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Numerical modelling of regional neotectonic movements in the northern Black Sea

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Abstract

The paper examines the manifestation of tectonic processes related to the Neogene and recent subsidence of the Black Sea margin and uplift of the Crimea mountains as well as current seismicity in the southern Crimea. A 2-D finite-element model, based on a seismic line across the northern Black Sea and the Crimean mountains, that predicts the present-day tectonic velocity field, using the available neotectonic data as boundary conditions, has been developed. The model results indicate that the observed structure and tectonic motions cannot be explained by forces imposed on the external side, boundaries of the model only. Rather, movements consisting of upwelling beneath the Black Sea and subsidence under the Crimea, interpreted as the effects of mantle processes, are necessary to explain the observed topographic evolution of the area. Zones where the ratio of the calculated shear stress to calculated pressure is greatest correspond to the position of the Southern Coast fault that controls the seismicity of the northern Black Sea margin.

Keywords: numerical modelling; state of stress; neotectonic movements; Black Sea

1. Introduction

The margin of the Black Sea along the southern edge of the Crimean Peninsula (Fig. 1) is characterised by significant recent onshore uplift (estimated to be up to ~2 mm/yr; Nikolaev, 1977), tectonic subsidence of the offshore sedimentary basin, and a zone of high seismicity associated with the Southern Crimean fault (SCF; Fig. 1). The geodynamic processes responsible must be intimately related to the present state of stress in the nearby lithosphere and it follows that, if the state of stress could be inferred from the neotectonic observations, the nature of the driving mechanisms may be revealed.

The calculation of the stress field within any specific region of the lithosphere requires a knowledge of the local structure — i.e., the distribution of density and rheological parameters, inferred from geological, geophysical and laboratory data — and the imposition of a set of boundary conditions, inferred from the regional tectonics of the area (e.g., Zoback, 1992). Numerical modelling studies of tectonic structures of different type and genesis has revealed, however, that the boundary conditions can strongly affect the computed stress and strain patterns (Mikhailov et al., 1991). Thus, the selection of appropriate boundary conditions is the key prob-

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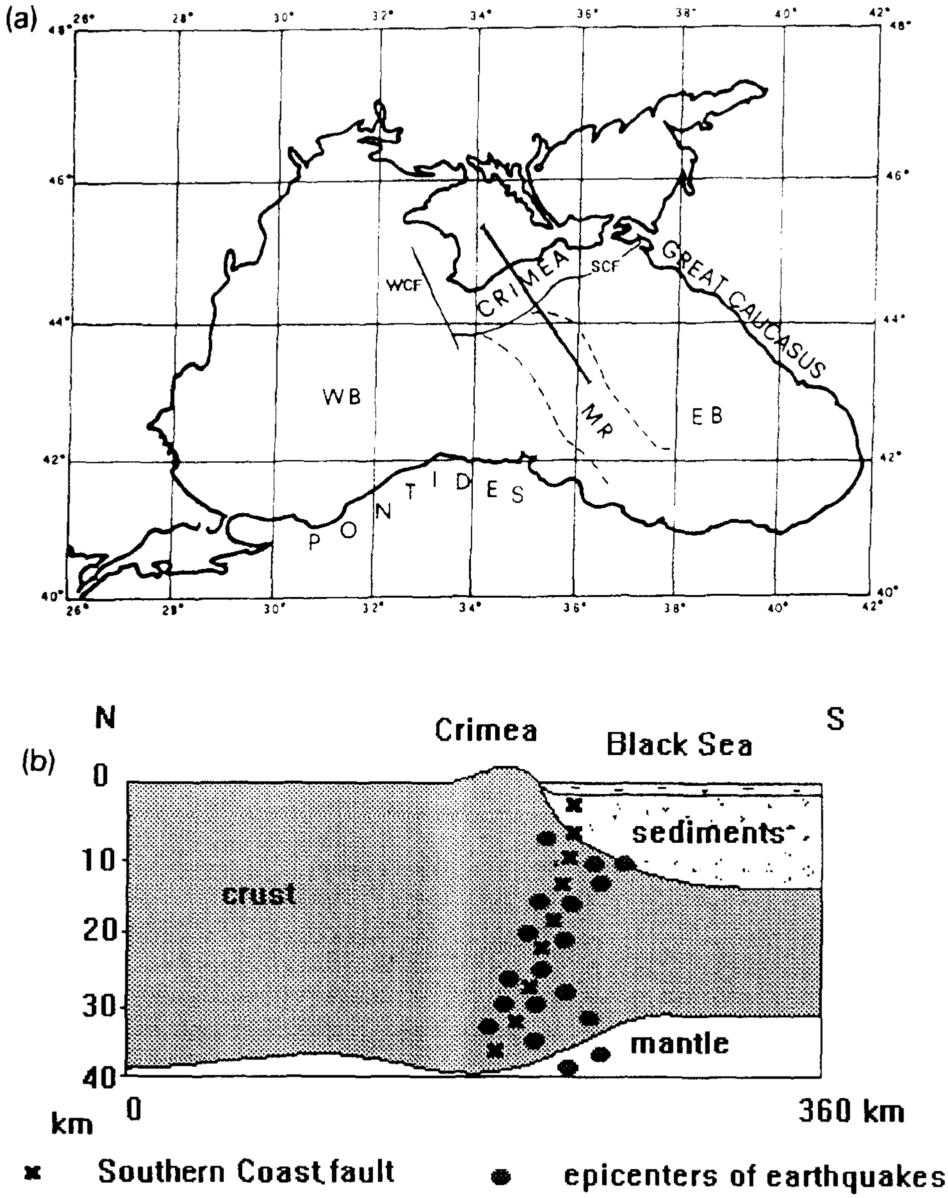

Fig. 1. Location of the study area; solid black line shows the position of the modelled crustal cross-section (Fig. 2). *SCF* = South Coast fault; *WCF* = West Coast fault; *WB* = western Black Sea Basin; *EB* = eastern Black Sea basin; *MR* = mid-Black Sea ridge.

lem when modelling the stress in any region of the lithosphere.

In this paper a method is presented, and applied to the Crimean margin of the Black Sea, in which the initial assumptions are validated by comparing the calculated value of the vertical velocity vector component at the surface with the topographic evolution trend derived from the observed neotectonic data. In so doing an attempt is made to simulate the complex present-day forces that can simultaneously support the uprising tendency of the Crimea mountains, subsidence of the offshore depression and explain the seismicity of this region.

2. Origin of the Black Sea

The Black Sea is an elliptical basin in the south-eastern part of Europe (Fig. 1) consisting of two depressions with oceanic crust separated by the Mid-Black Sea ridge which has a thinned continental crust (Letouzey et al., 1977; Tugolesov, 1985; Finetti et al., 1988). Maximum water depth is 2206 m (Ross et al., 1974). The thickness of probably Cretaceous to Holocene sediment cover in the western depression is more than 14 km and in the eastern one about 12 km. The tectonic processes responsible for the origin and evolution of the Black Sea are still debated but the hypotheses advanced fall into two basic categories.

One has been inspired by plate tectonic theory. Dewey et al. (1973), for example, consider the Black Sea as a remnant of a 'Tethyan' ocean separating Gondwana and Laurasia. Other authors suggest that the Black Sea represents the remnant of a back-arc marginal basin formed in late Mesozoic–early Tertiary times as part of the European continental paleo-margin behind the Balkan–Pontide–Lesser Caucasus foldbelt (Boccaletti et al., 1974; Adamia et al., 1977; Letouzey et al., 1977; Zonenshain and Le Pichon, 1986; Finetti et al., 1988). The results of simulation of syn-rift and post-rift stratigraphies and subsidence of Black Sea are presented in Robinson et al. (1995) and Spadini et al. (1996). The main differences of interpretation are in defining the age of the opening of the basin system. The authors agree, however, that the opening occurred after the closure of the Paleo-Tethys. The number, position and size of the original Neo-Tethys branches vary in the authors' interpretations depending on the results of various palinspastic reconstructions of the Jurassic and Cretaceous (Biju-Duval et al., 1977; Robertson and Dixon, 1984; Dercourt et al., 1986).

A second set of hypotheses considers the Black Sea as an area of oceanisation, i.e., replacement of continental crust with oceanic type crust following major vertical movements within the mantle (e.g., Muratov, 1972; Brinkmann, 1974; Beloussov et al., 1988). Opinions of the authors about the age of formation of the Black Sea vary considerably — from Jurassic or earlier (Muratov, 1972) to Paleocene (Tugolesov et al., 1985. Beloussov et al. (1988) considered that four major subsidence episodes took place during the evolution of the Black Sea: in the Late Cretaceous, associated with the formation of the deep Black Sea basin, and then in the Paleogene, pre-Pliocene and the Late Quaternary.

All the hypotheses about the origin of the Black Sea agree that the western and eastern sub-basins have different subsidence histories but that neotectonic evolution is related to the present-day regional tectonic setting. Beloussov et al. (1988) concluded on the basis of reflection seismic data (Tugolesov, 1985) that the Black Sea depression has subsided as a unit since 30 Ma. However, the high level of seismic activity in the southern Crimea indicates that tectonism remains active in the eastern sub-basin. Thus, two different geodynamic processes are likely responsible for the tectonic forces affecting the study area (the Crimean margin) during the neotectonic time frame of interest: those driving the subsidence of the Black Sea as a unit and those, presumably acting since the Eocene (reference) delimited by the West Crimean fault (Finetti et al., 1988; Okay et al., 1994).

3. Geological/geophysical cross-section

Fig. 1 shows the location of the simplified crustal cross-section (Fig. 2) along which stresses have been calculated. It is based on geological and geophysical data published by Subbotin (1975), Tugolesov (1985) and Beloussov et al. (1988) and is perpendicular to the trends of the Crimea mountains and the southern coast of the Crimean Peninsula. Also shown in Fig. 2 is the average vertical component of recent surface velocity — clearly demonstrating subsidence in the offshore depression and uplift in the Crimea mountains — derived from neotectonic data (Nikolaev, 1977) as well as data on recent movements (Meshersky, 1987) and on tectonic subsidence in the Black Sea (Tugolesov, 1985).

Artemjev (1975) concluded, given constraints on the thickness of the sedimentary and crustal layers from seismic data, that it was not possible to obtain a density distribution compatible with the observed seismic velocities leading to a state of isostatic equilibrium for the Black Sea and the Crimea mountains. Rather, a number of gravity models of this area, differing in their level of detail, all lead to the inference of positive isostatic anomalies in the Crimea region

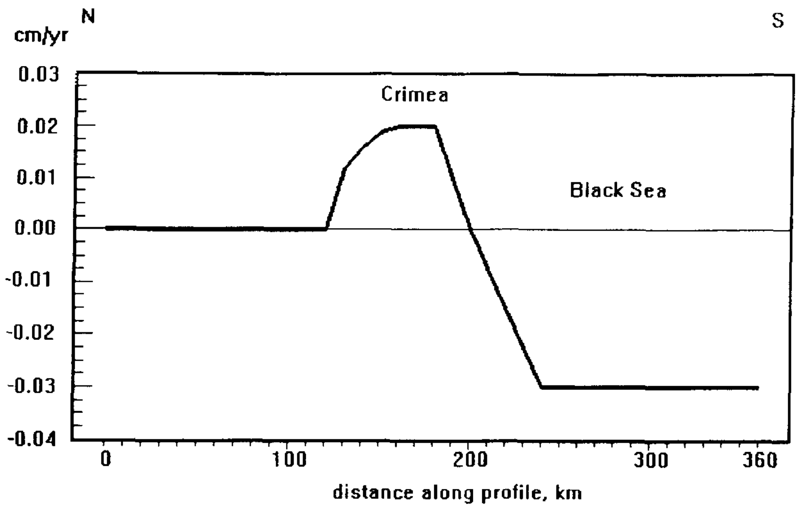

Fig. 2. Simplified crustal cross-section (located in Fig. 1), with the position of the Southern Coast fault and associated earthquake epicentres, used for the numerical modelling study and, above, the averaged curve of vertical neotectonic velocities along the cross-section (Nikolaev, 1977; Meshersky, 1987; Tugolesov, 1985) showing subsidence in the Black Sea depression and uplift in the Crimea mountains.

and negative ones in the Black Sea (Artemjev et al., 1972; Avdulov, 1979; Burianov et al., 1979).

Thus, the observed neotectonic data (Fig. 2) cannot be explained by isostatic responses, in which case, according to the isostatic anomalies the vertical displacements would be opposite to those observed. Furthermore, that there are no differences in the Neogene and younger subsidence rates of the eastern and the western parts of the Black Sea and of the Mid-Black Sea ridge, suggests that the offshore subsidence is not due to a post-extensional thermal subsidence mechanism only. It is deduced, therefore, that the observed neotectonic movements are at least in part driven by other tectonic processes. In order to isolate these for modelling, however, the observations have been adjusted in Fig. 2 for the expected isostatic effects of erosion, onshore, and sedimentation offshore, including the potential thermal subsidence contributing to the subsidence of the sedimentary basin.

4. Model theory

The rheologies used to describe the deformation of the lithosphere were those of a viscous fluid and an elastic continuum. Qualitatively, the results obtained for both cases are similar and, since the distance and time scales of the deformation processes in the study are large, the viscous fluid model will only be discussed further.

Ignoring the inertial terms, and assuming incompressibility of the material and body forces other than those of gravity, Stokes equations that can be written (for this stationary 2-D case) as:

$$\frac{\partial P}{\partial x_\alpha} + \rho\frac{\partial \varphi}{\partial x_\alpha} = \frac{\partial}{\partial x_\alpha}\left[\eta\left(\frac{\partial u_\alpha}{\partial x_\beta} + \frac{\partial u_\beta}{\partial x_\alpha}\right)\right],$$
$$\alpha,\ \beta = 1,\ 2 \qquad (1)$$

where $u_1 = u$, $u_2 = w$ are the components of velocity vector $\bar{u}$ of movements in the directions $x_1 = x$ and $x_2 = z$ (x is horizontal, z is upward), $P(x, z)$ is pressure, φ is gravity potential, $\partial\varphi/\partial x_\alpha = \{0, -g\}$ is the mass force vector, and $\eta(x, z)$ is viscosity.

For an incompressible fluid it is also true that:

$$div\,\bar{u} = 0 \qquad (2)$$

The boundary conditions to determine $u(x, z)$ and $w(x, z)$ are that stresses at the top of the model vanish, written as:

$$\sigma_{i,j} n_i = 0 \qquad (3)$$

where $\sigma_{i,j}$ is the stress tensor and n_i represents the components of a normal vector; that at the base of the model:

$$u = f(x)$$

$$w = -f'(x)(z-h) \tag{4}$$

where $f(x)$ is the desired function; and that on the vertical boundaries:

$$u = f(x_0)$$

$$w = -f'(x_0)(z-h) \tag{5}$$

$$u = f(x_1)$$

$$w = -f'(x_1)(z-h) \tag{6}$$

where (Fig. 3a) x_0 and x_1 are the positions of the left-hand and right-hand vertical boundaries, and h is the so called free mantle or floating level.

The equation used in Eqs. 4–6 to determine the desired function $f(x)$ can be obtained in different ways (cf. Myasnikov et al., 1993). It encompasses the properties of the employed physical model. For example, assuming homogeneous extension in the lithosphere, $f(x) = ax + b$, horizontal topography, homogeneous densities for the layers, and an absence of viscous flow leads to extension similar to that predicted by the stretching model of McKenzie (1978). When given the horizontal velocity at the bottom of the model, $f(x)$, the problem is complete.

In the present case the problem to be solved, however, is to determine the function from the condition that the velocity of movements at the surface of the model should fit observed values:

$$W(x) = w\,[x, \zeta(x)] \tag{7}$$

where $\zeta(x)$ is the topography of the surface (or sea bottom).

To satisfy the boundary conditions (4)–(6) and be consistent with the governing Eqs. 1, 2, it is essential that the height of the study area, H, should be much less than its length, L, i.e., $H/L = \epsilon \ll 1$, and that the top surface topography slope only gently, i.e., $\partial\zeta/\partial x \approx \epsilon$.

The numerical scheme used is based on the finite element method (Zienkiewicz, 1977). The function $f(x)$ is taken to be:

$$f(x) = \sum_{n=1}^{N} C_n f_n(x) \tag{8}$$

where $f_n(x)$ is a set of linearly independent functions, in the present case, represented by triangles of the unit height with the base $l = 2L/(N-1)$ with N being the number of triangles (Fig. 3b), thus ensuring that the problem is linear with respect to the coefficients C_n. For each $f_n(x)$; $(n = 1, 2, \ldots, N)$, $u(x, z)$ and $w(x, z)$ are determined under the condition $f(x) = f_n(x)$ giving the distribution of vertical surface velocity $w^{(n)}(x)$ corresponding to every unit function. $f_n(x)$

The problem of least squares approximation for the data given on a mesh x_k; $(k = 1, 2, \ldots, M)$ can be written as:

$$\sum_{k=1}^{M} \left[W_k - \sum_{n=1}^{N} C_n w_k^{(n)} \right]^2 = \min \tag{9}$$

where M is the number of nodes in the horizontal di-

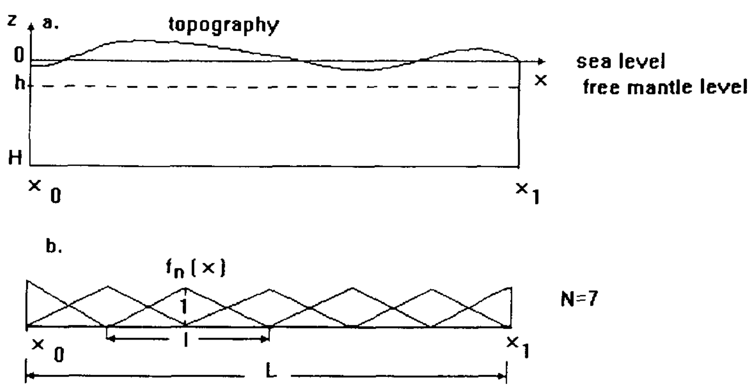

Fig. 3. (a) Geometric parameterisation of the model; density and viscosity distributions are assumed known. Horizontal velocities at the base of the model are given by the function $f(x)$, schematically represented in (b) as a system of linearly independent functions $f_n(x)$, with N being the number of triangles, l the baselength of a triangle, and the height of triangles being 1.

rection, W_k is the velocity of neotectonic movements $W(x)$ for node k, and $w_k^{(n)}$ is the calculated value of the vertical velocity for node k for unit function $f_n(x)$.

After the coefficients C_n have been determined, the function $f(x)$ can be found from Eq. 8, and the velocity field $\{u(x, z), w(x, z)\}$ can be computed as a solution to the direct boundary-value problem.

The number of triangles should be chosen so as to achieve the desired detail of topography evolution, while at the same time satisfying the boundary layer conditions. As initially a triangle has a fixed height equal to 1, the size of the triangle's base should provide the fulfilment of the above-mentioned conditions. It should also be noted that the base of the model should not be obligatory a physical boundary. It is an arbitrary level at which the boundary conditions are imposed. Furthermore, where the observed deformation is discontinuous in fault zones and zones of high fracture density (with slippage therefore tending to occur along faults), lower viscosity values can be assumed for the fault zones and zones of non-consolidated sediments in order to simulate the observations.

It is easily seen that the linear function $f(x)$ corresponding to a simple compression or tension at the side boundaries is just a special case of Eq. 8. Therefore, the solution whereby the vertical neotectonic movements of a region is governed purely by in-plane plate boundary forces is permissible.

The stress field is calculated on the basis of the slow flow pattern of the velocity solution to compare and correlate the resulting stress concentration zones with the zones of faulting and seismic activity observed. For descriptive purposes, the total pressure P, obtained by integrating:

$$\frac{\partial P}{\partial x} = \frac{\partial \tau_{xx}}{\partial x} + \frac{\partial \tau_{xz}}{\partial z}$$
$$\frac{\partial P}{\partial z} = \frac{\partial \tau_{xz}}{\partial x} + \frac{\partial \tau_{zz}}{\partial z} \tag{10}$$

the total shear stress τ, given by:

$$\tau = \sqrt{\left(\tau_{xx}^2 + \tau_{zz}^2 + \tau_{xz}^2\right)} \tag{11}$$

where τ_{ij} are the components of the viscous stress tensor, and the total shear to pressure ratio τ/P are calculated.

The shear to pressure ratio reflects the fracturability of rocks (Byerlee, 1968) and can be interpreted as a 'damage parameter'. Fractures appear when the τ/P value is high, and conversely, when τ/P ratio becomes comparatively low, they stop growing, and those that have been previously formed close. As such, a correlation can be made between zones of high τ/P and the occurrence of seismicity.

5. Application of the model

5.1. Geometry and mechanical parameters

A 2-D finite element model, comprising a 19×37 grid (vertical by horizontal) with grid intervals of 1 and 10 km, respectively, was constructed to represent the structural cross-section shown in Fig. 2. As such, the model length, $L = 360$ km, is appropriate to the length scale of the observed neotectonic movements of the northern Black Sea margin and the Crimea whereas the base of the model, $H = 18$ km, has been arbitrarily chosen to lie at an intracrustal depth in order to satisfy the stability criterion $H/L = \epsilon \ll 1$. The 2-D approach, in which it is assumed that the velocity vector projection in the direction perpendicular to the model plane is zero, is validated by the approximate east–west strike of the structures that have been active during the neotectonic stage and since the Black Sea depression has subsided as a unit during this time (Beloussov et al., 1988).

The function $f(x)$ was approximated by seven triangles ($N = 7$) each with base length $l = 120$ km (Fig. 3b).

The model input data are the density and viscosity distributions $\rho(x, z)$ and $\eta(x, z)$. In the present case, a homogeneous density model was employed. This was justified on the basis of preliminary model runs in which various proposed density distributions, including those of Artemjev (1975), Avdulov (1979) and Burianov et al. (1979) were employed. These results indicated that to predict surface velocities as vigorous and variable as those observed along the cross-section, the requisite boundary forces were so large that the internal body forces became negligible. The widely used value of 10^{23} Pa s (e.g., Turcotte and Schubert, 1982) was adopted as the crustal viscosity, including the bulk of the sedimentary column. To simulate the presence of active faults, lower viscosity values, of the order 10^{22} Pa s were presumed in fault zones. The same value was assigned to zones of unconsolidated sediments.

5.2. Results

Fig. 4 shows the resulting slow flow pattern for the model profile with the lengths of the arrows being proportional to the computed velocities. The boundary conditions were chosen in accordance with Eq. 9. The displayed vectors are drawn relative to a point of zero velocity on the right-hand boundary of the model, indicated by the square. For the present case, the calculated boundary conditions imply an intracrustal uplift beneath the oceanic part and downwarping beneath the Crimea. Movements on the side boundaries correspond to extension on the corresponding boundary (right or left depending upon the point of view). In the region below the continental slope, the velocity vectors change their orientation. From nearly vertical they become nearly horizontal.

Various aspects of the stress distribution arising from the flow pattern shown in Fig. 4 are presented as contour diagrams in Fig. 5. Fig. 5a displays the normal component of the stress tensor τ_{xx}, positive corresponding to tension, indicating that the crust beneath the Black Sea is in an extensional state with maximum extension being at the foot of the continental slope. The continental part of the model, in contrast, is in compression with the maximum lying immediately below Crimea. The state of stress in the continental crust north of Crimea is nearly lithostatic. Fig. 5b displays total pressure P calculated in accordance with Eq. 10 indicating that there is a zone of comparatively low total pressure at the foot of the continental slope. Fig. 5c shows the total shear stress calculated according to Eq. 11, indicating concentrations at both the foot of the slope and beneath Crimea and generally higher shear stress in the oceanic part of the model. Fig. 5d is a plot of the 'damage parameter', the shear to pressure ratio, with the greatest values being predicted at the foot of the slope.

6. Discussion

Although the velocities calculated at the base of the model pertain to a lower crustal depth, it

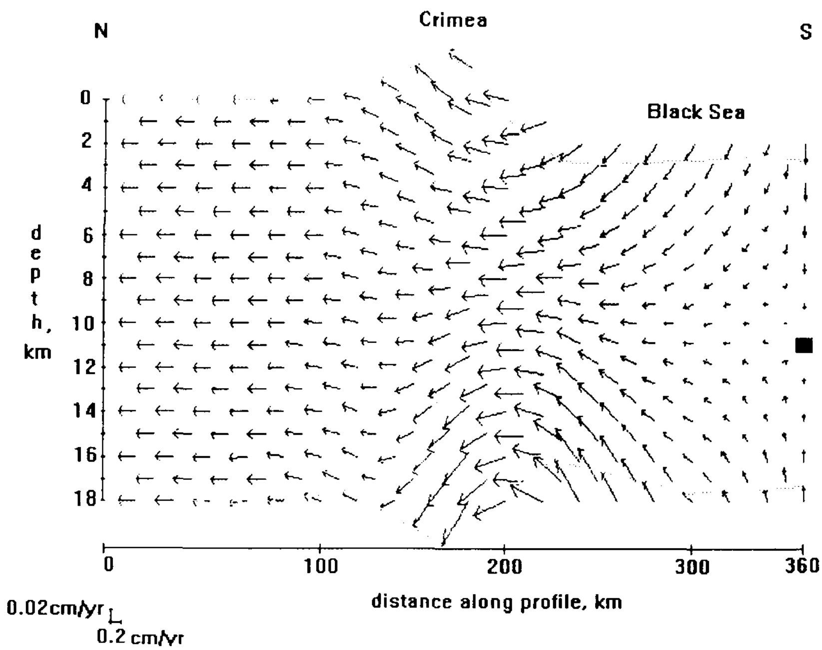

Fig. 4. Slow flow pattern for boundary conditions derived such that the vertical component of velocity on the surface best fits the observed data (Fig. 2). The displayed vectors are drawn relative to the point indicated by the square; the length of arrows is proportional to the computed velocity at that point.

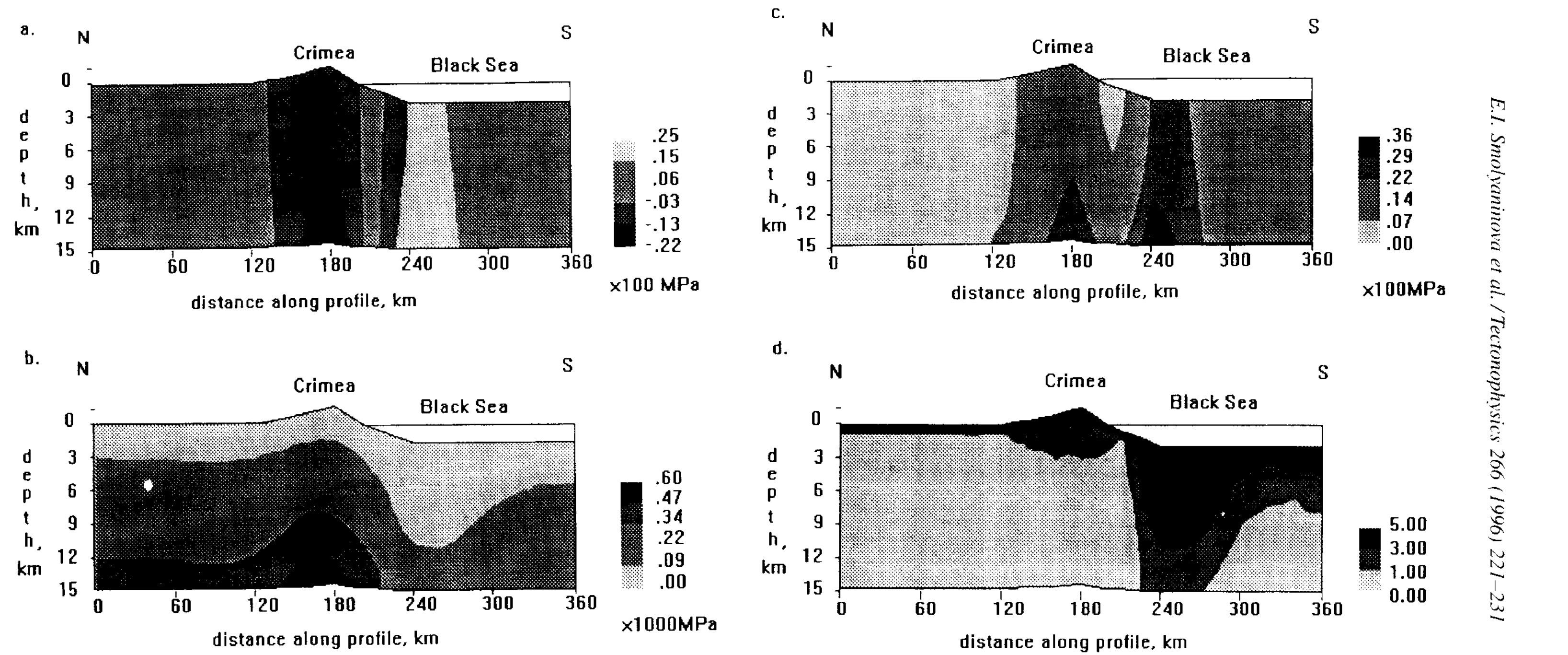

Fig. 5. Contour maps along the model cross-section of (a) normal stress component τ_{xx} of the stress tensor, (b) total pressure P, (c) total shear τ and (d) the shear stress to pressure ratio (or 'damage parameter').

can be assumed that they reflect similar movements in the upper mantle. The lithosphere and the underlying mantle are a closely linked, in particular with respect to vertical motions (cf. Kobozev and Myasnikov, 1987; Myasnikov et al., 1993). Flows in the mantle cause corresponding movements in the lithosphere and vice versa. The interplay is complicated by inhomogeneity of both interacting layers and strong feedback processes. In the present application, it follows that plate boundary forces cause the inferred mantle flows or, alternatively, that existing mantle movements cause the extension in the crust predicted by the model. Local mantle and/or lithospheric movements due to horizontal pressure gradients may be superimposed but, in general, the velocity distribution in the mantle can be inferred to be similar to $f(x)$.

The results indicate, in order to predict adjacent, rapid, vertical surface movements of opposite sign (uplift of the Crimea and subsidence in the depression), that $f(x)$ cannot be of the form $f(x) = a(x - x_1)$, which corresponds to a simple compression or tension applied to the side boundaries of the model such as described by Kolpakov et al. (1991) or, for example, found in models such as that postulated by Cloetingh and Kooi (1992). The implication is that the Neogene evolution of the Black Sea basin and its Crimean margin cannot be explained by in plane forces only although their presence and influence cannot be ruled out. Rather, the application of boundary forces at the model base is also necessary. These presumably reflect the complex regional neotectonic processes at play in the lithosphere and underlying mantle, responsible for the regional tectonic stress field during the past 10–30 Myr and related to the geological development of the Crimean margin during that time such as the formation of the present-day topography.

Fig. 4 illustrates that subsidence in the Black Sea depression can be caused, given the present model parameterisation, by movements resembling an upwelling mantle stream which in turn plunges beneath the Crimea to cause the uplift of the Crimea mountains. Although there is no supporting evidence of upwelling mantle in the geothermal data for the Black Sea, which is characterised by low heat flow values (Cermak and Hurtig, 1979), this may be reflection of thermal blanketing due to the high rate of sedimentation, which during the Quaternary has been more than 0.025 cm/yr (Finetti et al., 1988).

Fig. 5 shows that the model predicts a zone of abnormal stress at the foot of the continental slope. This area is characterised not only by the greatest shear stresses but also by the smallest total pressure, together resulting in a 'damage parameter' (τ/P) maximum in this area, coinciding with the seismically active Southern Coast fault (Fig. 1b). In terms of the present model, its position is fixed by the reversal of direction of the mantle flow velocity vector (Fig. 4). A secondary shear stress high is predicted below the continental part of the Crimea (Fig. 5c). However, the total pressure is much higher in this area than at the foot of the continental slope and, consequently, the predicted 'damage parameter' is low, a prediction consistent with the observed absence of significant seismicity there. Moreover, the predicted present-day state of compressional stress in the Crimea region is in consistent with the inferences of Finetti et al. (1988) based on seismicity data.

The results shown in Fig. 5 were obtained assuming no slippage along the Southern Coast fault. When slippage was 'allowed', by adopting lower viscosity values in the fault zone, the computed function (e.g., Fig. 4) did not change significantly although shear stresses in the fault zone were reduced. In consequence the previously found zone of high τ/P was negated, in fact transformed into a zone of lower τ/P than in the surrounding areas. The physical interpretation of this is that it corresponds to a phase of stress relaxation, as opposed to the case where no slippage along the fault was allowed which corresponded to a phase of stress accumulation. The results imply that it is the latter case that applies to the Southern Coast fault zone at the present time.

7. Summary and conclusions

Numerical modelling of the regional tectonic movements in the northern Black Sea has been presented, illustrating the possibility of using neotectonic data as well as upper lithosphere structure as input to such modelling. This results indicate complex regional tectonic movements in areas with vigorous surface movements of opposite signs (adjacent zones of uplift and subsidence). Mantle movements, in particular, may strongly affect the tectonic development in such regions.

An upwelling stream beneath the Black Sea and a stream downwelling under the Crimea are predicted to explain the uplift of the Crimean mountains and the subsidence of the adjacent sea floor during the neotectonic period (10–30 Myr). The calculated stress field lends support to these results. The maximum values of predicted shear stress to pressure ratio correspond to the location of the Southern Coast fault zone that controls the seismicity of the northern Black Sea implying that the seismic activity of this fault may be due to a change in the orientation of convective flows in the underlying mantle.

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