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Possible Mechanism of Horizontal Overpressure Generation of the Khibiny, Lovozero, and Kovdor Ore Clusters on the Kola Peninsula

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Abstract—The paper discusses questions related to the generation of increasing crustal horizontal compressive stresses compared to the idea of the standard gravitational state at the elastic stage or even from the prevalence of horizontal compression over vertical stress equal to the lithostatic pressure. We consider a variant of superfluous horizontal compression related to internal lithospheric processes occurrin in the crust of orogens, shields, and plates. The vertical ascending movements caused by these motions at the sole of the crust or the lithosphere pertain to these and the concomitant exogenic processes giving rise to denudation and, in particular, to erosion of the surfaces of forming rises. The residual stresses of the gravitational stressed state at the upper crust of the Kola Peninsula have been estimated for the first time. These calculations are based on the volume of sediments that have been deposited in Arctic seas beginning from the Mesozoic. The data speak to the possible level of residual horizontal compressive stresses up to 90 MPa in near-surface crustal units. This estimate is consistent with the results of in situ measurements that have been carried out at the Mining Institute of the Kola Science Center, Russian Academy of Sciences (RAS), for over 40 years. It is possible to forecast the horizontal stress gradient based on depth using our concept on the genesis of horizontal overpressure, and this forecasting is important for studying the formation of endogenic deposits.

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INTRODUCTION

The manifestation of horizontal overpressure is a crucial problem for mining geologists in the mining of ore deposits. The origin of these stresses currently remains unknown. They are frequently referred to as far-ranging pressure from the boundaries of lithospheric plates and occasionally to planetary stresses caused by the Earth's rotation. Publications of the late 20th century link these to residual stresses of preceding epochs (Ponomarev, 1969, 1971; Volokh et al., 1972; Herget, 1973; Sykes and Sbar, 1973; Voigth and Pierre, 1974; Markov 1980, 1985). In this paper, we present the results of forecasting horizontal overpressure, the generation of which is related to the possible long-term existence of residual stresses of the gravitational stressed state (GSS) in upper crustal rocks.

According to the classification proposed by Davidenkov (1936), two types of residual stresses are distinguished. Stresses of the first type are balanced at the macro- and megascopic levels within a sample or a structural block; in the second type, stresses are balanced at the microlevel close to mineral grains in size.

Our proposed method for estimating horizontal overpressure is based on the assumption that their sources are residual stresses of the first type. In contrast to residual stresses of the second type (Aitmatov, 1987), these stresses are not self-balanced at the mesoscopic level. Complete release of these stresses is possible after the release of lateral compression.

The estimation of stresses relies on the amplitudes of surface denudation. When the available data make it possible to recognize stages with different rates of rock exhumation, it becomes possible to recognize the stages of stress variation related to the formation of fracture reservoirs of mineral-forming fluids (Petrov et al., 2015). In terms of the genesis of horizontal overpressure considered in this paper, it is possible to forecast the horizontal stress gradient based on depth, and this is important for assessing deep-seated mineralization trends as deposits form (Prokof'ev and Pek, 2015). The possibility of forecasting the depth of variation in the geodynamic type of the stressed state, which in turn sharply modifies the conditions of the fluid regime in a mineral-forming system at an endogenic deposit, is also important for studying ore formation. Erosion removal proper is significant for comprehension of deep-seated ore formation under paleogeodynamic conditions.

HORIZONTAL OVERPRESSURE: MECHANISMS OF GENERATION

As mentioned above, the motion of lithospheric plates is considered now by mining geologists to be the most probable cause of the high values of horizontal compressive stresses (Brown and Hoek, 2002; Savchenko and Gorbatsevich, 2012). Two mechanisms of intracontinental stresses are known in lithospheric plate theory. In one of them, lateral pressure is responsible for these stresses, and in the other, they are shear stresses acting at base of the crust on the side of the moving mantle. Both lateral pressure and shear stresses on the side of the moving mantle are active forces in the zone of collision between active and passive lithospheric plates, e.g., the Indian and the Eurasian plates.

Simple physical reasoning allows us to suggest that at a great distance from plate boundaries, the resultant lateral compressive stresses (total GSS and far-ranging pressure) will tend to be distributed uniformly with depth. It can be expected that deviation from this rule would be observed near plate boundaries, where over- and underthrust faults are predominant. In addition, the level of these stresses should weaken with distance from plate boundaries due to widening of the arc of influence along which the compressive stresses act.

The second concept related to deep penetration of the moving mantle beneath a passive continental plate ensures transmission of horizontal compressive stresses for a great distance with a gradual increase in their level with distance from the axis of the oceanic ridge (rift axis). This implies that horizontal compression should increase with depth rather than weaken, because the forces creating compression are applied to the sole of the crust.

Voigth and Pierre (1974), Gudman (1987), and Rebetsky (2008a, 2008b) considered the geomechanical relationships leading to the possible formation residual stresses only within regions of the active GSS. The appearance of such stresses is related to (1) deformation of rocks under the action of their dead weight in the absence of lateral extension; (2) irreversible deformations at great depth, which create additional horizontal compression compared to the elastic model proposed by Dinnik (1926); (3) the presence of ascending movements in the Earth's crust accompanied by denudation of the surface and resulting in partial vertical unloading of rocks.

Later, Turcotte (1973, 1974) suggested that erosion of uplifted regions might be a cause of additional extension rather than additional horizontal compression. This phenomenon was related to the sphericity of the planet and the failure of the first of the aforementioned conditions at the stage of ascending movements in the crust. It was thought that theincreasing length of the arc in the ascending layer mostly compensates residual compressive stresses (Fig. 1a). The cooling of rocks in the process of uplifting results in thermoelastic shortening, which also leads to additional horizontal extension stresses. Calculations carried out by Haxby and Turcotte (1976) confirmed this hypothesis. This publication ended discussion of the residual stress effect.

In order to understand how we have to treat the results obtained by Haxby and Turcotte (1976), it is necessary to recall another hypothesis related to the effect of the Earth's sphericity on stresses in developing troughs (Kosygin and Magnitsky, 1948; Magnitsky, 1965). It has been suggested that fold formation in sedimentary basins is caused by the transition of a spherical surface through a chord where the spherical surface sags. This is similar to what Turcotte stated (1974) and assumes additional compressive stresses in subsiding regions caused by the sphericity of the planet (Fig. 1b).

In addition, one more important geological conclusion should be recalled, made at the beginning of the 20th century and known as the Karpinsky rule (Karpinsky, 1919), which emphasizes that near a growing uplift with intense erosion on its slopes, there are always subsiding basins that actively accumulate most of the sedimentary rocks. According to this rule, the processes in the basin and an on the rise actively interact in tectogenesis and cannot be considered separately.

The D. Turcotte and V.A. Magnitsky's hypotheses suggest that the vertical boundaries of bodies where differently directed vertical movements take place are not deformed. At the same time, it is quite clear that submerging and emerging crustal areas adjoin each other and vertical movements in one of them compensate those in the neighboring area (Goncharov, 1993). Thus, the hypothesis about nondeformed vertical boundaries dividing uplifted and subsided areas should be regarded as incorrect. As data on natural stresses show (Rebetsky, 2015), the deformation scheme presented in Fig. 1c looks more correct. The stress states in neighboring uplifted and subsided crustal areas were calculated in (Rebetsky, 2011). They take into account the effect of elastic change in volume on a viscous flow. In fact, the effect of residual GSS stresses is taken into account. It has been shown that a stress state of horizontal compression is formed in the uplifted crust, whereas the subsided areas are characterized by horizontal extension.

RESULTS OF INSTRUMENTAL STRESS MEASUREMENTS

The main results of in situ stress measurements carried out in mining works in various regions and in mines of the Kola Peninsula are considered below.

General Patterns of Stresses Measured in Mining Works

Brown and Hoek (1978), Herget (1973), Brady and Bzown (2004), Potvin et al. (2007), Kozyrev and

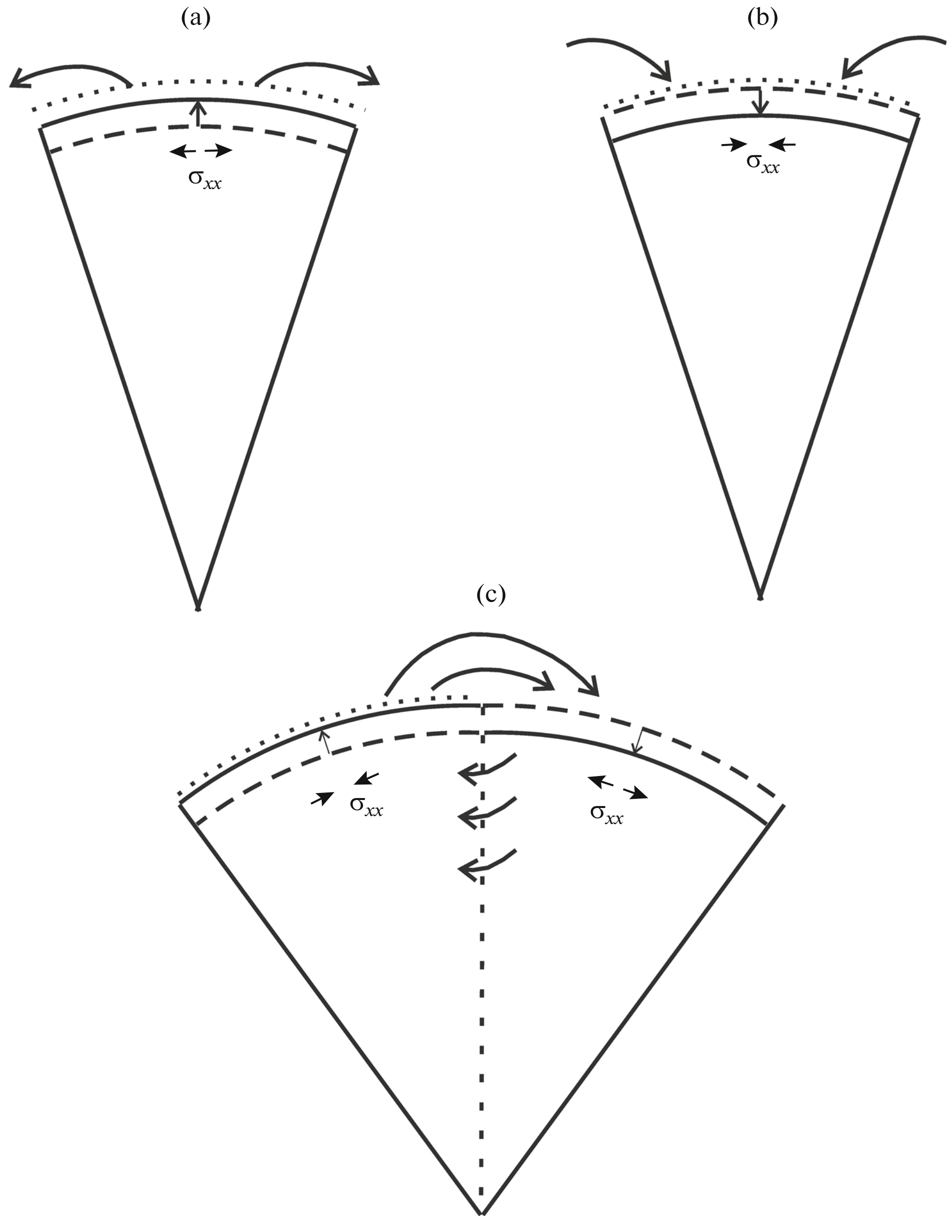

Fig. 1. Change in geometry of near-surface layers on the spherical Earth: (a) for rises and (b) for basins, after Turcotte (1974) and Magnitsky (1965), see text for explanation; and (c) crustal rises and basins combined into common cell. Arrows at surface show denudation of rises (a) and sedimentation in basins (b). Vertical dashed line and arrows (c) show that boundary between rise and basin is conventional, so that intracrustal flow of matter takes place here. Additional horizontal compression or extension (σ_{xx}) arising in regions of rise and basin is indicated according to models of D. Turcotte, A.V. Magnitsky and after results of natural stress study (Rebetsky, 2015).

Savchenko (2009), and Zubkov et al. (2010) have generalized measured stresses for many underground works and open pits on almost all continents. The result of these observations has shown that the depth distribution of vertical stresses, in general, approaches the lithostatic pressure. Stresses averaged in the lateral direction at each measurement point for a depth smaller than 1 km are characterized by the widest scattering from 0.3 to 3.5 relative to the vertical stresses (Fig. 2). The maximum values of horizontal compressive stresses decrease with depth, approaching 0.8–1.0 of the lithostatic pressure at a depth of 2–3 km (vertical dashed line). For each depth level, the minimum ratios of the mean lateral to vertical stress remain almost invariable at ~0.3.

The left boundary of the distribution zone of the in situ measurement results in Fig. 2 corresponds to the elastic behavior of rock massifs (Dinnik, 1926) in regions without surface denudation. Even sagging of the surface and sedimentation are possible in these

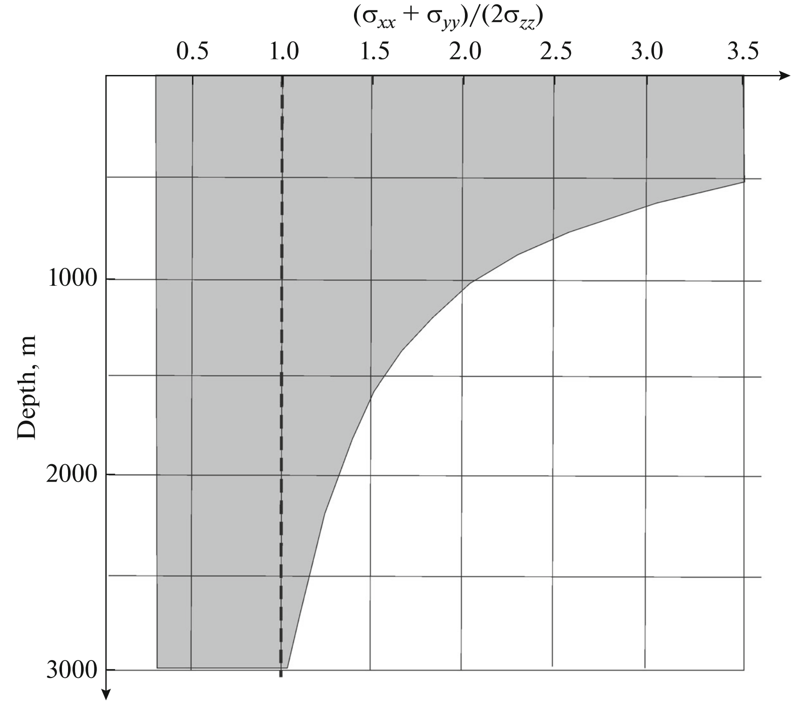

Fig. 2. Area of ratio of one-half of sum of horizontal stresses to doubled vertical stress obtained for large number of in situ measurements. Simplified after Brady and Bzown (2004). Vertical dashed line separates regions of stresses with geodynamic regimes of horizontal extension (right) and horizontal compression (left).

crustal areas. The right boundary of the in situ measurement zone at depths less than 500 m determines establishment of the limiting state (Rebetsky, 2008a, 2008b) in a geodynamic horizontal compression regime at a moderate fluid pressure level that reaches 0.3–0.4 of the lithostatic pressure for an internal friction coefficient of 0.55–0.60 and natural cohesive rock strength of 10–15 bar. A strength of 20–70 bar for minor samples of sedimentary rocks is always higher than that of a natural fractured massif. These parameters are quite appropriate for the state of rock massifs under near-surface conditions.

If the same rock strength is used to interpret the measurement results at a great depth (2–3 km), then the maximum ratio of horizontal to vertical stress (0.8–1.0) existing here corresponds to subcriticality, when the limit of fracture flowability is not overcome (no displacement along fractures). In other words, purely elastic behavior of rocks apparently takes place with a stressed state after Dinnik (1926). Brady and Bzown (2004) also noted, after Hast (1974), that vertical and horizontal stresses converge at depths greater than 2 km.

Data on stresses in deep-seated layers of crystalline crust obtained with geophysical methods for the off-borehole space of the deep OKU Hole in Finland have shown that smoothing of vertical and horizontal stresses take place at a depth of 1.2 km (Savchenko and Gorbatsevch, 2012). Special core analysis techniques for the Kola Superdeep Hole SG-3 have shown that the ratio of horizontal to vertical compression at depths of >2 km is close to 0.8 (Gorbatsevich and Il'chenko, 1999; Savchenko, 2004; Gorbatsevich and Savchenko, 2009).

In the data obtained, one fact attracts attention: beginning from a certain depth, the ratio of averaged lateral to vertical stress varies from >1 to <1. This implies that in the same areas of the crust, the horizontal compression regime takes place at upper levels, whereas the regime of horizontal extension is inherent to the lower levels, where maximum compression is oriented nearly vertically. If it is assumed that external forces are responsible for the formation of an anomalous stressed state, the question arises: why did these external forces not give rise to increasing horizontal compression at lower levels. Thus, the in situ data on the upper crustal units do not support the concept of stressed state formation due to boundary tangential stresses acting along the sole of the crust.

Stress Measurements at Mines of the Khibiny, Lovozero, and Kovdor Massifs

Measurements of stress by unloading method (Markov, 1977; *Upravlenie ...*, 1996) at various levels from 92 to 600 m at the Kirovsk, Yukspor, Central, and Rasvumchorr mines in the Khibiny massif showed a prevalence of horizontal over vertical compressive stresses. If vertical stresses are close, on average, to the lithostatic pressure, then horizontal stresses can exceed them by several times. It has been established that in more rigid host rocks, the stresses are higher than in ore by 50–100%. At various mines, the stresses in host rocks vary from 20 to 70 Mpa, and in ore, from 15 to 45 MPa.

The stresses increase close to geological disturbances and faults. For example, at the Rasvumchorr Mine, the horizontal stresses in the body of a fault (50–60 MPa) are twice the level of stresses of the mine at the same and even deeper levels. Such effects of a fault's influence have also been noted at the Karnasurt Mine of the Lovozero massif. The level of horizontal stresses measured at various mines under near-surface conditions (10–100 m) is 20–40 MPa and corresponds to the level of lithostatic pressure at a depth of 700–1300 m.

An insignificant increase in depth (a few hundred meters) results in increased measured stresses. For example, at the Kirovsk Mine, a depth increase by 230 m gives rise to growth of horizontal stresses by 17 MPa, whereas lithostatic pressure increases by ~7 MPa. For the Yukspor and Rasvumchorr mines, the increase in depth by 160–180 MPa led to increased horizontal stresses for 30 MPa. *With a further increase in depth, the increase in horizontal compressive stresses slows, and this results in a decrease in the ratio of horizontal to vertical compressive stresses from 5–10 MPa near the surface to 2–3 MPa at the level of valley bottoms in the Khibiny massif*. After achievement of the valley level, the stress state changes sharply (*Seismichnost* ..., 2002). Whereas Markov (1977) and *Upravlenie* ... (1996) only suggested, in the monograph (*Destruktsiya zemnoi*..., 2012) in the section written by mine mechanics from the Mining Institute, Kola Science Center, RAS, it is stated that at a depth of 1000–1500 m, *the geodynamic horizontal compression regime changes from horizontal compression to extension*.

At the mines of the Lovozero massif, the level of measured horizontal compressive stresses is somewhat higher than in the Khibiny massif (70–80 and 80–90 MPa for the Karnasurt and Umbozero mines, respectively). At the Zhelezny Mine of the Kovdor massif, the horizontal compressive stresses amount to 3–27 MPa (depth is 100–300 m) and are nonuniformly distributed throughout open pit (Rybin et al., 2012).

The results of very recent direct stress measurements at mines of the Khibiny and Lovozero massifs (*Upravlenie* ..., 1996) have shown that the *stressed state is close to axisymmetric* here (Fig. 3), with a tangential orientation of maximum compression. Note that an axisymmetric stressed state cannot be explained by remote influence of stresses from any direction. In this case, compression in the direction of external lateral pressure and extension in the orthogonally lateral direction should be observable.

Analysis of the results of in situ measurements in the Kola Region allows us to assume with a high probability that the stressed state regime in the crust is not related to the remote effect of horizontal compression from collision zones and rifting.

RESIDUAL GSS STRESSES

In this section, we discuss the main statements of the theory on calculating residual GSS stresses. They manifest themselves in emerging massif along with parallel denudation, but begin to form at the sedimentation stage, when rock undergoes subsidence.

Initial Stage of GSS Formation

The stressed state of the crust corresponding to action of only mass forces is represented by two deep levels, which characterize different rock reactions to loading. In the upper part of the crust (from hundreds of meters to 1–2 km depending on the fluid regime), the rocks under GSS conditions are deformed only elastically; irreversible deformations are formed at deeper levels. Rebetsky (2008a, 2008b) presents expressions for estimating GSS stresses at various depths. Figure 4a shows the change in stresses in the elementary bulk of rocks in a sedimentary basin that first undergoes subsidence, an increase in vertical and horizontal compressive stresses (sedimentation stage), and then elastic loading caused by exhumation.

In the upper part of the section—the region of purely elastic deformation—the most rapid linear growth with depth of the deviator stress level and maximum shear stresses is observed. The straight line OA corresponds to this part of the crust (Dinnik, 1926) (Fig. 4a). The vertical stresses for the GSS of the considered stage are always active and represent principal compressive stresses ($\sigma_{zz} = \sigma_3$), whereas the horizontal reactive stresses ($\sigma_{xx} = \sigma_{yy} = \sigma_1 = \sigma_2$) caused by lateral constraint of rocks ($\varepsilon_{xx} = \varepsilon_{yy} = 0$) determine the impossibility of their free deformation in the horizontal direction. The relationship between horizontal and vertical stresses is determined as $\nu/(1 - \nu)$, where ν is the Poisson ratio.

The growth of tangential GSS stresses with depth may lead in the end to violation of the elastic equilibrium of rocks due to the most probable causes: (1) brittle failure and (2) attainment of the elastic limit. A hypothesis exists (Nikolaevsky, 1979) that readily corresponds to the observed data: crustal rocks achieve a true plastic state (irreversible deformation by means of intragranular or intracrystalline dislocation flow) at the Moho discontinuity. The elastic equilibrium in rocks of the upper and middle crust is violated due to the formation and activization of multiple fractures as brittle strength defects variable in scale. The cataclastic fluidity that arises in rocks gives rise to irreversible deformation caused by fracture flow. At a suitable scale of averaging, the displacement along multitude of variously oriented fractures looks as plastic flow.

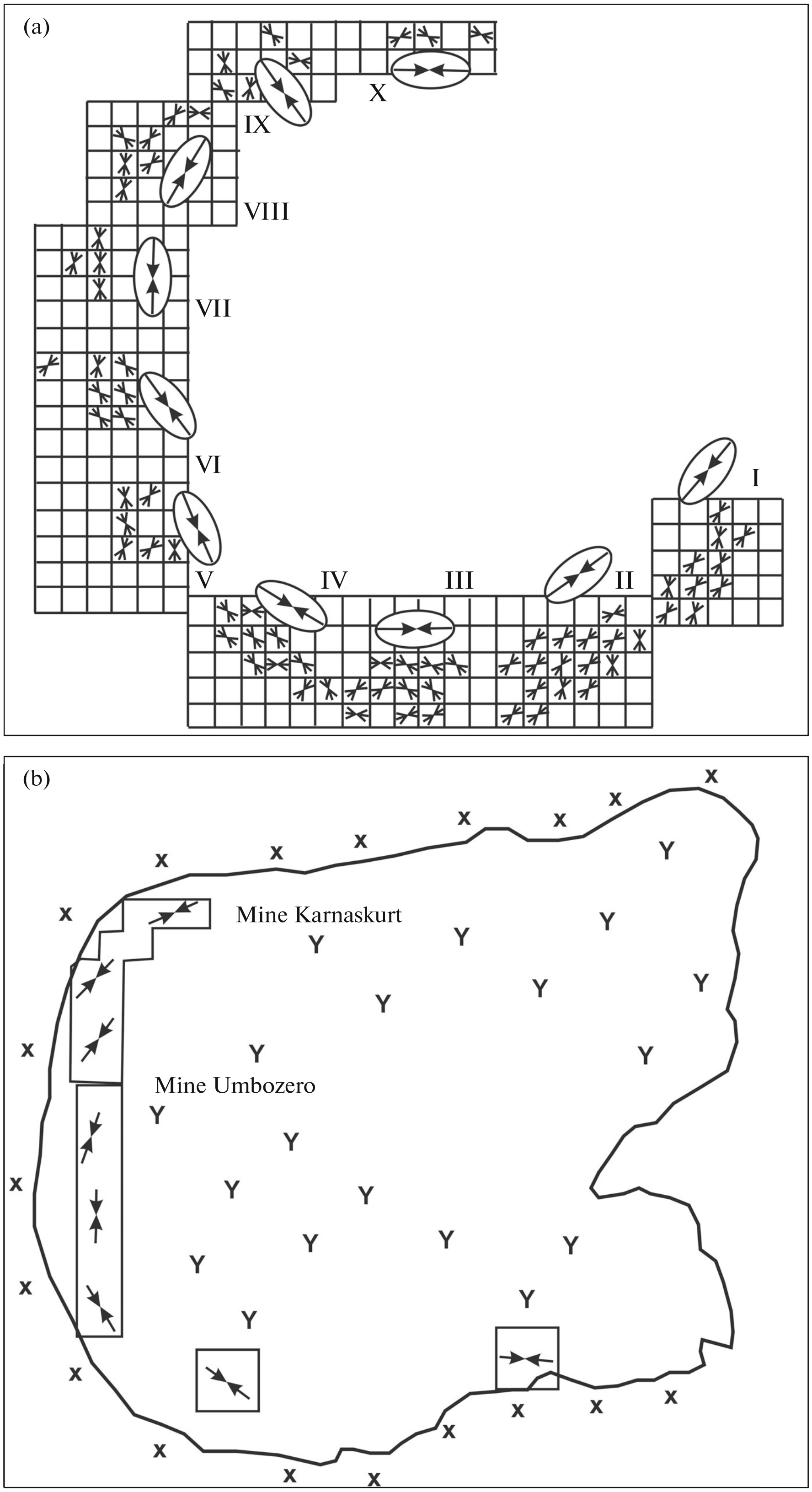

Fig. 3. Orientation of axes of maximum compression from results of in situ measurements in mines of (a) Khibiny and (b) Lovozero massifs, after *Upravlenie …*, (1996). Arrows indicate direction of maximum horizontal compression after results of in situ measurements; arrows within ellipses in panel (a) show averaged direction according to total determinations. Location of eudialytic lujavrite (Y) in Lovozero massif is shown in panel (b). Lujavrite, foyaite, and urtite occur elsewhere in this massif, as well as Archean gneiss, granite gneiss, and migmatites in the outer zone (X). Roman numerals are sequence numbers of integrated in situ measurements.

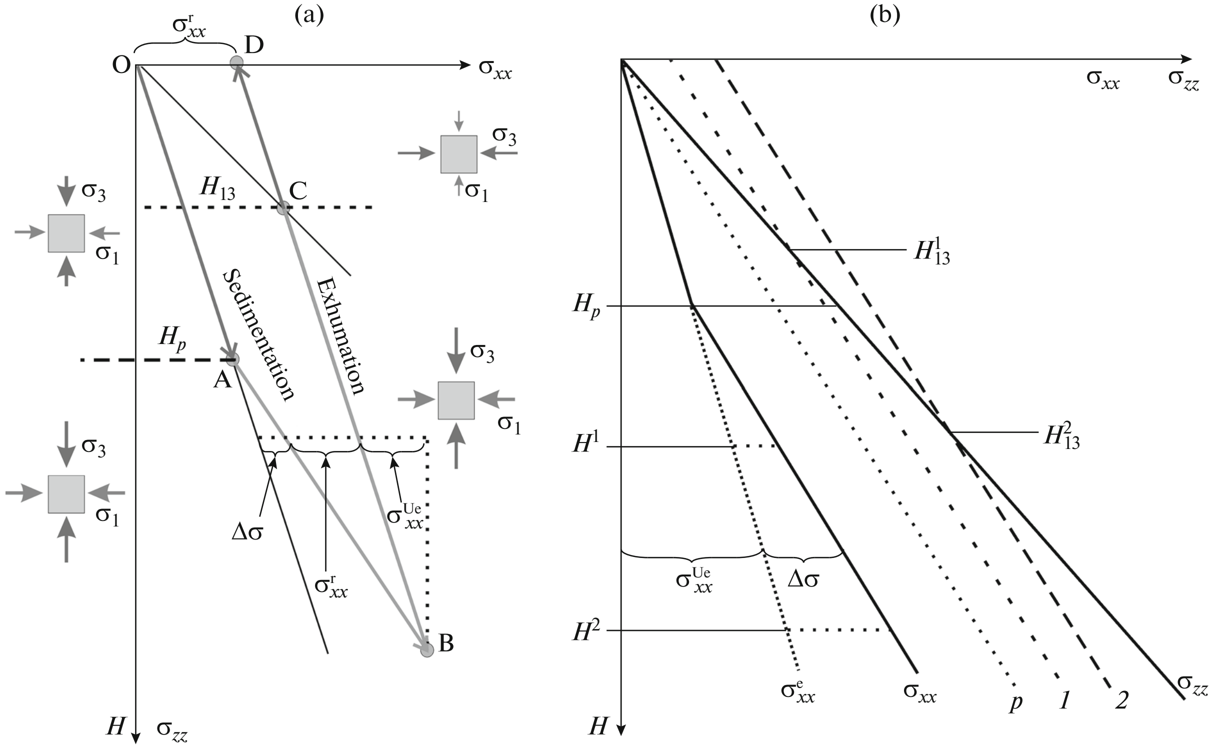

Fig. 4. Formation of residual GSS stresses in regions of ascending vertical movements accompanied by surface denudation (exhumation of rock). (a) Diagram of loading for elementary body within sedimentary basin in the process of sedimentation (arrow downward) and exhumation (arrow upward). Vertical coordinate axis: depth of elementary body in sedimentary basin and level of vertical compressive stresses (σ_{zz}). Horizontal coordinate axis: horizontal compressive stresses (σ_{xx}). Long-dash line is depth of roof of elastic–cataclastic deformation (H_p); short-dash line is depth of re-indexation of principal stresses (H_{13}) for the stage of rock exhumation; point **C** is depth, for which horizontal compressive stresses at exhumation stage coincide with vertical stresses ($\sigma_{xx}/\sigma_{zz} = 1$). Loading path : OA, purely elastic loading of body within region $H \leq H_p$; AB, loading at elastic–cataclastic deformation stage ($H \geq H_p$); BD, stage of unloading and exhumation of rock; CD, area of discharge where horizontal compression becomes greater than vertical $|\sigma_{xx}| > |\sigma_{zz}|$. Orientation of principal stress axes shown on left ($\sigma_1 > \sigma_3$, compression is negative) at sedimentation stage (initial GSS stage) and on right at exhumation stage. (b) Distribution of stresses over depth for initial GSS stage (solid lines for stresses σ_{xx} and σ_{zz}) and for exhumation stage for different variants of surface denudation amplitude (lines *p*, *1*, and *2* with dotted, short-dashed, and long-dashed lines corresponding to three depth levels: $H_p < H_1 < H_2$). Depths H^1_{13} and H^2_{13} correspond to level of reindexation of principal stresses for two different cases of rock exhumation (lines *1* and *2*). σ^{Ue}_{xx} are horizontal stresses released during elastic discharge (exhumation), $\Delta\sigma$ is additional compression arising under cataclastc flow during subsidence of rocks, σ^{r}_{xx} (a) are residual compressive stresses at rock exhumation stage, and σ^{e}_{xx} (b) elastic stresses in case of ideal elastic body.

In the region of cataclastic fluidity (below depth H_p), where the Coulomb–Mohr limit ratio is fulfilled, the growth of these stresses with depth slows. This is caused by the proximity of vertical stresses to the lithostatic pressure of rocks and the requirement that the elastic limit be fulfilled. Because of this, additional horizontal compressive stresses ($\Delta\sigma$) appear here. These stresses are added to compressive stresses that would act at this depth if the GSS were purely elastic. At depths $>H_p$, the ratio of normal horizontal to vertical stresses will be higher than in the case of a pure elastic state. In Fig. 4a, line AB, which is straight in the case of a constant ratio of fluid to lithostatic pressure, corresponds to this deep level of stresses.

Variations of the Stressed State for the Vertical Displacement of Rocks Accompanied by Surface Erosion

In the case of rock exhumation caused by ascending vertical movements and by surface denudation for the *dH* value, the vertical load acting on the elementary body of the geological medium decreases. According to mechanical equations, unloading will develop according to an *elastic law* (Fig. 4a, line BCD).

When rock occurs at a depth $H^0 \leq H_p$ above the part of the crust undergoing cataclastic (pseudoplastic, geomechanical) flow, then the purely elastic stress state takes place therein. In this case, elastic unloading will lead to a decrease in normal stresses acting in the horizontal and vertical directions in the $\nu/(1-\nu)$ ratio i.e., by the same law as for elastic loading (Dinnik, 1926). During such unloading, the vertical/horizontal ratio does not change. In the end, when a rock is completely exposed, the stresses acting therein are close to zero.

Another situation will take place for rocks that occur at a depth $H^0 > H_p$ before onset of exhumation (Fig. 4b, BCD is path of unloading). The relationship of horizontal and vertical stresses in these rocks does not fit the proportion $\nu/(1 - \nu)$, because additional compressive stresses appear in the course of cataclastic flow. Therefore, under elastic unloading caused by the exhumation of rock, the vertical stresses can be completely eliminated when the rock is exposed, in contrast to horizontal stresses.

Figure 4b shows the distribution of horizontal stresses with depth for the initial GSS stage, when the upper zone of the elastic state is combined with the lower zone of the elastic–cataclastic state (the counterpart of the elastic–plastic state). The deep distributions of these stresses for three different cases of surface denudation are also given here. In the first case, the layer initially located at depth H_p is then exposed at the surface (the upper boundary of the elastic–cataclastic layer). In this case, horizontal compressive stresses are lower overall than the vertical counterparts, and horizontal extension is a geodynamic type of stressed state. The two other stress states correspond to exhumation of rocks brought to the surface from great depth ($H_2 > H_1 > H_p$). In this case, the layer for which the level of horizontal compressive stresses is higher than of vertical counterparts, appears in the upper part of massif. Two deep-seated zones with different geodynamic types of stressed state always exist in this case as well. In the upper part of the massif (Fig. 4b), above a depth $H^k{}_{13}$ ($k = 1, 2$), the horizontal compressive stresses are higher than the vertical stresses, while below $H^k{}_{13}$ the horizontal compressive stresses are lower.

According to (Rebetsky, 2008a, 2008b), the level of residual GSS stresses does not depend on the elastic properties of rocks and is controlled by strength of the geological medium, including the weakening effect of fluid. As a result, *a weak dependence of residual stresses on the compositional heterogeneity of mountain massifs is combined with a strong dependence on the structure of the geological medium and fluid pressure in the fracture–porous space*.

The possible appearance and long-term existence of horizontal overpressure caused by residual elastic GSS deformations is an integral property of the upper crust and apparently of local areas in the middle crust related to the limited deformation of mountain massifs at the depth and level of deviator stresses, estimated at tens of MPa, which is much lower than the true rock fluidity limit (300–400 MPa). This makes it possible to retain these stresses for a sufficiently long time (Watts and Cochran, 1984).

A similar phenomenon in the subcrustal lithosphere does not take place due to the effect of dislocation creep on the deformation mechanism. As follows from petrophysical analysis of mantle inclusions from a depth of 60–200 km (Mercier, 1980), the level of deviator stresses is not high (5–15 MPa). The dislocation plasticity mechanism determines the impossibility of the long-term existence of residual elastic GSS deformation controlled by vertical motions. In the lower and middle crust under conditions close to pseudoplastic deformation (Nikolaevsky, 1979), the long-term existence of deviator stresses above a few tens of MPa is unlikely.

ESTIMATATION OF STRESSES BASED ON AMPLITUDES OF DENUDATION–EROSION REMOVAL FROM FENNOSCANDIA OVER THE MESOZOIC AND CENOZOIC

Geological and Geomorphic Structure

Thick Archean and Proterozoic sequences are predominant on the Kola Peninsula, which occupies the northeastern part of the Fennoscandian (Baltic) Shield. The oldest Archean rocks are represented by high-grade metamorphic and intensely deformed gneisses and granites. Proterozoic rocks are more diverse in composition: quartzite, crystalline schist, sandstone, marble, and subordinate gneisses are intercalated with greenstones. The Archean Belomorian and Proterozoic Karelian foldings are distinguished. Paleozoic rocks make up alkaline and alkaline–ultramafic intrusions; the Khibiny and Lovozero intrusive massifs are the largest throughout the Fennoscandian Shield.

The orographic pattern of the Kola Peninsula is rather simple. Its western part is uplifted and distinguished by a dissected topography. Single tops of the Khibiny, e.g., Mt. Chasnachorr (1191 m), and of the Lovozero Tundra exceed 1000 m in height. The eastern half of the peninsula is characterized by an even, wavy topography with spot heights of 150–250 m. According to geomorphic studies, the rate of the Holocene emergence of the Khibiny and Lovozero massifs is estimated at 1–3 mm/yr (Koshechkin, 1979).

The vertical component of tectonic movements is expressed in the general arching of the northeastern Fennoscandian Shield and in the growth of local domes therein. The topography of the Kola Peninsula formed over the long period of its continental evolution under conditions of stable uprising and continuous occurrence of crystalline rocks. As a result, planation surfaces formed. Denudation–tectonic, structural–denudation, and denudation landforms are dominant. They are represented by flat-topped mountain massifs, hills, low ridges, plateaus, rock-defended plains and lowlands.

The uplifted denudation surfaces broken by faults were transformed in the Quaternary by the activity of a glacier that originated in this territory. The postglacial isostatic emergence of the northwestern Kola Peninsula reached 150–170 m over a short time span, while the northeastern Kola Peninsula likely underwent hardly any uplift.

The maximum isostatic uplifts during the interglacial stage varied from tens of meters to 100–200 m. The massifs towered above the adjacent territory, and this is caused not so much by their stability to weathering, which is occasionally even lower than for country rocks, but by contouring of massifs by faults and their uplifting. In other words, these are horsts, which have continued to grow locally until now (at least some of them). As a result of selective denudation, a quite distinct topography hs been created on Archean granite and gneiss and on Proterozoic rocks: quartzite intercalating with greenstone. This is explained by dissimilar stability to destruction of rocks belonging to Belomorides and Karelides.

The Khibiny and Lovozero massifs are similar in origin: they are nepheline syenite intrusions of the central type. According to the results of geological studies, these unique alkaline intrusions 380–410 Ma in age are related to intracrustal magma injections without contact with the surface. The massifs, 1300 km^2 and 590 km^2 in area, are close to round in plan view; the depth of lower contact is ~ 7 km.

Before the formation of the alkaline intrusion, the considered part of the Fennoscandian Shield revealed a tendency of the Earth's crust to rise (Belousov, 1962; *Geology* ..., 2002; Zarkhidze and Musatov, 1989) with a marked spatiotemporal nonuniformity of block uplifts. Over subsequent geological epochs, the entire territory of the contemporary Kola Peninsula continued to undergo gradual uplift accompanied by destruction (weathering) and removal of the Earth's upper crust. Weathering in combination with removal of material by rivers and glaciers gradually gave rise to ancient intrusions free of covering rocks. Since the Oligocene (the onset of the neotectonic stage on the Fennoscandian Shield (*The Map* ..., 1985)), the body of a cooled intrusion has been raised by tectonic processes above the adjacent plains. The Khibiny massif reveals a conic ring structure. The textural and compositional zoning of the Khibiny massif was caused by heating and metasomatic reworking of the initially monotonic zonal foyaite body under the action of a foidolite melt that intruded along the ring fault (Arzamastsev et al., 1998). The Lovozero volcanic–sedimentary suite makes up tthe roof of the Khibiny and Lovozero massifs. The difference in spot elevations between them is 3300–3900 m.

The Kovdor massif occupies the next position in terms of size after the Khibiny and Lovozero massifs and is one of the best studied alkaline–ultramafic plutons on the Kola Peninsula (*Geology* ..., 2002). The massif has a concentrically oval shape in plan view and is somewhat elongated in the meridional direction. The massif cuts through gneiss of the Kola–Belomorian Complex and has eruptive contacts with country rocks. The Kovdor massif is characterized by intense supergene alteration (Sidorenko, 1958; Ternovoi et al., 1969). The preglacial weathering mantle covers more than 60% of the massif's area.

Algorithm for Estimating Denudation–Erosion Removal from Fennoscandia over the Mesozoic and Cenozoic

According to the concept stated in the preceding section, it is necessary to know the amplitudes of denudation–erosion scouring to estimate the superfluous horizontal compressive stresses. The amplitude of removal from the Kola Peninsula is estimated from the thickness of sediments filling the water reservoirs surrounding Fennoscandia as a whole. At the present-day level of Fennoscandia, Phanerozoic stratified rocks were hardly preserved at all except for scattered residues in separate depressions and fault zones. The most complete data on the rocks preserved in this territory were collected by Kirichenko (1970), who compiled a schematic map of Paleozoic rock remnants on the Fennoscandian Shield and showed spots with rocks varying in age from Cambrian to Permian. The thickness of volcanic–sedimentary rocks of the Upper Devonian–Carboniferous Kontozero Group is estimated by L.A. Kirichenko at 800 m only for the Carboniferous part of the section; the thickness of eroded rocks remains unknown. The rocks are preserved in a caldera.

The thickness of sediments in water reservoirs surrounding Fennoscandia is estimated from the atlas of lithologic–paleogeographic maps of the world, which represents maps of continents and oceans for all epochs of the Mesozoic and Cenozoic (Ronov et al., 1979). To calculate the amounts of material removed over a certain epoch, the maximum thicknesses of sediments of the corresponding age were summed for the water areas of the Barents, Norwegian, and Northern seas. It has been accepted that the orders of magnitude of sedimentation and removal areas are commensurable. In separate epochs, one-half the thickness of sediments was taken in order to subtract the removal from the north (present-day Spitsbergen Islands, etc.), west (Britain Peninsula), south (Bohemian massif), and other lands. The thicknesses of Triassic and certaim other sediments of the Norwegian and Northern seas are given below as an example of taking two sources of removal into account (Table 1). More detailed stratigraphic subdivisions are considered in the atlas *Geology and Mineral Resources of Russian Shelves* (*Geologiya* ..., 2004), but they are selective and do not permit calculation of removal in particular epochs. It is accepted that the maximum total thickness of sediments of certain age corresponds to the maximum thickness of removed material. The calculations do

Table 1. Averaged thickness of sediments gained in framework of Fennoscandia over Mesozoic and Cenozoic eras

Epoch	Thickness, m			
	Barents Sea	Norwegian Sea	Northern Sea	Total
T_1	500	560	500 : 2 = 250	1210
T_2	1000	–	250 : 2 = 125	1125
T_3	1000	2000 : 2 = 1000	250 : 2 = 125	2125
$\mathbf{T_1–T_3}$				**4460**
J_1	250	250 : 2 = 125	250 : 2 = 125	500
J_2	100	100 : 2 = 50	100 : 2 = 50	200
J_3	500	800 : 2 = 400	100 : 2 = 50	950
$\mathbf{J_1–J_3}$				**1650**
K_1	750	500 : 2 = 250	500 : 2 = 250	1250
K_2	100	500	1000 : 2 = 500	1100
$\mathbf{K_1–K_2}$				**2350**
Pg_1	480	500 : 2 = 250	100 : 2 = 50	780
Pg_2	100	500 : 2 = 250	250 : 2 = 125	475
Pg_3	100	300	250 : 2 = 125	525
$\mathbf{Pg_1–Pg_3}$				**1780**
N_1	–	250 : 2 = 125	100 : 2 = 50	175
N_2	7	500	6	513
$\mathbf{N_1–N_2}$				**688**

Total thickness 10 928 m

not take into account the thickness of Mesozoic and Cenozoic sedimentary rocks of the Russian Plate due to lacking data, nor the thickness of Quaternary sediments due to specific sedimentation related to glaciation, as well as a lack of complete information. We also failed to take into account the body of Late Mesozoic and Cenozoic volcanic material supplied to water reservoirs over four stages (Korago et al., 2010). The apparent scouring of Triassic sediments overlain by Middle and Upper Jurassic in the Barents Sea is also beyond the scope of this paper.

Table 2. Rate of removal from Fennoscandia over various cumulative time intervals

Epoch	Time, Ma	Thickness, m	Rate, mm/yr
$T_1–N_2$	250.0	10928	0.044
$T_3–N_2$	227.4	8593	0.038
$J_1–N_2$	205.1	6468	0.031
$K_1–N_2$	142.0	4818	0.034
$Pg_1–N_2$	65.5	2468	0.038
$Pg_2–N_2$	55.0	1688	0.031
$Pg_3–N_2$	33.7	1213	0.035
$N_1–N_2$	23.8	688	0.029

Amplitudes of Removal

As follows from the calculation results (Table 1), the order of magnitude of erosion of Fennoscandia over the Mesozoic and Cenozoic reaches 11 km (Sim, 2012). Based on the thickness of possible removal, the erosion rate over various time intervals were calculated (Tables 2, 3). The order of magnitude of the erosion rate over various cumulative time intervals is commensurable and changes from 0.029 to 0.044 mm/yr. They correspond to the data known for Europe, as well as calculated for local areas in the Fennoscandian Shield (Evzerov, 2001). This allows us to hope that the results are reasonable. The rate for separate, more detailed time intervals is not as stable (0.146 to 0.009 mm/yr). This is explainable by both the rough calculation of thickness for short time interval and specific sedimentation.

It is noteworthy that various methods for calculating the removal rate (Table 2) yield maximum values for the Triassic and Paleocene. The maximum removal rates over the total time from the Early Triassic to the Pliocene inclusive (0.044 mm/yr) and over the Triassic (0.099 mm/yr) were most likely caused by removal not only from Fennoscandian, but also from elsewhere to the Barents Sea. Thus, according to paleogeographic maps (Ronov et al., 1989), in the

Early and Middle Triassic, the entire Scandinavian Peninsula was a low land except for a narrow strip in its southwestern part. At the same time, Greenland, the North Urals–Pai-Khoi, and the Timan were elevated lands in the Early Triassic; i.e., removal into the Barents Sea from the east was substantial. In the Late Triassic, the Scandinavian Peninsula, the Timan, and the Urals were a land.

The significant areas covered by continental sediments are mapped in the eastern Barents Sea, the entire present-day Kara Sea, Novaya Zemlya, and the Timan–Pechora Plate, whereas at the coast of Fennoscandia, intercalation of marine and continental sediments has been established. The NE-trending axis of the trough extends in parallel to Novaya Zemlya; the 1000 m contour line is located approximately at an equal distance from Novaya Zemlya and the Murmansk coast. Because of this, only one-half the thickness of sediments was taken into account (Table 1). A maximum rate of the shield's emergence was related to the final Hercynian Orogeny in the Urals, and its rate is apparently overestimated due to the impossibility of taking into account sedimentation related to removal from the eastern orogenic structural units.

Thus, the results of this research make it possible to integrate the geological history of the Fennoscandian Shield in the Mesozoic and Cenozoic. Over the Late Triassic–Early Jurassic, the shield underwent a long-stage planation, and this is reflected in a minimum of removal rate in the Jurassic (0.026 mm/yr, see Table 3). The weathering kaolin mantle was formed at that time as the main source of material supplied into the South Barents Basin. By the onset of the Late Jurassic, only the roots of the weathering kaolin mantle were retained on the shield, so that only less weathered material than in the preceding period was removed from the shield in the Late Jurassic–Early Cretaceous (Evzerov, 2001). The second maximum of the emergence rate and, correspondingly, the intensity of erosion are noted in the Paleocene (0.038 and 0.074 mm/yr, see Table 2, 3), marked by manifestation of the Laramide Orogeny, which was accompanied by critical tectonic and paleogeographic changes. Relatively deepwater clay sediments in the center of the Northern Sea basin gave way to continental sandy sediments supplied from actively rising Fennoscandia.

It is known that the transition zone from the Fennoscandian Shield to the Russian Platform was flooded in the Paleogene by a marine basin, which created an initial marking level: remnants of Paleogene marine sedimentary rocks have been retained in northern Finland and the Kaliningrad region (Nikonov, 2010). Most likely these remnants are Eocene in age and coincide in time with maximum transgression in the western Arctic region (Zarkhidze and Musatov, 1989). The minimum erosion rate (0.031 and 0.022 mm/yr) is noted in the Eocene.

Table 3. Rate of removal from Fennoscandia over various time intervals

Period or Epoch	Time, Ma	Thickness, m	Rate, mm/yr
T	44.9	4460	0.099
J	63.1	1650	0.026
K	76.5	2350	0.031
Pg_1	10.5	780	0.074
Pg_2	21.3	475	0.022
Pg_3	9.9	522	0.053
N_1	18.48	175	0.009
N_2	3.51	513	0.146

The global tectonic and paleogeographic changes in Oligocene related to the Pyrenean and coeval tectonic phases were reflected in an increasing sedimentation rate in the framework of Fennoscandia (0.035 and 0.053 mm/yr). The areas flooded by the sea are reduced in western Eurasia. The maximum of the regression fell on the end of epoch, when the Polish–Pripyat Strait closed (Ronov et al., 1989). The Oligocene is a critical stage in the evolution of the West Arctic Shelf and the Arctic Ocean, with the formation of the typical morphostructure of the oceanic basin and large arches in the West Arctic Shelf.

In the Miocene–Pliocene, the territory of Fennoscandia underwent planation accompanied by the formation of a weathering hydromica mantle, which was redeposited at the end of the Neogene with concentration of a loparite placer in the Lovozero district. The increase in the erosion rate at that time was reflected in an anomalous rate in Pliocene (0.146 mm/yr, see Table 3) estimated by separate calculation for the Miocene and Pliocene. It was apparently related to the accumulation of large bodies of morainelike sediments (Patyk-Kara and Drushits, 2009) and a shortage of volcanic material supplied during the Late Cenozoic from western Spitsbergen and Novaya Zemlya (Korago et al., 2010).

The Neogene weathering mantle is extremely important for recognition of the neotectonic stage of the Fennoscandian Shield, because in neotectonic maps its onset is referred to the Oligocene (*Karta* ..., 1985). According to the yielded erosion rates and the aforementioned characteristics of the Oligocene, the onset of the neotectonic stage can be reasonably referred precisely to the Oligocene, whereas the erosion of the Neogene weathering mantle at the end of the Pliocene stage (Evzerov, 2001) can be regarded as either a substage of the neotectonic stage or as a local center for the ascension of separate blocks of the shield. In addition, the activation of recent movements at that time has been noted within a significant part of the Arctic region in northern Eurasia, e.g., in the north of the Western Siberian Plate and in the Polar Urals (Sim et al., 2008). These events might be related to the final opening of the Fram Strait between the Norwegian–Greenland and Eurasian basins (Khain and Limonov, 2004).

The results of calculating the erosion–denudation removal show that the order of magnitude of scouring of Fennoscandia over the Mesozoic and Cenozoic could have reached 11 km. The surface denudation rates averaged over various time intervals and, correspondingly, the exhumation rates vary from 0.01 to 0.15 mm/yr. These values correspond to the copious data published by Kukkal (1987). At the same time, they are almost two orders of magnitude lower than the exhumation rates of rocks from collisional belts obtained by thermobarometry and thermochronology (Burbank, 2002).

The rates of vertical movements of Fennoscandia substantially increased during the period of glacier thawing, achieved maximum values (8–12 cm/yr) by its complete disappearance 10000 years ago, and then slowed down to 2 cm/yr (6000–7000 years ago). The climatic variations over this time span could have facilitated an abrupt increase in the denudation rate up to a few millimeters per year. Hallet et al. (1996) published data on the glacial erosion rate, which varies from 1 to 30 mm/yr.

Estimation of the Tectonic Stress Level

In this section, the above theory of residual stress calculation is tested with data on the denudation of the Fennoscandian surface (Tables 2, 3), as well as on the amplitude of cover elimination since the start of the initial glaciation stage.

(1) We suggest that rocks that currently occur near the surface (down to ~100 m) were at a depth of 300 m shortly before the onset of Quaternary glaciations. A decrease in the depth of rock localization by 200 m was mainly caused by the grader effect of a glacier moving from the northwest to the southeast (20000 years of active glacier growth with an estimated glacial erosion rate of 10 mm/yr (Ramsay and Hackman, 1994). The surface denudation during the active glacial thawing period in Fennoscandia (5000 yr) reached 20–30 m for the rapidly growing Khibiny, Kovdor, and Lovozero massifs; 5 mm/yr is the rate of the active denudation phase in a waterlogged medium with sharp temperature fluctuations at high latitudes (Kukkal, 1987).

In calculating the stresses, we used the formulas from (Rebetsky, 2008a, 2008b) and the following parameters: rock density $\rho = 3$ g/cm^3, coefficient $\nu = 0.25$, cohesion value $\tau_f = 5$ MPa (the initial fracturing of rocks is taken into account here), and $k_f = 0.6$. The fluid pressure is determined according to the hydrostatic law of distribution. We assume that at the initial time instant (before Quaternary glaciation), there were no residual stresses in rock, and its stressed state was completely determined by the initial GSS stage. For the parameters accepted above, the depth of the roof of the cataclastically deformed layer (H_p) is ~ 0.6 km. In this case, the vertical compressive stresses at a depth of 300 m are –9 MPa and the horizontal compressive stresses are –3 MPa.

At the glaciation stage, the vertical load increases owing to ice cover, which is 2500–3000 m thick. We suppose that, like for contemporary Antarctica, a water lens existed near the sole of ice. This led to elevated fluid pressure in the fracture–porous space of rocks (150% of hydrostatic pressure). In this case, the depth of transition into a cataclastic state is less than 10 m. Thus, during the glaciation period, at a paleodepth of 300 m, rocks reached the limit of cataclastic flow (Rebetsky, 2008a, 2008b) and additional horizontal compressive stresses appeared. For the accepted thickness of ice, this more accurate definition can add 700 m to the paleodepth of 300 m ($H = 1$ km). Based on this depth value, we find the additional horizontal compressive stresses $\Delta\sigma \sim -8$ MPa. At this time instant, the vertical stresses will be close to –31 MPa, whereas horizontal compression taking into account additional stresses will be –18 MPa, i.e., a geodynamic horizontal extension regime.

As follows from this evidence, after completion of the glaciation period and thawing of the glacier, rocks that obtained additional compressive stresses at a depth of 300 m were brought to a depth of 100 m. The elastic unloading caused by denudation and thawing of ice released only stresses corresponding to the purely elastic deformation stage (unloading develops according to an elastic law), whereas additional compressive stresses remain in the rock. These stresses should be regarded as residual. Thus, the effect of glaciation currently assumes the formation of horizontal compressive stresses of about –9 MPa at a depth of 100 m; the level of vertical compression is about –3 MPa. The values of horizontal stresses are 2–5 times smaller than stresses measured in situ for the Khibiny and Lovozero massifs, but they are close to the stresses measured in the Kovdor massif.

(2) To calculate residual stresses caused by surface denudation during the preglaciation period, we use the data presented in Table 2. It is assumed that over the last 150 Ma (Late Mesozoic–Cenozoic) the surface denudation could have been 5 km; i.e., rocks that occur now at the surface were at a depth of 5 km from the surface150 Ma ago. We believe that the stressed state of the mountain massif at that time was controlled only by the action of mass forces. The calculations also do not take into account the stressed state induced by the additional tightening weight of the rocks that formed during the glaciation period.

In the calculations, we use the same properties and strength of the medium as in the preceding section. In this case, the roof boundary of cataclastic deformation of the crystalline crust is 0.6 km and the additional horizontal compressive stresses form at a depth of 5 km during cataclastic deformation. Using the formulas obtained from (Rebetsky, 2008a, 2008b), we find that the level of these additional compressive stresses is about 38 MPa. The values of complete stresses at this period combine 150 MPa of the vertical compressive stresses and 88 MPa of the the horizontal compressive stresses.

The tectonic uplift over 150 Ma was accompanied by denudation of rocks, which acquired additional compressive stresses at a depth of 5 km and were then drawn to the surface. The elastic unloading caused by denudation releases only stresses corresponding to the purely elastic stage of deformation, whereas additional compressive stresses remain in the rock. Thus, when coming out to the surface (depth is 100 m), the vertical compressive stresses will be 3 MPa, whereas horizontal compressive stresses will be close to 39 MPa.

(3) We apply similar calculations to the Triassic–Jurassic–Cenozoic. According to the data in Table 2, surface denudation could have been about 11 km; i.e., the rocks that now occur at the surface were at a depth of 11 km from the surface 250 Ma ago. In this case, the additional horizontal compressive stresses formed during the cataclastic flow will be already about 90 MPa. Thus, when coming out to the surface (depth is 100 m), the vertical compressive stresses will be 3 MPa, whereas horizontal compressive stresses will be already about 91 MPa.

DISCUSSION

The estimates of stresses presented above were calculated from the average amplitude of crust surface denudation of Fennoscandia as a whole. It can be suggested that rocks in this large territory were exhumed nonuniformly, which may be related to local areas of faster uplifted intrusive massifs, as well as to existing fault zones. From this point of view, the stress estimates obtained for the Khibiny and the Lovozero can be regarded as the lower limit of additional horizontal compressive stresses.

As compared with the Lovozero massif, the Khibiny massif is characterized by a significant range of mountaintop heights relative to the internal valleys and a sharper transition of slopes to the adjacent territories. From this viewpoint, a somewhat higher level of residual stresses in the Lovozero massif than in the Khibiny massif is expected. This is supported by the smaller inclination of layers in the Lovozero massif, which may be treated as large amplitudes of vertical upward compression of rocks.

The faster uplifting, which has no time for complete erosion, forms the topography. In this case, the exhumed rocks of one vertical section that had close amplitudes of elastic unloading, but at different depths, may be diverse in the level of additional horizontal compressive stresses. This is caused by rocks above the level of the valley topography undergoing additional unloading due to the diminished level of lateral compression. Thus, in uplifted regions with strikingly expressed mountainous topography, the level of horizontal compressive stresses related to the residual stressed state somewhat decreases compared to uplifts with a quiet, slowly rising surface.

The results of horizontal stress calculations over the glaciation period are consistent with in situ data for the Kovdor massif, but are several times less than for the Khibiny and Lovozero massifs, which underwent much higher amplitudes of uplifting. This implies that one effect of glacial erosion is insufficient to explain the measured stresses for these massifs.

To date, only the first results have been obtained in the framework of this concept. A more in-depth look at the proposed concept of a substantial contribution from residual stresses to the anomalously high level of horizontal compressive stresses may show that over the last hundreds of millions of years, nearly all crustal mountain massifs underwent periods of uplifting accompanied by erosion of slopes (mountain orogens) and denudation of large areas of relatively gentle topography (shields and plates). The rates of surface uplifting and erosion–denudation do not strongly differ in order of magnitude from each other, estimated at 0.001–1.0 mm/yr for time intervals of few to tens of millions of years (Ollieer, 1984), and they depend on the tectonic regime and climatic conditions (Kukkal, 1987). Thus, the rate of rock exhumation from a great depth on shields and platforms can be 0.01–0.1 mm/yr. For the first hundreds of millions of years, the possible level of residual stresses is estimated at 10–100 MPa. Because platforms and shields occupy large areas of continental plates, residual GSS stresses may prove a source that recharges planetary crustal horizontal compressive stresses.

CONCLUSIONS

The considered mechanism of generation of horizontal overpressure in a rock massif shows that under conditions of the sole action of mass (gravitational) forces, three types of domains exist. These are (a) domains of purely elastic deformation closely adjoining the crust roof; (b) domains of elastic–cataclastic flow, where due to overcoming of the limit of fracture fluidity, the level of horizontal compression approaches the lithostatic pressure but remains below it; and (c) domains of elastic unloading caused by exhumation of rocks, owing to the emergence of the surface and its denudation.

The depth of the roof in the domain of elastic–cataclastic deformation depends on the type of rocks (the Poisson ratio), strength parameters (internal friction and cohesion coefficients), and fluid pressure. In the domain of cataclastic flow with vertical orientation of the maximum compression axis (standard GSS), elastic compaction of rocks takes place; irreversible elongation deformation develops in the horizontal direction along with shortening in the vertical direction. As a consequence, additional horizontal compressive stresses form. Meanwhile, the vertical stresses overall remain the most highly compressive.

According to our calculations, residual stresses acting near the surface (at a depth close to 100 m) spread from 9 to 39 and even 91 MPa, indicating that substantial ranges of horizontal compressive stresses are possible in various rock domains depending on the evolution of the deformed state and specific conditions. The calculations of stresses carried out using a tectonophysical method, which takes the amplitudes of denudation into account, yield results close to the stress values measured instrumentally in situ for the Lovozero, Khibiny, and Kovdor massifs. This makes it possible to regard the residual GSS stress formation caused by denudation a possible and methodically substantied mechanism that explains high-level horizontal compressive stresses.

Note that calculations of residual GSS stresses caused by denudation may be useful for estimating the depth at which endogenic deposits form.

This result may be used in predicting the status of walls in deep open pits and underground works when mining ore deposits. The estimation of erosion removal and forecasting of superfluous horizontal stresses are of special importance in ore geology for understanding the depth and paleogeodynamic formation conditions of orebodies hosted in igneous rocks.

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