Geologie en Mijnbouw 65: 103–117 (1986)

Tectonics of passive margins: implications for the stratigraphic record

Sierd Cloetingh
Department of Theoretical Geophysics, Institute of Earth Sciences, University of Utrecht, P.O. Box 80.021, 3508 TA Utrecht, The Netherlands

Received 5 February 1986; accepted in revised form 27 May 1986

Keywords: passive margins, seismostratigraphy, apparent sea level fluctuations, Vail curves, tectonics, flexure, intraplate stresses, sedimentary basins

Abstract

Thermo-mechanical modelling demonstrates that tectonically induced vertical motions of the lithosphere can explain the major part of the short-term fluctuations in apparent sea level deduced from the seismostratigraphic record at passive margins. The interaction of intraplate stresses and the deflection of the lithosphere caused by sedimentary loading can produce apparent sea level changes of up to 100 metres at the flanks of passive margins. This mechanism is most effective for young passive margins subject to rapid sediment loading. Stress variations in the lithosphere of a few hundred bar are sufficient to explain most of the lowerings in sea level shown in the Vail et al. (1977) curves. To induce apparent sea level fluctuations with magnitudes greater than 50 metres, changes in stress level of more than 1 kbar are required, which must be related to major reorganizations at convergent plate boundaries, fragmentation of plates, or collision processes. By its nature, the tectonic model can explain contemporaneous fluctuations in apparent sea level in neighbouring depositional environments. In principle, it implies the possibility of regional correlations in different basin settings. Specific short-term fluctuations in the Vail et al. curves can be associated quantitatively with particular plate-tectonic reorganizations of lithospheric stress fields. The seismostratigraphic record may provide a new source of information on paleo-stress fields to be correlated with results of independent numerical modelling of intraplate stresses.

Introduction

Passive continental margins form an important class of sedimentary basins (Watkins & Drake 1983) and play a central role in orogeny. Previous work (Cloetingh 1982; Cloetingh et al. 1982, 1983, 1984) concentrated on the function of passive margins in the Wilson cycle of opening and closing of oceanic basins. From detailed comparison of stresses at passive margins and the strength of the underlying lithosphere, it was found that the age of the passive margin is a key factor in its conversion into an active margin. Our model studies have shown that the flexure induced by sediment loading dominates the stress state at passive margins. We have demonstrated further that the stress level induced increases with the age of the margin owing to the continuing accumulation of sediments at passive margins. An important feature following from rheological considerations (Cloetingh et al, 1982, 1984), which has been implemented in our thermomechanical models for the evolution of passive margins is that the strength of the lithosphere at the margin increases with age as well. We found that stresses generated at passive margins of large ocean basins are generally insufficient to induce lithospheric failure and transformation into active margins. This offers an explanation for the enigma that gravitationally unstable oceanic lithosphere at the margins of the Atlantic is not subject to subduction. Extensive sediment loading on the mechanically weak passive margins of newly rifted basins, however, creates a high stress level with respect to lithospheric strength and was shown to be an effective mechanism for basin closure. Our modelling demonstrated that if after a short evolution (a few tens of million years) subduction has not yet started, aging of the passive margin alone does not result in conditions more favourable for transformation into an active margin. These findings led us to propose a modification of the classical sequence of the Wilson cycle concept (Fig. 1), in which closure of newly rifted basins occurs by initiation of subduction of young oceanic lithosphere, with implications for tectonics and volcanism which differ from those associated with deep subduction zones. This reflects the absence of island arc volcanism (e.g. Trümpy 1982; Winterer & Bosellini 1981) and the occurrence of ophiolites, obducted fragments of young and hence gravitationally stable (Vlaar & Wortel 1976) oceanic lithosphere emplaced within a few tens of million years after their creation at spreading ridges in Wilsonian orogenies (Cloetingh et al. 1984).

In more recent work (Cloetingh et al. 1985) we concentrated on the subsidence history of passive margins. Passive margins are the sites of extensive sediment accumulation. According to recent estimates (Southam & Hay 1981), approximately one third of all Phanerozoic sediments are located on passive continental margins formed at the break-up of Pangea. Since that event, the total length of passive margins has been increasing with time to the present length of approximately 70.000 km (Southam & Hay 1981). Quantitative models of the evolution of passive margins are of increasing importance in evaluating petroleum potential, in particular in the light of the extensive seismostratigraphic studies that have been carried out during the last decade (Watkins & Drake 1983). A major contribution to the seismostratigraphic analysis of passive margins was made by Vail and his coworkers at Exxon (Vail et al. 1977, 1984). These authors produced a set of curves for sea level fluc-

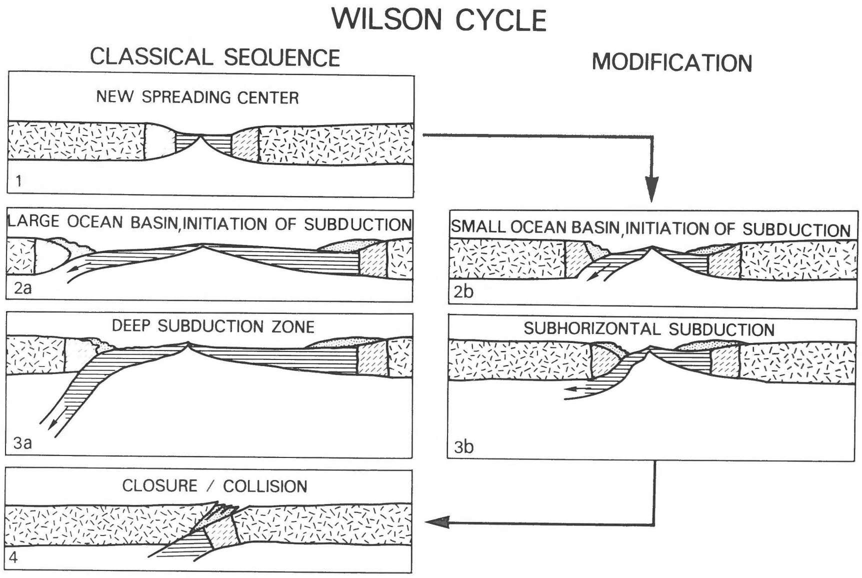

Fig. 1. Senarios for Wilson cycle. Left: Classical scenario, in which rifting is followed by formation of large oceanic basin. Initiation of subduction involves old oceanic lithosphere, producing deep subduction zone. Right: preferred scenario for Wilsonian orogenies, in which closure of newly rifted basin by initiation of subduction of young lithosphere occurs, with implications for tectonics and volcanism differing from those associated with deep subduction zones. Patterns indicate passive-margin sediments and continental, rift-stage, and oceanic lithosphere.

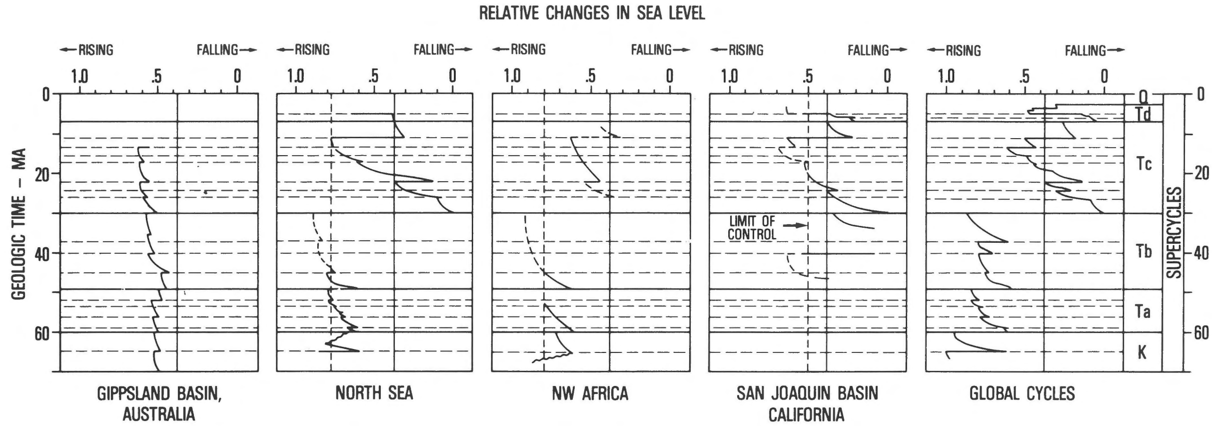

Fig. 2. Correlation of regional cycles of relative changes in sea level from sedimentary basins in four different continents. Although the global cycles have been constructed by averaging the regional cycles, they are heavily weighted in favour of the North Sea and North West Africa (After Vail et al., 1977).

tuations based on Exxon data from passive margins and intracratonic basins in different parts of the world.

Since the publication of Vail et al.'s (1977) global cycles of relative changes in sea level, there has been a continuous debate on the mechanism responsible for these fluctuations (e.g. Bally 1980, 1982). Recently Cloetingh et al. (1985) have proposed a new tectonic mechanism for sea level variations of about 1–10 cm/1000 years, with a magnitude of up to a few hundred metres. Their model explains these changes, provided that horizontal stresses of the order of a few kbars (1 kbar = 100 MPa) exist in the lithosphere and changes in these stress fields occur on geological time scales. In the present paper we will concentrate on the consequences of the associated passive margin tectonics for the stratigraphic record. We shall then investigate the evidence for changes in the intraplate stress field. Next we shall introduce a new approach to the analysis of the seismostratigraphic record, which arises from the tectonic model for regional sea level fluctuations presented here. This will be followed by a discussion of some of the new elements in passive margin and sedimentary basin research that follow from this work.

A new tectonic model for regional sea level fluctuations

Although the Vail et al. (1977) curves are based on data from various basins around the world, they have been heavily weighted in favour of North America, the Gulf Coast, the northern and central Atlantic margins and especially the North Sea. Their results, therefore, do not seem to be a true reflection of global changes; this is evident from Figure 2 which shows several regions that conform with the 'global' pattern neither in magnitude nor in timing of the cycles. Pitman & Golovchenko (1983) pointed out that glacial fluctuation in sea level is the only known mechanism that can cause changes in sea level at rates in excess of 1 cm/1000 years and with magnitudes in excess of 100 metres. These may, therefore, explain the Oligocene unconformities, but they cannot explain those major parts of the sea level curve where the glacial cycles are thought to have been insignificant (Thorne & Watts 1984). For example, with the exception of the Oligocene event, there is no evidence in the geological and geochemical records for significant Mesozoic and Cenozoic glacial events prior to middle Miocene (Frakes 1979). Alternatively, the sea level curves were attributed to tectonic processes (e.g. Bally 1980, 1982; Hallam 1984; Watts 1982; Veizer 1985) although the nature of this mechanism for explaining rapid short-term fluctuations remained obscure (Hallam 1984; Pitman & Golovchenko 1983; Watts 1982).

The tectonic model recently proposed by Cloetingh et al. (1985) represents the interaction between intraplate stresses and the deflections of the lithosphere caused by sedimentary loading (Fig. 3). This interaction can produce apparent sea level changes of more than 100 metres within a few Ma at the flanks of sedimentary basins. Therefore, we were able to argue that glacial fluctuations (Pitman & Golovchenko 1983) are not the only mechanisms capable of producing apparent sea level changes with a magnitude and rate inferred from the stratigraphic record. Cloetingh et al.'s model also explains contemporaneous fluctuations in apparent sea level in neighbouring basin settings. The action of changing horizontal stresses is not restricted to passive margins but also modifies the vertical movements within intracratonic basins. As such this mechanism may also provide a tectonic explanation for some of the observed correlations between the timing of the sea level changes in oceanic and intracontinental regions noted by Sloss (1979) and Bally (1980).

Calculated apparent sea level changes

The total deflection of the lithosphere at passive margins is dominated by sediment loading and thermal contraction (Sleep, 1971). In our modelling, the sediment load is represented by two adjacent triangular wedges, one on the continental shelf, the other on the continental rise. We adopt a reference model for sedimentary loading in which the maximum height of the sedimentary wedges corresponds to the thickness that would be attained provided that sedimentation has kept pace with the subsidence predicted by the boundary layer model of the underlying oceanic lithosphere (Turcotte & Ahern 1977; Wortel 1980). As a result the maximum thickness of the sedimentary wedges increases gradually in proportion with the square root of the age up to a maximum value of 7.3 km at 100 Ma (Fig. 3). This reference model gives a fair representation of the sediment loading histories and thicknesses at passive margins and agrees with the observation that sediment accumulation rates tend to decrease with time after the initial rifting phase (Southam & Hay 1981; Cloetingh 1982). Furthermore this sediment loading model allows us to equate onlap and offlap with rises and falls in sea level, respectively.

Studies of the flexural response of oceanic lithosphere to seamount loading and to bending during the subduction process (Caldwell & Turcotte 1979; Cazenave et al. 1980; Bodine et al. 1981; McAdoo et al. 1985) as well as studies of the distribution of the maximum depth of oceanic intraplate seismicity (Wiens & Stein 1983), show an increase in the effective flexural rigidity, or equivalent elastic thickness, of the oceanic lithosphere with age of the crust. Thus, the response of the oceanic lithosphere to the sediment load (Watts et al. 1982), and the interaction between this response and the in-plane stress, is time dependent not only because the sediment load builds up with time, but also because of the changing mechanical properties of the lithosphere.

The interactions between sediment loading, lithospheric thermal evolution and intraplate stresses are calculated using finite element techniques. We investigate the response of the lithosphere, adopting an equivalent elastic layer whose thickness increases with age according to a square root of age function (Bodine et al. 1981). Of interest here is the modification of the basin shape by variations in intraplate stress fields. Figure 4 (see inset) shows the effect for a sedimentary basin underlain by 30 Ma old oceanic lithosphere with a corresponding elastic plate thickness. A transition from tension of 1 kbar to compression of the same magnitude produces a net uplift (or apparent sea level fall) of up to about 40 metres at the edge of the basin. The deflections at other stress levels can be scaled in proportion to the magnitude of the applied horizontal stresses with a stress field of a few hundred bars, corresponding to a deflection of about 10 metres. The dependence of deflection on the age of the underlying lithosphere is demonstrated in Figure 4. The differential uplift ΔW, defined as the difference in deflection for the change in stress or in-plane force from tension to a compression of equal magnitude, is computed for

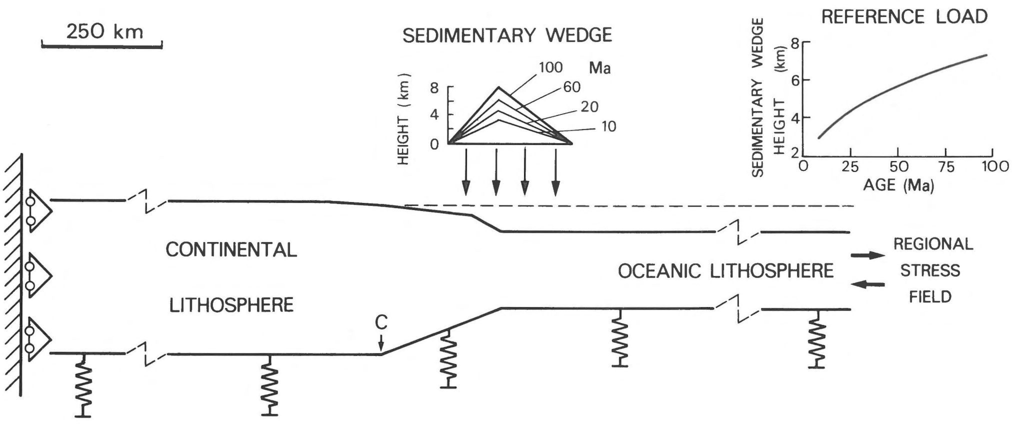

Fig. 3. Tectonic model for apparent sea level fluctuations. Variations in intraplate stress field effect the vertical displacements at a passive margin evolving through time, due to the thermal evolution of the lithosphere and the loading of this lithosphere by a wedge of sediments (see inset for reference model of sediment loading). A model is adopted with uniform elastic properties; rheological differences between continental and oceanic lithosphere are neglected for simplicity.

the edge of the basin as a function of variations in the in-plane force and stress. Curves (I)–(V) illustrate the deflection for changes in stress from 0.25 to −0.25 kbar, from 0.5 to −0.5 kbar, from 1 to −1 kbar, from 2 to −2 kbar, and from 3 to −3 kbar, respectively, in all cases with the same time-dependent reference sediment load (negative stress denotes compression). In these models the in-plane force increases with increasing lithospheric age because of the associated thickening. Curve (VI) illustrates the deflection, under the same reference load, for a change in the in-plane force from 5×10^{12} Nm^{-1} to -5×10^{12} Nm^{-1}, irrespective of lithospheric thickness.

On the basis of these calculations we conclude that the effectiveness of external forces in producing deflections increases with increasing thickness of deposited sediments but decreases with the increasing age of the lithosphere. Hence, external forces are most effective at those young passive margins that are subject to rapid sediment loading. External forces are least effective at old margins subjected to low rates of sediment loading. This is most clearly seen when a constant sediment load is placed on an oceanic lithosphere of increasing effective elastic thickness with age and when sediment loads greater than the reference model are placed on the lithosphere (curves VII–X , Fig. 4).

Similar changes in subsidence occur within the basins, although the relative subsidence there, on the order of a few hundred metres, is small in comparison with the total subsidence which is on the order of several kilometres. The estimates of the magnitude of the vertical displacement from our modelling of the deflection of a uniform elastic lithosphere are conservative. Introduction of a depth-dependent rheology (Goetze & Evans 1979) in the modelling would enhance the effectiveness of the action of variations in intraplate stress. The same is true for rheological weakening of the lithosphere due to flexural stresses (Cloetingh et al. 1982, 1984) induced by the sediment loading. The vertical movements within the basins become more complex when the effective flexural thickness is varied across the continental margin (Cloetingh et al. 1985). In this case an additional tilting of the crust is induced at the transition from oceanic to continental lithosphere. This tilting is a consequence of the change in the thickness of the layer that carries the intraplate stress, and occurs even in the absence of sediment loading. These tilts amplify the vertical displacement at the basin edge shown in Figure 4 when the effective elastic thickness of the continental lithosphere is less than that of the oceanic lithosphere. Both theoretical studies (Cloetingh & Nieuwland 1984) and observations (Vink et al. 1984; Steckler & Ten Brink, in press) of rifting processes in oceanic and continental

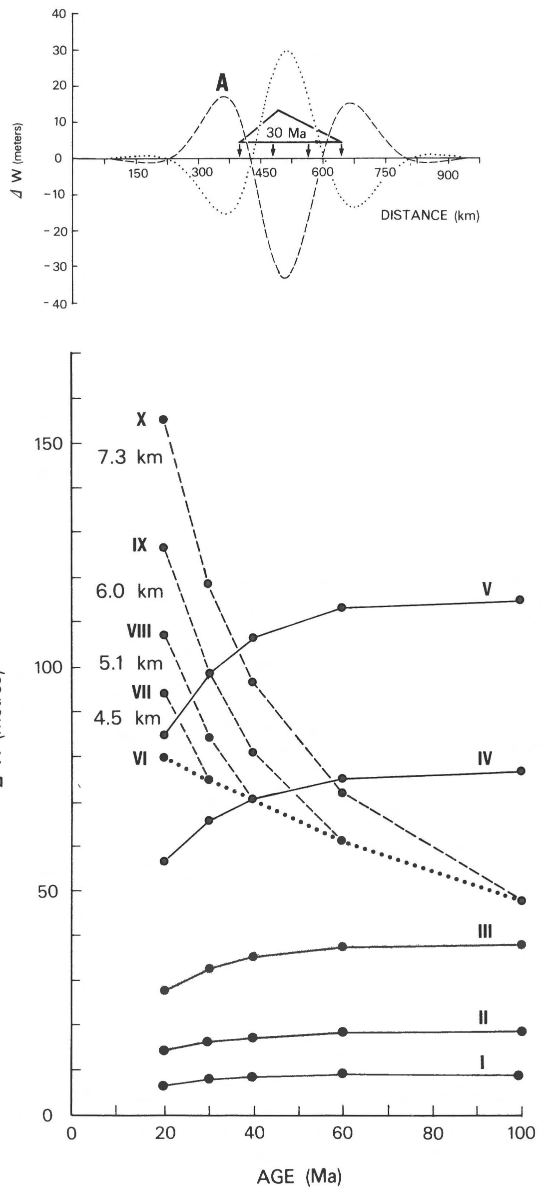

Fig. 4. Differential uplift (metres) at basin edge (position A indicated in inset) due to superposition of variations in regional stress field on flexure caused by sediment loading, plotted as a function of the age of the underlying lithosphere. Solid lines (I–V): results for a fixed stress transition from 0.25, 0.5, 1, 2, and 3 kbar tension to 0.25, 0.5, 1,2, and 3 kbar compression, respectively, superimposed on sediment loading according to the reference model. Dotted curve (VI): results for a transition from $5 \times 10^{12}\,Nm^{-1}$ tension to $5 \times 10^{12}\,Nm^{-1}$ compression, corresponding to stresses varying between ± 1.25 kbar at 100 Ma and ± 2.8 kbar at 20 Ma, superimposed on sediment loading according to the reference model. Dashed lines (VII–X): results from sediment loading models that are different from the reference model. Heights of the sedimentary wedges are specified in km. Inset: differential subsidence or uplift (metres) from the deflection due to sediment loading and thermal contraction of 30 Ma old oceanic lithosphere caused by an intraplate stress field of 1 kbar compression (dashed curve) and tension (dotted curve). Sign convention for deflections: uplift is positive, subsidence is negative.

lithosphere indicate that continental lithosphere at passive margins might be mechanically weaker than the adjacent oceanic lithosphere. The effect is, however, highly dependent on the assumed rheological contrast between oceanic and continental lithosphere and the degree of amplification or reduction will therefore vary for different basins.

Figure 4 demonstrates that variations in relative sea level of at least ten metres can be caused by regional changes in the in-plane stresses of the order of a few hundred bars. If these stress changes occur on a timescale of 10^6 years, then the associated sea level change rates are on the order of at least 1 cm/1000 years. The actual magnitude obtained for a given change in stress is controlled by the magnitude of the perturbation or deflection of the lithosphere at the time that the in-plane stress is applied, or in the context of passive margin evolution, on the rate of sedimentation, and on the response of the lithosphere to this sediment load.

Independent estimates for the magnitude of the apparent sea level changes

Different estimates exist for the magnitude of the relative sea level changes (Pitman 1978; Watts & Steckler 1979; Wise 1974; Bond 1978). Pitman (1978) has scaled the Vail et al. (1977) curves from calculations based on changes in the volume of mid-ocean ridges caused by changes in spreading rates and ridge lengths since Cretaceous times, and estimates a maximum change of 350 m in about 70 Ma. A number of other studies, however, strongly support a lower magnitude for these fluctuations. Lowrie et al.'s (1980) analysis of the Cretaceous magnetic quiet zone leads to a reduction of Pitman's spreading rates and therefore to a reduced sea level change. Watts & Steckler (1979) examined bore hole records for the eastern margin of North America and proposed an average fall in sea level of about 100–150 metres since Cretaceous time, a value largely consistent (see also Kominz, 1984) with estimates derived from studies of continental flooding (Bond 1978; Wise 1974). Thorne & Bell (1983) derived a eustatic sea level curve from histograms of North Sea subsidence which is also consistent with lower estimates of the amplitude of the sea level changes.

New evidence from studies of Oligocene-Miocene carbon isotope cycles and abyssal circulation changes (Miller & Fairbanks 1985) and from modelling subsidence at the United States passive margin (Watts & Thorne 1984) has provided revised quantitative estimates for the magnitude of the mid-Oligocene fall in sea level. The magnitude of this fall in sea level, by far the largest shown in the Vail et al. (1977) curves, is estimated by Miller & Fairbanks (1985) and Watts & Thorne (1984) to be at most 50–60 metres. The modified Vail et al. curve (see also Vail & Todd, 1981) for Jurassic sea levels (Vail et al. 1984) retains the same overall form as the original coastal onlap and offlap curve, but exhibits a general reduction of the magnitude, with some of the corresponding sea level changes more nearly symmetrical. These findings lead us to explore here the consequences of the possibility that the majority of the short-term sea level falls inferred from seismostratigraphy, have a characteristic magnitude of only a few tens of metres within a time interval of a few Ma (Aubry 1985). The outcome of this study is important in connection with the magnitude and rate of the underlying variations in intraplate stresses.

Evidence for fluctuations in intraplate stress fields

Several independent studies of lithospheric deformation in active continental margin and intraplate tectonic settings lead to the conclusion that horizontal stresses exist in the lithosphere and that these stresses may reach magnitudes up to a few kbars. Ward (1983), for example, examined the depths of oceanic earthquakes associated with the bending of the lithosphere prior to subduction. He found that the neural surfaces in young oceanic lithosphere could be elevated by as much as 15 km above the corresponding neutral surfaces in old oceanic lithosphere and concluded that the regional stress field had to be of the same order of magnitude as the bending stresses which have previously been shown to be on the order of a few kbars (Caldwell et al. 1976; McAdoo et al. 1978). Folding of the oceanic lithosphere in the southern part of the Bay of Bengal has been interpreted in terms of deformation caused by compressive forces on the order of several kbars (Weissel et al. 1980; McAdoo & Sandwell 1985). Investigations of the departures from isostatic equilibrium in several tectonic provinces within Australia have also led to the postulation of in-plane stresses at the level of a few kbars (Lambeck 1983; Lambeck et al. 1984). Studies of the formation of sedimentary basins also indicate horizontal (tensional) stresses on the order of a few kbars (Houseman & England 1986; Cloetingh & Nieuwland 1984).

Recent numerical modelling of stresses induced by plate-tectonic forces in the lithosphere (Wortel & Cloetingh 1981, 1983; Cloetingh & Wortel 1985, in press) has yielded greater insight into the underlying causes of the variations in stress level observed in different plates. The driving plate-tectonic forces are 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, and the pull acting on the downgoing slab in a subduction zone. The incorporation of the dependence of slab pull and ridge push on the age of the oceanic lithosphere in our modelling provided a quantitative basis for analysing, explaining and even predicting various deformational processes in lithospheric plates from the resulting stress field. Of particular interest is the comparison of stress fields in various plates. Our modelling of the stress field in the Indo-Australian plate demonstrated that the joint occurrence in this single plate of an exceptionally high level of compressive deformation in the plate's interior (McAdoo & Sandwell 1985; Bergman & Solomon 1985) and normal faulting in the near-ridge areas (Wiens & Stein 1984), is a transient feature unique to the present-day dynamic situation of the Indo-

Australian plate. The high level of intraplate seismicity in the Indo-Australian plate makes intraplate earthquakes a reliable indicator of the intraplate stress field (Bergman & Solomon 1985). Stress orientation data from Bergman & Solomon (1985) given in Figure 5b show a rotation of the observed stress field in the Northern Indian Ocean from N-S oriented compression in the north to a more NW-SE directed compression in the southeastern part of the Bay of Bengal region. A similar pattern is found in the calculated stress model (see Fig. 5a). Furthermore, the stress field as calculated gives a consistent explanation for the observed concentration of seismic activity and significant deformation in the oceanic crust (Geller et al. 1983; McAdoo & Sandwell 1985) in the northeastern Indian Ocean segment of the Indo-Australian plate. The stresses in the Indo-Australian plate are an order of magnitude greater than those we have calculated using the same technique for the Nazca plate. The latter are on the order of 500 bar (Wortel & Cloetingh 1985), a stress level more characteristic for plates not involved in continental collision or fragmentation processes.

An important feature of the stress field in the Indo-Australian plate is the occurrence of strong lateral *spatial variations* in the stress field along its trench systems (Fig. 6). The lateral stress variations along the Java-Sumatra trench (Cloetingh & Wortel 1985, 1986), and similar variations along the Peru-Chile trench (Wortel & Cloetingh 1985) are in excellent quantitative agreement with marine geophysical data on the structure of the trench regions (see Wortel 1986, this volume, for a more extensive discussion). The transition from a compressive stress field off and parallel Sumatra to a tensional stress field normal to Java, is the result of the contrast in age of the subducted lithosphere under Sumatra (40–70 Ma) and Java (140 Ma). Following the Sunda arc in an easterly direction, underthrusting of continental shelf occurs from just west of Flores onwards, which results in a compressive stress field off this trench segment. Rapid *temporal variations in stress* have occurred here, at the onset of the Banda arc collision, where a period during which the stress field was controlled by the slab pull associated with subduction of old oceanic

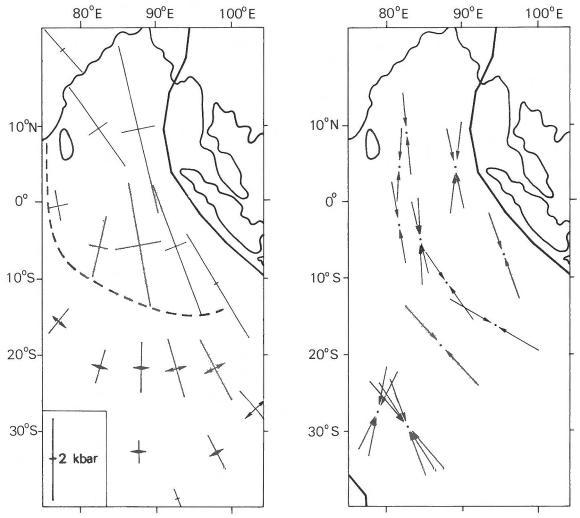

Fig. 5. Intraplate stress field in the northern Indian Ocean. a (left). Calculated stress field (after Cloetingh & Wortel, in press). Horizontal deviatoric stresses averaged over a uniform elastic plate with a reference thickness of 100 km are plotted. Symbols (↔) and (—|—) denote tension and compression respectively. The length of the arrows is a measure for the magnitude of the stresses. The dashed line is the southern limit of the observed deformation in the northern Indian Ocean (Geller et al., 1983). b (right). The orientation of maximum horizontal compressive stress inferred from a focal mechanism study by Bergman & Solomon (1985).

lithosphere, has been followed by a phase of net compressive resistance due to the arrival of buoyant continental lithosphere at the subduction zone. Five Ma ago a stress field dominated by tension changed to a stress field of predominantly compressional character. Such changes induced at the convergent boundaries propagate into the interiors of the plates, where they affect passive margins and intracratonic basins.

Rapid temporal changes in stress field are not limited to collision processes. Fragmentation of plates is also associated with drastic changes in stress state. An example of such an event is the break-up of the Farallon plate into the Cocos and Nazca plates. We have demonstrated (Wortel & Cloetingh 1983) that fragmentation of the Farallon plate took place under the influence of tensional stresss induced by age-dependent slab pull forces (see also Fig. 7). Figure 7 shows the significant differences in stress level associated with relaxation

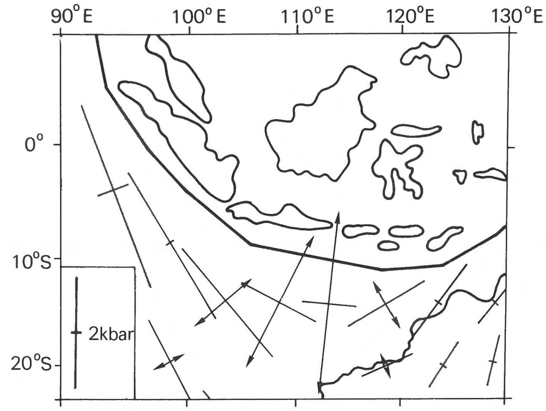

Fig. 6. Lateral variations in the intraplate stress field in the Indo-Australian plate segment close to the Sunda arc. Figure conventions as in Figure 5a.

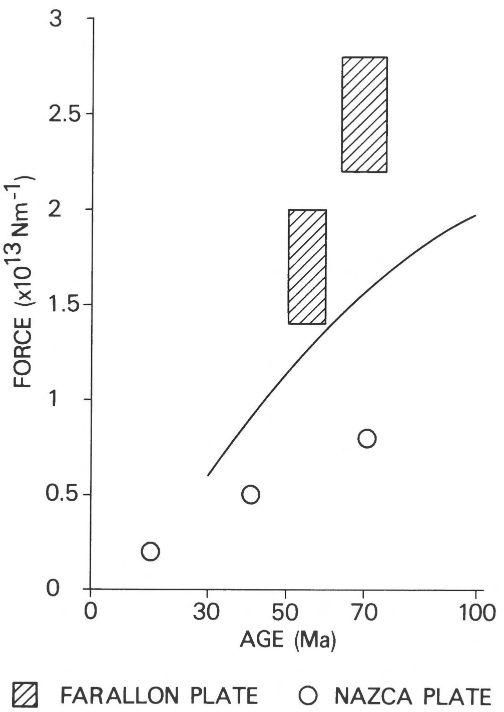

Fig. 7. The force required to induce tensional failure in oceanic lithosphere (rheology based on Goetze & Evans, 1979) plotted as a function of lithospheric age. Boxes indicate tensional paleostress levels of several kbars prior to the break-up of the Farallon plate into the Nazca and Cocos plates calculated by Wortel & Cloetingh (1981). Circles indicate intraplate stress levels generated in the present-day Nazca plate (Wortel & Cloetingh, 1983). The values referred to are representative for large parts of the Farallon plate and Nazca plate, respectively.

of tensional stresses of 2–3 kbar in the rifting process.

It has been shown in our previous work (Wortel & Cloetingh 1981; Cloetingh & Wortel 1985) that concentration of slab pull forces dominates the plate-tectonic stress field. Passive margins in plates not involved in collision or subduction processes are not subject to the influence of these slab pull forces. In such circumstances locally induced stresses can be much more important than the regional stress field induced by the remaining plate-tectonic forces. Thus, for passive margins located in the interiors of the American plates, stresses induced by sediment loading are an order of magnitude greater than the regional stress field associated with ridge push forces (Cloetingh et al. 1984). The latter is typically an order of magnitude of a few hundred bar. Under such circumstances, local adjustment of stresses at passive margins, e.g. by initiation of spreading in adjacent oceans, rarely involves changes of more than a few hundred bar.

The seismostratigraphic record: new source of information on paleo-stress

In the foregoing we have argued for a tectonic cause for apparent sea level changes. Others (e.g. Bally 1982; Watts 1982) have preceded us in this but were unable to identify a mechanism for lowering sea level. In particular, Bally (1982) has pointed out the strong correlation in timing of plate-tectonic reorganizations and lowerings in sea level shown in the Vail et al. (1977) curves.

The Vail et al. (1977) global curve, as already mentioned, is heavily weighted in favour of North America, the Gulf coast, the northern and central Atlantic margins and the North Sea (see Fig. 2). Therefore, the 'global' cycles strongly reflect the seismostratigraphic record of basins in a tectonic setting dominated by rifting events in the Northern and Central Atlantic Ocean. This applies in particular to the North Sea Basin where accurate timing of the different tectonic events is available (Ziegler 1982). Adopting a magnitude of 50 metres for the mid-Oligocene lowering in sea level, we have calibrated the Vail et al. (1977) generalised

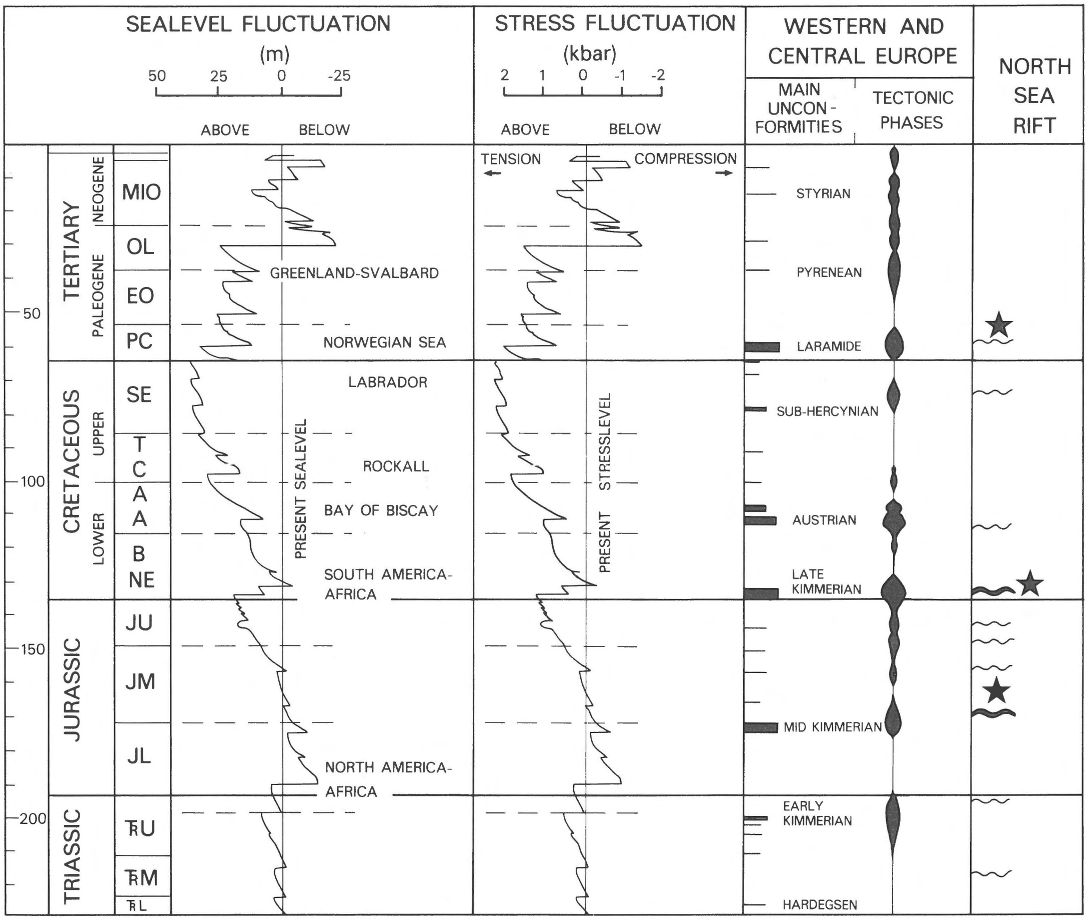

Fig. 8. Synthetic paleo-stress curve derived from the generalised seismostratigraphic record (Vail et al., 1977). Columns on the right hand side with the timing of tectonic events (after Ziegler, 1982) in western Europe and the North Sea Basin are given for comparison. Also indicated, along the sea level curve, are timings of the onset of sea-floor spreading in the Atlantic domain. Paleo-stresses are plotted relative to the present-day stress level. Thick and thin wavy lines denote major and minor rifting phases, respectively; stars denote volcanic activity.

sea level curve (see Fig. 8). On the basis of the outcome of the model calculations for the response of a uniform elastic layer summarized in Figure 4, we have inverted the information on changes in sea level in the calibrated Vail et al. (1977) curve to derive a paleo-stress curve. The result is displayed in Figure 8, where paleo-stresses are given relative to the present stress level. According to our modelling, falls in sea level at the edge of the basin are associated with relaxation of tensional intraplate stress or equivalently an increase in compressive stress (Fig. 8). As an example for comparison, Ziegler's (1982) timing of tectonic events in western Europe is given in Figure 8. This shows that the synthetic paleo-stress curve derived from the seismostratigraphic record mirrors the tectonic evolution of Northwestern Europe and the North Sea Basin: rifting episodes correspond to relaxation of tensional paleo-stresses.

Relative changes in sea level documented for the eastern North American margin concommitant with the break-up of the eastern American continental margin (Sheridan 1983) and early Cretaceous volcanism on the northeastern American margin (Jansa & Pe-Piper 1985) are anticipated to be the reflection of adjustment of stress associated with these tectonic events. Other areas where the timing of tectonic events and associated stress changes correspond with rapid changes in apparent sea level, include the south Pyrenean basin (Atkinson & Elliott 1986), the early Cretaceous foreland basins of western Montana (Suttner et al. 1986), the Late Cenozoic basins of the Mediterranean region (Meulenkamp 1982 and pers. comm. 1985) and the Australian passive margins (Cloetingh et al. 1985). Here, changes from transgression to regression at the onset of the Banda-arc and Himalayan collision events, associated with stress changes from tension to compression (Cloetingh & Wortel 1985), have been noted.

In general, stress variations of a few hundred bar associated with local adjustment of stress at passive margins and intracratonic basins would suffice to explain the largest part of their seismostratigraphic record. Stress changes of more than 1 kbar are required in order to induce sea level fluctuations with magnitudes of the order of 50 metres such as inferred for the mid-Oligocene event (Miller & Fairbanks 1985; Watts & Thorne 1984). These must be related to major reorganizations in lithospheric stress fields. In this context it is interesting to note that the mid-Oligocene fall in sea level is coincident with a global plate reorganization, presumably with a concomittant change in the paleo-stress state, in which the break-up of the Farallon plate into the Cocos and Nazca plates (Wortel & Cloetingh 1981) played a major part. The superposition of the effect of the tectonically induced fall in sea level and an important glacio-eustatic event might explain the exceptional magnitude of the mid-Oligocene fall in apparent sea level.

From the above, we see that the seismostratigraphic record can provide a new source of information for paleo-stress fields. Examination of the stratigraphic record of individual basins in a wide range of tectonic settings in connection with independent numerical modelling of paleostresses, such as carried out by Wortel & Cloetingh (1981), is required to fully exploit the potentials of this new avenue of research.

Discussion

By its character the tectonic mechanism discussed here can explain contemporaneous fluctuations in apparent sea level in neighbouring basinal settings. It is important to realize that the degree of correlation between the timing of sea level changes induced by fluctuations in intraplate stress fields depends primarily on the dimensions of the stress province. Numerical modelling (Wortel & Cloetingh 1985; Cloetingh & Wortel 1985) and observation of lithospheric deformation (McAdoo & Sandwell, 1985; Wiens & Stein 1984; Bergman & Solomon 1985; Zoback & Zoback 1980) show that the stress provinces can vary in size from that of an entire lithospheric plate to that of part of a plate with dimensions up to several thousand km. Boundaries of plate-tectonic stress provinces in plates do not necessarily coincide with the transition of oceanic and continental lithosphere at passive margins. Examples are found in the North American and Indo-Australian plates. The eastern and central parts of North America form a compressional stress province together with the Northern Atlantic, while the western part of the continent is subject to a tensional regime (Zoback & Zoback 1980). Western and central Australia together with the adjoining Indian Ocean share a regional compressional stress field while eastern Australia and the northern part of the Tasman Sea are under the influence of a tensional regional stress field (Cloetingh & Wortel 1985, in press). Small scale variations in stress can occur superimposed on long wavelength stress patterns. Examples of these are found in oblique shear zones (Ballance & Reading 1980) where alternations of compressional and tensional segments occur. The effects of such alternations in stress field are reflected in the seismostratigraphic onlap and offlap patterns along the eastern margin of Canada (A.J. Tankard, pers. comm. 1985), consistent with the tectonic mechanism discussed here. The regional character of the tectonic mechanism sheds new light on observed deviations from the Vail et al. (1977) global sea level curve. Although there was an initial tendency to apply the global curve literally to every individual basin there is an increasing number of published stratigraphic studies showing sequences of onlap and offlap events which deviate from the global cycles proposed by Vail et al. (1977). These include examples both from well studied areas in the northern (Hallam 1984; Harris et al. 1984) and southern hemisphere (Carter 1985; Chaproniere 1984). While such deviations from a global pattern are a natural consequence of the character of the tectonic mechanism, it does not rule out global events in the stratigraphic record. However, these are only to be expected when major reorganizations in stress field occur simultaneously in more than one plate, as conjectured for the early Cenozoic global plate reorganization (Rona & Richardson 1978), or when glacio-eustatics dominate.

An interesting outcome of our analysis is the dependence of the magnitude of the lowerings in sea level on the age of the margin. Thorne & Watts (1984) showed that relatively slow changes in sea level can cause unconformities at old passive margins while relatively rapid changes are required to form unconformities in a subsiding young margin. Note however, that the tectonic mechanism discussed in the present paper is most effective for young passive margins that are subject to rapid sediment loading.

Although we have concentrated in this paper on the relation between tectonics and stratigraphy at passive margins, the tectonic mechanism discussed is applicable in a wider range of sedimentary environments. Other settings where lithosphere is flexed under the influence of sediment loading occur at foreland basins (Quinlan & Beaumont 1984) and in the vicinity of volcanic islands (Ten Brink & Watts 1985). Despite its height of only a few hundred metres, the peripheral bulge flanking foreland basins is of particular stratigraphic interest (Quinlan & Beaumont 1984; Homewood 1986). The action of intraplate stresses of tensional or compressional character, of which the latter is more natural in this tectonic setting, can reduce or amplify the height of the peripheral bulge and, consequently, greatly influence the stratigraphic record at foreland basins. In this context it is interesting to note that recent studies of foreland basins (Atkinson & Elliott in press; Suttner et al. in press) give independent support for the role of intraplate stress fluctuations on basin stratigraphy.

Ten Brink & Watts' (1985) detailed study of the seismic stratigraphy of the flexural moat flanking the Hawaiian Islands provides an example of flexural deformation and associated onlap and offlap patterns in a distinctly different intraplate tectonic setting. The observed pattern of onlap, followed by offlap and rapid variations in onlap and offlap in the upper part of the stratigraphic record may be an expression of relaxation of tensional stresses during the formation of this part of the Hawaiian island chain followed by other phases of fluctuations in intraplate stress field. As such the seismic-stratigraphic record at Hawaii and other volcanic islands provides a potential source of information on fluctuations in the paleo-stress field in the Pacific plate, which lacks the stratigraphic record found at passive margins and foreland basins more characteristic for other plates.

Conclusions

The flexural behaviour of the lithosphere has important consequences for the tectonic and stratigraphic evolution of passive margins. Flexural stresses play a key role in the mechanics of the transformation of passive margins into active margins, and the interaction of intraplate stress with flexural loading forms a crucial element in basin stratigraphy. We have demonstrated the importance of the age of the oceanic lithosphere in passive margin tectonics and stratigraphy. Stress variations in the lithosphere of a few hundred bar can explain the major part of the seismostratigraphic record at passive margins. The tectonic mechanism for regional sea level fluctuations proposed by Cloetingh et al. (1985) explains the observed correlations (Bally 1982) between the timing of lowerings in sea level inferred from the seismostratigraphic record and tectonic events. By its nature, the model also explains contemporaneous fluctuations in apparent sea level in neighbouring basin settings. The action of changing intraplate stresses is not restricted to passive margins but also modifies the vertical movements within intracratonic basins, foreland basins and flexural moats flanking intra-oceanic volcanic complexes. For this reason the model may provide a tectonic explanation for some of the observed correlations between the timing of apparent sea level changes in oceanic and intracontinental regions. The seismostratigraphic record could provide a new source of information for paleo-stress fields. Examination of the stratigraphic record for individual basins in a wide range of tectonic settings in connection with independent numerical modelling of paleo-stresses, is required to fully exploit the potentials of this new avenue of research.

Acknowledgments

I thank Prof. N.J. Vlaar for constructive support for my studies of passive margin evolution and Dr. Rinus Wortel for close and stimulating co-operation in numerical modelling of intraplate stress fields. Prof. K. Lambeck and Dr. H. McQueen made significant contributions to the initial phase of the work on the tectonic causes of sea level changes. Dr. J. Thorne and Dr. J.E. Meulenkamp offered valuable suggestions. Drs. J. Veizer, U. ten Brink and D. McAdoo furnished preprints of their papers.

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