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NUMERICAL MODELS FOR THE THERMO-MECHANICAL EVOLUTION OF ATLANTIC-TYPE CONTINENTAL MARGINS

S. A. P. L. CLOETINGH

Department of Theoretical Geophysics, University of Utrecht, P.O. Box 80.021, 3508 TA Utrecht, The Netherlands

AND

G. WISSE

Computing Centre, Delft University of Technology, P.O. Box 354, 2600 AJ Delft, The Netherlands

SUMMARY

We have developed a versatile finite element approach to analyse stresses and displacements induced in thermally thinned and thickened geological structures. Material and geometric non-linearities form an important aspect of the finite element simulation. The changes in geometry are incorporated by a multipoint constraint procedure. The method takes into account temperature-dependent variations in rheological properties of the medium, including the presence of a brittle–ductile transition. We illustrate the application of the method for specific geophysical problems associated with the temperature history of the earth's crust at continental margins of the Atlantic type.

1. INTRODUCTION

The earth's outer shell, the lithosphere, is divided into a finite number of plates. The boundaries of these plates are defined by earthquake zones. Since earthquakes are manifestations of deformation, the plates themselves may be reasonably taken to be rigid. Two important categories of plate boundaries are mid-oceanic ridges and subduction zones. The upper part of the lithosphere is formed by the earth's crust. Although the lower part of the lithosphere is itself a part of the earth's mantle, it is distinguishable from the mechanically weak part of the mantle below it, which is the substratum upon which the lithosphere rests. Thermal modelling of the oceanic lithosphere constrained by observations of heat flow and bathymetry of the world ocean floor characterizes the lower boundary of the lithosphere by an isotherm of 1200°C .^{10,29} This corresponds to a thickness of a few tens of kilometres in the young oceanic lithosphere to about one hundred kilometres in the oldest, which is 200 million years (henceforth m.y.) old.^{10,29} (See Figure 1.) New lithosphere is created by sea-floor spreading: mantle material welling up at mid-oceanic ridges is added to the plates on either side. At subduction zones, which generally coincide with deep oceanic trench systems, the oceanic lithosphere is recycled into the mantle. To initiate sea-floor spreading in an ocean such as the Red Sea or the present-day Atlantic, the adjacent continents must separate, or rift. Rifting is associated with rapid heating of the lithosphere. It is an essential stage in the formation of ocean–continent boundaries flanking such oceans. Note that these Atlantic-type continental margins are not plate boundaries and are, therefore, also known as passive continental margins. Rifting is followed by cooling as the ocean basin widens.¹⁸ Consequently, the thermo-mechanical properties of the lithosphere at an Atlantic-type continental margin vary considerably

Figure 1. Temperature profiles for oceanic lithosphere with ages varying between 10 and 200 m.y. The lower boundary of the lithosphere is defined by a temperature of 1200 °C

during its geological evolution.^{8,9,27} This is also true of the system of forces acting on the lithosphere: sedimentation and erosion at the earth's surface and the forces associated with the sea-floor spreading itself.^{8,9,30}

The interaction of time-dependent forces, changing geometry and the temperature-dependent rheology (Section 2) of the earth's lithosphere at ocean-continent boundaries and its implementation in finite element (f.e.) analysis (Section 3) is the subject of the present paper. After a description of general features of the f.e. models (Section 3.1) we concentrate on the thermo-mechanical aspects of the formation of Atlantic-type continental margins by rifting (Section 3.2) and on their subsequent evolution (Section 3.3). In the last-mentioned section, we investigate whether oceans can be closed by the transformation of Atlantic-type continental margins into subduction zones, or Pacific-type continental margins, also known as active continental margins. Such an analysis draws heavily on the capabilities of finite element procedures. Quantitative models of the thermo-mechanical evolution of the lithosphere at passive continental margins such as described in Section 3.4 are of increasing importance in evaluating their petroleum potential.²⁶ For a more elaborate description of the geophysical aspects of the work, the reader is referred to References 5-9 and 30.

2. THE INFLUENCE OF TEMPERATURE ON THE RHEOLOGY OF THE LITHOSPHERE

Recent experimental and theoretical work on the deformation of olivine, the dominant constituent of the oceanic lithosphere, has provided a framework for understanding the rheology of the

lithosphere.^{4,15} In particular, the role that creep processes plays has been established. Creep in lithospheric processes is one of steady-state, owing to the long geological timescale involved. One very important creep process is the thermally activated climb of dislocations. This process obeys a non-linear relation between stress and strain-rate:¹⁴

$$\dot{\epsilon} = \dot{\epsilon}_1 \sigma^3 \exp(-Q_1/RT) \quad (1)$$

with Q the activation energy, R the universal gas constant, σ the differential stress, $\dot{\epsilon}$ the strain-rate (s^{-1}), and T the temperature (K). $Q_1 = 122$ kcal/mole for dry olivine. For σ measured in bars (1 MPa = 10 bar), the value of the pre-exponential constant $\dot{\epsilon}_1$ is $70 \text{ bar}^{-3} s^{-1}$. Power-law creep holds for conditions of high temperatures and relatively low deviatoric stresses.²

The constitutive equation (1) certainly does not hold for the transition between the elastic and ductile regimes in the lithosphere, where stresses are high and temperatures relatively low. Under these stress and temperature conditions, dislocation glide is the dominant creep process. For dislocation glide, the velocity of the dislocations, or the strain-rate, is limited by obstacles in the gliding plane. A much better description of the deformation in this region is provided by the Dorn-type creep formulation developed by Goetze and his co-workers.^{14,15} The constitutive equation for Dorn-creep, which holds for stresses in excess of 2 kbar, is

$$\dot{\epsilon} = \dot{\epsilon}_2 \exp(-Q_2(1 - \sigma/\sigma_p)^2/RT) \quad (2)$$

where $\sigma_p = 85$ kbar and $Q_2 = 128$ kcal/mole for dry olivine. The value of the pre-exponential constant $\dot{\epsilon}_2$ is $5.7 \times 10^{11} s^{-1}$. Brittle strengths in the uppermost part of the lithosphere increase linearly with depth according to Byerlee's law.¹⁵ For a specific strain-rate, equations (1) and (2) can be used to calculate the transition stress between the elastic-brittle upper part and the lower ductile part as a function of temperature. The result is given in Figure 2(a), adopting a strain-rate of $\dot{\epsilon} = 10^{-18} s^{-1}$ (representative for sediment loading on the oceanic lithosphere). Using the constitutive equations (1) and (2) and the temperature profiles derived from Crough's^{10,29} model for the thermal evolution of the oceanic lithosphere, strength profiles can be calculated for the oceanic lithosphere with various ages (Figure 2(b)) as a function of depth. The response of the

Figure 2. Rheological model for the oceanic lithosphere based on data for olivine. Three regions are present in the mechanically strong part of the lithosphere (MSL). The top is characterized by brittle fracture, the central core is in the elastic regime and the lower part is deforming according to ductile flow. (a) The transition stress between the elastic upper part and the ductile lower part of the lithosphere as a function of temperature based on two different flow laws for olivine given by Goetze.¹⁴ A strain-rate of 10^{-18} s^{-1} is adopted. The lower boundary of the MSL is defined by a stress level of 0.5 kbar. (b) Strength profiles through the upper part of the oceanic lithosphere for ages of 30 and 100 m.y. derived from Figure 2(a) by specifying the temperature function inside the lithosphere according to the thermal models given in Figure 1. Differential stresses ($\sigma_H - \sigma_V$) are plotted against depth. σ_H and σ_V are horizontal and vertical principal stresses, respectively. Sign convention for the stresses: tension positive, compression negative. An order of magnitude variation in strain rate corresponds to only a 1–2 km change in the thickness of the MSL.

oceanic lithosphere with depth-dependent rheological properties to loading is equivalent to that of an elastic plate with a (thinner) effective elastic plate thickness which increases according to a square-root-of-age function owing to the cooling of the lithosphere.

The rheology of the continental lithosphere is more complex than that of the oceanic lithosphere owing to the presence of a thick crust and a more complicated geological evolution. Recent independent evidence from widely different fields,²³ however, suggests that the major rheological characteristics recognized for the ocean lithosphere apply also to the continental lithosphere. We adopt an average model for continental rheology that is based on laboratory studies of rocks of the crust (quartzite and diabase) and upper mantle (olivine). Three layers are present in the model: a

15 km upper crust, a 15 km lower crust and a 120 km lithospheric mantle. Power-law creep parameters for quartzite²⁴ are adopted to describe ductile flow in the upper crust. For the ductile rheology of the lower crust, we adopt power-law creep parameters for diabase.²⁴ Ductile flow in the lithospheric mantle is described as above by Goetze's¹⁴ flow laws for olivine.

3. MODEL CALCULATIONS AND RESULTS

3.1. General features of the models

The non-linear rheology is implemented in the thermo-mechanical models by specifying lithospheric strengths σ_y at the integration points of the elements. In our analysis we are primarily interested in the flexural deformation in the mechanically strong upper part of the lithosphere (henceforth MSL), the lower boundary of which is defined^{4,5} by a strength of 500 bar (see Figure 2). Stresses and deformation induced in the mechanically weak underlying structure are not of primary interest. Nor are we concerned with the question of whether strain-hardening in creep will take place. Therefore, for the sake of simplicity we assume perfect plasticity throughout the analysis.

Flexural analysis requires elements with a displacement function order that is at least quadratic. An element with a linear displacement function (constant strain element) is obviously inadequate. The use of constant strain elements in flexural analysis does result in severe stress discontinuities between adjacent elements, because stresses are not allowed to vary within the separate constant strain elements. We have used elements with a quadratic displacement function throughout the present study. The use of linear strain elements also permits an accurate representation of variations of material properties with depth, since material properties are allowed to vary linearly within each separate element. The flexure of the lithosphere under the influence of loading is counteracted by hydrostatic restoring forces. These are incorporated into the analysis by modifying the stiffness matrix at the basal nodes of the model. This is accomplished by the implementation of an elastic foundation in the model. The differential stresses induced by flexure are to be compared with differential strengths. In regions where bending stresses are large, the largest principal stress σ_1 is equal to the maximum horizontal bending stress σ_H . The minimum principal stress σ_3 coincides with the vertical gravitationally induced stress σ_v . The intermediate principal stress σ_2 is perpendicular to the plane strain section. On these grounds the Tresca yield criterion ($\sigma_H - \sigma_v \leq \sigma_y$) would be useful. As the Tresca criterion is characterized by discontinuities at its boundary, difficulties in formulating an associated flow rule arise.³¹ Consequently, in numerical practice, use of the Tresca criterion generally has to be discarded. We therefore adopt the Von Mises yield criterion for associated flow. The differences between the yield surfaces of the two criteria (see References 21 and 31) are sufficiently small to permit accurate analysis.

The plane strain equations were solved using a secant stiffness procedure.^{19,31} If the strain energy difference between the estimated state and the recovered state satisfies the chosen tolerance, no recycling takes place. During recycles, the recovered strains are used as estimated strains for the stiffness evaluation. Equilibrium is ensured by adding the residual load corrections from the previous increment to the load vector. In non-linear analysis, the steps must be sufficiently close to ensure that these corrections are small in relation to the external forces. This prevents large deviations from the actual load-displacement curve.

Finite element networks with up to 3216 degrees of freedom are employed in our analysis. Analytical test solutions are not available for the deformation of a plate with a depth-dependent non-linear elastic-plastic rheology under the influence of distributed loads. Therefore, the accuracy of the model calculations presented in this paper was checked by convergence tests, made

Figure 3. Results of a convergence test on the coarsest model used in the analysis of Section 3.3 (mesh 1). Strength and the results of stress calculations are given as a function of depth (km). The lines inside the strength envelope, connecting the dots, are the stress distributions. Zero strength has been assumed for sediment lying on top of the lithosphere. Hatched areas in the upper and lower parts of the MSL denote failure by brittle fracture and ductile flow respectively.

by refining the finite element meshes. An example of such a test is given in Figure 3. This figure gives a characteristic lithospheric stress distribution induced by bending processes. Stresses are displayed as a function of depth down to the base of the MSL. Failure occurs by brittle fracture in the uppermost part and by ductile flow in the lowermost region of the MSL; the main part is in the elastic state. A comparison of the results of two model calculations carried out with meshes differing by a factor two in the number of elements shows that the mesh spacing (mesh 1) used throughout the work was sufficient for accurate analysis. Finally, the reaction forces of the model were checked to ensure equilibrium of external and internal forces.

3.2. Formation of an Atlantic-type continental margin

Doming of the lithosphere is caused by the effect of temperature perturbations on the density stratification of the upper mantle. In our models we incorporate the influence of the temperature perturbations on both the driving force for uplift and the temperature distribution and thus on the rheology of the overlying lithosphere.

The finite element model for doming is given in Figure 4. Uplift forces acting at the centre of the

Figure 4. The finite element mesh for the doming of thermally thinned lithosphere. On account of symmetry only half of the structure is incorporated in the model. At the left-hand boundary, only vertical displacements are allowed. The uplift forces are indicated by vertical arrows. Buoyancy forces counteracting the deflection are indicated by springs. The solid line is the strength envelope for a temperature-dependent rheology of the lithosphere including a brittle-ductile transition. Note the difference in the thickness and strength of the MSL inside and outside the area of uplift.

Figure 5. The uplift required to induce failure in the overlying lithosphere. 1 and 2 denote 100 m.y. old oceanic lithosphere reheated to a solidus at depths of 50 km and 35 km respectively. 3 denotes continental lithosphere thinned to 50 km. Included in the figure are geological estimates of the geometry of the uplift for the Baikal rift (BR), the Rhine graben (RG),²⁸ the Ethiopian rift zone (ER), and the Hoggar Swell (HS), Darfur Swell (DS), Cape Verde Rise (CR), and Hawaii Swell (HS).¹¹

structure are indicated by vertical arrows. We have calculated the mechanical properties of the rheologically thinned oceanic lithosphere using temperature profiles¹² for Hawaii, where many geological and geophysical data on uplift have been collected. Rheological profiles for the thinned continental lithosphere are calculated using geotherms for the thinned lithosphere of the well-documented Basin and Range province in western North America.¹⁶ The results of the calculations are summarized in Figure 5. As a result of its rheological layering, the effect of a temperature perturbation on the continental lithosphere is more severe than the effect on the rheology of the oceanic lithosphere. Strong temperature perturbations will create minima in the lithospheric strength distribution near the boundary between the upper and lower crusts, and near that between the lower crustal and mantle rocks. A temperature perturbation will produce the greatest shift in the position of the brittle-ductile transition for rocks with a low melting point. Owing to this effect, the amount of uplift required to induce failure in the thinned continental lithosphere is actually less than that necessary to generate failure in the oceanic lithosphere.

Depending on the width of the dome, uplifts of 2–3 km are capable of inducing lithospheric failure. A comparison of the results of our model calculations with continental uplift data provides an explanation for the observation¹¹ that relatively few thermally induced domes generate new rifts and the formation of new continental margins of the Atlantic type.

3.3. Thermo-mechanical evolution of Atlantic-type continental margins

Growth of the oceanic lithosphere at Atlantic-type continental margins and sediment loading usually takes place simultaneously.^{8,9,27} A one-step iterative solution is, therefore, not applicable. To simulate the increasing thickness of the lithosphere, at each time a layer of elements must be added at its lower boundary. Therefore, a compromise must be made between element size and time step length. The width of the elements is 25 km. For the oceanic lithosphere, element thicknesses range from 10 to 5 km, corresponding with time steps of 2.5 m.y. in the initial cooling phase, to 20 m.y. at the final cooling phase (see Figure 6). Note, however, that the complete mesh of the structure is generated at the start of the calculations. Elements that at a given time step are not physically part of the structure are assigned values for Young's modulus reduced by a factor of 100. For the case of the lithosphere floating on a weak substratum, we tie the elastic foundation (or springs) to the layer of elements at the lower boundary of the MSL. Thus, at each time step the tying is shifted to the base of the newly formed layer of mechanically strong elements. The number of mechanically weak elements is reduced and a load increment is added simultaneously with the generation of new strength profiles. The average bandwidth of the stiffness matrix is 99. The finite element calculations were carried out with a modified version of the Marc¹⁹ program. The corresponding cpu times per iteration step within an increment on an Amdahl V470 for the assembly of the stiffness matrix, the matrix solution and the stress recovery are 100, 200 and 70 s, respectively. The complete analysis described in this section required approximately 6 hours cpu time on an Amdahl V470. Models are constructed with ocean widths varying between 100 and 4000 km. Adopting a spreading-rate of 2 cm/year (as documented for the Atlantic Ocean¹⁸), an increase in the width of the ocean of 200 km corresponds with an increase in the age of the oceanic lithosphere at the adjacent boundary of 10 m.y. Rheological profiles in the transition between the

Figure 6. Central part of the f.e. mesh employed in the analysis of stresses induced in cooling lithosphere at ocean-continent boundaries. Layers of elements are added incrementally to the MSL whose lower boundary is indicated at various time steps (specified in m.y.). A comparison of the lower boundaries at 12.5 m.y. and 20 m.y. (given by dashed lines) illustrates the growth of the MSL in the horizontal and vertical directions. The height and width of the load on the oceanic and thinned continental (so-called rift-stage) lithospheres are given in km at some selected time steps (specified in m.y.).

oceanic and continental lithospheres have been constructed by linear interpolation between the rheologies of the oceanic and continental lithospheres. Extensive sedimentary deposits at ocean-continent boundaries result in flexural stresses in the underlying lithosphere. Two models for sediment loading are considered in the analysis: the reference model, in which sediment loading keeps up with the thermal subsidence of the underlying lithosphere (with sediment thicknesses growing to 9.4 km at an age of 200 m.y.), and the full load model, in which the full loading capacity of the oceanic lithosphere is taken up by sediment loading at the ocean-continent boundary (implying sediment thicknesses up to 16 km). Sedimentary wedges with widths of 150–250 km are incorporated in the models (see Figure 6). Large thermal stresses²⁰ (order of magnitude tens of kilobars) can be generated by lithospheric cooling following creation at spreading centres. However, such high stresses are probably relaxed near spreading centres very early in the evolution of the lithosphere.^{8,9} Therefore, we do not incorporate thermal stresses in the models.

Differential stresses induced by sediment loading are largest at the points of maximum flexure. The greatest stress maximum is located in the oceanic lithosphere close to the ocean-continent boundary. In Figure 3 the stress maximum for 100 m.y. is displayed for the reference load as a function of depth, down to the base of the MSL. For the reference model of sediment loading, stresses remain too low to result in rupture of the lithosphere. The ratio of the maximum stress generated and the maximum strength is essentially independent of age, since load, strength and thickness of the MSL all exhibit the same square-root-of-age behaviour owing to the thermal cooling of the lithosphere.

The situation is drastically different when the full loading capacity is taken up by sediment. The surplus of sediments added to the reference load will promote high stresses most effectively when deposited on a young (weak) continental margin. Full loading on Atlantic-type continental margins with ages below 20 m.y. leads to complete failure of the lithosphere. For ages in excess of 20 m.y. the relative amount of failure decreases markedly with age (Figure 7).

Hence, in general, external forces are required to induce failure in the oceanic lithosphere at continental margins of the Atlantic type.

Figure 7. Ratio of A_y (the hatched area within the strength envelope; see Figure 3) to A (total area of the envelope) as a function of lithospheric age for the full load model. For ages below 20 m.y., lithospheric failure and consequently conversion into an active margin is induced.

3.4. Stress fields induced by plate-tectonic forces: the effects upon Atlantic-type margin evolution

3.4.1. *Stress fields caused by plate-tectonic forces.* Several independent studies of lithospheric deformation in the active continental margin and intraplate tectonic settings have led to the conclusion that horizontal stresses exist in the lithosphere and that these stresses may reach magnitudes of up to a few kbars.¹⁷ Recent numerical modelling of stresses induced by plate-tectonic forces in the lithosphere^{30,7} 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 resulting 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 using the resulting stress field to analyse, explain and even predict various deformational processes in lithospheric plates.

One of the plates that has been investigated by us is the Indo-Australian plate.⁷ Figure 8 shows the finite element stress used in the analysis. We have approximated the spherical surface of the Indo-Australian plate by an assembly of 485 triangular membrane elements with a quadratic displacement field. The number of degrees of freedom is 2104. The maximum size of the elements is five degrees. We ignore vertical stresses in the plane stress approximation adopted here and calculate the horizontal stress field in the lithosphere which is integrated over the lithospheric

Figure 8. Finite element mesh of the Indo-Australian plate. The spherical surface of the plate is approximated by an assembly of 485 triangular membrane elements with a quadratic displacement field

thickness. The resulting field represents the non-lithostatic state of stress in the lithosphere. The outcome of our stress calculations can be easily related to estimates of the integrated depth-dependent lithospheric strength that follow from studies of the rheological properties of the lithosphere discussed in Section 2. In the present section we display stress values derived (from the values integrated over the lithospheric thickness) for an elastic plate with uniform elastic properties and a thickness of 100 km. We emphasize that this is a reference value only. The calculations of the stress field in the Indo-Australian plate are made with the Aska package of finite element routines.¹ Figure 9 illustrates the quality of the approximation of the spherical shell by the assembly of flat elements. Our modelling of the stress field in the Indo-Australian plate demonstrated that the exceptionally high level of compressive deformation in the plate's interior³ is a transient feature unique to the present-day dynamic situation of the Indo-Australian plate. We showed that stresses induced by slab pull forces associated with subduction of the oceanic lithosphere at various segments of the convergent boundaries of the Indo-Australian plate are transmitted into the interior of the plate, where they affect the lithospheres of both oceanic and continental character and Atlantic-type continental margins. The high level of intraplate seismicity in the Indo-Australian plate makes intraplate earthquakes a reliable indicator of the intraplate stress field. Stress directions inferred from earthquake source mechanism studies³ given in Figure 10 show a rotation of the stress field in the Northeastern Indian Ocean from a N-S orientated 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. Furthermore, the stress field as calculated gives a consistent explanation for the observed concentration of seismic activity and significant deformation in the oceanic crust¹³ in the northern Indian Ocean segment of the Indo-Australian plate.

Evidence from earthquake source mechanism studies indicates that large parts of the Australian continent are in a state of horizontal compression, with a magnitude of the order of 1–2 kbar¹⁷ for the stress field induced by plate-tectonic forces. As shown in Figure 11, the observed principal stress orientations are in different directions in different regions of the continent. In essence, the stress field varies from E-W compression in SW Australia to N-S compression in central Australia. In SE Australia, a more complex state of stress has been observed.¹⁷ Our modelling shows that the rotation of the stress field in the Australian continent is mainly the consequence of its

Figure 9. Illustration of the quality of the approximation of the spherical shell by the assembly of flat elements

Figure 10. Intraplate stress field in the northeastern Indian Ocean. (a)—left. Calculated stress field (after Reference 7). Principal horizontal non-lithostatic stresses averaged over a uniform elastic plate with a reference thickness of 100 km are plotted. Symbols ($\rightarrow$) and ($\leftarrow$) 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.¹³ (b)—right. The orientation of maximum horizontal compressive stress inferred from an earthquake source mechanism study by Bergman and Solomon³

geographic position relative to the surrounding active continental margins and the variations of the forces acting on the downgoing lithospheres in each. Significant variations in the stress field caused by plate-tectonic forces occur along the Australian continental margins of the Atlantic type. This is especially evident along the eastern Australian margin. There the calculated stress field is roughly perpendicular to the margin varying along the margin from NW–SE tension in the north to NE–SW tension and finally to E–W compression in the south.

In the following, we investigate the effect of the superposition of fluctuations in the stress field associated with plate-tectonic forces and the flexure induced by sediment loading on the passive margins and discuss the consequences for the subsidence history of the lithosphere at Atlantic-type continental margins.

3.4.2. The effect of stresses caused by plate-tectonic forces on the deflection of the lithosphere at Atlantic-type margins. In this section, we concentrate on the subsidence of sedimentary basins located on Atlantic-type continental margins. Fluctuations in sea-level are recorded in the sedimentary deposits at the edge of passive margins. As such, they can be recognized from seismic reflection profiles of the layering of sediments at passive margins. Therefore, sea-level fluctuations provide

Figure 11. Stress field induced by plate-tectonic forces in the Australian continent and peripheral areas. (a)—top: calculated stress field, after Reference 7. (b)—bottom: stress orientation data from earthquake source mechanism studies.¹⁷

Hatched areas mark the region with recent basaltic volcanism in northern Queensland and Victoria

important diagnostic tool is commercial hydrocarbon exploration at Atlantic-type continental margins. A major contribution to hydrocarbon exploration of passive margins was made by Vail and co-workers.²⁵ These authors produced a set of curves for sea-level fluctuations at rates of 1–10 cm/1000 years and magnitudes up to 150 m based on Exxon borehole and seismic reflection data from sedimentary basins located in passive margins in different parts of the world. Since the publication of Vail *et al.*'s²⁵ cycles of relative changes in sea-level, there has been a continuous debate on the mechanisms responsible for these fluctuations.²² Pitman and Golovchenko²² 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 m. These cannot, however, explain those major parts of the sea-level curve where there is no geological evidence for significant glacial events. Alternatively the sea-level fluctuations were attributed to tectonic processes, although the nature of this mechanism for explaining rapid short-term fluctuations remained obscure.^{22,27}

Figure 12. Tectonic model from apparent sea-level fluctuations: variations in the stress field caused by plate-tectonic forces affect the vertical displacements at a passive margin evolving through time owing to the thermal evolution of the lithosphere and loading of this lithosphere by a wedge of sediments. Rheological differences between continental and oceanic lithospheres are neglected for simplicity

In the present section we will show that consideration of lithosphere mechanics sheds new light on this problem. Of interest here is the modification of the shape of the lithosphere underlying sedimentary basins at Atlantic-type continental margins. Recently, Cloetingh *et al.*⁵ have investigated the interaction between stresses induced by plate-tectonic forces and the deflection of the lithosphere caused by sedimentary loading (Figure 12). Figure 13 shows the effect of this interaction for a sedimentary basin underlain by a 30 m.y. old oceanic lithosphere with a corresponding elastic plate thickness. A transition from an externally applied tension of 1 kbar to a compression of the same magnitude produces a net uplift (or apparent sea-level fall) of up to about 40 m 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 an external stress field of a few hundred bars corresponding to a deflection of about 10 m. The dependence of deflection on the age of the underlying lithosphere is demonstrated in Figure 13. The differential uplift, Δw , defined as the difference in deflection for the change in stress from tension to an equal magnitude compression is computed for the edge of the basin as a function of variations in the intraplate stress. Curves illustrate the deflections for changes from 2 to -2 kbar, from 1 to -1 kbar, from 0.5 to -0.5 kbar and from 0.25 to -0.25 kbar, respectively, always with the same reference, time-dependent sediment load (negative stress denotes compression).

Similar changes in subsidence occur within the basins, although the relative subsidence caused by an external stress field is small in comparison with the total subsidence which is of the order of several kilometres. The estimates of the magnitude of the vertical displacement inferred from our modelling of the deflection of a uniform elastic lithosphere are conservative. Introduction of a depth-dependent rheology¹⁵ in the modelling would enhance the effectiveness of the action of stress variations.

Figure 13 demonstrates that variations in relative sea-level of at least 10 m can be caused by regional changes in the in-plane stresses of the order of a few hundred bars. If these stress reorganizations occur at a time scale of 10^6 years, then the associated sea-level change rates are of the order of at least 1 cm/1000 years. The actual magnitude obtained for a given change in stress is

Figure 13. Differential uplift (metres) at basin edge (position A indicated in inset) due to superposition of variations in the stress field caused by plate-tectonic forces on flexure caused by sediment loading plotted as a function of the age of the underlying lithosphere. Results are given for a fixed stress transition from 2, 1, 0.5 and 0.25 kbar tension to 2, 1, 0.5 and 0.25 kbar compression respectively. Inset: differential subsidence or uplift (metres) from the deflection due to sediment loading and thermal contraction of 30 m.y. old oceanic lithosphere caused by an external stress field of 1 kbar compression (dashed curve) and tension (dotted curve). Sign convention for deflections: uplift is positive, subsidence is negative

controlled by the magnitude of the perturbation or deflection of the lithosphere at the time that the in-plane stress 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.

Stress fluctuations in the lithosphere of a few hundred bar are sufficient to explain most of the fluctuations in sea-level shown in the Vail *et al.*²⁵ curves. To induce apparent sea-level fluctuations with magnitudes greater than 50 m, changes in stress level of more than 1 kbar are required, which must be related to major stress reorganizations at active continental margins, fragmentation of lithospheric plates or continental collision processes.⁵ By its nature, the mechanical model implies the possibility of regional correlation of seismic reflection data in different sedimentary basin settings. Specific rapid fluctuations in the Vail *et al.*²⁵ sea-level curves can now be associated quantitatively with particular plate-tectonic reorganizations of lithospheric stress fields. Seismic reflection data from Atlantic-type continental margins could provide a new source of information for the lithospheric stress field in the geological past. Examination of commercial seismic reflection data in a wide range of sedimentary basins in connection with independent finite element

modelling is required to fully exploit the potentials of this new avenue of research in quantitative studies of Atlantic-type continental margins.

4. EVALUATION AND DISCUSSION

The finite element approach followed here is a powerful tool in analysing thermo-mechanical problems. The method has been shown to be effective in stress analysis of geological structures with a temperature-dependent rheology. An alternative method for the numerical investigation of problems associated with thermal thickening or thinning of the lithosphere is rezoning. In the cases analysed in Section 3.3, stresses change smoothly with time in the elements in the upper part of the structure. Therefore, regeneration of the mesh would allow a reduction of the number of elements in this part of the model. At the lower part of the mesh, however, the rezoning technique is inadequate, for it would smooth out the stress peaks, by a repeated extrapolation, averaging and interpolating. Moreover, the addition of stress-free material at the bottom of the structure is often difficult to model with this technique, as the extrapolation of stresses and strains does not generally result in a locally stress-free medium.

It is, therefore, to be expected that the alternative multipoint constraint procedure employed in our analysis will be useful in modelling a wide range of thermo-mechanical problems in geomechanics. Other fields of application may be found in the analysis of structures where growth is not taking place smoothly (e.g. in metal foundry), or where large pieces of material are added to or removed from the structure.

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