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GEODYNAMIC FACTORS OF METAMORPHISM AND THEIR MODELING: REVIEW AND ANALYSIS OF THE PROBLEM

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The dynamics of deep-seated geological phenomena (tectonics and magmatism) disturbs the mass and temperature balance, thus leading to metamorphism of rocks. Five types of metamorphism are recognized: contact, in low-pressure/high-temperature belts, burial (subsidence), Archean, and collisional. These types differ in thermodynamic conditions, paleogeothermal gradients, and metamorphism duration. They are usually combined with each other. These combinations are not random and reflect certain tectonic appropriateness in the crust evolution. Magmatic intrusions are considered the most important elements of an additional heat supply into the Earth's crust. In the absence of associated magmatism, rock temperatures during the Phanerozoic burial metamorphism usually did not exceed those typical of prehnite-pumpellyite subfacies/facies and, rarely, greenschist facies. The collisional metamorphism caused by the subduction or superimposed load manifested itself at nonestablished thermal equilibrium owing to the rapid setting and subsequent rapid exhumation of crustal blocks and erosion. At present, determination of *PTt*-trends without invoking supplementary information can be used only for approximate evaluation of tectonic situations and elucidation of the types (causes) of metamorphism.

Geodynamics, metamorphism, modeling

INTRODUCTION

Academician Vladimir Stepanovich Sobolev, an outstanding Russian mineralogist and petrographer, made a significant contribution to the study of metamorphism. In 1964, he published a first scheme of metamorphic facies which reflected thermodynamic conditions of mineral formation in the Earth's crust and upper mantle. A principally important feature of the scheme is the substantiation of the crucial role of pressure as a factor of rock metamorphism. This idea, having been developed by V. S. Sobolev since 1949, later permitted recognition of belts of "blue" (glaucophane) schists as one of the most important elements of the structure and evolution of the Earth's crust. Sobolev initiated generalization on the study of metamorphic facies. In 1966, a relevant map encompassing a large territory was first compiled under his leadership. It was the *Map of Metamorphic Facies in the USSR*. Later, with participation of international geological institutions, other maps were compiled: *Map of Metamorphism in Europe* (1973) and *Map of Metamorphism in Asia* (1977). In 1970–74, four monographs on metamorphic facies were published under Sobolev's supervision. They became a significant inventory of theoretical, experimental, and geological data on mineral associations, petrochemistry of metamorphic rocks, conditions of their formation, and abundance in the Earth's crust and upper mantle. Sobolev put emphasis on study of metamorphic formations, taking into account lithological properties of rocks and their metamorphic history.

The information obtained on the basis of studying and mapping of metamorphic rock facies is now actively used in geodynamic constructions, permitting judgement of the causes of metamorphism. This paper deals with analysis of the state of the art.

PROBLEM BACKGROUND

First judgements on the causes of metamorphism seemed to appear simultaneously with the notion of "rock metamorphism", but now they are of historical interest only. In the struggle of ideas between neptunism and plutonism, the concept of metamorphism under the effect of a "fire-liquid" magma appeared in the first half of the XIX century. Then came ideas of two main factors responsible for metamorphism: heat of intruded magma and deep subsidence of rocks. Geologists began to discriminate between contact and regional metamorphism [1–3]; metamorphism related to tectonic events and orogenic deformation was recognized as a special type [4, 5]. Further development of the concept of metamorphism in the first half of the XX century was reduced mainly to the development and improvement of these ideas [6–11], although problems of stress-metamorphism and metasomatism were also discussed [12–14]. In the middle of the century, Turner and Verhoogen [15] made a classical synopsis of the data on metamorphism. They distinguished contact metamorphism, regional metamorphism associating with deformations and granitoid magmatism in orogenic belts, and burial metamorphism. They described the mantle as a general source of magmatic and metamorphic heat. Winkler [16] shared their opinion as a whole, but he emphasized the role of an "excess" (overlithostatic) pressure of fluid in the formation of specific lawsonite-jadeite-glaucophane rocks in burial metamorphism.

To explain the causes of metamorphism, revolutionary ideas were formulated by Miashiro in terms of plate tectonics [17]. Using combinations of metamorphic facies, he deduced three facies types of metamorphism for low, medium (moderate), and high pressures. The first type is associated with magmatism in the basement of ancient volcanic arcs and on the continental margins and is characterized by high geothermal gradient; the third type is connected with zones of subduction and distinguished by low geothermal gradient; both types can spatially be united into "paired" metamorphic belts, with contrasting orogeny and subsidence [18]. The second type, intermediate in the values of geothermal gradient and distinguished by moderate pressures of metamorphism, has a vague genetic nature. Specific categories according to Miashiro are: granulite-amphibolite metamorphism of shields as ancient nuclei of continents, metamorphism in collisional orogenic belts, and oceanic metamorphism of sea-bottom basalts.

Having generalized the information on metamorphic facies, Dobretsov et al. [19] came to the conclusion that the fold belts formed under conditions of subsidence metamorphism, deformation, and subsequent fluid-magmatic heat advection; granulite and migmatite-gneissic shields of distinct depths resulted from Precambrian polymetamorphism; eclogite-glaucophane-schist complexes appeared in zones of deep-seated faults or subduction with combination of subsidence, tectonic stress, and fluid "overlithostatic" pressure. Later, Dobretsov [20] recognized oceanic metamorphism as a variety of burial metamorphism typical of basic rocks of mid-oceanic ridges and basement of island arcs.

The concept of paired belts of Miashiro was proved in the region of the "Pacific circle" and became popular among petrologists, but serious obstacles hindered its application to the folded intracontinental regions. Therefore, an assumption was made that these fold belts, especially those containing no ophiolites and glaucophane schists, could have appeared as a result of simple collision of continental plates or between a continent and island arc [21].

In connection with the problem of intracontinental collision, Oxburgh and Turcotte [22] as well as England and Richardson [23] paid attention to the great significance of the thermal restructuring of the Earth's crust as a result of large overthrusts. Glaucophane schists and eclogites on the continents began to be considered as evidence of old subduction zones [20, 24]. The progress in plate tectonics permitted an explanation for the subsidence of expanding terranes of the Earth's crust [25–27] and contribution to the subduction model [e.g., 28]. These and other ideas [for example, 29–32] have formed a base in definition of thermotectonic positions of metamorphism; tectonic aspect of metamorphism has become a part of geodynamics, and the causes of metamorphism has been tied up with dynamic processes and thermal state of the lithosphere.

For the last 15–20 years, using achievements in continuum mechanics (theory of elasticity, rheology, and hydrodynamics), heat conductivity, and mathematical modeling, geodynamics greatly advanced its frontiers in some directions important for understanding the causes of metamorphism. With emphasis placed on thermal and rheological properties of lithosphere, important studies were carried out, concerning the formation of basins and depressions in the Earth's crust under various tectonic regimes, movements of large blocks of the Earth's crust, followed by disturbance in mass and temperature distribution, and thermotectonic regime of orogeny including uplift of magmas and deformations of all scales [33–35; see also review: 36, 37]. Attempts were made to extrapolate the ideas of plate tectonics onto the Precambrian; models for growth of the continental Earth's crust were considered, and rates of a decrease in heat flow from Archean to Recent were estimated [38]; geochemical models for formation of the Archean crust were also proposed [39]. An important

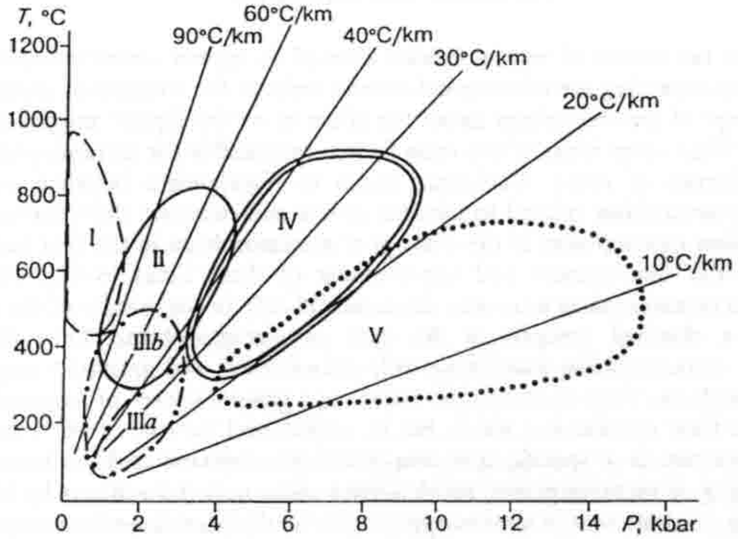

Fig. 1. ***PT*-diagram for types of metamorphism that differ in the values of geothermal gradient. I — contact, II — in low-pressure/high-temperature belts, III — burial (IIIa — in the absence of magmatic activity, IIIb — with additional supply of heat by magmatic intrusions to the basement of basin), IV — Archean, V — collisional.**

field of study has become the determination of *PTt*-trends on the basis of mineralogical data and modeling [40–43], imposing restrictions and contributing to tectonic reconstructions.

TYPES OF METAMORPHISM

The mantle convection flows interact with the lithosphere, causing mass redistribution and heat transfer. As a consequence of deep-seated geodynamic activity, this mass and heat transfer is expressed mainly in tectonics and magmatism, i.e., in movement of blocks of the Earth's crust, deformations, and melt injections. It disturbs the mass and temperature balance at depth in the Earth's crust, thus leading to rock metamorphism. The process either continues until the thermal equilibrium and/or mass balance are restored or, which occurs more frequently, is interrupted by another tectonic or magmatic event.

Based on the works of predecessors (see above), we can distinguish, at least, five types of metamorphism caused by tectonics and magmatic activity: (1) *contact*, (2) *in belts at low pressures* (*P*) *and high temperatures* (*T*), (3) *burial*, (4) *Archean*, and (5) *collisional*. They are implemented in different thermodynamic regimes and are, therefore, correlated with certain associations of metamorphic facies. In accordance with the existing nomenclature [44], different combinations of muscovite-hornfelsic, amphibole-hornfelsic, pyroxene-hornfelsic, and spurrite-merwinite facies, depending on temperature distribution, appear during contact metamorphism, in thermal aureoles around shallow (no more than 5 km from the surface) intrusive magmatic bodies. Low-*P*/high-*T* belt metamorphism, associating with moderate-depth (to 15 km) magmatism, leads to zonal andalusite-sillimanite (in metapelites) complexes, including "zeolite", greenschist, epidote-amphibolite, and amphibolite facies of "low"-pressure. On burial metamorphism, rocks of "zeolite" and, more rarely, of greenschist facies form in spreading zones. The continental shields (oldest stable terranes of the Earth's crust) are usually made up of rock complexes of granulite, amphibolite, epidote-amphibolite, and greenschist facies of "low"-pressure. Collisional metamorphism leads to the origin of disthene-bearing gneiss-schist complexes (in metapelites) or combinations of rocks of eclogite, disthene-schist, greenschist, and jadeite-glaucophane facies. These types are usually combined with one another, e.g., contact metamorphism is combined with low-*P*/high-*T* belt metamorphism, Archean one with burial metamorphism, the latter with collisional low-*P*/high-*T* belt metamorphism (a frequent variant in folded areas), etc. These combinations are not accidental; on the one hand, they indicate the presence of transitional tectonic regimes and, on the other, reflect certain geotectonic appropriateness in the Earth's crust evolution.

The first two types were caused by an additional supply of heat as magmatic melts intruded the Earth's crust. As a result, the geothermal gradient in metamorphic rocks significantly exceeded the "normal" (average) quantity. The third and fourth types are characterized by geothermal gradients close to the average value. Thus, the heat flow during metamorphism was within the norm. The last (fifth) type is distinguished by a geothermal gradient below the "normal" quantity; most likely, this is connected with a relatively short duration of events and failure in reaching thermal equilibrium between blocks of rocks at corresponding depths: Temperature changes more slowly than pressure does. The role of overlithostatic pressure without a radical growth of heat flow is not clear. These reasonings are schematically illustrated in *PT*-projection in Fig. 1; the boundaries of types are shown roughly.

Types and models of metamorphism are characterized below in the traditional succession: from low to high pressures.

I. Contact metamorphism. This type of metamorphism manifests itself in the upper part of the Earth's crust, at rather low pressures, in a quiet tectonic setting. Of all the types of metamorphism, this demonstrates the most remote and weak relation to the crust-mantle interaction; its nature is clear and is studied best of all. Contact metamorphism obviously depends on the heat of intruding magma. Studies of the geochemistry of contact-metamorphic rocks show that in the cases when mineral transformations were performed isochemically, the predominant way for heat transfer was conduction. This permitted us and other authors to develop relevant models, to describe the process in detail, to consider the dynamics of contact metamorphism, and to analyze its factors in the context of their importance [45–49 etc.]. Mathematically, the problem is written as a system of equations of heat conductivity, taking into account the position of phase boundaries, sinks and sources of heat. Modeling is made in terms of Stephan's problem, with the rates of phase transformations corresponding to the rate of temperature variations. The numerical solution of the problem yields a 3D distribution and evolution of temperatures near magmatic intrusive bodies of any shape. The study of limiting cases showed that the most important factors of contact metamorphism are: temperature at the contact of intrusion, initial temperature of the country rocks, and sizes of an intrusive magmatic body. Combination of these factors governs, as a rule, the level of maximum temperatures at different distances from the contact and character of metamorphism (distribution of *PT*-conditions). Pressure on contact metamorphism is usually controlled by the weight of the overlying rocks, i.e., with a rare exception, it is lithostatic.

Under shallow conditions (less than 1–1.5 km), the thickness of contact aureole is determined by the initial temperature (and composition) of magma and duration of heating. The latter is due chiefly to sizes of the intrusion. In the absence of convection and melt flow, the temperature at the contact depends on the solidus temperature and latent heat of crystallization. As a rule, higher temperatures are related to basic and intermediate magmas and lower temperatures, to acid magmas. Thermal transformations of rocks in the contact aureole proceed, as a rule, in the range from 300 to 700 °C. Endothermic metamorphic reactions in the country rocks can somewhat reduce the time of magmatic crystallization and slightly lower the temperature at the contact. Temperatures of more than 750 °C during contact metamorphism under shallow conditions are reached only as a result of magma motion (heat convection or force flow). The melt overheated above the solidus temperature may slightly increase the width of contact aureole; the number of metamorphic zones within the aureole has an insignificant effect on its total thickness. Sizes of aureole depend little on the type of mineral transformations (distribution of heat effect of the reaction in time) at the front of metamorphism. However, an increase in the initial temperature of the country rocks has a rather considerable effect on the facies character of metamorphism. Thus, with all other things being equal, an increase in initial temperature by 100 °C leads to a 15% increase in the width of contact aureole, the immediate contact temperature being 40–50 °C higher.

Our conclusions underline that the conductive heat transfer in the aureole rocks is predominant over the convective transfer. If there are indications of heat transfer by a rock-filtering fluid, the problem of metamorphism must be reduced to metasomatism [45, 46, 50, 51 etc.]. The modern methods of research, including mathematical simulation of temperature fields, permit these cases to be distinguished safely [48, 49].

Metasomatism expressed at the magmatic stage is usually involved with alkaline melts, rarely with acid melts, and extremely seldom with basic and intermediate melts; its development is local and is limited by permeability of enclosing rocks. The role of conductive heat transfer in the transformation of rocks at intrusive exocontacts seems generally to be much more significant than the role of rock heating with participation of a convecting magmatic fluid.

II. Low-*P*/high-*T* belt metamorphism. The so-called "low-pressure/high-temperature metamorphic belts" are in space and time associated with orogeny (uplift) and granitoid magmatism. Their tectonic nature is treated either as that of ancient island arcs, e.g., Abakuma Plateau in Japan [18, 21] and Connemara region in Ireland [52], or as magmacrystalline "cores" (arcs) of folded mountain ridges [53–55]. The main typical feature of metamorphism of this kind is an elevated heat flow; it could be due either to magma intrusions [54–56] or to filtration of mantle fluid [57–60]. These mechanisms could act either separately or in combination.

The mathematical model for rock metamorphism near a magmatic intrusive body is based on solving equations of heat conductivity on the assumption that the heat balance is kept when the boundaries move along the exo- and endocontacts. The heating of enclosing rocks follows the same behavior as described in the previous paragraph, with the exception of an increased initial temperature of the environment. The initial distribution of temperatures is given by the geothermal gradient, and subsequent changes are governed by the solidifying magmatic intrusion depending on depth of its occurrence, thickness, shape, temperature, and composition. Duration of metamorphism (and anatexis) is limited by heat resources of the magma chamber. Calculations for the middle part of the Earth's crust (below 20 km) show [56] that in the roof of a basic/ultrabasic intrusion 5 km in thickness, the width of the anatexis zone in metapelites can reach 3 km and that of metamorphism, about 5–7 km in total. The total duration of progressive and regressive stages of metamorphism and anatexis is estimated at 5–10 Ma. The granite melt can retain its mobility for millions of years after the basic/ultrabasic intrusion has been solidified. At the base of thick basic plutons, the anatexis is both more extensive and more intensive than at the roof (owing to higher initial temperatures of the medium).

Model for a fluid flow implies solving an equation of the convective heat transfer by volatiles filtered through a vertical permeable zone, given that Stephan's condition is fulfilled at the moving boundaries of phase transitions; the parameters taken into account are: temperature, heat conductivity, heat capacity and density of rocks, melts, and fluid, heat of phase transitions, flow rate of fluid, etc. Calculations for the middle part of the Earth's crust show [57] that the degree of metamorphism and melting of acid rocks within a linear permeable zone 2 km or more in width depends chiefly on the flow rate of fluid (at least, 10^{-9} g/(cm$^2\cdot$s)) and its temperature (more than ~650 °C). The distribution of maximum temperatures throughout the section of the Earth's crust corresponds to stationary regime, which comes into the existence about 4 Ma after the beginning of filtration. The thickness of the zone of metamorphism bounded from below by anatexis reaches 13 km; anatectites occur at a distance of 16–18 km from the Earth's surface. The growing width of the permeable zone (up to 70 km) leads to reduced heat loss into the wall rocks and favors the enlargement of zone of melting and metamorphism.

The intrusive model is characterized by the same initial thermal state for the upper part of the section of the Earth's crust, but with depth, approaching the basic/ultrabasic pluton, temperature quickly increases. In the model for fluid flow, an increase in rock temperature occurs at small depths (less than 5 km); the deeper parts of the section are characterized by a weak monotonous increase in temperature with a small thermal gradient: 15–20 °C/km. Paleogeothermal gradients, typical of zonal metamorphic complexes of folded zones of orogens, are satisfactorily correlated with calculated gradients in the model of magmatic intrusion and are poorly correlated with gradients in the model for fluid flow.

The combined model unites the above-mentioned mechanisms [58]. Their joint action leads to significant retardation of crystallization of a basic melt and to a longer period of preservation of anatectic acid melt as well as to a higher position of the anatexis boundary in the section. This creates conditions for a longer existence of large volumes of granite melts at depths of more than 10 km. Having a low density, these melts are lighter in weight than the overburden rocks and must float up. Thus, granite diapirs and granite-gneiss domes must form in regions with high-grade metamorphism. The quantitative role of convective heat transfer in a combined model depends on a particular situation, but in general it should not be great. As in the case of contact metamorphism, conductive heat transfer dominates in the formation of low-pressure/high-temperature metamorphic complexes.

As in the case of contact metamorphism, the pressure is determined chiefly by a load of the overlying rocks, i.e., is lithostatic.

III. Burial metamorphism. Metamorphism of sediments to be subsided ("buried") without relation to subduction appears, as a rule, under horizontal extension/spreading and thinning of the lithosphere in different geological situations. Spreading is caused by mantle flows. Another important mechanism of the formation of depressions in the Earth's crust (e.g., in the piedmont troughs, foreland) is believed to be the lithosphere flexure caused by overthrust loading. The rifting mechanism of subsidence is better investigated and understood.

A generalized geodynamic model of rifting is developed by McKenzie [26]. This model implies the extension and thinning of the lithosphere, uplift of the hot asthenosphere, restoration of isostatic equilibrium, cooling and thermal compaction. The model is formulated as an equation of heat transfer from the asthenosphere to the Earth's crust, taking into account variations in the thickness of lithosphere, rate of its spreading, and accumulation of sediments in the forming depression. Good agreement of the model calculations with the observed dynamics of tectonic subsidence obtained on the basis of stratigraphic and lithological information on the structure of the sedimentary section with the use of "backstripping" procedure is indicative of the evident validity of the model [61]. In terms of mathematical modeling, three versions of finite extension/spreading of the lithosphere were analyzed: "instant", accelerated, and at constant rate. It is established that the dynamics of subsidence of depressions and the evolution of heat flow can be roughly estimated by a model permitting "instant" extension of the lithosphere [26]. The reason is that the long-term subsidence of the depression after spreading occurred independently of the way (rate) of extension and was related mainly to thermal contraction [62].

As a rule (in the Paleozoic and Mesozoic), the extension of lithosphere plate during rifting amounted to 30–50%, depths of depressions were 6–9 km, temperature of rocks at the base of depressions reached the level of prehnite-pumpellyite facies/subfacies of metamorphism and, more rarely, the level of greenschist facies. However, exceptions are possible which are connected mainly with ancient (Precambrian) rifts where very deep depressions appeared (e.g., the Adelaid basin, Australia, reaching a depth of 15 km). A possible reason is either significant extension of the lithosphere plate (by a factor of 2–3 and more) or eclogitization of the basalt layer [33].

The model temperature estimates were made, ignoring possible magma injections into the rocks at the base of basins. However, if rifting was accompanied by basic magmatism, the role of this factor could have been quite important; because of the abundant magmatic injections (sills, dikes), the temperature of rocks at the bottom of the section could have increased by many tens and even some hundreds of degrees. As in the case of a model for magmatic intrusion (see above), the temperatures would grow near magmatic bodies, i.e., at the basement of basin and in sediments immediately lying on it (1–2 km thick), thus leading to epidote-amphibolite and/or greenschist facies metamorphism, depending on the total amount of intruding magma.

In the majority of cases, the evolution of rift depressions is explained by simple extension of the lithosphere plate. This is confirmed by a positive correlation between depths of basins and values (coefficients) of spreading. Sometimes, however, the observed extension of the lithosphere is insufficient to explain significant subsidence of the Earth's crust within relatively narrow graben-like structures. Then the mechanism of eclogitization of the basalt layer is used [33, 63]. The mechanism includes accumulation of the basic melt in the asthenosphere lens during the rifting stage, followed by its solidification in the form of eclogite and subsidence. The formation of dense mass in the lower part of the lithosphere leads to deepening of the graben in the Earth's crust, but the subsidence and heating of buried sediments is usually attended by metamorphism, which does not exceed the grade of greenschist facies (in the absence of additional heat supplied by the intruding magma).

Under the effect of the load of overthrusts in the foreland basins, the depth of the formed depression does not exceed 3 km [33]. Judging from model estimates and geological observations, the other cases of formation of depressions in the Earth's crust were not accompanied by a considerable subsidence of sedimentary rocks to great depths either [64–67]. The depths do not exceed 3–4 km, thus preventing the sediments from heating up to more than 100–150 °C and corresponding metamorphism.

Burial metamorphism is usually interrupted by uplift of a rock mass and erosion caused by other geological events. The duration of burial metamorphism can reach millions of years. The pressure under burial metamorphism does not usually exceed the lithostatic one.

IV. Archean metamorphism. Tectonic and geodynamic causes of metamorphism of the rocks making up the old shields are unclear and debatable; the ways of formation of the Archean-Proterozoic continental crust are not clear either. It is possible, however, to state that tectonic and magmatic activity in the Archean and Early Proterozoic was significantly different from the Phanerozoic activity. First of all, the subduction caused by plate collision seems to be very unlikely at that early stage of the Earth's evolution, but there is evidence of spreading extrapolated onto the formation of greenstone areas [38, 39].

Most likely, we can speak about two different complexes making up the old shields: granulitic (or granulite-migmatitic) and greenstone. They formed in different geological settings and have significantly distinct thermal histories. The former are thought to be produced in the process of formation of the primary basalt crust, its repeated melting, and accumulation of volcanoplutonic sialic masses in the form of ancient plates (blocks), whose thickness (as a result of horizontal movements) was probably commensurate with the thickness of the modern crust (up to 50 km). The latter seemed to appear as the basalt-komatiite filling of extended deep troughs during the splitting apart and spreading of the above plates [39].

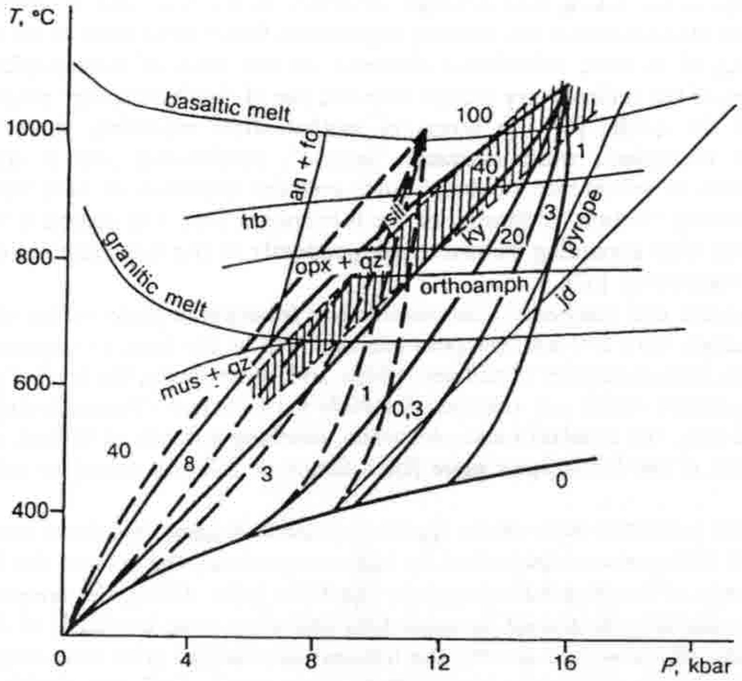

Fig. 2. ***PT*-diagram for one of the scenarios of the Archean metamorphism during the formation of granulites. There are two versions of the evolution of temperatures (as assemblages of isochrones – solid and dashed lines) with depth in the case of a constant heat source (with temperatures of 1000 and 1100 °C) at the lower boundary of the two-layer crust (see in the text and [33]); digits on isochrones mean time in Ma. Boundaries of stability of some minerals and their associations are shown after [44]: muscovite with quartz (mus + qz), orthoamphibole (orthoamph), pyrope, jadeite (jd), orthopyroxene with quartz (opx + qz), hornblende (hb), anorthite with forsterite (an + fo); hatched field is a zone of uncertainty at the boundary between stability fields of sillimanite (sil) and kyanite (ky). Boundaries of melting are shown for granite and basalt.**

Multistage polymetamorphism during the formation of granulites occurred in the Earth's crust of variable thickness, which resulted from tectonic "heaping" and subsequent extension/spreading of the Archean lithosphere; increased temperatures were caused by thinning of the lithosphere and intrusions of mantle magmas. The pressures ranged from 7 to 15 kbar and temperatures, from 700 to 1000 °C. Many stages of metamorphism are difficult to differentiate, but, evidently, various thermodynamic regimes coming to the existence were related to decompression and cooling, with or without erosion [68]. As a rule, the general geothermal gradient at the progressive stages of metamorphism was sufficiently close to the "normal" value; significant deviations depending on tectonic setting occurred at regressive stages.

A fundamental peculiarity of the Archean granulites is that nearly uniform high pressures and temperatures were attained over large areas of metamorphism. In terms of one-stage metamorphism (in a first approximation) this can be explained by the mechanism of lithostatic control of pressure and long-term warming of the rocks by a constant heat source, the convecting mantle magma. A relevant model taking into account the conductive heat transfer in the two-layer lithosphere with heat generation was devised by Artyushkov and Batsanin [33]. Figure 2 shows two versions of the dynamics of depth-dependent temperature variations in time for two cases of continental crust, 40 and 60 km thick. In the first case, the constant temperature at the lower boundary of the two-layer crust is equal to 1000 °C and in the second, to 1100 °C. During some millions of years, the temperature changes mainly in the lower part of the crust and remains nearly invariable in the upper. Then variations encompass the upper crust as well, especially during the period between 3 and 30 Ma. Melting of acid rocks occurs at depths of more than 20 km. After 15–40 Ma, the temperature distribution in the crust differs little from the stationary pattern.

Other possible explanations of *PT*-conditions for granulites must, likely, take into account models for heat transfer near large basic/ultrabasic plutons (see paragraph II) and deformations of the Earth's crust (see paragraphs III and V). Attention must be given, however, to one more circumstance. In some cases, e.g., in the Laplandian granulites [69], variations in paleotemperatures with depth during the same stage of metamorphism were very small. This is evidence of heat transfer by a filtering CO_2-enriched fluid [68]. Thus, a model for fluid flow (see above) is preferable to explain these peculiarities of metamorphism.

As regards the Archean greenstone rocks on shields, their formation may, likely, be interpreted in terms of burial metamorphism in trenches during extension/spreading of the lithosphere plate (see above). In this case, the depth of trenches and heat flow must be considerable (as opposed to Phanerozoic) to explain the appearance of greenschists and epidote amphibolites. Greenstone rocks are associated with granite-gneissic domes, which are, probably, connected with ultrametamorphism of a later stage of tectonic activity [38].

Kyanite-gneissic units of median masses are worthy of special note. They can be considered to be relict, heavily transformed blocks of the ancient crust which was similar to granulite complexes of shields [19, 20].

V. Collisional metamorphism. This kind metamorphism is due to horizontal compressing motions in the Earth's crust at the minimum intrusive magmatic activity. Collision of lithosphere plates causes their compression, deformation, formation of overthrusts and underthrusts, and thickening ("heaping") where they collide. On the continents these processes are accompanied by the formation of orogenic belts of collision type. It seems possible to recognize two kinds of collision: underthrust of oceanic lithosphere beneath a continental one, leading to subduction, and thrust of continental plates over one another [70, 71].

The first kind of plate collision is associated with metamorphism in "ophiolite belts" or in "belts of high pressures/low temperatures" of Miashiro, widely represented, in particular, on the periphery of the Pacific. These metamorphic belts are distinguished by their development in linear extended structures, common metabasic rocks among low-temperature metapelites, usually no granitoids, stability of specific minerals and their associations as indicators of high pressures and low temperatures, irregular and intensive deformation and foliation [18, 19, 72].

First attempts to model subduction-related processes were made about 20 years ago [28, 31, 73 etc.]; later, these efforts continued, and now numerous publications on this subject are available (the fundamental works are [74–80]). Subduction is supposed to be a process when an oceanic lithosphere plate moves under the continental plate by the action of horizontal force of the convecting mantle and gravity. The horizontal force must be high enough to overcome the push-out resistance of mantle rocks and, making the oceanic lithosphere heavier by eclogitization, to be able to subside the plate to a significant depth. While the plate descends, it is warming up, but millions of years are required to restore the equilibrium with the surrounding mantle. To escape its heating to the mantle temperature, the plate should descend "rapidly". "Accretional" [21] volcanosedimentary rocks entrained by the oceanic-plate during subduction can be subsided to depths of 40–50 km, where such minerals as glaucophane, lawsonite, jadeite-bearing pyroxene and pyrope-bearing garnet, aragonite, etc. form under the effect of high pressure at relatively low temperatures [19, 20, 52, 72]. Like subduction, the uplift and erosion of the rocks which were subjected to high-pressure metamorphism but did not reach the thermal equilibrium with the environment must be "rapid" (some centimeters a year) to prevent the minerals from replacement by higher-temperature phases. This is a special problem, which is under wide discussion [72, 81, 82]. The solution is likely to be sought in such a distribution of forces of compression (and stress) and ratio of densities and viscosities between a passive and an active plate and the surrounding mantle when conditions for the "rapid" reverse motion ("floating up") of accretionary rocks are provided after subduction. Under discussion is a model where a rigid foreign block ("breast-wall") interrupts the subduction of accretionary matter and changes the direction of motion to ascending [83]; more complicated models, taking into account "rapid" erosion of the upper crust, are also proposed [84]. These processes involved with intensive deformation lead to the appearance, in zones of subduction, of combinations of mineral associations reflecting the nonuniform distribution of pressure as a result of complicated metamorphic history. Erosion usually reflects a tectonic mosaic of rock blocks deformed and metamorphosed to a variable degree. At present, there is no satisfactory mathematical model for subduction taking into account all the above phenomena.

Thrusting of continental plates over one another also leads to collision metamorphism. Some kyanite-bearing metamorphic complexes of moderate and high pressures certainly belong to this category. In particular, there is a high probability that large overthrusts controlled the formation of the kyanite rocks from the Keiv Formation among the Karelides of the Kola Peninsula, as well as similar rocks from the Precambrian, Ufalei Massif in the Urals [19]. A typical feature of this metamorphism is either absence or weak development of temperature zoning. This indicates the rock metamorphism under load of a "cold" overthrust rock mass. Given that a "hot" plate (the temperature at its bottom significantly exceeds the temperature at the roof of underlying unit) "rapidly" thrusts over the "cold" plate, a temperature-"inverted" metamorphic zoning can form, which is also considered an important indication of collision [52]. Variations in P and T corresponding to this structural and thermal restructuring arise under conditions of a great diversity of initial mechanical, temperature, and geometrical factors owing to tectonics [43].

Duration of collisional metamorphism is usually governed by the time of "rapid" uplift of a rock block preventing high-pressure minerals from replacement; it does not exceed 10 Ma.

The collision processes are widely represented in the history of the Earth's crust; many folded areas of continents appeared as a result of these processes, the collisional metamorphism being combined, at different stages of evolution, with other types of metamorphism – orogenic and subduction. In addition, old prefolded metamorphic blocks were usually involved in collision. All these factors create serious difficulties in studying the geological past of fold belts [52].

The most intriguing events involved with collision took place when diamond- and coesite-bearing ultrahigh-pressure metamorphic rocks formed [85]. These rocks originated at a pressure of no less than 28–40 kbar in the Alps (Dora-Maira), in the Kokchetav Massif in Northern Kazakhstan, in the Dabie Shan (Eastern China), in the Western-Gneiss region of Southern Norway, etc.; if they were caused by subduction or intracontinental collision, the depth of subsidence should be at least 100 km. Taking into account that metamorphic rocks must ascend to the Earth's surface "rapidly" (see above), this provides a fresh view on the possibilities of collision mechanism in terms of subduction or load of large thrusts over the underlying rocks. Although many aspects of the models need to be improved, it is not necessary to use hypotheses of tectonic or fluid overlithostatic pressure to explain the origin of these rocks.

PTt-TRENDS AND TYPES OF METAMORPHISM

Variations of thermodynamic conditions in time within a single geological process, following the same direction of evolution, were named *PTt*-trajectories, *PTt*-trends, *PTt*-paths, etc. Proceeding from the physical essence (i.e., physical mechanism: heat supply by magmatic intrusions, extension/spreading of lithosphere, deformations of the Earth's crust, including subduction, overthrusts, underthrusts, etc.), each type of metamorphism should be characterized by specific changes of P and T in time, with the uplift, erosion, and cooling of rocks taken into account. This is confirmed by results of modeling of metamorphism of various types, which, however, remain within sufficiently simple situations [40–42, 52, 59, 72, 86–88 etc.]. On the other hand, changes in composition of mineral associations under rock metamorphism are also indicative of the evolution of P and T. A problem arose in a natural way – to relate the model *PTt*-trends with the trends observed from mineral associations so that we could analyze past tectonic events.

Proceeding from the model concepts and geological observations, we can distinguish two main tendencies in the *PT*-evolution of metamorphism: (a) when the maximum temperature (peak T) in time precedes the maximum pressure (peak P), the P and T parameters move "counterclockwise", (b) when the maximum pressure precedes the maximum temperature, the P and T parameters move "clockwise". The first tendency is realized during orogeny and heat supply by an intrusive magma [89, 90], which is usually accompanied by deformation and thickening of the crust [41]. The second tendency is most common under subduction [72], when "cold" rocks are subjected to metamorphism; they begin to warm up somewhat later, but the thermal equilibrium is not reached because of short duration of the event. Both *PTt*-trajectories are possible during overthrusting [41, 42, 91, 92]; overthrusts can also lead to an inverse metamorphic zoning [42]; complicated versions of *PT*-evolution appear during overthrusts, with uplift of the Earth's crust blocks and erosion taken into account [43]. Burial metamorphism, a result of extension/spreading of the lithosphere, is characterized by the counterclockwise motion of P and T in time [93]. During contact metamorphism, because the process is short-term, the progressive and regressive parts of the *PTt*-loop virtually coincide, since the warming and cooling of wall rocks proceed at constant pressure.

It is evident that model estimates are rather ambiguous; they can hardly be used for clearing up tectonic causes of metamorphism. As regards the reconstruction of *PTt*-trends from mineral constituents of metamorphic rocks, there are considerable specific difficulties. In mathematical terms, to reconstruct the *PTt*-trends is to reconstruct the functionals from dynamic equations using a set of data obtained at different times. This problem is difficult to solve, because necessary information is not available in a required volume. As applied to the issue under discussion, a precise definition of *PTt*-trend is difficult or even impossible, because the information on the preceding stage of metamorphism, "recorded" in coexisting minerals, is often "erased", distorted, or replaced by new information either as a result of "phase reactions" ("discontinuous"), when minerals appear and disappear, or as a result of exchange reactions ("continuous"), when the number of phases remains constant. This occurs the more rapidly, the closer to equilibrium is the mineral association. If the rock is completely recrystallized and is made up of an equilibrium mineral association, no reconstruction of *PTt*-trend is possible: Its state is determined by the *PT*-parameters existing at that time. Therefore, during prograde metamorphism (at high temperatures and in the presence of a water fluid) the information on the early history of a metamorphic rock is usually completely obliterated. On a local scale, only the retrograde part of the *PTt*-trend (in a descending branch of evolutional loop) can be restored from relict minerals. There is still a possibility to restore some part of *PT*-evolution, using spatially separated minerals or mineral associations of a rock (of the type of coronites or drusites). In this case, the processes of diffusional transport of chemical components are taken into consideration, and the problem becomes dynamic. In the framework of this problem, famous for its complexity, it is possible to choose a number of simpler problems solvable in a strict way. First of all, an assumption on local quasi-equilibrium may be retained in some cases. It is reasonable to suppose that the relatively far mineral grains in the rock retain information about an early metamorphic episode, whereas the near grains – about a late one. This serves as a basis for a widely used technique for estimating *PT*-parameters of metamorphism from the compositions of central ("cores") and/or marginal ("rims") parts of zonal mineral grains with the use of thermobarometry. It should be stressed that this approach is of some risk. Estimations of *PT*-conditions from mineral cores ignore the effects of intraphase redistribution of chemical components when ordering the solid phases and diffusion, which, judging by experiments, are quite rapid. Estimations of *PT*-parameters from adjoining rims of zonal minerals are also doubtful: It is unknown to which moment of the regressive stage of evolution they correspond. The use of different rims can lead to contradictory results [94], although principally they can be all correct but correspond to different time intervals. In the *PT*-projection, these different estimates can yield a single smooth curve and, together with other data, give a version of *PTt*-trend, which is not always reliable. A stricter approach is to solve a system of equations for the dynamics of transfer and the kinetics for chemical components of the rock to be metamorphosed, but there are difficulties related to the uncertainty of initial and boundary conditions, coefficients of diffusion, etc. To follow this way, we can, in some cases, have a progress in the restoration of *PT*-conditions of metamorphism, in particular, for zonal mineral structures of the coronite type [95].

Thus, at present the determination of *PTt*-trends cannot be considered a quite reliable tool to clear up tectonic conditions and to recognize types of metamorphism. The illustration in Fig. 3, shows on the basis of literature data that the evolution of *P* and *T* within the same type of metamorphism may occur both clockwise and counterclockwise (especially in metamorphic belts at high *T* and low *P*). Nevertheless, all the types differ distinctly in values of geothermal gradient. In addition to this information, other data should be used to analyze the tectonic causes of metamorphism. This promising problem needs further study.

CONCLUSIONS

Some conclusions follow from the above.

Metamorphism is the result of geodynamical processes; its study provides an insight into the thermal state and evolution of the lithosphere and important information permitting us to judge the connection of metamorphism with tectonic and magmatic phenomena in the Earth's history.

We can recognize *five types* of metamorphism on the basis of their tectonic causes: *contact*, *in low-pressure/high-temperature belts*, *burial*, *Archean*, and *collisional*; they differ in thermodynamic regimes and duration. The first two are distinguished by a higher geothermal gradient; the third and the fourth are characterized by a geothermal gradient close to the average value, i.e., normal; the latter has a typically lower geothermal gradient. The pressure in all types of metamorphism is controlled by a lithostatic load. The types of metamorphism usually occur in combination; these combinations are not accidental and reflect certain geotectonic appropriateness in the Earth's crust evolution.

Magmatic intrusions are considered to be the most important element (mechanism) of supply of additional heat to the rocks of the Earth's crust during metamorphism; the heat transfer related to fluid filtration plays a drastically subordinate role.

In the absence of accompanying magmatic activity, the rock temperatures during the Phanerozoic *burial metamorphism* did not exceed the level of prehnite-pumpellyite facies/subfacies, more rarely, the level of greenschist facies.

Collisional metamorphism typically manifests itself under the conditions of nonstationary *PT*-equilibrium as a result of "rapid" descent and uplift of the Earth's crust blocks and erosion.

At present, in the absence of additional information, the determination of *PTt*-trends cannot be

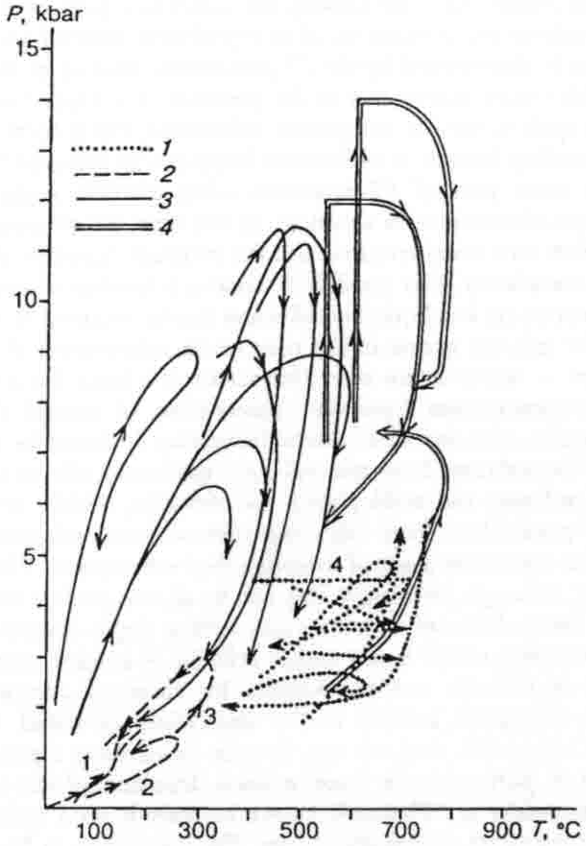

Fig. 3. A diversity of *PTt*-trends according to literature data for various types of metamorphism. *1* – metamorphism in low-pressure/high-temperature belts, after [55]; *2* – burial metamorphism; *3* – collisional metamorphism (subduction complexes), after [72]; *4* – metamorphism during formation of granulite complexes; supposed *PTt*-trajectories are shown within different tectonic events, after [42]. The digits show: 1 and 2 – *PT*-evolution of the basement rocks, during the development of the Danish and Dnieper-Donetsk basins, respectively, after [96]; 3 – *PTt*-trend for the rocks of the Welsh basin, after [93]; 4 – supposed *PTt*-trend for the rocks of the Tongulak metamorphic moderate-pressure complex in Altai, after [96].

successfully used to analyze tectonic situations and to establish types of metamorphism. This problem is required to be further developed.

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REFERENCES

[1] J. J. Sederholm, *Tscherm. Min. Petr. Mitt.*, vol. 12, p. 97, 1891.
[2] L. Milch, in: *Neues Jahrb.*, Beilage-Band 9, p. 101, 1894.
[3] Ch. R. Van Hise, *Treatise on Metamorphism. US Geol. Surv.*, vol. 47, 1904.
[4] F. Zirkel, *Lehrbuch der Petrographie. Ester Band*, Leipzig, 1893.
[5] A P. Karpinsky, in: *Selected works*, Moscow–Leningrad, vol. 1, p. 186, 1941.
[6] U. Grubenmann and P. Niggli, *Die Gesteinmetamorphose. I. Allgemeiner Teil*, Berlin, 1924.
[7] F. F. Grout, *Petrography and petrology*, New York–London, 1932.

[8] H. Rosenbusch, *Elemente der Gesteinlehre* [Russian translation], Leningrad – Moscow, 1934.

[9] A. Harker, *Metamorphism* [Russian translation], Moscow, 1937.

[10] T. F. W. Barth, C. W. Correns, and P. Eskola, *Die Entstehung der Gesteine. Ein Lehrbuch der Petrogenese*, Berlin, 1939.

[11] D. S. Korzhinskii, *Factors of mineral equilibria and mineralogical depth facies* [in Russian], Moscow, 1940.

[12] F. J. Turner, *Mineralogical and structural evolution of the metamorphic rocks* [Russian translation], Moscow, 1951.

[13] H. Ramberg, *The origin of metamorphic and metasomatic rocks. A treatise on recrystallization and replacement in the Earth's crust*, Chicago, 1952.

[14] D. S. Korzhinskii, in: *Fundamental problems of the study of magmatogenic ore deposits* [in Russian], Moscow, p. 335, 1955.

[15] F. Turner and J. Verhoogen, *Petrology of igneous and metamorphic rocks* [Russian translation], Moscow, 1961.

[16] H. G. F. Winkler, *Petrogenesis of metamorphic rocks* [Russian translation], Moscow, 1969.

[17] A. Miyashiro, *J. Petrol.*, vol. 2, p. 277, 1961.

[18] A. Miyashiro, *Metamorphism and metamorphic belts*, London, 1973.

[19] N. L. Dobretsov, V. S. Sobolev, N. V. Sobolev, and V. V. Khlestov, *Facies of regional high-pressure metamorphism* [in Russian], Moscow, 1974.

[20] N. L. Dobretsov, *Global petrological processes* [in Russian], Moscow, 1981.

[21] A. Miyashiro, K. Aki, and A. M. C. Sengor, *Orogeny*, Chichester – New York – Brisbane, 1982.

[22] E. R. Oxburgh and D. L. Turcotte, *Schweiz. Miner. Petrogr. Mitt.*, vol. 54, p. 641, 1974.

[23] P. C. England and S. W. Richardson, *J. Geol. Soc. London*, vol. 34, p. 201, 1977.

[24] W. G. Ernst, *Tectonophysics*, vol. 26, p. 229, 1975.

[25] J. H. Illies and K. Fuchs (Eds.), *Approaches to taphrogenesis*, Stuttgart, 1974.

[26] D. McKenzie, *Earth Planet. Sci. Lett.*, vol. 40, p. 25, 1978.

[27] E. V. Artyushkov, *Geodynamics* [in Russian], Moscow, 1979.

[28] M. N. Toksoz and P. Bird, *Tectonophysics*, vol. 41, p. 181, 1977.

[29] B. F. Windley, *The evolving continents*, Chichester – New York – Brisbane, 1977.

[30] A. E. Scheidegger, *Principles of Geodynamics*, Berlin, 1982.

[31] D. L. Turcotte and G. Schubert, *Geodynamics. Applications of continuum physics to geological problems*, New York, 1982.

[32] P. Morgan, *Phys. Chem. Earth*, vol. 15, p. 107, 1984.

[33] E. V. Artyushkov, *Physical tectonics* [in Russian], Moscow, 1993.

[34] L. I. Lobkovskii, *Geodynamics of spreading zones, Subduction, and two-layer plate tectonics* [in Russian], Moscow, 1988.

[35] G. Ranalli, *Rheology of the Earth*, London, 1995.

[36] O. Sorokhtin and S. A. Ushakov, *Nature of the Earth's tectonic activity* [in Russian], Moscow, 1993.

[37] K. R. Furlong and D. S. Chapman, *Rev. Geoph.*, vol. 25, p. 1255, 1987.

[38] A. Kroner (Ed.), *Precambrian plate tectonics*, Amsterdam – Oxford – New York, 1981.

[39] S. R. Taylor and S. M. McLennan, *Continental crust: its composition and evolution* [Russian translation], Moscow, 1988.

[40] R. A. Jamieson, *Geol. Rundschau*, vol. 80, p. 321, 1991.

[41] M. Brown, *J. Geol. Soc. London*, vol. 150, p. 227, 1993.

[42] F. S. Spear, *Metamorphic phases equilibria and pressure – temperature – time paths*, Washington, 1993

[43] C. Ruppel and K. V. Hodges, *Tectonics*, vol. 13, p. 17, 1994.

[44] N. L. Dobretsov, V. V. Reverdatto, V. S. Sobolev, et al., *Facies of metamorphism* [in Russian], Moscow, 1970.

[45] V. V. Reverdatto, V. N. Sharapov, and V. G. Melamed, *Contr. Miner. Petrol.*, vol. 29, p. 310, 1970.

[46] V. V. Reverdatto, V. N. Sharapov, and R. M. Slobodskoy, *Contr. Miner. Petrol.*, vol. 36, p. 195, 1972.

[47] A. N. Dudarev, V. A. Kudryavtsev, V. G. Melamed, and V. N. Sharapov, *Heat transfer in magmatogene processes* [in Russian], Novosibirsk, 1972.

[48] G. Buntbarth, in: *Equilibrium and kinetics in contact metamorphism: the Ballachulish igneous complex and its aureole*, Berlin – Heidelberg, p. 379, 1991.

[49] K. R. Furlong, R. B. Hanson, and J. R. Bowers, in: *Contact metamorphism*, Chelsea, p. 437, 1991.

[50] V. S. Golubev and V. N. Sharapov, *Dynamics of endogenic ore formation* [in Russian], Moscow, 1974.

[51] V. N. Sharapov and V. S. Golubev, *Dynamics of magma – rock interaction* [in Russian], Novosibirsk, 1976.

[52] B. W. D. Yardley, *An introduction to metamorphic petrology*, New York, 1989.

[53] M. D. Barton, D. A. Battles, G. E. Bebout, et al., in: *Metamorphism and crustal evolution, western conterminous United States*, New York, vol. VII, p. 110, 1988.

[54] R. B. Hanson and M. D. Barton, *J. Geophys. Res.*, vol. 94, no. B8, p. 10363, 1989.

[55] J. J. De Yoreo, D. R. Lux, and C. V. Guidotti, *Tectonophysics*, vol. 188, p. 209, 1991.

[56] V. V. Reverdatto and A. S. Kalinin, *Two-dimensional models of metamorphism and anatexis in folded regions of the crust. 1. Model of magmatic intrusive*, Geologiya i Geofizika (Soviet Geology and Geophysics), vol. 30, no. 6, p. 63(54), 1989.

[57] V. V. Reverdatto and A. S. Kalinin, *Two-dimensional models of metamorphism and anatexis in folded regions of the Earth's crust. 2. Model of fluid flow*, Geologiya i Geofizika (Soviet Geology and Geophysics), vol. 30, no. 8, p. 41(37), 1989.

[58] V. V. Reverdatto and A. S. Kalinin, *Two-dimensional models of metamorphism and anatexis in folded regions of the Earth's crust. 3. Combined fluid-magmatic model, comparison with other models, and analysis of the problem*, Geologiya i Geofizika (Soviet Geology and Geophysics), vol. 31, no. 6, p. 3(1), 1990.

[59] P. C. England and A. B. Thompson, *J. Petrol.*, vol. 25, p. 894, 1984.

[60] J. B. Brady, *J. Petrol.*, vol. 29, p. 1187, 1988.

[61] V. V. Reverdatto, V. S. Sheplev, and O. P. Polyanskii, *Petrologiya*, vol. 3, no. 1, p. 37, 1995.

[62] V. S. Sheplev and V. V. Reverdatto, *Dokl. RAN*, vol. 334, no. 1, p. 103, 1994.

[63] L. I. Lobkovsky, A. T. Ismail-Zadeh, S. S. Krasovsky, et al., *Tectonophysics*, vol. 268, p. 281, 1996.

[64] G. Einsele, *Sedimentary basins. Evolution, facies, and sediment budget*, Berlin – Heidelberg, 1992.

[65] R. Govers, *Dynamics of lithospheric extension: a modeling study*, Utrecht, 1993.

[66] T. Peper, *Tectonic control on the sedimentay record in foreland basin. Inferences from quantitative subsidence analyses and stratgraphic modelling: Thesis*, Amsterdam, 1993.

[67] R. Zoetemeijer, *Tectonic modelling of foreland basins, thin skinned thrusting, syntectonic sedimentation and lithospheric flexure: Thesis*, Amsterdam, 1993.

[68] S. L. Harley, *Geol. Mag.*, vol. 126, p. 215, 1989.

[69] M. V. Mints, V. N. Glaznev, A. N. Konilov, et al., *Early Precambrian of northeastern Baltic Shield: paleogeodynamics, structure and evolution of continental crust* [in Russian], Moscow, 1996.

[70] Ch. Beaumont, S. Ellis, J. Hamilton, and Ph. Fullsack, *Geology*, vol. 24, p. 675, 1996.

[71] S. Ellis, *Geology*, vol. 24, p. 699, 1996.

[72] W. G. Ernst, *Geology*, vol. 16, p. 1081, 1988.

[73] G. A. Houseman, D. P. McKenzie, and P. Molnar, *J. Geophys. Res.*, vol. 86, no. B7, p. 6115, 1981.

[74] K. R. Furlong, D. S. Chapman, and P. W. Alfeld, *J. Geophys. Res.*, vol. 87, no. B3, p. 1786, 1982.

[75] Y. Ida, *J. Geophys. Res.*, vol. 88, p. 7449, 1983.

[76] Z. Garfunkel, C. A. Anderson, and G. Schubert, *J. Geophys. Res.*, vol. 91, no. B7, p. 7205, 1986.

[77] P. G. Silver, R. W. Carlson, and P. Olson, *Ann. Rev. Earth Planet. Sci.*, vol. 16, p. 477, 1988.

[78] T. N. Jordan, A. L. Lerner-Lam, and K. C. Creager, in: *Mantle convection: Plate tectonics and global dynamics*, New York, p. 97, 1989.

[79] T. Lay, *Ann. Rev. Earth Planet. Sci.*, vol. 22, p. 33, 1994.

[80] D. Olbertz, *The long-term evolution of subduction zones: a modelling study. Thesis*, Utrecht, 1997.

[81] N. L. Dobretsov and A. G. Kirdyashkin, *Geodynamics of deep-seated process* [in Russian], Novosibirsk, 1994.

[82] A. I. Chemenda, M. Mattauer, and A. N. Bokun, *Earth Planet. Sci. Lett.*, vol. 143, p. 173, 1996.

[83] J. P. Platt, *Phil. Trans. Royal Soc. London*, vol. A321, p. 87, 1987.

[84] R. A. Jamieson and C. Beaumont, *Tectonics*, vol. 7, p. 417, 1988.

[85] R. G. Coleman and X. Wang (Eds.), *Ultrahigh pressure metamorphism*, Cambridge, 1995.

[86] A. B. Thompson and Ph. C. England, *J. Petrol.*, vol. 25, p. 929, 1984.

[87] R. J. H. Loosveld and M. A. Etheridge, *J. Metamorphic Geol.*, vol. 8, p. 257, 1990.

[88] R. A. Jamieson and C. Beaumont, in: *Evolution of metamorphic belts*, Oxford, p. 117, 1989.

[89] J. C. Schumacher, R. Schumacher, and P. Robinson, *Geol. Soc. London Special Publ.*, no. 34, p. 453, 1989.

[90] M. Sandiford, N. Martin, S. Zhou, and G. Fraser, *Earth Planet Sci. Lett.*, vol. 107, p. 164, 1991.

[91] F. S. Spear, M. J. Kohn, F. Florence, and T. Menard, *J. Metamorphic Geol.*, vol. 8, p. 683, 1990.

[92] T. R. Armstrong, R. J. Tracey, and W. E. Hames, *J. Metamorphic Geol.*, vol. 10, p. 415, 1992.

[93] D. Robinson and R. E. Beavins, *Earth Planet. Sci. Lett.*, vol. 92, p. 81, 1989.

[94] V. I. Shul'diner, *Contrasting thermal settings of granulite formation: PTt-paths and geotherms of the Ladoga (Baltic Shield) and Sutam (Aldan Shield) complexes*, Geologiya i Geofizika (Russian Geology and Geophysics), vol. 39, no. 6, p. 778(791), 1998.

[95] V. S. Sheplev and V. V. Reverdatto, *Dokl. RAN*, vol. 361, p. 392, 1998.

[96] V. V. Reverdatto and O. P. Polyanskii, *Dokl. AN*, vol. 325, p. 1017, 1992.

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