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PASSIVE MARGIN EVOLUTION, INITIATION OF SUBDUCTION AND THE WILSON CYCLE *

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ABSTRACT

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We have constructed finite element models at various stages of passive margin evolution, in which we have incorporated the system of forces acting on the margin, depth-dependent rheological properties and lateral variations across the margin. We have studied the interrelations between age-dependent forces, geometry and rheology, to decipher their net effect on the state of stress at passive margins. Lithospheric flexure induced by sediment loading dominates the state of stress at passive margins. This study has shown that if after a short evolution of the margin (time span a few tens of million years) subduction has not yet started, continued aging of the passive margin alone does not result in conditions more favourable for transformation into an active margin. Although much geological evidence is available in support of the key role small ocean basins play in orogeny and ophiolite emplacement, evolutionary frameworks of the Wilson cycle usually are cast in terms of opening and closing of wide ocean basins. We propose a more limited role for large oceans in the Wilson cycle concept.

INTRODUCTION

During the last few years much progress has been made in understanding the subsidence mechanisms at passive margins (Steckler and Watts, 1981; Beaumont et al., 1982). These studies have revealed that subsidence due to thermal cooling is strongly amplified by sediment loading at passive margins. The sediment loading capacity is greatest for mature passive margins, where 16 km of sediments can accumulate (Kinsman, 1975). These thicknesses are comparable with sediment thicknesses inferred from reconstructions of sedimentary sequences incorporated in orogenic belts (Stoneley, 1969). This observation together with geological arguments

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put forward by several authors (Dietz, 1963; Dewey, 1969; Cohen, 1982) supports the thesis that the mature stage of passive margin evolution is followed by a transition to an active margin. Passive margins are, therefore, expected to play a key role in the Wilson cycle of ocean opening and closing (Wilson, 1966).

An important problem in determining the possible ways in which the Wilson cycle may operate is the assessment of the size of the ocean that has opened and closed (McWilliams, 1981). As noted by Church and Stevens (1971), much of the geological evidence in collision orogens points to closing of smaller ocean basins rather than large oceans of the scale of the present Atlantic. Nevertheless, evolutionary frameworks of the Wilson cycle are usually cast in terms of opening and closing of wide oceans. Most authors (e.g., Turcotte and Schubert, 1982; Hynes, 1982) advocate a scenario with the development of a major oceanic basin with old and, hence, cold and gravitationally unstable oceanic lithosphere at its continental boundaries, before the basin can be closed.

Basing themselves on the increase with age of the gravitational instability of oceanic lithosphere and the absence of oceanic lithosphere of ages in excess of 200 m.y., these authors argue that eventually the lithosphere would become sufficiently unstable so that it founders and subducts spontaneously. Because there are no obvious examples of new trenches being formed today, the details of this process and the underlying mechanism are poorly understood (Dickinson and Seely, 1979; Flinn, 1982; Turcotte and Schubert, 1982).

A point systematically overlooked in discussions on initiation of subduction in general and in studies of the transformation of passive into active margins in particular, is that the considerable strength of the lithosphere (Kirby, 1980) inhibits its spontaneous foundering. Lithospheric failure is a prerequisite for initiation of subduction as the lithosphere is capable of supporting differential stresses of the order of several kilobars on geological timescales (Watts et al., 1980). It follows that an analysis of the mechanism underlying the passive to active continental margin transition must deal basically with the assessment of possibilities for lithospheric rupture.

The present work, therefore, focuses on the state of stress at passive margins, to investigate whether the stresses generated at passive margins are sufficiently high to induce lithospheric failure and transformation into an active margin. In previous model studies (Cloetingh et al., 1981; 1983) we have shown that flexure induced by sediment loading dominates the state of stress at passive margins. In that work we have demonstrated that, owing to the continuing accumulation of sediments at passive margins, the stress level induced increases with the age of the margin. An important new feature following from more detailed rheological considerations discussed briefly by Cloetingh et al. (1982) and implemented in thermo-mechanical models for the evolution of passive margins is that the strength of the lithosphere increases with age as well.

We found that the aging of passive margins alone does not make them more

susceptible to initiation of subduction. In general, the stresses generated at passive margins are not sufficient to induce lithospheric failure.

The results of our analysis provide insight into the conditions that govern the transition from passive margins into active margins. This applies in particular in the context of opening and closing of small oceanic basins.

MODELS

General features

We have constructed finite element models for passive margins with ages between 5 and 200 m.y. The model features are illustrated in Fig. 1. For all models, we take a half-spreading rate of 1 cm yr^{-1}, characteristic of oceanic lithosphere without attached downgoing slabs (Forsyth and Uyeda, 1975). Age-dependent lithospheric thicknesses are based on Crough's (1975) model for the oceanic lithosphere. A thickness of 150 km, inferred from a number of independent geophysical approaches (e.g., Sclater et al., 1981) is assigned to the continental lithosphere.

Eventual weakness zones associated with the initial break-up phase of passive margin evolution, possibly influence the stress pattern, in particular in the early post-rift phase. Both the extent in depth of such fault zones and the degree of decoupling are unclear. Modelling of subsidence at passive margins (Steckler and Watts, 1981) provides strong evidence that faults associated with the break-up phase are locked very early in the post-rift evolution, and are no longer active. Therefore, we refrain from incorporating mechanical weakness zones and discontinuities (fossil fault zones), and model the transition between oceanic and continental lithosphere as continuous. We adopt a width of 200 km for the transition taking into account the evidence for the presence of rift-stage lithosphere at the margin (Hutchinson et al., 1983).

The magnitudes of the forces associated with the ridge push, and the negative buoyancy of the oceanic lithosphere are calculated on the basis of Oxburgh and

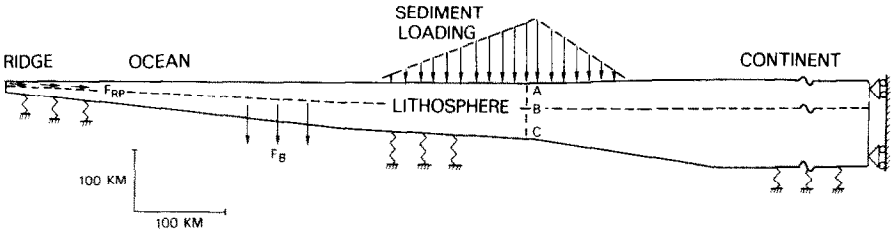


Fig. 1. Model features: geometry, rheology, system of forces and boundary conditions. The bottom of the mechanically strong part of the lithosphere (MSL) is indicated by a broken horizontal line. Young's modulus $E = 7 \times 10^{10}$ N/m^2 and Poisson's ratio $\nu = 0.25$. F_{RP} is the push exerted by the oceanic ridge, F_B is the negative buoyancy associated with the cooling of the oceanic lithosphere when it moves away from the spreading centre. Isostatic forces counteracting the deflection are indicated by springs.

Parmentier's (1977) model for the formation of oceanic crust, and Crough's (1975) model for the thermal evolution of oceanic lithosphere. Following Lister (1975), we model the ridge push not as a line force, but as a pressure gradient, excluding in this way artificial stress concentrations at the ridge. The integrated pressure gradient per unit width along the ridge for oceanic lithosphere with an age of 100 m.y. is 2.3×10^{12} N/m. We ignore drag at the base of the lithosphere. Zero horizontal displacements are prescribed for the right hand boundary of the model to simulate a ridge push transmitted through the continent from an adjacent oceanic plate.

Sediment loading

The sediment loading capacity of oceanic lithosphere increases with age through continued cooling and densification of the lithosphere (which controls the subsidence of the ocean floor (Sclater et al., 1971)). One might, therefore, expect a coupling of the thickness of the sedimentary sequence deposited at a passive margin with the age-dependent subsidence of the underlying lithosphere (Turcotte and Ahern, 1977). As a model for sedimentary loading we adopt two adjacent triangular sediment wedges at the continental shelf and the continental rise. As our references we assume that the maximum thickness of the wedges corresponds with the thickness that can be expected if the sedimentation has been keeping up with the subsidence of a boundary-layer model of the cooling oceanic lithosphere (Turcotte and Ahern, 1977; Wortel, 1980). This implies that the maximum thickness of the sedimentary wedge will gradually increase and reach a maximum of 9.4 km at 200 m.y. (see Fig. 2), following roughly a square-root-of-age relation. The *reference model* of sediment loading constitutes a fair average of the sediment loading histories and resulting thicknesses observed at passive margins (Cloetingh, 1982). Other authors (Southam and Hay, 1981) have reached similar conclusions, stating that "the general principle of sediment accumulation rates decreasing with time after the breaking apart of a passive margin, is confirmed by a large number of geophysical sections". The huge sediment accumulations found at deltas, however, exceed clearly the thicknesses depicted by the reference model. Therefore, a second model of sedimentation, the *full load* model, is adopted, in which the entire loading capacity of oceanic lithosphere is taken up by sediments. In this model, the maximum thickness of the sedimentary wedges reaches 16 km at 200 m.y. (see Fig. 2).

The average shelf width of passive margins is 108 km, while the average slope width is only 11 km. The total width of the sedimentary wedges is on the average 250 km, although for young margins, widths as small as 150 km are found (Southam and Hay, 1981). On the margin, sediments replace water. A reasonable value for the density difference is $\Delta\rho = \rho_{\text{sediments}} - \rho_{\text{water}} = 1.4$ g cm^{-3} (e.g. Kinsman, 1975).

The flexure of the lithosphere under the influence of loading is counteracted by isostasy. Isostatic forces proportional to the deflection due to loading, are simulated with an elastic foundation at the base of the model.

Rheology

We have derived lithospheric strength profiles from Goetze's flow laws for dry olivine (Goetze, 1978; Goetze and Evans, 1979) with an assumed strain-rate of 10^{-18} s^{-1} (characteristic for sedimentary basin development). Temperature profiles are based on Crough's (1975) model for the oceanic lithosphere. Similar to Bodine et al. (1981), we define the depth at which the strength is 500 bar as the lower boundary of the *mechanically strong upper part of the lithosphere* (MSL). Its thickness S_{me} and its maximum strength σ_{YT} are strongly dependent on age. Both increase according to a-square-root-of-age function from a few kilometers, respectively a few kilobars for young lithosphere to values of approximately 50 km and 10 kbar for old oceanic lithosphere (see also Fig. 5). An order of magnitude variation in strain-rate corresponds to only a 1–2 km change in the thickness of MSL. The models presented here deal only with gross features of passive margin evolution and no detailed modelling of a specific margin is intended. As input of (local) detailed stratigraphy is essential before gravity can be used as a successful constraint for flexural modelling, we refrain from incorporating gravity as a constraint. A recent investigation of gravity anomalies at passive margins (Karner and Watts, 1982), however, confirms the validity of our assumption that the mechanical properties of oceanic lithosphere at passive margins are not essentially different from the rheological properties of standard oceanic lithosphere (as inferred from studies of seamount loading). This

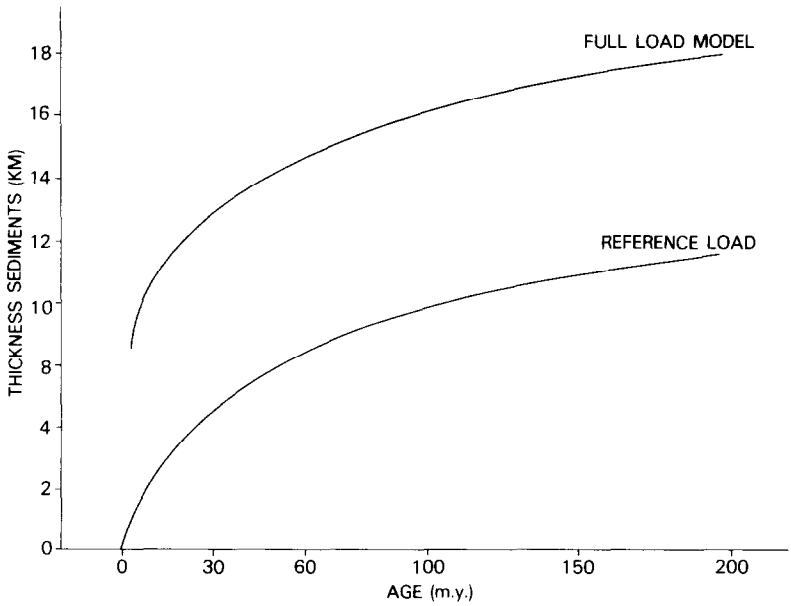


Fig. 2. The increase of sediment thickness versus age adopted for the reference load and full load models.

applies in particular to the increase of the thickness of MSL with the age of the passive margin (Karner and Watts, 1982).

A thickness of 60 km for the MSL of the adjacent continental lithosphere is inferred from studies of the loading of the continental crust (Cochran, 1980). Temperature profiles given by Zielinski (1979) show a gradual linear transition of 100–300 km from the ocean, towards the continent, across the margin. This applies in particular to the isotherms below 600°–700°C, which are most pertinent to the mechanical properties at the margin. Therefore, rheological profiles under the shelf and in the rift-stage lithosphere have been constructed by linear interpolation between the rheologies of the adjacent oceanic and continental lithosphere.

The increase in strength with depth inside the sedimentary section at the top of the lithosphere is usually linear (Bryant et al., 1981). However, even amongst seemingly similar sediment types, a considerable range of shear strengths versus depth has been measured at shallow levels. Further down large uncertainties in strength arise. Therefore, instead of making an uncertain estimate we attribute a zero strength to the sediments.

The hydrostatic pressure extended by the sedimentary load increases the differential strength of the underlying lithosphere. This effect has been incorporated in our rheological models. If we assume a state of thermal equilibrium, the presence of sediments with zero-strength results in a reduction of the thickness of the MSL (see Fig. 5). For reference loading the effect on the temperature distribution is very small at depths corresponding with the transition of the elastic to the ductile regime. For deltas, where rapid sedimentation often takes up the full loading capacity of the underlying lithosphere the situation is different. Initially, sudden deposition of a sedimentary sequence results in a zero-temperature and a non-zero pressure perturbation. With time, thermal equilibrium is restored. We have performed simple thermal calculations, using a finite difference scheme, to estimate this effect. The results indicate that within 10 m.y. thermal equilibrium is restored. Deviations from thermal equilibrium result in an under-estimate by only a few km of the thickness of MSL. We neglect this effect and assume thermal equilibrium throughout our analysis.

Finite element solution procedure

The central part of the finite element mesh employed in our study is shown in Fig. 3. The mesh has been generated by using the GIFTS (Kamel and McCabe, 1979) package as a pre-processor. The mesh has approximately 3000 degrees of freedom. In order to simmulate the increasing thickness of the lithosphere, at each time step a layer of elements must be added at the lower boundary. This is accomplished by generating the complete mesh of 525 8-node elements at the start of the calculations. To elements that at a specific time step are not physically part of the lithosphere, values for Young's modulus reduced by a factor of 100 are assigned.

The elastic foundation is tied to the layer of elements at the lower boundary of MSL. Thus, at each time step the tying is shifted to the base of the newly formed layer of mechanically strong elements. Simultaneously with the generation of new strength profiles, the number of mechanically weak elements is reduced and a load increment is added. The selected time steps which vary from 2.5 m.y. to 20 m.y. are equivalent to increments in the thickness of the lithosphere of the order of 5 km.

The finite element calculations were carried out with a modified version of the MARC (Marc, 1980) program. A secant stiffness procedure was used to solve the plane strain equations for plasticity. The depth-dependent rheology has been implemented via the integration points of the elements. This approach allows an accurate strength representation, without strength discontinuities between adjacent elements. For flexural analysis, constant strain elements are inadequate and an element is required with linearly varying strain over its surface. For this reason, the displacement field has to be at least quadratic and, hence, an eight-node quadrilateral element was chosen. The results of the analysis were checked and confirmed by convergence tests and an analysis of the internal reaction forces of the model. It should be noted that the finite element approach followed here is not hampered by the limitations and the assumptions of the classical thin plate theory, encountered in other methods of flexural analysis (Bodine et al., 1981). In particular, the assump-

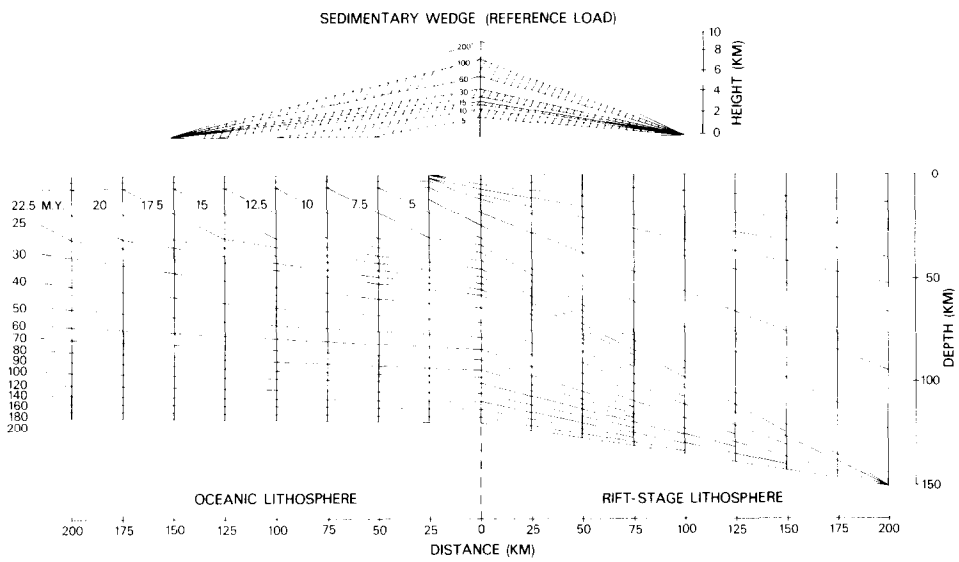


Fig. 3. Central part of the finite element mesh employed in this study. Layers of elements are incrementally added to the lithosphere, whose lower boundary is indicated at various time steps (specified in m.y.). A comparison of the lower boundaries at 12.5 m.y. and 20 m.y. (given by dashed lines) illustrates the growth of the lithosphere in horizontal and vertical directions. The height and the width of the sedimentary wedge are given in kilometers at some selected time steps (specified in m.y.).

tions of a zero shear stress neutral surface and of zero shear forces in the vertical plane are avoided.

For computational reasons, the emphasis of our analysis is on a set of model calculations, in which the loads are placed instantaneously on the lithosphere. Most of the earth's deltas have, measured on geological time-scales, been deposited instantaneously (within a few hundred thousand years) in particular during Pleistocene times (Nelson et al., 1970). For deltas the approximation of an instantaneously applied load is anticipated to be a very good one. We infer, however, that the stress maxima calculated at the bottom of the MSL tend to be overestimated for the reference model of sediment loading. To evaluate this effect more quantitatively, we have developed an additional class of models for a limited set of lithospheric ages, in which incremental loads are placed on the lithosphere.

RESULTS

The stress calculations are made for a passive margin in different stages of evolution in which the two sediment loading models and the depth-dependent rheologies are incorporated.

Reference load model

Figure 4 shows the results for the case of a reference load on a 100 m.y. old passive margin. The deformation of the lithosphere and the resulting stress field, with an order of magnitude of some kilobars, is dominated by the sediment loading: the contribution of the plate tectonic forces to the stress field is an order of magnitude smaller. Considering these orders of magnitude, it is justified to neglect here relatively minor effects of stresses due to crustal thickness inhomogeneities—less than 100 bar in oceanic lithosphere (Bott and Dean, 1972)—and stresses induced by (hypothetical) phase changes in the lithosphere under the influence of sediment loading (Neugebauer and Spohn, 1978). Differential stresses $\sigma_H - \sigma_V$ are largest at the points of maximum flexure. The largest stress maximum is located under the rise in oceanic lithosphere close to the transition to rift-stage lithosphere. The extreme lowermost and uppermost regions of the MSL fail. This is due to the high stresses developed at the top and at the base. The main part of the MSL remains in the elastic state. In Figs. 5a and 5b, the stress maxima for 100 m.y. and 30 m.y. are displayed for the reference load as a function of depth, down to the base of the MSL. Hatched areas indicate failure by brittle fracture in the uppermost part and by ductile flow in the lowermost region of the MSL. From a comparison of these figures, the dependence on age of the state of stress at passive margins is evident.

In order to summarize this dependence, we have plotted in Fig. 6 the maximum differential (tensional) stresses as a function of age for the four cases (30, 60, 100 and 200 m.y.) considered in the reference load models. The age-dependence is

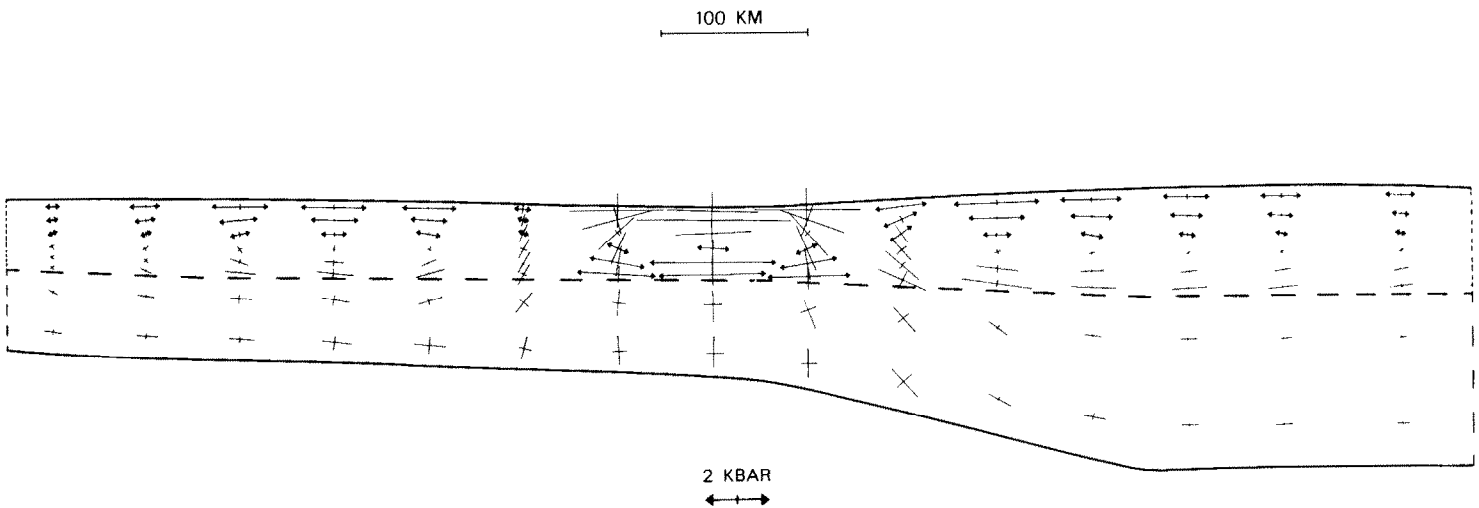


Fig. 4. Stresses calculated for a passive margin with an age of 100 m.y., based on the reference model of sediment loading, given in Fig. 2. Flexure caused by sediment loading forms the dominant deformation mode at the margin. Principal stresses denoted by arrows are plotted in the undeformed configuration. Stresses are plotted only for the parts of the lithosphere where the deformation is significant. Principal stresses are given in kilobars. Symbols (↔) and (+) denote tension and compression, respectively.

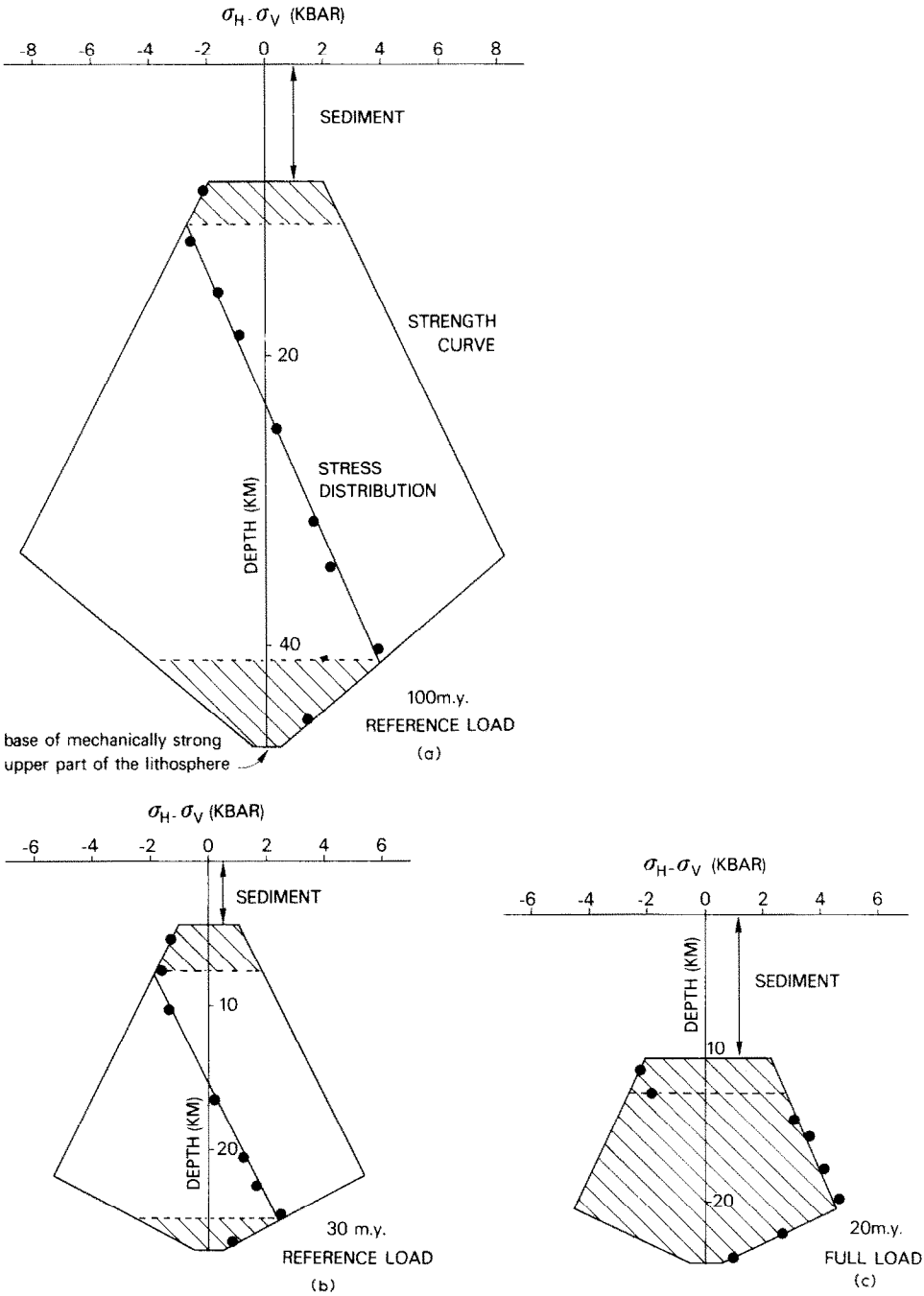


Fig. 5. Comparison of stresses generated at the margin with lithospheric strength. Strength envelope and results of stress calculations (solid dots) as a function of depth down to the base of MSL at the point of maximum flexure (*AB* of cross section *ABC* in Fig. 1; see also Fig. 4). The line inside the strength envelope connecting the solid dots is the stress distribution. Differential stresses ($\sigma_H - \sigma_V$) are plotted

strongest for ages below 100 m.y. From 30 to 100 m.y., an interval in which the sedimentary loading and the thickness of MSL and its strength increase, the differential stress maxima are also seen to increase with age. From 100 to 200 m.y., the strength and thickness of MSL show only a gradual increase. The increase in sediment load, according to our reference model, results only in a minor increase of the stresses. Interesting quantities are: the ratio of the maximum stress generated and the maximum strength in tension (σ_{YT}), the ratio of the thickness of the two parts of MSL which are in failure (S_y) and the thickness of MSL (S_{me}); and the ratio of A_y, corresponding with the hatched area inside the strength envelope (Fig. 5) and the total area A of the envelope. For the reference model of sediment loading, these quantities prove to be essentially independent on age (Fig. 6a).

Stresses are on a level too low to result in rupture of the lithosphere, no matter whether the margin is in a youthful or mature stage.

Full load model

The situation is drastically different when the full loading capacity is taken up by the sediments. For the full load analysis, we constructed seven models for ages between 15 and 200 m.y. The surplus load of sediments added to the reference load will be most effective in creating high stresses when deposited on a young (weak) margin.

This is demonstrated in Fig. 5c, which shows the stresses at the point of maximum flexure generated by full loading on a 20 m.y. old passive margin. Complete failure of the lithosphere is induced. Of the three ratios $(\sigma_H - \sigma_V)/\sigma_{YT}$, S_y/S_{me} and A_y/A, we select the latter as the most meaningful quantity to illustrate the evolution of the stress pattern under full loading conditions. Quantity A (or A_y) is a combined measure of the thickness and the strength of the mechanically strong part of the lithosphere. The ratio A_y/A is plotted in Fig. 6b as a function of the age of the margin. Figure 6b and results of numerical calculations made for ages below 20 m.y. (not shown here) show that full loading on passive margins with ages below 20 m.y. leads to complete failure of the lithosphere. For ages in excess of 20 m.y., the relative amount of failure strongly decreases with age.

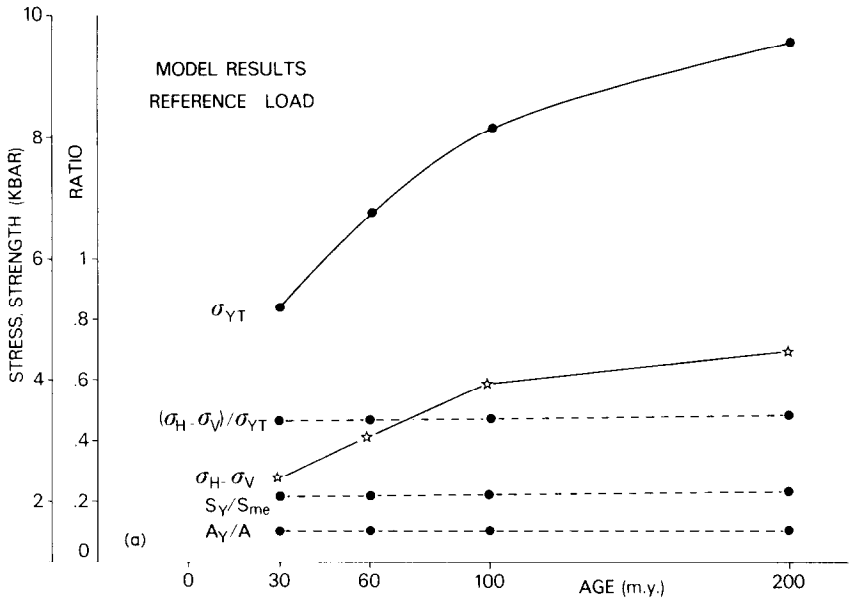
Investigation of the assumption of instantaneously applied loads

We have investigated the influence on the results presented in the previous section of the assumption of instantaneously applied loads. To this end, we have constructed

versus depth. Sign convention for the stresses: tension positive, compression negative. Zero-strength has been assumed for the sediments. Hatched areas in the upper and lower part of the MSL denote failure by brittle fracture and ductile flow respectively.

a (top), b (lower left). Results for the reference model of sediment loading for ages of 100 and 30 m.y.

c (lower right). Results for the full load model for 20 m.y. The horizontal dashed line indicates the neutral surface just before complete failure of 20 m.y. old lithosphere takes place.



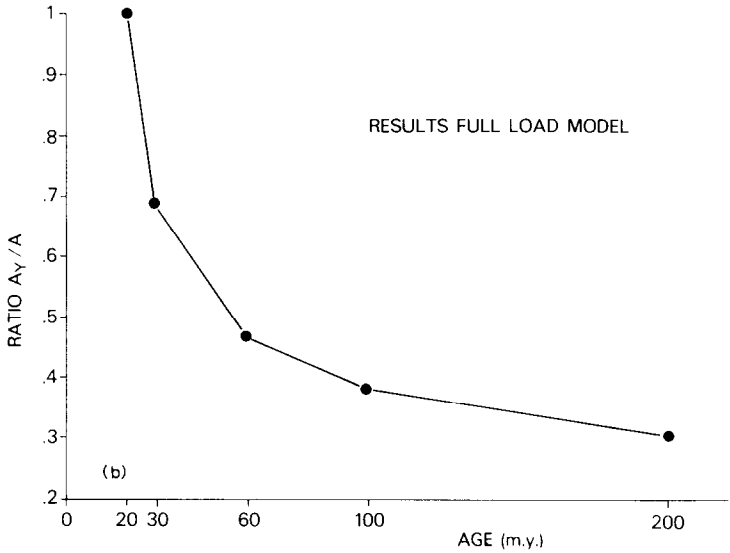


Fig. 6a (top). Model results for the reference load: maximum differential stresses ($\sigma_H - \sigma_V$), tensile strength σ_{YT}, their ratio, the relative thickness of MSL in failure (S_y/S_{me}), and the ratio of A_y (the hatched area inside the strength envelope; see Fig. 5) and A (total area of the envelope) plotted as a function of lithospheric age. Stresses remain on a level too low to result in rupture of the lithosphere, no matter whether the margin is in a youthful or mature stage.

b (bottom). Results for the full load model: ratio of A_y and A as a function of lithospheric age. For ages below 20 m.y. lithospheric failure and consequently initiation of subduction is induced.

models for a limited set of lithospheric ages, in which incremental loads are placed on the lithosphere. Considering the results obtained with instantaneously applied loads, we restrict ourselves to the youthful stages of margin evolution (ages up to 20 m.y.). Compared with the results of the calculations in which we applied an instantaneous load, a relatively small part of the lithosphere is in failure. Flexural stresses are more evenly distributed over the thickness of MSL, since the position of the neutral plane moves downward with each new loading step. With time, several stress maxima are developing at different levels within MSL compared with the case of instantaneously applied loads (Fig. 6a), the differential stress maxima are lower. The differential stress maxima show a relatively small increase with age, which results for ages in excess of 5 m.y. in values for the ratio $(\sigma_H - \sigma_V)/\sigma_{YT}$ of approximately 0.35. Although the value of 0.35 for $(\sigma_H - \sigma_V)/\sigma_{YT}$ is slightly lower than the value of 0.45 found with the models described in the previous section, the essence of the conclusions reached with models assuming instantaneously applied loads is not altered.

For the full load model, the deviations from the results obtained with loading in small steps from the results given in the previous section are even smaller. Loading with small time steps does not change the age (20 m.y.) at which complete lithospheric failure is induced.

IMPLICATIONS FOR THE WILSON CYCLE

It has been shown here that stresses generated at passive margins of large ocean basins are generally insufficient to induce lithospheric failure and transformation into active margins. This offers an explanation for the enigma that gravitationally unstable oceanic lithosphere at the margins of the Atlantic is not subject to subduction. Furthermore, this finding might provide an explanation for the observation that in the Pacific frequently renewal of previously existing subduction zones occurs (Karig, 1982) instead of the formation of a new subduction zone. It is widely held that the present Pacific coastline of Northern and Southern America has been the site of semi-continuous eastward subduction since the late Proterozoic (Windley, 1977). At that time different plate configurations and a different thermal regime might have offered conditions more suitable for initiation of subduction. Due to steeper temperature gradients in the lithosphere (McKenzie and Weiss, 1980), the oceanic plates must have been considerable weaker and thinner, implying a considerable lower stress level required for plate rupture.

From our results, we conclude, however, that extensive sediment loading (following the full load model) might provide an effective mechanism for closure of young, and presumably small, ocean basins. As mentioned before, closing of small oceanic basins plays an important role in the process of mountain building. Usually, precise data on the timing of the closing events are absent, which inhibits a quantitative comparison with our model results. Nevertheless, much of the available geological

evidence is clearly contradictory to activation of passive margins in a late stage of their evolution. As observed by McWilliams (1981) not all (Proterozoic) Pan-African and older mobile belts mark the sites of major ocean closure, but rather formed without the destruction of vast amounts of oceanic lithosphere. On the base of an extensive survey of the tectonic framework of Central and Western Europe, Zwart and Dornsiepen (1978), point out that the Variscan orogens are due to collision and closing of short-lived oceans of minor size. Furthermore, glaucophane schists and paired metamorphic belts, which phenomena are usually associated with subduction of old and, hence, cold oceanic lithosphere, are rare in the Atlantic region (Miyashiro, 1980). Various investigations of Alpine orogeny (e.g., Trümpy, 1981; Zwart and Dornsiepen, 1978) have provided strong evidence for closure of small oceanic basins in an early stage after opening. For the Central Alps, closure of oceanic basins *within* the first 100 m.y. after their opening has been demonstrated by Frisch (1979). Winterer and Bosellini (1981) have documented the evolution of a thickly sedimented Jurassic passive margin in the Southern Alps. These authors argue for closing of the oceanic basin *within* 50 m.y. after opening.

Frequently, however, only indirect and debatable arguments are available for the estimation of the size of the proposed oceans. This applies in particular to the "type locality" of the Wilson cycle, the Iapetus ocean, which is of unknown width (Windley, 1977). In fact, the best evidence for the Iapetus ocean is the occurrence of ophiolite belts (Zwart and Dornsiepen, 1978). Several authors (e.g. Dewey, 1976; Nicolas and Le Pichon, 1980) favour obduction of thermally young immature oceanic lithosphere to account for the short time gap documented between ophiolite formation and ophiolite emplacement. Boudier and Michard (1981) found that the Semail ophiolite of Oman was emplaced on the adjacent passive margin 20 m.y. after creation at the spreading ridge. A small ocean basin rather than a large Atlantic type setting has been proposed for other Tethyan ophiolites and for Pan-African ophiolites (Le Blanc, 1981).

Emplacement of young, gravitationally stable (Vlaar and Wortel, 1976) oceanic lithosphere on the adjacent continent during the transformation of a passive margin of a small ocean basin, into an active margin offers an explanation for these observations.

To cast at forehand, evolutionary frameworks of the Wilson cycle in terms of opening and closing of wide ocean basins seems sometimes to be more inspired by an actualistic comparison with the present size of the Atlantic Ocean, than by a consideration of more pertinent geological observations. During its evolution, the Atlantic Ocean passed through a transition from a narrow to a wide ocean basin, without the formation of a system of subduction zones at its margins. Apparently optimal loading conditions for transformation of passive into active margins were not fulfilled at this stage, while further aging has not made the passive margins more susceptible to initiation of subduction. To rely too heavily on such an actualistic analogon might, therefore, be very misleading. Moreover, estimated widths of oceans

inferred from palinspastic restorations are frequently automatically increased with an additional 1000 km inferred from the lengths of slabs consumed in modern circum-Pacific subduction zones, associated with subduction of old oceanic lithosphere (e.g. Williams, 1980). Such reconstructions, exclude a priori the possibility of the closing of a young oceanic basin and might even lead to an overlooking of eventual interesting consequences of the subduction of young lithosphere (England and Wortel, 1980). This study has shown that if after a short evolution of a passive margin, subduction has not yet started, continued aging of the passive margin alone, does not result in conditions more favourable for transformation into an active margin. Therefore, we propose a more critical appraisal of the role large oceans play in the Wilson cycle concept.

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