in: Maurice Ewing Symposium
Series 3
American Geophysical Union
pp. 218-234, 1979

SUBSIDENCE AND EUSTASY AT THE CONTINENTAL MARGIN OF EASTERN NORTH AMERICA

A. B. Watts and M. S. Steckler

Lamont-Doherty Geological Observatory and Department of Geological Sciences
of Columbia University, Palisades, New York 10964

Abstract. Biostratigraphic data from the COST B-2 well off New York and four deep commercial wells off Nova Scotia have been used to remove the effect of sediment loading at the Atlantic-type continental margin off the East Coast, North America. The resulting subsidence contains terms due to both "tectonic" and "eustatic" effects. By assuming the tectonic subsidence is thermal in origin these effects can be separated. The "eustatic" effects have been isolated by least squares fitting an exponential curve to the subsidence data. The resulting sea-level curve shows a maximum rise in sea-level during the Late Cretaceous (about 75-80 m.y.B.P.) which probably does not exceed 150 meters. The tectonic subsidence has been interpreted in terms of a simple thermal model for the cooling lithosphere. Based on this model the thermal thickness of the lithosphere and the total amount of crustal thinning are estimated. These estimates which are consistent with surface ship gravity and GEOS-3 altimeter measurements are used to define the structural elements which control the tectonic evolution of the margin.

Introduction

The Atlantic-type continental margin off the East Coast of North America comprises a substantial thickness ($\gtrsim$ 12 km) of seaward dipping Mesozoic and Tertiary sediments (for example, Sheridan, 1974; Schlee et al., 1976; Jansa and Wade, 1974). Biostratigraphic data from deep commercial boreholes in the outer continental shelf (Scholle, 1977; Gradstein et al., 1975; Jansa and Wade, 1975) show that a large proportion of these sediments were deposited in shallow-water environments. Such large thicknesses of shallow-water sediments cannot be caused by the effects of sediment loading alone and other factors must contribute to the observed subsidence.

A number of authors (Sleep, 1971; Artemjev and Artyushkov, 1971; Bott, 1973; Falvey, 1974; McKenzie, 1978) have discussed the origin of the subsidence of Atlantic-type margins. A useful approach (Sleep, 1971; Watts and Ryan, 1976; Steckler and Watts, 1978) is to account quantitatively for the effect of sediment loading and isolate that part of the subsidence which is not caused by the weight of the sediments. Sleep (1971) corrected for sediment loading using stratigraphic data from more than 35 commercial boreholes in the U.S. coastal plain. He showed that the subsidence not caused by sediment loading was exponential in form and that it was probably thermal in origin. Watts and Ryan (1976) used biostratigraphic data from the Gulf of Lion in the western Mediterranean and showed that the subsidence was similar in form to the empirical ocean ridge curve, and Steckler and Watts (1978) used biostratigraphic data from the COST B-2 well off New York and showed that the subsidence could be interpreted in terms of a simple thermal model for the lithosphere.

In order to quantitatively evaluate and remove the effect of sediment loading two procedures should be followed (Steckler and Watts, 1978). First, use biostratigraphic data to reconstruct the sedimentary section during the development of the margin. Second, "backstrip" the sedimentary section to obtain the subsidence not caused by the weight of sediments.

The resulting subsidence of the margin, however, is affected by world-wide changes in sea-level through time. These changes act as an additional load of water on the basement. Sleep (1971) and Watts and Ryan (1976) did not correct for sea-level changes. Steckler and Watts (1978) corrected for sea-level changes using Pitman's (1978) curve, based on changes in ridge crest volumes and Vail et al s (1977) curve based on seismic stratigraphy. Steckler and Watts (1978) showed that this correction was large and that it significantly altered the shape of the subsidence curve.

There is a difficulty, however, in determining a sea-level curve because "tectonic" and "eustatic" effects cannot be easily separated (for example, Fairbridge, 1961; Hallam, 1963). For example, Pitman (1978) assumed a single, constant hypsometric curve for the continents through time. Differential changes in continental hypsometry, however, may have significantly altered this curve (Bond, 1978).

There is now good evidence that the subsidence of Atlantic-type margins not caused by the effects of sediment loading is thermal in origin (Sleep, 1971; Watts and Ryan, 1976; Steckler and Watts, 1978). Thermal models predict that the subsidence is a function of age and is exponential in form. It should therefore be possible at a margin to separate the thermal or "tectonic" part of the subsidence and isolate that part caused by "eustatic" changes in sea-level.

The purpose of this paper is to use biostratigraphic data from five wells off the East Coast, North America (Fig. 1) to quantitatively understand the origin of the subsidence of Atlantic-type continental margins. By assuming the subsidence is thermal in origin we can separate "tectonic" and "eustatic" effects. The "eustatic" effects are interpreted in terms of a new sea-level curve and the "tectonic" effects in terms of a simple thermal model for the cooling lithosphere. This model is used to place constraints on the evolution of the margin.

Data Reduction

Biostratigraphic data from the five wells off the East Coast, North America (Fig. 1) have been used to remove the effect of sediment loading at the margin. We have used the "backstripping" technique described by Steckler and Watts (1978) assuming the Airy model of isostasy. In this case "backstripping" each stratigraphic horizon can be summarized by the equation

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

where ρ_m is the average mantle density, ρ_w is the average water density, and ρ_s is the average sediment density. Y is the depth to basement without sediment and water loads and represents the subsidence caused by "tectonic" effects. S* is the sediment thickness corrected for compaction, W_d is the water depth at the time of deposition, and Δ_{SL} is the elevation of mean sea-level. The right hand term in equation 1 therefore represents the subsidence caused by "eustatic" effects.

The five wells (Fig. 1) are located in different tectonic settings along the margin (Jansa and Wade, 1975; Scholle, 1977). The Naskapi N-30 (Fig. 2) and Mohawk B-93 wells are located on the slowly subsiding La Have platform, and the Sable Island C-67 and COST B-2 (Fig. 2) wells are located in deep sedimentary troughs. The Oneida O-25 (Fig. 2) well is located near a hinge zone separating these two tectonic environments.

We summarize the stratigraphy of each of the five wells in Figure 3. The wells penetrated sediments which range in age from Middle to Latest Jurassic (Fig. 3; Williams, 1975; Smith et al., 1976). The sediments comprise a sequence of mainly sands and shales with only a small amount of carbonate rocks.

To correct for the effect of compaction (equation 1) we utilized downhole sonic and density logs for each well to determine the variation of porosity with depth (Fig. 4). For the COST B-2 well we used the sonic and density log porosity values tabulated in Rhodehamel (1977), and for wells off Nova Scotia we examined the original Schlumberger sonic and density logs. By assuming that the porosity versus depth curve has remained constant we calculated the thickness and average density of each horizon as it appeared through geologic time. In these calculations we used a uniform grain density of 2.70 g/cm^3 for the wells off Nova Scotia and 2.65 g/cm^3 for the COST B-2 well (Rhodehamel, 1977). We also assumed that the porosity values at the base of these wells could be extrapolated to great depth.

The paleo-water depths required for each well (equation 1) are summarized in Figure 5. Estimates of paleo-water depth for the COST B-2 well are based on dinoflagellates and benthonic foraminifera (Smith et al., 1976) and estimates for the wells off Nova Scotia are based on unpublished multidisciplinary studies (F.J. Paulus, personal communication).

We have not applied a correction for variations in water load due to changes in sea-level through time. Equation 1 can therefore be rewritten

$$Y' = Y + \Delta_{SL} \left[\frac{\rho_m}{(\rho_m - \rho_w)} \right] \qquad (2)$$

where Y' is the depth to basement through time corrected for sediment loads, but still contains terms due to both "tectonic" and "eustatic" effects.

Figure 6 shows the total sediment accumulation and the calculated Y' for each of the wells through time. The dotted region represents that part of the subsidence caused by sediment loading and the hatched region represents that part caused by "tectonic" and "eustatic" effects. The relative smoothness of Y' shows that "backstripping" successfully removes variations in the total sediment accumulation through time that are due to changes in the supply of sediments and local sea-floor sedimentary processes.

Eustasy

In order to examine the evolution of the margin it is necessary to separate the "tectonic" and "eustatic" effects from Y .

A number of studies have now been carried out (Sleep, 1971; Watts and Ryan, 1976; McKenzie, 1978; Steckler and Watts, 1978) that suggest the tectonic subsidence at margins is thermal in origin. Sleep (1971) has suggested the subsidence is caused by cooling of the lithosphere following uplift and sub-aerial erosion whereas McKenzie (1978) has postulated the subsidence is caused by cooling of the lithosphere following "stretching" and thinning at the time of initial rifting. McKenzie (1978) has shown, however, that for all thermal models the subsidence of the

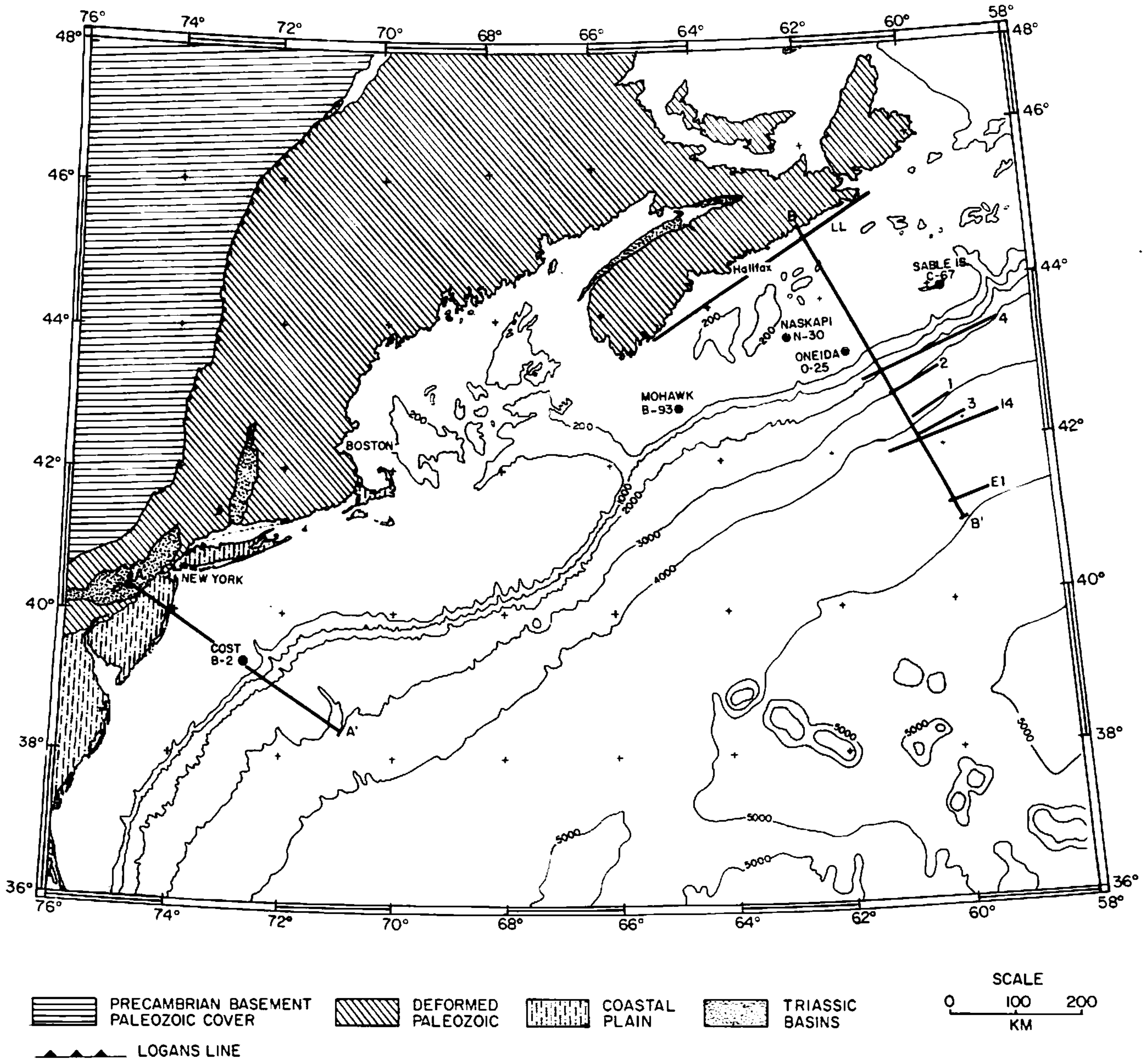

Figure 1. Location map of the COST B-2 well (Scholle, 1977) and the wells offshore Nova Scotia used in this study (Williams, 1975; Jansa and Wade, 1975; A. W. Bally, personal communication). The summary geology and generalized bathymetry are based on the tectonic map of North America (King, 1969). The heavy lines with bars indicate the location of the schematic geological structure sections of the margin shown in Figure 2 and the heavy lines without bars indicate the location of seismic refraction stations 1, 2, 3, 4 of Keen et al. (1975), E1 of Ewing and Ewing (1959), 14 of Keen and Loncarevic (1976) and LL of Barrett et al. (1964).

margin can be characterized by a simple exponential curve for ages greater than 20 to 30 m.y. after initial rifting.

These considerations suggest, therefore, that the "tectonic" and "eustatic" effects can be separated by fitting an exponential curve to Y' We then interpret the exponential component of the subsidence as the "tectonic" effect and the difference between the exponential and Y' as the "eustatic" effect. We fitted an exponential curve to the data within each well by least squares.

In addition, for the three wells which did not reach basement (Oneida 0-25, Sable Island C-67,

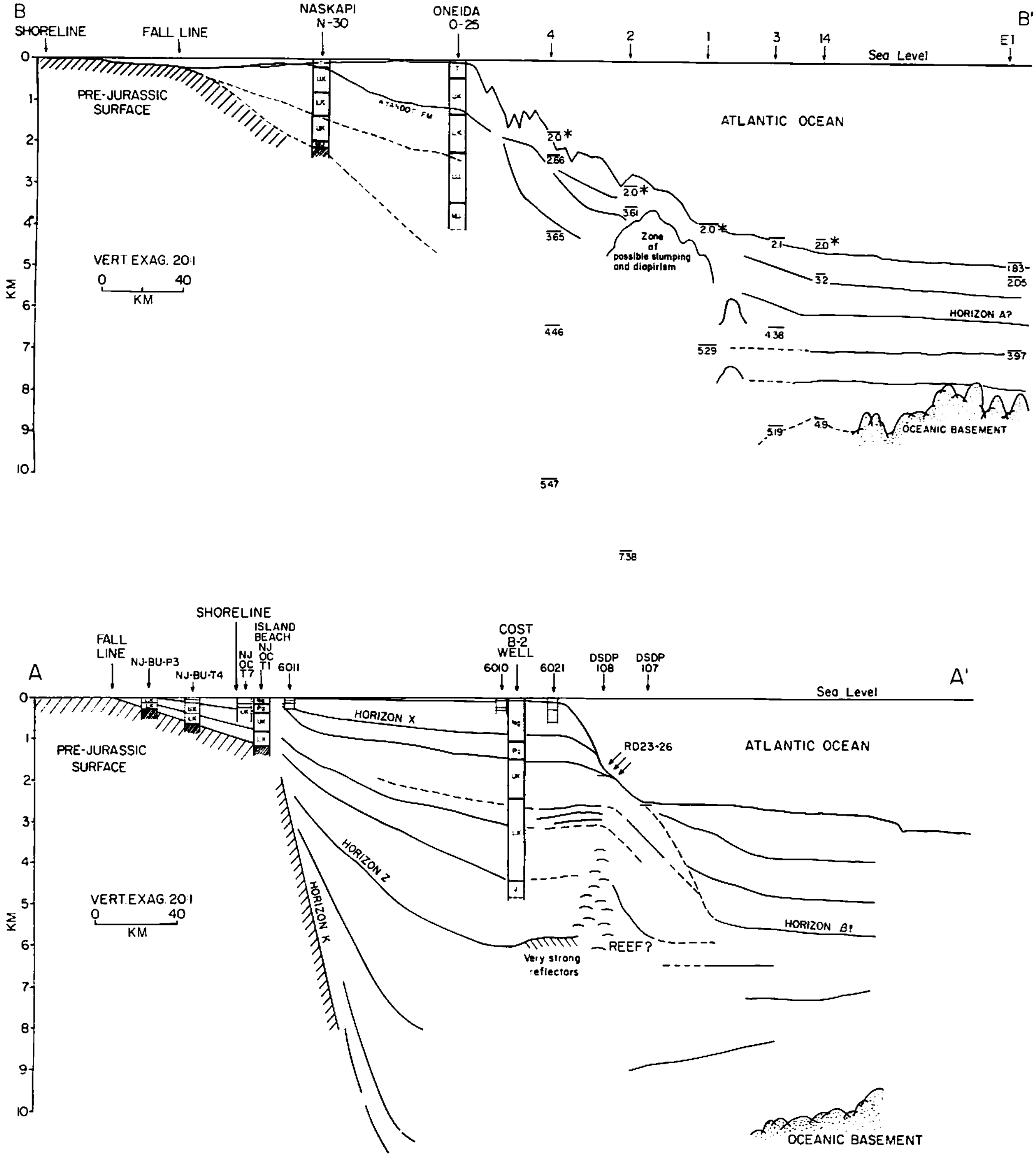

Figure 2. Summary geologic cross-section of the continental margin off New York (profile A-A', Fig. 1) and Nova Scotia (profile B-B', Fig. 1). The summary stratigraphy of COST B-2, Oneida O-25 and Naskapi N-30 wells is based on Scholle (1977) and Williams (1975). In profile A-A' the summary stratigraphy of other wells in the coastal plain is based on Maher and Applin (1971) and in the Outer Continental Shelf is based on Hathaway et al. (1976). Solid lines indicate prominent seismic reflectors identified on adjacent multi-channel seismic profiles based on Given (1977) and King and MacLean (1974) for profile B-B' and Grow et al. (lines 2 and 9; in press) for profile A-A'. The sources and location of seismic refraction profiles is given in Fig. 1.

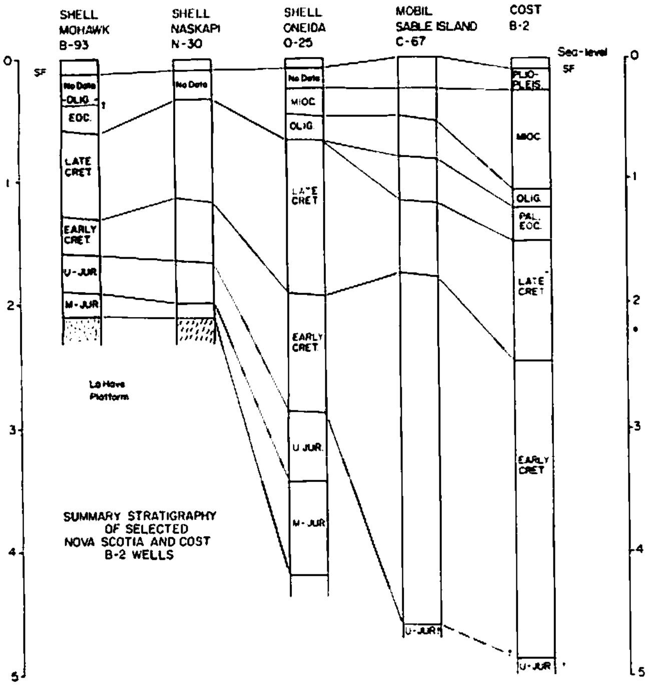

Figure 3. Generalized stratigraphy of the deep commercial wells which are used in this study (Fig. 1). The depth to stratigraphic horizons is based on Scholle (1977) for the COST B-2 well and on Williams (1975) for the wells off Nova Scotia. The slope of the horizons represents the uncertainty in its position.

COST B-2) we considered the effect of the additional data point at the base of the sediments beneath the well. Since this data point is at the time of initial rifting a complete thermal model must be used. For this point we assumed the cooling plate model of Parsons and Sclater (1977). As pointed out by McKenzie (1978) and Steckler and Watts (1978) the ocean ridge model may not strictly apply to a continental margin. At a margin the entire lithosphere has probably not been heated to the solidus temperature. A constraint on this data point is needed, however, in order to determine the long-term trend in sea-level.

In order to apply the model we made the following assumptions: 1) The depth to basement at the Oneida O-25 well is 5.3 km based on a nearby multi-channel seismic profile (Given, 1977), at the Sable Island C-67 well is 8.2 km based on the contours on basement map of the Scotian shelf (Geological Survey of Canada map 1400 A), and at the COST B-2 well is 12.8 km based on nearby multi-channel seismic profile USGS Line 2 (Schlee et al., 1976; Mattick, 1977). 2) The earliest sediments beneath the well were deposited at sea-level. 3) The subsidence of the margin began 195 m.y.B.P., corresponding to the time of extensive basaltic activity in the North Atlantic region at the Triassic-Liassic boundary (Van Houten, 1977). The sea levels, Δ_{SL}, required to explain the differences between these exponentials and Y' (equation 2) are shown in Figure 8. Figure 8a indicates the differences from an exponential curve least squares fitted to the data within the well and Figure 8b indicates the differences in which the additional data at the base of the well has been included. Since Mohawk B-93 and Naskapi N-30 both reached basement, the differences for these wells in Figure 8 are the same. Although there is scatter in the data, a similar pattern of sea-level changes can be distinguished for each well (Fig. 8). With the exception of the Sable Island C-67 well, sea-level appears to rise between the Early and Late Cretaceous and fall between Late Cretaceous and Miocene. A smaller, but rapid, rise in sea-level appears to occur during the Latest Jurassic. The shapes of the two sea-level curves are similar, therefore the sea-level curve is only weakly dependent on the time constant τ and the assumptions of the model.

The well data was then corrected for sea-level using the curve in Figure 8b (equation 2) and the previous procedure was repeated until the differences from the best fitting exponential were satisfactorily reduced. We chose the sea-level curve in Figure 8b because it includes the long-term trend in sea-level. Figure 9 shows the sea-level curve from Figure 8b and the "first estimate" of the sea-level curve derived after iteration. The sea-level curves in this figure have been plotted with present day sea-level as the origin.

The well data which includes the "first estimate" sea-level curve together with the best fitting exponential curves is shown in Figure 10. The inclusion of the sea-level correction has resulted in a noticeably improved fit of an exponential curve to the well data, as can be seen by comparing Figures 7 and 10. There are, however, departures from an exponential curve which cannot be explained by sea-level changes. We attribute these departures to local "tectonic" effects. Prior to the Late Cretaceous the Sable Island C-67 well departs from the curve. This well is located on a possible basement block (North Sable high, Jansa and Wade, 1975) as indicated by dip reversals on nearby seismic reflection profiles. There is also evidence on these profiles of salt diapirism. These factors indicate local tectonic activity which could have disturbed the lower part of the well. The COST B-2 well departs from the curve mainly during the Oligocene when extensive erosion of the shelf may have occurred (Grow et al., in press). The flexural effects of this erosion have been modeled by Steckler and Watts (1978) and can adequately account for the observed departures from the curve.

We compare in Figure 11 the "first estimate" curve to other sea-level curves. The fine solid line is the sea-level change based on the Schuchert-Wise percentage continental flooding estimates and the hypsometric curve for North America given by Wise (1974, fig. 5). The heavy

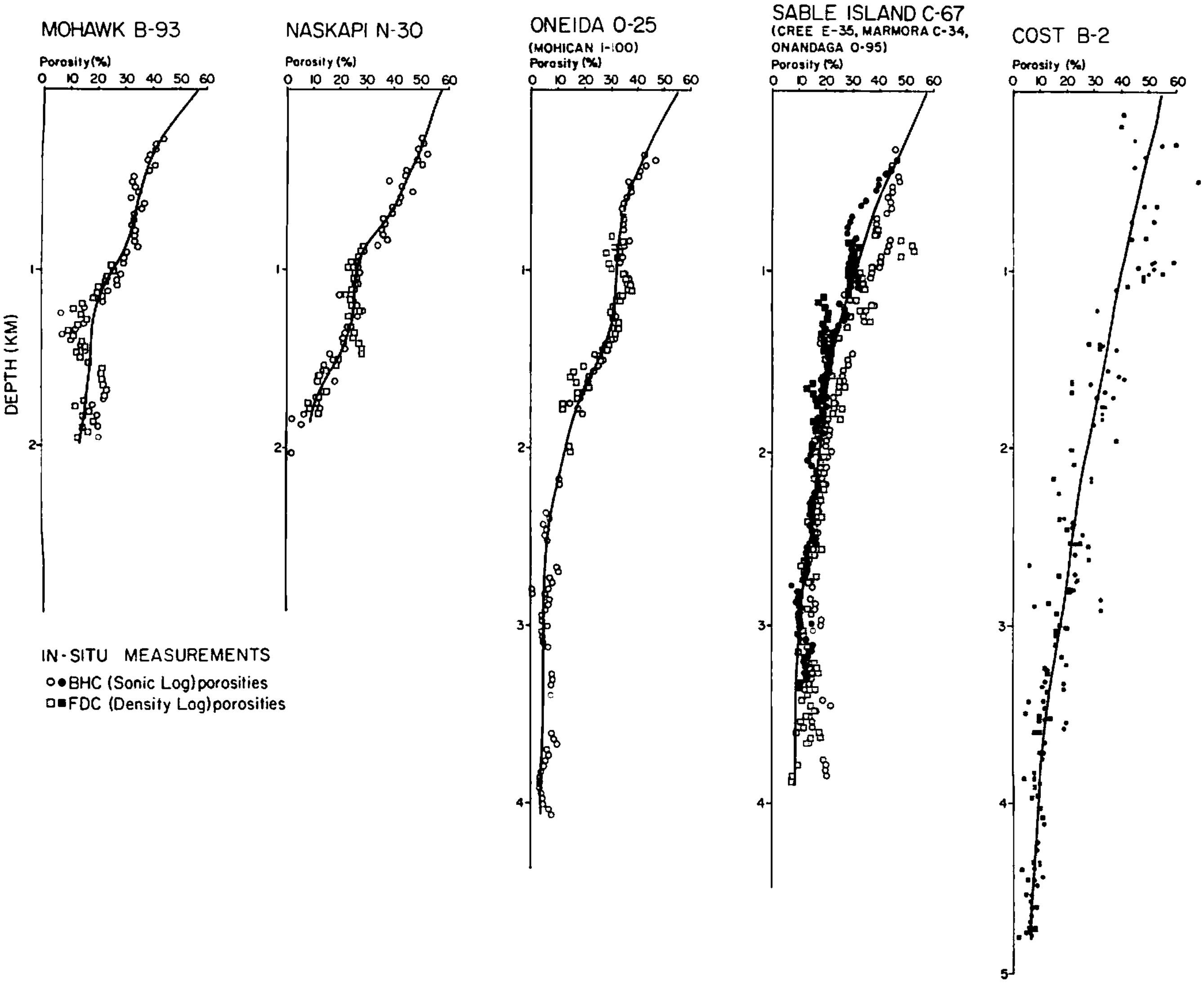

Figure 4. Porosity-depth curves for the five wells used in this study. The sources of the porosity data are discussed in the text. The porosity data for Oneida O-25 and Sable Island C-67 are based on the nearby well logs indicated in the figure. The solid lines indicate the curve used in the computation of the reconstructed sedimentary section.

dashed line is the adjusted sea-level curve of Bond (1978, fig. 4B) based on more recent percentage continental flooding estimates and different hypsometric curves for each continent. The discrepancies between Bond (1978) and Schuchert-Wise are caused by differences in the hypsometric curve assumed through time. There is good agreement that the maximum rise in sea-level in the Late Cretaceous based on estimates of percentage continental flooding and the "first estimate" sea-level curve is 100 to 150 meters. The dashed line with open circles is the sea-level change obtained by Pitman (1978) and Vail et al. (1977). The maximum rise in sea-level based on this curve is greater than 300 meters, which is significantly larger than the "first estimate" sea-level curve.

The differences between using the Pitman (1978) and Vail et al. (1977) sea-level curves and the "first estimate" curve are illustrated in Figure 12 for the Oneida O-25 well. The upper curve in this figure is Y' which includes the "tectonic" subsidence uncorrected for "eustatic" effects. The lower two curves in figure 9 show Y, the "tectonic" subsidence at the margin, calculated using Pitman (1978) and Vail et al. (1977) and the "first estimate" curve. This figure shows the "first estimate" curve successfully improves the fit of the tectonic subsidence to an exponential curve while the Pitman (1978) and Vail et al. (1977) curve implies a nearly linear tectonic subsidence. A

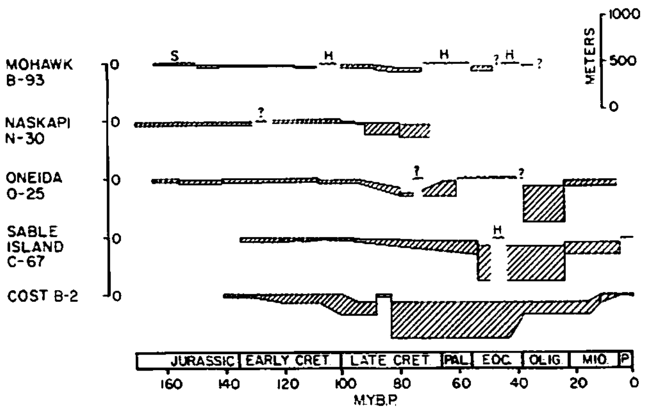

Figure 5. Estimates of the change in water depth of deposition for each of the wells through geological time. The sources of this data are discussed in the text. H indicates a hiatus in the stratigraphic record and S indicates subaerial deposition.

subsidence curve of this form cannot easily be explained by current tectonic models for the evolution of Atlantic-type continental margins.

Thermal Parameters

We have used the "first estimate" sea-level curve to correct Y' and obtain Y, the tectonic subsidence at each well (Fig. 10). By comparing this subsidence to the cooling plate model, which Parsons and Sclater (1977) have shown explains the subsidence of oceanic crust out to ages of at least 160 m.y., the thermal parameters of the lithosphere beneath each well can be estimated. The advantage of this model is that the subsidence follows a relatively simple pattern. At first, the subsidence is proportional to $t^{\frac{1}{2}}$ and then, after a time dependent upon the thickness of the thermal lithosphere, the subsidence is arrested and decays exponentially to a constant value. The parameters that characterize the subsidence curve can be estimated by plotting the tectonic subsidence Y on two simple graphs, Y versus $t^{\frac{1}{2}}$ and log Y versus t. The rate of subsidence on a $t^{\frac{1}{2}}$ plot normalized to the total subsidence (C_1/C_3 of Parsons and Sclater,

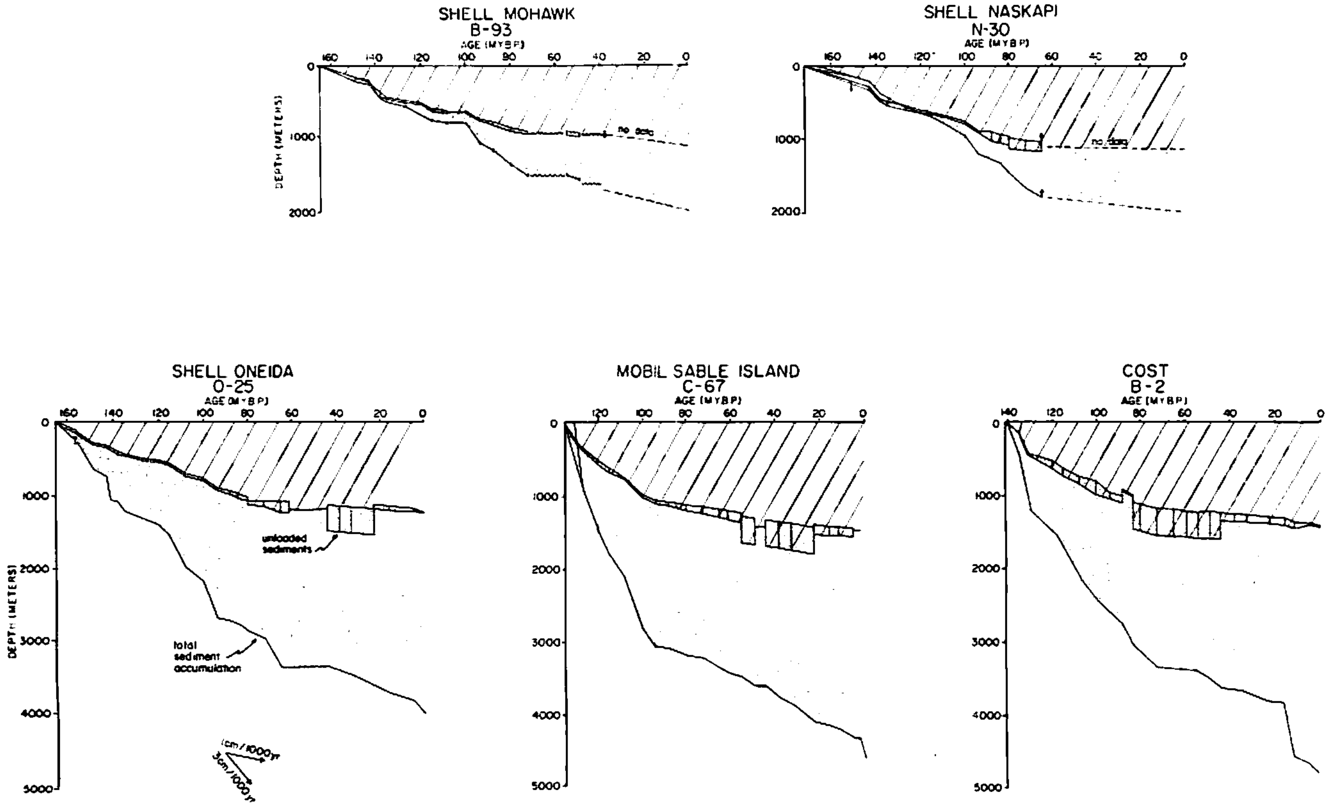

Figure 6. Plot of Y' and the total sediment accumulation through time for each of the wells (Fig. 1). The dotted region indicates that part of the subsidence caused by sediment loading and the hatched region indicates that part of the subsidence caused by other "tectonic" and "eustatic" effects. The range of Y' reflects uncertainties in the water depth (Fig. 5) and the position of individual stratigraphic horizons (Fig. 3).

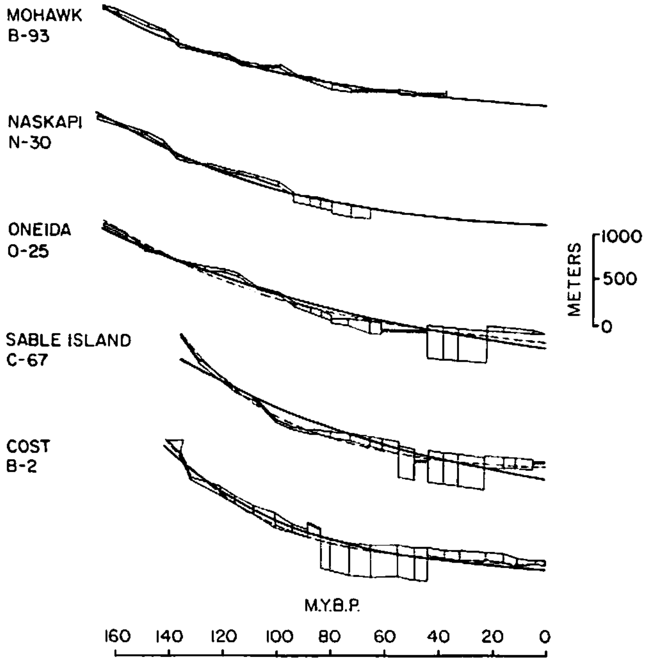

Figure 7. Plot of Y' (Fig. 6) compared to simple exponential curves. The dashed line represents an exponential curve least squares fitted to data within each well. The solid line represents an exponential curve obtained by a least squares fit to the observed data including the data point at the base of the sedimentary section.

1977), the breakdown of the linear part of Y versus $t^{\frac{1}{2}}$ (C_2), and the time constant of the exponential decay (C_2') can be used to obtain independent estimates of the thermal thickness of the lithosphere. In addition, from simple isostatic considerations (Steckler and Watts, 1978), the total tectonic subsidence D_o (C_3) can be used to estimate the total amount of crustal thinning that has occurred at the margin.

We have obtained estimates of the thermal parameters for the COST B-2 and the Oneida 0-25 wells. These wells have been selected because they are considered the most reliable of the five wells studied. Both the Naskapi N-30 and Mohawk B-93 wells are located on relatively thin sedimentary sections (Fig. 3) which penetrated basement. The oldest sediments in these wells are of Middle Jurassic age and are significantly younger than the inferred age of the opening of the Atlantic. The Sable Island C-67 well was not considered because of the poor fit of the tectonic subsidence to an exponential curve (Fig. 10).

Figure 13 shows the plot of the tectonic subsidence Y against $t^{\frac{1}{2}}$ and log Y against t for the COST B-2 and Oneida 0-25 wells. The linear part of the $t^{\frac{1}{2}}$ plot is generally well constrained for the Oneida 0-25 well but poorly constrained for the COST B-2 well. The tectonic subsidence data for the COST B-2 well extends into only a small part of the linear portion of the plot. However, even with this uncertainty the tectonic subsidence for each well appears to deviate from a straight line. Thus estimates of the slope of the straight line (C_1) and the age$^{\frac{1}{2}}$ of the departure from the straight line (C_2) can be determined for each well. The straight line in the plots of log Y against t was determined by a least squares fit to the tectonic subsidence using D_o (C_3) as an independent parameter. The fit to a straight line is generally good and gives an estimate for the time constant of the exponential term (C_2') and D_o (C_3) for each well.

The estimates of the parameters obtained from the plots (Table 1) have been used to compute the thermal thickness of the lithosphere, a, and the amount of crustal thinning that has occurred beneath each well. The COST B-2 well, which is located in a deep sedimentary trough, yields a small estimate for the lithospheric thickness (108 to 130 km), and a large estimate for the amount of crustal thinning. The Oneida 0-25 well, which is located just landward of the "hinge zone" (Fig. 2; Jansa and Wade, 1975) yields a large estimate for the lithospheric thickness (160 to 170 km) and a small estimate for the amount of crustal thinning.

Lithospheric models

We have examined the validity of these estimates of the crustal thinning and thermal lithospheric thickness at the COST B-2 and Oneida 0-25 wells by comparing their computed gravitational effect to observed surface-ship gravity data and GEOS-3 satellite radar altimeter measurements over the margin.

We constructed models for the crustal structure of the margin along profiles A-A' and B-B' (Fig. 1) which include the COST B-2 and Oneida 0-25 wells and compared their gravitational effect to observed surface-ship gravity measurements (Fig. 14). The distribution of the sediments for each profile is based on nearby multi-channel seismic reflection data summarized in Figure 2. The sediment structure thus determined was held constant and only the configuration of the 'Moho' surface was varied between the models. In the upper models (Fig. 14), the 'Moho' is based on an Airy model of compensation for the bathymetry and sediments and an assumed compensation depth of 32 and 30 km for Nova Scotia and New York respectively. The crust and mantle densities assumed are the same as used in the "backstripping" calculations. There is generally a poor agreement between the computed gravity effect based on an Airy model and the observed profiles for each margin. In the lower models (Fig. 14), the 'Moho' surface was varied until a "best fit" was obtained between the computed and observed effects.

The dashed vertical line beneath the Oneida 0-25 and COST B-2 wells in Figure 14 is the

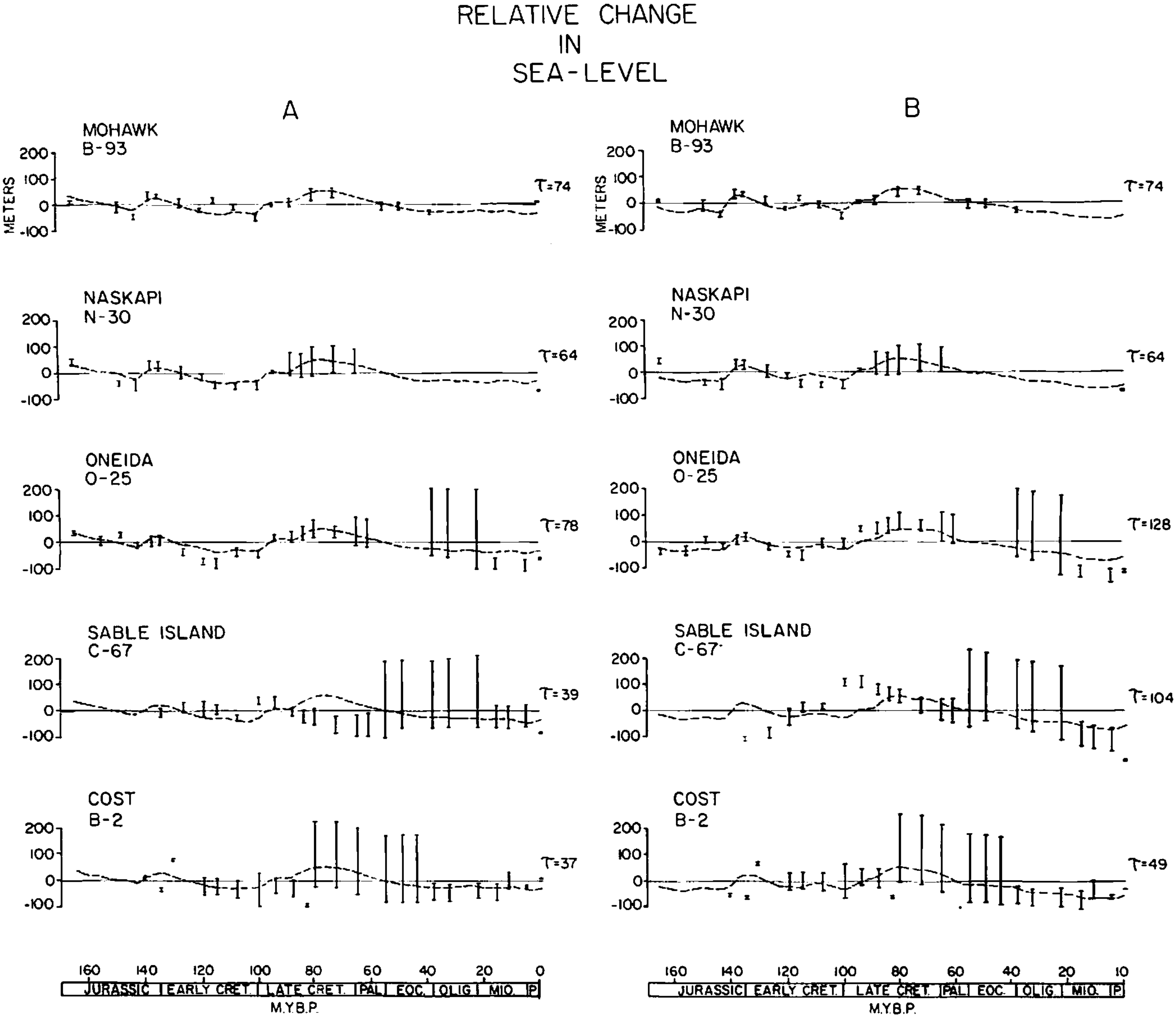

Figure 8. Differences between Y' and the simple exponential curves in Figure 7 expressed as relative changes in sea-level. Column A are the differences from an exponential curve based on data within each well. Column B are the differences from an exponential curve including the data point at the base of the sedimentary section. The dashed line is the average sea-level curve for each column. The time constant τ in m.y. is indicated to the right of each curve.

estimate of the crustal thickness based on thermal calculations (Table 1). This estimate was obtained by subtracting the assumed compensation depth from the amount of crustal thinning, based on D_0. Although gravity data cannot be used to uniquely determine the depth to 'Moho' beneath each well there is good agreement between the crustal thickness based on gravity and thermal modeling (Fig. 14). Thus gravity data are consistent with a small amount of crustal thinning beneath the Oneida O-25 well and a large amount of crustal thinning beneath the COST B-2 well.

Available seismic refraction measurements (Ewing and Ewing, 1959; Barrett et al., 1964; Keen and Loncarevic, 1966; Keen et al., 1975) do not constrain the crustal thickness beneath the Oneida O-25 and COST B-2 wells (Fig. 14). However, they are in agreement with the crustal structure determined in the models for the continental rise off Nova Scotia.

Although gravity data is a useful constraint on the shallow structure of the margin, it provides little information on the deep structure. Recent advances in satellite radar altimetry (Leitao and McGoogan, 1975; Martin and Butler,

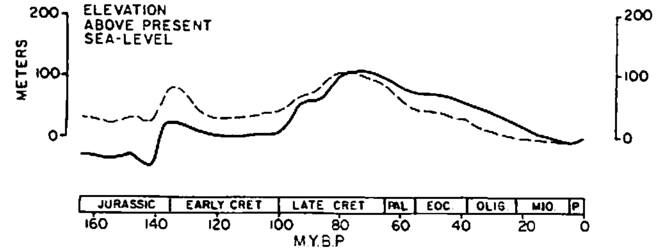

Figure 9. Elevation of sea-level above present day. The dashed line is the average sea-level curve from Figure 8B. The solid line is the "first estimate" sea-level curve determined by iteration. The origin is at present day sea-level.

1977) enable the height of the marine geoid above the reference ellipsoid to be determined to an accuracy of about one meter. The geoid is more sensitive than gravity to long wavelength features in the gravitational field and may therefore provide more information on deep structures. We have used recent determinations of the geoid height derived from GEOS-3 altimeter measurements over the margin off New York and Nova Scotia to better constrain our estimates of the thermal thickness of the lithosphere (Table 1).

The geoid height associated with the "best fit" gravity models in Figure 14 was computed and compared to the observed GEOS-3 geoid height (Fig. 15 a,b). There is an excellent agreement between the observed and computed geoid off New York, but a discrepancy of over 3 meters in the step in the geoid across the margin off Nova Scotia. We have found that reasonable variations in the density of the crust and upper mantle cannot explain this discrepancy, suggesting that its source is located deeper within the lithosphere.

The thermal models predict there may be a significant difference in the lithospheric thickness between the continental shelf (160-170km) and the ocean (128 km, Parsons and Sclater, 1977) off Nova Scotia but little or no difference in the corresponding region off New York (108-130 versus 128 km). To test this possibility we computed the geoid effect of a model which included these differences in the lithosphere for the margin off Nova Scotia. We used the cooling plate model to determine the temperature within the plate and then computed densities using $\rho = \rho_0 (1-\alpha T(Z))$ where α is the coefficient of thermal volume expansion and $T(Z)$ is the temperature as a function of depth. We assumed the same values of α and T_m, the temperature at the base of the plate, as those determined by Parsons and Sclater (1977). The lithospheric model was compensated by adjusting the 'Moho' surface in order to ensure a good fit to the gravity data. This procedure increased the relief of the 'Moho' slightly and increased the depth to the base of the crust from 32 to 33.2 km.

The geoid effect of the lithospheric model is compared to the observed geoid off Nova Scotia in Figure 15c. This figure shows the step in the computed geoid is about 5 to 6 meters, which is in good agreement with the observed geoid. Thus, GEOS-3 altimeter measurements are consistent with similar thermal thicknesses beneath the COST B-2 well and the oceanic lithosphere and with significant differences beneath the Oneida 0-25 well and the oceanic lithosphere.

Discussion

We have determined the tectonic subsidence at five deep wells in the Atlantic-type continental margin off the East Coast, U.S. The tectonic subsidence can be explained by simple models for the cooling of the lithosphere. The total amount of subsidence appears to be determined by the amount of heating and thinning that occurs during the rifting process. Therefore, the structural evolution of the margin is mainly controlled by the amount of initial thinning and the subsequent tectonic subsidence.

The pattern of crustal thinning appears to be similar for both the margin off New York and Nova Scotia (Fig. 14). In a seaward direction across the margin the continental crust gradually thins to a hinge zone. At the hinge zone, the crust thins rapidly to thicknesses similar to that of the oceanic crust. Off New York, the hinge zone corresponds to a major fault identified on multi-channel seismic reflection profile Line 2 (Schlee et al., 1976) as "Horizon K". The hinge zone off Nova Scotia, corresponds to a region of faults and flexures in the basement observed on multi-channel seismic reflection profiles (Jansa and Wade, 1975).

The crustal thinning also appears to be associated with a thinning of the entire lithosphere. Thermal calculations at the Oneida 0-25 well, which is located just landward of the hinge zone, indicate a significantly larger thickness for the lithosphere than at the COST B-2 well, which is located seaward of the hinge zone.

A number of authors have discussed the origin of lithospheric thinning at Atlantic-type margins. Sleep (1971) and Turcotte et al. (1977) have suggested the thinning is caused by uplift and erosion. We have previously shown (Steckler and Watts, 1978) that this mechanism cannot satisfactorily explain the large amount of crustal thinning at the COST B-2 well. McKenzie (1978) has suggested the thinning is caused by "stretching" or "necking" of the lithosphere. There are two observations of basement structure which support the suggestion of regional extension at a margin. First, the presence of normal faulting at the young, starved margin of the Bay of Biscay (De Charpel et al., this volume). Second, the presence of extensive dyke intrusion in the Precambrian basement underlying the former continental shelf of the proto-Atlantic (Rast, in preparation).

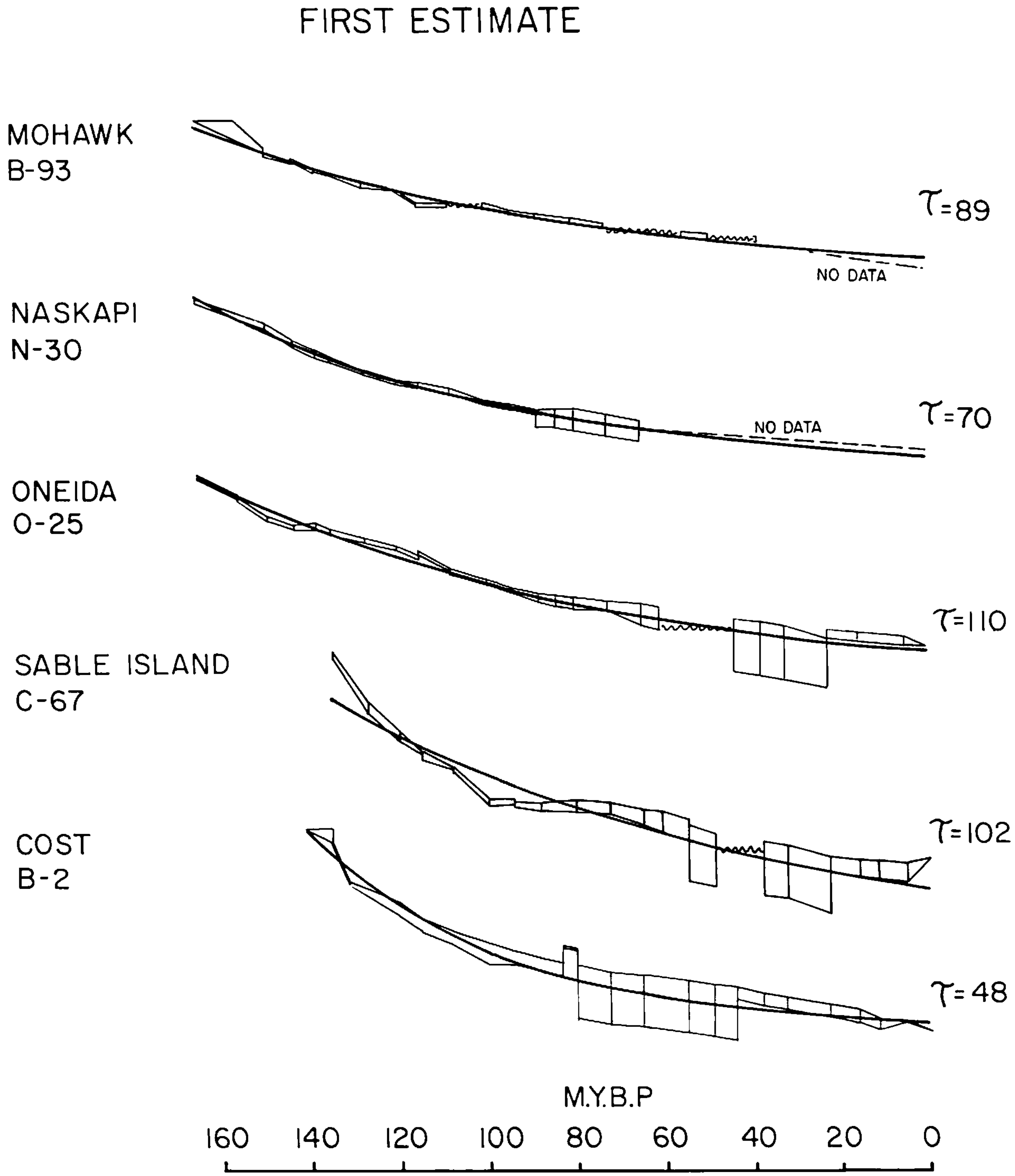

Figure 10. Tectonic subsidence Y of the margin at each well obtained by correcting Y' for "eustatic" effects using the "first estimate" sea-level curve (Fig. 9). The heavy line is the "best fit" exponential which describes the tectonic subsidence for each well.

The tectonic subsidence is a result of the cooling of the thinned lithosphere. As a margin subsides sediments accumulate so that the regions of largest crustal thinning are associated with the largest thickness of sediments. Although we have assumed a simple Airy model of loading in our "backstripping" calculations we would expect that the scheme of loading would change during

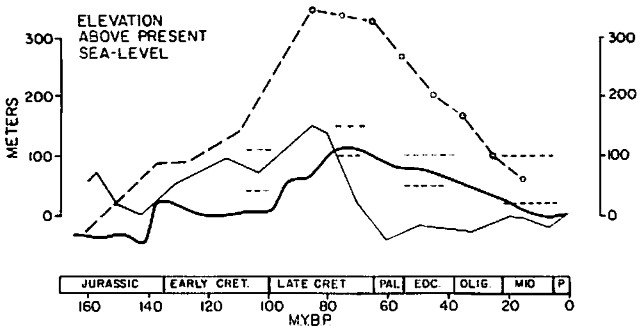

Figure 11. Comparison of recent estimates of sea-level elevations from the present to 165 m.y.B.P. The heavy line is based on the "first estimate" sea-level curve determined in this paper. The circles are from Pitman (1978) and the dashed line is a smoothed curve based on Vail et al. (1977). The fine line is from the Schuchert-Wise estimates of percentage flooding of North America and an hypsometric curve for North America (Wise, 1974). The dashed heavy line is from Bond (1978) and shows the range of sea-level elevations based on percentage continental flooding estimates with different hypsometric curves for each continent.

the evolution of the margin. Initially, sediments load a relatively hot and weak lithosphere. Later in margin evolution the sediments load a relatively cool and rigid lithosphere. Thus the deposition of sediments landward of the hinge zone may be controlled more by lithospheric

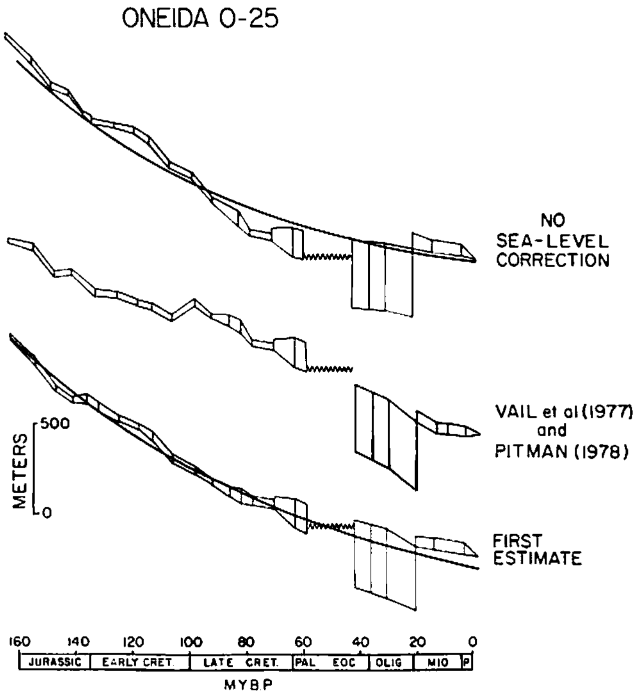

Figure 12. Comparison of the subsidence at the Oneida O-25 well with no sea-level correction curve (upper curve), the Pitman (1978) and Vail et al. (1977) sea-level curve (middle curve) and the "first estimate" sea-level curve (Fig. 9) (lower curve). The heavy lines on the upper and lower curve are the "best fit" exponentials.

TABLE 1. SUMMARY OF THERMAL PARAMETERS DETERMINED FOR THE ONEIDA O-25 AND COST B-2 WELLS

CONSTANT	FORMULA (Parsons & Sclater, 1977)	ONEIDA O-25	COST B-2
Slope Depth vs $(\mathrm{Age})^{\frac{1}{2}}$	$C_1 = \frac{2\rho_o \alpha T_m}{(\rho_o-\rho_w)}\left[\frac{K}{\pi}\right]^{\frac{1}{2}}$	172 $M/(M.Y.)^{\frac{1}{2}}$	512 $M/(M.Y.)^{\frac{1}{2}}$
Asymptote of Depth vs $(\mathrm{Age})^{\frac{1}{2}}$, D_o	$C_3 = \frac{\rho_o a\alpha T_m}{2(\rho_o-\rho_w)}$	2550 M.	5280 M.
Breakdown of linear $(\mathrm{Age})^{\frac{1}{2}}$ relation	$C_2 = a^2/9K$	110-130 M.Y.	70-80 M.Y.
Slope of log (elevation) vs age	$C_2' = \frac{a^2}{\pi^2 K}$	110 M.Y.	48 M.Y.
Thickness of the lithosphere	a	160-170 KM	108-130 KM
Crustal thinning	$T_c = D_o \frac{(\rho_c-\rho_w)}{(\rho_o-\rho_c)}$	10 KM	21 KM

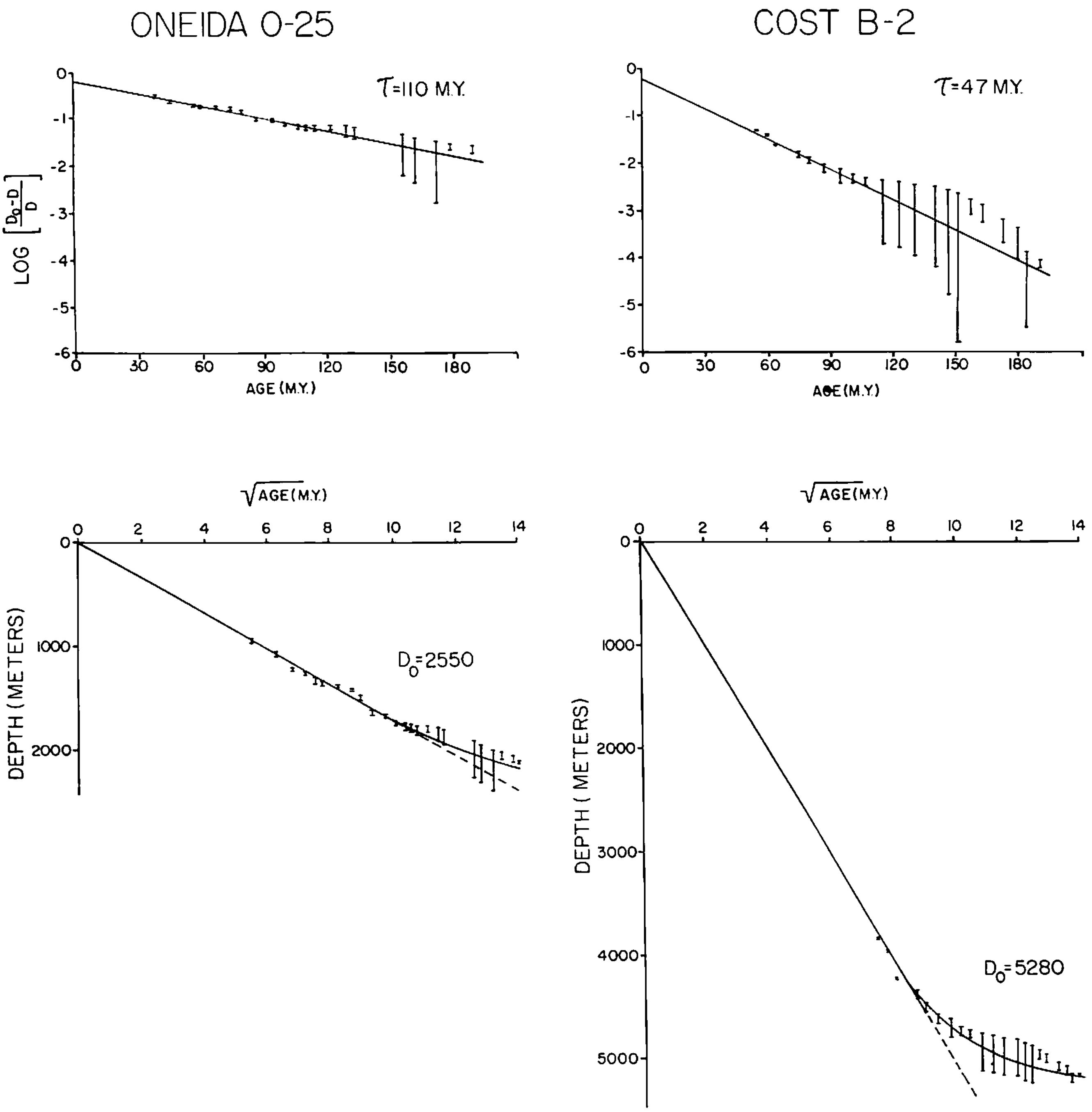

Figure 13. Plot of tectonic subsidence Y against age$^{\frac{1}{2}}$ and log Y against age for the Oneida 0-25 and COST B-2 wells.

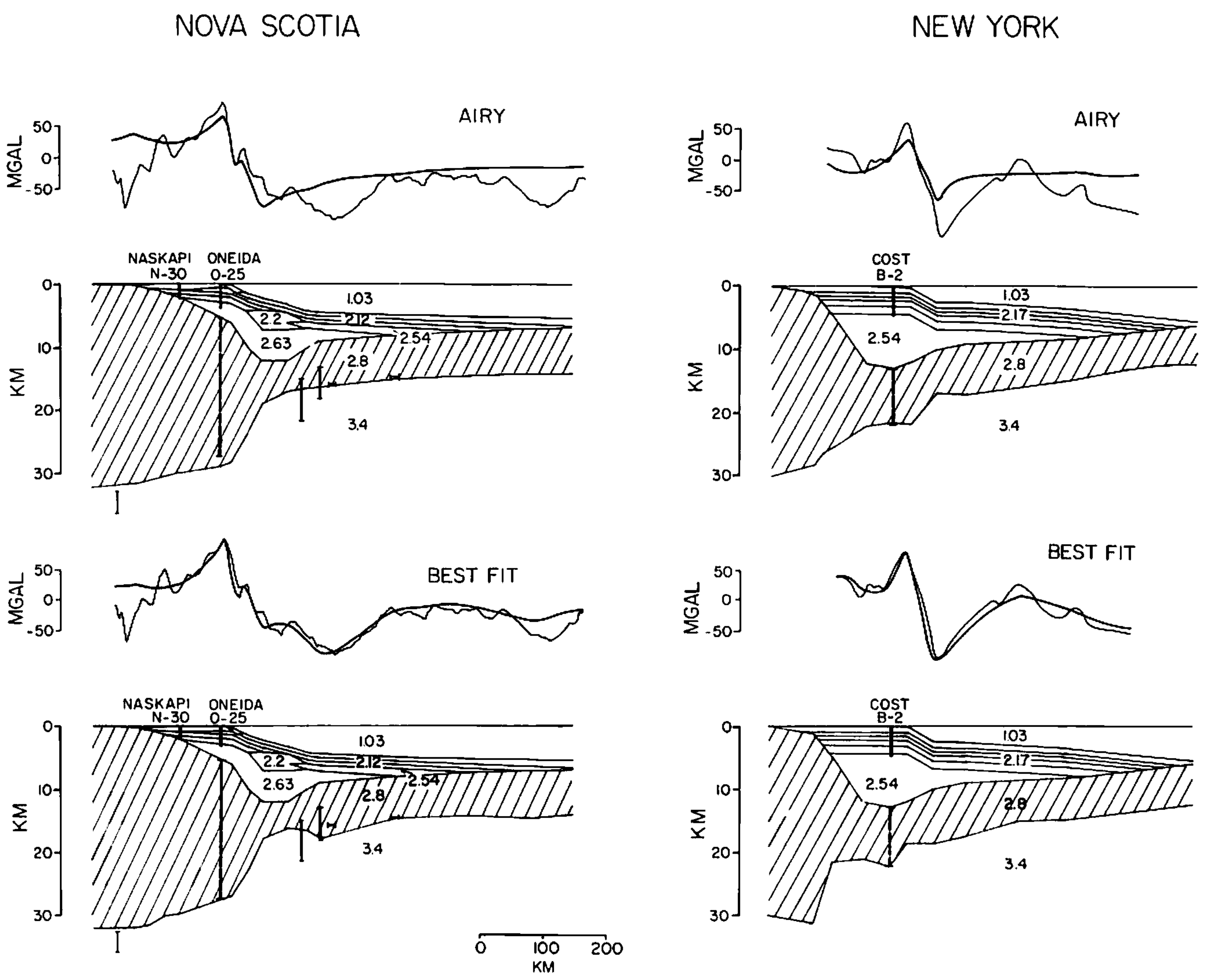

Figure 14. Computed gravity effect of simple models of the margin off New York and Nova Scotia compared to observed free-air gravity anomaly data. The data off Nova Scotia was obtained from R/V Vema cruise 23 and the data off New York was obtained from a map compiled by Ewing (1978). The upper model is based on Airy isostasy and lower model is based on a "best fit" model. The total mass is the same for the computed profiles of each margin. The solid vertical lines indicate the depth to 'Moho' from the seismic refraction lines in Fig. 1. The dashed vertical lines indicate estimates of the crustal thickness based on thermal calculations.

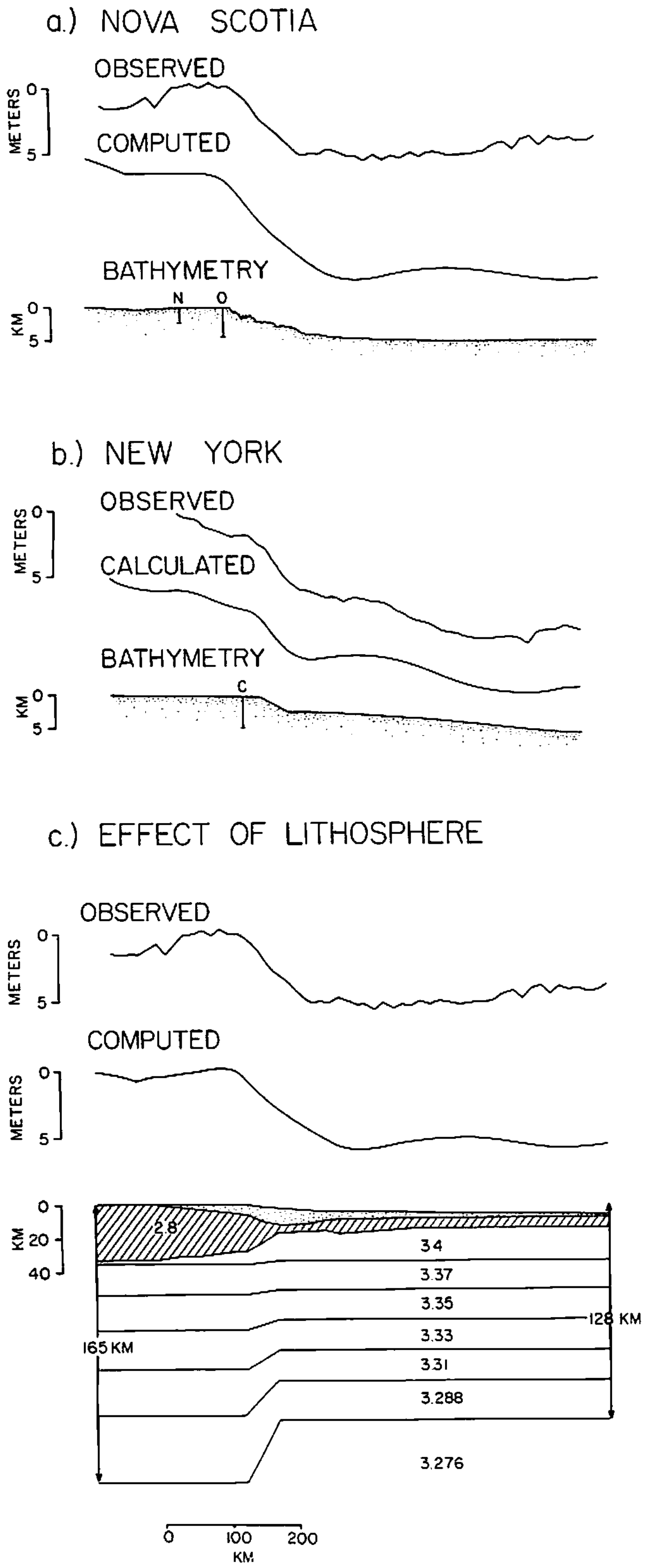

Figure 15. a) and b). Comparison of computed geoid heights associated with the "best fit" models in Figure 13 with observed geoid heights based on the GEOS-3 satellite radar altimeter. The profile off Nova Scotia was obtained on October 30th, 1975 during Orbit #2875 and off New York was obtained on August 9th, 1975 during Orbit #1724. Note that the fit for the New York profile which includes only the crustal structure is generally good but that the fit for the Nova Scotia profile is poor.
c). Comparison of computed geoid heights of a lithospheric model based on thermal calculations to observed geoid heights for the Nova Scotia margin. The fit for this profile, which includes the effect of the lithosphere, is greatly improved.

flexure than by crustal thinning. For example, at Naskapi N-30 and Mohawk B-93 sediments are not deposited until the Middle Jurassic.

The position of the present-day shelf edge does not appear to be controlled by the tectonic subsidence (Fig. 14). The margin off New York, has built out seaward of the maximum sediment thickness. Off Nova Scotia, however, the shelf edge has not built out as far and is located landward of the maximum sediment thickness. Furthermore, off Nova Scotia the trend of the shelf edge and of the locus of the maximum sediment thickness differ. Adjacent to the La Have platform the maximum sediment thickness occurs beneath the continental rise while further to the northeast the maximum sediment thickness occurs beneath the outer continental shelf. These observations suggest the position of the shelf edge is not controlled by the tectonic subsidence; rather, it is probably controlled by the rate of sediment influx and near-bottom sedimentary processes.

In our models a major change in crustal thickness occurred at the hinge zone. It is not known whether the hinge zone represents the boundary between continental and oceanic crust. The crust beneath the COST B-2 well, for example, could be either oceanic, continental, or "transitional" in origin. The hinge zone does, however, represent the location of a major transition in the thermal and mechanical properties of the margin. It is the hinge zone, we believe, that has played the major role in controlling the structural evolution of the margin.

Acknowledgments. We are grateful to A.W. Bally of Shell Oil Company and F.J. Paulus of Superior Oil Company for help in providing down-hole geophysical logs and to W.B.F. Ryan for suggesting we use the well data to examine sea-level. Discussions with G. Karner, D.P. McKenzie and W.C. Pitman have been helpful. G. Bond, V. Ewing, and R. Fairbridge critically read the manuscript and made a number of helpful suggestions. This research was supported by National Science Foundation grant OCE 77-10647 and National Aeronautic and Space Administration grant NAS 6-2519.

Lamont-Doherty Geological Observatory Contribution Number 2802.

References

Artemjev, M.E. and E.V. Artyushkov, Structure and isostasy of the Baikal rift and the mechanism of rifting, J. Geophys. Res., 76, 1197, 1971.

Barrett, D.L., M. Barry, J.E. Blanchard, M.J. Keen and R.E. McAllister, Seismic studies on the eastern seaboard of Canada: The Atlantic Coast of Nova Scotia, Can. J. Earth Sci. 1, 10-12, 1964.

Bond, G., Speculations on real sea-level changes and vertical motions of continents at selected times in the Cretaceous and Tertiary periods, Geology, 6, 247-250, 1978.

Bott, M.H.P., Shelf subsidence in relation to the evolution of young continental margins; in: Implications of Continental Drift to the Earth Sciences, v.2, D.H. Tarling and S.K. Runcorn, (eds.), Academic Press, London, 675-683, 1973.

DeCharpel, O., L. Montadart and D. Roberts, Rifting, subsidence and crustal attenuation in the N.E. Atlantic continental margins, in press, this volume.

Ewing, J. and M. Ewing, Seismic refraction measurements in the Atlantic Ocean Basins, Bull. Geolog. Soc. America, 70, 291-318, 1959.

Ewing, V., Free-air gravity map of the U.S. Atlantic margin, E.O.S. abstract, 59, 378, 1978.

Fairbridge, R.W., Eustatic changes in sea level, in: Physics and Chemistry of the Earth, L.H. Ahrens, F. Press, K. Rankama, S.K. Runcorn, (eds.), Pergamon Press, London, 4, 99-185, 1961.

Falvey, D.A., The development of continental margins in plate tectonic theory, APEA Bull., 58, 95-106, 1974.

Given, M.M., Mesozoic and Early Cenozoic geology, Bull. Can. Petrol. Geol., 25, 63-91, 1977.

Gradstein, F.M., G.L. Williams, W.A.M. Jenkins and P. Ascoli, Mesozoic and Cenozoic stratigraphy of the American continental margin, eastern Canada, in: Canada's Continental Margin and Offshore Petroleum Exploration, C.J. Yorath, E.R. Parker and D.J. Glass, (eds.), Can. Soc. Petrol. Geol. Memoir, 4, 103-130, 1975.

Grow, J.A., R.E. Mattick, J.S. Schlee, Multichannel seismic depth sections and interval velocities over Outer Continental Shelf and Upper Slope between Cape Hatteras and Cape Cod, in: Geological Investigations of Continental Margins, J.S. Watkins, L. Montadert and P.W. Dickerson, (eds.), A.A.P.G. Memoir, in press.

Hallam, A., Major epeieogenic and eustatic changes since the Cretaceous and their possible relationship to crustal structure, Amer. J. Sci., 261, 397-423, 1963.

Hathaway, J.C., J.S. Schlee, C.W. Poag, P.C. Valentine, E.G.A. Weed, M.H. Bothner, F.A. Kohut, F.T. Manheim, R. Schoen, R.E. Miller, and D.M. Schultz, Preliminary summary of the 1976 Atlantic margin coring project of the U.S. Geological Survey, U.S. Dept. Int. Geol. Surv. Open File Report 76-844.

Holmes, A., Principles of Physical Geology, Ronald Press, New York, 1288p, 1965.

Jansa, L.F. and J.A. Wade, Geology of the Continental Margin off Nova Scotia and Newfoundland, in: Offshore Geology of Eastern Canada, Geol. Surv. of Canada, paper 74-30, v.2, 51-105, 1975.

Keen, C.E., M.J. Keen, D.L. Barrett and D.E. Heffler, Some aspects of the ocean-continent transition at the continental margin of eastern North America, in: Offshore Geology of Eastern Canada, Geol. Surv. of Canada, paper 74-30, v.2, 189-197, 1975.

Keen, C.E. and B.D. Loncarevic, Crustal structure on the eastern seaboard of Canada: Studies on the Continental Margin. Can. J. Earth Sci., 3, 65-76, 1966.

King, L.H. and B. MacLean, Geology of the Scotian shelf and adjacent areas, Marine Sci. Paper series No. 7, G.S.C. Paper No. 74-31, 31p., 1976.

King, P.B., Tectonic map of North America, U. S. Geol. Survey, Dept. of Interior, Washington, D.C. #G67154, 1976.

Leitao, C.D. and J.T. McGoogan, Skylab Radar Altimeter: Short wavelength perturbations detected in ocean surface profiles, Science 186, 1028-1029, 1969.

Maher, J.C. and E.R. Applin, Geologic framework and petroleum potential of the Atlantic coastal plain and continental shelf, U.S. Geol. Survey Prof. paper 659, 98p, 1971.

Martin, C.F. and M.L. Butler, Calibration results for the GEOS-3 altimeter, NASA Contractor Rept. CR-141430, 86p., 1977.

Mattick, R.E., Geologic setting, in: Geological Studies on the COST B-2 well, U.S. Mid-Atlantic Outer Continental Shelf Area, P.A. Scholle, (ed.), Geol. Surv. Circ. 750, 4, 1977.

McKenzie, D.P., Some remarks on the development of sedimentary basins, Earth Planet. Sci. Lett., 40, 25-32, 1978.

Parsons, B. and J.G. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res., 82, 803-827, 1977.

Pitman, W.C., III, The relationship between eustasy and stratigraphic sequences of passive margins, Bull. Geol. Soc. America, 89, 1389-1403, 1978.

Rast, N., Precambrian Meta-Diabases of southern New Brunswick, the opening of Iapetus Ocean?, in preparation.

Rhodehamel, E.C., Sandstone porosities, in:

Geological Studies on the Cost B-2 well, U.S. Mid-Atlantic Outer Continental Shelf Area, P.A. Scholle, (ed.), U.S. Geol. Survey Circ. 750, 23, 1977.

Schlee, J., J. C. Behrendt, J.A. Grow, R.M. Robb, R.E. Mattick, P.T. Taylor and B.A. Lawson, Regional geologic framework off northeastern United States, A.A.P.G. Bull. 60, 926-951, 1976.

Scholle, P.A., Geological studies on the COST B-2 well, United States Mid-Atlantic outer continental shelf area, in: Geological Studies of the COST B-2 well U.S. Mid-Atlantic Outer Continental Shelf Area, P.A. Scholle, (ed.), U.S. Geol. Surv. Circ. 750, 1-3, 1977.

Sheridan, R.E., Atlantic continental margin of North America, in: Geology of continental margins, C.A. Burk and C.L. Drake, (eds.), Springer-Verlag, New York, 391-407, 1974.

Sleep, N.H., Thermal effects of the formation of Atlantic continetal margins by continental breakup, Geophys. J.R. astr. Soc., 24, 325-350, 1971.

Smith, M.A., R.V. Amato, M.A. Furbush, D.M. Pert, M.E. Nelson, J.S. Hendrix, L.C. Tamm, G. Wood, Jr., and D.R. Shaw, Geological and operational summary, COST No. B-2 well, Baltimore Canyon trough area, Mid-Atlantic Outer-Continental Shelf (OCS), U.S. Geol. Surv. open file rept. 76-774, 79p. 1976.

Steckler, M.S. and A.B. Watts, Subsidence of the Atlantic-type continental margin off New York, Earth Planet. Sci. Lett. 41, 1-13, 1978.

Turcotte, D.L., J.L. Ahern and J.M. Bird, The state of stress at continental margins, Tectonophysics, 42, 1-28, 1977.

Vail, P.R., R.M. Mitchum Jr., and S. Thompson III, Part Four: Global cycles of relative changes of sea level, in: Seismic stratigraphy - applications to hydrocarbon exploration, A.A.P.G. Memoir 26, 83-98. 1977.

Van Houten, F.B., Triassic-Liassic deposits of Morocco and eastern North America: Comparison, A.A.P.G. Bull. 61, 79-99, 1977.

Watts, A.B. and W.B.F. Ryan, Flexure of the lithosphere and continental margin basins, Tectonophysics, 36, 25-44, 1976.

Williams, G.L., Dinoflagellate and Spore Stratigraphy of the Mesozoic-Cenozoic, Offshore Eastern Canada, in: Offshore Geology of Eastern Canada, Geol. Survey of Canada paper 74-30, v.2, 107-161, 1975.

Wise, D.U., Continental margins, freeboard and the volumes of continents and oceans through time, in: The Geology of Continental Margins, C.A. Burk and C.L. Drake, (eds.), Springer-Verlag, New York, 45-58, 1974.