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GEOLOGY

Petrological Data Allow Estimating the Amplitudes of Crustal Uplifts Caused by Retrograde Metamorphism

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Abstract—Analysis of the morphology of the recent uplifts on Precambrian cratons and geological–geophysical data on the structure of the crust and mantle indicate that these structures were formed due to expansion of the crustal rocks as a result of retrograde metamorphism. This occurred due to the contribution of large volumes of deep fluids to the complexes of the Early Proterozoic rocks, which underwent high-grade metamorphism in the lower crust. Later, these complexes were moved to the shallower depths after denudation of thick overlap sequences from the craton's surfaces. The calculation of the volumetric expansion effects using *P–T* diagrams for the main types of metamorphic rocks shows that this mechanism could have prompted the uplifts of the Precambrian crust in recent time with amplitudes from 100–200 to 1000–1500 m.

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Rapid crustal uplifts with the amplitudes from 100–200 m to 4–5 km occurred in the Pliocene–Quaternary on ~90% of the total area of continents [1, 2]. These uplifts have formed most of the present-day mountains and high plateaus including old crystalline shields. Judging from the huge total volume of the relief formed, the recent uplifts are the most powerful event in the continental lithosphere. They were abundant on the Precambrian crust with amplitudes from 100–200 m on the Precambrian cratons (East European Craton and northwest Africa) and to 1000–1500 m on the northwestern Siberian Craton, Aldan Shield, East Antarctic, and eastern and southeastern Africa [1, 2].

The Precambrian crust is widespread on most continents (~70% of their total area), and the nature of their recent uplifts is of great interest. Large uplifts of the continental crust are most often explained by its strong shortening; however, no shortening occurred on the Precambrian cratons during the last 0.5 Ga. Vertical lithospheric displacements are also explained by mantle convective flows, which forming the dynamic topography at the top of the asthenosphere.

Modeling of this event [3] leads to the results that are quite different from the real distribution of the recent uplifts on the continents [4]. The thickness of the mantle part of the lithosphere beneath the Precambrian cratons is rather high (150–250 km) [5]. This excludes the occurrence of the recent uplifts due to delamination of the dense mantle lithosphere or astenospheric upwelling to the base of the crust. In most cratons, the crust is underlain by a thick layer of the mantle lithosphere and not by a hot material [5].

The recent uplifts cannot be explained by the ascent of melt to the crust, which was extracted from mantle plumes (magmatic underplating). Only in some areas (e.g., Cameroon Volcanic Line and northeast Africa) do asthenospheric bulges reach the base of the crust.

Under these conditions, expansion of rocks in the crust can be proposed as the most likely cause for the recent uplifts on Precambrian cratons [4, 6]. This is also evident from strong lateral heterogeneities of the uplifts. On the maps of the recent uplifts [1, 2], they are often bounded by steep slopes from 100 m to 1 km high and from 20–30 to 100 km wide [6]. Their formation indicates the expansion of rocks at low depths (within the crust). The expansion was explained by metamorphism, which has been manifested over several million years as a result of the infiltration of a large volume of mantle fluids to the crust [4, 6].

Rocks, which were formed at the pressure of 0.2–0.8 (locally, up to 1.0) GPa, typically occur on the surface of the basement of the Precambrian cratons (e.g.,

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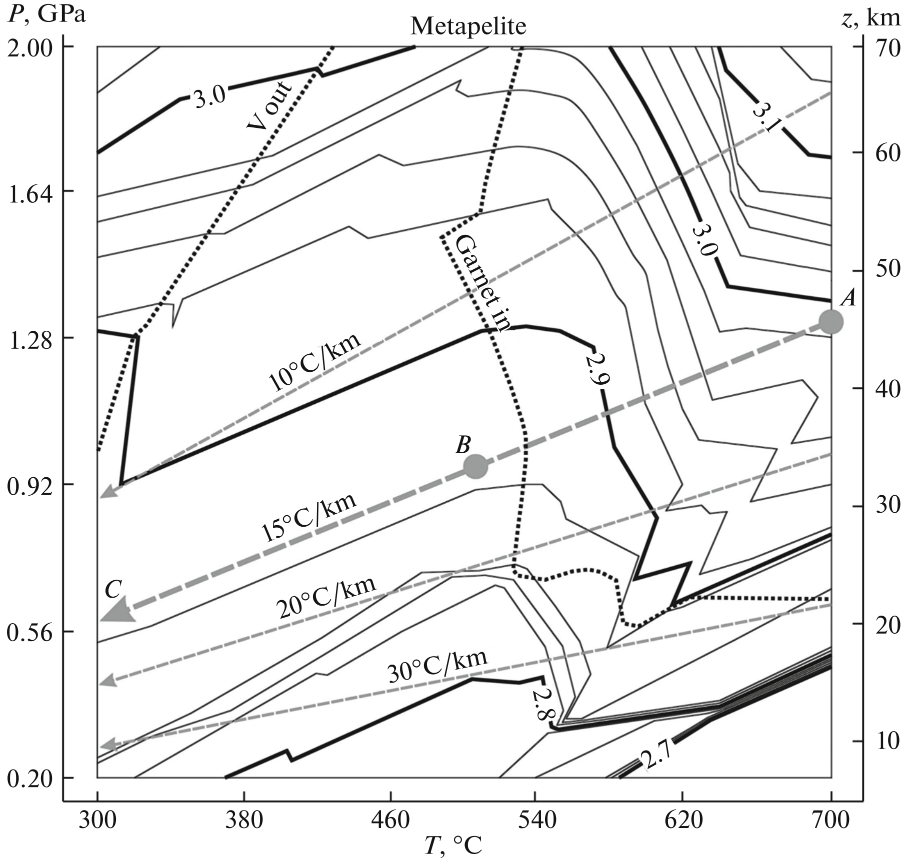

Fig. 1. *P–T* parameters of metamorphism for metapelite with isochores (solid lines) calculated using the PERPLE_X program (here and in Figs. 2, 3). The modal composition of a Mix-2 fine-grained mixture is as follows (wt %): quartz 18, illite 45, chlorite 22, plagioclase 12, accessory minerals (mostly Fe and Ti oxides) 3 [8]. The step of the density scale (difference between isochore lines) is 0.02 g/cm^3. Here and in Fig. 3, the bulk chemical composition of mixture is after [8]. The dotted line shows the boundaries of *P–T* areas without a free water-bearing fluid phase and with garnet-bearing assemblages. Dashed lines with arrows show the trajectories corresponding to exhumation of metamorphic rocks under conditions with different geothermal gradients.

Kola Peninsula and Aldan Shield) [7]. This indicates that 7 to 35 km of rocks were denuded from the surface of the cratons. In this case, the middle and lower crust of the cratons should include the rocks, which were formed 7–35 km deeper (as well as at significantly higher *P* and *T*) than in the present-day position.

Figures 1–3 show the *P–T* diagrams for the main types of metamorphic rocks of the continental crust, metapelites, metamafic rocks, and gneisses with isochores of densities calculated in the PERPLE_X program [8, 9]. In the right-hand upper part of the diagram at high *P* and *T* typical of the lower crust, the density of rocks is significantly higher than in their left-hand part, where *P* and *T* correspond to the middle crust. The rocks are slowly rising because of the surface denudation, and the rocks of the lower crust reach the areas of lower *P* and *T*, where their density should be much less under the state of equilibrium. The new state of equilibrium of rocks should be the result of metamorphism, which is rapid only in a presence of fluids [10]. The dry rocks emerged to shallower depths remain metastable and preserve their high initial density. The infiltration of a large volume of fluid into the rocks enhances rapid secondary metamorphism (diaphthoresis) [11], which is accompanied by hydration of rock-forming minerals, expansion of rocks, and upwarping of the crustal surface.

The magnitude of uplifts can be estimated using the *P–T* diagrams with isolines of the rock density ρ, g/cm^3. For example, the average density of metapelite ρ_1 is ~3.0 and ρ_2 is ~2.9 g/cm^3 in the right-hand upper and lower parts of the diagram (in the lower and middle crust), respectively (Fig. 1). Let us assume that a rock layer in the lower crust with average density ρ_1 and h_1 thickness emerged as a result of surface denudation to the middle crust, remained dry, and preseve its initial density. The infiltration of the fluids into the crust can lead to metamorphism of the layer. Let us suggest that the average density of the rock in the layer in the new state of equilibrium decreased to ρ_2, whereas the thickness of the layer increased to h_2, because the mass of the substance of the layer remains

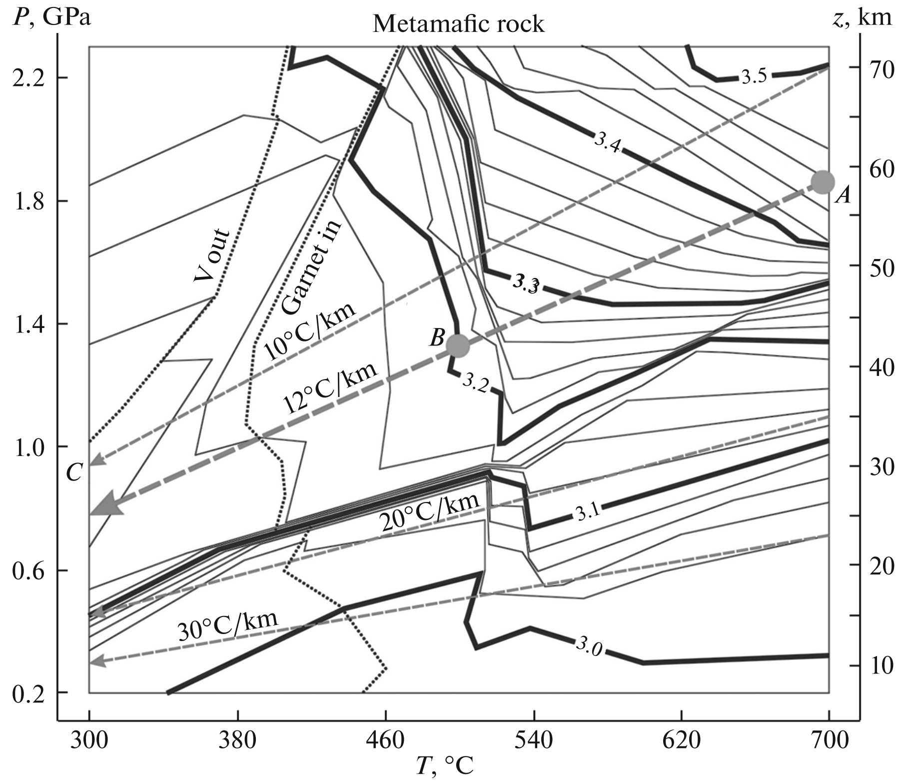

Fig. 2. *P–T* parameters of metamorphism of metamafic rock and isochores. The bulk chemical composition of the mixture is after [9]. For symbols, see Fig. 1.

constant ($\rho_1 h_1 = \rho_2 h_2$), rather than $h_2 = (\rho_1/\rho_2)h_1$. The expansion of the layer results in the uplift on the Earth's surface $\zeta = h_2 - h_1$ or

$$\zeta = [(\rho_1 - \rho_2)/\rho_2]h_1. \quad (1)$$

For the above values of ρ_1 and ρ_2 and the initial thickness of the layer h_1 = 15 km, ζ is ~500 m. Similar Neotectonic uplifts occurred over a significant part of Africa, Australia, and the Siberian and Dekan cratons, as well as in many other areas [1, 2].

In Fig. 2, ρ_1 and ρ_2 of metamafic rock are 3.35 and 3.15 g/cm^3 in the upper right- and lower left-hand parts of the *P–T* diagram, respectively. For these values of ρ_1 and ρ_2 at h_1 = 15 km, it follows from (1) that ζ is ~950 m, which is comparable with the recent uplifts of the crust on the Aldan Shield, southwest Africa, and Brazilian Shield.

In Fig. 3, ρ_1 and ρ_2 of orthogneiss are 3.25 and 2.85 g/cm^3 in the upper right- and lower left-hand parts of the *P–T* diagram, respectively. At h_1 = 15 km, it follows from (1) that ζ is ~1600 m. This is a rather high value comparable with the recent uplifts in the Precambrian areas (East Antarctic, Kongo and Kaapwaal cratons, Putorana Plateau).

For more precise estimations, we need to determine the average densities of the rock layer in its lower and upper positions. To calculate the average density of the layer, we should know the increase in temperature *T* with the depth *z*. In every certain area, the composition of rocks at different depths is known approximately. In rocks of different composition, the radiogenic heat release can vary by an order of magnitude producing large uncertainty in plotting the geotherms *T*(*z*) [12]. In addition, the density of rock of the same type can depend strongly on the contents of Fe, Al, H_2O, and other components [13]. Taking this into account, it makes sense to consider the simplest models. Here, we consider the processes in the lower and middle crust with low heat release. At constant temperature conductivity *T*(*z*), it can be approximated by linear functions as it is often done (e.g., [14]). The geotherms calculated for the Slave Craton [15] can serve as the example (Fig. 4). The distribution of *T*(*z*) on geotherm 1 for the epoch of 2.5 Ga at a depth of 30–45 km is almost linear. A similar situation is typical of the middle crust at a depth of 15–30 km on the present-day geotherm 2.

Let us consider the *P–T* diagram for metapelite as the example (Fig. 1). Let the temperature gradient in the crust be 15°C/km corresponding to the *AC* geotherm in Fig. 1. At average density of the crust of 2.9 g/cm^3, *T* of 700°C is achieved at point *A* at *P* of 1.254 GPa and the depth of 46.7 km. In Fig. 1, the layer *AB* 15.57 km thick in the lower crust corresponds

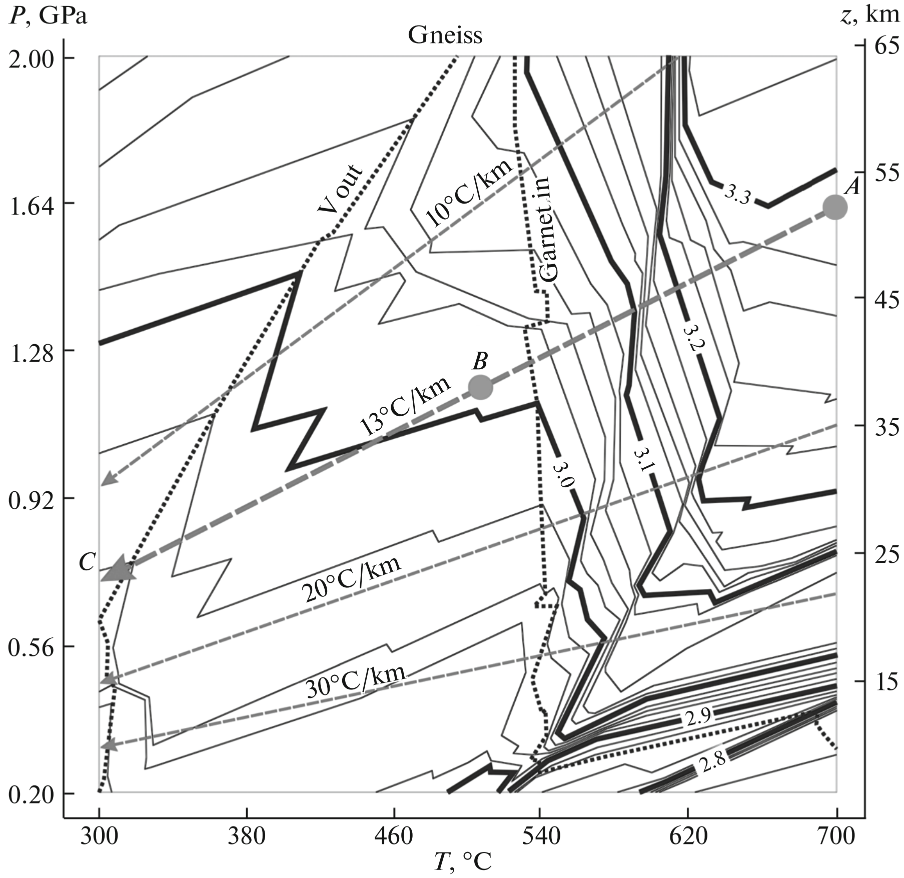

Fig. 3. *P–T* parameters of metamorphism of typical felsic gneiss from the metamorphic complex of Fjørtoft Island (Western Gneiss region, Norway) with isochores. The bulk composition of gneiss is after [8]. For symbols, see Fig. 1.

to the temperature range of 500–700°C. Its lower part includes the rocks of the amphibolite facies in contrast to rocks of epidote–amphibolite facies in the upper part. The average density ρ_1 of the layer calculated on the basis of the diagram of Fig. 1 is 2.94 g/cm^3. Let us assume that, as a result of denudation, the *AB* layer has become on the *BC* level of the *AC* geotherm, where *T* is 300–500°C. The average density ρ_2 of the layer decreases to 2.86 g/cm^3 after the infiltration of fluids into the layer, establishment of a new mineral equilibrium, and formation of the green schist facies in the upper part of the layer. According to relation (1), this provides the uplift with the amplitude ζ of 436 m on the crustal surface, which is close to the above rough estimation. It is suggested that the temperature gradient in the *BC* layer has the same value as in the *AB* layer in the Early Precambrian. In reality, the crust at that period was cooled by 100–200°C, whereas the temperature gradient in the middle crust of the cratons is typically higher than in the lower crust (Fig. 4). Thus, the geotherm of the *BC* layer will have another form. The task, however, is simplified, because the isolines in the left-hand part of Fig. 1 with a density step of 0.02 g/cm^3 are distributed over a large pressure ranges of 0.3–0.4 GPa. Thus, the inevitable uncertainty in the location of geotherms hardly results in errors exceeding 0.02 g/cm^3.

A similar approach can be used for Figs. 2 and 3, where the density isolines near the *BC* area of the *AB* geotherm are also separated by wide pressure intervals. For example, the average density ρ_1 of metamafic rock in the *AB* layer 14.2 km thick at dT/dz of 12°C/km is 3.34 g/cm^3. After displacement of the layer to the position *BC* and establishment of a new mineral equilibrium in the presence of fluid, the density decreases to ρ_2 of 3.1 g/cm^3. According to (1), this provides an uplift of the crust by ζ = 809 m. For the Fjørtoft gneiss, it is supposed that dT/dz is 13°C/km (Fig. 3). At a average density of the crust ρ of 3.1 g/cm^3 in the *AB* layer 14.4 km thick, the average density ρ_1 is 3.34 g/cm^3. In the *BC* position, the average density ρ_2 of the layer of diaphthorated rocks is 2.86 g/cm^3. This expansion will result in uplift of the crust by ζ = 1720 m.

Our estimations are somewhat approximate, especially taking into account that most crustal areas include blocks of different composition. Nonetheless,

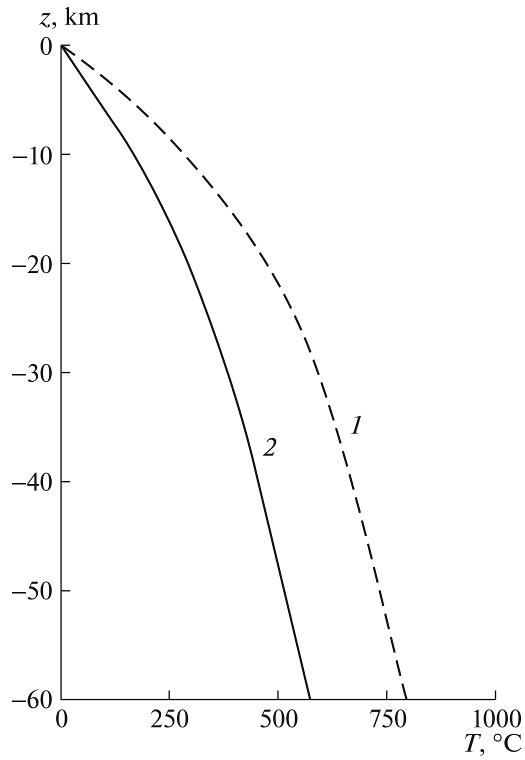

Fig. 4. Temperature profiles of the crust of the Slave Craton (Canadian Shield, modified after [15]). 1, Calculated geotherm for the differentiated crust at 2.55 Ga; 2, present-day geotherm.

they show that expansion of rocks due to metamorphism in the presence of deep fluids could produce large recent uplifts of the Precambrian crust, which is widespread on the continents.

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