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INTRAPLATE STRESSES: A TECTONIC CAUSE FOR THIRD-ORDER CYCLES IN APPARENT SEA LEVEL?

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ABSTRACT: Thermo-mechanical modeling demonstrates that tectonically induced vertical motions of the lithosphere may provide an explanation for third-order cycles in apparent sea level deduced from the seismic stratigraphic record of passive margins. The interaction of fluctuations in intraplate stresses and the deflection of the lithosphere caused by sedimentary loading can produce apparent sea-level changes of as much as 100m at the flanks of passive margins.

In general, stress variations of a few hundred bars associated with local adjustment of stresses at passive margins suffice to explain a significant part of the stratigraphic record associated with short-term variations in sea level on the order of a few tens of meters. To induce short-term apparent sea-level fluctuations with magnitudes on the order of 50m or more, which occur less frequently in the record, changes in stress level in excess of one kbar are required. These larger fluctuations in apparent sea level could be related to major reorganizations of lithospheric stress fields due to rifting and fragmentation of plates, dynamic changes at convergent plate boundaries, or collision processes. A fluctuating horizontal stress field in the lithosphere can explain contemporaneous changes in apparent sea level in neighboring depositional environments. In principle, it implies the possibility of regional correlations in different basin settings. Specific short-term fluctuations in the curves of Vail and others, (1977; 1984) can be associated with particular plate tectonic reorganizations of lithospheric stress fields. The seismic stratigraphic record may provide a new source of information on paleo-stress fields which can be correlated with results of independent numerical modeling of intraplate stresses.

INTRODUCTION

During the last decade, major advances in quantitative analysis of the sea-level record in sedimentary basins have resulted from studies by Vail and his coworkers at Exxon (Vail and others, 1977; Vail and Todd, 1981; Vail and others, 1984). At the same time, there has been a continuous debate over the mechanisms that cause third-order cycles in sea level (e.g., Bally, 1980, 1982). Recently, Cloetingh and others (1985a,b) have proposed a new tectonic mechanism for apparent sea-level variations of about 1-10 cm/1,000 yr, with a maximum magnitude of a few hundred meters. Their model can explain these third-order cycles in sea level, provided that horizontal stresses exist in the lithosphere and that changes in these stress fields occur on time scales of a few million years and longer. In this paper, we will concentrate on the consequences of the action of intraplate stresses for the stratigraphic record. We shall then investigate the evidence for changes in the intraplate stress field. This will be followed by a discussion of some new elements in passive margin and sedimentary basin research that follow from this work. Mechanisms for long-term changes in sea level (e.g., Kominz, 1984; Heller and Angevine, 1985) fall beyond the scope of the present paper.

A TECTONIC MECHANISM FOR SHORT-TERM SEA-LEVEL FLUCTUATIONS

Pitman and Golovchenko (1983) pointed out that glacially induced fluctuation in sea level is the only known mechanism that can cause changes in sea level at rates in excess of 1 cm/1,000 yr and with magnitudes in excess of 100 m. Glacial-eustatic changes may, therefore, explain the Oligocene unconformities, but cannot explain those third-order cycles of sea level where glacial events are thought to have been insignificant (Thorne and 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 record has been attributed to tectonic processes (e.g., Bally, 1980, 1982; Watts, 1982; Hallam, 1984; Veizer, 1985; Hallam, this volume), although the nature of these mechanisms for explaining rapid short-term fluctuations remained unclear (Watts, 1982; Pitman and Golovchenko, 1983; Hallam, 1984; Miall, 1986).

We interpret apparent fluctuations in sea level, or onlap/offlap sequences, as expressions of regional, and possibly global, tectonic processes rather than in terms of eustatic changes. It is recognized, however, that the latter are always potentially present and that a separation of the two processes may not be possible without independent supporting evidence concerning, for example, climatic change.

The tectonic model proposed by Cloetingh and others (1985a,b) represents the interaction between intraplate stresses and the deflections of the lithosphere caused by sedimentary loading (Fig. 1). This interaction can produce apparent sea-level changes of more than 100 m within a few million years along the flanks of sedimentary basins. Therefore, Cloetingh and others (1985a,b) argued that glacial fluctuations (Pitman and Golovchenko, 1983) are not the only mechanisms capable of producing third-order cycles in apparent sea level with a magnitude and rate comparable to those inferred from the stratigraphic record. The model of Cloetingh and others can explain contemporaneous fluctuations in apparent sea level in neighboring basin settings. The action of changing horizontal stresses is not restricted to passive margins but also modifies the vertical movements within intracratonic basins and foreland basins. This tectonic mechanism may provide an 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).

The basin stratigraphy and apparent sea-level record can exhibit considerable variations associated with changes in the stress field. The model of Cloetingh and others (1985a,b)

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FIG. 1.—Tectonic model for fluctuations in apparent sea level. Variations in intraplate stress field affect 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.

requires that intraplate stress fields be of sufficient magnitude to modulate the deflections of the lithosphere caused by other subsidence processes and that they have variations on time scales of a few million years and longer. In what follows, we review evidence for the existence of such intraplate stress fields.

Calculated apparent sea-level changes.—

The total deflection of the lithosphere at rifted margins is dominated by sediment loading and thermal contraction (Sleep, 1971). Subsidence at a rifted margin is controlled by the stretching of crust or lithosphere and by subsequent cooling and sediment loading. The subsidence curves will therefore be characterized by a decreasing subsidence rate with time, the magnitude of subsidence being a function of lithospheric properties, driving force parameters, the rate of cooling, and the rate of sedimentation. To model passive-margin processes, we consider, for convenience, an elastic oceanic lithosphere that cools with time according to the boundary layer model of thermal conduction. In our modeling, the sediment load is represented by two adjacent triangular wedges, one on the continental shelf, the other on the continental rise. Sedimentation is assumed to be sufficiently rapid to keep up approximately with subsidence (Turcotte and Ahern, 1977). Because thermal subsidence is proportional to the square root of age, the maximum thickness of the sedimentary wedges grows with the same time dependence. This leads to a maximum sedimentary thickness of 7.3 km at 100 Ma (Fig. 1). This reference model gives a fair representation of the sediment-loading histories and thicknesses at rifted margins and agrees with the observation that sediment accumulation rates tend to decrease with time after the initial rifting phase (Southam and Hay, 1981). Although certainly an oversimplification (see Pit-

man, 1978, and Pitman and Golovchenko, 1983, for a discussion), the assumption that sedimentation keeps up with subsidence allows us to equate onlap and offlap with rises and falls in sea level, respectively.

Analyses of the flexural response of oceanic lithosphere to tectonic processes (Bodine and others, 1981; McAdoo and others, 1985) as well as seismo-tectonic studies (Wiens and 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 and others, 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 here 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 and others, 1981). Of particular interest is the modification of the basin shape by variations in intraplate stress fields. Figure 2 (see inset) shows the effect for a sedimentary basin underlain by 30-Ma oceanic lithosphere with a corresponding elastic plate thickness. A transition from tension of 1 kbar (1 kbar = 100 MPa) to compression of the same magnitude produces a net uplift (or apparent sea-level fall) of as much as about 40 m at the edge of the basin, immediately landward of the principal sediment load. The deflections at other stress levels are proportionally scaled to the magnitude of the applied horizontal stresses, with a stress field of a few hundred bars corresponding to a deflection of about 10 m. Evidence for the existence of intraplate stress fields with magnitudes as much as a few kilobars is presented later.

The dependence of deflection at the edge of the basins on the age of the underlying lithosphere is demonstrated in Figure 2. The differential uplift Δw , defined as the difference in deflection for a change in stress or in-plane force from tension to compression of equal magnitude, is computed for 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 $5 \times 10^{12} \text{ Nm}^{-1}$ to $-5 \times 10^{12} \text{ Nm}^{-1}$, irrespective of lithospheric thickness.

The deflection of the lithosphere at the edge of the basin is of particular stratigraphic interest. External compression (or, equivalently, relaxation of tension) narrows the basin, causes relative uplift at the edge, and manifests itself as stratigraphic offlap. An apparent fall in sea level results, possibly exposing the sediments to produce an erosional or weathering horizon. External tension (or equivalently, relaxation of compression) widens the basin, causes relative subsidence at the edge, and manifests itself as stratigraphic onlap. This situation is characterized by an apparent rise in sea level, and renewed deposition, with a corresponding facies change, is possible. Figure 3 displays in an idealized fashion the stratigraphy for the edge of a sedimentary basin underlain by 40-Ma lithosphere, as predicted on the basis of the calculations shown in Figure 2. The synthetic stratigraphy at the basin edge is schematically shown for the following three situations: (a) widening of the basin with cooling in the absence of an intraplate-stress field (Fig. 3a), (b) the same case with a superimposed transition to 500-bar compression at 30 Ma (Fig. 3b), and (c) the case of a stress change to 500-bar tension at 30 Ma (Fig. 3c). Watts and others (1982) have pointed out the differences in stratigraphy predicted at the basin edge for elastic and viscoelastic models of the lithosphere in the absence of intraplate stresses. These authors showed that for elastic models, the width of the basin increases, inducing progressive overstepping of sediments caused by the increase in the thickness of the elastic layer since basin formation. They demonstrated that viscoelastic models for the lithosphere predict narrowing of the infill of the basin, with younger sediments restricted to the basin center. Care should be taken, however, in rigidly classifying basins as elastic or viscoelastic strictly on the basis of this criterion. The stratigraphy we predict for a basin located on an elastic lithosphere and under the influence of a compressional-stress regime during deposition of younger strata will be very similar to the stratigraphy characteristically attributed (Watts and others, 1982) to the response of viscoelastic lithosphere to surface loading alone.

Our modeling demonstrates that the incorporation of intraplate stresses in elastic models of basin evolution can, in principle, predict a succession of onlap and offlap patterns such as observed along the flanks of passive margin

FIG. 2.—Differential uplift (meters) 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 $5 \times 10^{12} \text{ Nm}^{-1}$ tension to $5 \times 10^{12} \text{ Nm}^{-1}$ 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 kilometers. Inset: differential subsidence or uplift (meters) from the deflection due to sediment loading and thermal contraction of 30-Ma 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.

FIG. 3.—Idealized stratigraphy at the edge of a basin underlain by 40-Ma lithosphere predicted on the basis of the calculations shown in Figure 2. (a) Onlap associated with cooling of the lithosphere in the absence of an intraplate stress field. (b) A transition to 500-bar compression at 30 Ma induces a short-term phase of offlap at that time. (c) A transition to 500-bar tension causes an additional short-term phase of onlap at 30 Ma.

of the United States (Fig. 4). Such stratigraphy can be interpreted as the natural consequence of the mechanical widening and narrowing of basins by fluctuations in intraplate stresses, superimposed on the longer term broadening of the basin with cooling since its formation.

The calculations show 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. 2).

Similar changes in subsidence occur within the basins, although the relative subsidence there, on the order of a few hundred meters, is small in comparison with the total subsidence, which is on the order of several kilometers. The sign and magnitude of the apparent sea-level change will be a function of the sampling point, which may provide a means of testing the model or distinguishing this mechanism from eustatic contributions. The estimates of the magnitude of the vertical displacement from our modeling of the deflection of a uniform elastic lithosphere are conservative. Introduction of a depth-dependent rheology with a brittle-ductile transition (Goetze and Evans, 1979) in the modeling would enhance the effectiveness of the action of variations in intraplate stress (Cloetingh and others, 1982). The same is true for rheological weakening of the lithosphere due to flexural stresses (Cloetingh and others, 1982, 1983) induced by sediment loading. In areas that combine the presence of thick sedimentary loads and a high level of intraplate stress, such as the Bay of Bengal (Cloetingh and Wortel, 1985), this can even cause folding of the lithosphere (McAdoo and Sandwell, 1985). Using Seasat altimeter data and a depth-dependent rheology of the lithosphere, McAdoo and Sandwell (1985) showed that present compressional stress levels in the northeastern Indian Ocean (Fig. 5) can cause the observed basement undulations with wavelengths of roughly 200 km and amplitudes as much as 3,000 m. In general, however, for deformation not reaching the buckling limit, the effect of depth-dependent lithospheric rheology will be less dramatic (Cloetingh, in prep.).

The vertical movements within the basins become more complex when the effective flexural thickness is varied across the continental margin (Cloetingh and others, 1985a). 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, possibly due to changes in lithospheric thinning for rifted basins, and it occurs even in the absence of sediment loading. These tilts amplify the vertical displacement at the basin edge (shown in Fig. 2) when the effective elastic thickness of the continental lithosphere is less than that of the oceanic lithosphere. Studies of rifting processes in oceanic and continental lithosphere (Vink and others, 1984) 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 2 demonstrates that variations in relative sea level of at least 10 m can be caused by regional changes in the in-plane stresses on the order of a few hundred bars. If these stress changes occur on a time scale of 10^6 years, then the associated sea-level change rates are on the order of at least 1 cm/1,000 yr. The actual sea-level variation 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 basin evolution, by the rate of sedimentation and the response of the lithosphere to this sediment load.

FIG. 4.—Stratigraphic cross section of the shelf of the United States Atlantic margin at Cape Hatteras. The shelf break is about 40 km from the left of the figure. Ages of formation boundaries are given in million years (after Sleep and Snell, 1976).

Independent estimates for the magnitude of apparent sea-level changes.—

Different estimates exist for the magnitude of relative sea-level changes (e.g., Bond, 1978; Pitman, 1978; Watts and Steckler, 1979). Pitman (1978) has scaled the curves of Vail and others (1977) from calculations based on changes in the volume of mid-ocean ridges caused by changes in spreading rates and ridge lengths since Cretaceous times. He estimates a maximum change of 350 m in about 70 million years. Watts and Steckler (1979) examined borehole records for the eastern margin of North America and proposed an average long-term fall in sea level of about 100–150 m since Cretaceous time, a value largely consistent (see also Kominz, 1984) with estimates derived from studies of continental flooding (Bond, 1978). Hardenbol and others (1981) used a method similar to the one employed by Watts and Steckler (1979) on well data from offshore northwest Africa. They estimated mid-Cretaceous sea level to be about 300 m above present sea level. As noted by Falvey and Deighton (1982), it is unreasonable to expect that the well sequence of Hardenbol and others (1981) with 95 million years of missing section, should provide quantitatively better estimates than the well sequences analyzed

by Watts and Steckler (1979), which had only a few tens of millions of years of missing sections. Thorne and Bell (1983) derived a eustatic sea-level curve from histograms of North Sea subsidence that is 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 and Fairbanks, 1985) and from modeling subsidence at the passive margin of the United States (Watts and 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 curves of Vail and others (1977), is estimated by Miller and Fairbanks (1985) and Watts and Thorne (1984) to be at most 50–60 m. The modified curve of Vail and others (1984; see also Vail and Todd, 1981) for Jurassic sea levels retains the same overall form as the original coastal onlap and offlap curve, but exhibits a general reduction in 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 a significant number of the short-term sea-level fluctuations inferred from seismic stratigraphy have a characteristic magnitude of only a few

FIG. 5.—Intraplate stress field in the northeastern Indian Ocean. (a) Calculated stress field (after Cloetingh and Wortel, 1985). Principal horizontal non-lithostatic stresses averaged over a uniform elastic plate with a reference thickness of 100 km are plotted. Symbols ($\longleftrightarrow$) and ($\longleftarrow\rightarrow$) denote tension and compression, respectively. The length of the arrows is a measure of the magnitude of the stresses. The dashed line is the southern limit of the observed deformation in the northeastern Indian Ocean (Geller and others, 1983). (b) The orientation of maximum horizontal compressive stress inferred from a focal mechanism study by Bergman and Solomon (1985).

tens of meters within a time interval of a few million years (Aubry, 1985; S.O. Schlanger, pers. commun., 1986). We propose that such fluctuations may reflect the magnitude and rate of the underlying variation in intraplate stresses.

EVIDENCE FOR FLUCTUATIONS IN INTRAPLATE STRESS FIELDS

Independent studies of lithospheric deformation in active continental margin and intraplate tectonic settings have provided strong evidence for the existence of horizontal stresses in the lithosphere and for stress levels that may reach magnitudes of as much as a few kilobars. 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 kilobars (Geller and others, 1983; McAdoo and Sandwell, 1985). Departures from isostatic equilibrium in several tectonic provinces within Australia have been attributed to in-plane stresses at the

level of a few kilobars (Lambeck and others, 1984). The formation of sedimentary basins by lithospheric stretching also requires horizontal (tensional) stresses on the order of a few kilobars (Houseman and England, 1986).

Recent numerical modeling of stresses induced by plate-tectonic forces in the lithosphere (Wortel and Cloetingh, 1981, 1983; Cloetingh and Wortel, 1985) 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 into the modeling of the dependence of slab pull and ridge push on the age of the oceanic lithosphere provided a quantitative basis for analyzing, explaining, and predicting various deformational processes in lithospheric plates from the resulting stress field.

Of particular interest is the comparison of stress fields in various plates. Modeling of the stress field in the Indo-Australian plate (Cloetingh and Wortel, 1985) has shown that the joint occurrence in this single plate of an exceptionally high level of compressive deformation in the plate's interior (Bergman and Solomon, 1985; McAdoo and Sandwell, 1985) and normal faulting in the near-ridge areas (Wiens and Stein, 1984) is a transient feature unique to the present dynamic situation of the Indo-Australian plate. The high level of intraplate seismicity in the Indo-Australian plate enables the intraplate stress field to be estimated through the determination of earthquake focal mechanisms (Bergman and Solomon, 1985). Stress orientation data from Bergman and Solomon (1985) given in Figure 5b show a rotation of the observed stress field in the northern Indian Ocean from North-South-oriented compression in the north to a more Northwest-Southeast-directed compression in the southeastern part of the Bay of Bengal region. A similar pattern is found in the calculated stress model (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 and others, 1983; McAdoo and 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, which is characterized by a low level of intraplate seismicity. There, stresses are on the order of 500 bars (Wortel and 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 and Wortel, 1985), and similar variations along the Peru-Chile trench (Wortel and Cloetingh, 1985) are in excellent quantitative agreement with marine geophysical data on the structure of the trench regions. There is a close

correspondence between the extent to which grabens are developed on the seaward trench slope, and hence the style of accretionary tectonics in the trench region (Hilde, 1983), and the calculated regional stress field. The transition from a compressive-stress field off and parallel to 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, under-thrusting of continental shelf occurs from just west of Flores onward, 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 lithosphere has been followed by a phase of net compressive resistance due to the arrival of buoyant continental lithosphere at the subduction zone. A stress field dominated by tension changed to a stress field of predominantly compressional character 5 Ma. As demonstrated by observed paleo-stress orientations in western Europe (Letouzey and Trémolières, 1980), 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. Geologic evidence also points to episodic tectonic events on time scales of a few million years (Megard and others, 1984), possibly caused by the response of individual plates to longer term global readjustments of the plate motion patterns. Rifting or fragmentation of plates is also associated with drastic changes in stress state. An example of such an event is the breakup of the Farallon plate into the Cocos and Nazca plates. We have demonstrated (Wortel and Cloetingh, 1983) that fragmentation of the Farallon plate took place under the influence of tensional stresses induced by age-dependent slab pull forces. Strong support for the stress modeling procedure and the fragmentation hypothesis formulated for the Farallon plate is provided by Warsi and others (1983). In their marine geophysical studies of the Nazca plate off Peru, these authors discovered that the Mendana Fracture zone, oriented perpendicular to the Peruvian Trench, is in an incipient stage of spreading, very much along the lines suggested for the origin of the Cocos-Nazca spreading center (Wortel and Cloetingh, 1981). Concentration of slab pull forces dominates the plate-tectonic stress field (Wortel and Cloetingh, 1981; Patriat and Achache, 1984; Cloetingh and Wortel, 1985). 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, other sources of stress (e.g., membrane stresses; Turcotte and Oxburgh, 1976) 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 and others, 1982). The latter is typically on the order of a few hundred bars (Richardson and others, 1979). Under such circumstances, local adjustment of stresses at passive mar-

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.

gins (e.g., by initiation of spreading in adjacent oceans) rarely involves changes of more than a few hundred bars.

DISCUSSION

In the foregoing, we have argued for a tectonic cause for short-term apparent changes in sea level. Others (e.g., Bally, 1980; 1982; Watts, 1982) have proposed a tectonic explanation but were unable to identify a specific mechanism for short-term lowering in sea level. In particular, Bally (1982) has pointed out the strong correlation in timing of plate-tectonic reorganizations and some lowerings in sea level shown in the curves of Vail and others (1977).

The tectonic mechanism discussed here can explain contemporaneous fluctuations in apparent sea level in neighboring 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 will depend primarily on the dimensions of the stress province. Numerical modeling (Cloetingh and Wortel, 1985; Wortel and Cloetingh, 1985) and observation of lithospheric deformation (Zoback and Zoback, 1980; Wiens and Stein, 1984; Bergman and Solomon, 1985; McAdoo and Sandwell, 1985) show that the stress provinces can vary in size from that of an entire lithospheric plate to that of a small part of a plate. Boundaries of plate-tectonic stress provinces in plates do not necessarily coincide with the transition of oceanic to 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 and 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 and Wortel, 1985).

Ziegler (1982) and Sclater and Christie (1980) have pointed out the existence of a correlation between tectonic phases in the North Sea basin and changes in the history of opening of the Atlantic and Tethyan oceans (see also Schwan, 1985, for a more general discussion). Changes in spreading rates in the Atlantic and Tethyan regions are probably caused by or associated with changes in plate-tectonic forces. Both modeling of lithospheric stress fields and measurements of paleo-stress directions (Letouzey and Trémoières, 1980; J. Letouzey, pers. commun. 1986) strongly suggest that the induced stress changes are propagated over great distances from the plate boundaries into the northwestern European basins, which provides a dynamic explanation for the observed correlations. Furthermore, a causal relation might exist between these stress changes and the timing of salt diapirism in the North Sea basin (Thorne, 1986).

Small-scale variations in stress can occur superimposed on long-wavelength stress patterns. Examples of these are found in oblique-slip zones (Ballance and Reading, 1980), where alternations of compressional and tensional segments occur. The effects of such alternation in stress fields are

reflected in the seismic stratigraphic onlap and offlap patterns along the eastern margin of Canada (A.J. Tankard, pers. commun., 1985) and are consistent with the tectonic mechanism discussed here.

The regional character of the tectonic mechanism sheds new light on some observed deviations from the global sea-level curve of Vail and others (1977). These include examples from well-studied areas in both the northern (Hallam, 1984; Harris and others, 1984) and southern (Chaproniere, 1984; Carter, 1985) hemispheres. While such deviations from a global pattern are a natural consequence of the character of the tectonic mechanism, they do not preclude the presence of global events in the stratigraphic record. These are to be expected when major reorganizations in stress fields occur simultaneously in more than one plate, as conjectured for the early Cenozoic global plate reorganization (Rona and Richardson, 1978), or when glacio-eustatic changes dominate.

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

Relative changes in sea level documented for the eastern North American margin concomitant with the breakup of the eastern American continental margin (Sheridan, 1983) and early Cretaceous volcanism on the northeastern American margin (Jansa and Pe-Piper, 1985) may reflect the adjustment of stress associated with these tectonic events. Similarly, high-frequency oscillations in Cretaceous sea levels having characteristic periods of a few million years, which cannot be correlated with fluctuations in spreading rates and ridge lengths (S.O. Schlanger, pers. commun., 1986), could be the result of adjustment of stresses in the process of the opening of the South Atlantic. Other areas where the timing of tectonic events and associated stress changes correspond to short-term changes in apparent sea level include the south Pyrenean basin (Atkinson and Elliott, pers. commun., 1985), the Illinois basin (De Vries Klein, 1986), the late Cenozoic basins of the Mediterranean region (Meulenkamp, 1982; Meulenkamp and Hilgen, 1986), and the Australian passive margins (Cloetingh and others, 1985a). Here, changes from transgression to regression (Steele, 1976) at the onset of the Banda-arc and Himalayan collision events, associated with stress changes from tension to compression (Cloetingh and Wortel, 1985), have been noted.

In general, stress variations of a few hundred bars associated with local adjustment of stress at passive margins and intracratonic basins would suffice to explain a significant number of the short-term variations in sea level inferred from the seismic stratigraphic record. Stress changes of more than 1 kbar are required in order to induce sea-level fluctuations with magnitudes on the order of 50 m such as inferred for the mid-Oligocene event (Miller and Fairbanks, 1985; Watts and Thorne, 1984). These must be

related to major reorganizations in lithospheric-stress fields. As pointed out by Engebretson and others, (1985), major plate reorganizations generally occur with characteristic intervals of a few tens of millions of years. The duration of the individual events, and hence the associated stress changes, is, however, much shorter and on the order of at most a few million years (Engebretson and others, 1985). 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 concomitant change in the paleo-stress state, in which the breakup of the Farallon plate into the Cocos and Nazca plates (Wortel and 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 (see also Schlanger and Premoli-Silva, 1986).

Although we have concentrated 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 in foreland basins (Quinlan and Beaumont, 1984) and in the vicinity of volcanic islands (Ten Brink and Watts, 1985). Despite its height of only a few hundred meters, the peripheral bulge flanking foreland basins is of particular stratigraphic interest (Quinlan and Beaumont, 1984). 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 of foreland basins. Intraplate stresses may also affect the postulated interaction (Quinlan and Beaumont, 1984) of adjacent foreland arches.

The detailed study by Ten Brink and Watts (1985) 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 initial onlap followed by offlap and the occurrence later of 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 an ensuing period of fluctuations in intraplate stress fields. 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 of other plates.

From the above, we anticipate that the seismic stratigraphic record may provide a useful new source of information for paleo-stress fields. As has been shown recently (Cloetingh, 1986; Lambeck and others, 1987), a synthetic paleo-stress curve derived from the seismic stratigraphic record mirrors the tectonic evolution of northwestern Europe and the North Sea basin: rifting episodes correspond to relaxation of tensional paleo-stresses. Examination of the stratigraphic record of individual basins in a wide range of

tectonic settings in connection with independent numerical modeling of paleo-stresses, such as that carried out by Wortel and Cloetingh (1981), is required to exploit fully the potential of this new approach.

CONCLUSIONS

The interaction of intraplate stress with flexural loading forms a new element in basin stratigraphy. Stress variations in the lithosphere of a few hundred bars can explain a significant part of the seismic stratigraphic record at passive margins. The tectonic mechanism for regional sea-level fluctuations proposed by Cloetingh and others (1985a,b) explains observed correlations (Bally, 1980, 1982) between the timing of lowerings in sea level inferred from the seismic stratigraphic record and tectonic events. The model also explains contemporaneous fluctuations in apparent sea level in neighboring 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.

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