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published in

Mathematical Geophysics: a Survey of Recent Developments in Seismology and Geodynamics
1988

document version

Publisher's PDF, also known as Version of record

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citation for published version (APA)

Cloetingh, S. A. P. L., & Wortel, M. J. R. (1988). On the mechanics of plate boundary formation. In *Mathematical Geophysics: a Survey of Recent Developments in Seismology and Geodynamics* (pp. 363-387). Reidel Publ. Co..

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Chapter 16

On the mechanics of plate boundary formation

S.A.P.L. Cloetingh and M.J.R. Wortel

We discuss the mechanics underlying the formation of new plate boundaries. Analysis of the relation between intraplate stress fields and lithospheric rheology leads to greater insight into the role that initiation of subduction and initiation of rifting play in the tectonic evolution of the lithosphere. Concentration of slab-pull forces is an effective mechanism for passive rifting of oceanic and continental lithosphere. Initiation of subduction at passive margins requires in general the action of external plate-tectonic forces, which will be most effective for young passive margins pre-stressed by thick sedimentary loads. Plate reorganizations occur predominantly by the formation of new spreading ridges, because stress relaxation in the lithosphere takes place much more efficiently through this process than through the formation of new subduction zones.

1. Introduction

Initiation of rifting and initiation of subduction are key elements in plate-tectonic schemes for evolution of the lithosphere. The importance of rifting processes is reflected in the considerable body of work devoted to this subject during the last decade (e.g. Palmason, 1982; Morgan and Baker, 1983), which has been motivated in part by the role of rifting in the formation of sedimentary basins. The mechanics of the initiation of new subduction zones has remained relatively unexplored (Cloetingh et al., 1982, 1983; Flinn, 1982) by comparison.

Advances in our understanding of lithospheric rheology (Goetze and Evans, 1979; Kirby, 1983) and in the modelling of intraplate stress fields (Wortel and Cloetingh, 1981, 1983; Cloetingh and Wortel, 1985, 1986) allow a unifying approach to the dynamics

underlying the formation of new plate boundaries. In the present paper we concentrate on a comparison of the stresses and strengths in the lithosphere. In our studies we have followed a finite element approach, which allows us to take full account of depth-dependent rheological properties of the lithosphere and laterally varying forces. We investigate whether the stresses generated by the different geodynamic processes are capable of inducing lithospheric failure. Such an analysis will be shown to lead to a better understanding of the differing roles that initiation of subduction and rifting play in plate reorganization processes.

2. Intraplate stress fields and the rheology of the lithosphere

Lithospheric plates are in motion under the influence of plate-tectonic forces: the ridge push, which results from the elevation of the spreading ridge above the adjacent ocean floor and the thickening of the lithosphere with cooling (Hales, 1969); the drag exerted at the base of the lithosphere (Richter, 1973) due to viscous shear forces associated with the motion relative to the underlying asthenosphere; and the pull acting on the downgoing slab in a subduction zone (McKenzie, 1969). In addition to these plate-tectonic forces, there are also forces acting on the lithosphere which induce local stresses. A number of these mechanisms have been reviewed by Turcotte and Oxburgh (1976) and Cloetingh (1982). Among the locally induced stresses those associated with flexure and those induced by cooling of the lithosphere in near-ridge areas stand out in magnitude of stress level. In plates that are involved in collision or subduction processes, however, plate-tectonic forces dominate the stress field. This dominance is greater if there is a concentration of stresses due either to lithospheric age variations encountered along trench systems or to angular and curved geometry of plate boundaries (Wortel and Cloetingh, 1981; Cloetingh and Wortel, 1985).

In the first phase of modelling lithospheric stress fields resulting from plate-tectonic forces, models were tested against focal mechanism data of intraplate earthquakes to infer the relative and absolute importance of various possible driving and resistive forces (Solomon et al., 1975; Richardson et al., 1979). These and several other studies (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977) resulted in the gross understanding that the ridge push and the slab pull were the main driving forces. Two developments, however, were necessary to set the stage for more advanced stress field modelling: 1. Recognition that the ridge push is not to be considered as a constant boundary force but as an integrated pressure gradient (Lister, 1975), which introduces the dependence of the force on the age of the lithosphere. 2. Deeper understanding of the variations in the subduction process of oceanic lithosphere and the associated variations in plate-tectonic forces (Vlaar and Wortel, 1976; Wortel and Vlaar, 1978; Molnar et al., 1979; Wortel, 1980; 1984). These studies demonstrated that age-dependent variations in length, depth and sinking rate affect the pull that a downgoing slab exerts on a converging plate.

By implementing these new features and insights in stress modelling Wortel and Cloetingh (1981, 1983, 1985) and Cloetingh and Wortel (1985, 1986) showed that the dynamical basis of their modelling procedure enabled the resulting stress field to be used to analyze, explain and even predict various deformational processes in lithospheric plates.

In the present study we elaborate on the tectonic effects of intraplate stress fields in the context of *rheological models* for the lithosphere. Stresses in the lithosphere are supported

by its mechanically strong upper part, the rheological structure of which consists in its simplest form of two sections: a top section in the brittle regime, where strength increases rapidly with pressure according to Byerlee's law (Byerlee, 1968, 1978), and a lower section where the effects of temperature dominate and stresses are limited by ductile flow (Goetze and Evans, 1979). Following Bodine et al. (1981) we take here the depth at which the ductile strength is 500 bar as the lower boundary of the mechanically strong part of the lithosphere (MSL). Studies of the flexural response of the *oceanic* lithosphere to seamount loading and to bending at trenches (Caldwell and Turcotte, 1979; Bodine et al., 1981), Seasat altimetry (McAdoo et al., 1985) and analysis of the depth distribution of oceanic intraplate seismicity (Wiens and Stein, 1983) all indicate an increase in thickness of the MSL from a few km near the spreading ridge to approximately 50 km at an age of 100 Ma, which is consistent with extrapolation of results of laboratory experiments on dry olivine by Goetze and Evans (1979). Similarly, according to predictions of rock-mechanics data the strength of the oceanic lithosphere increases according to a square root function of age.

The rheology of *continental* lithosphere involves many uncertainties, and is obviously more complex than the rheology of oceanic lithosphere. 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 the depths and magnitudes of earthquakes in continental lithosphere is consistent with a depth-dependent rheology (Meissner and Strehlau, 1982; Sibson, 1983). Studies of the flexural rigidity of continental lithosphere (Kuznir and Karner, 1985) also agree with the existence of an MSL, the thickness of which is temperature-dependent and consistent with the extrapolation of laboratory data on crustal (Kirby, 1983; 1985) and mantle (Goetze and Evans, 1979) rocks. In the present paper we adopt an average model for continental rheology with a 15 km thick upper crust, 15 km lower crust and 120 km lithospheric mantle. Compared to its apparently minor influence on the rheology of oceanic lithosphere (McAdoo et al., 1985), the weakening effect of water on the rheology of continental upper crustal rocks is more substantial (Brace and Kohlstedt, 1980). Therefore, power-law creep parameters for quartzite (Shelton and Tullis, 1981), which mechanical behavior appears to be 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 have adopted power-law creep parameters for diabase given by Shelton and Tullis (1981). Such a rheology with a relatively strong lower crust made up of mafic minerals, is supported by the occurrence of earthquakes in the lower crust in various continental rift zones (Shudofsky et al., 1987; Fuchs et al., 1987). Intraplate stresses must exceed the strength of the lithosphere in order to give rise to the formation of new plate boundaries. We shall elucidate this by comparing the outcome of the modelling of intraplate stresses with rheological models of the lithosphere. We shall then consider the implications for the initiation of rifting and subduction.

3. Initiation of rifting

In this section we discuss some mechanical aspects of the initiation of passive and active rifting (Turcotte and Emmerman, 1983, see Figure 1). The active rifting model has been strongly advocated in the early seventies (Sleep, 1971), especially in the context of studies on the subsidence of passive continental margins. Since then, the frequently observed

absence of evidence for erosion and the widespread occurrence of listric normal faults on seismic profiles of passive margins gave rise in the late seventies to the increasing popularity of passive rifting models in the interpretation of subsidence of sedimentary basins (see e.g. Watkins and Drake, 1983). More recent studies (Hellinger and Sclater, 1983) have, however, shown that the occurrence or absence of erosion at the rift shoulders cannot be uniquely interpreted in favor of active or passive rifting mechanisms. In the arguments concerning passive and active rifting mechanisms, relatively little attention has been paid to the rheological properties of the lithosphere. Only recently has the role that thermo-mechanical properties play in the process of passive rifting been appreciated (e.g. Vierbuchen et al., 1983; Sawyer, 1985).

Figure 1. Schematic representation of mechanisms for active rifting and passive rifting. Forces are indicated by arrows. a. (left): active rifting. b. (right): passive rifting of lithosphere.

With a few exceptions (e.g. Cloetingh and Nieuwland, 1984; Houseman and England, 1986), studies of active rifting have been restricted to simplistic rheological models of the lithosphere, adopting either uniform viscoelastic, a purely viscous, or a uniform elastic rheology for the lithosphere. Here we concentrate on model calculations which we performed to investigate the interplay between the forces responsible for rifting of the lithosphere and its material properties. We investigate whether the stresses generated are capable of inducing failure and rifting in the lithosphere. 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.

3.1 Active rifting

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 such as reported by Detrick et al. (1981). We have incorporated in the models the influence of asthenospheric upwelling on the temperature distribution and thus on the

rheology of the overlying MSL. In the Hawaiian case the thickness and strength of the MSL is reduced by the reheating process from values appropriate for 100 Ma old oceanic lithosphere (50 km, respectively 8 kbar) to values that vary from 20 to 30 km and 3.5 to 5.5 kbar, respectively. Figure 2 shows part of the finite element mesh used to calculate the response to uplift of rheologically attenuated lithosphere. Details of the finite element calculations are given in Cloetingh and Nieuwland (1984). The results of the calculations are summarized in Figure 3. The Figure shows that even with the reduction in thickness and strength of the MSL in the region where uplift is taking place, the amounts of uplift required to induce lithospheric failure are generally in excess of geological estimates. For a comparison with oceanic swells, the selected case of Hawaii is well constrained by geological 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). According to our modelling, however, a 4 km uplift is necessary to induce lithospheric failure. Therefore, we suggest that active rifting alone is not sufficient to induce lithospheric failure in the Hawaiian case. In addition to uplift, tensional stresses associated with plate reorganization in the Pacific (Watts et al., 1980) seem to be required. This agrees with the mechanism for passive rifting worked out for the Farallon plate (Wortel and Cloetingh, 1981; see below).

Figure 2. Central part of finite element mesh used for analysis of active rifting of a rheologically attenuated lithosphere. Note the differences in strength and thickness of the mechanically strong part of the lithosphere inside and outside the area of uplift.

Figure 3 also gives the results of the calculations carried out for the case of uplift of rheologically weakened 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 lithosphere. Strong temperature perturbations will create minima in the lithospheric strength distribution near the boundaries between upper and lower crust and between lower crust and lithospheric mantle. This is due to differences in melt temperature 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 lithospheric failure in attenuated continental lithosphere considerably, relative 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 failure. These estimates are consistent (see Figure 3) 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 Rhinegraben systems. The calculated estimates are conservative in that we have assumed that the lithosphere is deformed at locations unaffected by earlier rifting events. Geological evidence for repeated rifting events, however, is abundant. Weakness zones created by a repetition of rifting events might provide favorable locations for the creation of new failure zones. On the other hand, evidence from rifted continental margins (Steckler and Watts, 1981) shows that faults associated with earlier break-up phases are locked and mechanically healed within a few Ma after the event if the source of heat is removed. Several other effects might further reduce the estimates of the amount of uplift and the forces required to induce lithospheric failure. These include erosion of continental lithosphere during elevation and the effect of gravitationally induced forces associated with uplift (Artyushkov, 1973). The stresses induced by the latter mechanism are a factor of five smaller than the flexural stresses associated with uplift and doming.

Figure 3. The uplift required to induce failure in the overlying lithosphere, when the effect of the temperature perturbation on the rheology is taken into account. Symbols (1) and (2) denote 100 Ma old oceanic lithosphere reheated to a solidus at depths of 50 km and 35 km, respectively. 3 denotes continental lithosphere thinned to 50 km (Lachenbruch and 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 (Crough, 1983).

In summary, modelling has shown that the lithosphere will not fail through doming unless this process 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 (Vlaar, 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 weakened lithosphere. A comparison of the results of the 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 domes has (Crough, 1983).

3.2 Passive rifting

Wortel and Cloetingh (1981, 1983) proposed a new mechanism for passive rifting of oceanic plates and the subsequent inception of spreading centres under the influence of age-dependent plate-tectonic forces. This mechanism was based on a case study of the well-constrained plate-tectonic evolution of the eastern Pacific, in which the break-up of the Farallon plate into the Cocos plate and the Nazca plate at 25-30 Ma ago (Hey, 1977) played a central role. Using reconstructed geometries and lithospheric age distributions (Handschumacher, 1976) appropriate for the plate configuration just prior to break-up they were able to calculate the distribution of (age-dependent) ridge-push and slab-pull forces on the plate and, consequently, the induced stress field.

The calculated paleo-stress field (σ) is dominated by tensional stresses (see Figure 4). To induce failure in the vertical column of the MSL the integrated force $F = \int \sigma dz$ must exceed the integrated depth-dependent strength $A = \int \sigma_y dz$, where σ_y is the strength profile and z the depth. In Figure 5 we have plotted values for F corresponding with tensional stresses representative of large parts of the Farallon plate. A comparison of these data with (age-dependent) integrated strengths A (solid line) shows that the tensional stresses generated in the plate are capable of inducing its observed rupture and fragmentation. Since the dynamic situation of the Farallon plate and the resulting stress field at that time can, in itself, explain both location and timing of the inception of spreading, it is unnecessary to invoke hotspot activity as a mechanism for its fragmentation (Wortel and Cloetingh, 1981).

Also plotted in Figure 5 are the magnitudes of the integrated forces in the present-day Nazca plate (Wortel and Cloetingh, 1983). Figure 5 shows that the magnitude of the forces in the Nazca plate falls consistently below the integrated strength curve, for ages in excess of 30 Ma, which explains why the Nazca plate has not been subjected to the same fate as the Farallon plate. Note that only the youngest and therefore mechanically weakest part of the plate is near failure. In confirmation of this theory, work on the Nazca plate by Warsi et al. (1983) has shown that a fragmentation process is now active along the Mendana fracture zone, where Wortel and Cloetingh's (1983) model calculations predicted tensional stresses.

Cloetingh and Wortel (1985, 1986) modelled the state of stress in the Indo-Australian plate in order to investigate quantitatively observed variations in tectonic style. As noted by Bergman and Solomon (1985) intraplate earthquakes are a reliable indicator of the intraplate stress field because of the high level of intraplate seismicity in the Indo-Australian plate. Cloetingh and Wortel's (1985, 1986) modelled stress field gives a quantitative explanation for the high level of the stress field, the dominance of compression in the plate's interior (Bergman and Solomon, 1985), the selective occurrence of normal faulting seismicity parallel to its spreading centres (Stein et al., 1987), and the strong variations in observed stress directions. Furthermore, Cloetingh and Wortel (1985, 1986) showed that tensional stresses induced by slab-pull forces associated with subduction of old oceanic lithosphere at various segments of the convergent boundaries of the Indo-Australian plate are transmitted into the interior of the plate, where they affect lithosphere of both oceanic and continental character. This is particularly the case for eastern Australia

Figure 4. Calculated stress field in the Farallon plate under the reconstructed conditions of 30 Ma ago, prior to the break-up of the Farallon plate. The plotted principal horizontal non-lithostatic stresses are calculated for a 100 km thick elastic plate. The shaded area indicates the tensional zone which developed into the Cocos-Nazca spreading centre (Galapagos Rift). After Wortel and Cloetingh, 1981.

(see Figure 6), an area characterized by recent volcanic activity, probably associated with a regional tensional stress regime (Duncan and McDougall, 1987). Furthermore, the measured continental flux of crustal Helium in the Great Artesian Basin of Queensland is consistent with fracturing of the crust under the influence of tensional stresses (Torgersen and Clarke, 1985).

The extent of continental thinning is a critical factor for its relative strength with respect to oceanic lithosphere (Steckler and Ten Brink, 1986, see Figure 7). In areas that have undergone earlier phases of lithospheric thinning such as rift zones and the hinge zones of passive continental margins, severely thinned continental lithosphere is probably substantially weaker than oceanic lithosphere and preferential rifting of continental

Figure 5. The force required to induce tensional failure in oceanic lithosphere with a depth-dependent rheology (Goetze and Evans, 1979) plotted as a function of lithospheric age (solid line). Boxes indicate stress levels prior to the break-up of the Farallon plate into the Nazca plates, calculated by Wortel and Cloetingh (1981). Circles indicate (intraplate) stress levels in the present-day Nazca plate (Wortel and Cloetingh, 1983). The values given are representative of large parts of the Farallon plate and Nazca plate, respectively.

lithosphere might occur (Vink et al., 1984; Steckler and Ten Brink, 1986). Thermo-mechanical models show that in this case stresses of the order of a few kilobars are required (Houseman and England, 1986; Cloetingh and Nieuwland, 1984). Fragmentation of plates under the influence of slab-pull forces is therefore seen to provide a mechanism for passive rifting of both oceanic and continental lithosphere. Due to its great lithospheric strength, rifting of old oceanic lithosphere is not expected to occur on a large scale. This is in agreement with the observation that, with two exceptions (Stein and Cochran, 1985; Mammerricks and Sandwell, 1986), evidence for rifting of old oceanic lithosphere is absent. Whether continents in the interiors of plates break up under the influence of plate-tectonic forces is dependent on their specific thermo-mechanical structure, the position of the continental fragments in the plate relative to the surrounding trench systems, and the variation of the forces acting on each downgoing slab. While we do not wish to fully discuss here the stress regime in the overriding plate (see e.g. England and Wortel, 1980), we conjecture that variations in age-dependent slab-pull forces and associated stress fields of the type discussed here will prove to be of considerable importance in analysis of rifting

Figure 6. Regional stress field in the Australian continent and peripheral areas. a (top) : calculated stress field, after Cloetingh and Wortel (1986). b (bottom) : Stress orientation data from focal mechanism studies (Lambeck et al., 1984). Hatched areas mark the regions with Late Pliocene to present basaltic volcanism in northern Queensland and Victoria (Duncan and McDougall, 1987). The site of excessive He degassing (Torgersen and Clarke, 1985) is marked by a star.

processes in the overriding plate.

Figure 7. The force required to induce lithospheric failure in oceanic and continental lithosphere as a function of age. The curves are calculated adopting depth-dependent rheologies. Curves for continental lithosphere are calculated for two different thermal thicknesses A of continental lithosphere (after Steckler and Ten Brink, 1986).

4. Initiation of subduction zones

The mechanisms that initiate the formation of new subduction zones are not fully understood (Dickinson and Seely, 1979; Flinn, 1982; Turcotte and Schubert, 1982; Kobayashi and Sacks, 1985). This is in part because there are no obvious present-day examples of new trench formation. Dickinson and Seely (1979) summarize the possible mechanisms for initiation of oceanic lithosphere subduction and distinguish two classes: (1) plate rupture within an oceanic plate or at a passive margin and (2) reversal of the polarity of an existing subduction zone, possibly following a collision of an island arc with a passive margin (see also Mitchell, 1984). A third mechanism is initiation of subduction by inversion of transform faults into trenches (Uyeda and Ben-Avraham, 1972).

We deal here with mechanisms for the formation of new convergent plate boundaries rather than those for polarity reversal. Geological evidence for margins with widely different ages (e.g. Dewey, 1969; Cohen, 1982) support the thesis that passive continental margins in particular might be potential sites for the formation of convergent plate boundaries. Passive margins may, therefore, be expected to play a central role in the Wilson cycle of the opening and closing of oceans (Wilson, 1966).

Several authors (Turcotte and Schubert, 1982; Hynes, 1982) advocate a scenario based on the development of a major oceanic basin with old, and hence cold and gravitationally unstable (Oxburgh and Parmentier, 1977; Vlaar and Wortel, 1976), oceanic lithosphere at its continental boundaries before the basin can be closed. This seems to be more based on a comparison with the present-day size of the Atlantic Ocean rather than by an evaluation of geological observations. The same may be said of arguments for spontaneous foundering and subduction of oceanic lithosphere older than 200 Ma (Hynes, 1982), based on the current absence of oceanic lithosphere older than 200 Ma and the increase with age of the gravitational instability of oceanic lithosphere. The transformation of passive margins into active ones by spontaneous foundering of old unstable lithosphere is inhibited by the great strength of oceanic lithosphere (Kirby, 1980; McAdoo et al., 1985). We focus, therefore, on the state of stress at passive margins and investigate whether the stresses generated there are sufficiently high to induce lithospheric failure and subsequent transformation into an active margin.

4.1 Models for passive to active margin transition

Our model studies (Cloetingh, 1982; Cloetingh et al., 1983) have shown that, in general, flexure induced by sediment loading dominates the stress state at passive margins. In that work we have demonstrated that the continuing accumulation of sediments at passive margins leads to an increase in the induced stress level with the age of the margin. Finite element stress calculations were made for a passive margin in different stages of evolution (Figure 8) for two different sediment loading models using a depth-dependent rheology inferred from extrapolation of rock-mechanics data (after Goetze and Evans, 1979).

As a *reference* we adopted a sediment loading model in which the maximum thickness of the sedimentary wedge at passive margins followed a square-root-of-age relation (Turcotte and Ahern, 1977; Wortel, 1980). This model represents a fair average of the sediment loading histories and resulting thicknesses observed at passive margins (Southam and Hay, 1981). In the second model, the *full load* model, the entire loading capacity of oceanic lithosphere is taken up by the sediments. Sediment loading extends from the continental shelf to the continental rise, with maximum sediment thickness in the outer shelf/continental slope region. Lithospheric strength profiles combining the effect of pressure on brittle behaviour (Byerlee, 1968, 1978) and temperature- and stress dependence on ductile deformation 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) were adopted (Cloetingh et al., 1982, 1984). Investigation of gravity anomalies at passive margins (Karner and Watts, 1982) has shown 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. It should be noted that the models discussed here deal only with the gross features of passive margin evolution which are pertinent to the problem under consideration.

Figure 9 shows the stress field for the case of a reference load on a 100 Ma old passive margin. Differential stresses ($\sigma_H - \sigma_V$) are greatest at the transition from oceanic lithosphere to rift-stage lithosphere. In Figure 10 the stress maximum for a 100 Ma old passive margin is displayed as a function of depth, down to the base of the MSL. The

Figure 8. Central part of the finite element mesh employed for analysis of stress state on a passive margin in the interior of a plate not involved in collision or subduction processes. Layers of elements are incrementally added to the lithosphere, whose lower boundary is indicated at various time steps (specified in Ma). A comparison of the lower boundaries at 12.5 Ma and 20 Ma (given by dashed lines) illustrates the growth of the lithosphere in horizontal and vertical directions. The height and the width of the sedimentary wedge, corresponding to sedimentary loading according to the reference model on outer shelf, slope and rise, are given in kilometres at some selected time steps (specified in Ma).

lowermost and uppermost regions of the MSL fail due to the high stresses developed here. However, the main part of the MSL remains in the elastic state as stresses are too low to result in rupture of the lithosphere. The ratio of the maximum stress generated to the maximum strength is essentially independent of lithospheric age since load, strength and thickness of MSL all exhibit the same square-root-of-age behaviour due to the thermal evolution of the lithosphere.

The situation is very much different when the full loading capacity is taken up by the sediments. The surplus of sediment added to the reference load promote high stresses most effectively when deposited on a young (weak) passive margin. This is demonstrated in Figure 11 where we have plotted the ratio of A_y , corresponding to the area of the strength envelope in failure, and the total area A , corresponding to the integrated strength of the MSL. Figure 11 shows that full loading on passive margins with ages below 20 Ma leads to complete failure of the lithosphere. For ages in excess of 20 Ma the relative amount of failure decreases rapidly with age. Therefore, if subduction has not commenced at a passive margin during a youthful stage, continued aging alone will not result in conditions more favourable for the initiation of subduction. This explains the apparent paradox of why gravitationally unstable oceanic lithosphere at the margin of the Atlantic is not subject to subduction (see Kobayashi and Sacks, 1985).

Figure 9. Stresses calculated for a passive margin at an age of 100 Ma based on the reference model of sediment loading. 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 there is significant deformation.

These findings lead us to propose the modification of the classical sequence of the Wilson cycle concept shown in Figure 12, in which closure of a newly rifted basin occurs by initiation of subduction of young lithosphere, with implications for tectonics and volcanism which differ appreciably from those associated with deep subduction zones (Vlaar, 1983; Cloetingh et al., 1984; Vlaar and Cloetingh, 1984). In particular, it explains the absence of island arc volcanism in Wilsonian orogenies (e.g. Trumpy, 1982) and the emplacement of ophiolites, fragments of young gravitationally stable oceanic lithosphere, on the adjacent continent during the transformation of a passive margin, features that do not conform to the standard concepts of subduction of oceanic lithosphere and the long timespan of the classical form of the Wilson cycle. An interesting analogy for the destruction of passive margins of small oceanic basins exists in the form of marginal basins, whose evolution is characterized by a short timespan (less than 20 Ma, Taylor and Karner, 1983) between their creation and collapse. As pointed out by these authors, the tectonic setting of several back-arc basins suggests that they represent more than just a passive response to kinematic boundary conditions. Taylor and Karner (1983) therefore argue that neither the global nor local models adequately explain the conditions necessary for back-arc basin formation. They conclude that better understanding of the processes associated with the initiation of subduction and rifting is required.

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-day Atlantic. Investigations of Alpine orogeny (Trumpy, 1982) have provided strong evidence for closure of small oceanic basins at an early stage after opening, as have studies of the evolution of the Appenines (Winterer and Bosellini, 1981). In fact, the Alpine basins, being characterized by young oceanic lithosphere, transcurrent faulting, extensive sediment loading, and a compressive tectonic regime, were ideally suited for the transformation of passive into active margins. Frequently, however, only indirect and

Figure 10. Comparison of stresses generated at 100 Ma old passive margin, under reference loading, with lithospheric strength. Strength envelope and results of stress calculation (solid dots) as a function of depth down to the base of MSL at the point of maximum flexure (see also figure 9). The stress distribution is given by the line inside the strength envelope connecting the solid dots. Differential stresses ($\sigma_H - \sigma_V$) are plotted against depth. Sign convention for the stresses: tension positive, compression negative. Zero-strength is assumed for the sediments. Differences between brittle strengths in compression and tension are ignored. Hatched areas in the upper and lower part of the MSL denote failure by brittle fracture and ductile flow, respectively.

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; Spray, 1984) favour obduction of

Figure 11. Ratio of A_y (the hatched area within the strength envelope, see figure 10) to A (total area of the envelope) as a function of lithospheric age for the full load model. For ages below 20 Ma lithospheric failure and subsequently initiation of subduction is induced.

thermally young immature oceanic lithosphere to account for the short time gap documented between ophiolite formation and ophiolite emplacement.

Similarly, near-horizontal subduction during closure of a small ocean basin provides an important element in the dynamical evolution of foreland thrust belts (Stockmal et al., 1986). Our findings also shed light on the far greater degree of basement fragmentation observed under foreland basins in collisional settings compared to their Andean analogs (Allen et al., 1986). Evolutionary frameworks of the Wilson cycle in terms of opening and closing of wide oceans often seem 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 reached 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 analogy might, therefore, be very misleading. Moreover, estimated widths of oceans inferred from palinspastic restorations are often routinely increased by an additional 1000 km meant to represent the lengths of slabs consumed in modern circum-Pacific subduction zones, associated with subduction of old oceanic lithosphere (e.g. Williams,

WILSON CYCLE

Figure 12. Scenarios for the Wilson Cycle. Left: classical scenario, in which rifting is followed by the formation of a large ocean basin. Initiation of subduction involves old oceanic lithosphere, which results in a deep subduction zone. Right: preferred scenario for Wilsonian orogenies, in which closure of the newly rifted basin occurs through initiation of subduction of young oceanic lithosphere leading to either shallow angle subduction or to obduction of gravitationally stable buoyant lithosphere. Implications for tectonics and volcanism strongly differ from those associated with deep subduction zone. Patterns indicate passive-margin sediments and continental, rift-stage, and oceanic lithosphere.

1980). Such reconstructions exclude a priori the possibility of the closing of a young oceanic basin and might even lead one to overlook possible interesting consequences of the subduction of young lithosphere. Therefore, we suggest that a more critical appraisal is made of the role large oceans play in the Wilson cycle concept.

4.2 The role of external forces

In the preceding section the flexural stresses induced by sediment loading were shown to be of the order of several kbars. As noted previously, the regional stress field caused by plate tectonic forces is dominated by concentration of slab-pull forces. For passive margins located in the interiors of plates not involved in subduction or collision processes, the local stress field induced by sediment loading dominates the stresses induced by the remaining plate-tectonic forces, which in general is of the order of a few hundred bars (e.g. Richardson et al., 1979). We have shown that under such conditions evolution of passive margins to maturity will not in itself lead to initiation of subduction.

Figure 13. Regional stress field in the northeastern Indian Ocean. a (left). Calculated stress field after Cloetingh and Wortel (1986). The dashed line is the southern limit of the observed deformation in the northeastern Indian Ocean (Geller et al., 1983). Plotted are principal horizontal non-hydrostatic stresses averaged over a uniform elastic plate with a reference thickness of 100 km. b (right). The orientation of maximum horizontal compressive stress inferred from a focal mechanism study by Bergman and Solomon (1985).

Thus, in general, external forces in addition to sediment loading are required to cause rupture of the lithosphere. Figure 5 has given a comparison of the force required to induce complete lithospheric tensional failure and calculated intraplate tensional stress levels in the eastern Pacific. An example of a different tectonic setting is found in the northeastern Indian Ocean, where a high level of compressional stress (Figure 13) is induced by the focusing of resistive forces associated with the Himalayan collision zone and the subduction of young oceanic lithosphere in the northern part of the Sunda arc (Cloetingh and Wortel, 1985; 1986). The response of the oceanic lithosphere in the northeastern Indian Ocean to the compressional stresses is, however, not one of the lithosphere failure, but one of buckling (Weissel et al., 1980; McAdoo and Sandwell, 1985; Zuber, 1987). Figure 14 shows that the estimates of the forces required to buckle the ocean floor in the

northeastern Indian Ocean given by McAdoo and Sandwell (1985) are in excellent agreement with the calculated stress levels shown in Figure 13. As noted by Cloetingh et al. (1983) the folding of the oceanic lithosphere caused by regional compressional stresses amplified by sediment loading might be a preparatory stage for the initiation of a new subduction zone in the northeastern Indian Ocean. More recently, strong independent evidence for the presence of a nascent diffuse plate boundary in the area has been presented by Wiens et al. (1985, 1986) and Wiens (1986) based on an inversion of current plate velocities and a study of the seismo-tectonics of the area.

Figure 14. Buckling load versus age for oceanic lithosphere with a depth-dependent rheology inferred from rock-mechanics studies (Goetze and Evans, 1979) (solid curve) and for fully elastic oceanic lithosphere (dashed curve) (after McAdoo and Sandwell, 1985). Boxes indicate the stress levels calculated for the area in the northeastern Indian Ocean (Cloetingh and Wortel, 1986, see figure 13) where folding of oceanic lithosphere under the influence of compressional stresses has been observed (McAdoo and Sandwell, 1985; Geller et al., 1983).

5. Discussion

We have shown that concentration of slab-pull forces provides an effective mechanism for fragmentation of plates and the formation of new spreading centres. We have also demonstrated that evolution of passive margins to maturity will not in itself lead to

initiation of subduction and shown that, in general, external forces in addition to sediment loading are required.

The action of external forces will be most effective when young passive margins are pre-stressed by thick sedimentary wedges. McKenzie (1977) demonstrated that a large slab pull is required to overcome the resistive forces active in the process of trench formation. The compressive stresses necessary to sustain the further development of subduction zones that involve young stable lithosphere (McKenzie, 1977; England and Wortel, 1980), may be provided by stress concentration of plate tectonic forces during plate reorganizations (Cloetingh and Wortel, 1985) and are an order of magnitude smaller than the stresses required to rupture the lithosphere.

Existing weakness zones located within plates might be more suitable sites for initiation of subduction than passive margins. As such, spreading ridges (Turcotte et al., 1977) and transform faults (Uyeda and Ben-Avraham, 1972) have been suggested. A spreading ridge may be thought of as the extreme case (0 Ma) of a young passive margin, and represents a rupture zone in the lithosphere. The former view is consistent with the results of a survey of recently initiated subduction zones in the Pacific (Karig, 1982), which shows that subduction zones initiated during the Neogene are frequently on the sites of transform faults. In other cases there has been rejuvenation of pre-existing subduction zones or geometrical adjustment of plate boundaries (Okal et al., 1986; Kroenke and Walker, 1986). It is widely held that the present Pacific coastline of North and South America has been the site of semi-continuous eastward subduction since the Late Proterozoic (Windley, 1977). The different plate configurations and thermal regimes at that time may have provided conditions more suitable for initiation of subduction.

Thus, in general, we do not expect initiation of subduction of oceanic lithosphere at passive margins to play a leading role in plate reorganizations such as documented by Rona and Richardson (1978). In an oceanic plate attached to a subduction zone the pull acting on the subducting slab can be concentrated to a sufficient stress level to induce the formation of new spreading centres (Wortel and Cloetingh, 1981; 1983). Therefore, we conjecture that plate reorganizations occur primarily through the formation of new spreading ridges, since stress relaxation in the lithosphere occurs much more easily via this process than through the formation of new subduction zones. During such a process new subduction zones might be subsequently created at the sites of already present transform faults, when the new spreading direction has a component perpendicular to the direction of the transform fault.

Better understanding of the mechanics of formation of new plate boundaries in general, and the processes underlying the transition of passive margins into active margins in particular, will enhance our insight into the evolution of sedimentary basins. We have shown (Cloetingh et al., 1985; Cloetingh, 1986) that the associated reorganizations of lithospheric stress field are recorded in the stratigraphic record of sedimentary basins. Specific short-term fluctuations (on time scales of a few Ma and longer) in apparent sea levels inferred from passive margins and intracratonic basins can now be associated quantitatively with particular plate-tectonic reorganizations. Alternatively, the seismostratigraphic record might provide a new source of information on paleo-stress fields (Cloetingh, 1986). Furthermore, the opening and closure of oceanic basins, with the associated changes in the area/age distribution of the ocean floor, has been shown (Heller

and Angevine, 1985) to be the main controlling feature on the occurrence of long-term (time scales of tens of Ma) sea level cycles. Further work on modelling of paleo-stress field alongs the lines set out by Wortel and Cloetingh (1981) is required to more fully document differences in the roles of initiation of rifting and the initiation of subduction in the tectonic evolution of the plates' interiors.

6. Conclusions

Lateral variations in the age of the downgoing lithosphere at subduction zones are an effective mechanism for fragmentation of oceanic and continental plates and formation of new spreading centres. Aging of passive margins will not in itself lead to a spontaneous initiation of subduction. In general, the formation of new subduction zones at passive margins requires a focusing of external plate tectonic forces. The action of these external forces will be most effective when young passive margins are pre-stressed by thick sedimentary wedges. Plate reorganizations probably take place primarily through the formation of new spreading ridges, because stress relaxation in the lithosphere occurs much more effectively via this process than through the formation of new subduction zones.

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