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Subsidence and tectonics of Atlantic-type continental margins

Atlantic continental margin
Backstripping studies
Tectonic subsidence
Stretching

Marge continentale atlantique
Suppression de l'effet des sédiments
Subsidence tectonique
Étirement

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ABSTRACT

The sediments which accumulate at a continental margin following initial rifting represent a load on the lithosphere which should sag due to its weight. Backstripping studies in which sediment, as well as water, loads are removed from a margin during different geological intervals of time have now been carried out at both relatively young and old continental margins. These studies show that a number of factors affect the subsidence and tectonics of margins which include eustasy, the flexural strength of the basement, compaction, and palaeobathymetry. These factors cannot, however, account for all the observed features of the subsidence and tectonics of margins and other processes must be involved. The most likely other contributor to the observed subsidence is thermal contraction, following stretching of the margin at the time of initial rifting. Simple stretching models appear to be able to explain the exponential character of the tectonic subsidence of margins although a number of problems still remain. The most important of these are the relative proportion of syn and pre-rift to post-rift sediments and the amount of crustal thinning across the margin.

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RÉSUMÉ

Tectonique et subsidence des marges continentales de type atlantique.

L'accumulation des sédiments post-rift sur une marge continentale représente une charge sur la lithosphère, qui devrait fléchir sous ce poids. Des études visant à supprimer l'effet de charge due à l'eau et aux sédiments ont été menées tant sur les marges anciennes que sur les marges relativement jeunes.

Ces études montrent qu'un certain nombre de facteurs jouent sur la tectonique et la subsidence des marges, parmi lesquels l'eustasie, la flexure du substratum, la compaction et la paléobathymétrie. Ces facteurs ne peuvent expliquer néanmoins l'ensemble des caractéristiques de la subsidence des marges continentales, pour lesquelles il faut rechercher d'autres causes. Parmi celles-ci, la plus probable est une contraction thermique après un étirement de la marge au moment du rifting.

Des modèles d'étirement simples expliquent le caractère exponentiel de la subsidence tectonique des marges, bien qu'ils laissent un certain nombre de problèmes non résolus, le plus important étant la proportion relative de sédiments anté- et synrift par rapport à ceux déposés après le rifting et le taux d'amincissement à travers la marge.

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INTRODUCTION

The continental margins of Atlantic-type which border the Atlantic, Indian and Arctic oceans are characterized by substantial thicknesses of seaward dipping sediments. These sediments range in age from Triassic to Recent for the margin off North America and Africa (Schlee *et al.*, 1976 ; Sheridan, 1976 ; Von Rad, Arthur, 1979) to Late Tertiary to Recent for the margins of the Red Sea and Western Mediterranean (Lowell *et al.*, 1975 ; Biju-Duval *et al.*, 1974). The relatively old margins off North America and Africa, which are characterized by thicknesses in excess of > 10 km, appear to overlie a faulted basement. The relatively young margins of the Red Sea and Western Mediterranean, on the other hand, are characterized by thinner sediments (< 3 km) which overlie a faulted continental basement.

Dietz (1963) suggested that the loading of sediments on the continental slope and rise would produce a regional downwarp or flexure of the adjacent continental crust. Gunn (1944), Walcott (1972), Cochran (1973), and Watts and Ryan (1976) have quantitatively shown that substantial thicknesses of shallow and deep water sediments may accumulate at a margin simply as a result of sedimentary loading. They used simple flexural models with the lithosphere responding to loads as an elastic plate overlying a weak fluid.

Biostratigraphic data from deep commercial boreholes in the outer continental shelf off eastern North America (Scholle, 1977 ; Gradstein *et al.*, 1975 ; Jansa, Wade, 1975), Northwest Africa (Von Rad, Arthur, 1979) and the Western Mediterranean (Cravatte *et al.*, 1974) show that much of these sediments were deposited in continental or neritic to inner shelf marine environments. While large thicknesses of relatively shallow water sediments cannot be produced by sedimentary loading alone, all hypotheses for the origin of the subsidence of continental margins are in agreement that sedimentary loading must contribute to a major portion of the subsidence.

Isostatic considerations suggest that the maximum thickness of sediments which can accumulate at any time during margin evolution is about 2 1/2 times the available water depth. If flexure is important, the maximum sediment thickness will be less, but the subsidence will occur over a larger region due to the strength of the lithosphere. Thus the effect of sediment loading at a margin is to amplify and otherwise modify the subsidence due to other causes.

A useful approach to the problem of the subsidence of continental margins, therefore, is to quantitatively account for the effects of sedimentary loading during margin evolution (Watts, Ryan, 1976 ; Steckler, Watts, 1978 ; Keen, 1979). In this manner, that part of the subsidence which is not caused by the weight of the sediments can be isolated, so that any similarities in the structural features of margins that are masked by sediments can be seen.

Sleep (1971) showed that the subsidence of the US coastal plain, when corrected for the effects of sediment loading, could be fit by an exponential decay with a time constant of 50 MY Watts and Ryan (1976) and Steckler and Watts (1978) referred to the subsidence corrected for sediment loading as the tectonic subsidence. They showed that the tectonic subsidence was similar in form to that of a mid-ocean ridge.

A number of studies have now compared the tectonic subsidence at a margin to the subsidence calculated based on simple cooling models for the lithosphere. Steckler and Watts (1978), Watts and Steckler (1979) and Keen (1979), using commercial wells off eastern North America, showed the subsidence could be explained in terms of a simple cooling plate model (Parsons, Sclater, 1977).

Recently, McKenzie (1978) proposed a model in which the subsidence is caused by cooling of the lithosphere following uniform extension and thinning at the time of rifting. Christie and Sclater (1980) have shown this model may explain the post Mid-Cretaceous subsidence of the North sea basin and Royden and Keen (in press) have applied this model to well data off Labrador and eastern North America.

In order to quantitatively evaluate and remove the effects of sediment loading at a margin two procedures should be followed. First, biostratigraphic data should be used to reconstruct the sedimentary section during evolution of the margin. Second, the sedimentary section should be "backstripped" and the subsidence of the margin not caused by the weight of the sediments obtained. The information required to reconstruct the sedimentary section through time consists of sediment thicknesses corrected for the effects of compaction, water depth of sediment deposition (paleobathymetry), and the long-term eustatic component of sea-level (Van Hinte, 1978 ; Steckler, Watts, 1978). "Backstripping" (Watts, Ryan, 1976) is the removal of sedimentary and water loads through time and requires information both on the densities within the sediment column and the long-term (> 10^6 years) mechanical behavior of the lithosphere.

The purpose of this paper is to briefly review the contribution of sediment and water loading to the subsidence of Atlantic-type continental margins. We will use biostratigraphic data, mainly from the margin off eastern North America, to illustrate how factors such as sediment compaction, paleobathymetry, sea-level changes and the mechanical response of the basement contribute to the subsidence. The overall objective of the paper is to better understand the origin of the tectonic subsidence of Atlantic-type continental margins.

SEDIMENTS

The development of seismic techniques which use the multi-channel array along with commercial and DSDP/IPOD drilling have greatly improved knowledge of the sedimentary structure of Atlantic-type continental margins. There have now been numerous geophysical studies of these margins although only a few have attempted to integrate both the seismic and well data (for example, Schlee *et al.*, 1976 ; Von Rad, Arthur, 1979).

Figure 1 *a* summarizes the sedimentary structure of the continental margin off eastern North America based on available seismic and well data. Each profile of the margin (Fig. 1 *b*) shows that the coastal plain sediments gradually increase in thickness from the fall line to a hinge zone (Jansa, Wade, 1975 ; Watts, Steckler, 1979 ; Austin *et al.*, 1980) where the depth to continental basement rapidly increases. The nature of the basement seaward of the hinge zone is obscured by the large thickness of sediments. Acoustic basement, which has been identified as the upper surface of oceanic layer 2, occurs at distances as far as 100 to 300 km seaward of the hinge zone (Fig. 1 *b*).

Studies of biostratigraphic data from commercial wells off eastern North America suggests a large increase in the amount of crustal thinning occurs across the hinge zone

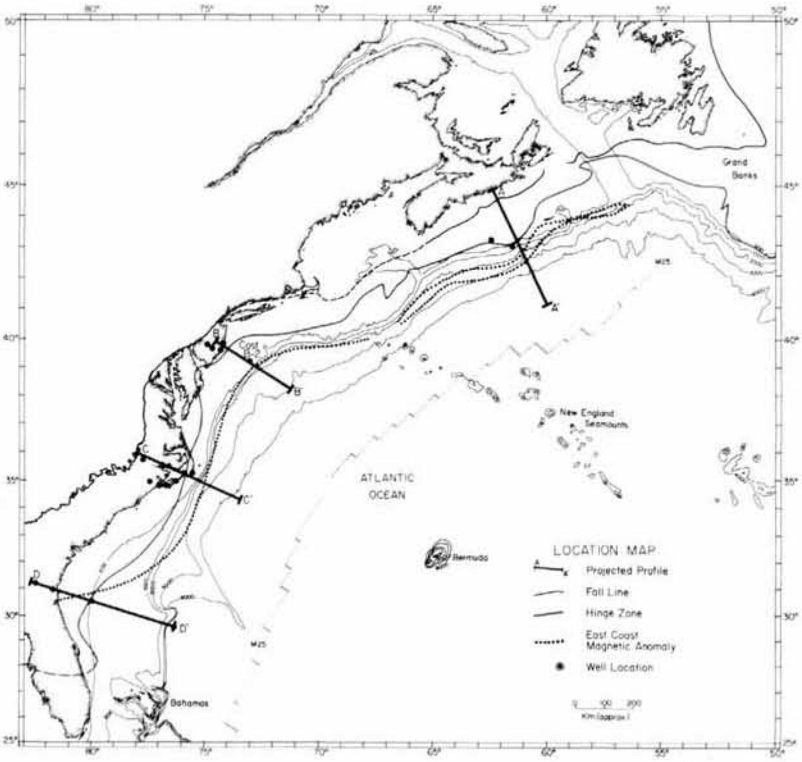

Figure 1 a

Location map of profiles off eastern North America. The position of the fall line is based on King (1969) on land and Jansa and Wade (1975) and Austin et al. *(1980) offshore. The location of the crest of the East Coast Magnetic Anomaly is from Rabinowitz (1974) and G. Karner (pers. comm.) and magnetic anomaly M-25 (155 m.y. B.P.) is from Schouten and Klitgord (1977). The position of the hinze zone is based on Maher and Applin (1971), Jansa and Wade (1975), Dillon* et al. *(1979) and Austin* et al. *(1980). Bathymetry is from Uchupi (1971). The heavy lines locate the geologic profiles in Figure 1b.*

(Watts, Steckler, 1979). Gravity and geoid modelling also confirm that at this location both the crust and thermal lithosphere thin from typical continental values to values more typical of oceanic crust (Watts, Steckler, 1979). Unfortunately, only limited seismic refraction data is presently available off eastern North America to test these models. The evidence that is available suggests the hinge zone is a major thermal and mechanical boundary and that its location (Fig. 1 *a*) along the margin has played an important role in its tectonic evolution.

Figure 1 *b* shows, however, that there are large variations in the overall distribution of sediments off eastern North America. Off New York and Florida the shelf break in slope extends seaward of the hinge zone while off Halifax and Cape Hatteras it is slightly landward of the hinge zone. The larger distance to the shelf break off New York may be due in part to the supply of terrigeneous and bioclastic sediments to the margin in the Early Jurassic, prior to the deposition of Horizon β (Tucholke, Mountain, 1979). In fact, the shelf break off New York appears to have extended even further seaward than at present and may have been eroded by 20-30 km by counter-currents which began in the middle Tertiary (Grow *et al.*, 1977).

These considerations of the sedimentary structure off eastern North America suggest that while the overall tectonic evolution of the margin is controlled by the hinge zone, the position of the shelf break and the width of the shelf depend on the availability of sediments and surficial geologic processes. Thus, these effects should be isolated and removed if the thermal and mechanical properties of the margin, and how they vary across a margin, are to be better understood.

SUBSIDENCE

The sediments which accumulate at a continental margin during and after rifting provide the best record of its tectonic evolution. The stratigraphy of the shelf region records the vertical movements which dominate the evolution of the margin. The sediment thicknesses observed at a margin are a consequence of several factors, including compaction, sea-level changes and the nature of the response of the basement to sediment and water loads. In order to obtain the tectonic subsidence at a margin, it is necessary to account for these factors.

The procedures which should be followed have been discussed previously in the geohistory analysis of Van Hinte (1978) and the "backstripping" approach of Steckler and Watts (1978). Available biostratigraphic and seismic reflection profile data are used to reconstruct the stratigraphy of the margin for different intervals of geologic time. In reconstructing the sedimentary thickness the effects of compaction, sea-level changes and variations in water depths of deposition should be included. Then, the sediment and water loads are "backstripped" for various intervals of geological time using different models for the response of the basement to these loads.

The general equation for "backstripping" can be written

$$y = \phi * \left| S* \frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)} - \Delta_{SL} \frac{\rho_w}{(\rho_m - \rho_w)} \right| + W_d + \Delta_{SL} \qquad (1)$$

where

y = depth of basement without the effects of sediment and water loads ;
$S*$ = sediment thickness corrected for compaction ;
ρ_s = mean density of sediments ;
ρ_m = mean density of mantle ;
ρ_w = mean density of water ;
Δ_{SL} = sea level relative to the present day ;
ϕ = basement response function.

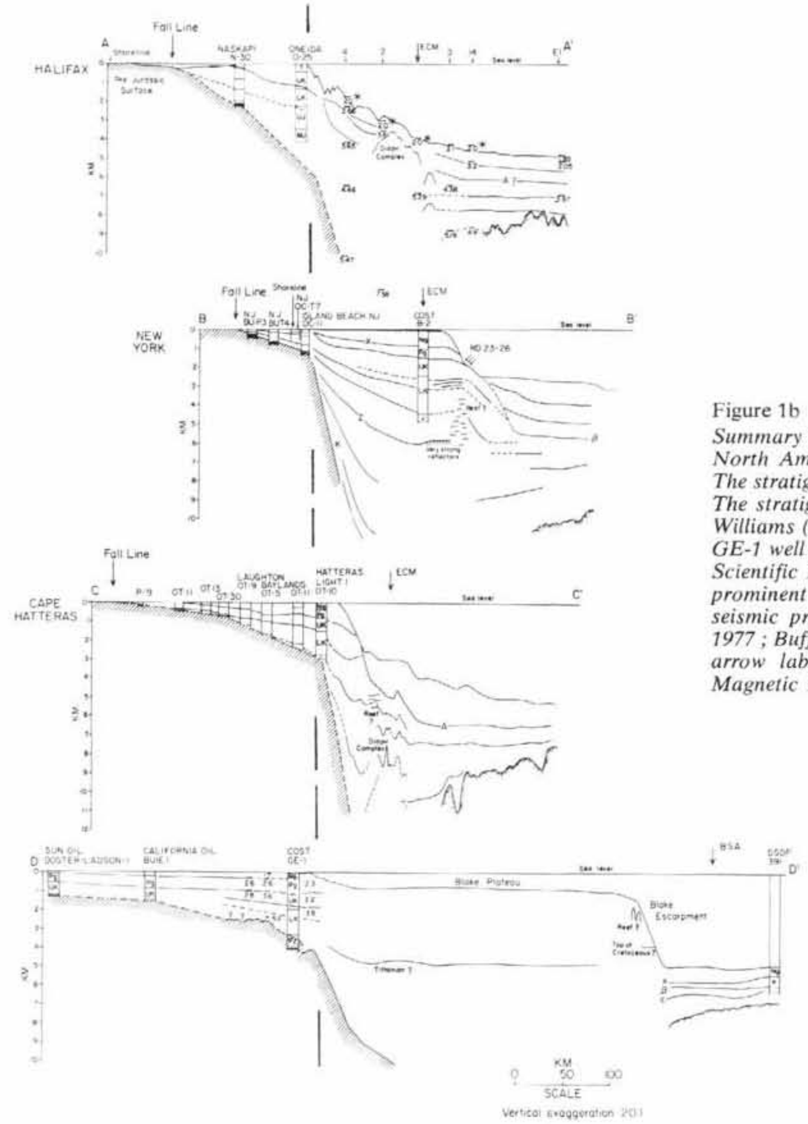

Figure 1b

Summary geologic cross sections along the continental margin off North America aligned along the hinge zone (heavy dashed line). The stratigraphy of the land boreholes are from Brown et al. *(1972). The stratigraphy of the wells offshore from Nova Scotia are from Williams (1975), the COST B-2 well from Scholle (1977), the COST GE-1 well from Amato and Bebout (1978) and DSDP site 391 from Scientific Party (1976). The solid lines in the four sections indicate prominent seismic reflectors indentified on nearby multi-channel seismic profiles (Given, 1977 ; Grow* et al., *1979 ; Grow, Markl, 1977 ; Buffler* et al., *1979 ; Dillon* et al., *1979 ; Sheridan, 1976). The arrow labeled ECM refers to the position of the East Coast Magnetic Anomaly and BSA to the Blake Spur Anomaly.*

If the basement responds to the sediment and water load in a manner similar to a thin elastic plate overlying a weak fluid then

$$\phi = e^{-\lambda x}(\mathrm{Cos}\,\lambda x + \mathrm{Sin}\,\lambda x)\,, \qquad (2)$$

where

$$\lambda = \sqrt[4]{\frac{(\rho_m - \rho_s)\,g}{4\,D}}\,, \quad D = \frac{ET_e^3}{12(1-\sigma^2)}\,, \qquad (3)$$

and

λ = flexural parameter ;
D = flexural rigidity ;
T_e = elastic thickness ;
E = Youngs Modulus ;
σ = Poisson's ratio ;
g = average gravity.

If, however, the basement responds to the load locally then $\lambda = 0$ and $\phi = 1$. Equation 2 shows that relatively wide loads will be locally supported while narrow loads are supported by the strength of the basement. The two terms in brackets in equation (1) therefore represent the effect of the response of the basement to the weight of sedimentary infill and changes in sea-level, while the last term represents the water depth relative to the present day sea level. The last term is therefore that part of the tectonic subsidence which has not been filled in and loaded by sediments.

Compaction correction

As the weight of overburden increases sediments will expel pore fluids and compact. Thus the present day thicknesses of older strata at a margin do not reflect the actual amount of subsidence during those time intervals.

The measure of the amount of compaction in sediments is the porosity, which is defined as the ratio of the pore volume to the total volume. In order to correct for compaction it is therefore necessary to know how the porosity varies with depth during the evolution of the margin. Lithologic logs (sonic, density, porosity) only provide information, however, on the present day variation of porosity with depth.

The actual procedures which should be used to correct for compaction are complicated since the processes by which sediments compact are not fully understood. The compaction of sediments (for example, von Engelhardt, 1977) is not only a mechanical process which depends on the depth of burial of the sediments. Chemical processes of mineral solution and recrystallization play an increasingly larger role with depth. These processes, like the mechanical process, gradually cause a reduction in the pore volume of sediments.

Steckler and Watts (1978) assumed the simple model of a constant porosity versus depth profile through geologic time. In order to compute the thickness of sediments at any time in the past they removed the sediments younger than the age considered and allowed the remaining sediments to decompact by sliding them up the porosity profile. The sediment thickness corrected for compaction is then given by

$$S^* = \Sigma_i \frac{(1 - \phi_i)}{(1 - \phi_i)} T_i , \qquad (4)$$

where

ϕ_i = porosity of the interval prior to sliding up the profile ;
ϕ'_i = porosity after sliding up the profile ;
T_i = interval thickness.

Sclater and Christie (in press) used a similar approach to Steckler and Watts (1978) but allowed each lithology to slide up its own porosity versus depth curve. They fitted exponential curves to the available data for each lithology and used these curves to correct for compaction in the North sea basin. There is much observational evidence, in fact, suggesting there is an exponential decrease in porosity with depth for sandstones and chalks. Plots of porosity versus depth curves for shales, however, show considerable scatter.

We illustrate in Figure 2 the effect of using different porosity versus depth profiles in the calculation of sediment thicknesses for data obtained in the COST B-2 well off New York (Smith *et al.*, 1976). Figure 2 *a* shows the best fitting estimate for porosity versus depth profile (b) along with upper (a) and lower bounds (c) for the profile and

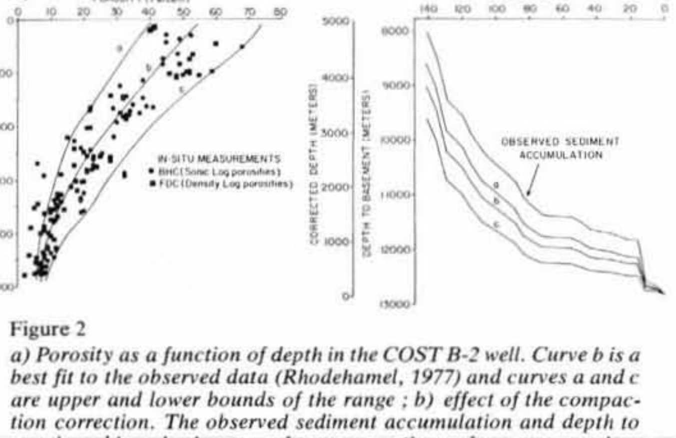

Figure 2

a) Porosity as a function of depth in the COST B-2 well. Curve b is a best fit to the observed data (Rhodehamel, 1977) and curves a and c are upper and lower bounds of the range ; b) effect of the compaction correction. The observed sediment accumulation and depth to stratigraphic horizons after correcting for compaction. Alternatively, curves a, b and c show the calculated depth to basement through time for each porosity profile.

Figure 2 *b* shows the total sediment accumulation in the well along with the sediment thickness corrected for compaction using each of the porosity versus depth profiles. The vertical axis in Figure 2 *b* indicates either the corrected depth to sedimentary horizons in the well or the depth to basement through time. Figure 2 shows that the size of the compaction correction is substantial, reaching nearly 1 km for sediments which formed 140 MY B.P.

The calculations of the compaction corrections in Figure 2 assume, however, the porosity remained constant beneath the base of the well and the porosity versus depth profile is independent of lithology. The lack of knowledge of the porosity between the base of the well and the basement severely hampers the accuracy of the compaction correction. Sediments such as sandstones and shales compact with increasing depth in a similar way but because of a number of factors, including the depositional history and thickness of the overburden, this compaction may vary. We believe, however, that the form of the subsidence curves shown in Figure 2 would not be altered significantly if a different porosity versus depth profile had been used for each lithology in the well.

Paleobathymetry

The tectonic subsidence of a margin provides a depression in which sediments infill. The sediments do not necessarily, however, fill the depression to sea-level. The water depth which remains, relative to present day sea level, is therefore, an important part of the tectonic subsidence of the margin. Unfortunately, estimates of the water depth through time (palaeobathymetry) are difficult to obtain and constitute a major uncertainty in obtaining the tectonic subsidence of a margin.

Although the distribution of some faunal assemblages seem to be related directly to depth (for example, Van Hinte, 1978) the availability of direct water depth indicators is limited. Estimates are made either by direct comparisons to present day occurrences of certain species or assemblages, or by quantitatively determining the relative abundance of, for example, benthonic/planktonic forams and radiolarians or ostracods. In general, estimates of water depths are most accurate in regions where recent faunal assemblages are well known (for example, Gulf Coast, US, Eastern Mediterranean) and in sediments formed in shallow-water environments (neritic and/or shelf facies). Estimates are less precise in older sediments and in sediments formed in deep-water environments.

Figure 3 summarizes estimates of the palaeobathymetry for six wells off eastern North America. The two most landward wells, Mohawk B-93 and Naskapi N-30, are associated with the shallowest estimates of paleobathymetry and the three most seaward wells, Oneida O-25, Sable Island C-67 and COST B-2 wells are associated with the largest estimates. Early in the history of the margin, during the most rapid subsidence, the supply of sediments seems to have been adequate to keep the water depths shallow. Later in the history of the margin, however, hiatuses and large variations in water depths occur. These differences probably arise because, as the subsidence slows, the position of the shoreline becomes sensitive to the rate of change in sea-level which can result in large changes in water depths (Pitman, 1978).

The small scale variations in paleobathymetry in Figure 3 probably are due to local sedimentary processes. The hiatus

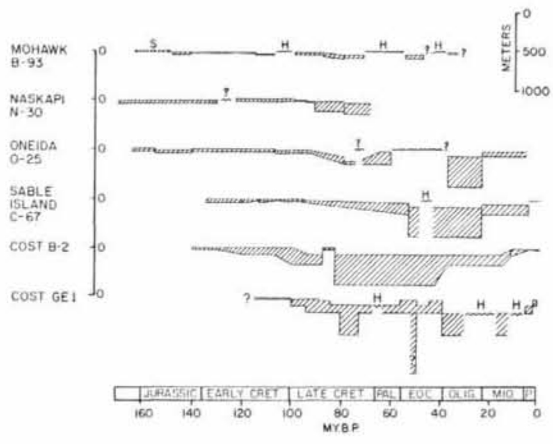

Figure 3

Paleobathymetry estimates for offshore wells off eastern North America (Fig. 1). Data for the Nova Scotian wells are from F. J. Paulas (pers. comm.), the COST B-2 well is from Smith et al. *(1976) and the COST GE-1 well is from Amato and Bebout (1978). H indicates a hiatus in the stratigraphic record and S indicates subaerial deposition.*

in the Sable Island C-67 well during a time of deep water deposition may be due to sediments bypassing the shelf and being deposited on the rise. The shallowing of the water depths in the Miocene at the COST B-2 well is due to a rapid influx of sediments to the shelf while the earlier shallowing in the Eocene/Oligocene may be due, in part, to erosion of the shelf edge by Mid-Tertiary counter-currents (Grow *et al.*, 1977 ; Steckler, Watts, 1978).

Sea level

The variations in sea-level which occur through time (for example, Pitman, 1978) contribute in two main ways to the tectonic subsidence of the margin. The height of sea-level is the reference surface for paleobathymetry and sediment thickness estimates. Thus changes in sea-level with respect to the present day are required in order to provide a reference surface for the tectonic subsidence. In addition, the excess water associated with a highstand in sea-level acts as a load on the basement and depresses it. Steckler and Watts (1978) and Watts and Steckler (1979) modelled this effect by assuming the basement responds to the water load as an Airy-type crust. This is justified in the case of the Late Cretaceous sea-level highstand because of its large areal extent. However, near the shoreline flexural effects are important (Chappell, 1974) and should be taken into account.

The magnitude of sea-level changes through time is a subject of controversy at the present time (Vail *et al.*, 1977 ; Pitman, 1978 ; Watts, Steckler, 1979 ; Bond, 1978). Pitman (1978) estimated that sea-level has fallen by about 350 meters since the Late Cretaceous using changes in the volume of mid-ocean ridge crests through time. This estimate appears to agree with that of Sleep (1971), based on the present elevation of Cretaceous sediments in a tectonically undisturbed region of the continental interior. Watts and Steckler (1979) estimated that sea-level has fallen by less than about 200 m since the Late Cretaceous using well data off eastern North America. Their method yields a minimum estimate, but their values are in better agreement with the magnitudes estimated from continental flooding (Wise, 1974 ; Bond, 1978).

Figure 4 illustrates the effect of using a different sea-level curve in the calculation of the tectonic subsidence for the Oneida O-25 well off eastern North America. This figure shows that changes in sea-level can contribute in a major way to the tectonic subsidence at the well. For example, if the Pitman (1978) curve is used the tectonic subsidence increases linearly with time. However, if the Watts and Steckler (1979) curve is used the margin subsides exponentially with time.

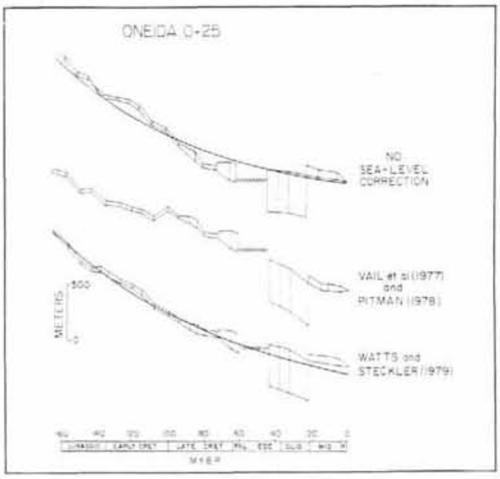

Figure 4

Plot of tectonic subsidence for the Oneida 0-25 well based on different assumed sea-level curves through time. The upper curve is the tectonic subsidence without a sea-level correction. The middle curve assumes the Vail et al. *(1977) first-order sea-level curve and the Pitman (1978) sea-level curve. The lower curve is based on the sea-level curve of Watts and Steckler (1979). The form of the tectonic subsidence depends strongly on the sea-level curve used. If the curve of Pitman (1978) is used the tectonic subsidence is a linear function of time but if the curve of Watts and Steckler (1979) is used the subsidence is an exponential function of time.*

Basement response

The response of the basement to sediment and water loads at a continental margin have been modelled either as Airy-type or as flexure of a thin elastic plate overlying a weak fluid (Gunn, 1944 ; Walcott, 1972 ; Cochran, 1973 ; Chappell, 1974 ; Watts, Ryan, 1976 ; Steckler, Watts, 1978 ; Watts, Steckler, 1979 ; Keen, 1979). Seismic reflection profiling suggests that active faulting accompanies the early stages of rifting (Bœuf, Doust, 1975 ; Ponte, Asmus, 1976 ; Given, 1977 ; de Charpal *et al.*, 1978) and that an Airy type model, in which the sediment and water loads are locally supported by the basement, is most applicable early in the rifting history. The presence of gently dipping postrift sediments and a wide coastal plain, however, suggest that a flexure model is more applicable later in margin evolution. We illustrate in Figure 5 the response of the basement to sediment loading using both an Airy-type and flexure model. Figure 5 *a* is based on the Airy model and 5 *b* is based on the flexure model with $T_e = 15$ km. In both models it is assumed the tectonic subsidence (shown by a dashed line in Fig. 5) increases steeply seaward of the original shore-line and reaches a constant value of 5 km beneath the continental shelf and slope. The shape of the resulting sedimentary basin for the Airy model exactly matches that of the tectonic subsidence. In the flexure model, however, the strength of the basement causes the basin to extend over a broader region, forming a coastal plain sequence even though there is no tectonic subsidence in the region.

In most current models for the evolution of Atlantic-type continental margins (Sleep, 1971 ; Falvey, 1974 ; McKenzie, 1978) the post-rift tectonic subsidence is caused by thermal contraction following heating and thinning of the lithosphere at the time of rifting. Studies of the response of the lithosphere to surface loads, such as seamounts and oceanic islands, show the flexural rigidity of the lithosphere is a strong function of temperature (Watts, 1978). Loads formed on young lithosphere are associated with small values of the flexural rigidity while loads formed on old lithosphere are associated with higher values. Thus we would expect that the flexural rigidity at a margin would increase with time as the basement cools.

A recent compilation of the results of flexure studies in the oceans (Watts *et al.*, 1980) shows that loads which form on or near a ridge crest can be adequately explained by $T_e = 5$ km while loads formed off-ridge can be explained by $T_e = 25$ km. Figure 5 *c* shows a simple model in which the first 2/3 of the sediments at a continental margin loads a 5 km thick elastic plate and the remaining 1/3 loads a 25 km thick plate. The main effect of using a model in which the basement rigidity increases with time is that the younger sediments overstep the older sediments. This is most easily seen (Fig. 5 *c*) at the edges of the basin and should be easily recognized in observed seismic reflection profiles and well data at margins.

The proportions of the sediments used in Figure 5 *c* correspond closely to that of the Jurassic and Cretaceous/Tertiary sections off eastern North America. Thus Figure 5 suggests the Jurassic should pinch out beneath the

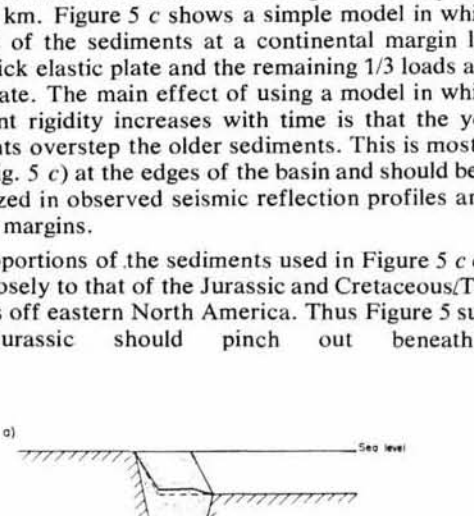

Figure 5

Theoretical cross-sections across a continental margin due to different models of sediment loading. The assumed tectonic subsidence in each model is delineated by the dashed line and reaches a maximum value of 5 km beneath the shelf. a) *Airy model ;* b) *Flexure model with constant elastic thickness ;* c) *Flexure model with varying elastic thickness. The solid lines indicate the positions of a theoretical stratigraphic horizon in which 2/3 of the tectonic subsidence lies below and 1/3 above. This corresponds approximately to the position of the Jurassic/Cretaceous boundary on the east coast of North America. The basement response for the three models is discussed in the text.*

Cretaceous/Tertiary sediments, forming a subcrop at the point X (Fig. 5), near the hinge zone. This subcrop is observed along most of the margin off eastern North America (Fig. 1 *b*). For example, neither the Island Beach well nor the COST GE-1 well (Profiles BB′ and DD′, Fig. 1 *b*) sampled Jurassic and the Hatteras Light-1 only penetrated a small thickness of Upper Jurassic before reaching basement (Profile CC′, Fig. 1 *b* ; Brown *et al.*, 1972 ; Amato, Bebout, 1978).

A further characteristic of flexural models is an outer high in the stratigraphy at the point labeled Y beneath the shelf edge (Fig. 5 *b*, *c*). This high is a consequence of the nature of the flexural response of the basement and the shelf edge and does not necessarily imply the existance of a basement ridge beneath the shelf edge (for example, Burk, 1967). For a particular stratigraphic horizon in the sedimentary section, however, it does suggest relative uplift compared to adjacent regions. The location of the outer high is of stratigraphic interest, particularly since wide shelves (for example off New York and Blake Plateau, Fig. 1 *b*) appear to be associated with buried reef complexes at or near the shelf edge.

These features of the flexure model (Fig. 5 *b*, *c*), are significant and should be taken into account in determining the tectonic subsidence of the margin. Unfortunately, there has been little quantitative modelling of flexure at margins (Walcott, 1972 ; Cochran, 1973) and none of the backstripping studies (Watts, Ryan, 1976 ; Steckler, Watts, 1978), for example, have yet considered the effect of the outer high or a time varying response of the basement on the tectonic subsidence.

Finally, in order to backstrip the sediments it is necessary to estimate the mean density of the sediments, ρ_s, (equation 1) through time. As the sediments compact their average density will increase and hence their loading ability increases. The mean density is therefore most easily obtained from

$$\rho_s = \Sigma_i \frac{(\phi_i \rho_w + (1 - \phi_i) \rho_g) T_i}{S^*}, \tag{5}$$

where ρ_g = sediment grain density and ϕ_i, T_i and S^* have been previously defined.

TECTONIC SUBSIDENCE

A number of authors have discussed the origin of the subsidence of Atlantic-type continental margins. Although most hypotheses appeal to one or more mechanisms, they may be divided into three main groups.

1) Stress based models in which differential loading produces seaward creep of lower crustal rocks and subsidence on the shelf (Bott, 1973).

2) Deep crustal metamorphism in which subsidence is produced by an increase in the density of rocks in the lower crust (Falvey, 1974 ; Spohn, Neugebauer, 1978).

3) Models in which subsidence is the result of thermal contraction after thinning of the crust. This thinning can occur by uplift and erosion (Sleep, 1971 ; Kinsman, 1975 ; Turcotte *et al.*, 1977), extension and necking of the crust (Artemjev, Artyushkov, 1971 ; McKenzie, 1978) or pervasive dike intrusion (Royden *et al.*, 1980).

The advantage of the procedures discussed in this paper (see also Steckler, Watts, 1978 and Van Hinte, 1978) is that the effects of sediment and water loads and surficial geologic processes can be removed, allowing the tectonic subsidence to be isolated. It is the tectonic subsidence that any model for the origin of continental margins must be able to explain.

We have previously shown that backstripping well data from offshore eastern North America reveals a tectonic subsidence curve which exponentially decreases with time. The curve is smaller in amplitude and smoother than the curve of sediment accumulation through time. Watts and Steckler (1979) showed that the shape and amplitude of the tectonic subsidence curve varies across the margin. The Oneida 0—25 well (profile AA', Fig. 1 *b*), which is located landward of the hinge zone, is best fit by an exponential decay with a time constant of 110 m.y. and an amplitude of 2,5 km, while the COST B-2 well (profile BB', Fig. 1 *b*), which is seaward of the hinge zone, is better explained by a smaller time constant of 48 m.y. and a larger amplitude of 5,3 km.

Watts and Steckler (1979) interpreted the tectonic subsidence at the COST B-2 well in terms of a simple thermal model for the cooling lithosphere. They assumed the total thickness of sediments which accumulated after rifting at the well is 12,8 km, based on multi-channel seismic line 2 (Grow *et al.*, 1977), and that the sediments beneath the well were formed at or near sea-level. They used a cooling plate model and estimated the thermal parameters which best explained the tectonic subsidence at the well. The subsidence in their best fitting model, however, was associated with a slope of 512 meters/$MY^{1/2}$ which exceeds that of a normal midocean ridge. One possibility therefore is that not all the tectonic subsidence at the well is of thermal origin. For example, there may have been an initial subsidence at the margin due to heating and thinning at the time of rifting.

McKenzie (1978) has considered a model in which the lithosphere undergoes a passive extension at the time of rifting. This extension causes necking or thinning of the lithosphere, which subsequently cools and subsides with time. Since isostatic compensation is assumed both before and after extension there is an initial subsidence which depends on the initial thickness of the assumed crust.

We show the tectonic subsidence data for the COST B-2 well compared to two different thermal models in Figure 6. The solid line is the cooling plate model of Parsons and Sclater (1977) using the best fitting parameters of Watts and Steckler (1979). The dotted line is the stretching model of McKenzie (1978) using $\beta = 6$. The main difference between the models is in the nature of the initial subsidence. For the cooling plate model it is assumed the crust is initially at or near sea-level but in the stretching model the crust is initially 2,2 km below sea-level. Unfortunately, there is presently too little information on the age and environments of deposition of the earliest sediments off eastern North America to distinguish between these models.

These uncertainties in the nature of the thermal model at the COST B-2 well complicate estimates of the thermal history at the margin. For example, Figure 7 shows the palaeotemperatures computed in the sediments for each thermal model based on a simple model in which

$$T(t, z) = T_{surface} + \int_0^z \frac{Q(t)\,dz}{K_s}, \qquad (6)$$

where

$T(t,z)$ = temperature in the sediments as a function of time and depth ;

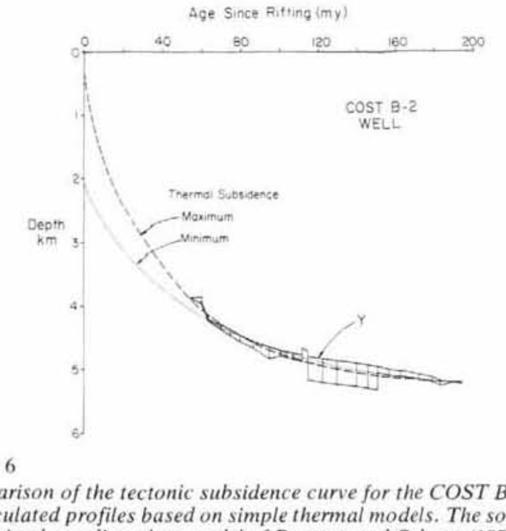

Figure 6

Comparison of the tectonic subsidence curve for the COST B-2 well to calculated profiles based on simple thermal models. The solid line is the simple cooling plate model of Parsons and Sclater (1977) with the parameters given in Watts and Steckler (1979). The dashed line is the stretching model of McKenzie (1978) with $\beta = 6$. In both models the total thickness of sediments at the well has been assumed to be 12,8 km. In the cooling plate model it is assumed the tectonic subsidence is entirely of thermal origin. In the stretching model only part of the tectonic subsidence is of thermal origin. In this model, there is an initial subsidence of 2,2 km at the time of rifting.

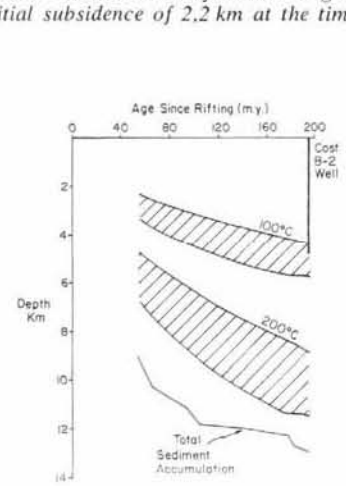

Figure 7

Palaeotemperatures in the sediments which accumulated at the COST B-2 well as a function of time since rifting. The temperatures are based on the cooling plate model (upper curve) and the stretching model (lower curve) shown in Figure 6. $\rho_0 = 3,4$ g/cm^3, $K_s = 5,0 \times 10^{-3}$ cal/°C-cm-sec. and other parameters as in Parsons and Sclater (1977).

K_s = thermal conductivity of the sediments ;

$Q(t)$ = surface heat flux as a function of time in the cooling basement.

The values of $Q(t)$ for the cooling plate model were computed using equation (10) of Parsons and Sclater (1977) and the thermal parameters determined by Watts and Steckler (1979). For the stretching model, equation (7) of McKenzie (1978) was used with $\beta = 6$. Figure 7 shows there are large differences in the depth to the 100 and 200°C isotherms for each model. Unfortunately, surface heat flow measurements and estimates of the geothermal gradients in the well (2,4°C/100 meters, are too uncertain to distinguish between these thermal models.

Available free-air gravity anomaly and geoid data of the margin, however, are consistent with the amplitude of the tectonic subsidence at the COST B-2 well. Simple isostatic considerations indicate that the corresponding amount of crustal thinning that has occurred at the margin is given by

$$T = \frac{D_0(\rho_m - \rho_w)}{(\rho_m - \rho_c)}, \tag{7}$$

where

ρ_c = mean density of the crust ;
D_0 = Œsymptote of the tectonic subsidence.

For either thermal model in Figure 6 $D_0 \simeq 5{,}3$ km so that the amount of crustal thinning is about 21 km. Thus, if the initial crustal thickness off eastern North America is 30 to 35 km these results indicate that a substantial amount of crustal thinning has occurred beneath the well. Figure 8 shows that gravity anomaly data over the margin is consistent with this amount of thinning. In addition, geoid data can be adequately explained by this crustal model and does not require any difference in the thickness of the lithosphere beneath the shelf and the adjacent ocean basin.

The most direct evidence for the nature of the initial subsidence of continental margins, however, has come from studies of biostratigraphic data from well sedimented young margins. Watts and Ryan (1976) showed that the tectonic subsidence of the Gulf of lion and Sardinia margin, which formed by the rotation of Corsica from France about 25 MY B.P., is similar to that of a midocean ridge. They found that the subsidence was most rapid for the most landward well. More recently, Steckler and Watts (in press) have used a model in which both a vertical and horizontal transfer of heat is considered. They used a modified form of McKenzie's (1978) stretching model in which different amounts of stretching occur across the margin. The amounts of stretching required in their models were in the range $\beta = 2{,}6$ to 6, which was larger than expected from present thickness of the pre-rift Oligocene sediments which outcrop in France and underlie the margin. They attributed the large amounts of stretching to either a relatively high geothermal gradient at the time of rifting or to an active heating of the margin.

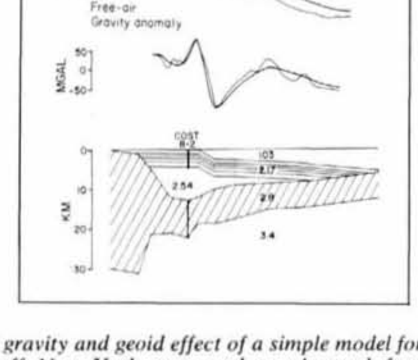

Figure 8

Computed gravity and geoid effect of a simple model for the crustal structure off New York compared to observed free-air gravity anomaly and GEOS-3 altimeter data (from Watts, Steckler, 1979). The dashed vertical line indicates the estimate of the crustal thickness beneath the COST B-2 well based on the thermal models in Figure 6.

SUMMARY

This paper has reviewed some of the factors affecting the subsidence and tectonics of Atlantic-type continental margins. We have shown that sediment and water loading, paleobathymetry and eustatic changes in sea-level and basement response contribute in a major way to the subsidence of these margins. Although these factors do not account for all the observed subsidence they amplify and modify the shape of the tectonic subsidence at a margin. Backstripping techniques are therefore required to quantitatively account for these factors and to isolate the tectonic subsidence.

Future studies in the next decade should emphasize both old margins, with their relatively long record of subsidence, and younger margins, which contain important information on the early history of margin development. Backstripping techniques are required in order to better understand the form of the tectonic subsidence and how this subsidence varies across a margin. However, greater accuracies will be needed in applying these techniques, particularly with regard to paleobathymetry, compaction effects and the effect of a basement response which varies with time. Studies of biostratigraphic from deep wells on the continental shelves, in combination with seismic studies with large aperture arrays, appear to hold the most promise of better understanding the origin of the subsidence of continental margins during the next decade.

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