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# State of Stress at Passive Margins and Initiation of Subduction Zones

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**We present the results of a finite element analysis of the state of stress at passive continental margins, which we performed to investigate whether these margins are potential sites for initiation of subduction. The state of stress is determined by the effect of sediment loading at the rise, and to a lesser extent by plate tectonic forces. For a model of sedimentation at the margin, coupled to the subsidence of the cooling oceanic lithosphere, stresses up to 3 kbar can be generated. Stresses of this order may cause failure of the lithosphere and initiation of subduction. In general, however, an additional cause (e.g. plate interaction and geometrical focusing effect) not included in the modelling is required to start the subduction process. If, after 100 million years of evolution, subduction has not started, continued aging of the passive margin alone does not result in conditions significantly more favorable for initiation of subduction.**

Initiation of oceanic lithosphere subduction is one of the major issues in geodynamics. However, as stated by Dickinson and Seely (1979) "the mechanisms that initiate plate consumption at new subduction zones and thus generate arc-trench systems are incompletely known." These authors summarize the possible mechanisms for initiation of oceanic lithosphere subduction, and distinguish the following two classes: (1) Plate rupture, either within an oceanic plate, or at a passive margin, and (2) Reversal of the polarity of an existing subduction zone, eventually after a collision of an island arc with a passive margin (see also Speed and Sleep, 1982; Speed, this volume). Another class, although not mentioned by these authors, is initiation of subduction by inversion of transform faults into trenches (see below). We concentrate here on the mechanisms for the formation of a new plate boundary and refrain from dealing with the mechanisms of polarity reversals.

A key factor generally determining the possibility of subduction of oceanic lithosphere is its gravitational stability. The total stability of the oceanic lithosphere is the sum of the positive buoyancy of its stable petrological stratification and the negative buoyancy resulting from thermal contraction upon cooling (Oxburgh and Parmentier, 1977). Wortel (1980) shows that for a model of oceanic lithosphere consistent with observations of heat flow and topography the system is stable for ages less than 30 million years, due to the stabilizing effect of the density changes accompanying the formation of oceanic crust (according to Oxburgh and Parmentier's model). As a result of its further cooling and thermal contraction, the lithosphere becomes unstable for ages above 30 million years (Oxburgh and Parmentier, 1977). Considering the age-dependence of the lithospheric instability, it is reasonable to expect that the chances for initiation of oceanic lithosphere subduction increase with age (Vlaar and Wortel, 1976). However, to allow initiation of subduction of gravitationally unstable lithosphere, stress conditions must be favorable for the creation of a failure zone within the lithosphere. Failure of oceanic lithosphere and initiation of subduction might preferentially take place at existing weakness zones. As such, transform faults have been advocated by several authors (Uyeda and Ben-Avraham, 1972; Uyeda and Miyashiro, 1974; Dewey, 1975). Turcotte, Haxby, and Ockendon (1977) argue favorably for initiation of subduction by mantle material penetrating the lithosphere at oceanic spreading centers. Although at ridge-transform intersections high stresses might be generated (Fujita and Sleep, 1978), this young lithosphere is still gravitationally stable. Initiation of subduction of stable oceanic lithosphere is hampered by resistive forces active in trench formation (McKenzie, 1977). Apparently, spreading centers are not the most appropriate sites for initiation of subduction.

Geological evidence and speculation for margins with widely different ages, ranging from early Proterozoic to recent (Dewey, 1969; Hoffman, 1980; Williams, 1979; Karig et al, 1980), support the thesis that particular passive margins might be potential sites for the formation of plate boundaries. There are several reasons why such a process might preferentially take place at a passive margin. Important factors are the contrast in mechanical properties across the margin (in general much greater than the contrast between oceanic lithosphere adjacent to a transform fault) and pre-stressing the lithosphere by sediment loading at the margin. The above consideration led us to focus here on the possibilities for initiation of oceanic lithosphere subduction at a passive margin.

The state of stress at a passive margin is determined by its local geometric and rheologic lithospheric properties, and by the system of forces acting on the lithosphere. Of these features the thickness of the oceanic lithosphere (Parsons and McKenzie, 1978), its rheological stratification (Caldwell and Turcotte, 1979; Bodine, Steckler, and Watts, 1981), the push exerted by the elevation of the oceanic ridge (Richter and McKenzie, 1978; England and Wortel, 1980), and the forces associated with the (negative) buoyancy of the lithosphere (Vlaar and Wortel, 1976) are a function of the oceanic lithosphere's age. The sediment loading capacity of oceanic lithosphere increases with age, through continued cooling and densification of the lithosphere. One might expect a coupling between the height of the sedimentary column deposited at the passive margin and the age-dependent thermal subsidence of the underlying oceanic lithosphere (Turcotte and Ahern, 1977).

In previous studies of the state of stress at passive margins (Walcott, 1972; Bott and Dean, 1972; Turcotte, Ahern, and Bird, 1977; Neugebauer and Spohn, 1978), possible implications of age-dependent properties of the oceanic lithosphere were not considered. Present work attempts to study the interrelations between age-dependent forces, geometry, and rheology and to decipher their net effect on the state of stress at passive margins using the finite element technique for stress calculations. Models presented here deal only with gross features of passive margin evolution and no detailed modelling of a specific margin is intended. As detailed (local) stratigraphy is essential before gravity anomalies can be used as a successful constraint for flexural modeling, we refrained from incorporating gravity data as a constraint in the present work. Subsequently we discuss the implications for initiation of subduction.

## MODELS FOR THE STATE OF STRESS AT A PASSIVE MARGIN

We constructed finite element models for a passive continental margin in four different stages of evolution, at 30, 60, 100 and 200 million years respectively. For all models, a half-spreading rate of 1 cm/year is taken, characteristic for oceanic lithosphere without attached downgoing slabs (Forsyth and Uyeda, 1975). Calculations were carried out with the MARC finite element package (MARC, 1980) using linear strain quadrilateral elements. For more details on the numerical aspects of the work, refer to Cloetingh and Wisse (1981). The model features are summarized in Figure 1.

### Forces

We adopt a triangular sediment wedge at the continental rise as a sedimentary loading model. As our reference, we assume that the maximum sediment thickness at the continental rise corresponds with the thickness that results if the sedimentation keeps up with the subsidence of a boundary layer model of the cooling oceanic lithosphere (Turcotte and Ahern, 1977; Wortel, 1980). For the maximum height of the sedimentary wedge, this implies an increase from 4.5 km at 30 million years to 9.4 km at 200 million years, following roughly a square-root of age relation (Figure 2). The width of the sedimentary wedge is taken to be 300 km, a value typical for wedges at continental rises (Sheridan et al, 1979), and only excess densities are considered. As sediments at the rise replace water we take a $\Delta\rho = 2.4\text{-}1 = 1.4$ g/cu cm. Although related to

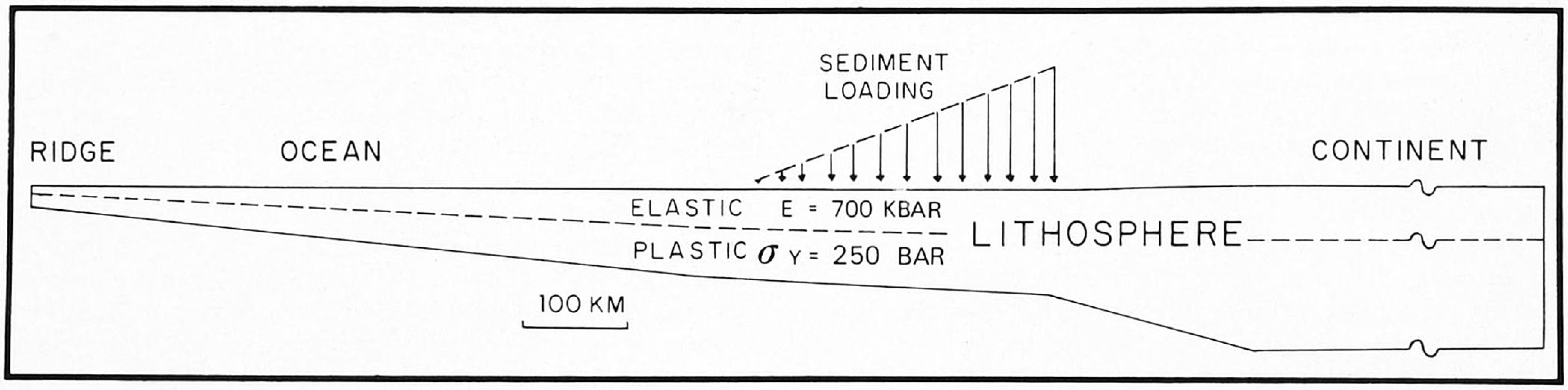


**Figure 1** — Geometry and other features of the models illustrated for a passive margin 100 m.y. old. Vertical scale equals horizontal scale.

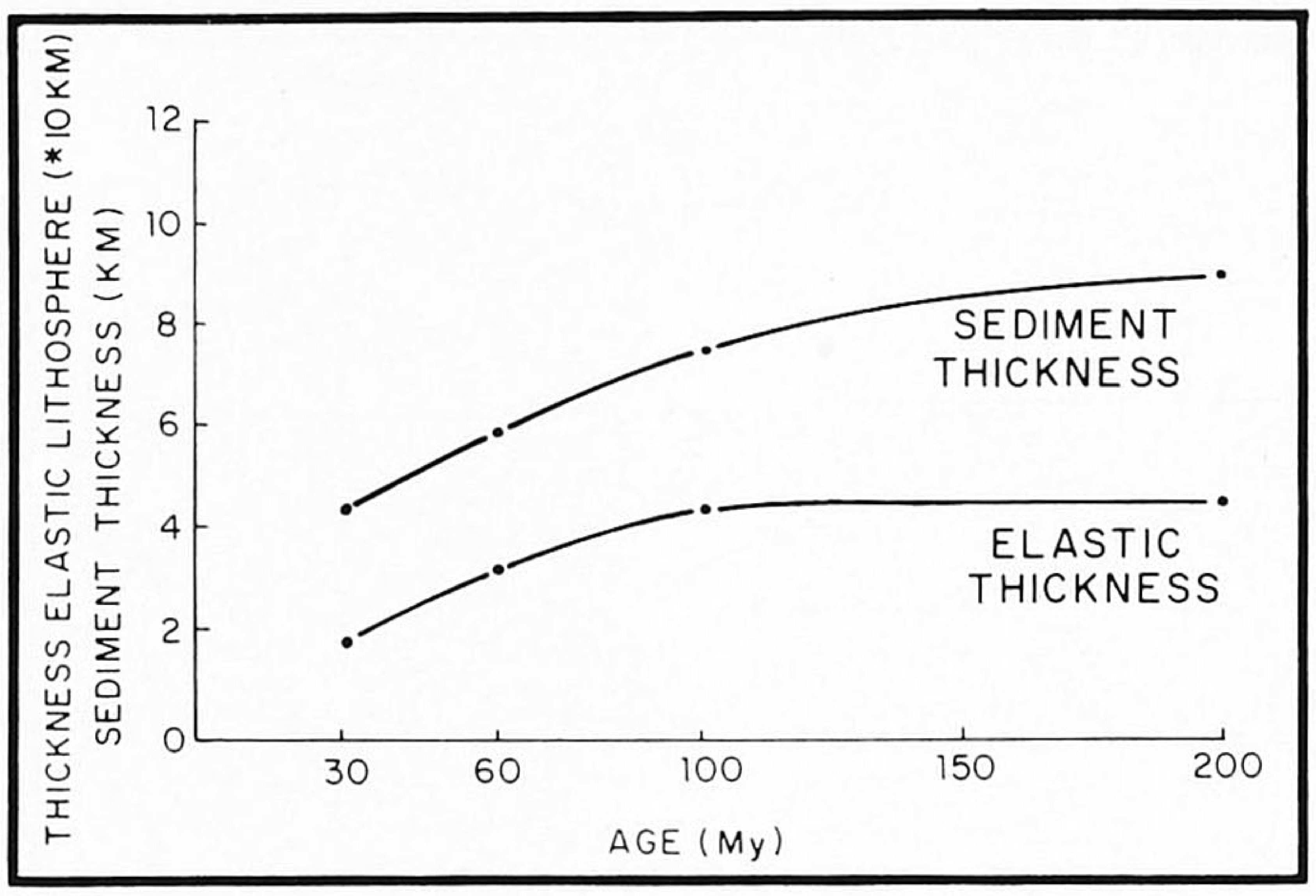


**Figure 2** — Maximum height of the sedimentary prism and the thickness of the elastic part of the lithosphere as functions of age.

age in a somewhat similar way as sediment loading at the rise, sediment loading at the shelf is more difficult to generalize. The position of the shelf edge is not controlled by tectonic subsidence but probably by the local features of sedimentary processes (Watts and Steckler, 1979). On some margins (e.g. off Nova Scotia) the shelf edge has not built out far, it is located landward of the maximum sediment thickness, and the maximum sediment loading is at the rise. The amount of lithospheric thinning under the shelf varies strongly (Watts and Steckler report 10 km thinning off Nova Scotia and 20 km thinning off New York) and is therefore difficult to incorporate in our modeling. Therefore, although in more detailed modelling of specific margins careful attention should be given to sediment loading at the shelf, we refrain from incorporating this feature in our models.

Isostatic forces proportional to the deflection of the lithosphere under loading are included by modifying the stiffness matrix at the model's basal nodes. To suppress rigid body translation, zero horizontal displacements are prescribed for the right hand boundary of the model. The horizontal extent of the continental lithosphere is sufficient to ignore any artificial effects of this boundary condition in the region of interest.

Of the plate tectonic forces implemented in the models, 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 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 artificial stress concentrations at the ridge. The integrated pressure gradient per unit width along the ridge for oceanic lithosphere 100 million years old is $2.3 \times 10^{12}$ N/m. For completeness we assume a resistive drag at the lithosphere's base, acting opposite the spreading direction with a magnitude per unit area of 1% of the pressure exerted by the elevation of the ridge (Solomon, Sleep, and Richardson, 1977).

### Lithospheric thickness and rheology

Age-dependent lithospheric thicknesses are based on Crough's (1975) oceanic lithosphere model and are taken from Wortel (1980). A thickness of 150 km, inferred from a number of independent geophysical approaches (e.g. Pollack and Chapman, 1977), is assigned to the continental lithosphere.

Turcotte, Ahern, and Bird (1977) argue that oceanic and continental lithosphere at a passive margin are effectively decoupled due to a marginal fault system. Since depth and decoupling of such a fault zone are unclear, we model the transition between oceanic and continental lithosphere as continuous. We adopt a width of 200 km for the transition, taking into account the reported evidence for the presence of rift-stage crust at the margin (e.g. Hutchinson et al, this volume). Implications of fault systems present at the margin are briefly discussed further on.

Based on evidence from studies on seismicity (Chapple and Forsyth, 1979) and flexure (McAdoo, Caldwell, and Turcotte, 1978) of oceanic lithosphere at trenches, we model the oceanic lithosphere as a plate with an elastic upper layer (Young's modulus $E = 7 \times 10^{10}$ N/m$^2$, Poisson's ratio $\upsilon = .25$) and a perfectly plastic lower part. From studies on steady-state flow of rocks (Mercier, Anderson, and Carter, 1977), we adopt a yield stress of 250 bar for the lower part of the oceanic lithosphere. Based on data on the width of the outer rise seaward of oceanic trenches, Caldwell and Turcotte (1979) argue that the thickness of the elastic upper part of the oceanic lithosphere is a strong function of its age. The thickness increases approximately proportional to the square-root of age function, from a few km near the ridge to 45 km for ages round 100 million years, with negligible increase beyond 100 million years (Figure 2). These values are confirmed by estimates inferred from microrheology of olivine by Beaumont (1979). One might argue that if the base of the elastic lithosphere coincides with the depth to an isotherm, the thermal effect of sediment loading may reduce the elastic thickness of the lithosphere. However, since the reduction in elastic thickness is relatively small (at most 15%) we do not incorporate this effect in the models. As a result, the stresses inferred from our calculations are somewhat conservative.

Because of cooling, the lithosphere's elastic upper part thickens with age by a continuous transition from a plastic to an elastic constitution. Stresses induced by loading, which are higher than the yield strength, can only be accommodated once the material has undergone the plastic-elastic transition. For our reference model of sediment loading, this implies that only the part of the sedimentary wedge deposited in an interval between ages $t_1$ and $t_2$ effectively determines the state of stress at $t_2$ in the bottom part of the elastic lithosphere (that is created between $t_1$ and $t_2$). This follows from the fact that loading prior to $t_1$ only induces stresses up to the yield strength in the plastic lithosphere. Therefore, the stresses calculated with our "static" models for the bottom of the elastic lithosphere tend to be overestimated. Numerical experi-

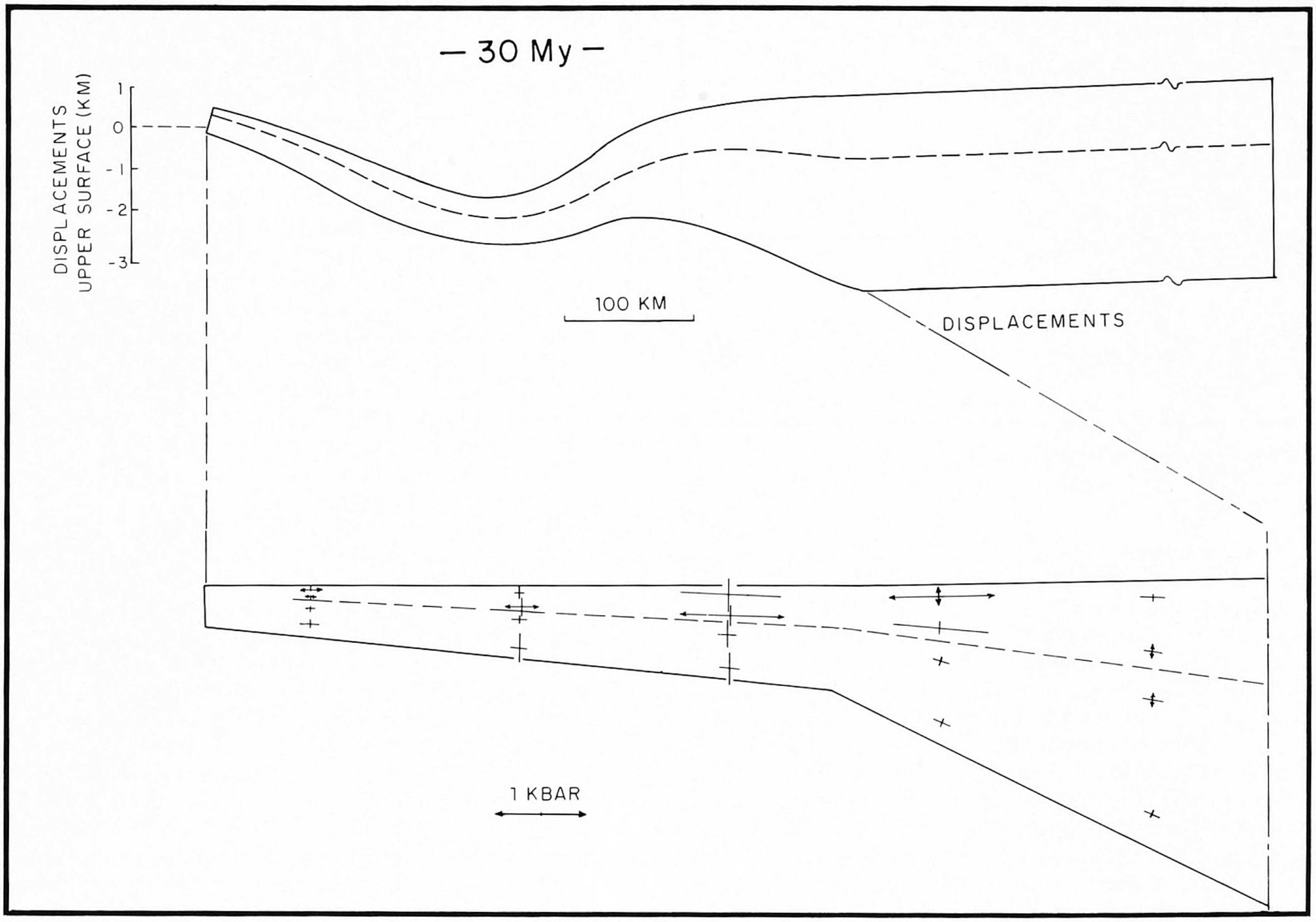


**Figure 3a** — Displacements and stresses calculated for a passive margin 30 m.y. old, based on the reference model of sediment loading given in Figure 2. Stresses are plotted only for the parts of the lithosphere where displacements are significant. These sections are bounded by the dashed lines. The rheological boundary between the elastic upper part and the plastic lower part of the lithosphere is marked by a broken line. Above: Displacements of the lithosphere (km). Note that the scale of displacements (vertical axis) and geometry (indicated by a horizontal bar at the bottom) are not the same. Below: Principal stresses (kbar) at the margin. Line with arrow on each end denotes tension, single line denotes compression. Scale indicated at the lower part of the figure.

ments showed an overestimation in the magnitudes of the stress maxima, for the four ages considered, varying between 15 and 22%.

A thickness of 50 km is assumed for the elastic part of the adjacent continental lithosphere, based on Haxby, Turcotte, and Bird (1976).

## Results

Stress calculations for a passive margin are made in four different stages of evolution. The reference model of sediment loading and the age-dependence of the elastic thickness and other features (as given in Figures 1 and 2) are incorporated into these calculations. The results for ages of 30 and 100 million years are given in Figures 3a and 3b respectively. The deformation of the lithosphere and the resulting stress field (order of magnitude a few kbar) is dominated by sediment loading at the rise; the contribution of the plate tectonic forces to the stress field is of smaller magnitude. Differential stresses are largest at the points of maximum flexure. Tensional stresses up to 3 kbar are generated at the base of the elastic lithosphere. Note once again that the magnitudes of tensional stresses at the bottom of the elastic lithosphere are somewhat overestimated. Stresses are concentrated in the elastic part of the lithosphere due to dividing the lithosphere into an elastic and a plastic part (e.g., Kusznir and Bott, 1977). The displacements outside the bending area shown in Figure 3b express the equilibrium of the forces associated with the negative buoyancy of the (unstable) oceanic lithosphere and the opposing isostatic forces. The resulting depth profile is a well-known feature of oceanic basins (e.g. Parsons and Sclater, 1977). A comparison of Figures 3a and 3b shows that the state of stress at a passive margin depends on the age of the adjacent oceanic lithosphere. Note that as sediment loading on the shelf is related to age in a similar way as sediment loading at the rise, incorporating this feature in the models would not alter this conclusion. To summarize this dependence

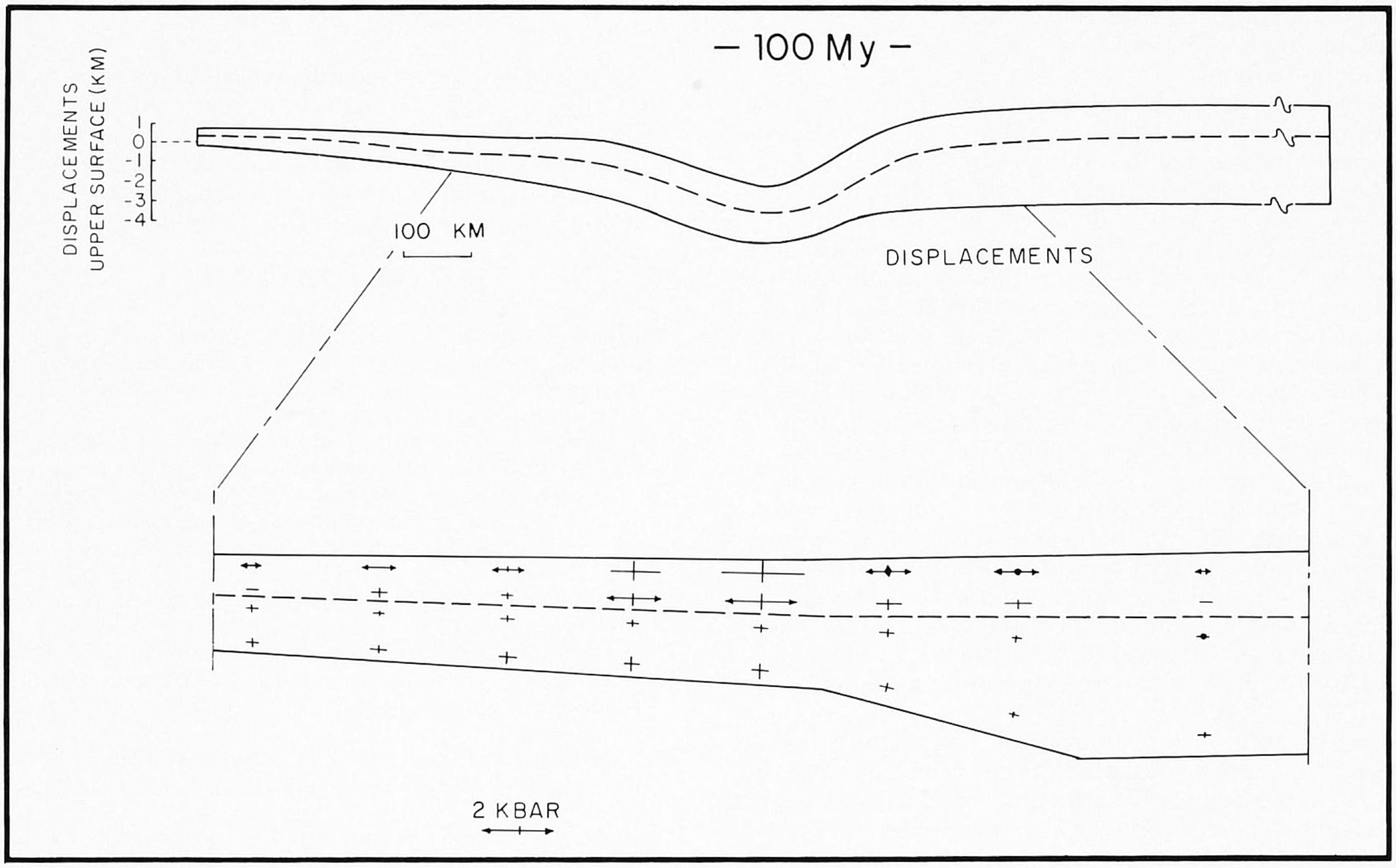


**Figure 3b** — Displacements and stresses calculated for a passive margin 100 m.y. old. Figure conventions as in Figure 3b.

we plotted (Figure 4) the maximum differential (tensional) stresses ($\sigma_1 - \sigma_3$) as a function of age for all four cases. The age-dependence is strongest for ages below 100 million years. From 30 to 100 million years, an interval during which both the sedimentary loading and the elastic thickness increase (Figure 2), the differential stress maxima also increase with age. From 100 to 200 million years, the elastic part of the lithosphere is at a constant thickness (Figure 2). The increase in sediment load, according to our time-dependent sedimentation model, results only in a minor increase of the stresses.

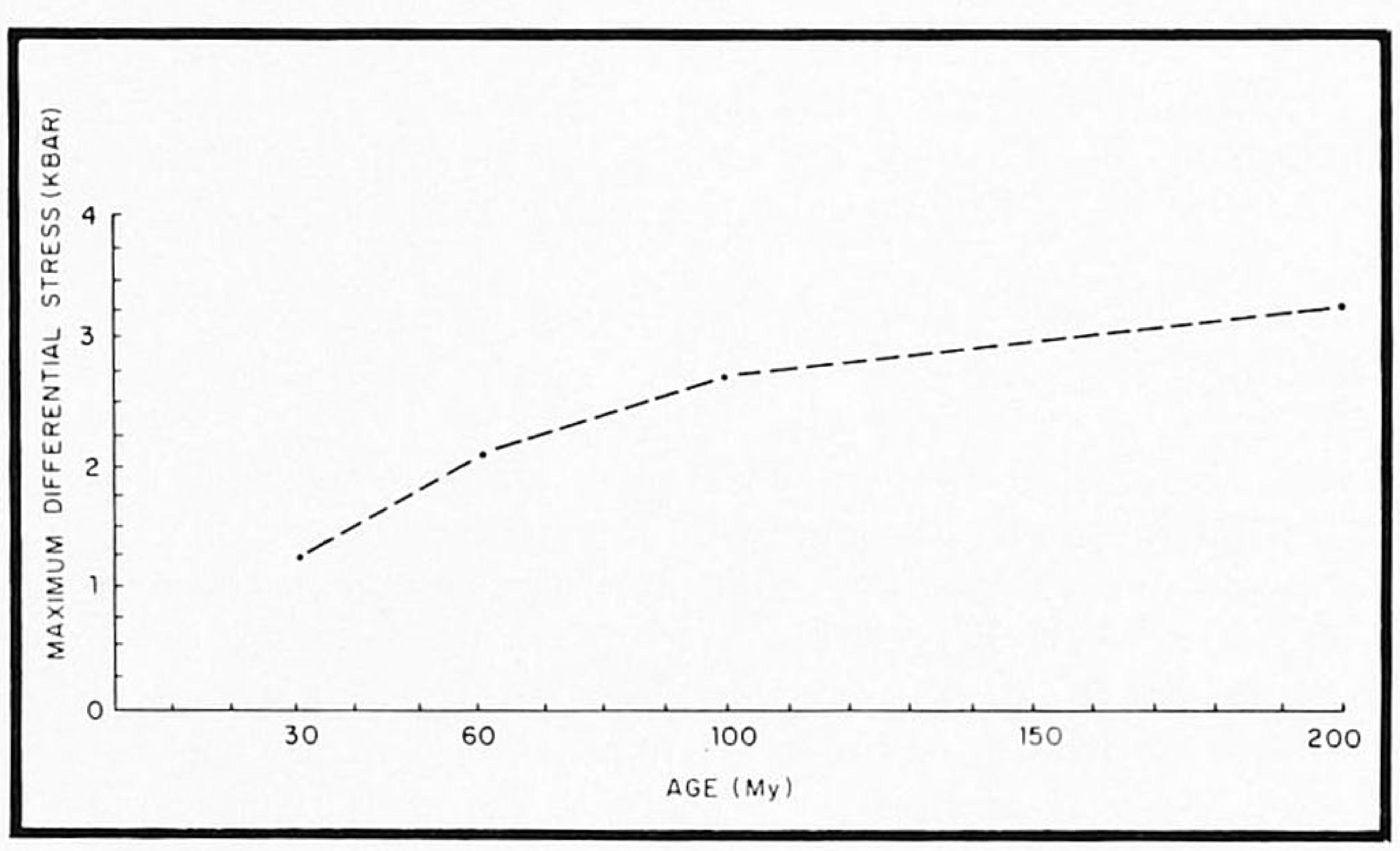


**Figure 4** — Maximum differential stresses as a function of age.

## DISCUSSION

Our calculations for the reference model of sediment loading show that stresses up to 3 kbar can be generated at a passive margin. However, can stresses of this magnitude create failure zones in the lithosphere? The critical point in answering this question is the strength of the elastic part of the lithosphere. Marine geophysical work on seamount loading and lithospheric bending at trenches (Watts and Cochran, 1974; Caldwell and Turcotte, 1979; Watts, Bodine, and Steckler, 1980) shows that oceanic lithosphere can support stresses of several kbars for long periods of geological time ( >50 million years). Pertinent laboratory experiments on olivine (e.g., Ashby and Verrall, 1978; Evans and Goetze, 1979; Kirby, 1980) point to yield stresses of about 3 to 10 kbar. Consider the laboratory data as an upper limit on the strength of the lithosphere (Paterson, 1979); nevertheless, a gap remains between the magnitude of the stresses and the strength of the elastic part of the lithosphere. The stresses generated according to our models are for a reference model of sediment loading. Deviations from the reference model occur in thicker sediment loads (e.g. the eastern U.S. margin with up to 10 km of sediments at the continental rise) and virtually no sediment loading (e.g. the starved southwest European margin). As a result of the age-dependence of the thickness, a surplus sediment load, above the load adopted in the reference model, effectively creates higher stresses when deposited on a young margin.

Several other mechanisms, contributing to stress and not considered here, might be envisaged: thermo-elastic stresses due to cooling of the plate (Turcotte, Ahern, and Bird, 1977), stresses due to crustal thickness inhomogeneities (Bott and Dean, 1972), and stresses generated by phase changes in the lithosphere under the influence of sediment loading (Neugebauer and Spohn, 1978). However, thermo-elastic stresses (order of magnitude several kbars) are probably largely relieved in the early stages of lithospheric evolution. The exact mechanism of relaxation of thermo-elastic stresses is uncertain, but fracture zones formation (Turcotte, 1974) is one possible mechanism. Stress contributions due to changes in crustal thickness at the ocean-continent boundary are of relatively minor importance. This mechanism results in shear stresses about a few hundred bars in the continental crust, while the contribution to the stresses in the oceanic lithosphere is negligible (Bott and Dean, 1972). The effect of a hypothetical phase change is even smaller, only a few tens of bars (Neugebauer and Spohn, 1978).

From the previous information you might conclude that the stresses calculated for our models do not provide upper bounds. On the other hand the strength of the lithosphere may be locally reduced by the occurrence of fossil fault systems at passive margins. So in some circumstances, stresses might be generated at passive margins quite close to, or even exceeding, lower bounds of estimates on the lithospheric strength; this allows the creation of failure zones in the lithosphere and initiates the subduction process.

In general, an additional cause must trigger the subduction process. Deviations from the two-dimensional situations represented by our models may cause geometrical focusing. At certain stages in the process of global plate reorganization, plate tectonic stresses may be concentrated locally to a level comparable with the stresses resulting from sediment loading. The short episode (latest Cretaceous-Eocene) of activation of the northern Iberian passive margin inferred by Boillot et al (1979) might have been caused by this mechanism. The northeastern Indian Ocean might provide a similar example of passive continental margin deformation. Here the onset of the deformation seems to be related to the Himalayan orogenic stage of the collision between India and Asia (Weissel, Anderson, and Geller, 1980). In this region continental rise sediments up to 12 km thick are present (Curray and Moore, 1974). This margin apparently combines numerous features, making it an interesting case for further study of the preparatory stages of subduction initiation.

In summary, our calculations show that tensional stresses up to 3 kbar can be generated at passive margins. If subduction has not begun after 100 million years of evolution, continued aging of the passive margin alone does not provide conditions significantly more favorable for initiation of the subduction process.

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