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Volkovsky massifs



The Nizhny Tagil and of the Uralian Platinum Belt, and related deposits



RUSSIAN ACADEMY OF SCIENCES ZAVARITSKY INSTITUTE OF GEOLOGY AND GEOCHEMISTRY (IGG UB RAS)

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The guidebook illustrates two distinct types of ore-bearing massifs within the 900-km-long Uralian Platinum Belt: the Nizhny Tagil clinopyroxenite-dunite massif and related platinum bedrock and placer deposits, and the Volkovsky gabbro massif comprising copper sulfide deposit with PGE mineralization. Structure of the Tagil zone and general characteristic of magmatism, questions of age and genetic models are briefly outlined. The proposed excursion stops include inspection of different ultramafic and mafic lithologies from natural bedrock outcrops and open-pits.

The materials of the guidebook are of a broad interest for geologists, earth scientists and students.

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INTRODUCTION

One of the objectives of the 12th International Platinum Symposium (12 IPS) is to acquaint the participants with the most important types of platinum-group elements deposits. The post-symposium field trip #5 aims to visit two distinct types of ore-bearing massifs: the Nizhny Tagil clinopyroxenite-dunite massif that contains platinum-bearing chromitite within dunite, and associated platinum placer deposits, and the Volkovsky gabbro massif comprising copper sulfide deposits with platinum-group element (PGE) mineralization. Both massifs form part of the 900-km-long Uralian Platinum Belt (UPB), which is represented by a chain of fourteen mafic-ultramafic massifs in the eastern slope of the Middle and Northern Urals.

It is intended that participants will become familiar with geological position, structure and ultramafic rocks of the Nizhny Tagil massif. We will visit sites of bedrock platinum deposits with inspection of small-scale mining operations remained from the last century and the modern open-pit mine that expose ultramafic lithologies (i.e., dunite and chromitite). We intend to visit the Chauzh placer deposit, which is located within the limits of the Nizhny Tagil massif.

The Volkovsky gabbro massif is a host to the one-and-only economic copper sulfide deposit within the Platinum Belt of the Urals. It is currently open-pit operated. Participants will visit the open-pit, where primary relationships between both Pd-bearing bornite-chalcopyrite ore and titanomagnetiteapatite ore with gabbro can be clearly observed.

Discovery and development history of the Ural platinum deposits

The Urals is the Motherland of the Russian platinum industry. In 1819, a "white metal" was discovered in gold placers of the Verkh-Iset, Nev'yansk and Bilimbaevsk fields, which turned to be osmiridium. The first platinum placer was discovered in 1824 along the Orulikha river, the left tributary of the Barancha river. In the following, platinum was found along the Martian river near Mount Solov'yeva, in the Isovsky river basin and other places.

The settlement of Uraletz closely located to the Nizhny Tagil massif was founded in 1836 as the Avrorinskly field to mine placer platinum in the basins of the Chauzh and Martian rivers. In 1892, bedrock platinum manifestations were found in Krutoy Log on the northern slopes of the Mount Solov'yeva. Later, in 1909, the Gospodsky mine plant was opened, which produced 200 kg of platinum during its operation. Platinum content in chromitites in the upper deposit horizons reached 400 g/t.

At the boundary of the 19th-20th centuries, the Ural placers of Nizhny Tagil, northern Kytlym and Isovsky areas gave about 80% of the world platinum mining with an annual production of 2-3 metric tons. Starting from 1911, the output began to fall and in 1922 it accounted to 41.2% of the world platinum. The total amount mined during 100 years since the mining started is estimated ca. 260 metric tons. Taking into account unreported metal the estimate of platinum mining accounts to about 330 metric tons.



GEOLOGICAL STRUCTURE OF THE TAGIL ZONE

The Nizhny Tagil clinopyroxenite-dunite and the Volkovsky gabbro massifs are located in the southern part of the UPB at the latitude of Nizhny Tagil (150 km north-west of Yekaterinburg).

The UPB is located in the western part of the Tagil area, adjoining from the east, along the Main Uralian Fault (MUF) to the Central Uralian megazone (Fig. 1), where the Precambrian (predominantly Meso- and Neoproterozoic) crystalline basement of the Urals is exhumed. To the west it is boarded by the Western Uralian megazone, composed mainly of Paleozoic shelf and bathyal sediments, resting on the Precambrian crystalline basement. The Central and Western Uralian megazones correspond to the Paleocontinental sector in the Paleozoic history of the Urals. The Tagil zone is a syncline composed of predominantly island-arc volcanogenicsedimentary complexes. Significantly melanged ophiolites occur only at its eastern and western periphery.

Fig. 1. Structural zones of the Urals (modified after *Puchkov*, 2010) Paleocontinental sector: A – Preuralian foredeep (Permian preflysh, flysh and molasses), B – West Uralian megazone (Paleozoic shelf and deep-water passive margin sediments), C – Central Uralian megazone (predominantly Meso- and Neoproterozoic crystalline basement). Paleooceanic sector: D – Tagilo-Magnitogorsk megazonezone (Ordovician-Early Carboniferous complexes of oceanic crust and ensimatic island arc), E – East Uralian zone (zone of late Paleozoic orogeny). Abbreviation: MUF – Main Uralian Fault; UPB – Uralian Platinum-bearing Belt.

Volcanogenic-sedimentary complexes

The section of the western trough of the Tagil limb in its Mid-Ural part (Fig. 2) begins with Ordovician basalts interbedded with jasperoids, spatially associated with a complex of parallel dikes of probably oceanic nature. Two varieties of basalts (high-titanium ophiolite, and low-titanium island-arc series) are identified in the *Vyisky Suite* ($O_{2-3}vs$) of presumably Ordovician age within the MUF zone. To the East and probably higher in the section, Ordovician basalts are replaced by an early island-arc contrasting basalt-natrolite pyritiferous series (*the Kabansky suite*, O_3 - S_1kb), basalt-andesitoid volcanogenic-sedimentary complexes of *the Pavdinsky Suite* ($S_1 pv$), and higher in the section by carbonate rocks. The overlying *Imennovsky Suite* (S_{1-2} im) in the central part of the syncline is composed of basalts, andesites and their tuffs, replaced in the east by reef limestones. The section of volcanogenicsedimentary complexes in the axial part of the syncline ends with rocks of the volcanogenic-sedimentary *Turinsky Suite* (S_2 - D_1 tr), which is dominated by coarse clastic red-coloured polymictic sediments, while lavas are represented by trachyandesites, trachytes and trachybasalts.

The studies have shown (*Puchkov*, 2010 and references cited therein) that the Tagil zone volcanism developed from homogeneous tholeiitic to differentiated calc-alkaline and subalkaline shoshonite, which is in agreement with particularities of modern island arcs and is associated with a decrease in the degree of the mantle substrate melting.

Plutonic complexes

Magmatic complexes of the Mid-Ural part of the Tagil zone are presented by two main types: dunite-clinopyroxenite-gabbro massifs (the Kachkanar complex O_3 k) and overwhelmingly prevailing gabbros (the Tagil-Kytlym complex S_1 tk) in the west, and gabbro-granitoid (the Arbatsky S_1 a and Verkhny-Tagil S_2 vt complexes) and gabbro-syenite (the Kushvinsky complex S_2 - D_1 k) massifs that are comagmatic to the relevant volcanic rocks in the east. These massifs together with volcanics are believed to form volcano-plutonic associations of a single basaltoid magmatism cycle (*Fershatater et al.*, 1984; *Fershtater*, 2013). Granitoids transect all rocks of the Tagil zone, except the Turin Suite (S_2 - D_1 tr), implying that they are the youngest magmatic formations.

The dunite-clinopyroxenite-gabbro and gabbro-norite complexes are divided by the strata of structurally nonhomogeneous metamorphic rocks, i.e. amphibolites and amphibole-pyroxene hornfels (Fig. 2). Both have tectonic contacts with a metamorphic "frame". Their position relative to each other is the same all over the Platinum Belt: dunite-containing massifs are located in the west, and gabbro-norite ones in the east. Due to the synform nature of the Tagil area, a different depth for these complexes can be assumed.

Granitoids and syenites in the massifs of the Platinum Belt predominantly occupy an extreme position in the west, tending to the axial part of the Tagil syncline. Here they are mainly in contact with a complex of fine-grained metamorphic rocks, i.e. apovolcanogenic hornfels formed in the endocontact of gabbro-norite intrusions (*Malakhova*, 1968). In the east granitoids transect Silurian volcanogenic strata and are overlain by the Turin suite sediments.

As demonstrated on an example of the largest Tagil massif, gabbro-plagiogranite, gabbro-dioritegranite and gabbro-syenite formations successively substitute each other from south to north and from west to east. Based on the geological observations and paleontological age of comagmatic volcanic rocks, the spatial formational series is heterochronous. This is reflected by the state geological map of the Russian Federation, scale 1:200 000 (Fig. 2).

Age

The main age range (460-400 Ma), defined by different isotopic methods for all magmatic complexes of the Tagil zone without exception (*Efimov* et al., 2005; *Bosch* et al., 2006, *Popov & Belyatsky*, 2006, among others) generally coincides with the age of host calc-alkaline volcanogenic complexes.



Volcanogenic-sedimentary complexes

West Uralian zone

D ₁₋₂ c	Carbonate strata: limestone, dolomite, carbonaceous shale
D ₁ tk	Takatinsky Suite: Sandstone, gravelstone, sericite schist
S ₂ jaz	Jazvinsky Suite: marbled limestone
R ₃ kr	Kyrminsky Suite: quartzitic sandstone, quartz and quartz-feldspar sandstone, carbon-bearing-bearing shale
	Central Uralian zone
O ₁₋₂ pl	Palnichninsky Suite: sericite-quartzite slate, carbon-quartz slate, quartzite
$R_3 kp_2$	Kolpakovsky Suite, the upper subsuite: metabasalt, chlorite-sericite-quartz schist
$R_3 kp_1$	Kolpakovsky Suite, the lower subsuite: chlorite-sericite-quartz schist, quartzity schist
	Tagilo-Magnitogorskaya zone
S_2-D_1 tr	Turinsky Suite: trachybasalt, trachyandesite, trachyte and tuff
S ₁₋₂ im	Imennovsky Suite: andesite-basalt, basalt, andesite, hyaloclastite, tuff, tuff conglomerate intercalated with siliceous siltstone and limestone
S ₁ pv	Pavdinsky Suite: tuffstone, tuff, dacite and andesite-dacite tuff, andesite-basalt tuff, andesite, carbon-siliceous shale, limestone.
O_3-S_1kb	Kabansk Suite: dacite, rhyolite, tuff, lava breccia, dolerite, basalt, green shale
O ₂₋₃ vs	Vyisky Suite: metabasalt, green schist, carbon-quartz slate
	Intrusive rocks
	Intrusive rocks Kushvinsky gabbro-syenite complex
$\epsilon_2 S_2 - D_1 k$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry
$\frac{\epsilon_2 S_2 - D_1 k}{p \gamma_3 S_2 v t}$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite
$\frac{\epsilon_2 S_2 - D_1 k}{p \gamma_3 S_2 v t}$ $q \delta_2 S_2 v t$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite
$\frac{\epsilon_2 S_2 - D_1 k}{p\gamma_3 S_2 vt}$ $q\delta_2 S_2 vt$ $qvn_1 S_2 vt$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite quartz-bearing gabbronorite, amphibole gabbro
$\epsilon_2 S_2 - D_1 k$ $p\gamma_3 S_2 vt$ $q\delta_2 S_2 vt$ $qvn_1 S_2 vt$ $p\gamma_3 S_1 a$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite quartz-bearing gabbronorite, amphibole gabbro Arbat gabbro-plagiogranite complex plagiogranite
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$\epsilon_2 S_2 - D_1 k$ $p\gamma_3 S_2 vt$ $q\delta_2 S_2 vt$ $qvn_1 S_2 vt$ $p\gamma_3 S_1 a$ $hv S_1 tk$ $v_2 O_3 k$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite quartz-bearing gabbronorite, amphibole gabbro Arbat gabbro-plagiogranite complex plagiogranite Tagil-Kytlymsky gabbro complex gabbronorite, amphibole gabbro Kachkanarsky gabbro-clinopyroxenite complex melanocratic gabbro
$\epsilon_2 S_2 - D_1 k$ $p\gamma_3 S_2 vt$ $q\delta_2 S_2 vt$ $qvn_1 S_2 vt$ $p\gamma_3 S_1 a$ $hv S_1 tk$ $v_2 O_3 k$ $v_1 O_3 k$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite quartz-bearing gabbronorite, amphibole gabbro Arbat gabbro-plagiogranite complex plagiogranite Tagil-Kytlymsky gabbro complex gabbronorite, amphibole gabbro Kachkanarsky gabbro-clinopyroxenite complex melanocratic gabbro clinopyroxenite, olivine clinopyroxenite, dunite
$\epsilon_2 S_2 - D_1 k$ $p\gamma_3 S_2 vt$ $q\delta_2 S_2 vt$ $qvn_1 S_2 vt$ $p\gamma_3 S_1 a$ $hv S_1 tk$ $v_2 O_3 k$ $v_1 O_3 k$	Intrusive rocks Kushvinsky gabbro-syenite complex Syenite, syenite porphyry Verkh-Tagilsky gabbro-diorite complex plagiogranite quartz diorite, diorite, gabbro-diorite quartz-bearing gabbronorite, amphibole gabbro Arbat gabbro-plagiogranite complex plagiogranite Tagil-Kytlymsky gabbro complex gabbronorite, amphibole gabbro Kachkanarsky gabbro-clinopyroxenite complex melanocratic gabbro clinopyroxenite, olivine clinopyroxenite, dunite

Fig. 2. Geological scheme of the Nizhny Tagil massif area (compiled using sheet O-40-XXIV of the state geological map of the Russian Federation, scale 1 : 200 000)

However, the Vendian (about 560 Ma) dating obtained by U-Pb and Sm-Nd methods for olivineanorthite gabbro and dunite (*Efimov* et al., 2010; *Krasnobaev* et al, 2007), as well as the Archean age for zircons from dunites (*Malitch et al.*, 2009; *Fershtater* et al., 2009) favor for the multistage formation of the Uralian Platinum Belt.

The determined by geological data age differences between granitoids of gabbro- plagiogranite (S_1), gabbro-diorite-granite (S_2) and gabbro-syenite ($S_2 - D_1$) complexes have not been confirmed by U-Pb dating of zircon. The total age range of zircons from granitoids, considering the error, is 436-400 Ma (*Efimov* et al., 2005; *Fershtater*, 2013). It overlaps with the age interval determined for zircons of gabbro-norites and surrounding near-contact hornfels. This fact is not in accordance with geological observations and can have several contradictory interpretations, which will not be addressed here. Nevertheless, geochronology issues of plutonic complexes, composing the Uralian Platinum Belt, require special attention in their further systematic study.

Models

The lack of preserved magmatic structures and traces of primary layering, as well as the absence of volcanic olivine gabbro analogues leads many researchers to question the magmatic nature of the dunite-clinopyroxenite-gabbro complex and consider it as a result of the mantle blocks intrusion in the lower crust due to the continental crust destruction during the rifting. The interpretation of dunite-clinopyroxenite-gabbro complex as a platform formation is favored by a complete analogy in composition between dunites in UPB and dunites of the Kondyor massif in the southeast of the Siberian platform. Most consistently these ideas are presented in the papers by Efimov (1984, etc.), who has also shown that the near-contact hornfels around these units have a tectonic relationship with volcanogenic environment. This implies that dunites have distinct origin and were emplaced to the upper crust within dunite-clinopyroxenite-gabbro blocks (*Efimov*, 1993).

At the same time, *Savelyeva* et al. (1999) have shown that widely spread plastic deformations do not completely eliminate magmatic structures, textures and composition features of the UPB rocks. According to *Savelyeva* et al. (1999), the UPB is a single plutonic complex formed during the continental crust destruction between the Early and Late Silurian. The lower part of the plutonic assotiation is composed of zonal ultramafic rocks, replaced in the upper part of the section by melanocratic olivine gabbro with dunitic "protrusions". The plutonic section is completed by trachytoid gabbro and gabbro-norites.

Most supporters of the Platinum Belt magmatic origin, however, do not tend to unite dunite-clinopyroxenite-gabbro and mainly gabbro (gabbro-norite) complexes. It is assumed that the source of parental picritic melts in the former case could only be a calcium-enriched and crustally-contaminated mantle material, whereas gabbro-norites, which are close in their composition to calc-alkaline highalumina basalts, are interpreted to be derived from the depleted mantle (*Fershtater et al.*, 1999).

Considering that the Uralian Platinum Belt massifs are located amid geological complexes formed mainly in the suprasubduction geodynamic setting, and the formation age range of the constituent rocks coincides with the time of the island-arc calc-alkaline magmatism of the Tagil zone, many researchers attribute these massifs to the island arc formations (*Ivanov & Shmelev*, 2006; *Puchkov*, 2010; *Fershtater*, 2013), not excluding their multistage origin and substantial prolonged evolution over time.

NIZHNY TAGIL CLINOPYROXENITE-DUNITE MASSIF AND RELATED PLATINUM DEPOSITS

General characteristics of the massif

The Nizhny Tagil massif is located in the Middle Urals, 50 km southwest of the city Nizhny Tagil, and covers an area of about 47 km² (Fig. 3). The Nizhny Tagil massif is almond shaped, shear bounded, and enclosed by Ordovician-Silurian metasediments to the west and predominantly mafic igneous rocks to the east. To the west the Nizhny Tagil massif is boarded by diabases, basaltic (andesitebasaltic) porphyrites and their tuffs with lenses and interbeds of siliceous-carbonaceous shales of the Ordovician-Silurian age. In close proximity to the massif, near to the Main Ural Fault Zone (MUF), volcanic rocks have been transformed into green shales. They have submeridional strike, where tight folds with gentle plunge of the crest lines in the southern (south-east) direction are observed. This indicates the overthrust nature of the zone structure. To the west of the MUF, continental-slope and shelf carbonate-terrigenous-shale sediments of the Lower-Middle Paleozoic age are observed. The massif contact with volcanogenic strata is tectonic. To the east it is separated from the overlying, significantly larger and predominantly gabbroic Tagil-Baranchinsky massif by a narrow band of migmatized apodiabasic amphibolites and pyroxene-amphibole hornfels (so called kytlymites) with interbeds and lenses of metasedimentary rocks of uncertain age. Kytlymite banding (mineral flatness) conforms to the massif contours, while its linearity has a steep east dipping (Fig. 3). Further to the east, these rocks are gradually replaced by gneissoid-banded or schistous amphibolites with green hornblende and epidote formed under lower-temperature metamorphism. According to their geochemical data, at the west surrounded rocks are characterized by the features of island-arc formations with a distinct HREE deficiency in the REE composition, whereas framing at the east, regardless of the metamorphic transformation degree, exhibit the features (e.g. low titanium content, enrichment by a series of mobile elements, etc.) typical of ophiolite basaltoids of suprasubduction type (Shmelev, 2007; Shmelev & Sedler, 1997). According to their composition and geological position, metabasites of the eastern frame are comparable with rocks of parallel diabase dikes of the ophiolite association (Fershtater et al., 1999).

The Nizhny Tagil massif is elongated in the submeridional direction with a length of about 14 km, and a width of outcrops up to 6 km (Fig. 3). The central core of the massif is composed of dunite of about 28 km². By size it is the largest of the ten known dunite bodies of the Uralian Platinum Belt (Fig. 3). Dunite is surrounded by a continuous rim of olivine clinopyroxenite with variable thickness from 100 m to 2000 m in the south-western part of the massif. The external contact of pyroxenite with host rocks is smooth, and the internal one with dunite is characterized by a winding surface. Besides, clinopyroxenite form several lenticular bodies in dunite with a length up to 0.5-1.5 km. In the north-eastern part of the massif (the Chauzh and Zotikha rivers) within the clinopyroxenite rim, there are small sites composed of porphyry melanocratic pseudoleucitic tylaite and dunite (*Fershtater & Pushkarev*, 1992). According to geological mapping, the Nizhny Tagil massif has an east dip with a steep bedding of the eastern contact and a flatter one of the western contact. According to the geophysical data interpretation, the massif structure can be traced to a depth of 10-15 km, where it gradually flattens and tapers out (*Sokolov*, 1989).

The age of dunite-clinopyroxenite-gabbro massifs of the Uralian Platinum Belt is a complex geological and isotopic-geochemical problem, which has been actively investigated and debated in recent years. The massifs have "cold" tectonic relationships with the host volcanogenic-sedimentary rocks,



Fig. 3. Structural geological map of the Nizhny Tagil Uralian-type massif (after Shmelev & Filippova, 2010). Numbers in circles refer to the excursion field stops

while high-temperature dynamothermal hornfel halos were brought to the level of erosion together with gabbro-ultramafic bodies and are alien to the geological environment (Efimov, 1999). The age data obtained for the rocks and minerals by different methods, cover the range from the Archaean through the Carboniferous period (*Bea et al.*, 2001; *Pushkarev et al.*, 2003; *Bosch et al.*, 2006; *Efimov et al.*, 2005; 2008; *Popov & Belyatsky*, 2006; *Malitch et al.*, 2009; *Fershtater et al.*, 2009; *Petrov et al.*, 2010; etc.). Based on the regular repetition of the radiological data obtained by different methods, as well as on some geological concepts, we can assume that the main events leading to the formation of gabbro-ultramafic UPB massifs took place during the interval of 570-400 Ma. It reflects in general a very long and complex formation history of the massifs and the entire UPB zone. An interesting fact is continuous discovery of the Archean-age zircons in dunites, and the former are considered either as relic-xenogenic (*Bea et al.*, 2001; *Fershtater et al.*, 2009) or as syngenetic to ultramafic minerals, reflecting the age of the dunitic mantle (*Knauf*, 2009; *Malitch et al.*, 2009; *Krasnobaev & Anfilogov*, 2014, among others).

Internal Structure

The zonal structure, caused by the existence of a dunite core and a clinopyroxenite rim, is the main but not the only element of the zonal structure of the massif. Thus, the systematic data on variations in olivine grain size and composition along the latitudinal profile across the Solov'yev mountain, demonstrate that marginal dunite adjacent to clinopyroxenite have smaller grain size and higher Fe content (Fig. 4). The main expansion of chromite schlieren is confined to the internal parts of the dunite bodies, to the zone of high-Mg recrystallized dunite and the areas of pegmatoid and miarolitic varieties (*Ivanov*, 1997). The pegmatoid variety is characterized by giant olivine crystals, up to several centimeters in size. Miarolitic dunite contains cavities, on the walls of which well-faceted crystals of olivine (pseudomorphically substituted by chrysotile), chromite, pyroxene, phlogopite, chlorite, calcium garnets and other minerals may occur. Both ultramafic pegmatites and miarolitic dunites must have formed with an active involvement of the fluid that provided exchange reactions and recrystallization phenomena at substantially lower temperatures as compared with fine-grained dunite of the outer zones (*Ivanov*, 1997). Subsequently it was shown that the coarse-grained dunite from the central zones



Fig. 4. Variations of the average grain size and the iron content of olivine in dunite at latitudinal section through the Nizhny Tagil massif (after Chashchukhin et al., 2002)

have lower temperatures of olivine-chromite equilibrium (about 900°C) in comparison with the finegrained marginal differences (1100°C), suggesting their later formation (*Chashchukhin et al.*, 2002). Most chromite schliren enriched in platinum-group minerals (PGMs) are confined to the same central zones or to the transition zone between coarse-grained dunite with magnesian olivine in the central parts and fine-grained dunite with Fe-rich olivine at the periphery.

Systematic studies on petrologic structure of the massif (*Savelyev* et al., 2001; *Shmelev & Filippova*, 2010) have shown that banding and mineral flatness bedding in ultramafic rocks could help reconstruct the contours of the large-scale funnel shaped subconcentric structure (Fig. 3).

In the dunite core, the banding and concordant mineral flatness form a subconcentric synform megastructure of the closed (?) type, stretched in the northeastern direction (Fig. 3). In the western part of the megastructure, these elements have an east-southeast dip at angles of 55-80°. In the eastern part, a steep (70-80°) southeast dip prevails, which changes to the east and northeast closer to the massif periphery. In the southern part the banding unfolds smoothly and acquires a sublatitudinal strike with a predominantly steep dip in the northern direction. The mineral linearity is oriented primarily on the flatness dip.

In the clinopyroxenite rim, the banding (mineral flatness) is generally concordant with the orientation of planar structures in dunite. In the northern part, these elements are parallel to the strike of the contact with dunite and dip steeply under the massif (Fig. 3). In the pyroxenite "bays" (north of Solovyova Mountain) their orientation is also subordinate to the contact configuration. In the southeastern part of the massif (Uraletz village), a fan-like change of the dip angles (from 10° to 90°) is observed. This change is due to the formation of synform-antiform folds of the northwest strike, which were previously identified as independent open folds and flexures (*Savelyev et al.*, 2001). They are definitely associated with the observed here complicated bay-like morphology of the contact between dunite and clinopyroxenite. In the southern part of the complex, close to the roof (Mount Malaya Shul'pikha), a stable subhorizontal or shallow (15-35°) occurrence of the banding with a dip in the eastern direction predominates (Fig. 3).

In general, the subconcentric structural pattern shows its conformity to the petrographic zoning of the massif (Fig. 3), which is a consequence of their joint formation. This structural pattern is complicated by the existence of isolated clinopyroxenite bodies that have an intersecting orientation with respect to the planar structure of the host dunite and look like peculiar apophyses of the major mass of the clinopyroxenite rim.

Microstructural heterogeneity

Ultramafic rock-forming minerals exhibit differences in their morphology, relationship character and grain size, which allow distinguishing protogranular, porphyroclastic and mosaic-granoblastic microstructure types in dunite (*Shmelev & Filippova*, 2010). The first type is interpreted as the initial one, belonging to the class of protomagmatic adcumulative structures. The other two types are more recent, formed during high-temperature plastic deformation and sintectonic recrystallization. Similar to dunite, there are microstructures of protogranular (adcumulative), porphyroclastic and mosaic-like types in clinopyroxenite. The analysis of spatial position in the massif of ultramafic rocks with different microstructure types allowed reconstructing the petrostructural pattern with elements of zonally symmetric structure.

The rocks with protogranular microstructures in the dunite core compose a significant in size central part (more than 50% of the area). Furthermore, varieties with idiomorphic grain structures (similar to the primary ones) occupy an arbitrary structural position, existing in both central and marginal portions of the massif, near to the contact with clinopyroxenite. More to the west, protogranular dunite is replaced by porphyroclastic variety and then by mosaic dunite, which are located in a fairly wide (400-700 m) near-contact band. To the east of the central region, a similar symmetric band of mosaic and transition type of dunite can be traced. Within the region of protogranular dunite, small-size sites of rocks with porphyroclastic microstructures are observed.

In the area of clinopyroxenite rim, variations of protogranular type (I) are located in the "shielded" (remote from the external contact) sites, making a single unit with dunite of similar structure. Toward the periphery of the massif, they are replaced by porphyroclastic (II), and then by mosaic (III) varieties.

Magmatic and tectonic stages can be identified in the history of the massif formation (*Shmelev & Filippova*, 2010). *The magmatic stage* of the massif formation was determined by dynamic differentiation and stratification of the rising magnesian melt (suspension) with the zonal structure formation. During submagmatic flow and adcumulative crystallization, protogranular microstructures of ultramafic rocks with distinct linear ([100]-type) orientation of olivine, mineral flatness and subconcentric megastructure contours take place. Formation of lenticular bodies in dunite and clinopyroxenite from the apical parts of the massif occurred in a static environment. *The tectonic stage* of formation. During the high-temperature plastic flow, the formation of porphyroclastic and mosaic microstructures of ultramafic rocks, dynamometamorphic zoning and mineral flatness and linearity, concordant with earlier elements, occurred. The massif intrusion was followed by the marginal dynamometamorphism of the host rocks with the advent of hornfels and kytlymites. In the subsequent tectonic movements of the massif during Urals collision, the host rocks underwent low-temperature deformations and metamorphism, while ultramafic rocks underwent local antigoritization and foliation.

Thus, the Nizhny Tagil clinopyroxenite-dunite massif, characterized by the zonal structure, subconcentric planar structure with distinct signs of the rocks flow and dynamometamorphic zoning, should be referred to a special formation class of igneous diapirs that are gradually transformed into solid-state diapirs during crystallization and cooling.

Rock characteristics

Dunite is represented by massive medium- grained and large-grained serpentinized rocks with scattered impregnation of idiomorphic chromite, easily visible on a yellowish-brown weathering surface. On average, the degree of cellular serpentinization of dunite on the surface reaches at least 50-60%. The amount of serpentine decreases with depth, and from a depth of about 400 meters in deep drill holes unserpentinized dunite predominates (Volchenko et al., 2007). The olivine composition corresponds to forsterite (Fo_{90.92}), average CaO content of 0.2 wt.% and NiO = 0.18-0.20 wt.% (Ivanov, 1997, among others). Accessory chromite comprises about 40 wt.% Cr₂O₃, less than 10 wt.% Al₂O₃ and 1% wt.% TiO, (Ivanov, 1997; Shmelev, 2007, etc.). Ivanov & Rudashevsky (1987) have shown that olivine and chromite compositions depend on the structural type of dunite, which in turn, reflect the recrystallization degree of primary dunite. From fine-grained to coarse-grained and pegmatoid varieties of dunite, the magnesian content of olivine and chromspinel, and the chromium content in chromspinel are increasing, i.e. olivine and chromite in dunite pegmatites and miarolite dunite are close in composition to the minerals of platinum-bearing chromite schlieren. As noted above, chromite segregations in space are confined to the areas of coarse-grain dunite and dunite pegmatites. Chromite segregations, schlieren and veins, sometimes forming ore zones, are associated with platinum mineralization distinguished as the Nizhny-Tagil industrial genetic type of late-magmatic platinum deposits (Ore deposits of the USSR, 1978).

Clinopyroxenite is represented by dark gray, dark greenish-gray fine-grained and medium-grained rocks. Usually they consist of 80-95% clinopyroxene (moderately aluminous diopside, 2-4 wt.% Al_2O_3 , with varying Fe content f = 0.1-0.2), 5-15% olivine with similarly varying Fe content, interstitial pargasite, and fine titanomagnetite impregnation. In comparison with olivine, clinopyroxene usually forms larger and idiomorphic grains. Monomineral clinopyroxenite and wehrlite are rare and occur closer to the dunite core. Magnetite varieties are confined to the marginal zone of the massif. Structural

elements of the rocks are represented by thin banding, formed by an alternation of olivine (less frequently magnetite) enriched and depleted layers. Clinopyroxenite is little affected by the secondary transformations (antigoritization, etc.). The chemical composition of prevailing olivine clinopyroxenite corresponds to the olivine-clinopyroxene cotectics at a pressure of 15-20 kbar (*Fershtater & Pushkarev*, 1987).

Pseudoleucitic tylaite was first described in the Nizhny Tagil massif by *Fershtater & Pushkarev* (1992), and later on discovered and described as part of the Kytlym massif in the Northern Urals (*Pushkarev*, 2000; *Krause et al.*, 2007). Tylaite is a melanocratic porphyry olivine- clinopyroxene rock, first described in the Urals on Mount Tylaysky Stone by *Duparc & Tihonowich* (1920). Porphyry phenocrysts in amount of 30-40% are presented by idiomorphic zonal clinopyroxene, often with an hourglass structure (Fig. 5). Clinopyroxene phenocrysts contain olivine, phlogopite F-bearing phlogopite and magnetite inclusions, located concordantly to the growth structures of the host mineral. The rock groundmass is composed of clinopyroxene, olivine (10-15%), high-Ti phlogopite, titanomagnetite, and green spinel. Interstices between the colored minerals (10-20%) are composed of cryptic andesine aggregate, K-feldspar and orthoclase-nepheline intergrowths interpreted as pseudoleucite. Tylaite is close in chemical composition to subalkalic picrite basalt with high CaO/Al₂O₃ ratio that exceeds 1. Tylaite is found only in association with dunite and clinopyroxenite and is lacking in essentially gabbro massifs of the UPB. According to calculations, the primary melt, parent for the Ural-Alaskan type massifs, approach a composition close to that of tylaite (*Fershtater & Pushkarev*, 1987; 1992; *Pushkarev*, 2000).

The geochemical features of ultramafic rocks and gabbro of the Nizhny Tagil massif (Fig. 6) differ significantly from similar (in mineral composition) rocks of ophiolite complexes and platform layered intrusions. However, they have common geochemical characteristics with zoned Ural-Alaskan and Aldan-type complexes known in other regions (Aldan Shield, Koryakiya, Southeast Alaska, etc.). Thus, dunite is characterized by concave (U-shaped) REE distribution trends at 0.1 of chondrite-normalized values in the middle of the spectrum. Clinopyroxenite show a slight increase in chondritenormalized values from La to Nd, peaking at Nd, Sm, and Eu and then gradually decreasing towards to Lu. In general, clinopyroxenite is characterized by high La/Yb ratio, equal to 2-3. Pseudoleucite tylaite is characterized by an even higher REE fractionation ratio, exceeding 3. In general, these features suggest the leading role of fractional crystallization in the formation of ultramafic and mafic rocks of the Nizhny Tagil and other UPB massifs (*Fershtater et al.*, 1999; *Shmelev*, 2007).



Fig. 5. Pseudoleucitic tylaite with porphyry structure. Centrally located are zonal porphyry phenocrysts of clinopyroxene with inclusions of olivine, phlogopite and ore minerals. The bulk contains olivine and clinopyroxene grains and cryptic aggregates of andesine, orthoclase and orthoclase-nepheline intergrowths. Field of view 5*3 mm, cross polarized light



Fig. 6. Distribution of rare earth elements in (A) rocks of the Nizhny Tagil massif and (B) surrounding formations (after Shmelev, 2005). A) Du, dunite; Cpx, clinopyroxenite; dotted line, magnetite bearing Cpx; 1555, metagabbro from the western contact; 1575a, pseudoleucitic tylaite. B) Kytlymite from the Eastern environment (Kytl) and island-arc volcanic rocks (Arc-v) from the Western environment. N-MORB – average composition of oceanic basalt. Dotted line shows the composition of ophiolite gabbro of the Horasyursky massif (after Shmelev, 2005)

Chromium-platinum mineralization of the Nizhny Tagil massif

The Nizhny Tagil massif is the most productive target for platinum mineralization among other clinopyroxenite dunite massifs of the Uraianl Platinum Belt. According to *Ivanov* (1997), about 1600 chromitite ore pockets, veins and schlieren within the Nizhny Tagil massif were known by 1950. Most of these bodies do not exceed 50 cm in length and a few centimeters in thickness. A considerable part of chromitites is enriched in platinum-group elements (PGE), although they do not have any economic value due to their small size. There are certain regularities in the allocation of chromite segregations within the dunite body area. For example, *Vysotsky* (1913), and *Zavaritsky* (1928) have noted that most of the chromium-platinum occurrences are localized along the western and south-western slope and piedmont of the Solov'yev mountain (Fig. 3). This area also includes all the largest platinum bedrock deposits, i.e., Gosshahta, Krutoy Log, Alexandrovsky Log, etc. Geologically, this is an area of transition between the marginal small-grained and central coarse-grained dunites. It is here that miarolite dunites and dunite pegmatites are often observed. Their formation is associated with the presence of a fluid that reduces the temperature of mineral crystallization.

Chromite bodies are characterized by a large variety of shapes and textural types. There are several classifications based on both structural-morphological and genetic features of the ore bodies (*Zoloev et al., 2001, etc.*). However, it seems the most appropriate to identify syngenetic and epigenetic ores based on their relationship with the host dunite (*Zavaritsky,* 1928). Syngenetic chromite accumulations are characterized by gradual transitions to the surrounding rock and are close to dunite in terms of their timing and formation conditions. Chromitite and dunite are involved in plastic deformations. The temperature of olivine-chromite equilibrium of dunite and syngenetic chromitite of the Nizhny Tagil massif is similar, i.e. 1100-1200°C (*Chashchukhin et al.,* 2002), which corresponds to the conditions of ultramafic subsolidus equilibrium.

Epigenetic ores usually form vein and lenticular bodies with sharp contacts with the host dunite (Fig. 7). The central zones of the ore bodies are often characterized by high porosity and increasing grain size of the chromite. Typical for this type of ores are breccia textures, in which rounded or sharp-edged fragments of variously serpentinized dunite are cemented by chromite-silicate aggregate. The contact between epigenetic chromitite and host dunite has a narrow (5-10 mm, rarely more) pale green serpentine rim containing no relics of olivine and being depleted in chromite. The rim thickness does not depend on the ore body sizes. Olivine in the matrix of epigenetic ores is usually absent or occurs as rare inclusions in chromite. Cement is presented as an aggregate of hydrous minerals (e.g., serpentine, chlorite, hydrogarnet,



Fig. 7. Longitudinal sections and cross-sections of the Gosshahta platinum deposit. Projection on a vertical plane along the ore body dip (azimuth 132 40). Oriented cross-sections for 10 different Gosshahta sections (modified after Volchenko et al., 2007) with the use of primary geological materials of "Uralzoloto" trust, 1933)

carbonate, etc.). The lack of olivine, in our opinion, is due to the saturation of the residual material with an aqueous fluid with a temperature below 500°C, which creates conditions either for its full serpentinization or for direct crystallization of low-temperature hydroxyl-bearing silicates in the equilibrium with an aqueous fluid. However, in the deep horizons of the "Gosshahta" deposit, weaker development of low-temperature water-bearing minerals in the ore cement was recorded (*Zavaritsky*, 1928; *Volchenko et al.*, 2007). The fluid presence in the ore-forming system is confirmed by inclusions of hydroxyl-bearing minerals in platinum-group minerals and chromspinelides (*Anikina et al.*, 1999; *Pushkarev et al.*, 2007; *Johan*, 1997; *Auge et al.*, 2005, among others).

Chromitite is divided by a polygonal net of thin mineralized cracks oriented on normal to the flatness of the schlieren. The cracks are filled by zonal colloform aggregates of the same minerals or cryptocrystalline (amorphous) ophite. These cracks usually do not go beyond the outer serpentinite rim surrounding ore bodies. It has been suggested (*Zavaritsky*, 1928; *Betekhtin*, 1935) that the contraction nature of the cracks were formed by shrinkage of the crystallizing ore material and filled with the residual material.

Epigenetic chromite does not bear any traces of high-temperature plastic deformation characteristic of the host dunite. However, the signs of chromite cataclase with the formation of brittle fractures, healed by low-temperature silicate and even platinum minerals, are frequently observed (*Volchenko et al.*, 2007). The temperature of olivine-chromite equilibrium for this type of ores is less than 750-800°C (*Chashchukhin et al.*, 2002). Observations of the morphology of epigenetic ore bodies, their structural and textural features and the nature of the relationship with dunite show that the ore formation occurs in almost totally consolidated dunite. However, chromium-platinum ores never go beyond dunite limits. Consequently, the ore-controlling structures are related to formation of the dunite and are not associated with later "cut-through" tectonic disturbances or brittle rock deformations. According to *Zavaritsky* (1928) and *Betekhtin* (1935), epigenetic ores are the main concentrator of platinum-group minerals.

All varieties of morphological types of the Nizhny Tagil massif ore bodies as shown by *Zavaritsky* (1928) can be classified as isometric, column-like and vein-like bodies. Most of the "Gosshahta" ore bodies as described by Betekhtin (1935) are represented by flattened lenses or "streams" steeply dipping to the east (at an angle of about 70°), up to two meters along dip and from half a meter to one meter along strike with a few centimeters in thickness (Fig. 7). The bodies are deposited either in groups, or as isolated lenses and irregularly shaped pockets disconnected with each other. In the footwall of the deposit two larger bodies from 10 to 20 m in length were recorded, one of which had a cross-sectional shape of an irregular star, with an area up to 0.5 square m, and the other was characterized by isometric cross-section up to 30 cm in diameter. However, whatever the morphology ore bodies have, they taper out relatively quickly and are surrounded by dunite on all sides, which is clearly seen in the sketches of *Vysotsky* (1913), *Zavaritsky* (1928), *Volchenko* (2007) and others. Closely located chromitite lenses and veins show features of independent and isolated bodies. The total thickness of steeply dipping ore zone of the "Gosshahta" deposit is about 5 meters.

A similar geological structure is characteristic of field N_{2} 4-00 Krutoy Log (Fig. 8), where originally the system of chromite ore schlieren was worked out by three pits (Vysotsky, 1913). Later on, the mines were merged into one open-pit, at the bottom of which horizontal mine galleries were tunneled.

Chromespinelide composition and ore bodies zoning. It is well known that chromespinel ores compared with accessory chromite in platinum-bearing dunite have higher magnesium and chromium content, which is often used as a proof of their earlier (early magmatic, cumulative, etc.) formation. The conducted study of chromespinelide composition of epigenetic ore bodies have revealed the presence of zoning (Table 1, Fig. 9). In its most general form, this zoning is expressed in increasing chrome, aluminum, magnesium and Fe³⁺/Fe_{total} content and decreasing iron content in chromospinel from the ore bodies edges to the center and, respectively, from the host dunite toward to the ore body. The Cr/(Cr +Al) value in chromespinelide does not vary, reflecting synchronous and



Fig. 8. Geological section and horizon plans of platinum deposit № 4-00 Krutoy Log (Vysotsky, 1913)

Table 1

Chemical composition of chromespinelide through the ore body number 111 (analyses 49-64), ore body № 111-A (analyses 23-44) and ore body № 35 (analyses 35-1 – 35-15), wt. %

Semple (Fig. 9) 49 50 51 52 33 54 55 66 57 58 59 60 61 62 63 TOQ, 0.54 0.48 0.48 0.50 0.48 0.50 0.48 0.50 0.48 0.48 0.44 0.44 0.51 0.44 0.45 0.45 <td< th=""><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th><th></th></td<>																
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Sample (Fig. 9)	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63
Al_Q, 7.7 7.31 7.17 7.69 7.29 7.29 7.55 7.28 7.45 7.44 7.40 7.44 7.11 7.19 7.47 Cr_Q, 49.48 49.27 48.64 49.27 48.64 49.27 48.20 48.25 Fe0* 33.50 32.77 33.73 30.21 29.94 29.71 30.49 30.75 32.69 33.43 33.33 33.73 <th< td=""><td>TiO₂</td><td>0.54</td><td>0.48</td><td>0.49</td><td>0.52</td><td>0.51</td><td>0.49</td><td>0.50</td><td>0.48</td><td>0.50</td><td>0.48</td><td>0.48</td><td>0.49</td><td>0.50</td><td>0.48</td><td>0.54</td></th<>	TiO ₂	0.54	0.48	0.49	0.52	0.51	0.49	0.50	0.48	0.50	0.48	0.48	0.49	0.50	0.48	0.54
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Al ₂ O ₃	7.57	7.31	7.71	7.69	7.52	7.29	7.55	7.28	7.45	7.44	7.40	7.44	7.11	7.19	7.47
Fe0* 33.50 32.77 34.52 34.52 31.97 30.21 29.94 29.71 30.49 30.75 32.69 33.42 34.30 34.30 MnO 0.84 0.57 0.58 0.60 0.51 0.48 0.50 0.50 0.57 0.58 0.60 0.58 0.53 0.50 0.59 0.59 0.58 0.68 0.68 0.60 0.58 0.58 0.58 0.58 0.58 0.58 0.58 0.58 0.59 0.50 0.56 0.59 0.50 0.58 0.57 0.58 0.58 0.58	Cr ₂ O ₃	49.48	49.27	48.46	48.85	49.81	50.79	50.74	51.02	51.26	50.56	50.82	50.61	48.96	49.02	48.82
MnO 0.58 0.59 0.60 0.61 0.57 0.53 0.50 0.51 0.48 0.50 0.57 0.58 0.60 0.58 MgO 8.48 8.27 7.85 8.28 9.49 9.94 10.01 10.01 9.94 10.028 9.93 10.028 9.93 10.028 9.91 10.028 9.93 10.028 9.93 10.028 9.921 9.972 9.72 9.77 9.77 10.028 0.58 0.56 0.56 0.56 0.56 0.58 0.56 0.51 0.56 0.54 0.55 0.55 0.51 0.51 0.54 0.54 0.54 0.54 0.54 0.54 0.54 0.54 0.54 <td>FeO*</td> <td>33.50</td> <td>32.77</td> <td>34.52</td> <td>34.56</td> <td>33.37</td> <td>31.97</td> <td>30.21</td> <td>29.94</td> <td>29.71</td> <td>30.49</td> <td>30.75</td> <td>32.69</td> <td>33.42</td> <td>34.30</td> <td>34.30</td>	FeO*	33.50	32.77	34.52	34.56	33.37	31.97	30.21	29.94	29.71	30.49	30.75	32.69	33.42	34.30	34.30
MgO 8.45 8.25 7.85 7.95 8.28 9.49 9.44 10.47 10.68 10.63 9.99 8.48 8.64 8.13 7.99 Cymma 100.12 98.67 96.61 100.18 100.02 99.27 99.7 Fe ¹ //Mg/Fe ¹⁺ 0.38 0.38 0.38 0.38 0.38 0.44 0.41 0.48 0.50 0.58 0.58 0.60 Fe ¹ //Fe ¹⁺ 0.38 0.38 0.38 0.38 0.38 0.38 0.38 0.32 0.83 0.83 <	MnO	0.58	0.59	0.60	0.61	0.57	0.53	0.50	0.51	0.48	0.50	0.50	0.57	0.58	0.60	0.58
Cymma 100.12 98.67 99.63 100.18 100.06 100.56 99.71 100.08 100.1 99.41 100.28 99.71 99.72 90.72 0.83 0.38 0.38 0.40 0.42 0.44 0.41 0.33 0.38 0.38 0.48 0.40 0.42 0.42 0.42 0.42 0.42 0.42 0.43 0.44 0.44 0.45 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44	MgO	8.45	8.25	7.85	7.95	8.28	9.49	9.94	10.47	10.68	10.63	9.99	8.48	8.64	8.13	7.99
$ \begin{array}{c} \label{eq:relation} \hline Fe^{+\prime}/(Mg+Fe^{+\prime}) & 0.58 & 0.58 & 0.66 & 0.60 & 0.58 & 0.53 & 0.50 & 0.48 & 0.47 & 0.48 & 0.50 & 0.58 & 0.58 & 0.59 & 0.60 \\ \hline Fe^{+\prime}/Fe & 0.39 & 0.38 & 0.38 & 0.38 & 0.38 & 0.40 & 0.40 & 0.42 & 0.42 & 0.44 & 0.41 & 0.37 & 0.41 & 0.39 & 0.38 \\ \hline Cr/(Cr+Al) & 0.81 & 0.82 & 0.81 & 0.81 & 0.82 & 0.84 & 0.8$	Сумма	100.12	98.67	99.63	100.18	100.06	100.56	99.44	99.7	100.08	100.1	99.94	100.28	99.21	99.72	99.7
Fe ^h /Fe 0.39 0.38 0.38 0.38 0.40 0.40 0.42 0.44 0.41 0.37 0.41 0.39 0.38 Cr/(Cr+A) 0.81 0.82 0.83 3.7 7.7 7.7 7.87 7.89 7.42	Fe ²⁺ /(Mg+Fe ²⁺)	0.58	0.58	0.60	0.60	0.58	0.53	0.50	0.48	0.47	0.48	0.50	0.58	0.56	0.59	0.60
Cr/(Cr+Al) 0.81 0.81 0.81 0.82 0.83 0.81	Fe ³⁺ /Fe	0.39	0.38	0.38	0.38	0.38	0.40	0.40	0.42	0.42	0.44	0.41	0.37	0.41	0.39	0.38
Somple (Fig. 9) 64 23 24 25 26 27 28 29 30 31 32 33 34 35 36 TiO, 0.48 0.6 0.62 0.55 0.56 0.51 0.56 0.57 0.57 0.54 0.54 0.54 0.57 Cr,O, 48.32 45.42 45.51 46.6 48.65 47.73 47.93 50.31 48.57 48.21 49.47 49.71 49.19 48.36 48.35 FeO* 35.12 35.48 37.5 30.72 29.75 30.1 29.64 29.25 29.11 19.91 0.62 11.09 10.9 10.62 11.09 10.9 10.62 11.09 10.62 10.5 0.5 0.5 0.47 0.49 45.0 44 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.0 44.	Cr/(Cr+Al)	0.81	0.82	0.81	0.81	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.81
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Sample (Fig. 9)	64	23	24	25	26	27	28	29	30	31	32	33	34	35	36
Al,O, 6.94 7.14 6.83 7.76 7.59 7.72 7.75 7.8 7.42 7.97 7.74 7.78 7.89 7.77 Cr,O, 48.32 45.42 45.51 46.6 48.65 47.73 47.93 50.31 48.57 48.21 49.47 49.71 49.19 48.36 48.35 FeO* 35.12 35.48 37.5 30.72 29.75 30.1 29.64 28.4 29.25 29.5 29.1 29.52 29.3 30.22 30.42 MnO 0.72 0.62 0.73 0.53 0.52 0.49 0.49 0.5 0.47 0.5 0.5 0.5 0.5 0.6 0.47 0.49 0.49 0.9 97.03 88.64 98.5 7.67 7.65 97.19 98.26 97.85 97.42 98.49 99 97.93 98.64 98.5 96.8 97.67 97.65 97.10 98.26 97.85 97.42 98.49 99 97.93 98.64 98.5 97.63 0.41 0.44 0.44 0.44 </td <td>TiO₂</td> <td>0.48</td> <td>0.6</td> <td>0.62</td> <td>0.55</td> <td>0.56</td> <td>0.61</td> <td>0.56</td> <td>0.54</td> <td>0.56</td> <td>0.59</td> <td>0.57</td> <td>0.54</td> <td>0.54</td> <td>0.61</td> <td>0.54</td>	TiO ₂	0.48	0.6	0.62	0.55	0.56	0.61	0.56	0.54	0.56	0.59	0.57	0.54	0.54	0.61	0.54
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Al ₂ O ₃	6.94	7.14	6.83	7.76	7.59	7.72	7.79	7.5	7.8	7.42	7.97	7.74	7.78	7.89	7.77
FeO* 35.12 35.48 37.5 30.72 29.75 30.1 29.64 28.4 29.25 29.11 29.52 29.3 30.22 30.42 MnO 0.72 0.62 0.73 0.53 0.52 0.49 0.49 0.5 0.47 0.5 0.5 0.5 0.5 0.47 0.49 MgO 6.55 8.26 7.26 10.64 10.6 11 10.78 11.01 11.2 11.2 10.87 10.99 10.62 10.49 0.44	Cr ₂ O ₃	48.32	45.42	45.51	46.6	48.65	47.73	47.93	50.31	48.57	48.21	49.47	49.71	49.19	48.36	48.35
MnO 0.72 0.62 0.73 0.53 0.52 0.49 0.49 0.5 0.47 0.5 0.5 0.5 0.5 0.5 0.47 0.49 MgO 6.55 8.26 7.26 10.64 10.6 11 10.78 11.01 11.2 11.2 10.87 10.99 10.62 11.09 10.9 Cymma 98.13 97.52 98.45 96.8 97.67 97.65 97.19 98.26 97.85 97.42 98.49 99 97.93 98.64 98.5 Fe ¹ /(Fe 0.36 0.44 0.42 0.47 0.45 0.46 0.44 0.43 0.46 0.45 0.46	FeO*	35.12	35.48	37.5	30.72	29.75	30.1	29.64	28.4	29.25	29.5	29.11	29.52	29.3	30.22	30.45
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	MnO	0.72	0.62	0.73	0.53	0.52	0.49	0.49	0.5	0.47	0.5	0.5	0.5	0.5	0.47	0.49
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	MgO	6.55	8.26	7.26	10.64	10.6	11	10.78	11.01	11.2	11.2	10.87	10.99	10.62	11.09	10.9
Fe ²⁺ /(Mg+Fe ²⁺) 0.66 0.58 0.63 0.46 0.47 0.45 0.46 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.44 <td>Сумма</td> <td>98.13</td> <td>97.52</td> <td>98.45</td> <td>96.8</td> <td>97.67</td> <td>97.65</td> <td>97.19</td> <td>98.26</td> <td>97.85</td> <td>97.42</td> <td>98.49</td> <td>99</td> <td>97.93</td> <td>98.64</td> <td>98.5</td>	Сумма	98.13	97.52	98.45	96.8	97.67	97.65	97.19	98.26	97.85	97.42	98.49	99	97.93	98.64	98.5
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Fe ²⁺ /(Mg+Fe ²⁺)	0.66	0.58	0.63	0.46	0.47	0.45	0.46	0.45	0.44	0.44	0.46	0.46	0.47	0.45	0.46
Cr/(Cr+Al) 0.82 0.81 0.82 0.81 0.83 0.85 0.53 0.57 0.50 0.57 0.56 0.57 0.56 0.57 0.56 0.57 0.56	Fe ³⁺ /Fe	0.36	0.44	0.42	0.47	0.44	0.47	0.45	0.43	0.46	0.47	0.43	0.44	0.43	0.46	0.46
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Cr/(Cr+Al)	0.82	0.81	0.82	0.80	0.81	0.81	0.81	0.82	0.81	0.81	0.81	0.81	0.81	0.80	0.81
TiO_0.570.560.620.620.590.620.590.60.480.500.500.510.480.48Al_Q_37.417.557.497.217.357.347.187.457.137.377.487.637.657.59Cr_Q_348.7748.3647.3343.0844.9644.6344.6345.6650.5151.0151.3650.8751.8552.03FeO*30.3432.5733.2139.5236.8337.5436.6136.7931.5730.0630.2230.2929.5028.19MnO0.50.50.570.760.670.670.660.630.480.450.410.410.40MgO10.89.249.236.257.186.987.77.839.6710.5511.3111.1111.4611.55Cymma98.3998.7898.4597.4497.5897.7897.3798.9699.449.480.440.440.44Fe ²⁺ /(Mg+Fe ²⁺)0.460.420.430.420.410.410.430.430.410.420.450.440.44Fe ²⁺ /(Fe0.460.420.430.420.410.410.430.430.440.430.820.820.820.820.820.82Cr/(Cr+Al)0.820.810.800.800.810.800.830.820.450.44	Sample (Fig. 9)	37	38	39	40	41	42	43	44	35-1	35-2	35-3	35-4	35-5	35-6с	
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	TiO	0.57	0.56	0.62	0.62	0.59	0.62	0.59	0.6	0.48	0.50	0.50	0.51	0.48	0.48	
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Al ₂ O ₂	7.41	7.55	7.49	7.21	7.35	7.34	7.18	7.45	7.13	7.37	7.48	7.63	7.65	7.59	
FeO* 30.34 32.57 33.21 39.52 36.83 37.54 36.61 36.79 31.57 30.06 30.22 30.29 29.50 28.19 MnO 0.5 0.5 0.57 0.76 0.67 0.67 0.66 0.63 0.48 0.45 0.41 0.41 0.41 0.40 MgO 10.8 9.24 9.23 6.25 7.18 6.98 7.7 7.83 9.67 10.55 11.31 11.11 11.46 11.55 Cymma 98.39 98.78 98.45 97.44 97.58 97.37 98.96 99.84 99.94 101.28 100.82 10.35 100.24 Fe ²⁺ /[Mg+Fe ²⁺] 0.46 0.42 0.43 0.42 0.41 0.41 0.43 0.43 0.41 0.42 0.45 0.44 0.44 0.43 Cr/(Cr+Al) 0.82 0.81 0.80 0.80 0.81 0.80 0.83 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82	Cr ₂ O ₂	48.77	48.36	47.33	43.08	44.96	44.63	44.63	45.66	50.51	51.01	51.36	50.87	51.85	52.03	
MnO 0.5 0.5 0.76 0.67 0.67 0.66 0.63 0.48 0.41 0.41 0.41 0.41 MgO 10.8 9.24 9.23 6.25 7.18 6.98 7.7 7.83 9.67 10.55 11.31 11.11 11.46 11.55 Cymma 98.39 98.78 98.45 97.44 97.58 97.37 98.96 99.84 99.94 101.28 100.22 101.24 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.54 0.53 0.67 0.63 0.64 0.60 0.60 0.52 0.48 0.45 0.44 0.44 0.43 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.42 0.43 0.42 0.41 0.41 0.43 0.43 0.41 0.42 0.45 0.44 0.44 0.43 0.42 0.44 0.43 0.41 0.42 0.44 0.43 0.44 0.43 0.42 0.45 0.44 0.43 0.42 0.45 0.44	FeO*	30.34	32.57	33.21	39.52	36.83	37.54	36.61	36.79	31.57	30.06	30.22	30.29	29.50	28.19	
MgO 10.8 9.24 9.23 6.25 7.18 6.98 7.7 7.83 9.67 10.55 11.31 11.11 11.46 11.55 Cymma 98.39 98.78 98.45 97.44 97.58 97.78 97.37 98.96 99.84 99.94 101.28 100.82 101.35 100.24 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.54 0.53 0.67 0.63 0.64 0.60 0.52 0.48 0.45 0.44 0.44 0.43 Fe ³⁺ /Fe 0.46 0.42 0.43 0.42 0.41 0.41 0.43 0.41 0.42 0.45 0.44 0.44 Cr/(Cr+Al) 0.82 0.81 0.80 0.80 0.81 0.80 0.83 0.82 <td>MnO</td> <td>0.5</td> <td>0.5</td> <td>0.57</td> <td>0.76</td> <td>0.67</td> <td>0.67</td> <td>0.66</td> <td>0.63</td> <td>0.48</td> <td>0.45</td> <td>0.41</td> <td>0.41</td> <td>0.41</td> <td>0.40</td> <td></td>	MnO	0.5	0.5	0.57	0.76	0.67	0.67	0.66	0.63	0.48	0.45	0.41	0.41	0.41	0.40	
Cymma 98.39 98.78 98.45 97.44 97.58 97.78 97.37 98.96 99.84 99.94 101.28 100.82 101.35 100.24 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.54 0.53 0.67 0.63 0.64 0.60 0.52 0.48 0.45 0.46 0.44 0.43 Fe ³⁺ /Fe 0.46 0.42 0.43 0.42 0.41 0.41 0.43 0.41 0.42 0.43 0.44 0.43 0.41 0.42 0.44 0.44 0.43 Cr//(Cr+Al) 0.82 0.81 0.80 0.80 0.80 0.81 0.80 0.83 0.82 <td>MgO</td> <td>10.8</td> <td>9.24</td> <td>9.23</td> <td>6.25</td> <td>7.18</td> <td>6.98</td> <td>7.7</td> <td>7.83</td> <td>9.67</td> <td>10.55</td> <td>11.31</td> <td>11.11</td> <td>11.46</td> <td>11.55</td> <td></td>	MgO	10.8	9.24	9.23	6.25	7.18	6.98	7.7	7.83	9.67	10.55	11.31	11.11	11.46	11.55	
$ \begin{array}{c c c c c c c c c c c c c c c c c c c $	Сумма	98.39	98.78	98.45	97.44	97.58	97.78	97.37	98.96	99.84	99.94	101.28	100.82	101.35	100.24	
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	Fe ²⁺ /(Mg+Fe ²⁺)	0.46	0.54	0.53	0.67	0.63	0.64	0.60	0.60	0.52	0.48	0.45	0.46	0.45	0.44	
Cr/(Cr+Al) 0.82 0.81 0.81 0.80 0.81 0.80 0.83 0.82	Fe ³⁺ /Fe	0.46	0.42	0.43	0.42	0.41	0.41	0.43	0.43	0.41	0.42	0.45	0.44	0.44	0.43	
Sample (Fig. 9) 35-6r 35-7 35-8 35-9 35-10c 35-10r 35-11c 35-11r 35-12 35-13c 35-13r 35-14 35-15 TiO2 0.47 0.47 0.48 0.49 0.48 0.49 0.50 0.46 0.44 0.48 0.48 0.49 0.50 Al_2O3 7.54 7.55 7.53 7.69 7.56 7.49 7.54 7.53 7.65 7.44 7.14 7.51 7.37 Cr_2O3 51.28 52.10 51.95 52.04 51.29 51.43 50.93 50.70 50.41 50.97 51.64 50.25 FeO* 29.69 29.44 29.38 29.85 30.52 30.52 30.35 31.43 31.19 30.47 31.63 30.81 31.67 MnO 0.44 0.41 0.42 0.42 0.45 0.47 0.47 0.45 0.45 0.47 MgO 11.09 11.47 11.46 11.0	Cr/(Cr+Al)	0.82	0.81	0.81	0.80	0.80	0.80	0.81	0.80	0.83	0.82	0.82	0.82	0.82	0.82	
TiO2 0.47 0.47 0.48 0.49 0.48 0.49 0.50 0.46 0.44 0.48 0.47 0.43 0.43 0.43 0.43 0.41 0.41 0.41 0.42 0.42 0.45 0.47 0.47 0.47 0.45 0.45 <th< td=""><td>Sample (Fig. 9)</td><td>35-6r</td><td>35-7</td><td>35-8</td><td>35-9</td><td>35-10c</td><td>35-10r</td><td>35-11c</td><td>35-11r</td><td>35-12</td><td>35-13с</td><td>35-13r</td><td>35-14</td><td>35-15</td><td></td><td></td></th<>	Sample (Fig. 9)	35-6r	35-7	35-8	35-9	35-10c	35-10r	35-11c	35-11r	35-12	35-13с	35-13r	35-14	35-15		
$ \begin{array}{ c c c c c c c c c c c c c c c c c c c$	TiO ₂	0.47	0.47	0.48	0.49	0.48	0.49	0.50	0.46	0.44	0.48	0.48	0.48	0.50		
Cr ₂ O ₃ 51.28 52.10 51.95 52.04 51.29 51.43 50.93 50.70 50.41 50.91 50.97 51.64 50.25 FeO* 29.69 29.44 29.38 29.85 30.52 30.52 30.35 31.43 31.19 30.47 31.63 30.81 31.67 MnO 0.44 0.41 0.41 0.42 0.42 0.44 0.42 0.47 0.47 0.45 0.45 0.47 MgO 11.09 11.47 11.46 11.07 11.21 10.74 10.91 10.34 10.62 10.75 9.81 10.59 10.42 Cymma 100.51 101.44 101.21 101.76 10.91 10.78 100.52 100.48 101.48 100.68 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.43 0.44 0.43 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.44 0.42 </td <td>Al₂O₃</td> <td>7.54</td> <td>7.55</td> <td>7.53</td> <td>7.69</td> <td>7.56</td> <td>7.49</td> <td>7.54</td> <td>7.53</td> <td>7.65</td> <td>7.44</td> <td>7.14</td> <td>7.51</td> <td>7.37</td> <td></td> <td></td>	Al ₂ O ₃	7.54	7.55	7.53	7.69	7.56	7.49	7.54	7.53	7.65	7.44	7.14	7.51	7.37		
FeO* 29.69 29.44 29.38 29.85 30.52 30.52 30.35 31.43 31.19 30.47 31.63 30.81 31.67 MnO 0.44 0.41 0.41 0.42 0.42 0.44 0.42 0.45 0.47 0.47 0.45 0.45 0.47 MgO 11.09 11.47 11.46 11.07 11.21 10.74 10.91 10.34 10.62 10.75 9.81 10.59 10.42 Cymma 100.51 101.44 101.21 101.56 101.48 101.11 100.65 100.91 100.78 100.52 100.48 101.48 100.68 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.46 0.48 0.47 0.49 0.48 0.47 0.52 0.48 0.49 Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.43 0.44 0.44 0.41 0.42 0.44 Cr/(Cr+Al) 0.82 0.82 <td>Cr₂O₃</td> <td>51.28</td> <td>52.10</td> <td>51.95</td> <td>52.04</td> <td>51.29</td> <td>51.43</td> <td>50.93</td> <td>50.70</td> <td>50.41</td> <td>50.91</td> <td>50.97</td> <td>51.64</td> <td>50.25</td> <td></td> <td></td>	Cr ₂ O ₃	51.28	52.10	51.95	52.04	51.29	51.43	50.93	50.70	50.41	50.91	50.97	51.64	50.25		
MnO 0.44 0.41 0.41 0.42 0.42 0.44 0.42 0.45 0.47 0.45 0.45 0.47 MgO 11.09 11.47 11.46 11.07 11.21 10.74 10.91 10.34 10.62 10.75 9.81 10.59 10.42 Cymma 100.51 101.44 101.21 101.56 101.48 101.11 100.65 100.91 100.78 100.52 100.48 101.48 100.68 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.46 0.46 0.48 0.47 0.49 0.48 0.47 0.52 100.48 101.48 100.68 Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.44 0.44 0.44 0.44 0.42 0.42 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82	FeO*	29.69	29.44	29.38	29.85	30.52	30.52	30.35	31.43	31.19	30.47	31.63	30.81	31.67		
MgO 11.09 11.47 11.46 11.07 11.21 10.74 10.91 10.34 10.62 10.75 9.81 10.59 10.42 Cymma 100.51 101.44 101.21 101.56 101.48 101.11 100.65 100.91 100.78 100.52 100.48 101.48 100.68 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.46 0.46 0.48 0.47 0.49 0.48 0.47 0.52 0.48 0.49 Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.43 0.44 0.44 0.41 0.42 0.44 Cr/(Cr+Al) 0.82 0.83	MnO	0.44	0.41	0.41	0.42	0.42	0.44	0.42	0.45	0.47	0.47	0.45	0.45	0.47		
Сумма 100.51 101.44 101.21 101.56 101.48 101.11 100.65 100.91 100.78 100.52 100.48 101.48 100.68 Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.45 0.46 0.46 0.47 0.49 0.48 0.47 0.52 0.48 0.49 Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.41 0.42 0.44 Cr/(Cr+Al) 0.82 0.83 0.82 <td< td=""><td>MgO</td><td>11.09</td><td>11.47</td><td>11.46</td><td>11.07</td><td>11.21</td><td>10.74</td><td>10.91</td><td>10.34</td><td>10.62</td><td>10.75</td><td>9.81</td><td>10.59</td><td>10.42</td><td></td><td></td></td<>	MgO	11.09	11.47	11.46	11.07	11.21	10.74	10.91	10.34	10.62	10.75	9.81	10.59	10.42		
Fe ²⁺ /(Mg+Fe ²⁺) 0.46 0.45 0.46 0.46 0.48 0.47 0.49 0.48 0.47 0.52 0.48 0.49 Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.43 0.44 0.43 0.44 0.43 0.44 0.44 0.41 0.42 0.44 Cr/(Cr+Al) 0.82 0.	Сумма	100.51	101.44	101.21	101.56	101.48	101.11	100.65	100.91	100.78	100.52	100.48	101.48	100.68		
Fe ³⁺ /Fe 0.44 0.44 0.43 0.45 0.43 0.44 0.43 0.44 0.44 0.41 0.42 0.44 Cr/(Cr+Al) 0.82	Fe ²⁺ /(Mg+Fe ²⁺)	0.46	0.45	0.45	0.46	0.46	0.48	0.47	0.49	0.48	0.47	0.52	0.48	0.49		
Cr/(Cr+Al) 0.82 0.82 0.82 0.82 0.82 0.82 0.82 0.82	Fe ³⁺ /Fe	0.44	0.44	0.44	0.43	0.45	0.43	0.44	0.43	0.44	0.44	0.41	0.42	0.44		
	Cr/(Cr+AI)	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.82	0.83	0.82	0.82		

Orebody Nº 111: 49-53, 61-64 – accessory chromospinelide of dunite; 54, 59, 60 – chromospinelide from serpentinite rim; 55-58 – chromospinelide ore. Ore body number 111-A: 23, 24, 40-44 – accessory chromospinelide of dunite; 25, 26, 39 – chromospinelide from serpentinite rim; 27-38 – chromospinelide ore. Orebody Nº 35-1: 13-15 – serpentinite rim. 2-12 – chromospinelide ore; c – center of the grain, r – rim of the grain. Position of chromespinel grains for analysis is shown in Fig. 9. Analyses performed on the X-ray microanalyzer JXA-5 at the Institute of Geology and Geochemistry, Ural Branch of RAS, Ekaterinburg (Analyst: V.G. Gmyra). FeO * – all iron is in form of FeO.

proportional change in the elements concentrations in chromespinel. This phenomenon is well known in mafic and ultramafic volcanic and intrusive rocks, when Cr and Al during chromite crystallization are proportionally replaced by Fe³⁺ and Ti, and Mg is replaced by Fe²⁺. However, in this case, the process goes clearly in the opposite direction and leads to an increase of magnesium, chromium and aluminum concentrations in chromespinels at the expense of iron and titanium with a simultaneous increase of magnesium content in coexisting olivine from early to late crystallization stages. This is opposite to the known magmatic trend, but is consistent with the changes in the composition of minerals during dunite recrystallization (*Ivanov & Rudashevsky*, 1987). We can assume that both dunite recrystallization and the formation of ore segregations are due to interrelated processes. The composition of chromespinel from serpentinite rims, surrounding epigenetic ores, indicates that these rims are elements of the overall ore zoning.



Fig. 9a. Photographs of epigenetic chromite segregations 111, 111-A and 35 in the Nizhny Tagil clinopyroxenite-dunite massif. Chromitites are surrounded by serpentinite zones (light). Dots indicate the site of chromite grains analyzed



Fig. 9b. Variation curves of chromespinelide composition variation in the profiles across the ore schlieren in samples 111, 111-A and 35. The profile of sample 111-A is created using the average values of measurements in two chromite grains from a single point: 1 – dunite, 2 – chromite, 3 – serpentine rim. The arrows show variations in chromite composition from the zonal grain center to the rim. Numbers on profiles correspond to those of Table 1 (after *Pushkarev et al.*, 2007)

PGE geochemistry

The PGE content in chromitite is significantly higher than that in dunites (Table 2, 1480.7-2682.3 and 40 mg/t, respectively). The iridium-platinum specialization of the chromitite at Nizhny Tagil is well expessed by "M-shaped" PGE pattern (Fig. 10), typical for chromitites of zoned ultramafic massifs of the Eastern Siberia (*Malitch*, 1999). Os-isotopic composition of the chromitite is characterized by minor variations ($^{187}Os/^{188}Os = 0.1217-0.1260$, Table 2). Low $^{187}Os/^{188}Os$ values are clearly indicative of of a near-to-chondritic source of the PGE.

PGM assemblages

PGMs are present in chromitites either as thin (1-20 microns) inclusions in chromespinel grains, or they form interstitial xenomorphic grains, cementing chromespinels. PGM from interstitial association tend to form large grains and their clusters that have several tens or hundreds of microns to several millimeters in size. The maximum recorded weight of such platinum ore clusters from the Ural depos-

Table 2

Rock	Sample	Os	lr	Ru	Rh	Pt	Pd	PGE total	Re	¹⁸⁷ Os/ ¹⁸⁸ Os
Chuomitito	NT-9-1	3	60	29	80	1297	11./7	1480.7	0.09	0.1217
Chromitite	NT-9-2	3	103	30	94,3	2434	18.0	2682.3	0.13	0.1260
Dunite*	NT-1	<5	<7	<2	<2	40	<20	40	n/a	n/a

Concentrations of PGE and rhenium in ultramafic rocks of the Nizhny Tagil massif, ppb

Note. *after Malitch (1999), n/a - not analyzed.



Fig. 10. Mantle-normalized PGE patterns in chromitite from the Nizhny Tagil massif. Mantle composition according to McDonough & Sun (1995)

its amounted to 427 grams. The list of PGMs identified in the Ural occurrences has several dozens of names and is constantly updated.

The majority of PGM from chromitite and associated platinum placers of the Nizhny Tagil massif are represented by Pt-Fe alloys, which dominate over osmium and iridium alloys and Ru-Os sulfides (Fig. 11). Dominant PGMs include isoferroplatinum, native platinum, tetroferroplatinum and tulameenite (*Ore deposits of the USSR*, 1978; *Cabri & Genkin*, 1991; *Lazarenkov et al.*, 1992; *Cabri et al.*, 1996; *Genkin*, 1997; *PGE mineralization ...*, 2001; *Garuti et al.*, 2002; *Malitch & Thalhammer*, 2002; *Auge et al.*, 2005; *Pushkarev et al.*, 2007; *Volchenko et al.*, 2007; *Malitch et al.*, 2011, among others).

Recent studies confirmed that the majority of PGM of chromitite from the Alexandrovsky Log are represented by Pt-Fe alloy grains with sizes ranging from 10 to 2000 microns (*Malitch et al.*, 2011; *Badanina et al.*, 2014). These Pt-Fe alloys show significant compositional variations. The dominant PGM is iron-platinum alloy having compositions close to Pt_2Fe (where Pt = at.% total of platinum-group elements, Fe = at.% total of Fe, Ni and Cu; Table 3, analyses 1, 2, 5, 6; Figs. 11, 12).

Os-Ir(-Ru) alloys, Ru-Os sulphides and Rh-Ir sulpharsenides are also present occurring as small inclusions in Pt-Fe alloy grains in subordinate amounts (Fig. 11). Os-Ir(-Ru) alloys correspond to minerals of osmium and iridium (Figs. 11a; 11c; 11e and 11g; Table 3, analyses 7 and 10). Laurite, besides ruthenium and sulfur, contains minor concentrations of osmium and iridium, varying in the range 0.38-1.24 at % and 0-0.98 at %, respectively (Fig. 11b and 11d; Table 3, analysis 4). PGE sulpharsenides are represented by hollingworthite (RhAsS) – irarsite (IrAsS) solid solution series.

The secondary low-temperature PGM paragenesis is represented by solid solution of tetraferroplatinum (PtFe) – tulameenite (PtFe_{0.5}Cu_{0.5}) series, as well as tetraferroplatinum (Table 3, analyses 3, 9, Figs. 11, 12).

No mineral with Pt_2Fe stoichiometry is known in the synthetic Pt-Fe system (*Massalski*, 1992). However, Pt-Fe alloys with a chemical composition close to Pt_2Fe are known in zoned ultramafic complexes from the Urals and Eastern Siberia, ophiolite massifs and differentiated ultramafic-mafic intrusions (*Sluzhenikin*, 2000; *Cabri et al.*, 1996; *Malitch & Thalhammer*, 2002; *Weiser et al.*, 1999; *Malitch & Kogarko*, 2011, among others).

An earlier X-ray diffraction study of Pt-Fe alloys from placers of the Guli, Kondyor, Inagli and Nizhny Tagil massifs (*Malitch & Thalhammer*; 2002; *Evstigneeva*, 2009) has led to the conclusion that all the studied Pt-Fe minerals, except for one tetraferroplatinum sample, are ferroan platinum according to classification of *Cabri & Feather* (1975). They are characterized by a disordered face-centered cubic structure (*Fm3m*) and Fe content between 20.52 and 36.16 at. %. Thus, Pt-Fe alloys at Nizhny Tagil that are compositionally similar to Pt_2Fe likely correspond to ferroan platinum.

Based on the geological position of PGE mineralization and in compliance with the identified geochemical specialization of PGMs (*Vysotsky*, 1913; *Lazarenkov et al.*, 1992; *Malitch*, 1999; *Zoloyev et al.*, 2001; *Malitch et al.*, 2011; *Johan et al.*, 1989; *Malitch & Thalhammer*, 2002; *Mochalov et al.*, 2002; *Tolstykh et al.*, 2011; *Sidorov et al.*, 2013, etc.), the bedrock source of the studied Pt-Fe alloys, enriched with refractory PGE, are chromitites and coarse dunites; while the source of Pt-Fe alloys with high palladium and rhodium concentrations are clinopyroxenites.

Osmium isotopic composition

Laurite inclusions have ¹⁸⁷Os/¹⁸⁸Os values between 0.12256 ± 0.00006 and 0.12284 ± 0.00009 , with a weighted mean of 0.12269 ± 0.00012 (2 sigma, n = 4), and ¹⁸⁷Re/¹⁸⁸Os lower than 0.00006, whereas Os-Ir alloys have ¹⁸⁷Os/¹⁸⁸Os values ranging from 0.12164 to 0.12259 with a mean of 0.12221 and a standard deviation of 0.00040, n=4 (*Badanina et al.*, 2014). The osmium isotope results identify a restricted range of similar ¹⁸⁷Os/¹⁸⁸Os values for laurite and Os-Ir alloys (Fig. 13) that are consistent within uncertainty with 'unradiogenic' ¹⁸⁷Os/¹⁸⁸Os values defined for the chromitite (0.1217-0.1260, *Malitch et al.*, 2011). The observed Os-isotope similarity between PGM and chromitite suggests that



Fig. 11. Features of the internal texture of PGM from chromitite (a-d, g, h) and placers (e) of the Nizhny Tagil massif. (a, sample NT-5-2; b, sample NT-5; c, sample 264; g, sample NT-5; d, sample 13; e, sample NL-1; g, sample 45; h, sample 31; Pt_2Fe – ferroan platinum; PtFe – tetraferroplatinum; Pt (Fe, Cu) – solid solution of tetraferroplatinum – tulaminite series; (Os, Ir) – iridian osmium; (Ir, Os) – osmian iridium; LR – laurite; CRT – chromite; images in back-scattered electrons

Table 3

Selected electron microprobe analyses of PGM from the Nizhny Tagil massif

Analysis	1	2	3	4	5	6	7	8	9	10
Sample	257	264	264	NT-5	13	NL-1	NL-1	45	45	45
Figure	11a	11в	11в	11г	11д	11e	11e	11ж	11ж	11ж
Mineral	Pt ₂ Fe	Pt ₂ Fe	Pt(Fe,Cu)	LR	Pt ₂ Fe	Pt ₂ Fe	(Ir,Os)	Pt ₂ Fe	Pt(Fe,Cu)	(Os,Ir)
					w	rt. %				
Fe	11,99	11.79	15.56	0.00	12.01	11.61	0.00	12.44	14.35	0.00
Ni	0.41	0.49	0.55	0.00	0.62	0.37	0.00	0.41	0.32	0.00
Cu	0.84	0.66	5.72	0.00	0.36	0.70	0.00	0.45	7.93	0.00
Ru	0.00	0.00	0.00	55.64	0.00	0.00	3.51	0.00	0.00	7.91
Rh	0.00	0.00	0.00	0.00	0.73	1.21	3.53	0.96	0.69	0.00
Pd	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Os	0.00	0.00	0.00	4.87	0.00	0.00	14.96	0.00	0.00	61.08
lr	1.58	1.69	0.00	2.25	4.20	6.06	74.91	3.41	3.17	30.92
Pt	85.15	85.30	78.01	0.00	82.62	80.12	2.29	81.67	72.69	0.00
S	0.00	0.00	0.00	37.15	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.97	99.93	99.84	99.91	100.54	100.07	99.20	99.34	99.15	99.91
					a	t. %				
Fe	31.59	31.23	35.82	0.00	31.45	30.61	0.00	32.64	32.82	0.00
Ni	1.03	1.23	1.20	0.00	1.54	0.93	0.00	1.02	0.70	0.00
Cu	1.95	1.54	11.57	0.00	0.82	1.62	0.00	1.03	15.93	0.00
Ru	0.00	0.00	0.00	31.52	0.00	0.00	6.32	0.00	0.00	13.97
Rh	0.00	0.00	0.00	0.00	1.04	1.73	6.25	1.37	0.86	0.00
Pd	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Os	0.00	0.00	0.00	1.47	0.00	0.00	14.32	0.00	0.00	57.32
lr	1.21	1.30	0.00	0.67	3.20	4.64	70.96	2.60	2.11	28.71
Pt	64.22	64.94	51.41	0.00	61.94	60.47	2.14	61.34	47.59	0.00
S	0.00	0.00	0.00	65.74	0.00	0.00	0.00	0.00	0.00	0.00
PGE total	65.43	65.99	51.41	33.66	66.18	66.84	100.00	65.31	50.55	100.00
Fe+Cu+Ni	34.57	34.01	48.59	0.00	33.82	33.16	0.00	34.69	49.45	0.00

Note. Pt_2Fe – ferroan platinum, Pt(Fe,Cu) – tetraferroplatinum (PtFe) – tulameenite ($PtFe_{0.5}Cu_{0.5}$) series, LR – laurite, (Ir,Os) – osmian iridium, (Os,Ir) – iridian osmium.



Fig. 12. Chemical composition of Pt-Fe alloys from chromitite of the Nizhny Tagil massif in the ternary diagram Pt+(Ir,Os,Rh,Pd) – Cu+Ni – Fe, at.% (after Badanina et al., 2014)



Fig. 13. Back-scattered electron images of internal texture of PGM from chromitite of the Nizhny Tagil massif (after *Badanina et al.*, 2014). Holes denote laser ablation MC-ICP-MS analyses. $Pt_2Fe - ferroan$ platinum, $Pt(Fe,Cu) - solid solution series tetraferroplatinum (PtFe) - tulameenite (<math>PtFe_{0.5}Cu_{0.5}$), LR - laurite, (Ir,Os) – osmian iridium, (Os,Ir) – iridian osmium, CRT – chromite, numbers in numerator and denominator correspond to ¹⁸⁷Os/¹⁸⁸Os value and the measurement error, respectively



Fig. 14. Concordia diagram for zircon from dunite of the Nizhni Tagil massif. Figures (b), (c), (d) and (e) provide a more detailed presentation about the location of dara points for zircons of the Neoarchaean, Paleoproterozoic, Mesoproterozoic and Paleopaleozoic age clusters; MSWD – mean square deviation, P – probability of conformity

the osmium isotope budget of chromitite is largely controlled by laurite and Os-rich alloy. The average T_{RD} ages of laurite and Os-rich alloys from the Nizhny Tagil massif correspond to the late Riphean (ca. 870±50 Ma).

Frequently the Uralian Platinum Belt has been considered as a chain of in situ gabbroic intrusions, where dunite was assumed to represent a cumulate of gabbroic magma. Contrary opinion considers the UPB as a complex tectonically integrated structural unit, which comprise distinct in origin formations (e.g. mantle residual material, products of its abyssal transformation, crystalline metabasites of problematic nature, material of abyssal crystallization of tholeitic magma, products of prograde metamorphism of volcanic crust and later granitoids). The new Os-isotope data support the concept of the genetic autonomy of the Uralian dunite 'cores' (*Malitch et al.*, 2009; *Badanina et al.*, 2013; *Krasnobaev & Anfilogov*, 2014) and allow considering them as the most ancient formations, tectonically integrated in the structure of the Uralian Platinum Belt.

Age characteristics of zircon from dunite of the Nizhny Tagil massif

Based on a detailed study of the internal structure and physical characteristics of zircon grains from dunite of the Nizhny Tagil massif, at least five age clusters have been identified (Fig. 14): Neoarchaean $(2781 \pm 56 \text{ Ma}, \text{Fig. 14a})$, Paleoproterozoic $(2487 \pm 33 \text{ Ma}, \text{Fig. 146} \text{ and } 1881 \pm 9 \text{ Ma}, \text{Fig. 14b})$, Mesoproterozoic (1172 \pm 9.8 Ma, Fig. 14a) and Mesopaleozoic (414.8 \pm 3.9 - 473 \pm 3.7 Ma, Fig. 14 μ), respectively. The studied zircons of the Nizhny Tagil massif are typical of dunites with wide development of plastic deformation structures, persistent olivine composition (Fo₉₂), and U-shaped type of REE distribution, generally characteristic of the mantle derivatives. The U-Pb age data of dunite zircons are characterized by discrete age clusters, which in turn are different from U-Pb age of the host volcanogenic rocks in the region of Nizhny Tagil massif. Characteristically, the ancient, but statistically representative U-Pb dating was also obtained for dunites of dunite-clinopyroxene massifs of the Platinum Belt of the Urals and East Siberia (Malitch et al., 2009; Malitch et al., 2012, unpublished data). Considering the genetic unity of platinum-bearing dunites of the Urals and the Aldan and Maimecha-Kotui provinces (Efimov, 1966), the revealed proximity of Archean U-Pb dating of zircons is not coincidental. Widespread ancient U-Pb ages of zircon in platiniferous zonal complexes, located within the mobile belt of the Urals and Siberian platform, are apparently due to the similarity of their geodynamic evolution. These results are consistent with the long-term evolution of platiniferous dunite and high prospectivity of finding noble metal mineralization in weakly eroded zoned Uralian-type massifs.

FIELD TRIP STOPS

STOP 1: Modern open-pit mine, N 57º40'05,54" E 59º38'19,45"

Modern Solov'yovogorsky dunite quarry serves as a depository for the production of refractory materials (periclase) for metallurgical plants in Nizhny Tagil. The open-pit uncovers homogeneous small- and medium-grained varieties of dunite from the south-western part of the massif (Fig. 15). Chromite schlieren, mostly developed in the eastern side of the open-pit, can be viewed and sampled from outcrops. Mineralized miarola with chromespinelide, chromediopside, sodium phlogopite, chromepargasite, native copper, rhodium pentlandite, magnetite and silicate-oxide microspherules were discovered in these rocks in the past (*Pushkarev et al.*, 2002; *Pushkarev et al.*, 2007). They reflect an active fluid involvement in the "low-temperature" pneumatolytic mineral formation, including that of chromespinelide.

STOP 2: Southeastern part of the massif (clinopyroxenite and dunite contact). N 57°39'41, 98" E 59°38'50, 62"

In the Uralets village on a small hillside a contact between clinopyroxenite and dunite, which form a linear bowing within clinopyroxenite, is exposed. Clinopyroxenite in the western contact has banding with individual thin dunite bands with a strike of 310-320° and a steep dip to the north-east. To the east they are abruptly replaced by a conventional dunite. Closer to the eastern contact with clinopyroxenite, the dunite is characterized by large greenish spots of serpentine. More to the east of the section, spotted variety of dunite is abruptly replaced by wehrlites. A detailed study of the contact rocks (600 m to the north-west) has discovered some signs of structural unity between them. First, dunite and clinopyroxenite are characterized by protogranular microstructures close to the primary ones (with high olivine protocrystals in dunite). Second, they demonstrate a concordant orientation of mineral flatness, which coincides with the contact plane orientation, and it is characterized by the identical linear orientation of olivine [100] and clinopyroxene [001] axes, resulting from the magmatic flow process.

STOP 3: The Chauzh placer deposit

The Chauzh placer was opened in 1830. Its periodic operation by individual prospectors and working teams began in 1831. Between 1911 and 1952, the placer was mainly mined by dredging technique. The total amount of metal mined within the period of the placer operation:

1) Muscular production, 1831-1879: platinum – 3019 kg, gold – 23 kg.

2) Steam dredge, 1911-1931, the upper placer: rock mass - 1,416,000 m³, platinum - 246 kg.

3) Dredge, the Kirov mine (northern placers), 1926-1952: rock mass production – 16 mln. m³, platinum – 2400 kg.

The work in the southern part of the Chauzh-Martian site (Kosogorovsky Log, Shvetsov Log, Belogorsky Log) were resumed in 2005. Since 2010, the mining of platinum placers upstream the Chauzh and Mart'yan rivers has been performed by JSC Artel of prospectors «Gold valley».

STOP 4: Central part of the massif, N 57º40'16,14" E 59º38'29, 99"

This is the main ore zone that unites the Krutoy Log (Fig. 16) and Syrkov Log deposits. It is a transition zone between coarse-grained dunite of the Solov'yev mountain and fine-grained marginal dunite. The geological scheme clearly shows that this zone corresponds to the region of the southern part of subconcentric structure (Fig. 3), which is characterized by the predominant sublatitudinal orientation of chromite schlieren with a northern dip (70-90°). Besides, there is the second conjugate (?) schlieren system of submeridional orientation. Porphyroclastic type of structure with signs of plastic deformation of olivine grains (kink bands, wavy extinction) predominates. Relict idiomorphic protocrystals are also common. Various structural and morphological types of chromite segregations can be observed in large-boulder dunites that remained after the Krutoi Log (Fig. 17) and Syrkov Log platinum placer washing. Most features of epigenetic chromitite described above, i.e. isolation of ore segregations in dunite, serpentinite rims, contraction cracks, porous texture can be observed here. A 10-15 minute walk (one way) to the Krutoy Log deposit (Fig. 16), close to which the first platinum ore accumulations were discovered in the late XIX century.



Fig. 15. Panoramic view of the dunite open pit



Fig. 16. View of the Krutoy Log deposit



Fig. 17. Schlieren of chromite in coarse-grained dunite, Krutoy Log deposit

GEOLOGICAL CHARACTERISTICS OF THE VOLKOVSKY MASSIF

The massif is located on the northern extension of a large intricately structured Tagil-Baranchinsky massif, composed mainly of gabbroids (Fig. 18). To the west of the massif, prevailing are metamorphosed Late Ordovician (*Mariinsky Suite* ($O_3 mr$) basalts, amphibolites, hornfels and kytlymites, to the east there are Silurian-Early Devonian volcanogenic-sedimentary sequences of andesite-basalt ($S_1 pv Pavdinsky$ and $S_{1,2}$ im Imennovsky suites) and basalt-trachytic (S_2 - D_1 tr Turinsky suite) formations (Fig. 18).

According to the geological mapping, the massif consists of three blocks or intrusive gabbro bodies, each having its own concentric-zonal structure (Fig. 19). The West block is composed of gabbronorite (*the Tagil-Kytlym complex* S_1 *tk*), and the Central and South blocks are composed of olivine gabbro with subordinate clinopyroxenite and olivinite (the Volkovsky complex S_2 V).

Quartz diorite, forming a large body in the central part of the massif, is similar in composition to quartz diorite located south of the Tagil granitoid massif and assigned to the Verkh-Tagil complex ($S_2 vt$) (Fig. 18). The intrusive character of gabbro and diorite contacts, expressed in the formation of eruptive breccias with diorite cement, as well as their composition distinction in the petrochemical diagrams, suggest gabbro and diorite formations to be of different age (*Maegov*, 1999). This view is reflected in the geological map (Fig. 18). However, most researchers (*Kashin*, 1948; *Vorobyeva et al.*, 1962; *Shteinberg & Eremina*, 1963; *Poltavets et al.*, 2006; *Eugeosynclinal* ..., 1984) consider quartz diorite of the Volkovsky massif as the product of magmatic differentiation of the single gabbro-diorite intrusion.

The contact relationship of syenite (the Kushvinsky complex S_2 - $D_1 k$) with gabbro of the Volkovsky complex is also ambiguous. The presence of syenite dikes in gabbros provided the basis for preconceived ideas about their intrusive relationships and a younger age of syenite (*Vorobyeva* et al., 1962; *Malakhov*, 1966; *State Geological...*, 1999; *Borozdina* et al., 2010). In contrast, G.B. Fershtater combines the rocks of the Volkovsky gabbroic and neighboring Kushvinsky syenite massif into a single differentiated gabbro-syenite series (Eugeosynclinal ..., 1984; Fershtater, 2013).

Attempts to solve the existing problems with the help of U-Pb dating of zircon of the Volkovsky massif gabbro (*Krasnobaev et al.*, 2007; *Fershtater*, 2013; *Anikina et al.*, 2014) and the Kushvinsky massif syenite (*Fershtater*, 2013) unfortunately have not given any definite result.

It has turned out that olivine gabbro of the Volkovsky massif contain four zircon populations of different age (Fig. 20): $2682\pm37 - 972\pm18$ Ma; $655\pm15 - 565\pm9$ Ma; 450 ± 12 Ma and 343 ± 8 Ma. The isotopic Hf-Nd systematics has shown that zircon aged 450 ± 12 Ma is likely to correspond to the rock formation time. Vein-shaped bodies of two-pyroxene labrador-bearing gabbro, intersecting olivine gabbro in the southwestern part of the massif, contain zircon aged 428 ± 7 Ma, most likely related to their crystallization time. A similar age of 436 ± 21 Ma (MSWD = 1.5) has been obtained for these gabbro using Sm-Nd technique for monomineral isochron (*Anikina et al.*, 2014).

Zircon from syenite of the Kushvinsky complex with the age of 440-428 Ma has proved to be identical in the age parameters, geochemistry and morphology to zircon from gabbro of the Volkovsky massif (*Fershtater*, 2013).

According to our data, granite-aplite dike bodies in olivine gabbro of the Volkovsky massif contain zircon aged 410 ± 10 Ma. The composition of zircon inclusions and the data of Hf-Nd systematics leave no doubt that this zircon is syngenetic to the rock, while its age is in good agreement with the reliable U-Pb dating of zircon from various granitoids of the Tagil zone (*Efimov et al.*, 2005; *Fershtater*, 2013). The age obtained does not contradict to numerous geological evidence of later granitoid complexes formation in relation to gabbroid rocks of the Uralian Platinum Belt.



Volcanogenic-sedimentary complexes



S₁₋₂ **Im**

 $S_1 pv$

The Turinsky Suite: trachybasalt, trachyandesite, trachyte and tuff

The Imennovsky Suite: andesite-basalt, basalt, andesite, hyaloclastite, tuff, tuff conglomerate intercalated with siliceous siltstone and limestone

The Pavdinsky Suite: tuffstone, tuff, dacite and andesite-dacite tuff, andesite-basalt tuff, andesite, carbon-siliceous shale, limestone.



The Mariinsky Suite: metamorphosed basalt, amphibolite, hornfel

Intrusive rocks:

Kushvinsky gabbro-syenite complex: Syenite, syenite porphyry

Verkh-Tagilsky gabbro-diorite complex: quartz diorite, diorite, gabbro-diorite

Volkovsky clinopyroxenite-gabbro complex:

Subalkaline gabbro, monzonite



 $\varepsilon_1 S_2 - D_1 k$



amphibole-pyroxene gabbro



clinopyroxenite

v S₁tk Tagil-Kytlymsky gabbro complex: gabbronorite, amphibole gabbro



Kachkanarsky gabbro-clinopyroxenite complex:O3 k



clinopyroxenite, olivine clinopyroxenite Contact-metamorphic rocks:



skarn



albitite, quartz-sericite

faul

faults

Fig. 18. Geological scheme of the Sredneural'sk Area (modified after State Geological Map of the Russian Federation, scale 1 : 200 000, Sheet-40-XXVIII, 1999)



Volcanogenic-sedimentary complexes

 S_2-D_1 tr The Turinsky Suite: trachybasalt, trachyandesite, trachyte and tuff

O, **mr** The Mariinsky Suite: metamorphosed basalt, amphibolite, hornfel

Intrusive rocks:



Fig. 19. The geological scheme of the Volkovsky gabbro-diorite massif (after Fominykh & Klevtsov, 1984)



Fig. 20. Concordia diagram for zircon from the olivine-anorthite gabbro of the Volkovsky massif



Fig. 21. Apatite-magnetite clinopyroxenite (southern block of the Volkovsky massif)

COPPER SULFIDE MINERALIZATION

The Volkovsky copper sulfide deposit is confined to the Central block of the same-name massif, while the Baronsky gold-palladium deposit is localized at the periphery of its South block. The deposits differ only in the amount of sulfide mineralization at comparable palladium and gold concentrations.

Characteristics of ore-bearing rocks

Detailed studies of the internal structure of the Volkovsky massif Central block by deep exploration holes (Maegov, 1999, etc.) have shown that the predominant type of rocks here are texturally homogeneous olivine gabbros. From top to bottom of the section, with decreasing olivine ferrosity and increasing plagioclase basicity (from An_{75-80} on top to An_{95} at the botton), they undergo structural changes from sub-ophitic to granoblastic. Labradorite gabbro is scarce in the upper parts of the section, mainly near the contact with diorite.

A characteristic feature of the Volkovsky complex gabbro, which distinguishes them from all other gabbros of the Urals Platinum Belt, is their anomalous phosphorus enrichment (up to 3 wt.% of P_2O_5). At the same time, high phosphorus concentrations are mainly observed in the upper part of the section and increase extremely unevenly towards the external contacts between gabbro and syenite. The maximum phosphorus content is characteristic of the sites with titanomagnetite mineralization.

According to their chemical composition, gabbro are referred to low and moderately alkiline $(K_2O + Na_2O = 0.68-3.28 \text{ wt.}\%)$, average -1.36 wt.%) (Fig. 21), high sodium $(K_2O/Na_2O = 0.06-0.59 \text{ average} - 0.19 \text{ wt.}\%)$, moderately low aluminous, moderately low titanium, moderately ferrous (f=0.46) rocks. They are characterized by relatively low ratios: Mg/(Mg+Fe) £ 0.55, Ni/Co = 0.34-1.70 1 μ Cr/V = 0.10-0.32 (Fig. 22).



Fig. 22. Contents of major components and strontium in the rocks of the Volkovsky massif.

Southern Block: 1 – labrador-bearing gabbro; 2 – olivineanorthite gabbro. Central block: 3 – olivine gabbroanorthite (after *Maegov*, 1999; *Bognibov & Balykin*, 1990); 4 – monzogabbro (after *Maegov*, *1999*); 4 – gabbro with apatite-magnetite mineralization (*Bognibov & Balykin*, 1990) MgO' = 0,55*Fe2O3 + 0,5*FeO + MgO The chemical composition of olivine gabbro and other rocks of the Volkovsky massif is shown in Table 4. Among homogeneous ophitic or granoblastic gabbro there are zones of rocks with taxitic texture due to irregular alternation of different melanocratic and granular sections. These zones form discontinuous bands that are subconcordant to the concentric structure of the massif and hidden layered gabbro. Associated with them is apatite -titanomagnetite, copper sulfide and gold-palladium mineralization in various parts of the massif.

Ore body structure and ore composition

The Volkovsky deposit is an arcuate ore zone, stretching almost 5 km along the north-eastern edge of the Central gabbroic block of the massif (Fig. 23). Inside the zone, taxitic areas with mineralization are developed unevenly, their length along strike and dip ranges from 600 to 800 meters, and thickness from 10 to 60 m. The ore bodies dip to the West (30-50°) in the southern part and to the South-West (74-80°) in the northern part of the deposit.

The ore bodies are composed of densely impregnated, sometimes solid apatite-titanomagnetite ores, gradually turning into coarse-grained clinopyroxenite or areas with high titanomagnetite impregnation. Copper mineralization spatially coincides with titanomagnetite mineralization, but develops unevenly – from minor sulfide impregnation to the formation of chalcopyrite-bornitetitanomagnetite and chalcopyrite-bornite ores.

Structural relationships of the main Cu-containing mineral, i.e. bornite, with other minerals are quite complicated. Its smallest inclusions are observed in crystals of non- saussuritized twinned plagioclase, in clinopyroxene, which has already shown a net of magnetite dissolution, and in apatite. The second form is mosaic development of bornite in clustures of cleavage pyroxene net, which is possibly synchronous in its formation time with the beginning of amphibolization. The third form is graphic intergrowth of bornite with digenite, enclosed in plagioclase and pyroxene.

Another leading mineral, i.e. titanomagnetite, occurs as rounded inclusions (ore drops) in pyroxene, plagioclase and apatite (Kashin, 1948),



Fig. 23. Geological scheme of the Volkovsky deposit (modified after Geological map, scale 1:10 000)

Table 4

Chemical composition (wt. %) the main types of rocks Volkovsky massif

Rock				Olivine	gabbro				Quartz	diorite	Monzo	gabbro	Sye	nite
№ n.n	1	2	3	4	5	6	7	8	9	10	11	12	13	14
SiO ₂	41,00	40,86	38,02	43,02	43,58	40,17	39,11	42,54	57,82	61,03	47,70	49,01	55,98	54,70
TiO ₂	0,46	0,53	0,67	0,58	0,99	0,85	0,91	0,34	0,88	0,58	1,69	1,24	0,30	0,48
Al ₂ O ₃	22,08	14,24	16,62	20,76	11,26	13,76	16,50	24,39	16,46	15,70	15,08	17,13	20,97	23,24
FeOtot.	9,59	8,76	11,96	9,30	15,59	13,59	17,84	6,99	7,04	6,53	12,12	9,50	3,03	3,68
MnO	0,15	0,14	0,13	0,12	0,20	0,11	0,11	0,06	0,13	0,10	0,31	0,24	0,10	0,08
MgO	6,10	8,11	9,15	5,80	11,65	8,31	8,19	5,37	3,74	3,31	6,19	5,09	0,39	1,61
CaO	15,79	17,73	15,58	15,34	14,80	17,79	15,67	16,14	6,94	5,24	9,65	8,77	5,12	7,04
Na ₂ O	1,36	0,93	0,85	1,23	0,72	0,62	0,62	0,84	5,00	4,92	2,53	4,00	3,60	4,39
K ₂ O	0,22	0,40	0,26	0,09	0,06	0,10	0,06	0,05	0,82	1,50	1,10	1,03	6,13	2,69
P ₂ O ₅	1,87	3,32	2,58	0,19	0,18	0,26	0,22	0,15	0,36	0,32	1,42	1,04	не.опр.	не опр.
Total	98,62	95,02	95,82	96,43	99,03	95,56	99,23	96,87	99,19	99,23	97,79	97,05	95,62	97,91
Sr (ppm)	3230	1130	1440	1420	600	820	840	1530	1280	1340	1860	2200	не.опр.	не опр.

Note: 2-12 (after Mayegov, 1999); 13,14 (after Malakhova, 1968).



Fig. 24. Schlieren disseminated bornite (A) and massive chalcopyrite (B) in gabbro from the Volkovsky deposit



Fig. 25. Morphology of ore minerals

A. The grains of titanomagnetite (Ti-mt) surrounded by fine-grained chlorite mass with fine inclusions of bornite (Bn) and kotulskite (PdTe). Titanomagnetite is overgrown by rim of grossular-andradite (Gar), bornite (Bn) and chalcopyrite (Cp). B. Bornite with lamellae of chalcopyrite and inclusion of cobaltite (CoAsS) and kotulskite. Bornite from the edges corroded and replaced by secondary magnetite and garnet (grossular-andradite series). Ap – apatite; Ilm – ilmenite; Amf – amphibole (actinolite); ChI – clinochlore

as well as graphic intergrowths with ilmenite and non-metallic minerals (pyroxene and amphibole). These kinds of bornite and titanomagnetite effusions are referred to their rare, earliest high-temperature modifications.

Disseminated gabbro impregnation with grain sizes, ranging from millimeter fractions to 5 mm, composes the main volume of titanomagnetite and copper-sulfide ores. Based on the analysis of structural and textural relationships of ore minerals and the patterns of their chemical composition changes, it has been established (*Poltavets et al.*, 2006) that the main ore mineral effusions in this period began in the following time sequence, that could be traced with the lowering temperature: titanomagnetite, ilmenite, bornite, digenite, bornite + chalcopyrite, chalcopyrite, covelline, pyrite. Chalcopyrite, the second most common sulfide, is also found in all gabbro varieties, mainly in the form of impregnation with a grain size up to 4 mm.

The deposit has widely developed schlieren-like and vein-like isolations from 0.5 to 15 cm in size, with bornite, chalcopyrite and bornite-chalcopyrite composition (Fig. 24). This is the latest sulfide

generation that replaces early silicates and titanomagnetite in cleavage and fracture, accompanied by secondary amphibole (tremolite), chlorite and epidote (Fig. 25).

Variations in sulfur isotopic composition of bornite (n = 3), chalcopyrite (n = 6) and pyrite (n = 1) from copper ores of the later generation and clinozoisitic rocks of the Volkovsky deposit are relatively few: δ^{34} S= -1,2 ... 3,1 ‰ and correspond to sulphur from magma chambers (0±5 ‰).

Platinum-group minerals

The Volkovsky deposit ores are accompanied by precious-metal mineralization, presented by palladium tellurides (merenskite, kotulskite and keithconnite) and gold. Besides, hessite (Ag, Pd), Te occurs in rare cases. According to Volchenko et al (1996), the total platinum and gold content in copper sulfide ores (Cu = 0.6-4.6 wt.% and S = 1.0-1.5 wt.%) and copper sulfide-apatite-titanomagnetite ores (Cu = 0.3-0.7 wt.% and S = 0.5-1.2 wt.%) ranges from 1 to 7 g/t and from 0.1 to 0.3 g/t, respectively. Precious metal minerals are in the form of inclusions 0.001-0.025 mm in size, mainly in sulfides (bornite, chalcopyrite), rarely confined to titanomagnetite, clinopyroxene and hornblende (Murzin et al., 1988; Poltavets et al., 2006) and, most likely, are associated with the latest-generation bornite and chalcopyrite.

The Baronsky deposit sulfides, tellurides and palladium arsenides with gold are disseminated in the rock and not accompanied by significant accumulations of sulfides, although they are in constant association with them (*Zoloyev et al.*, 2001; *Zaccarini et al.*, 2004; *Anikina et al.*, 2004). This mineralization is obviously associated with lowtemperature metamorphic parageneses (prehnitepumpellyite facies, T 300-350°C, P 1.5-6.5 kbar), and the accompanying sulfides are similar to the late-generation bornite-chalcopyrite ores of the Volkovsky deposit (Fig. 26).



Fig. 26. Bornite (Bn) with lamellae of chalcopyrite (Cp) and composite inclusion of isomertieite (Pd₁₁As₂Sb₂), kotulskite (PdTe), palladium and silver telluride (Pd,Ag,Te), and electrum. Bornite is replaced by chlorite (ChI) and magnetite (Mt)



Fig. 27. Calculated isotopic composition of oxygen and hydrogen fluid in equilibrium with hydrous silicates: 1 – with amphibole hornblendite, amphibole-pyroxene gabbro from Platinum Belt; 2 – with copper sulfide ore with amphibole; 3 – with clinocoizite; 4 – with amphibole clinopyroxenite; 5 – with veined clinochlore. Diagram shows the field of the isotopic composition of water reservoirs involved in the hydrothermal mineralization according to *Faure* (1989) and *H\vec{e}fs* (1983)

Mineralization genesis

The majority of researchers support a magmatic model of titanomagnetite-copper sulfide mineralization. According to this model, water was generated by water-containing (6-8 wt.%) basalt chamber as a result of crystallization differentiation, while ore deposition occurred in the final stages of the intrusion formation from the residual fluid-enriched melts (*Poltavets et al.*, 2006). The role of the residual fluid is reduced here to the autometasomatism and gabbro recrystallization processes to form its taxitic differences. At the same time, the metasomatic nature of the most recent sulfide-quartz-calcite veinlets is recognized, and metasomatizing fluids are associated with the intrusion of post-ore quartz diorite bodies. In other concepts of the Volkovsky type copper sulfide mineralization, i.e. metasomatic and hydrothermal ones (*A.A. Efimov, L.P. Efimova, E.S. Kontar et al.*), the fluid role is fundamental. Thus, according *Efimov et al.* (1999, 2002), the original magma was "dry" and already crystallized gabbro experience water metamorphism (metasomatism). The fluid responsible for the formation of noble-metal and copper-sulfide mineralization, was abyssal and appeared in the most recent epoch of the Platinum Belt formation along with granitoid melts.

The oxygen and hydrogen isotopic composition of rock-forming silicates from ore-bearing rocks of the Volkovsky and Baronsky deposits and the calculated isotopic composition of their equilibrium fluid corresponds to the juvenile source, which is consistent with both suggested mineralization hypotheses (*Murzin et al.*, 2006).

Whereas the isotopic composition of the fluid, forming low-temperature mineral assemblages (tremolite, chlorite, clinozoisite, carbonate), corresponds to the metamorphic water (Fig. 27), this fluid could be responsible for a substantial redistribution of ore components in taxitic areas and for the formation of separated in space later ore parageneses: copper sulfide and gold-palladium.

The cross-cutting patterns of the late copper sulfide (bornite-chalcopyrite) mineralization at the Volkovsky deposit in relation to the early apatite-titanomagnetite mineralization has been determined by geochemical findings of copper, titanium and phosphorus halos (*Fominykh & Klevtsov*, 1984). Special studies of precious metal distribution within the deposit have not been conducted.

FIELD TRIP STOPS

The Volkovsky deposit has been known since 1912. Its geological survey mapping and prospecting were carried out during 1927–1943. The detailed exploration of the Volkovsky deposit and its reserves calculation were completed in 1987. When calculating reserves at a cut-off copper grade of 0.3% for open and 0.4% for underground works in copper-iron-vanadium ores and at 14% of iron in iron-vanadium ores, the total of 371 ore bodies were outlined within the ore zone with a total length of about 5 km. Of these, 134 were composed of copper-iron-vanadium, 187 of iron-vanadium and 50 contain both types of ores. The ore bodies were steeply dipping with dip angles of 40-70° WSW.

Average ore content is as follows: Cu (0.63%), Fe (14.90%), V_2O_5 (0.24%), P_2O_5 (3.22%), Au (0.11g/t), Ag (3.17 g/t), Pd (0.12 g/t), Te (1.02 g/t), Se (8.98 g/t), S (0.41%), and F (0.10%).

The industrial development of the deposit began in 1982. At low metal levels the ores are easily enriched by flotation, giving a quality copper concentrate in the output. The deposit ores are open-pit operated. The ore is crushed at a crushing-and-sorting plant, and then delivered by rail to Krasnoural'sk for further processing at the "Svyatogor" processing plant (former Krasnouralsky copper smelting plant).

STOP 1: Viewing platform, open-pit mine of the Lavrovo-Nikolayevsky area (Fig. 28)

The open-pit mining and concurrent oxidized ore mining began in 1964. Off-balance oxidized ores were stockpiled in mine dumps (3168000 metric tons of oxidized ore with 0.85% Cu and 7.68% Fe), from which between 1975 and 1987 about 1400 metric tons of cementation copper were obtained by heap leaching.

The industrial ore production was carried out from 1982 to 2005, and 24.5 mln tons of ore were produced, including:

- Oxidized ore: 3.1 mln. tons;

- Mixed + primary: 20.8 mln. tons;

- Iron-Vanadium: 0.5 mln. tons.

In 2005, the open-pit of the Lavrovo-Nikolayvsky area reached the final project section and was closed.

STOP 2: North-West area of the Volkovsky deposit (the new open-pit, Fig. 29)

The ore reserves of the North-West area is estimated at 11.3 million tons at the following average content of precious metals (g/t): Pd 0.12; Au 0.1; Ag 3.35. The open-pit for the mining of the North-West area became operational in 2010. The project parameters of the open-pit until the end of mining are given in the Table 5.

The most significant ore body within the open-pit is number 1. Its total length is 1.572 m along strike, and steep dip of 70-50° to the west. The length of the ore body along dip varies from 197 m to 920 m. Its thickness varies greatly from 5 m to 71 m. The disseminated ores have mainly copper-iron-vanadium specialization.

Project parameters of the North-West open-pit

Table 5

Open-pit parameters	Indices
Average surface elevation, m	310
Open-pit depth, m	160
Open-pit bottom mark, m	150
Open-pit bottom width, m	66
Open-pit bottom length, m	670
Open-pit bottom area, thousand m ²	43
Open-pit surface length, m	1035
Open-pit surface width, m	450
Open-pit surface area, m ²	413
Open-pit volume, thousand m ³	32690
Ore reserves, thousand tons	12300



Fig. 28. A: Bay-Nicholas quarry (photo taken in 2009); B: Massive bornite-chalcopyrite ore (ore body number 67)



Fig. 29. Northwestern quarry (photo taken in 2014)

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34	Kaapteauft florare-pereas- ofmannenit niepers, Quare- betkauphiboldiork,	GS	Fannacrise examps a newannen. Thouschiefer und Sandsteine,	е х
34	Kuapuesuli porozoodmanko- nek ziopers. Quazeanphi- boldioriz.	69	Spennerae ennuz. Kieselschiefer.	0 8
34	Knapusnuk Goreronak go- pera. Quarzbiotitdiorit.	70	Respuerse a 4, porosana. Quarsite.	•
35	Аплятовидный біотитовый гранить. Біосідугаліс.	Di	Мраноровидные извест- няки нажислевонскаго возрастя. Kalkateine,	A
36	опантовадный ботито-ро- говообязатовый гранкть, Biotitamphibolgranit.	DĨ	Доложитоные навестняка в. доронскаго взораста. Dolomitische Kalksteine.	
n	ланитовиданий роговооб- вляковый гранать, Атгрий- bolgranit.	Qı	Horriaionesosica orao- menia. Postpliorene Abla- gerangen.	erenset ofpast
38	Гранофировый гранитовый аллить (акьбитовый). Gra- nophyrische Granitaplite.	Q:	Conpercames oraoacenis plars a dozors. Recente Ablagerungen.	But T