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On the Initiation of Subduction Zones

SIERD CLOETINGH,^{1,2} RINUS WORTEL¹ and N. J. VLAAR¹

Abstract—Analysis of the relation between intraplate stress fields and lithospheric rheology leads to greater insight into the role that initiation of subduction plays in the tectonic evolution of the lithosphere. Numerical model studies show that if after a short evolution of a passive margin (time span a few tens of million years) subduction has not yet started, continued aging of the passive margin alone does not result in conditions more favorable for transformation into an active margin.

Although much geological evidence is available in supporting the key role small ocean basins play in orogeny and ophiolite emplacement, evolutionary frameworks of the Wilson cycle usually are cast in terms of opening and closing of wide ocean basins. We propose a more limited role for large oceans in the Wilson cycle concept. In general, initiation of subduction at passive margins requires the action of external plate-tectonic forces, which will be most effective for young passive margins prestressed by thick sedimentary loads. It is not clear how major subduction zones (such as those presently ringing the Pacific Basin) form, but it is unlikely they form merely by aging of oceanic lithosphere. Conditions likely to exist in very young oceanic regions are quite favorable for the development of subduction zones, which might explain the lack of preservation of back-arc basins and marginal seas.

Plate reorganizations probably 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.

Key words: Initiation of subduction, passive to active margin transition, preservation of back-arc basins, emplacement of ophiolites, mechanics of plate boundary formation, Wilson cycle, plate reorganizations.

Introduction

Initiation of oceanic lithosphere subduction is a key element in plate-tectonic schemes for the evolution of the lithosphere. The mechanisms that initiate the formation of new subduction zones are, however, not fully understood (DICKINSON and SEELY, 1979; FLINN, 1982; TURCOTTE and SCHUBERT, 1982; KOBAYASHI and SACKS, 1985). This is in part because, with a possible exception in the western Pacific (OKAL *et al.*, 1986; KROENKE and WALKER, 1986), there are no obvious

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present-day examples of the formation of new major trench systems. 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 consuming plate boundaries rather than those for polarity reversal. Geological evidence for margins with widely different ages (e.g., DEWEY, 1969; COHEN, 1982) support the notion that passive continental margins in particular might be potential sites for the formation of consuming 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).

The Role of Passive Margins in the Wilson Cycle

Several authors (TURCOTTE and SCHUBERT, 1982; HYNES, 1982) advocate a model based on the development of a major oceanic basin with old and, hence, cold and gravitationally unstable (VLAAR and WORTEL, 1976; OXBURGH and PARMENTIER, 1977) oceanic lithosphere at its continental margins before the basin can be closed. This seems to be more inspired by an actualistic comparison with the present size of the Atlantic Ocean rather than by a consideration of more pertinent geological observations. The same may be said of arguments for spontaneous foundering and subduction of oceanic lithosphere older than 200 Ma (HYNES, 1982), made on the basis of the temporary absence of oceanic lithosphere older than 200 Ma and the increase with age of the gravitational instability of oceanic lithosphere. The transformation of passive into active margins by spontaneous foundering of old unstable lithosphere is inhibited by the great strength of oceanic lithosphere (KIRBY, 1980; MCADOO *et al.*, 1985). We, therefore, focused on the state of stress at passive margins to investigate whether the stresses generated there are sufficiently high to induce lithospheric failure and transformation into an active margin.

Models for Passive to Active Margin Transition

Our numerical 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, owing to the continuing accumulation of sediments at passive margins, the induced stress level increases with the age of the margin. An important rheological feature (BODINE *et*

al., 1981; CLOETINGH *et al.*, 1982) implemented in our thermo-mechanical models for the evolution of passive margins (CLOETINGH *et al.*, 1982; 1984) is that the strength of the lithosphere at the margin also increases with age. Finite-element stress calculations were made for a passive margin in different stages of evolution incorporating two different sediment loading models and the depth-dependent rheology inferred from extrapolation of rock-mechanics data (after GOETZE and EVANS, 1979).

As a *reference* model 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), reaching a maximum of 9.4 km at 200 Ma. This model constitutes 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. This model is more appropriate for the huge sediment accumulations found at deltas, which clearly exceed the thicknesses depicted by the reference model. In this model the maximum thickness of the sedimentary wedges reaches 16 km at 200 Ma. For both the reference and full load models, sediment loading extends from the continental shelf to the continental rise, with maximum sediment thickness in the outer shelf/continental slope region (see Figure 1). Lithospheric strength profiles were adopted that combine the effect of pressure on brittle behavior (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). Such a rheology provides a good first-order description of the mechanical behavior of oceanic lithosphere, and is consistent with seismotectonic studies (WIENS and STEIN, 1983) and observations of the response of oceanic lithosphere to tectonic processes (MCADOO *et al.*, 1985). Similar to BODINE *et al.* (1981) we defined the depth at which the strength is 500 bar as the lower boundary of the *mechanically strong upper part of the lithosphere (MSL)*. Both its thickness and its maximum strength increase according to a square-root-of-age function from a few kilometers and a few kilobars, respectively, for young lithosphere to values of approximately 50 km and 10 kilobar for old oceanic lithosphere. 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 gross features of passive margin evolution, pertinent to the problem under consideration.

Figure 2 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 3 the stress maximum for a 100 Ma old passive margin is displayed as a function of depth, down to the base of

Figure 1

Model features geometry, rheology and system of forces acting on a passive margin in the interior of a plate not involved in collision or subduction processes. The bottom of the mechanically strong part of the lithosphere (MSL) is indicated by a broken horizontal line. F_{RP} is the push exerted by the oceanic ridge, F_B is the negative buoyancy associated with the cooling of the oceanic lithosphere when it moves away from the spreading center. The position of the sedimentary wedge corresponds to sedimentary loading on outer shelf, slope and rise.

Figure 2

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. Symbols ($\leftrightarrow$) and ($---$) denote tension and compression, respectively.

Figure 3

Comparison of stresses generated at 100 Ma old passive margin, under reference loading with lithospheric strength. Strength curve envelope and results of stress calculation (solid dots) are plotted as a function of depth down to the base of MSL at the point of maximum flexure (see also Figure 2). 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. Hatched areas in the upper and lower part of the MSL denote failure by brittle fracture and ductile flow, respectively.

the MSL. The lowermost and uppermost regions of the MSL fail due to the high differential 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. To induce complete lithospheric failure, the integrated force $F = \int \sigma dz$ must exceed the integrated depth-dependent strength $A = \int \sigma_y dz$, where σ is the stress distribution, σ_y the differential strength profile and z the depth. 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 behavior due to the thermal evolution of the lithosphere.

The situation is very much different when the full loading capacity is taken up

Figure 4

Ratio of A_y (the hatched area within the strength envelope, see Figure 3) to A (total area of the envelope) as a function of lithospheric age for the full load model where the entire loading capacity of the lithosphere is taken up by passive margin sedimentation. For ages below 20 Ma lithospheric failure and consequently initiation of subduction is induced.

by the sediments. The surplus of sediment added to the reference load promotes high stresses most effectively when deposited on a young (weak) passive margin. This is demonstrated in Figure 4 where we have plotted the ratio of A_y , corresponding to the area of the strength envelope in failure, to the total area A , corresponding to the integrated strength of the MSL. Figure 4 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 favorable for the initiation of subduction. Similarly, stress levels required to be produced by external forces in order to create a new consuming plate boundary in this manner will increase with the growth of the oceanic basin as well. These findings unravel the enigma that gravitationally unstable oceanic lithosphere at the margin of the Atlantic is not subject to subduction (see KOBAYASHI and SACKS, 1985).

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 kilobars. In previous work (WORTEL and

CLOETINGH, 1981, 1983; CLOETINGH and WORTEL, 1985, 1986) we have demonstrated that the *regional* stress field caused by plate-tectonic forces (FORSYTH and UYEDA, 1975; CHAPPLE and TULLIS, 1977) is dominated by concentration of slab-pull forces. For passive continental 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, the evolution of passive margins to maturity will not in itself lead to an initiation of subduction.

In general, external forces in addition to sediment loading are required to cause rupture of the lithosphere. A comparison could be made of the force required to induce complete lithospheric tensional failure and the estimates of age-dependent slab-pull and ridge-push forces (ENGLAND and WORTEL, 1980; DAVIES, 1983). However, as noted by CHAPPLE and TULLIS (1977), such comparisons do not take into account that the level of the regional stress field can be considerably magnified due to the geometrical shape of the plate boundaries and the complexity of boundary forces along lithospheric plates. More specifically, we have demonstrated the key importance of incorporating plate-tectonic forces due to variations in the age of the downgoing lithosphere at convergent boundaries in lithospheric stress modelling (WORTEL and CLOETINGH, 1981, 1985; CLOETINGH and WORTEL, 1985). These findings have important implications for the formation of new plate boundaries. In fact, we have demonstrated in our model study of the paleo-stress field of the Farallon plate (WORTEL and CLOETINGH, 1981) that the level of the tensional stress field caused by concentration of plate-tectonic forces was of sufficiently high magnitude to cause the inception of the Galapagos spreading center and fragmentation of the Farallon plate into the Cocos plate and the Nazca plate at 25–30 Ma ago (see Figure 5). 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, for example, the case for the zone of concentrated extensional seismicity in young oceanic lithosphere in the equatorial Indian Ocean (STEIN *et al.*, 1987) and for eastern Australia, an area characterized by recent volcanic activity, probably associated with a regional tensional stress regime (DUNCAN and MCDUGALL, 1988). Preferential rifting of continental lithosphere might occur due to its relative weakness compared to oceanic lithosphere (VINK *et al.*, 1984; STECKLER and TEN BRINK, 1986; SHUDOF-SKY *et al.*, 1987). 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 rifting of both oceanic and continental lithosphere. Due to its great strength, rifting of old oceanic

Figure 5

The force required to induce tensional failure in oceanic lithosphere plotted as a function of lithospheric age (solid line). Boxes indicate stress levels prior to the break-up of the Farallon plate into the Cocos and 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 is not expected to occur on a large scale. This is in agreement with the observation that, with two exceptions (STEIN and COCHRAN, 1985; MAMMERICKS and SANDWELL, 1986) evidence for rifting of old oceanic lithosphere is absent.

A different situation arises in a tectonic setting dominated by compressional stresses, where even higher stress levels are required to exceed lithospheric strength. An example of this is found in the northeastern Indian Ocean, where an exceptionally high level of compressional stresses (Figure 6) is induced by the focusing of resistive forces associated with the Himalayan collision and 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 high level of in-plane compressional stress approaching lithospheric strength is, however, not one of lithosphere failure, but one of buckling (WEISSEL *et al.*, 1980; MCADOO and SANDWELL, 1985; ZUBER, 1987). Figure 7 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 6. 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

Figure 6

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). Dashed area indicates the location of the proposed (WIENS *et al.*, 1985) diffuse plate boundary separating an Indian plate from an Australian plate. Plotted are principal horizontal non-hydrostatic stresses averaged over a uniform elastic plate with a reference thickness of 100 km. Note the high level of the stress field associated with the unique dynamic situation of the Indo-Australian plate. b (right) The orientation of maximum horizontal compressive stress inferred from an earthquake focal mechanism study by BERGMAN and SOLOMON (1985).

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 separating an Indian plate from an Australian plate has been presented by WIENS *et al.* (1985, 1986) based on an inversion of plate velocities and a study of the seismo-tectonics of the area. Hence, it seems that in the Indo-Australian plate stress levels induced by plate-tectonic forces are of sufficient magnitude to induce the inception of a new consuming plate boundary, even in old oceanic lithosphere. It is important to note that this process is the expression of the unique present-day dynamic situation of the Indo-Australian plate, which reflects for a major part the dramatic impact of the Eurasia-India collision, which is probably one of the most important tectonic events during the last 250 Ma of Earth history. For these

Figure 7

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). Box indicates the stress levels calculated for the area in the northeastern Indian Ocean (CLOETINGH and WORTEL, 1985, 1986, see Figure 6) where folding of oceanic lithosphere under the influence of compressional stresses has been observed (MCADOO and SANDWELL, 1985; GELLER *et al.*, 1983).

reasons, the mechanism underlying the formation of a nascent plate boundary in the central Indian Ocean is likely the exception rather than the rule for the creation of new subduction zones.

Closure of Small Oceanic Basins

Therefore, initiation of subduction of old oceanic lithosphere requires extremely high levels of intraplate stress to be induced by external forces, probably in association with unique plate-tectonic settings such as presently encountered in the northeastern Indian Ocean. As demonstrated by our modelling of stresses at passive margins, the level of stresses to be produced by plate-tectonic forces drops dramatically with decreasing age of the lithosphere, favoring initiation of young oceanic lithosphere subduction at the margins of small oceanic basins. These findings lead us to propose the modification of the classical sequence of the Wilson cycle concept shown in Figure 8, in which closure of newly rifted basins by initiation of subduction of young lithosphere occurs. This has 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). For example, a notable feature of shallow-angle subduction of young oceanic lithosphere is the induction of a compressional regime in the overriding plate, contrasting with the

Figure 8

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 Wilson cycle, 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 (VLAAR, 1983). 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.

general association of high-angle subduction of old oceanic lithosphere with an extensional regime in the overriding plate (ENGLAND and WORTEL, 1980). Closure of small oceanic basins explains the frequently observed 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 time span 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 time span between their creation and collapse (less than 20 Ma, TAYLOR and KARNER, 1983). The modelling described in the present paper has demonstrated that the conditions likely to exist in these very young oceanic regions are quite favorable for the development of new subduction zones. These findings might have important consequences for evaluating the movement and accretion of terranes and other continental crust fragments.

As noted by CHURCH and STEVENS (1971), much of the geological evidence in collision orogens points to closing of small ocean basins rather than large oceans of the scale of the present-day Atlantic. Usually, precise data on the timing of the closing events are absent, which inhibits a quantitative comparison with our model results. Nevertheless, much of the geological evidence is clearly contradictory to activation of passive margins in a late stage of their evolution. As observed by MCWILLIAMS (1981), not all (Proterozoic) Pan-African and older mobile belts mark the sites of major ocean closure. Rather they are formed without the destruction of vast amounts of oceanic lithosphere. On the base of a survey of the tectonic framework of central and western Europe, ZWART and DORNSIEPEN (1978) pointed out that the Variscan orogens are also due to collision and closing of short-lived oceans of minor size. Investigations of Alpine orogeny (TRUMPY, 1982) have provided strong evidence for closure of small oceanic basins in 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 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). The latter situation is to a large extent caused by the orientation of the spreading, which has hampered reliable paleomagnetic estimates for the width of the Iapetus Ocean. Paleomagnetic evidence, however, certainly does not rule out that large Paleozoic oceans were consumed (e.g., MCELHINNY and VALENCIO, 1981). In fact, the best evidence for the Iapetus Ocean has come from the occurrence of ophiolite belts (ZWART and DORNSIEPEN, 1978). Several authors, (e.g., DEWEY, 1976; NICOLAS and LEPICHON, 1980; SPRAY, 1984) favour obduction of thermally young immature oceanic lithosphere to account for the short time gap documented between ophiolite formation and ophiolite emplacement.

Similarly, flat-angle subduction (VLAAR, 1983) 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 seem sometimes to be more inspired by an actualistic comparison with the present size of the Atlantic Ocean, than by a consideration of more pertinent geological observations. During its evolution, the Atlantic Ocean passed through a transition from a narrow to a wide ocean basin, without the formation of a system of subduction zones at its margins. Apparently, optimal loading conditions for transformation of passive into active margins were not fulfilled at this stage, while further aging has not made the

passive margins more susceptible to initiation of subduction. To rely too heavily on such an actualistic analogon might, therefore, be very misleading. Moreover, estimated widths of oceans inferred from palinspastic restorations are frequently automatically increased with an additional 1000 km inferred from the lengths of slabs consumed in modern circum-Pacific subduction zones, associated with subduction of old oceanic lithosphere (e.g., WILLIAMS, 1980). Such reconstructions exclude *a priori* the possibility of the closing of a young oceanic basin and might even lead to overlooking the eventual interesting consequences of the subduction of young lithosphere. Our model studies have shown that if subduction has not started after a short evolution of a passive margin, continued aging of the passive margin alone does not result in conditions more favorable for transformation into an active margin. That requires a concentration of external forces in a unique dynamic situation such as the one in which the present-day Indo-Australian plate is involved. Therefore, we propose a more critical appraisal of the role large oceans play in the Wilson cycle.

Discussion

We have demonstrated that evolution of passive margins to maturity (200 Ma) will not in itself lead to initiation of subduction. In general, external forces in addition to sediment loading are required to cause rupture of the lithosphere.

The action of external forces will be most effective when young passive margins are prestressed by thick sedimentary wedges. The compressive stresses necessary to sustain the further development of subduction zones involving 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. Spreading ridges (TURCOTTE *et al.*, 1977) and transform faults (UYEDA and BEN-AVRAHAM, 1972) have been suggested. A spreading ridge should be thought of as the lower bound (0 Ma) of a young passive margin, and forms a rupture zone inherent in the lithosphere. This view is consistent with the results of a survey of recently initiated subduction zones in the Pacific (KARIG, 1982) showing that zones initiated during the Neogene are frequently on the sites of transform faults or else are rejuvenated pre-existing subduction zones or geometrical adjustments (OKAL *et al.*, 1986; KROENKE and WALKER, 1986). However, it is not clear how the major subduction zones that presently ring the Pacific have been initiated. For example, it is widely held that the present Pacific coastline of Northern and Southern America has been the site of semi-continuous eastward subduction since the Late Proterozoic (WINDLEY, 1977).

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 oceanic lithosphere subduction at passive margins to play a leading role in the 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 centers (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 the former 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 the mechanics of the 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 in the evolution of sedimentary basins. We have shown (CLOETINGH *et al.*, 1985; CLOETINGH, 1986, 1988) that the associated reorganizations of lithospheric stress fields are recorded in the stratigraphic record of sedimentary basins. Specific short-term fluctuations (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. Conversely, the seismostratigraphic record of sedimentary basins might provide a new source of information on paleo-stress fields (CLOETINGH, 1986; LAMBECK *et al.*, 1987) and, hence, on the dynamics of plate reorganizations. Furthermore, the opening and closure of oceanic basins, with the associated changes in the area/age distribution of the ocean floor during the Wilson cycle, 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 fields along 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.

Conclusions

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 prestressed by thick sedimentary wedges. Conditions likely to exist in very young oceanic lithosphere are

quite optimal for the development of new subduction zones, which might explain the lack of preservation of back-arc basins and marginal seas. It is not clear how major subduction zones, such as those presently ringing the Pacific, form. Probably, plate reorganizations primarily take place 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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