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GEOCHEMISTRY

Mathematical Modeling of Overthrusting Fault as a Cause of Andalusite–Kyanite Metamorphic Zoning in the Yenisei Ridge

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Regional structural features of the Yenisei Ridge are traditionally displayed as a NW-trending linear system of tectonic sheets that are thrust over one another and separated by faults where collision of blocks occurs (Fig. 1). As a result, low- and moderate-pressure series of regional metamorphism turn out to be juxtaposed. Moderate-pressure metamorphism related to collision [1] is locally superimposed on low-pressure metamorphic rocks. Therefore, andalusite is progressively replaced with kyanite, and new mineral assemblages of moderate pressure and deformational microstructures are formed. The replacement of andalusite by kyanite in the Yenisei Ridge is not a common phenomenon, because the stationary continental geotherm, as a rule, does not intersect the andalusite–kyanite equilibrium line.

According to the results of mineralogical–petrographic and structural studies and the geothermobarometric data obtained for Fe–Al-metapelites in the Transangara part of the Yenisei Ridge (basins of the Eruda and Chirimba rivers), Middle Riphean (~1100 Ma) low-pressure metamorphism proceeded at 3.5–4.0 kbar and 540–560°C. This was recorded in mineral assemblages of greenschist and epidote–amphibolite facies of the Korda Formation. Subsequently, these rocks were affected by the Late Riphean (~850 Ma) collision-related prograde metamorphism near the Panimba overthrust (P = 4.5–6.7 kbar; T = 540–600°C). Upon approaching the thrust, one can recognize three zones of superimposed amphibolite-facies metamorphism of moderate pressure. They are distinguished by proportions of relict and newly formed minerals and by the degree of deformation. In total, they make up a band (5–7 km wide and no less than 20 km long), which is bounded in the east by the thrust. The Mesoproterozoic metacarbonate rocks of the Penchenga Formation (~1600 Ma) occur behind this fault. The calculation of reactions that proceed with great volumetric and small entropic effects [3] and the calculated PT trends of the evolution confirm the gradual increase in pressure (from 1.0 to 2.5 kbar) applied to metapelites of the Korda Formation upon approaching the thrust—without a significant rise in temperature (no higher than 20 ± 15°C). This testifies to the almost isothermal subsidence of rocks during collision of crustal blocks.

To explain the origin of the observed metamorphic zoning, we have proposed a tectonic model and made essential thermophysical calculations that take into account the real physical parameters of metapelites and metacarbonates (radioactive heat production and thermal conductivity) [4, 5]. The gradually increasing lithostatic pressure was attributed to tectonic thickening of the earth's crust in the Panimba overthrust zone. Consequently, metapelites of the Korda Formation were overlapped by metacarbonates of the Penchenga Formation 5–7 km thick. The overthrusting without appreciable temperature rise is related to the specific behavior of stationary geotherms in different rock types with contrasting heat-producing and thermophysical properties. Our model explains the character of metamorphic evolution of metapelites (the gradual replacement of andalusite by kyanite, the enrichment of garnet grains in the grossular component from the center to the edge of grains, the gradual increase in lithostatic pressure with an insignificant rise in temperature, and so on) [5].

Advancing the proposed interpretation of the collision-related metamorphism in the 1D approximation [5, 6], we have developed a 2D model of deformation of the lithosphere by thrusting.

The principal difficulties in mathematical modeling of geological processes are related to considering the substantial nonlinearity of equations for the mechanics

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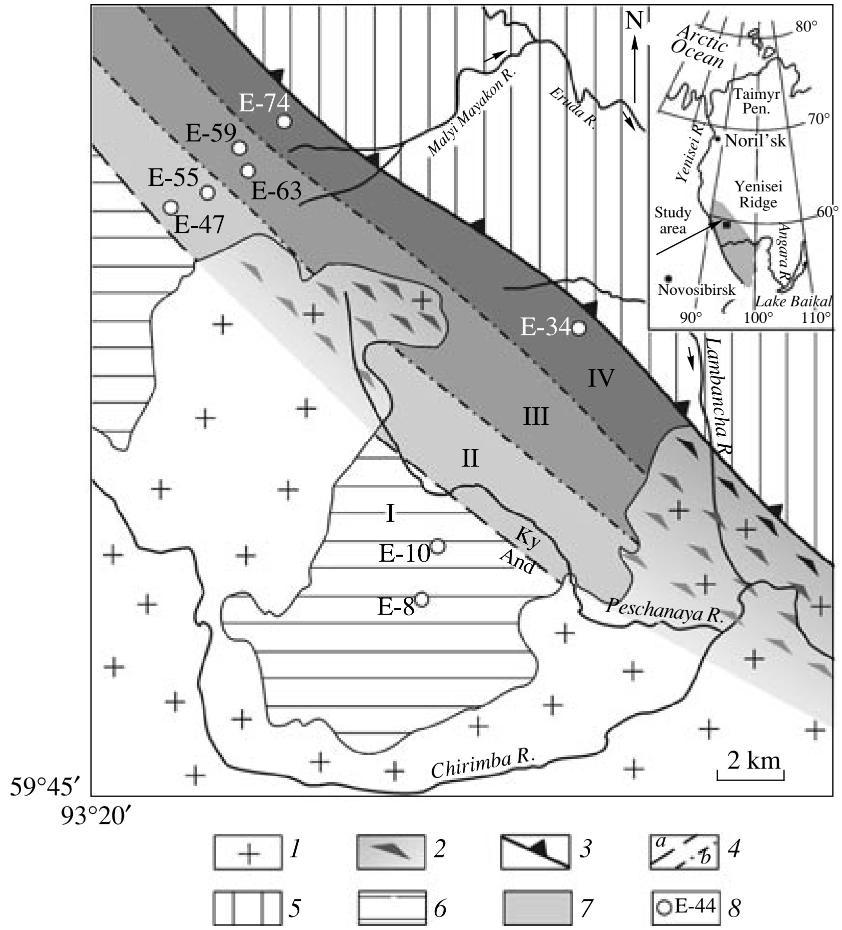

Fig. 1. Schematic geological map indicating location of collision-related metamorphic zones at the Yeruda and Chirimba interfluve, Transangara region of the Yenisei Ridge [1]. (*1*) Granite of the Chirimba pluton; (*2*) cataclastic and blastocataclastic zones in granite; (*3*) suture of the Panimba overthrust (ticks are oriented down the dip); (*4a*) andalusite–kyanite isograde and (*4b*) boundaries between metamorphic zones in metapelitic rocks; (*5*) metasiltstones and metacarbonate rocks of the Penchenga Formation; (*6*) metapelites of the Korda Formation (products of regional metamorphism); (*7*) collision-related metamorphic zones in metapelitic rocks; (*8*) sample location.

of deformed solids (MDS) for an adequate description of real geological processes. One can recognize three main (physical, contact, and geometric) types of nonlinearity of MDS equations. Physical nonlinearity implies the nonlinearity of determining relationships: equations of the link between components of stress and strain tensors and/or their rates. Contact nonlinearity means a procedure for determining an apriori unknown contact boundary and contact forces for the bodies coming into contact. Finally, geometric nonlinearity comes into play when we take into account the variation of body geometry in the deformation process under given loads and displacements.

In this work, we solve problems with consideration of physical and contact nonlinearities, ruling out geometric nonlinearity of deformation.

The equations of quasistatic deformation that describe geological processes are solved by numerical methods. The method of finite elements is used for spatial discretization of the equations [7–12]. This method is the most suitable for the numerical solution of problems of mathematical physics with arbitrary domains. The idea of the method consists in approximation of unknown functions by interpolating polynomials determined in local sites (finite elements). Integration of the equations by time is carried out with a stepwise iteration refinement of the solution by the Newton–Rafson method. Since the deformation rate is relatively low, the MDS problem is solved as a quasistatic one; i.e., the inertial terms in equations of motion are omitted. The equations of quasistatic deformation of solids are given in [7]. The mathematical modeling was carried out with a MSC.Marc 2005 software that takes into consideration the required types of nonlinearity of MDS equations. The geometric modeling and stepwise integrations were performed with the MSC.Patran 2005 and MSC.Marc 2005 software packages, respectively [8].

The thrusting of a plate over absolutely rigid body was modeled in approximation of the 2D plane strain

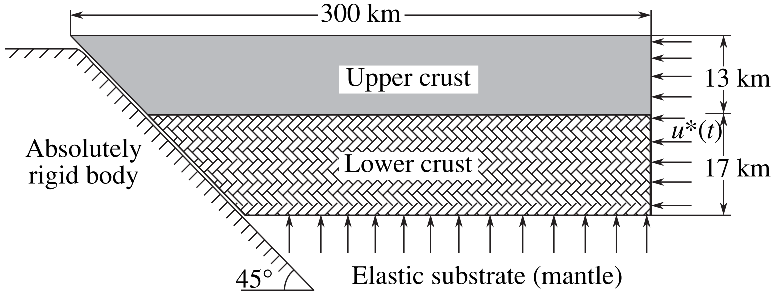

Fig. 2. Boundary conditions, initial geometric parameters of the plate before loading, and scheme of mechanical influence.

state. It was assumed that the continental lithosphere is a domain (300 × 30 km) of two blocks, which were jointly subjected to lateral compression with a rate of 0.15 cm/yr. The domain simulating a thrust was bounded by the contact between the converging blocks, one of which was set as a rigid immobile buttress with a given slope of the contact. The geometric parameters of the deformed plate and boundary condition of the problem are presented in Fig. 2.

Special attention was paid to the choice of rheological parameters that would adequately describe the upper lithosphere. The material of the lithosphere was deemed an initially elastic substance that takes into account rock failure in the process of deformation. The physical parameters of the material were chosen as corresponding to the properties of the upper and lower crust. It was assumed that the moving plate consists of two layers with different properties (the brittle upper layer and the elastoplastic lower layer) in rigid cohesion. The upper layer is simulated by initially isotropic elastic material that undergoes brittle failure at the critical cracking stress σ_t. Further deformation occurs on the decreasing segment of the diagram of uniaxial deformation of the material with the tension-softening modulus E_S [8]. The lower crust is simulated by an elastoplastic material with work hardening. In the course of plastic deformation, the hardening was given by tangent module E_t [7].

The modeling of a plate thrusting over an absolutely rigid body was realized in the following way. Motion and deformation of the plate were considered within a time interval of 50 Ma. It was assumed that the body force corresponding to the plate weight varied from the zero value to the final value $P = \rho g$ (N/m^3) over the time interval necessary for reaching the equilibrium state ($\rho = 3000$ kg/m^3 is the average density of the crust, and $g = 9.81$ m/s^2 is the acceleration of gravity). In our case, this interval was 50 yr. Further, from the moment of complete application of the gravity force ($t = 50$ yr) to the moment of 50 Ma, we set a proportional displacement $u^*(t)$ of the right edge of the plate (block) from 0 to 75 km; i.e., the right block moves with a constant velocity $V \approx 0.15$ cm/yr. It is assumed that the lower surface of the plate is supported by an elastic substrate with rigidity K_f (Winkler boundary condition [13]), which linearly increases from 0 to 1.2 MPa and afterwards remains constant, like a weight. Thus, we simulate the hydrostatic pressure of the mantle on the base of the moving plate. The upper surface of the plate is considered to be free of stress. When the plate comes into contact with an absolutely rigid body, the left surface of the plate at the contact is subjected to corresponding forces. The plate interacts with the absolutely rigid body in contact zones, which were determined by the solution of the problem with an apriori unknown contact boundary. At the initial moment ($t = 0$), the absolutely rigid body and plate were separated by a horizontal gap (100 m), which was fitted in such a way that it provided the contact between interacting plates (blocks) after application of load.

Material of the lower layer of the plate is supposed to be elastoplastic with isotropic strengthening [7] and the following mechanical constants: Young's modulus $E = 50$ GPa, Poisson's ratio $\nu = 0.25$, yield stress $\sigma_y = 0.5$ GPa, and tangent modulus of elasticity $E_t = 0.02E = 1$ GPa. The material of the upper layer is supposed to be initially isotropic (linearly elastic) with a brittle failure capacity. Uniaxial stress–strain diagram is characterized by the tension-softening modulus $E_s = 1000$ Pa. We applied the following mechanical constants: Young's modulus $E = 5$ GPa, Poisson's ratio $\nu = 0.25$, and ultimate stress at tension $\sigma_t = 0.02$ GPa.

The calculations were carried out with an adaptive time step. Note that the loading modeling includes all the considered mechanisms of nonlinearity: plastic deformation of the lower crust, development of fractures in the overriding plate, and its contact interaction with an absolutely rigid body (the left margin of the plate overlies the absolutely rigid body). The stress-and-strain configuration of the plate as a result of complete loading was accepted as the initial state that marked the onset of displacement of the right margin of the plate.

Figures 3 and 4 show the deformed configurations of the moving plate at various moments. They illustrate the results of modeling of the overriding plate and the uplift of its left margin in the absence of erosion, and all the destroyed elements of the overthrusted sequence are

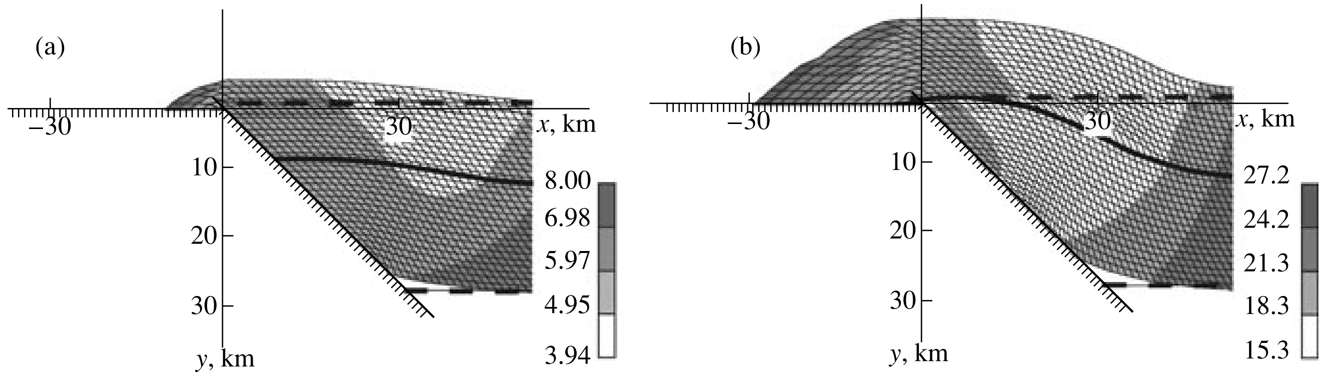

Fig. 3. Contour lines of displacements (km) of the deformed plate at different moments: (a) $t = 1.555 \cdot 10^7$ yr; (b) $t = 5 \cdot 10^7$ yr. The solid line is the boundary between the upper elastic-brittle and the lower elastoplastic crust; the dash line indicates the initial position of the plate.

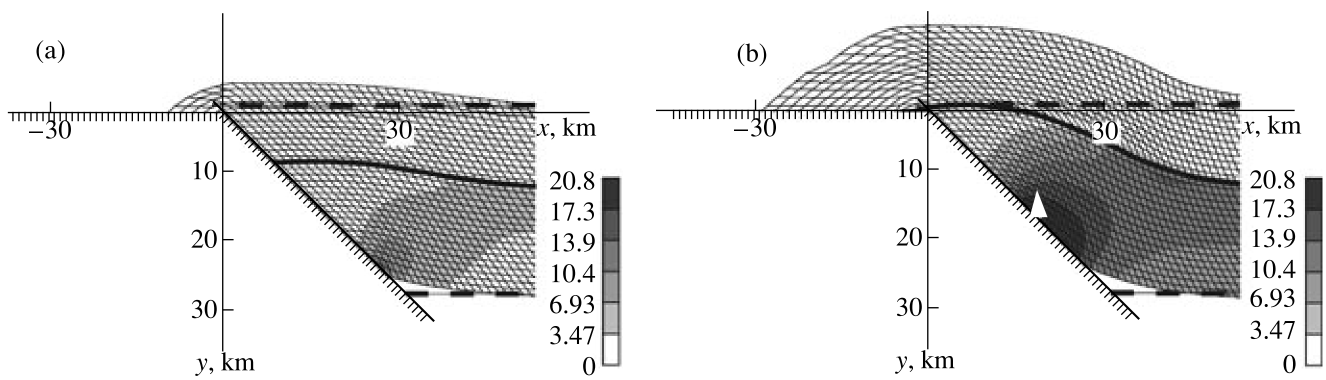

Fig. 4. Contour lines of effective stress (GPa) of the deformed plate at different moments: (a) $t = 1.555 \cdot 10^7$ yr; (b) $t = 5 \cdot 10^7$ yr. Triangle indicates the maximum stress region.

retained. The boundary between the upper and the lower layers is shown by the heavy line in Figs. 3 and 4. At $t = 1.555 \cdot 10^7$ yr, the left upper point of the plate came into contact with the upper surface of the rigid body and began to slip along this surface. Figure 3 shows the deformed configurations of the plate at $t = 1.555 \cdot 10^7$ and $5 \cdot 10^7$ yr and the levels of displacements. Figure 4 demonstrates the distribution of effective stress σ_e [7] at the same moments. In Fig. 4b, the upper value of the scale is accepted as 20.8 GPa, and the maximum effective stress (44.5 GPa) is marked by a triangle to provide more detailed visualization. The thrust rate was estimated at ~300 m/Ma on the basis of petrological data, which indicates an almost isothermal pressure released in the course of evolution of the metamorphic complex near the Panimba overthrust in the Yenisei Ridge [5]. This assessment is consistent with estimates based on modeling of the Tien Shan and Himalayan orogenic systems [14, 15]. Comparison of petrological and model estimates of rock exhumation in the course of thrusting of one plate over another was one of the tasks of the modeling. As follows from Fig. 3, the displacement in the thrust zone was $15.3 \cdot 10^3$ m over the course of 50 Ma; i.e., the thrust rate is 300 m/Ma, which is consistent with estimates based on mineralogical geothermobarometry. The maximum stress is recorded at the contact between the overriding and underlying plates in the lower crust. The values of maximum stress depend on the rheology of the lower crust.

Thus, based on petrological and geological data, we carried out mathematical modeling of thrusting of a thick rock sequence over the underlying plate. The thrust rate is comparable to the value based on an almost isothermal increase in pressure in the process of collision-related metamorphic zoning near the Panimba overthrust in the Yenisei Ridge [5]. The modeling yielded estimates of the stress field in the thrust zone and displacements of points in the course of deformation of the overriding plate. The structure of the deformed crust in the thrust zone is qualitatively consistent with the imbricated nappe structure of the Alps and Himalayas. The length of the nappe is ~90 km and approximately corresponds to the displacement of the distal margin of the lithospheric plate. The modeling calculations show that the displacement of thrust front for 25–30 km in 50 Ma produced a crustal uplift zone 80–90 km wide. The vertical displacement of the upper surface and M boundary in the distal zone of the thrust zone is 800–900 m. Thus, the relative thickening of the plate under compression was no more than 1.8/30 = 0.06. The complete erosion of the thrust zone might provide exhumation of rocks from a depth of no less than 15 km. This value is controlled by the rate of plate collision.

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REFERENCES

1. V. V. Reverdatto and V. S. Sheplev, Geol. Geofiz. **39**, 1679 (1998).
2. I. I. Likhanov, O. P. Polyansky, P. S. Kozlov, et al., Dokl. Akad. Nauk **375**, 509 (2000) [Dokl. Earth Sci. **375A**, 1411 (2000)].
3. I. I. Likhanov and V. V. Reverdatto, Petrologiya **10**, 541 (2002) [Petrology **10**, 479 (2002)].
4. I. I. Likhanov, O. P. Polyansky, V. V. Reverdatto, et al., Geol. Geofiz. **42**, 1205 (2001).
5. I. I. Likhanov, O. P. Polyansky, V. V. Reverdatto, and I. Memmi, J. Metamorph. Geol. **22**, 743 (2004).
6. I. I. Likhanov, O. P. Polyansky, and V. V. Reverdatto, in *Tectonics and Geodynamics of the Continental Lithosphere* (GEOS, Moscow, 2003), pp. 348–351 [in Russian].
7. S. N. Korobeinikov, *Nonlinear Deformation of Solids* (Sib. Otd. Ross. Akad.Nauk, Novosibirsk, 2000) [in Russian].
8. *MARC Users Guide. Volume A. Theory and User Information* (MSC Software Corp., Santa Ana, 2005).
9. O. C. Zienkiewich and R. L. Taylor, *The Finite Element Method* (McGraw-Hill, London, 1997).
10. K.-J. Bathe, *Finite Element Procedure in Engineering Analysis* (Prentice Hall, Englewood Cliffs, 1982).
11. A. I. Golovanov and D. V. Berezhnoi, *The Finite Element Method in Mechanics of Deformed Solids* (DAS, Kazan, 2001) [in Russian].
12. B. D. Annin, S. N. Korobeinikov, and A. V. Shutov, in *Strain State and Seismicity of the Lithosphere* (Sib Otd. Ross. Akad. Nauk, Novosibirsk, 2003) [in Russian].
13. Yu. N. Rabotnov, *Mechanics of Deformed Solid* (Nauka, Moscow, 1979) [in Russian].
14. J. P. Avouac and E. P. Burov, J. Geophys. Res. **101**, 17747 (1996).
15. O. Vanderhaeghe, S. Medvedev, P. Fullsack, et al., Geophys. J. Int. **153**, 27 (2003).