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ON THE MECHANICS OF LITHOSPHERIC STRETCHING AND DOMING: A FINITE ELEMENT ANALYSIS[1]

SIERD CLOETINGH[2] & FRANK NIEUWLAND[2]

ABSTRACT

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To investigate the thermo-mechanical aspects of lithospheric thinning and rift formation we have constructed finite element models for doming and stretching. We have found that the magnitude of stretching forces required to induce failure in oceanic lithosphere is proportional to lithospheric age. Updoming of the lithosphere caused by a temperature perturbation in the upper mantle can be an effective mechanism for rifting only when updoming is associated with a reduction in the thickness and the strength of the mechanically strong part of the lithosphere. These findings explain why relatively few uplifts generate new rifts.

INTRODUCTION

In recent years considerable attention has been paid to processes associated with lithospheric thinning and rift formation (e.g. FISCHER & JUDSON, 1975; RAMBERG & NEUMANN, 1978; ILLIES, 1981; MORGAN & BAKER, 1983). This is not surprising, considering the implications of lithospheric thinning for sedimentary basin formation and evolution (WATKINS & DRAKE, 1983) Two models dominate the present-day literature on this topic. In the first model updoming of the lithosphere takes place due to a temperature perturbation in the upper mantle (Fig. 1a). Subsequently, the thickness of the uplifted lithosphere is reduced by subaerial erosion. Additional thinning of the lithosphere upon asthenospheric upwelling may occur as a result of its effect on the temperature distribution of the overlying lithosphere. This model has been strongly advocated in the early seventies (SLEEP, 1971; KINSMAN, 1975), especially in the context of studies on the subsidence of passive continental margins.

The fact, however, that evidence for erosion is often absent, together with the observation of listric faults on high quality seismic sections of sediment-starved passive continental margins (e.g. MONTADERT ET AL., 1977), gave strong support to an alternative mechanism, in which lithospheric thinning takes place by stretching (Fig. 1b). This mechanism was originally proposed by ARTEMJEV & ARTYUSHKOV (1971) and analysed by MCKENZIE (1978). McKenzie's investigation in particular initiated a flow of papers on subsidence studies of passive margins and intra-cratonic basins, applying the stretching hypothesis (e.g. ROYDEN & KEEN, 1980; SCLATER ET AL., 1980; LE PICHON & SIBUET, 1981). However, several workers pointed out that unrealistically large amounts of lithospheric thinning are required to explain the observed subsidence (LE PICHON ET AL., 1982; ZIEGLER, 1982).

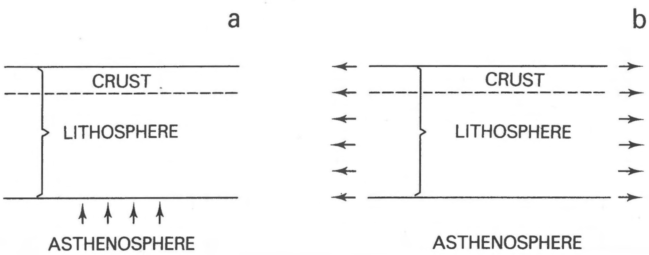

Fig. 1
Schematic representation of the two mechanisms for lithospheric thinning investigated in this paper. Forces are indicated by arrows. a (left): Lithospheric doming. b (right): Stretching of the lithosphere.

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[2] Dept. of Theoretical Geophysics, Institute of Earth-Sciences, University of Utrecht, Budapestlaan 4, 3584 CD Utrecht, The Netherlands.

In the arguments concerning the doming and stretching hypotheses, very little attention has been paid up till now to the material properties of the lithosphere. This is particularly striking in view of the significant progress which has been made in understanding lithospheric rheology. Only very recently, the role thermo-mechanical properties play in the process of lithospheric stretching has been appreciated (BEAUMONT ET AL., 1982; VIERBUCHEN ET AL., 1983; ENGLAND, 1983). Previous studies of the mechanical response of the lithosphere to doming (NEUGEBAUER, 1978; BRIDWELL, 1978; BOTT & KUSZNIR, 1979; WITHJACK, 1979) have been made for simplistic rheological models of the lithosphere, adopting either uniform viscoelasticity, a purely viscous or a uniform elastic rheology for the lithosphere. Therefore, at present the challenge still exists to quantify stresses induced by doming (CROUGH, 1983).

In the present paper we report on model calculations which we performed to investigate the interplay between the forces responsible for thinning of the lithosphere and its material properties. In this study we follow a finite element approach (see CLOETINGH ET AL., 1982, 1983; CLOETINGH, 1982), which allows us to take into full account depth-dependent rheological properties of the lithosphere and laterally varying forces. We investigate whether the stresses generated by the different mechanisms are capable to induce failure and rifting in the lithosphere. To that aim we restrict our attention to the instantaneous response of the lithosphere to the applied forces. The subsequent evolution of the mechanically thinned lithosphere is a subject that falls outside the scope of the present paper.

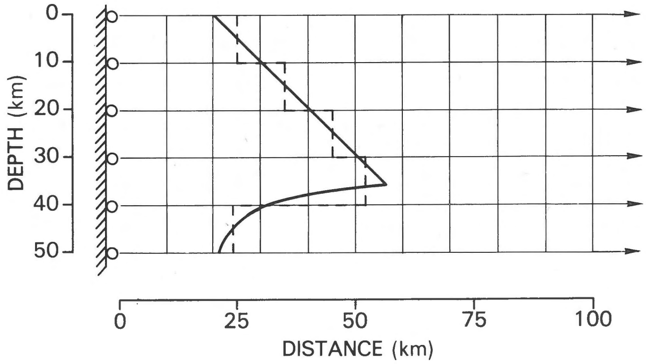

Fig. 2
The finite element configuration for lithospheric stretching of 100 Ma old oceanic lithosphere. We consider the case of instantaneous uniform extension with a non-uniform rheology of the lithosphere including a brittle-ductile transition. Arrows denote the stretching forces. Due to symmetry only half of the structure is incorporated in the model. At the left hand boundary only vertical displacements are allowed. The solid line is the strength envelope. Maximum tensile strength is 8 kbar. The dashed line is the average strength per element implemented in the finite element calculations.

MODEL CALCULATIONS AND RESULTS

General features of the models

In our study we investigate the stress fields generated by doming and stretching for both oceanic and continental lithosphere. Stresses induced in the lithosphere are supported by its mechanically strong upper part, the rheological structure of which consists of two sections: a top section in the brittle regime, with a strength increasing strongly with pressure, and a lower section where the effects of temperature become dominant and stresses are limited by ductile flow (GOETZE & EVANS, 1979). 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).

Thermal models for the *oceanic* lithosphere (CROUGH, 1975) indicate that the base of this part of the lithosphere coincides with an isotherm of about 600-700°C. Studies of the flexural response of the lithosphere to seamount loading and to bending at trenches (BODINE ET AL., 1981) and an analysis of the depth distribution of oceanic intra-plate seismicity (OKAL, 1983), yield an increase in thickness of the MSL from a few km near the spreading ridge to approximately 50 km at an age of

Table 1

Layering	Petrology	Depth range (km)	Young's modulus ($\times 10^2$ kbar)	Density (G cm^{-3})	N	Parameters Creep-laws: Q_w (kcal mol^{-1})	$\dot{e}_o$ (s^{-1}bar^{-N})	Q_D (kcal mol^{-1})	σ_P (kbar)	$\dot{e}_{o2}$ (s^{-1})
Upper crust	quartzite	0-15	2.5	2.6	2	40	1.1×10^{-5}			
Lower crust	diabase	15-30	3.5	2.9	3.4	62	9×10^{-8}			
Lithospheric mantle	olivine	30-150	7.0	3.35	3	122	70	128	85	5.7×10^{11}

Elastic constants, densities and parameters of ductile flow laws incorporated in the simplified rheological model for the continental lithosphere. A strain-rate $\dot{e}$ of $10^{-14}s^{-1}$ is adopted throughout the work. N, Q_w and $\dot{e}_{o1}$ are the exponent, activation energy and pre-exponential constant used in the constitutive equation for power-law creep:
$\dot{e} = \dot{e}_{o1}\, \sigma^N \exp[-Q_w/RT]$ where σ is the differential stress (bar), R the universal gas constant and T the temperature (K). $\dot{e}_{o2}$ and σ_P are constants used in the Dorn-creep equation: $\dot{e} = \dot{e}_{o2} \exp\{ -[Q_D(1-\sigma/\sigma_P)^2/RT] \}$ where Q_D is the activation energy for Dorn-creep. Creep data are taken from the following sources: quartzite (Shelton & Tullis, 1981); diabase (Shelton & Tullis, 1981); olivine (Goetze, 1978). Note the considerable differences in activation energies for creep in upper- and lower crust and upper mantle. Relaxed values (Anderson & Minster, 1980) for Young's modules compiled by Stacey (1977) are used.

100 Ma. These findings are consistent with rheological models (CLOETINGH, 1982) based on the laboratory experiments on dry olivine by GOETZE (1978) and GOETZE & EVANS (1979). Rheological profiles for oceanic lithosphere with widely different ages have been constructed in CLOETINGH (1982) and CLOETINGH & WORTEL (1982), where a more detailed account of the rheology of oceanic lithosphere can be found. For olivine, the ductile flow can be described by a power-law creep relation and a Dorn creep law for stresses below and in excess of 2 kbar, respectively (GOETZE, 1978). These particular creep laws are specified in Table 1. The depth of the brittle-ductile transition (see Fig. 2) is a function of temperature. At that depth the maximum strength in tension (σ_{YT}) is reached. σ_{YT} increases according to a square-root function of lithospheric age (CLOETINGH, 1982) from a few kilobars for young oceanic lithosphere to values of approximately 8 kbar for 100 Ma old oceanic lithosphere.

The *rheology of continental lithosphere* is subject to considerable uncertainties, and is obviously more complex than the rheology of oceanic lithosphere. Recent independent evidence from widely different fields, however, suggests that the major rheological characteristics recognized for oceanic lithosphere apply also to continental lithosphere. The distribution of depths and magnitudes of earthquakes in continental lithosphere is consistent with a depth-dependent rheology (MEISSNER & STREHLAU, 1982). Flexural studies of sedimentary basins located on continental lithosphere (WATTS ET AL., 1982) agree with the existence of an MSL, which thickness is temperature dependent and which cleary shows an increase with geological age. A value of 150 km inferred from a number of independent geophysical approaches (SCLATER ET AL., 1981; MORGAN, in press) is assigned to the total thickness of the continental lithosphere.

We adopt the average model for continental rheology given in Table 1, that is based on recent laboratory studies of the deformation of rocks of the crust and upper mantle. Three layers are present in the model: 15 km upper crust, 15 km lower crust and 120 km lithospheric mantle. Compared with its apparently minor influence on the rheology of the oceanic lithosphere (McNUTT & MENARD, 1982), the weakening effect of water on the rheology of continental upper crustal rocks is more substantial (BRACE & KOHLSTEDT, 1980). Therefore, power-law creep parameters for quartzite (SHELTON & TULLIS, 1981), which mechanical behaviour appears to lie between that of wet and dry granite (CARTER ET AL., 1981), are adopted to describe ductile flow in the upper crust. For the ductile rheology of the lower crust, we adopt power-law creep parameters for diabase, given by SHELTON & TULLIS (1981). Ductile flow in the lithospheric mantle is described as above by GOETZE's (1978) flow laws for dry olivine. Although clearly a simplification, the model incorporates the gross features of continental rheology. Therefore, it is considered a proper starting point for the stress analysis presented in this paper.

The *finite element method* (see for an elaborate discussion ZIENKIEWICZ, 1977) provides an adequate tool for incorporating the material properties discussed above into models for lithospheric stretching and doming. These models are constructed for a vertical 2-dimensional (plane-strain) cross section through the lithosphere. We have used quadrilateral elements with a quadratic displacement field (linear strain). This element type is particularly suited for modelling lithospheric processes (CLOETINGH, 1982). The rheology has been implemented in the models by attributing a constant strength to each different element. By employing at least five layers of elements in the vertical direction of the MSL a sufficiently accurate approximation of the yield envelope was achieved. The finite element calculations were carried out with a modified version of the NONSAP package (BATHE ET AL., 1974). The plane strain equations for material non-linearity were solved using incremental time-step procedures. Convergence tests and checks of the internal reaction forces of the models showed that the finite element meshes presented in the following sections were sufficiently fine to permit an accurate analysis.

Mechanical models for stretching

Fig. 2 illustrates several features of the finite element models for stretching. Note that owing to the symmetry of the problem, the mesh is restricted to the right half of the modelled structure. For the same reason, at the left boundary only vertical displacements are allowed. The width of the structure varies from 100 to 300 km. Also shown is the finite element representation of the material properties of the MSL. We consider the case of instantaneous uniform extension (SCLATER ET AL., 1980) with a non-uniform rheology of the lithosphere including a brittle-ductile transition.

JARVIS & MCKENZIE (1980) have demonstrated that extension can be considered to be instantaneous if the stretching event lasts less than 20 Ma. We have investigated lithospheric deformation at the onset of the stretching process. Lithospheric failure begins at the upper and lower boundaries of the MSL where strengths in tension (σ_{YT}) are relatively low. While loading proceeds, stresses are concentrated in the central part of the MSL. We have calculated the stretching force F required to generate failure in a vertical column of the MSL (Fig. 3). This is given by the following expression:

$$F = \int \sigma_{YT} dz, \tag{1}$$

where the integration is over the thickness z. Force F is linearly related with age, for ages up to 70 Ma with a small deviation from linearity above 70 Ma, the latter due to a slower increase of lithospheric growth. Fig. 3 shows that stretching forces with magnitudes varying from 5×10^{12} N/m to 20×10^{12} N/m are required to induce failure of oceanic lithosphere. Based on the similarity of the thermal structures of continental and old oceanic lithosphere (SCLATER ET AL., 1981) it is to be expected that the mechanical response of continental lithosphere to stretching does not deviate signifi-

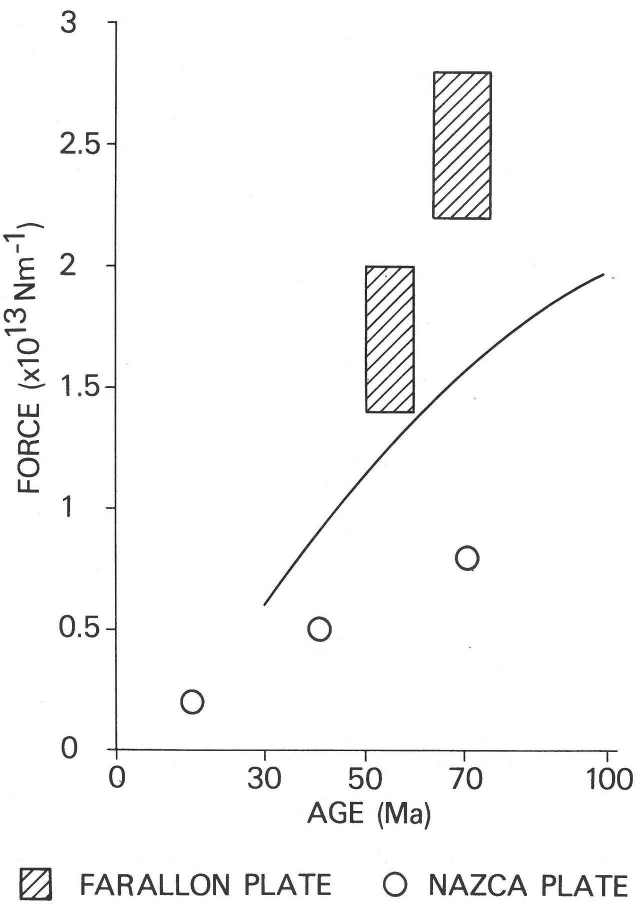

Fig. 3
The force F required to induce lithospheric failure in oceanic lithosphere plotted as a function of lithospheric age. Boxes indicate stress levels prior to the break-up of the Farallon plate into the Nazca plate and the Cocos plate, calculated by Wortel & Cloetingh (1981). Circles indicate (intra-plate) stress levels generated in the present day Nazca plate (Wortel & Cloetingh, 1983). The values referred to are representative of large parts of the Farallon plate and Nazca plate, respectively.

cantly from that observed for old oceanic lithosphere (ages in excess of 100 Ma).

It should again be noted that our calculations are restricted to the case of instantaneously applied stretching forces. If stretching does not take place instantaneously (JARVIS & McKENZIE, 1980), the effect of cooling on the rheology of the stretched lithosphere must be considered. ENGLAND (1983) has shown that for stretching events of 20 Ma and longer, the stiffening of the lithosphere, due to cooling may even inhibit further thinning. Therefore, the results of our calculations provide underestimates of the magnitudes of the stretching forces required for lithospheric failure.

The estimates of the magnitudes of the stretching forces and the associated stresses are particularly interesting in view of the results of an earlier study of rift formation in oceanic lithosphere by WORTEL & CLOETINGH (1981). They showed that a lateral variation of forces along a subduction zone could generate stretching forces with magnitudes varying from 14×10^{12} N/m to 28×10^{12} N/m (see Fig. 3) inside the oceanic plate attached to a subducting slab. In Fig. 3 we have plotted values for stretching forces induced by this mechanism in the present-day Nazca plate (WORTEL & CLOETINGH, 1983) and its predecessor the former Farallon plate (WORTEL & CLOETINGH, 1981). A comparison of these data with the model curve shows that stretching forces generated in the Farallon plate were capable to induce its observed break-up. Furthermore, the magnitude of the forces in the Nazca plate falls consistently below the model curve which explains why the Nazca plate has not undergone the same fate as the Farallon plate.

From the above we may, therefore, conclude that at least one mechanism is available to generate stretching forces of the right order of magnitude to induce lithospheric failure and rift formation in oceanic lithosphere. However, no satisfactory mechanism has been worked out sofar that would generate significant stretching in an *intra-cratonic* setting (e.g. SCLATER ET AL., 1980).

Mechanical models for doming

Updoming of the lithosphere is caused by the effect of temperature perturbations on the density stratification of the upper mantle. Magnitudes of the temperature induced buoyancy forces acting on the lower boundary of the lithosphere, are based on thermal calculations reported by WORTEL & WELTEVREDE (1983).

Two classes of models for doming are considered. In the first models we restrict ourselves to the role of temperature perturbations in the asthenosphere as a driving force for uplift of the lithosphere. Subsequently, we incorporate in the models the influence of asthenospheric upwelling on the temperature distribution and thus on the rheology of the overlying MSL. Though doming has previously been advocated as a mechanism for crustal attenuation (e.g. VAN DER LINDEN, 1975), a quantitative investigation of this process has been lacking.

The finite element model for doming is given in Fig. 4a. Symmetry has been taken into account using the technique discussed in the previous section. Uplift forces acting at the centre of the structure are indicated by vertical arrows. The finite element mesh is finest at positions where uplift is maximal. The width of the region where uplift forces are acting varies from 25 to 400 km. The total width of the structure is taken to be sufficient to ensure negligible vertical displacements at the edges. The modelled widths vary as a function of lithospheric age (corresponding with different values for the flexural parameters) from 500 to 700 km. In the process of updoming, water (density ϱ_w) is replaced isostatically by mantle material (density ϱ_m). To incorporate isostasy in the models, springs are placed at the lower boundary of the models with stiffness $K = (\varrho_m - \varrho_w)g$, where g is the acceleration due to gravity, following a technique described by CLOETINGH & WORTEL (1982).

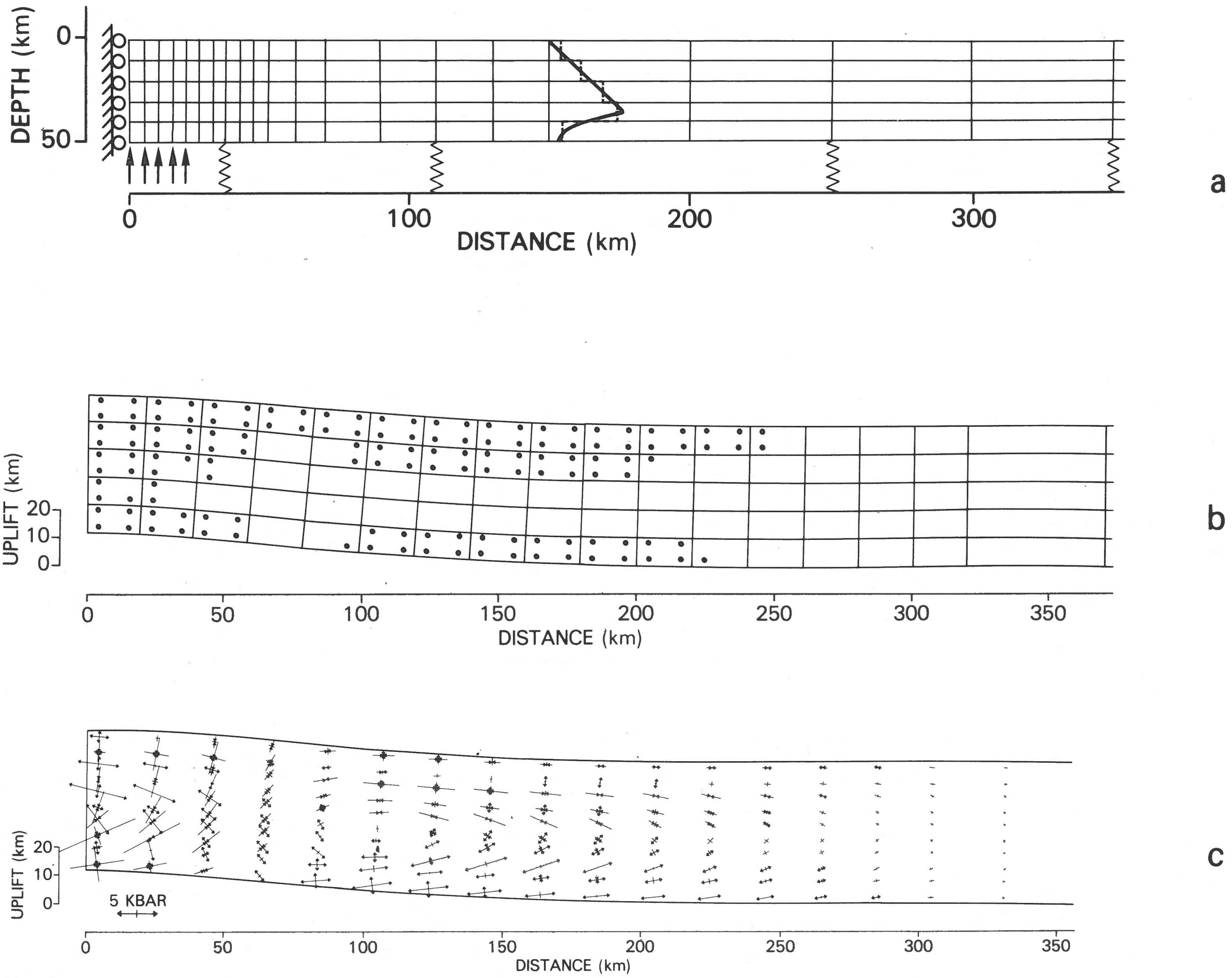

Fig. 4
Model features for doming. a (upper): The finite element mesh for doming of a 100 Ma old oceanic lithosphere. The uplift forces are indicated by vertical arrows. Isostatic forces counteracting the deflection are indicated by springs. Width of the uplift is 50 km. Figure conventions as in Fig. 2. b (middle): Uplift and associated failure of the lithosphere. Note that the scales of the vertical displacements and of the geometry are not the same. Black dots denote integration points where stresses are induced in excess of lithospheric strength. c (below): Stresses induced by updoming according to Fig. 4a. Principal stresses in kbar. Symbols (⟷) and (—|—) denote tension and compression, respectively.

Models for uplift of oceanic lithosphere have been constructed for lithospheric ages varying from 30 to 200 Ma. Figs. 4b and c. illustrate some characteristic features of the mechanical response to doming of a 100 Ma old oceanic lithosphere. Failure induced by doming takes place preferentially in regions that combine a low lithospheric strength and a large uplift. These regions are primarily found at the upper and lower boundaries of the MSL at the point of maximum flexure. For large uplifts stresses can be generated, that are sufficient to induce failure in a vertical column through the lithosphere (Fig. 4b). The stress distribution is shown in Fig. 4c. Note that maximum stresses are not found at the lower and upper boundaries of the MSL as would be the case for a uniform elastic rheology. The width of the region of updoming is an important parameter for the stress distribution. The greater the width of the uplift the lower the curvature. Therefore, stresses will decrease, as a function of width, to the isostatic limit. This is reached for widths exceeding the flexural parameter of the overlying lithosphere (CLOETINGH, 1982).

These points are illustrated in Fig. 5, which summarizes the results of the model calculations for various lithospheric ages. Young oceanic lithosphere is most vulnerable to failure due to uplift. However, the uplift required to induce lithospheric failure is more than 4 km even in the case of young lithosphere and that clearly exceeds geological estimates (WITHJACK, 1979; CROUGH, 1983). Fig. 5 also includes the results obtained for uplift of continental lithosphere. As could be expected, the mechanical responses of old oceanic lithosphere and continental lithosphere to uplift are similar. This explains the observed (CROUGH, 1983) overall similarity in shape and size of oceanic and continental swells.

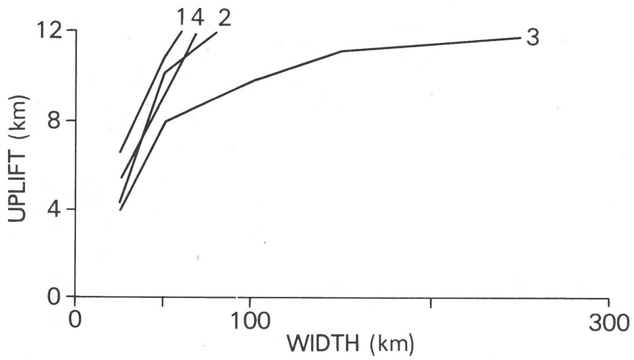

Fig. 5
Uplift required to induce failure in the overlying lithosphere plotted as a function of the width of the area of uplift. 1, 2, 3: oceanic lithosphere of respectively 100, 50 and 30 Ma old, with values for maximum tensile strengths and thicknesses of the mechanically strong part of the lithosphere of 8 kbar, 6.5 kbar, 5 kbar and 50 km, 42 km, 30 km, respectively. 4: continental lithosphere, with maximum tensile strength and thickness of the mechanically strong part of the lithosphere of 7 kbar and 55 km.

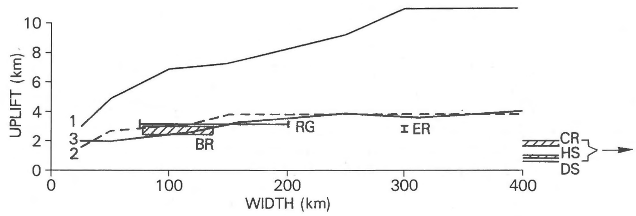

Fig. 7
The uplift required to induce failure in the overlying lithosphere, when the effect of the temperature perturbation on the rheology is taken into account. 1 and 2 denote 100 Ma old oceanic lithosphere reheated to a solidus at depths of 50 km and 35 km respectively (Wortel & Weltevrede, 1983). 3 denotes continental lithosphere thinned to 50 km (Lachenbruch & Sass, 1978). Included in the figure are geological estimates of the geometry of the uplift for the Baikal rift (BR), the Rhine graben (RG) (Withjack, 1979), the Ethiopian rift zone (ER), and the Hoggar Swell (HS), Darfur Swell (DS), Cape Verde Rise (CR) and Hawaii Swell (HS) (Crough, 1983).

We have calculated the material properties of rheologically attenuated oceanic lithosphere using temperature profiles given by WORTEL & WELTEVREDE (1983) for the Hawaii swell. In the Hawaiian case the thickness and strength of the MSL is reduced from values appropriate for 100 Ma old oceanic lithosphere (50 km, resp. 8 kbar) to values that vary from 20 to 30 km and 3.5 to 5.5 kbar, respectively. Rheological profiles for attenuated continental lithosphere are calculated using geotherms for thinned lithosphere of the Basin and Range Province given by LACHENBRUCH & SASS (1978).

Fig. 6 shows the finite element mesh used to calculate the response to doming of rheologically attenuated lithosphere. The results of this class of calculations are summarized in Fig. 7. That figure shows that the reduction in thickness and strength of the MSL in the region where uplift is taking place, strongly decreases the amounts of uplift required to induce lithospheric failure, which brings them much closer to geological estimates for uplift. For a comparison with oceanic swells we select the case of Hawaii, where uplift is particularly well constrained by geological and geophysical data. The width of the Hawaii swell is of the order of 400-600 km with a maximum uplift of 1.5 km (DETRICK ET AL., 1981; WORTEL & WELTEVREDE, 1983). According to our modelling, however, a 4 km uplift is necessary to induce lithospheric failure. Therefore, we suggest that uplift alone is not sufficient to induce lithospheric failure in the Hawaiian case. In addition to uplift, tensile stresses associated with plate reorganizations in the Pacific (WATTS ET AL., 1980) seem to be required. This agrees with the mechanism worked out for the Farallon plate (WORTEL & CLOETINGH, 1981).

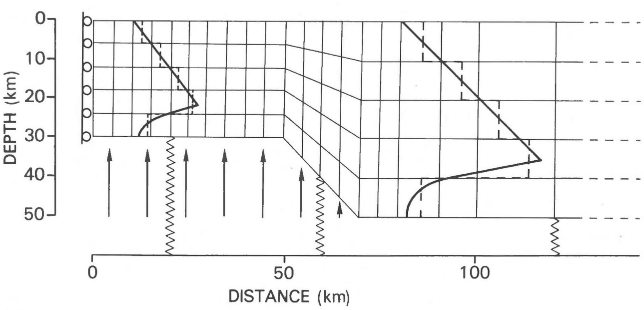

Fig. 6
The finite element mesh for the updoming of a rheologically attenuated lithosphere. Note the differences in the thickness and strength of the mechanically strong part of the lithosphere inside and outside the area of uplift. Figure conventions as in Figs. 2 and 4a.

Figure 7 also gives the results of the calculations carried out for the case of doming of rheologically attenuated continental lithosphere. As a result of its rheological layering the effect of a temperature perturbation on continental lithosphere is more severe than the effect on oceanic rheology. Strong temperature perturbations will create minima in the lithospheric strength distribution near the boundaries op upper and lower crust, and of lower crust and lithospheric mantle. This is due to different melt temperatures of upper crustal, lower crustal and mantle rocks. A strong temperature perturbation will shift the position of the brittle-ductile transition most effectively for rocks with a low melting point. This effect reduces the amount of uplift required to induce failure in attenuated continental lithosphere considerably, relatively to the uplift necessary to generate failure in oceanic lithosphere. Depending on the width of the dome, uplifts of 2-3 km are capable of inducing lithospheric failure. These estimates are consistent (see Fig. 7) with uplift data of major continental rift zones summarized by WITHJACK (1979) and CROUGH (1983). This applies in particular to data from the Baikal rift and the Rhine graben systems.

DISCUSSION

Our calculations show that the age of the oceanic lithosphere is one of the main controlling factors with respect to the effectiveness of stretching and doming as mechanisms for lithospheric thinning and rift formation. We have demonstrated that forces of the order of 10^{13} N/m are required to induce lithospheric failure by stretching. As shown, forces of that magnitude can be generated during plate reorganizations in oceanic lithosphere by processes described by WORTEL & CLOETINGH (1981). Such processes form a mechanically sound alternative for theories of rift formation in oceanic lithosphere based on hot spots. For continental lithosphere, however, at present no mechanism for stretching is available to induce stresses of the required order of magnitude.

Our approach is conservative in that we have assumed that the lithosphere is deformed at locations unaffected by earlier doming and/or stretching events. Geological evidence, however, for repeated rifting events is abundant. In particular, studies of the East-African rift system (e.g. BURKE & WHITEMAN, 1973; NOLET & MUELLER, 1982) point to a long history of rifting events since Precambrian. This could have been responsible for a considerable reduction of the thickness of the MSL to values of the order of 12 km determined by SHUDOWSKY (1982).

Weakness zones induced by such a repetition of uplift and/or stretching events might provide favourable locations for the creation of new failure zones. On the other hand, evidence from rifted continental margins shows (STECKLER & WATTS, 1981) that faults associated with earlier break-up phases are locked and mechanically healed within a few million years after the event and after the source of uplift is removed. Therefore, it is anticipated here that unless repeated deformation takes place with only short intervals (order of a few Ma), the assumption concerning the absence of previously induced weakness zones in the models is reasonable.

There are several (second order) effects not incorporated in the models, which might further reduce the estimates of the amount of uplift and the forces required to induce lithospheric failure. Of these, we have mentioned the weakening effect of a 'wet' rheology of the lithosphere. Erosion of continental lithosphere during elevation will further reduce the required uplift. The same can be said for the effects of gravitationally induced forces associated with uplift, which have been investigated by ARTYUSKOV (1973) and BOTT & KUSZNIR (1979). The stresses induced by this mechanism are a factor 5 smaller than flexural stresses associated with doming. Considering this order of magnitude it was justified to neglect them here.

We have shown that the lithosphere will not fail through doming unless that is associated with weakening and thinning of its mechanically strong part. Such strong perturbations of the thermal structure of the lithosphere can be produced by lithospheric doubling and shifting recently proposed by VLAAR (1982, 1983). We have found that narrow domes of a few hundred km wide with an uplift of the order of 2 km are capable of inducing failure in rheologically attenuated lithosphere. This applies in particular to young oceanic lithosphere (age less than 30 Ma) and to continental lithosphere.

A comparison of the results of our model calculations with continental uplift data offers an explanation for the observation (CROUGH, 1983) that relatively few thermally induced domes generate new rifts. Particularly striking in this respect is that uplift of the order of 1 km at the Hoggar swell has not been followed by rifting, whereas uplift of the order of 2 km in the Ethiopian dome has (CROUGH, 1983).

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