Numerical modeling of neotectonic movements and the state of stress in the North Caucasus foredeep

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Received 9 October 2000; revised 28 January 2002; accepted 7 February 2002; published XX Month 2002.

[1] This paper examines modern tectonic processes and the state of stress in the North Caucasus along three profiles that cross its eastern, central, and western parts. Using the models of crustal structure along these profiles, we solved the following inverse problem: to find boundary conditions (in this study, velocity at side boundaries and bottom of the model) that provide the best fit of calculated vertical movements at the top of the model and vertical neotectonic movements during the late Quaternary. Our modeling shows that during the late Quaternary, geodynamics of the eastern and central parts has been controlled by regional intraplate compression normal to the Greater Caucasus belt. The western part has been strongly influenced by tectonic processes taking place in the Eastern Black Sea. The calculated slow flow patterns and stress fields showed good agreement with the distribution of seismic events, results of focal mechanisms inversions, and GPS data. *INDEX TERMS:* 8102 Tectonophysics: Continental contractional orogenic belts; 8107 Tectonophysics: Continental neotectonics; 8164 Tectonophysics: Evolution of the Earth: Stresses—crust and lithosphere; 7215 Seismology: Earthquake parameters; *KEYWORDS:* Stress distribution, inverse problem, North Caucasus, foredeep basin, neotectonics, fault plane solution

1. Introduction

[2] Quantitative estimates of regional and local tectonic stress fields within the lithosphere are an important problem for both theoretical geodynamics and practical applications. A great number of papers are devoted to analytical or numerical solution of the direct problem - to calculate stress fields within a given area (or space) when boundary conditions are given. Our approach consists in a solution of the inverse problem: to reconstruct tectonic forces acting at the boundaries of a study area, which induce such movements at the top of the model that fit best in a least squares sense to rates of movements at the Earth's surface (given from geomorphology, GPS, etc.). This inverse method has been developed for 2-D structures using finite element technique, so any peculiarities of the crustal structure along a profile (inhomogeneous distribution of the density and mechanical properties, faults, weak or strong zones, etc.) can be taken into account. The obtained results can be compared with different data on state of stress of the lithosphere including distribution of seismicity, stress field determined from inversion of focal mechanisms, structural field data etc.

[3] Inverse method has been used to estimate regional and local stress fields along profiles crossing different regions: the Baltic Shield [*Kolpakov et al.*, 1991], the Northern Black Sea [*Smolyaninova et al.*, 1996], and the Southern Beaufort Sea [*Stephenson and Smolyaninova*, 1999]. In this study we applied the method to a very complicated area-the transition from the active Caucasus mountains to the stable Scythian plate. Calculations along three profiles crossing different tectonic structures are presented. Geodynamics of this region is extremely varied as it is governed not only by the collision of the Eurasian and Afro-Arabian plates, but also by active tectonic processes in the Eastern Black Sea and Southern Caspian basins. To take into account the complex interaction of all these processes, we suggest that tectonic forces act not only within the lithosphere (intraplate forces) but are acting on its base as well (mantle-induced forces).

[4] *Bird* [1996] developed a similar approach when studying the Alaska region, where he solved a 2-D plain problem, averaged along a vertical axis. However, he did not solve the inverse problem, but instead analyzed results of multiple calculations for different boundary conditions. This approach cannot guarantee a best fit solution.

2. Geological Framework

[5] The wide transition zone from the Greater Caucasus mountain belt to the Scythian plate, referred to below as the North Caucasus foredeep, includes three main tectonic structures (from east to west): the Terek-Caspian trough, the Stavropol high, and the Azov-Kuban trough (Figure 1). All these tectonic units differ from each other in their origin and evolution, as well as in their present-day structure and geodynamics [*Panov*, 1976; *Sedenko*, 1976; *Gamkrelidze*, 1986]. The main thrust zone of the Greater Caucasus belt is situated at its southern slope, deeping to the North [*Philip et al.*, 1989], so the North Caucasus foredeep can be

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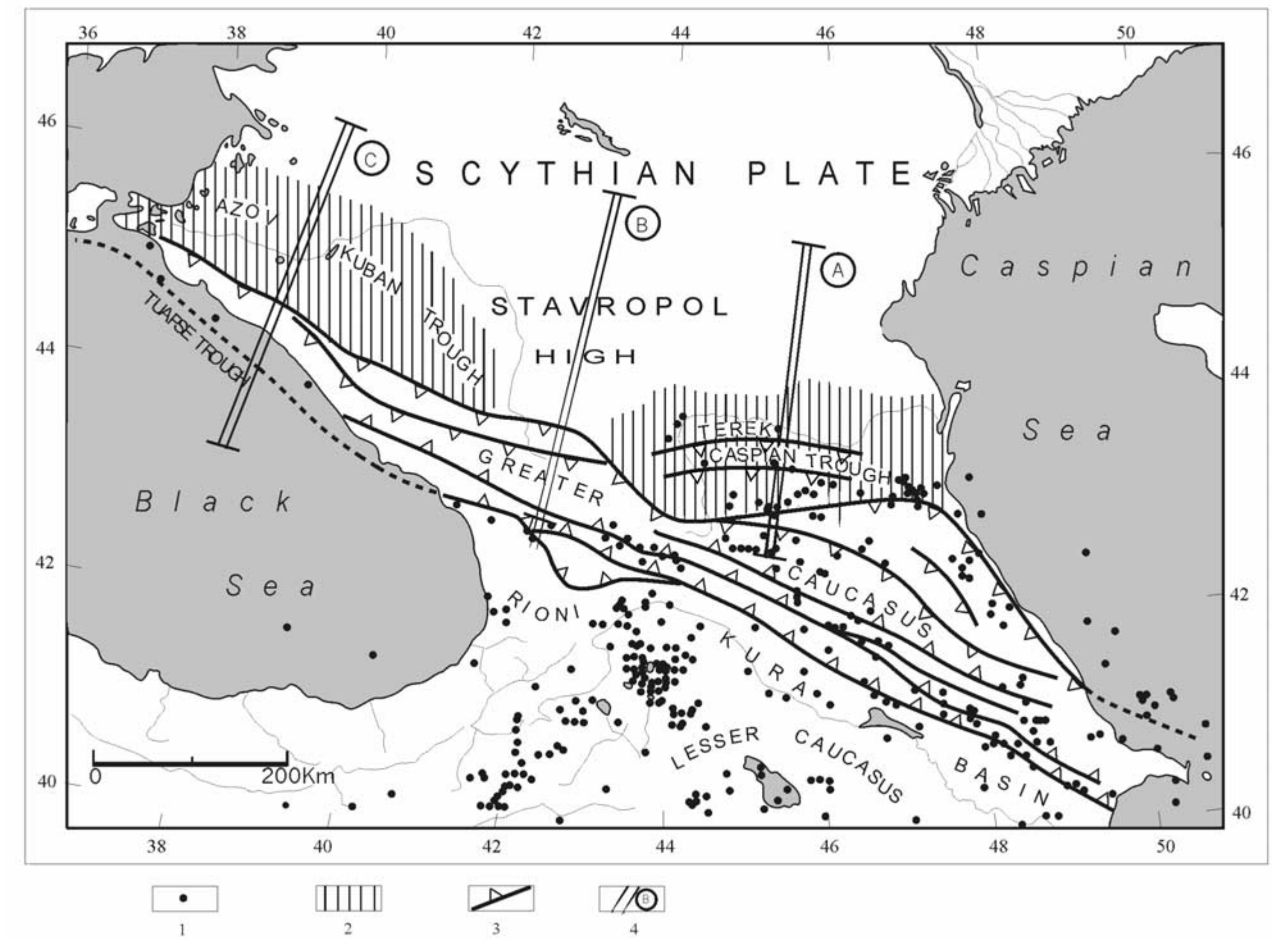

Figure 1. Sketch map of the main structural units of the study area. 1. - epicenters of seismic events with available focal mechanism; 2. - Azov-Kuban and Terek-Caspian troughs; 3.- the major thrust faults; 4.- location of profiles A, B and C shown in Figure 2.

considered as the main back thrust system of the Caucasus mountains.

[6] The structure of the eastern part - the Terek-Caspian trough, is more complex. It incorporates the Terek and Sundja ridges, whose deep structure is not clear, as resolution of seismic data is not sufficient to show the position of main interfaces below these ridges. Crustal-scale profiles across the Terek-Caspian trough based on DSS, gravity data [*Pavlenkova*, 1980; *Krasnopevtseva*, 1984; *Artemiev et al.*, 1985], and available geological information are shown in Figure 2a (position of Figures 2a–2c is shown in Figure 1). More detailed profiles of sedimentary layers based on shallow seismic and borehole data for all three tectonic regions were published by *Mikhailov et al.* [1999b]. The western part of the North Caucasus foredeep – the Azov-Kuban trough, is characterized by a relatively simple asymmetrical structure with a gentle and wide northern flank, and a very narrow and steep southern one that corresponds to the location of the main backthrust system (Figure 2c). The central area – the Stavropol high, is distinguished by the absence of a clear foredeep basin in the Mesozoic and Cenozoic sedimentary stratigraphy (Figure 2b). The present-day topography of the Stavropol high has been forming since the Pontian (5–7 My) which is coeval with the main uplift of the Greater Caucasus [*Panov*, 1976; *Sedenko*, 1976].

[7] A characteristic feature of the North Caucasus region is that the deepest basins have been developed not in front of the highest central part of the mountain belt, but in front of its relatively low western and eastern parts (the Azov-Kuban and Terek-Caspian troughs). In front of the highest central part of the Greater Caucasus there is not a clearly developed foredeep basin. This relationship between height of mountains and depth of foredeep basins is inconsistent with the widely used model considering subsidence of a foredeep as a result of elastic flexure of the lithosphere under weight of mountain ridge topography. This flexural model was originally suggested by *Price* [1973], and later tested numerically using different models: elastic plate with constant [*Jordan*, 1981] or temperature-dependent elastic thickness [*Beaumont*, 1981], visco-elastic plate [*Karner and Watts*, 1983], and plate with variable thickness [*Stockmal*

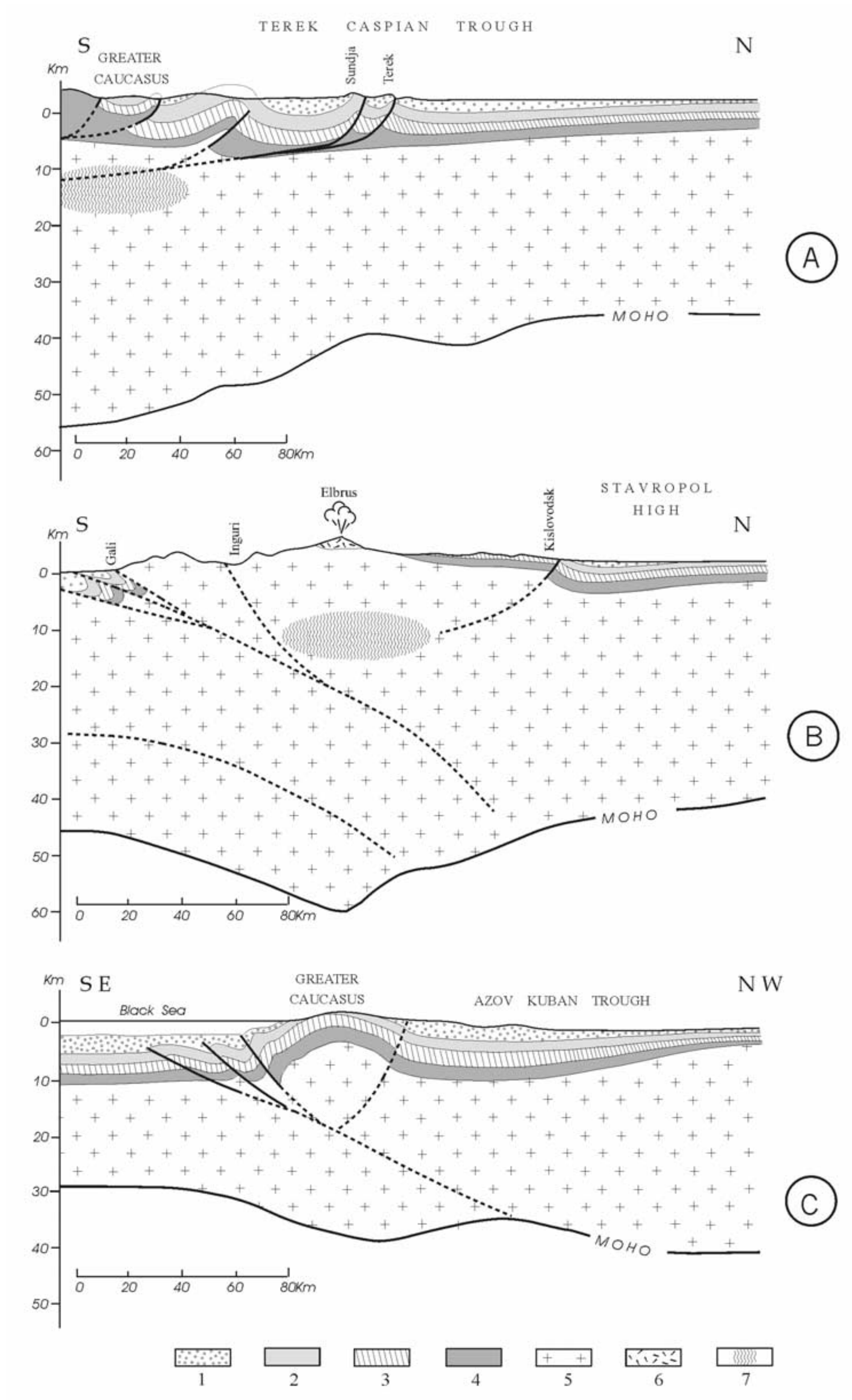

Figure 2. Simplified cross sections along profiles A, B, C (Figure 1) based on borehole, seismic reflection and DSS data. All the data not substantial for the modeling are omitted. For more detailed sections of the sedimentary cover, see *Mikhailov et al.* [1999b]. 1, N-Q sediments; 2, Pa_3-N_1; (Maykopian); 3, Pa_{1-2}-K; 4 -T-J; 5, crystalline crust; 6, volcanism; 7, low-velocity zones in the crust according to DSS data.

et al., 1986]. A common feature for many orogenic belts is that loading due to topographic weight appears to be insufficient to explain observed subsidence in the foredeep. Existence of a "hidden" loading (density inhomogeneities) was suggested, e.g., for the Alps [*Karner and Watts*, 1983] and Apennines [*Royden and Karner*, 1984]. Involving such a load can hardly be admitted for the North Caucasus region: according to the gravity modeling [e.g., *Artemiev et al.*, 1985] the lithosphere in this region has lower average density as compared to adjacent territories.

[8] *Ershov et al.* [1999] made an attempt to explain the formation of the North Caucasus foredeep by elastic flexure solely by changing the thickness of both the crust and the lithosphere below the Greater Caucasus. To find a solution, they had to assume huge changes of lithospheric thickness (28 km of thickening below the eastern Greater Caucasus and 60 km of thinning below the Central Greater Caucasus, thus they obtained 88 km of asthenospheric uplift within a distance of 150–200 km). The adopted value of effective elastic thickness of 60 km seems to be also too big for the Alpine mountain belt. Even for such an extreme geodynamical model, they arrived to the best fit solution, which differs from the observed Bouguer anomalies by more than 100 mGal [*Ershov et al.*, 1999, Figure 9]. Thus this attempt at explaining the formation of the North Caucasus foredeep in the framework of an elastic flexure model seems to be questionable.

[9] All the difficulties of describing formation of foredeep basins only by elastic flexure has a major consequence in searching some other mechanisms of subsidence [e.g., *Artyushkov et al.*, 1996, *Zhang and Bott*, 2000].

[10] A detailed comparative study based on the analysis of numerous subsidence curves calculated for different parts of the North Caucasus foredeep, showed that foredeep formation was characterized by alternating relatively short (some several million years) uplift events and longer (up to 10 Myr) subsidence events [*Mikhailov et al.*, 1999b]. Although the amplitude of these events differs from place to place, the majority of events were synchronous throughout the North Caucasus region. The first period of uplift, marking the beginning of orogenic belt- foredeep basin formation, took place between 39.6 and 36.0 My, i.e., before deposition of the Oligocene-Lower Miocene Maykopian series. The subsequent uplift events occurred in the intervals 16.6–15.8, 14.3–12.3, and 7.0–5.2 My, respectively. According to neotectonic data, the Quaternary period is also marked by an uplift event. The present paper focuses on the Late Quaternary uplift event.

[11] Comparing uplift events in the foredeep basin with faulting and folding deformations in the Greater and Lesser Caucasus, *Mikhailov et al.* [1999b] concluded that periods of uplift or close to zero subsidence coincided with regional compressional events. Such a correlation has been found in many regions. For example, when studying the pre-Carpathian foredeep, *Artyushkov et al.* [1996] arrived at similar conclusions regarding the coincidence of uplift events in the foredeep basins and periods of regional compression. To explain such behavior of foredeep development an alternative model of the North Caucasus foredeep formation was suggested [*Mikhailov et al.*, 1999a]. According to this model, disturbance of the mechanical and thermal equilibrium in the Earth's outer shell (including the lithosphere and asthenosphere) due to intraplate compression triggers small-scale flow in the asthenosphere, which restores equilibrium in the Earth's outer shell. Its observable manifestations are uplift of the whole orogenic domain, including the foredeep, during compressional events, followed by subsidence in the foredeep basins during the restoration of equilibrium that postdates the compressional events. Additional subsidence may result from thrusting, elastic flexuring, and other tectonic processes.

[12] The present-day along-strike structural variations of the North Caucasus foredeep have been predetermined by inhomogeneous lithospheric structure at the beginning of the Jurassic. According to *Panov* [1976], the Stavropol high had been an uplifted area during Early Jurassic and possibly earlier; thus it may be distinguished by a different crustal structure. Distribution of sedimentary rock thickness and facies showed that the Liassic extension of the Greater Caucasus trough had not been uniform either: two deep basins were formed to the east and west from a relatively shallow central part of the trough [*Panov*, 1976]. Strong along-strike variations in the thickness of the Lower Jurassic sediments, as well as alkaline diabase dykes of the same age stretching from north to south, support the idea of *Stampfly and Pillevuit* [1993] that the Early Jurassic extension of the Greater Caucasus trough was accompanied by strong left lateral transform movements, which led to the formation of pull-apart basins.

[13] The present-day geodynamics of the North Caucasus region is even more complex. It is governed by all the previous geological history as well as more recent superimposed external factors. These external factors are significant and complicated as the study area is situated to the north of the collision zone of the Eurasian and Afro-Arabian plates. Moreover, the Black Sea depression to the west and the Caspian Sea region to the east also have been zones of active tectonic processes during the Neogene-Quaternary.

[14] The distribution of seismic activity reflects the complex geodynamics of the North Caucasus area. Almost all earthquake epicenters here are located under the Greater Caucasus mountains. According to the seismic catalogue of the United Institute of Physics of the Earth, Russian Academy of Sciences [*Ulomov*, 1993], there are only a few epicenters in the Azov-Kuban trough and the Stavropol high. In contrast, the Terek-Caspian trough exhibits more seismic activity. The epicenter distribution of focal mechanism solutions based on the data from the UIPE catalogue is showed in Figure 1. Even if availability of focal solutions is controlled by two factors, the magnitude value and network configuration, distribution of events in Figure 1 demonstrates stronger seismic activity in the eastern part of the North Caucasus (Terek-Caspian trough), than in the central and western parts (Stavropol high and Azov-Kuban foredeep) where no focal solutions have been calculated. From our point of view, the observed rapid eastward increase of seismic activity cannot be explained by eastward increase of convergence rate [*DeMets et al.*, 1990] between the Afro-

Arabian and Eurasian plates, because east to west changes of convergence rate are relatively small [*Reilinger et al.*, 1997]. This investigation is an attempt to gain insight into the factors controlling the complex present-day geodynamics, the state of stress and seismicity in the North Caucasus foredeep.

3. Description of Numerical Method

[15] One of the questions that arise when performing numerical modeling is choosing the constitutive law of the media. In our modeling we use finite element technique, thus there are no restrictions on the rheological laws involved. Nevertheless, the more complicated model we use, the more parameters we have to assign. For example, when using brittle – ductile rheology, one should assign composition of the crustal layers and lower lithosphere, distribution of temperature, values of principal stresses and so on. For the Greater Caucasus, there are no detailed seismic data to constrain parameters of the media. We believe that it is better to use a simple model to describe the media, instead of a complicated one, which is not based on observed data. The object of the present study is to determine tectonic movements in the Northern Caucasus foredeep during the Quaternary; that is, we deal with comparatively large time and spatial scales. This enables us to consider a viscous model to describe the media. In this study the sought-for functions are components of the velocity vector at the base and side boundaries of the model. For this kinematic problem, the main regional features of strain and relative stress distribution do not depend critically on the constitutive law that is used. Solutions are similar even for inhomogeneous viscous and elastic media. Thus, to determine the velocity field within the model region, we solved Stokes equations, assuming material incompressibility and an absence of body forces other than those due to gravity.

[16] To calculate stress and strain fields in a specific area of the lithosphere, it is necessary to know the distribution of density and mechanical properties in the area, and a set of boundary conditions. Boundary conditions are determined by such outgoing forces as intraplate forces or mantle drag, and thus they were previously estimated only qualitatively on the basis of numerous, sometimes conflicting geodynamical hypothesis. At the same time, boundary conditions are precisely the things that are of interest in order to understand the mechanism of structure formation as well as for numerical modeling. Here we present the main ideas of the approach to the quantitative estimation of boundary conditions without any preliminary assumptions regarding the nature of acting outgoing forces.

[17] The principal idea of the approach is that boundary conditions are found as a result of the inverse problem solution, so that calculated movements on the surface match the observed data (neotectonic or GPS). As both the horizontal and vertical components of the velocity vector cannot be found independently [e.g., *Mikhailov*, 1993], we used an equation linking these two components following the suggestion of *Braun and Beaumont* [1989]. At first, their approach was suggested for modeling of sedimentary basins formed as a result of intraplate extension. This model can be applied to modeling of orogenic belt- foredeep basin system by changing extension for compression. In this model the process of lithospheric structure formation subdivides into two stages – extension in the absence of gravity and subsequent isostatic rebound. For the first stage, *Braun and Beaumont* [1989] suggested that (1) during deformation of the lithosphere, the horizontal component of velocity vector U does not depend upon the vertical coordinate and (2) there exists some specific "necking" level z_n (further referred to as the neutral level), which remains horizontal during deformation of the lithosphere in the absence of gravity. Thus, at the first stage W the vertical component of the velocity vector in the lithosphere is related to the horizontal one by the equation: $W(x, z) = -(z - z_n)\partial U/\partial x$, where x and z are the horizontal and the downward vertical coordinates, respectively. This equation has been extensively used for modeling of different sedimentary basins (for bibliography, see *Cloetingh et al.* [1995]). In basin modeling, depth to the neutral level was treated as a free parameter; its value is supposed to be dependant on composition, age, and thermal state of the lithosphere. Considering the deformation of a thin inhomogeneous elastic plate by intraplate forces in the absence of gravity (the first stage), *Mikhailov* [1999] demonstrated that its deformation obeys the above mentioned equation, and found dependence of the depth to z_n on the distribution of mechanical properties with depth. For example, when the lithosphere contains two rigid layers of the same thickness situated at the top of the crust and below the Moho, the level z_n is situated in the middle crust equidistant from both layers. When there is one rigid layer at the top of the crust (in the case of tectonically active areas having high thermal gradient), the level z_n is situated at a shallow depth in the middle of the rigid layer, close to the Earth's surface.

[18] Depth to the neutral level z_n critically determines the behavior of the model at the second stage. Indeed, disturbance of isostatic equilibrium as a result of deformation at the first stage (usually referred as a "load") is linearly proportional to the distance between the neutral (z_n) and free-mantle (z_{fm}) levels [e.g., *Mikhailov*, 1999]: $q(x) = \rho_a g(z_n - z_{fm})\partial U/\partial x$, where ρ_a is density of the asthenosphere, g is gravity acceleration, and level z_{fm} is the so-called free mantle or floating level. Definition of this level is inferred from the condition of isostatic equilibrium. Physical explanation of the free mantle level is that of an equilibrium level to which an inviscid mantle would rise when a well is drilled through the crust to the depth of the mantle. The load $q(x)$ causes deflection of the lithosphere at the stage of isostatic rebound. Thus kinematics of the lithosphere in the model of *Braun and Beaumont* [1989] is complex: it is pure shear in the first stage, when in the second one it includes rotation as a result of lithospheric flexure.

[19] For a thin lithosphere or a lithosphere having high thermal gradient the levels z_n and z_{fm} are close and thus the load q is close to zero. In this case, the lithosphere holds the local isostatic equilibrium during deformation by intraplate forces, even if its effective flexural rigidity is comparatively

high. In general, these levels do not coincide, and the lithosphere restores to isostatic equilibrium, either locally or regionally, depending on its flexural rigidity. Thus elastic flexure at the second stage depends on two parameters: depth to the neutral level and effective flexural rigidity. It is worth noting that the problem of estimating the both these parameters using gravity data has no unique solution [*Mikhailov*, 1999].

[20] Thus, at the first stage (deformation by external forces in the absence of gravity) we used the equation $W(x, z) = -(z - z_n)\partial U/\partial x$ to link the horizontal and vertical components of the velocity vector at two vertical side boundaries and at the base of the model. (It is important to emphasize that this equation is valid not only for deformation by intraplate forces. *Myasnikov and Savushkin* [1978], *Kobozev and Myasnikov* [1987], and *Myasnikov et al.* [1993] arrived at the same equation when considering formation of regional structures in the lithosphere by mantle-induced forces.) The surface boundary of the model is supposed to be stress-free. When depth to the neutral level is assigned, we arrive at the problem of estimating one of the components of the velocity vector: W or U (strictly speaking $\partial U/\partial x$) under the condition that provides a best fit to uplift and subsidence rates obtained from neotectonic or GPS data. To solve this problem numerically, the sought for function (function W in our consideration) at the bottom was presented as an expansion in a series of elementary functions (in this study these functions were triangles, shifted from one another horizontally at a half their width with corresponding boundary conditions at the sides). Using the finite element method [*Zienkiewich and Taylor*, 1989], the problem of deformation of viscous media was solved for every elementary function. In particular, the vertical component of velocity was calculated at the top of the model. After that, the amplitude of every triangle can be found under the best fit (in the least squares sense) to the observed movements at the top of the model. As for a Newtonian media this problem is linear, we arrived at a system of linear equations, which was solved by least squares method. For detail description of the procedure see [*Smolyaninova et al.*, 1996].

[21] It should be emphasized that, in general, when the neutral level does not coincide with the free mantle level, the load $q(x)$ and elastic flexure are to be calculated for every elementary function as well. Flexural rigidity can be either assigned or treated as a free parameter. This elastic flexure should be added to the vertical component of the velocity vector at the top of the model found as a result of the FEM solution. The neotectonic and geodetic data for the Northern Caucasus region, discussed in the next section, have not shown subsidence in the Eastern and Central parts and only small subsidence in its western part during the late Quaternary (see below). Taking into account these data, we concluded that rigid (effective elastic) layers in the Greater Caucasus lithosphere are only located at the top of the crust. Thus we considered the neutral level being close to the free mantle level, and considered the load $q(x)$ being negligibly small, at least for the late Quaternary. This makes it possible to explain why there is no foredeep in front of the highest mountains at the Stavropol high.

[22] It should be stressed that the bottom boundary of the model needs not coincide with any physically interpretable boundary. The formal statement of the problem, which was used to establish boundary conditions, is based on the thin layer approximation, because the horizontal components of velocity vector in the lithosphere below the bottom of the model supposed to be independent of z. As a result, the particular choice of the depth to this bottom boundary of the model is not important for the solution. The main point is that it has to be deeper than the main structural inhomogeneities in the crust, while to comply with the thin layer approximation the ratio of horizontal to vertical dimensions of the model must be much greater than unity (for more details, see *Stephenson and Smolyaninova* [1999]).

4. Geological/Geophysical Settings and Description of the Adopted Models

[23] The three profiles crossing the western, eastern, and central parts of the North Caucasus foredeep were chosen for the modeling. For the location of the profiles see Figure 1. These profiles will be informally referred below as Terek-Caspian or profile A, Stavropol (B) and Azov-Kuban (C) profiles, respectively.

[24] It should be mentioned that a large body of borehole and reflection seismic data reveal in details the upper structure of the studied area. In contrast, data on deep structure of the North Caucasus foredeep are scarce. There are several 30–40 year old DSS profiles, which have been reinterpreted many times. Thus here we placed the simplified sections based on both the DSS [*Pavlenkova*, 1980; *Krasnopevtseva*, 1984] and reflection seismic profiles accounting for the borehole data from *Mikhailov et al.* [1999b]. All the data, which are ambiguous and not substantial for the modeling, were omitted (Figures 2a–2c).

[25] Vertical movements, which are to be given at the surface of the model, can be governed in tectonically active regions, as the Caucasus, by several tectonic factors of different origin. Among them are:

[26] Vertical movements due to regional compression by intraplate forces, which are acting nearly horizontally [*Zoback*, 1992].

[27] Movements at the surface governed by mantle activity. Convective movements in the mantle can develop as a result of mechanical or thermal instability in the Earth's outer shell in areas of compression or extension [see, e.g., *Buck*, 1986; *Mikhailov et al.*, 1996] or due to phase transitions [e.g., *Artyushkov et al.*, 1996].

[28] Thermal subsidence due to disturbance of thermal equilibrium (heating) when the Greater Caucasus trough was formed in the Mesozoic.

[29] Topography changes due to sedimentation and denudation, and corresponding isostatic movements.

[30] The aim of this investigation is to estimate the influence of active tectonic forces (first two factors). Thus corrections for the third and forth factors, as well as for subsidence due to sediment compaction, must be introduced.

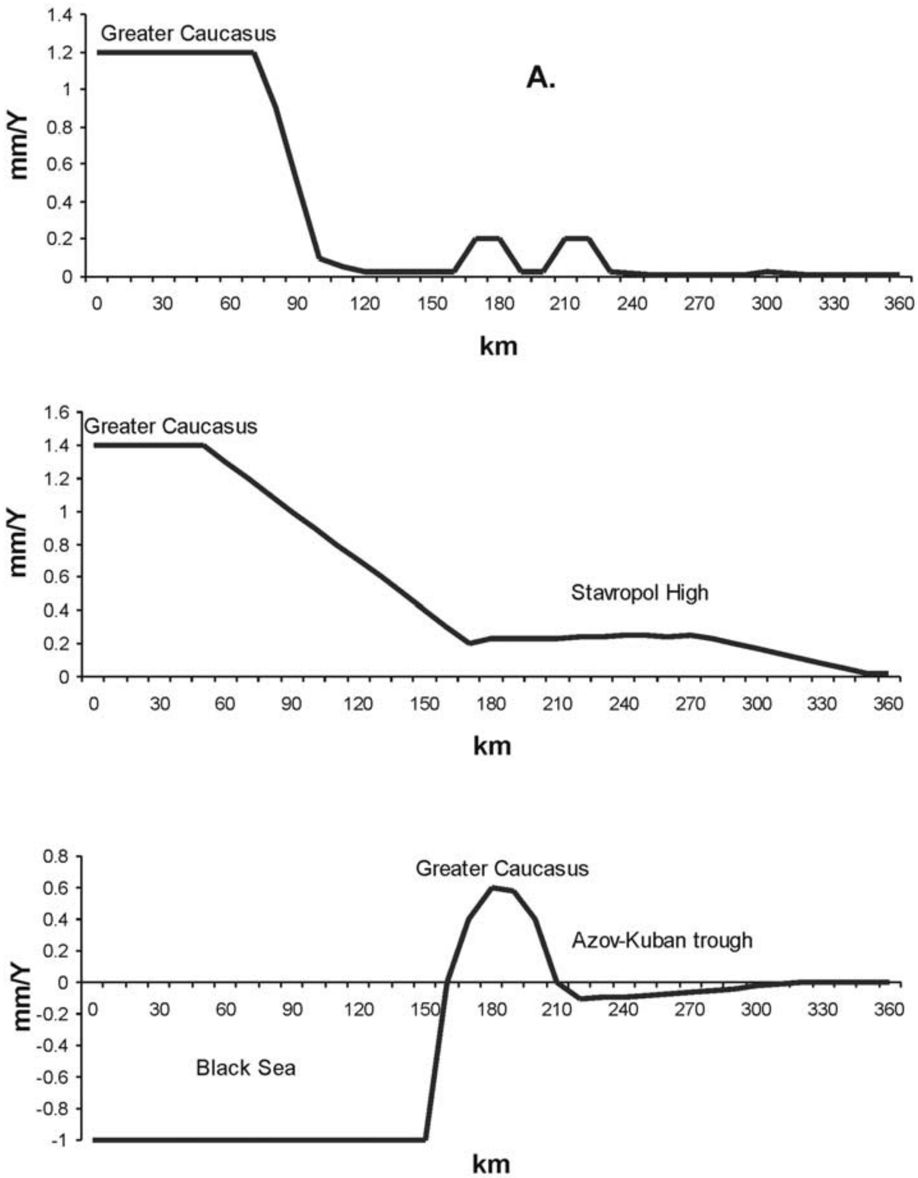

Figure 3. Vertical movements at the surface along profiles A, B, C shown in Figure 1 based on late Quaternary geomorphology data [*Beloussov and Enmann*, 1999], subsidence curves [*Mikhailov et al.*, 1999b], and marine reflection seismic profiles [*Tugolesov*, 1985].

[31] It is also very important to determine the time interval for which vertical movements are evaluated. Vertical movements consist of components of different length. For the Caucasus, the measured amplitude of vertical movements strongly depends upon duration of the time period chosen: even the sign of obtained values for movements can change.

[32] On the map of neotectonic deformations [*Nikolaev*, 1977] the interval of data averaging is about 20 my. The Terek-Caspian and Azov-Kuban troughs on this map are depicted as subsidence areas, which according to sediment thickness were subsiding during this period at a rate of 0.025–0.05 mm/yr (in the vicinity of profiles A and C, Figure 1). On the other hand, according to *Mikhailov et al.* [1999b], there were at least three uplift periods followed by rapid subsidence during this time interval (and also Quaternary uplift events). Thus we conclude that neotectonic movements averaged for the time period about 20 My may be used in modeling the formation of the whole orogen-foredeep system. However, to get information on modern dynamics and state of stress we have to incorporate only data on recent movements.

[33] In order to get information on recent movements, we analyzed topographic data for the central part of the North Caucasus foredeep, accounting for the denudation and

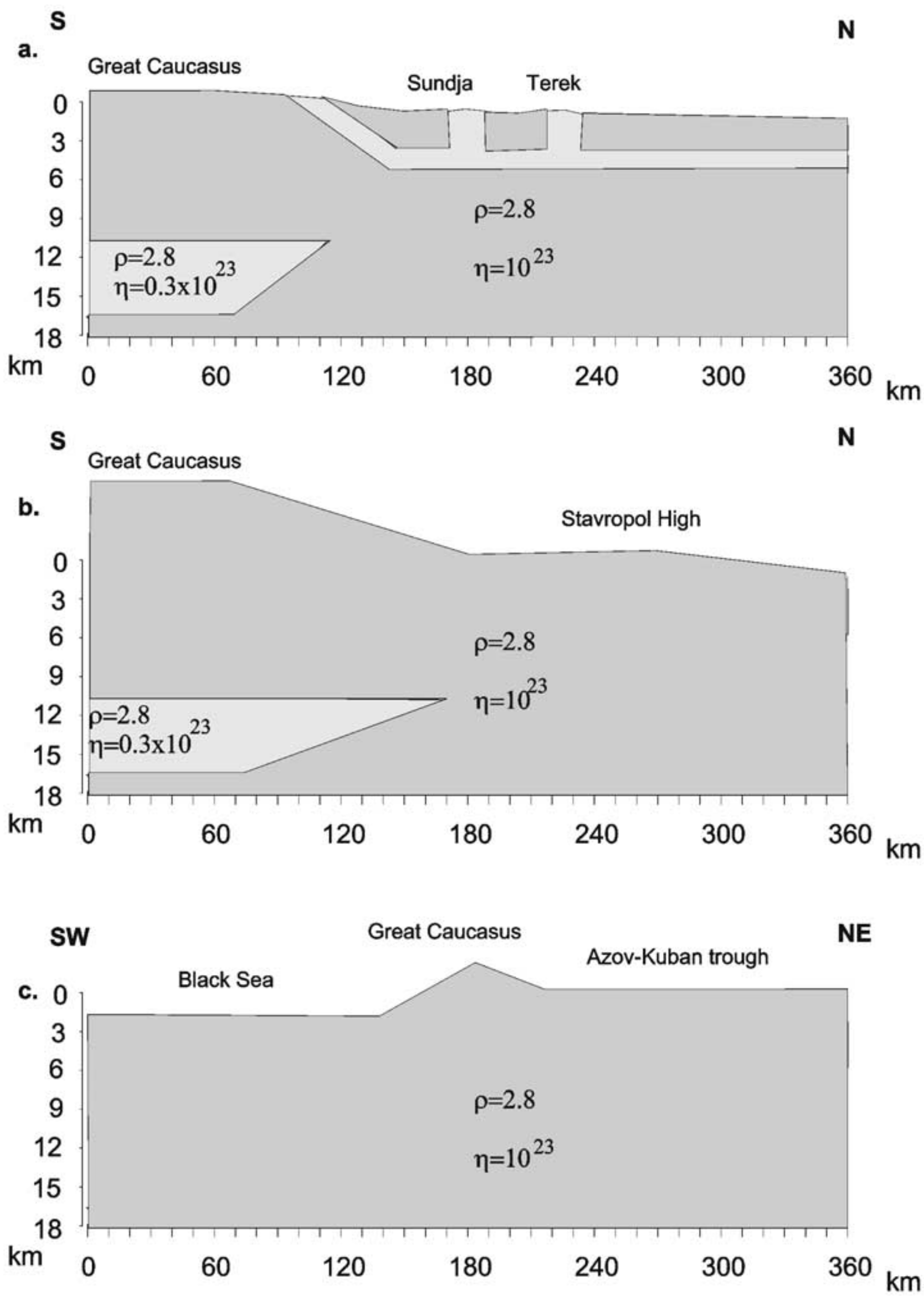

Figure 4. Density (g/cm^3) and viscosity (Pa s) models adopted for modeling along profiles A, B, C (Figure 1). Lower-viscosity zones (marked light) below the Caucasus correspond to lower seismic velocity zones given in accordance with DSS profiles [*Pavlenkova*, 1980; *Krasnopevtseva*, 1984]; in the Terek-Caspian trough these zones correspond to lower-viscosity Oligo-Miocene Maykopian clays and highly deformed rocks of the Terek and Sundja ridges.

sedimentation [*Beloussov and Enmann*, 1999]. According to these estimations, the Stavropol high rose up to 200–400 m during the Pliocene, while the uplift of the area to the north from the Stavropol high (Figure 1) was about 50–100 m in relation to the averaged topography formed by the end of Apsheronian (1.6–0.6 My) time. This leads to average values of Quaternary vertical movements of about 0.04 ± 0.06 mm/yr for the northern part of the profile and 0.17 ± 0.06 mm/yr for the Stavropol high (accounting for the fact that the accuracy of initial topography estimates is at least ±100 m). The rate of vertical movements for the Greater Caucasus during the last 1.8 My is about 1.4 mm/yr. The rate distribution along the profile B for the Quaternary is shown in Figure 3b.

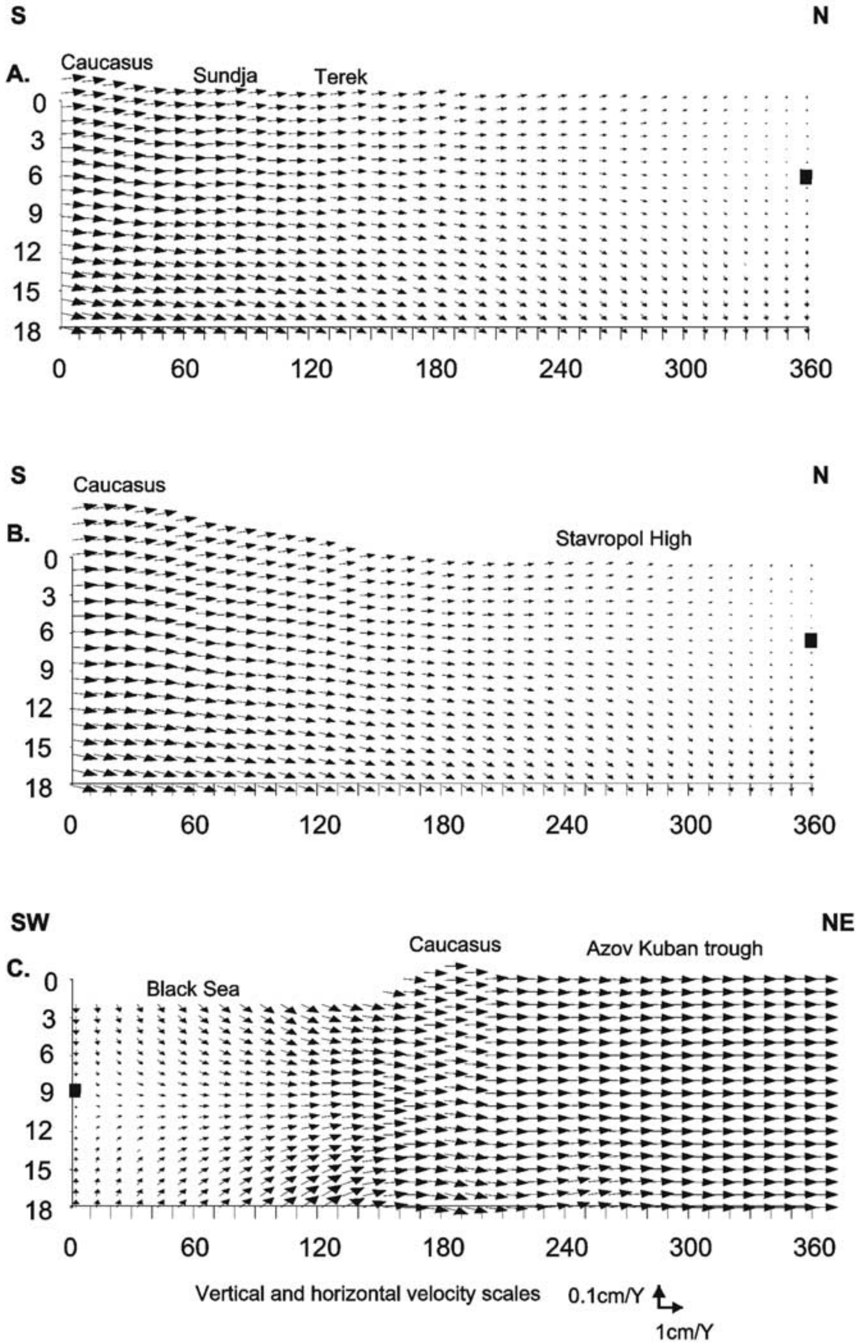

Figure 5. Calculated slow flow patterns for profiles A, B, C shown in Figures 1 and 2 in relation to a fixed point marked by square. The length of arrows corresponds to the absolute values of the velocity. To show uplifting movements below the Black Sea, the fixed point on profile C was stated at the left side of the profile. For profiles A and B it was stated at the right side. Mention the rapid changes of the velocity in a transition zone from the Western Caucasus to the Black Sea (profile C). See section 8 for more details.

[34] These data were also compared to results of repeated geodetic measurements [*Enmann and Nikonov*, 1993; *Beloussov and Enmann*, 1999]. Analysis of different maps shows that amplitude of movements on the maps strongly varies from year to year depending upon the data itself as well as upon methods of processing. All the estimations of rates of vertical movements based on the repeated geodetic measurements are significantly discrepant with the Quaternary data. Taking into account that these geodetic data are not accurate enough and that they are probably controlled by local processes (and, thus are not related to the sought-for tectonic signal, which controls modern state of stress and

seismicity) we incorporated in our modeling only the Quaternary data (or late Quaternary data when available).

[35] For profile A, running across the Terek-Caspian trough we used data based on the subsidence curves analysis [*Mikhailov et al.*, 1999b] as well as data from L.V. Panina (personal communication) and maps of the Quaternary vertical movements [*Beloussov and Enmann*, 1999]. The existence of an uplifted area to the south of the 44°N latitude due to strong uplifting of the Terek and Sundja ridges is indicated by all these data. The territory to the north of the ridges is characterized by zero uplifting or insignificantly small values of subsidence. However, borehole data that record the late Quaternary sediments (0.7 My) reveal uplifting movements there. The summarized data used for the modeling along this profile are shown in Figure 3a.

[36] For the Azov-Kuban trough, average rates of subsidence for the Quaternary, accounting for isostasy and compaction, are about 0.1 mm/yr [*Mikhailov et al.*, 1999b]. Uplift rates of the Greater Caucasus in the vicinity of the intersection with profile C (Figure 3c) are up to 0.6 mm/yr [*Nikolaev*, 1997], and subsidence in the Black Sea depression, accounting for thermal subsidence and sediments loading, is in the order of 1.0 mm/yr [*Tugolesov*, 1985].

[37] The amplitude of thermal subsidence was estimated using the analysis of the history of subsidence of the Terek-Caspian trough [*Mikhailov*, 1993]. It was shown there that tectonic movements in the trough since the Jurassic can be represented as a sum of the two components: (1) long-term exponentially reducing total subsidence of the trough as a whole and (2) local movements varying from place to place. If we assume that the long-term subsidence component is determined by thermal subsidence due to extension of the Greater Caucasus area in the Jurassic, we can estimate a correction, which should be added to rates of tectonic movements for the Quaternary. This correction is calculated to equal to 0.004 mm/yr, i.e., negligibly small as compared to absolute values of Quaternary rates.

[38] To describe the inner structure of the study area in the present model, we have to assign distributions of density and viscosity. For all the profiles we tried several density models and various assumptions on viscosity distribution. It was found that for all three profiles, the existing density variations between different models are not essential, as no strong gravity anomalies are observed in the study areas [e.g., *Artemiev et al.*, 1985]. On the contrary, to match the observed neotectonic data and seismicity in this area of active tectonics, regional tectonic forces (boundary conditions) have to be very strong. As a result, boundary conditions are the main factor controlling movements at the surface of the model and disturbance of the stress fields in the Earth crust. Different density distributions are able to cause some local peculiarities in stress fields, but regional movements remain the same. Taking into account uncertainties of density estimates, the homogeneous density model with an average density of 2.8 g/cm^3 was adopted in the final version.

[39] Distribution of viscosity appears to be more important for results. The inverse problem has been stated as a kinematic one: the sought-for function is the vertical component of the velocity vector at the side and bottom boundaries of the model when the vertical component of the velocity vector at the top of the model is given. Therefore the inverse problem solution does not depend on the absolute value of viscosity. Qualitatively, stress field corresponding to velocity field obtained also does not depend upon the absolute values of viscosity: the location of zones of maximum and minimum stress do not vary when changing absolute values of viscosity. Absolute values of stress components are in direct proportion to the adopted viscosity. Thus, for the present case we assumed viscosity values of the order of 10^{23} Pa s to obtain reasonable values of stress. What appeared essential is the existence of viscosity drops reflecting changes in mechanical properties of rocks (although not always). For the western part, the adopted homogeneous viscosity provided the best results (Figure 4c). For the central part, the introduction of the zone of low viscosity (Figure 4b) reflecting the zone of low seismic velocity, appeared necessary to support higher uplift in the Greater Caucasus belt compared to the Stavropol high. The existence of such zones in the vicinity of profiles A and B do agree with the DSS data [*Pavlenkova*, 1980; *Krasnopevtseva*, 1984]. For the Terek-Caspian region, the best fit between the calculated and observed data at the surface was obtained when we assumed 3 orders of magnitude viscosity drop in the zones of Oligo-Miocene Maykopian clays and in the zones of highly deformed Terek and Sundja ridges (Figure 2a), where rock strength is reduced.

[40] Figure 4 displays the geometry and properties of the crust and upper mantle used in the final model calculations. For finite element purposes the cross sections comprise a 19 × 37 grid (vertical by horizontal) with grid intervals along z and x of 1 and 10 km, respectively. The base of the model is at 18 km. It was chosen so because it meets the requirements that the thickness of the model should be many times less than its length, which was equal to 360 km (see section 3). Moreover, it is the maximum depth for which reliable structural information is available.

[41] The position of the free mantle level depends on average density of the crust and density contrast at the crust-mantle discontinuity. According to the gravity modeling of the Greater Caucasus region [*Artemiev et al.*, 1985], we placed this level at a depth of 6 km for the Terek-Caspian and Stavropol profiles and at 8 km for the Azov-Kuban profile. According to *Smolyaninova et al.* [1996], the exact position of the free mantle level does not qualitatively change the results, as long as the free mantle level does not exceed the thickness of the model.

5. Results of Numerical Modeling

[42] Figure 5 shows the best fitting slow flow pattern results for the model profiles. The arrow length at each point is proportional to the computed velocity. The displayed vectors are drawn relative to a fixed point of zero velocity, indicated by a square. These points were placed at the depth of the free mantle level.

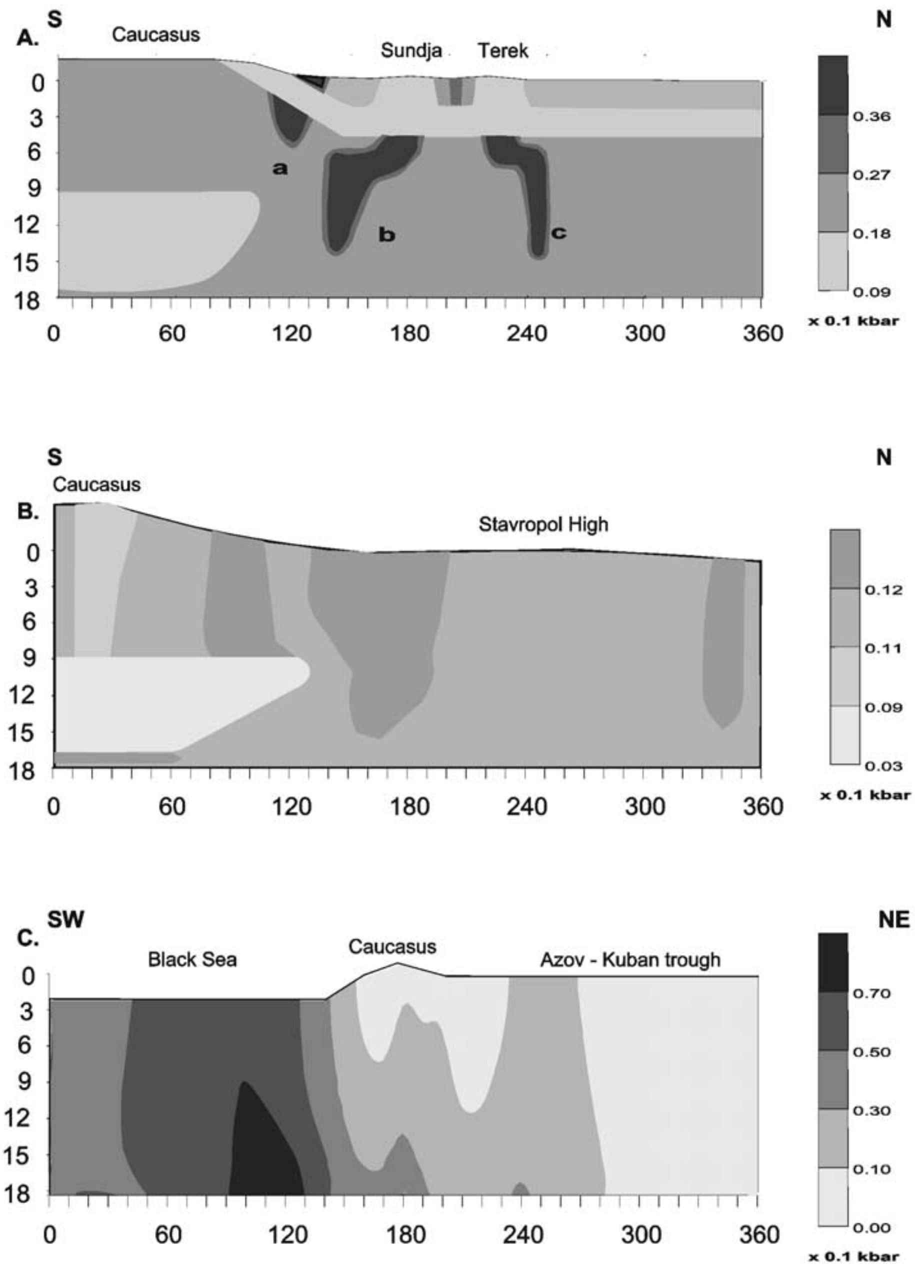

Figure 6. Calculated total shear stress (the square root of the sum of the squares of the deviatoric stress tensor components) for profiles A, B, C (Figure 1). Mention that scales used to show stress fields in sections A, B, and C are different. Zone of high values of the shear stress in profile A marked by "a" coincides with the Vladikavkaz fault. Higher shear stress values in profile C marks the Tuapse trough and southern border of the Azov-Kuban trough.

[43] The calculated boundary conditions imply simple compression on the side boundaries of the model at the rate of 1 cm/yr for profiles running across the Terek-Caspian trough and the Stavropol high. In contrast, rather complex boundary conditions were obtained for profile C (Figure 5). They incorporate extension at the side boundaries and some effects at the bottom of the model, which can be explained as uplifting flows of the mantle material under the Black Sea and downwarping flows under the western Greater Caucasus and Azov-Kuban trough.

[44] Figure 6 shows the total shear stress for the profiles (the square root of the sum of the squares of the deviatoric stress tensor components). For the Terek-Caspian trough (Figure 6a) one zone of maximum calculated shear stress coincides with the main Vladikavkaz fault (marked by letter "a"), and two other zones are disposed within basement

Table 1. Results of the Focal Mechanisms Inversion[a]

Subregions	s_1	s_2	s_3	R
Azov-Kuban (profile C)	N355° E5°	N235° E81°	N86° E8°	0.57
Stavropol High (profile B)	N12° E7°	N281° E6°	N151° E80°	0.78
Terek (profile A)	N11° E8°	N280° E5°	N158° E81°	0.77
Dagestan	N209° E1°	N119° E1°	N338° E89°	0.73

[a] Deviatoric stresses are as follows: s_1, compressional; s_2, intermediate; s_3, extensional; $R = s_2 - s_1/s_3 - s_1$, shape ratio.

under the foredeep ("b" and "c"). In the lighter zones that correspond to lower-viscosity areas, stresses are low and the deformation rate is high. These zones coincide with the Terek and Sundja anticlines and the zone of low seismic velocity under the Caucasus.

[45] Stresses in the central part of the foredeep (Figure 6b) are comparatively low. Maximum values here are at the foot of the Caucasus and are framing the Stavropol high. These maximum values are much less than the maximum values for the eastern and western parts of the North Caucasus foredeep.

[46] Figure 6c shows distribution of total shear stress component for the Azov-Kuban profile. The maximum value of this stress component is at the foot of the Greater Caucasus in the Black Sea. This maximum coincides with the Tuapse trough (Figure 1), an area which is very well investigated by reflection seismics [*Tugolesov*, 1985] and is characterized by many active faults and folds and high seismicity. According to our calculations, the Black Sea depression and the Azov-Kuban trough appeared to be exposed to extension, which is much stronger in the Black Sea, while the area of maximum compression correlates with the Greater Caucasus belt. The zone of transition from compression to extension corresponds in the south to the Tuapse trough, and in the north to the southern part of the Azov-Kuban trough.

6. Results of Focal Mechanism Inversion

[47] Present-day state of stress along the Greater Caucasus has been determined using inversion of focal mechanisms of shallow events, which were selected from the UIPE database, which spans from 1939 onward. The inversion procedure that was used is the one, which has been developed at Orsay University [*Carey-Gailhairdis and Mercier*, 1987, 1992]. As with other stress calculations, inversion is based on a very simple mechanic assumption, which considers that the shear stress resolved on a fault plane is parallel and of the same sense as the fault slip. Additionally for focal mechanisms, the procedure permits, prior to the inversion, to select one of the two nodal planes as the fault plane.

[48] Inversions of focal solutions essentially permit determination of the present-day stress regime of the Northern Caucasus. In fact, a major problem arises from the scarcity of seismic events in the studied region. Most of the few central and western Great Caucasus focal solutions are located along the southern part of the mountain range (see previous section). Indeed, the eastern Greater Caucasus; i.e., mainly Dagestan, is the only area where there is significant seismic activity and consequently many available focal solutions. Basically, focal mechanism inversions were performed within four sub-regions of the Northern Caucasus: western ("Azov-Kuban"), central ("Stavropol High"), eastern ("Terek"), and easternmost ("Dagestan"). Inversion results presented in Table 1 yield N-S to NE-SW trending compression axes; the compression orientations shift eastward from N-S to NE-SW. The obtained results demonstrate thus that the stress regime in the Northern Caucasus is compressional and nearly orthogonal to the strike of the mountain belt. This is in very good agreement with the overall stress field determined in our modeling (see previous section) and also justifies the use of a 2-D approach to this modeling.

7. Discussion

[49] It was already mentioned above that the distribution of regional stress, obtained as a result of the inverse problem solution, is mainly determined by mechanical properties of the lithosphere (effective viscosity, in this case). In particular, the location of the zone of lower effective viscosity under the Greater Caucasus (Figures 4a and 4b) made it possible to explain their uplift by intraplate compression applied at the side boundaries of the study area. This zone was located by seismic and confirmed by heat flow data. The Central and Eastern parts of the Caucasus are characterized by higher values of heat flow up to 70–80 mWt/m^2 [*Cermak and Hurtig*, 1979] and are areas of volcanic activity. These high values cannot be explained exclusively by erosion and uplifting of the Greater Caucasus but most likely are an indication of higher temperature in the middle crust and probably in the lower crust and upper mantle. This would lead to reduced rock strength [*Ranalli*, 1995]. When going westward, heat flow becomes considerably less (<50 mW/m^2) and no volcanic activity is observed. This can be a possible explanation for the absence of a low seismic velocity (weak) zone in profile C.

[50] In general, we can conclude that by varying the distribution of mechanical properties in the lithosphere we can change our estimates of regional stress within certain limits. For example, it appears impossible to find out any distribution of mechanical properties that can help explain the observed neotectonic movements along profile C, thus suggesting that only intraplate compression or extension are presently active. In many cases, when we deal with comparatively extended areas of alternating uplift and subsidence, it is necessary to involve processes acting at the

bottom of the model, such as various movements in the mantle, phase transitions, etc. The alternating areas of uplift and subsidence at the surface of the crust may be due to small-scale convection in the asthenosphere (see above). Formation of orogenic belts and rift zones is associated with considerable horizontal variations of pressure and temperature within the lithosphere and upper mantle. Numerical modeling demonstrates, that recovering of equilibrium is accompanied by development of convective cells in the asthenosphere [*Mikhailov et al.*, 1996, 1999a]. This small-scale convection produces areas of local compression and extension in the lithosphere; as a consequence, the inner structure of rifts and orogens becomes more complicated. Direction of flow in the small-scale convection cells in the areas of a transition from comparatively thin to a thicker lithosphere, matches the slow flow pattern in Figure 5c, i.e., upwelling streams are under the areas with thin crust (Black Sea depression) and downwarping is under the comparatively thick crust (western Greater Caucasus). A similar picture was obtained also when modeling modern movements and state of stress for the northern Black Sea-Crimea mountains area [*Smolyaninova et al.*, 1996].

[51] Distribution of shear stress obtained from modeling favor low stress contrast in the western and central parts of the North Caucasus foredeep (Figures 6b and 6c). Almost the whole foredeep appears to be under compression, except its western part, Azov-Kuban, where stresses are close to zero. Average shear stress in the crust becomes less westward.

[52] The obtained results are in agreement with present-day estimates of strain rates obtained from seismological and GPS data. In particular, compilation of GPS data for the Caucasus region [*Reilinger et al.*, 1997] shows that part of the shortening in the Caucasus region is accommodated to the north of the main thrust zone of the Great Caucasus. Although data for this area are not very well constrained, showing small displacements in comparison to the error bars, we can conclude that the shortening rate of 10 mm/yr estimated by numerical modeling for the central and eastern parts of the foredeep does not contradict these GPS data.

[53] In comparison with profiles B and C (Figure 6), profile A demonstrates sharper stress contrasts, which are in agreement with an increase in the level of seismic activity in this part of the North Caucasus region. The position of the high shear stress zones marked in Figure 6a as "b" and "c" is governed by the location of vertical low-viscosity zones in the vicinity of the Terek and Sundja ridges. These zones were introduced to take into account westward propagation of active folding there. Because the crust of the central part of the foredeep was considered to be homogeneous (Figure 4), the location of these zones is not well constrained and cannot be related to any specific deep-seated faults (existence of possible peculiarities of mechanical properties may cause perturbations of the stress field). These high shear stress zones suggest only that some compressional deformations are accommodated within the basement of the Terek-Caspian foredeep basin. High shear zone "a" in Figure 6 is better constrained, as this zone appears to be due to the interaction of two factors: the variation of the regional topography and position of the low-viscosity zone under the Greater Caucasus. Thus zone "a" may be correlated with the Vladikavkaz fault zone – the main boundary thrust of the northern side of the Greater Caucasus.

[54] Comparing the length of velocity vectors for the northern and southern parts of the foredeep along all the three profiles (Figure 5), we can conclude that the strain rate along the North Caucasus foredeep increases eastward. Variations in the length of velocity vectors at the northern side of the Greater Caucasus increases from profile "C" to "A". In the Azov-Kuban trough, deformations are small, as the length of the velocity vectors remains nearly constant. In the western Caucasus (central part of the profile "A"), the absolute values of the velocity vectors also show negligibly small variation, thus most of the shortening should occur at the southern slope of the Greater Caucasus, even if a small part of deformations may be accommodated at the northern slope. These results are in good agreement with the distribution of seismicity in this region (Figure 1). Farther to the east, the zone of comparatively high deformations becomes wider (profile B), occupying the northern gentle slope of the Greater Caucasus and the southern part of the Stavropol high. Some shortening seems to be accommodated within the Stavropol high as well. On profile A, the zone of high deformations becomes more extended, correlating with the fact that seismic events in this area stretch further to the north.

[55] It should be emphasized that we used the Quaternary data on vertical movements at the surface according to which the Terek-Caspian trough is an uplifting area. It is obvious that before the Quaternary this area mainly subsided. Thus, during the Quaternary, compression may have extended considerably northward. Formation of the Terek and Sundja ridges supports this statement.

[56] Strictly speaking the regional-scale problem for the whole North Caucasus foredeep calls for 3-D modeling. In the present case we resolved only the projections of acting forces (of rates determined movements causing these forces) in the direction of the profiles.

8. Conclusion

[57] We applied the numerical method of quantitative estimation of regional and local stress field using data on velocity of vertical movements at the top of the lithosphere to investigate deformations and state of stress in the North Caucasus region. The results presented are first-order estimates. A more detailed result could be obtained with more data on the deep structure and composition of the lithosphere and on the recent tectonic evolution of the region, which may be available with the next CPD profile running across Stavropol high and the Great Caucasus ridge to be completed in 2003.

[58] Our analysis shows that during the late Quaternary the western, eastern and central parts of the North Caucasus foredeep were exposed to differing regional tectonic

forces. Geodynamics of the eastern and central parts has been controlled by regional compression due to intraplate forces acting in an approximately N-S direction. The western part has been strongly influenced by tectonic processes taking place in the Eastern Black Sea depression, which are likely governed by upwelling of upper mantle material below the Eastern Black Sea, and by its downwarping below the western Greater Caucasus. The calculated slow flow patterns and stress distributions within the crust are in good agreement with GPS and seismological data.

[59] **Acknowledgments.** This work started as a part of the INTAS 94-3881 project. Financial support of INTAS is kindly appreciated. Funding of this research was provided also by the Russian Foundation for Basic Research (Project N98-05-64621) as well as the Peri Tethys Program. Critical comments of Yu.Yu. Podladchikov and Philippe Davy are greatly acknowledged. We also thank Pilar Garcia for correction of English and useful comments and suggestions.

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