.y.	(60)	(72)	(85)	(117)	(128)	(139)
Flexural loading (eroded shelf)						
C_1 m/(m.y.)$^{1/2}$	537	543	571	125	133	135
C_3 meters	6340	6400	6462			
C_2' m.y.	67	73	79	129	134	139
C_2 m.y.	(70)	(80)	(90)	(126)	(135)	(143)

Definition of constants (after Parsons and Sclater [26]):

$C_1 = \dfrac{2\rho_0\alpha T_1}{(\rho_0 - \rho_w)}\left(\dfrac{K}{\pi}\right)^{1/2}$ slope of depth versus (age)$^{1/2}$

$C_3 = \rho_0 a\alpha T_1/2(\rho_0 - \rho_w)$ intercept of depth versus (age)$^{1/2}$

$C_2 = a^2/9K$ breakdown of linear (age)$^{1/2}$ relation

$C_2' = \pi^2 K \log_{10} e/a^2$ slope of log (elevation) versus age

Late Eocene in the flexural model gives the largest estimate of the thickness and appears to give a somewhat better fit to the data.

Although we have not tested non-thermal models we believe that the consistency in the estimates for the equilibrium thickness is good evidence that the "tectonic driving force" is thermal in origin. Thus, the subsidence at the COST B-2 well can be explained by a simple thermal model of cooling of an initially thinned lithosphere and crust subsequently loaded by sediments.

4. Discussion

We determine estimates of the total subsidence of the margin, D_0, from the subsidence history at the COST B-2 well. Although we use an ocean ridge-type model, these estimates should not change significantly for other thermal models. The importance of D_0 is that it is determined by the amount of initial crustal thinning.

A minimum estimate for the amount of crustal thinning beneath the COST B-2 well can be obtained assuming an Airy isostatic model. If the depth of compensation is at the base of the original crust, the thinning is given by:

$$T = \frac{D_0(\rho_m - \rho_w)}{(\rho_m - \rho_c)} \qquad (3)$$

where ρ_c = average crustal density and the other parameters are as previously given. For the values of D_0 determined (Fig. 8) this gives estimates of the amount of crustal thinning which range from 20.6 km for local loading to a minimum of 24.9 km for the flexural model with eroded shelf. If the original crustal thickness is similar to the 30–40 km found beneath the eastern Appalachians [28] then these results imply that a substantial proportion of the crust has been thinned. Any model for the origin of

Atlantic-type margins must be able to produce these amounts of thinning.

There are two main hypotheses which have been proposed to explain the thinning at these margins: sub-aerial erosion and crustal stretching.

Sub-aerial erosion. In this hypothesis the crust is domed over a broad region as a result of thermal expansion and is thinned by sub-aerial erosion of the uplifted region [9,11]. The maximum amount of crustal thinning is obtained for instantaneous erosion of the uplifted region [29, fig. 2] and is approximately given by:

$$T_{sa} = (T_L - T_C)\left[\frac{\rho_m - \rho_m'}{\rho_m - \rho_c}\right] \tag{4}$$

where T_L = the original thickness of the lithosphere and ρ_m' = average density of heated lithosphere. With $(\rho_m - \rho_m') = 0.08$ g/cm^3 [30] and $T_L = 150$ km which corresponds to a surface heat flow of 1.4 HFU in eastern U.S.A., we get $T_{sa} = 18.5$ ($T_c = 30$ km) and $T_{sa} = 16.9$ km ($T_c = 40$ km). Comparing these results with those computed from D_0 suggests that sub-aerial erosion cannot fully explain the amount of crustal thinning observed beneath the COST B-2 well.

Crustal stretching. In this hypothesis the crust is extended during the early stages of rifting by regional tension [12,13]. To estimate the amount of extension we need to know how D_0 varies across the margin. Based on estimates of D_0 from the flexural models (Fig. 5) the extended area occurs as far landward as the Island Beach well. Thus, the amount of extension at the margin is at least a factor of 2 and locally beneath the COST B-2 well may exceed a factor of 5. The amount of extension that occurs at a rifting margin is difficult to determine. However, if this extension occurs during the 20-m.y. period prior to the initial subsidence at 195 m.y. B.P. [27] the average strain rate is on the order of 10^{-14} to 10^{-15} s^{-1}, which is of the same order of magnitude as geological processes such as shortening in orogenic belts [32].

5. Conclusions

(1) Biostratigraphic data from the COST B-2 well have been used to determine the contribution of sediment loading to the subsidence of the margin off New York and to isolate that part of the subsidence due to other forces.

(2) These other forces can be explained by a simple thermal model of a cooling lithosphere. This model gives an estimate for the thermal thickness of the lithosphere beneath the well in the range 113–139 km.

(3) The asymptotic value of the subsidence, D_0, gives an estimate of the amount of crustal thinning that has taken place at the margin. A substantial amount of thinning beneath the well must have taken place and lies in the range 20–25 km.

(4) Isostatic considerations suggest this amount of crustal thinning cannot be caused by sub-aerial erosion alone and other processes must be involved. The most likely other process involved is lithospheric extension during the initial phase of rifting.

Acknowledgements

T. Crough, M. Langseth, D. McKenzie, W. Pitman, W. Ryan and N. Sleep made a number of helpful suggestions and comments. This work was supported by National Science Foundation grant NSF OCE-77-10647.

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