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On a tectonic mechanism for regional sealevel variations

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No satisfactory tectonic explanations have yet been offered for the apparent sealevel fluctuations of about 1 cm/1000 years and a magnitude of up to a few hundred meters that have been proposed by Vail et al. [1]. We propose a mechanism that does appear to be able to explain these changes if horizontal stresses of the order of a few kilobars occur in the lithosphere and if changes in these stress fields occur on geological time scales. The proposed model is one of an interaction between these stresses and the deflections of the lithosphere caused by sedimentary loading. Apparent sealevel changes of up to 100 m can be produced at the flanks of the sedimentary basins by this interaction. The mechanism is most effective for young margins that are subject to rapid sediment loading.

By its nature, the tectonic model can explain contemporaneous fluctuations in apparent sealevel in neighbouring depositional environments. In principle, it implies the possibility of regional correlations in different basinal settings.

1. Introduction

Since the publication of Vail et al.'s [1] global cycles of relative changes in sealevel, there has been a continuous debate on the mechanisms responsible for these fluctuations (e.g. [2]). Three orders of cycles were recognized by Vail et al., with durations of 200–300 Ma (first order), 10–80 Ma (second order) and 1–10 Ma (third order) respectively. A characteristic feature of these curves is the occurrence of slow increases in relative sealevel followed by periods of rapid falls on timescales of less than 10^6 years. As has been noted by several authors [2,3], much of the source data leading up to the Vail et al. curves is unpublished and it is not obvious whether these rapid falls are real reflections of apparent sealevel changes or artifacts of the method of analysis. Other studies [4,5] show a more symmetric pattern, for the rises and falls, with both occurring in time intervals of a few million years.

Although the Vail et al. curves are based on data from different basins around the world, they have been heavily weighted in favour of North America, the Gulf Coast and the northern and central Atlantic margins. Their results may, therefore, not be a true reflection of global changes; in particular their fig. 5 shows several regions that do

not conform with the “global” pattern in both the magnitude and timing of the cycles. The Gippsland Basin of southeastern Australia, for example, is quite anomalous [1,6] and some recent stratigraphic studies from other Australian margins indicate that the sharp sealevel drops characteristic of the Vail et al. curves are generally absent ([7]; G.C.H. Chaproniere, personal communication, 1984). Further examples of regional deviations from the global pattern, in particular from the Lower and Middle Jurassic of Western Europe and the North Sea are given by Hallam [5].

Widely different estimates exist for the magnitude of the relative sealevel changes [8–11]. Pitman [8] has scaled the 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 [12] analysis of the Cretaceous magnetic quiet zone leads to a reduction of Pitman's spreading rates and therefore to a reduced sealevel change. Watts and Steckler [9] examined bore hole records for the eastern margin of North America and proposed an average fall in sealevel of about 100–150 m since Cretaceous time, a value

that is largely consistent with independent estimates derived from studies of continental flooding [10,11]. Furthermore, Vail and Todd [13] no longer equate relative changes of coastal onlap with relative changes of sealevel. The modified Vail et al. curve for the Jurassic sealevels [13] has kept the same overall form as the original coastal onlap and offlap curve, with an overall reduction of the magnitude and with some of the corresponding sealevel changes being more symmetrical. By recalibrating the Vail et al. curves such that the post-Cretaceous change is 150 m, the shorter term fluctuations in relative sealevel are of the order of 50–150 m maximum in magnitude.

The cause of these changes has remained debatable. Pitman and Golovchenko [14] concluded that glacial fluctuation in sealevel represents the only known mechanism that can cause worldwide changes at rates in excess of 1 cm/1000 years and with magnitudes in excess of 100 m. These can therefore explain the Oligocene unconformities, but they cannot explain those major parts of the sealevel curve when the glacial cycles are thought to have been insignificant [15]. For example, with exception of the Oligocene event, there is no evidence in the geological and geochemical records for significant Mesozoic and Cenozoic glacial events before middle Miocene. Instead, the sealevel curves have been attributed to tectonic processes [2,5,14,16–19] although the nature of this mechanism for explaining rapid short-term fluctuations has remained obscure [5,14,16]. Pitman and Golovchenko [14] conclude "It must be emphasized that our inability at present to propose a mechanism that will cause lowering in sealevel of the rate and magnitude suggested by Vail and others does not mean that such events did not occur. We can only conclude that either there is a sufficient, but as yet unknown, mechanism or that the interpretation of the sedimentary patterns in terms of sealevel is incorrect". Their analysis of possible mechanisms is, however, incomplete and we propose another tectonic mechanism that does appear to be capable of producing variations in onlap and offlap in sedimentary basins of the required magnitude and timescale. We argue that variations in regional stress fields acting within inhomogeneous lithospheric plates are capable of producing vertical movements of the Earth's surface or the apparent sealevel changes

[20–22] of a magnitude equal to those deduced from the stratigraphic record. In this paper we examine the vertical displacements that can occur at a passive margin that is evolving through time due to the thermal evolution of the lithosphere, the loading of this lithosphere by sediments and a superimposed horizontal stress field that varies in magnitude or sign through time, possible in response to changing driving forces or boundary conditions at the plate margins [23,24].

The success of the proposed model rests on the magnitude of these horizontal stresses being comparable to the flexural stresses created by the sediment loading, that is, of the order of a few kilobars (1 kbar = 100 MPa). A number of studies [25–31] point to this being so, and this is discussed in section 4 below together with evidence for such changes on a time scale of a few million years.

2. Models

The essential characteristic of the model is a lithosphere whose deflections or deformations are magnified or reduced by horizontal forces acting on the plate. More specifically, the initial deformation considered here is the response of the plate to a sedimentary load deposited on a passive continental margin. This response is then modified by applied in-plane forces of compression or tension to produce a sequence of apparent sealevel changes. Studies of the flexural response of oceanic lithosphere to seamount loading and to bending during the subduction process [32–34], as well as studies of the distribution of the maximum depth of oceanic intraplate seismicity [35], are suggestive of 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, and the interaction between this response and the in-plane force, 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 in-plane forces are examined here using finite element models of two types (Fig. 1). The first set of models deals with the response of a uniform elastic layer whose thickness increases with age according to a square

Fig. 1. Model features: geometry, rheology, systems of forces and boundary conditions. Young's modulus $E = 7 \times 10^{10} \text{ N m}^{-2}$ and Poisson's ratio $\nu = 0.3$. Buoyancy forces counteracting the deflection are indicated by springs. Densities of sediments and mantle are 2.4 and 3.3 g cm^{-3} respectively. The height and the width of the sedimentary wedge (see inset for reference model of sediment loading) are given in kilometers at some selected time steps (specified in Ma). (a) Model with uniform elastic thickness. (b) Model with lateral variations in elastic thickness across a passive margin.

root of age function [34]. In the second group, lateral variations have been incorporated into the model through a variable effective elastic thickness of the type that is sometimes suggested for passive margins [32]. The layer is, for convenience, assumed to be elastic and to overlie an inviscid fluid so that there are no drag forces at the base of the plate. If stress relaxation occurs within the lithosphere, then some of the vertical displacements will grow with time [20–22]. The flexure of the lithosphere by the load is supported partially by a buoyancy force that is proportional to the deflection caused by loading. Zero horizontal displacements are prescribed for the left-hand boundary.

We are concerned here only with the gross features of the interactions between the various forces near the continental margin side of the basin, with the objective of testing the hypothesis that this interaction can produce the magnitudes

and rates of the fluctuations in relative sealevel that have been inferred from the seismic stratigraphy records. No detailed modelling of a specific basin and continental margin is intended.

2.1. Sediment loading

The sediment load is represented by two adjacent triangular wedges, one on the continental shelf, the other on the continental rise (see Fig. 1). We adopt as reference load one in which the maximum height of the wedges corresponds to the thickness that can be expected if sedimentation has kept up with the subsidence predicted by the boundary layer model of a cooling oceanic lithosphere [36]. As a result, the maximum thickness of the sedimentary wedges increases gradually with the square root of age up to a maximum value of 7.3 km at 100 Ma . This reference model presents a

fair representation of the sediment loading histories and thicknesses at passive margins [37] and agrees with the observation that sediment accumulation rates tend to decrease with time after the initial rifting phase of the margin [38].

The average shelf width of passive margins is observed to be about 110 km, while the average slope width is only about 10 km. The average observed width of the sediment load is about 250 km although it may be as little as 150 km for young margins [38]. We adopt an asymmetric triangular load, centred on the shelf edge and of width 250 km (see Fig. 1). Within the basin, the sediments displace water and the effective density of the load is taken to be 1.4 g cm^{-3} (e.g. [39]). For the sake of simplicity, the effects of the thermal blanketing of the sediments [40] and of lateral heat conduction across the basin edges [41] are ignored.

2.2. Models with lateral structural variations across the continental margin

During the initial breakup phase of passive margin evolution, zones of weakness develop which could control or influence the subsidence patterns during the early post-rift phase. The extent and depth of the fault zones, as well as the degree of uncoupling on them remains unclear, but a number of observations [42] suggests that these faults are locked very early in the post-rift evolution. We therefore model the plate as one without discontinuities, and in which the transition from oceanic to continental lithosphere occurs over a distance of 200 km [43]. This can be justified by the temperature profiles given by Zielinski [44] which illustrate the essentially linear transition from continental to oceanic lithosphere over a distance of 100–300 km. This applies particularly to the isotherms less than 600–700°C, those which are most pertinent to the mechanical properties of the lithosphere.

Estimates of the flexural rigidity of continental lithosphere vary considerably, partly because they are dependent on the adopted models of the original emplacement and subsequent evolution of the topographic loads. Thus Quinlan and Beaumont [45] estimate an equivalent elastic thickness of more than 60 km for eastern North America while Stephenson [46] has estimated only 25–30 km for

some of the older terrains of North America. This latter estimate is comparable with estimates for central and eastern Australia [20,47].

2.3. Finite element model

The finite element calculations yield profiles of the total deflections due to the sediment loading of the thermally evolving lithosphere, and of the effects on the deflection of regional in-plane tension or compression. A Choleski-decomposition scheme is used to solve the equilibrium equations for a two-dimensional elastic body assuming plane strain. Constant strain elements are inadequate for flexural analyses and instead we use an eight-node quadrilateral element in which strain varies linearly and the displacement field varies quadratically over the surface. The adopted finite element mesh has approximately 400 degrees of freedom and the results have been checked and confirmed (a) by comparison with analytical solutions for the case of a line load, (b) by convergence tests using a mesh with more than 2000 degrees of freedom, and (c) by an analysis of the internal reaction forces of the model.

To simulate the increasing effective elastic thickness of the oceanic lithosphere with age, the finite element mesh has to be modified at each time step of the computation. This is accomplished by freezing in the deformation and increasing the thickness of the elastic layer, at each step. Simultaneously, a load increment is added at each time step. Because the effect of aging is more pronounced in young lithosphere, the time steps are progressively increased such that the lithospheric thickness increases by 15–30% in each time step.

3. Results

3.1. Uniform elastic plate model

Fig. 2a illustrates the deflection through time of the lithosphere with an age-dependent equivalent elastic thickness loaded by sedimentary wedges whose age-dependent heights are prescribed by the reference model discussed above. With time, both the wavelength and amplitude of the flexure increase due to the combined effect of the thickening of the oceanic lithosphere with increasing age and the growth of the sediment load with time.

The effect on the deflection of adding a horizontal compression or tension is illustrated in Fig. 2b. The total deflection of the lithosphere is dominated by the sediment loading but of interest here is the modification of the deflection by the inplane force, particularly near the edges of the basin. The adopted nominal in-plane force is $5 \times 10^{12} \text{ N m}^{-1}$, corresponding to a stress of 2.3 kbar for a plate thickness appropriate to 30 Ma old lithosphere. A transition from a tension of $5 \times 10^{12} \text{ N m}^{-1}$ to a compression of the same magnitude, produces a net uplift (or apparent sealevel fall) of up to about 100 m at the edge of the basin (see Fig. 2c). The deflections at other stresses can be scaled in proportion to the magnitude of the applied horizontal stress, with a stress field of a few hundred bars corresponding to a deflection of about 10 m.

The dependence of deflection on age and rate of loading is summarized in Fig. 3. The differential uplift Δw , defined as the difference in deflection for the change in stress or in-plane force from tension to an equal magnitude compression is computed for the edge of the basin as a function of variations in the in-plane force and stress. Curves I–III illustrate the deflections for changes in stress from 1 to –1 kbar, from 2 to –2 kbar and from 3 to –3 kbar respectively, always with the same reference, time-dependent, sediment load (negative stress denotes compression). In these models the in-plane force increases for increasing lithospheric age because of the associated thickening. Curve IV illustrates the deflection, for the same reference load, for a change in the in-plane force from $5 \times 10^{12} \text{ N m}^{-1}$ to $-5 \times 10^{12} \text{ N m}^{-1}$, irrespective of lithospheric thickness.

Fig. 2. Deflections calculated for the uniform elastic plate models. Sign convention for deflections: uplift is positive, subsidence is negative. (a) The deflection of the lithosphere at various stages of its evolution (specified in Ma) under the influence of sediment loading according to the reference model. (b) Effect of variations in regional stress field on the deflection of a plate with a thickness corresponding to an age of 30 Ma. Solid line: deflection due to reference model sediment loading alone. Dashed and dotted lines show the effect on the basin shape of superimposing $5 \times 10^{12} \text{ N m}^{-1}$ axial compressional and tensional forces respectively (corresponding to a regional stress field of 2.3 kbar). (c) Differential subsidence or uplift (metres) from the deflections plotted in (b) caused by regional stress fields of $5 \times 10^{12} \text{ N m}^{-1}$ axial compression (dashed curve) and $5 \times 10^{12} \text{ N m}^{-1}$ axial tension (dotted curve).

Fig. 3. Differential uplift (metres) at basin edge (position *A* indicated in Fig. 2c) 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–III*): results for a fixed stress transition from 1, 2 and 3 kbar tension to 1, 2 and 3 kbar compression respectively, superimposed on sediment loading according to the reference model. Dotted curve (*IV*): results for a transition from 5×10^{12} N m⁻¹ tension to 5×10^{12} N m⁻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 (*V–VIII*): results for sediment loading models that are different from the reference model. Heights of the sedimentary wedges are specified in km.

On the basis of these calculations we conclude that the effectiveness of the in-plane force in producing deflections increases with increasing thickness of deposited sediments and decreases with the increasing age of the lithosphere. Hence the in-plane force is most effective at those young passive margins that are subject to rapid loading. It is least effective at old margins subjected to low rates of sediment loading. This is most clearly seen where a constant sediment load is placed on an oceanic lithosphere of increasing effective elastic thickness with age and where sediment loads greater than the reference model are placed on the lithosphere (curves *V–VIII*, Fig. 3).

Similar changes in subsidence occur within the basins, although the relative subsidence here, of

the order of a few hundred meters, is small in comparison with the total subsidence which is of the order of several kilometers.

3.2. Effect of lateral variations in elastic thickness across the continental margin

The vertical movements within the basins become more complex when the effective flexural thickness is varied across the continental margin according to the model illustrated in Fig. 1b. In particular, when an in-plane stress field is applied, a tilting of the crust is induced at the transition from oceanic to continental lithosphere [48]. This tilting is a consequence of the change in thickness of the layer that carries the in-plane force and occurs even in the absence of sediment loading. Fig. 4a illustrates the effect for a sediment starved margin with 30 Ma old oceanic lithosphere adjacent to continental lithosphere of either 50 or 25 km thickness. For the 50 km thick continental plate and a change in the horizontal force from 5×10^{12} N m⁻¹ tension to 5×10^{12} N m⁻¹ compression, the tectonic subsidence is about 100 m at the continental side of the margin and about 170 m of uplift occurs at the oceanic side of the margin. These tilts are reduced for the thinner continental plate and would change sign if the effective elastic thickness of the continental lithosphere was actually less than that of the oceanic lithosphere. The effect of therefore highly model dependent as well as being time dependent if stress relaxation occurs in the lower lithosphere.

4. Discussion

4.1. Magnitude of vertical displacements

The model calculations demonstrate that variations in relative sealevel of about 100 m can be caused by regional changes in the inplane stresses of the order of a few kilobars. 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 force is applied; that is, in the context of passive margin evolution, on the rate of sedimentation and on the response of the lithosphere to this sediment load. In the uniform elastic models, the mechanism is most effective when the loads are

Fig. 4. Differential subsidence (metres) due to variations in regional stress field for models with lateral variations in rheological structure across passive margins (Fig. 1b). Curves *I* and *II* show the effect on the basin edge of superimposing $5 \times 10^{12} \text{ N m}^{-1}$ axial compressional and tensional forces respectively. Results are given for two models with effective elastic thicknesses (EET_c) for continental lithosphere at 25 and 50 km. (a)

superimposed rapidly on young oceanic lithosphere where a change in stress regime can produce the apparent fluctuations in sealevel of the order of 100 m [1,8–11]. For the model with a variation in effective elastic thickness across the margin, the changes in stress regime can cause a tilting of the crust near the edge of the basin. This can also give rise to apparent sealevel changes of the order of 100 m and is superimposed on the preceding effect.

4.2. Evidence for kbar regional stress fields

Several independent studies of lithospheric deformation at 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 of a few kilobars. Ward [25], for example, examined the depths of oceanic earthquakes associated with the bending of the lithosphere prior to it being subducted. He found that the neutral surfaces in young lithosphere could be elevated by as much as 15 km above the corresponding neutral surfaces in old 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 of the order of a few kilobars [26,49]. The deformation of the lithosphere prior to subduction has also been interpreted in some instances as implying in-plane stresses of a few kilobars [49]. Broad basement undulations seen in seismic reflection profiles for the southern part of the Bay of Bengal have been interpreted in terms of deformation caused by compressive forces of the order of several kilobars [50]. Studies of the departures from isostatic equilibrium in several tectonic provinces within Australia have also lead to the postulation of in-plane stresses at the level of a few kilobars [20,27]. Studies of the formation of sedimentary basins also indicate horizontal tensional stresses of the order of a few kilobars [51,52].

Of more concern than the magnitude of these stress levels are the changes in magnitude and the time interval in which these changes can occur. An

Tilting effect calculated for a starved continental margin of 30 Ma. Position *C* as indicated in Fig. 1b. (b) Results for a 30 Ma old margin loaded with sediments according to the reference load model.

important factor in controlling stress in the lithosphere is the time-dependent behaviour at convergent margins. One example of this is illustrated by the onset of the Banda arc collision where a period during which the stress would have been controlled by slab pull associated with subduction of old oceanic lithosphere [53,54] was followed at about 5 Ma ago by a phase of net resistance [24] due to the arrival of the buoyant continental lithosphere at the subduction zone [55,56]. Simultaneously, substantial tectonic changes appear to have occurred at adjacent convergent margins of the Indo-Australian plate; examples are the rapid uplift of the highlands of New Guinea [57] and a reorganization of movements and deformation along the Solomon and New Hebrides boundaries (R.J. Musgrave, personal communication, 1984). A significant change in the stress state may therefore have occurred at about 5–6 Ma ago and models of the stress field within the plate set up by boundary forces [24,58] indicate that these changes in stress may propagate into the interiors of the plates. As such, they effect passive margins in the Indo-Australian plate, in particular those around Australia. A change from transgression to regression has in fact been noted in the Otway Basin of southern Australia [6] at this time, in late Miocene/early Pliocene. We also anticipate that the onset of the Himalayan collision [59], part of an early Cenozoic world-wide plate reorganization [60], induced significant changes in the regional stress within the plates on the time scale of a few million years. Both the Gippsland and Otway Basin show changes in relative sealevel in early Eocene time.

Such rapid changes in the stress field are not limited to the Indo-Australian plate. Megard et al. [61] documented at least four pulses of crustal compression during the late Cenozoic evolution of the Ayacucho intermontane basin in the Andes of Central Peru, phases that were separated by periods, in which the stresses relaxed or were possibly replaced by tensional stress. A more global plate reorganization, presumably with a concomitant change in stress state [62], occurred in the Pacific region at about 30 Ma ago when the Farallon plate broke up into the Nazca and Cocos plates [62] and the Galapagos spreading centre developed.

Less dramatic changes in stress regime, of a continuous rather than abrupt character, may oc-

cur when the age of the lithosphere as it enters the trench system—and consequently the age-dependent slab pull force [54]—changes with time [63,64].

If these stress reorganizations occur on a time scale of 10^6 years, and are sufficiently large to produce an apparent sealevel change of 100 m, then the sealevel change rates are of the order of $10 \text{ cm}/10^3 \text{ years}$. If they occur on a timescale of 10^7 years as may be more appropriate, then these rates are of the order of $1 \text{ cm}/10^3 \text{ years}$. Hence not only glacial fluctuations, as argued by Pitman and Golovchenko [14], may be capable of producing apparent sealevel changes in excess of 1 cm/1000 years as well as magnitudes of about 100 m. Various external geologic factors may contribute to the sharpness of the falls in apparent sealevel deduced by Vail et al. [1] from unconformities [13]. Some marine unconformities, however, can also be attributed to extremely low rates of deposition [65]. Therefore, although the tectonic model proposed here is in itself not characterized by any inherent asymmetry between the rise and fall in sealevel, this does not exclude asymmetry in its seismostratigraphic expression.

By its nature, the model can explain contemporaneous fluctuations in apparent sealevel in neighbouring basinal settings. The action of changing horizontal stresses will not be restricted to the passive margins but will also modify the vertical movements within intracratonic basins. As such this mechanism may provide a tectonic explanation for some of the observed correlations between the timing of the sealevel changes in oceanic and intracontinental regions noted by Sloss [66] (see also [2]).

The evidence for temporal changes in regional stress fields requires further examination before specific changes in the Vail et al. apparent sealevels can be associated with particular tectonic reorganizations. That horizontal stresses occur in the lithosphere, and that major changes occur at plate boundaries, indicates that the proposed mechanism plays an important role in understanding the apparent sealevel fluctuations proposed by Vail et al. and others.

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