

Mineralogy, Geochemistry, and Origin of the Nezhdaninsk Gold Deposit (Sakha-Yakutia, Russia)

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Abstract—The Nezhdaninsk gold deposit is located at a junction of four regional faults in the South Verkhoyansk synclinorium. The ore bodies are the mineralized cataclitic zones and the veins. They are hosted by metamorphosed terrigenous sandstone-slate sequences intruded by stocks and dikes of silicic and intermediate rocks. The age of the magmatic rocks was found to vary from 154 to 94 Ma. The deposit was formed during three stages. Gold-bearing ore bodies were formed during the second stage. A regeneration of quartz and sulfides, and a crystallization of silver-bearing mineral assemblages occurred during the third stage. The gold deposition began at the metasomatic alteration of the host rocks. The altered rocks consist of quartz, sericite, carbonates, and disseminated pyrite and arsenopyrite. These sulfides contain “invisible” chemically bound gold. Native gold crystallized in quartz veins. The native gold is associated with coarse-grained arsenopyrite and pyrite, and is coeval with sphalerite–galena–sulfosalt aggregates. The fluid inclusion study showed that the ore formation occurred from two immiscible fluids. One of them was a H₂O + CO₂-dominated fluid with N₂, CH₄, and soluble chlorides (up to 4.5 wt % equiv NaCl). The second fluid was a vapor phase consisting mainly of CO₂ and CH₄. The stable isotope (S, C, and O) study in the minerals suggests that sulfur, water, and CO₂ of magmatic origin prevailed in fluids. They were mixed with components incorporated from the host rocks. The magmatic and host rocks interacted with the isotopically light and possibly meteoric waters. The mineral crystallization occurred at temperatures of 360 to 175°C, and under pressures of 1.2 to 1.7 kbar from mineral-forming fluids with a pH equal to 5.5–6.1. The CO₂ mole fraction of the fluids was from 0.25 to 0.3. The formation of the economic gold ores was linked with the postaccretion tectono-magmatic activation of the South Verkhoyansk synclinorium in the Early to Late Cretaceous.

INTRODUCTION

The Nezhdaninsk deposit is a mesothermal low-grade disseminated and lode deposit located within a terrigenous sequence containing organic matter. The deposit is situated on the western slope of the Suntar-Khayata Mountain Range about 440 km east of Yakutsk. Mesothermal vein and disseminated gold deposits hosted by terrigenous sequences are beginning to play an important role in the gold production of the Russia (Safonov, 1997). This tendency is characteristic of other countries that involve newly discovered gold deposits in mining. Therefore, an understanding of processes responsible for the formation of these deposits is of interest both from academic and industrial points of view.

Some information on the geological setting of the Nezhdaninsk deposit, the structural control of ores, mineral composition, and zonality have been previously published (Gamyranin, 1991; Yanovskii, 1993; Silichev, 1970; Silichev and Belozertseva, 1981). However, data on mineral paragenetic assemblages of

altered rocks and ores, mineral chemistry, fluid inclusions, and the stable isotope ratio of minerals have not been completely studied. The absence of this data made it difficult to reconstruct the formation conditions and to reveal the source of a fluid and its constituents. The problem of linking deposits of this type with magmatism and metamorphism is debatable (Bortnikov *et al.*, 1996; Kerrich, 1990; Sillitoe, 1997).

This paper contains the original results of a study of the mineral composition, fluid inclusions, stable isotopes, and age of intrusive rocks obtained by the authors, and a compilation of data on the geology of the deposit and the ore district. These data allow us to propose a model for the formation of the Nezhdaninsk gold deposit.

GEOLOGICAL STRUCTURE OF THE REGION AND DEPOSIT

The Nezhdaninsk deposit is located within the South Verkhoyansk synclinorium (Fig. 1a) that is a part

of the Verkhoyansk–Kolyma orogenic belt (Parfenov *et al.*, 1993). The South Verkhoyansk synclinorium is composed of the sandstone–slate sequence of Carboniferous to Jurassic ages, deposited on the Siberian passive continental margin. The Mesozoic folding of these rocks resulted from the collision of the southern part of the Verkhoyansk passive margin with the Okhotsk microcontinent. There are several belts of ultra- and mesoabyssal granite within the Verkhoyansk fold belt. The Rb–Sr ages of the large batholite-like massifs of biotite–garnet granites are dated at 140 to 150 Ma, while Ar–Ar dating gave 120 to 123 Ma (Parfenov, 1995). A generation of these granites was due to the melting of the craton margin during its subduction under the Kolyma–Omolon superterrain (Parfenov *et al.*, 1993). The Ar–Ar age of the small granite massifs is from 92 to 97 Ma. Their emplacement is related to the Okhotsk subduction. Accretion of the Okhotsk–Chukotsk belt is assumed to have triggered an activation of the Verkhoyansk orogen.

The present day geological structure of the Verkhoyansk synclinorium resulted from the compli-

cate geodynamic history of the northeastern territories of Russia.

A number of global events is responsible for processes of tectonics, magmatism, and ore formation in the South Verkhoyansk synclinorium. The northern part of the South Verkhoyansk synclinorium was influenced by the Udsak subduction that reactivated the submeridional Minorsk (Kiderikinsk) fault. According to gravity modeling, a thickening of the Earth's crust up to 45 km occurred during this period (Parfenov, 1995). The next activation of the region resulted from the formation of a new Benoff–Zavaritskii zone (the Okhotsk subduction) east of the Nezhdaninsk ore deposit. The thickening of the Earth's crust and the generation of magma chambers are related to this phenomenon.

The Nezhdaninsk deposit is located in the Dybinsk anticline flexure-like fold at a junction of four systems of fault zones (Fig. 1b). One limbs of the fold dips gently to the west, while the other limb dips steeply to the east. The deposit is hosted by Lower Permian rocks. Rocks of the lower Djuptagansk suite are siltstones interlayered with sandstones. Siltstones contain up to 2 wt % of organic matter and from 5 to 10 wt % of car-

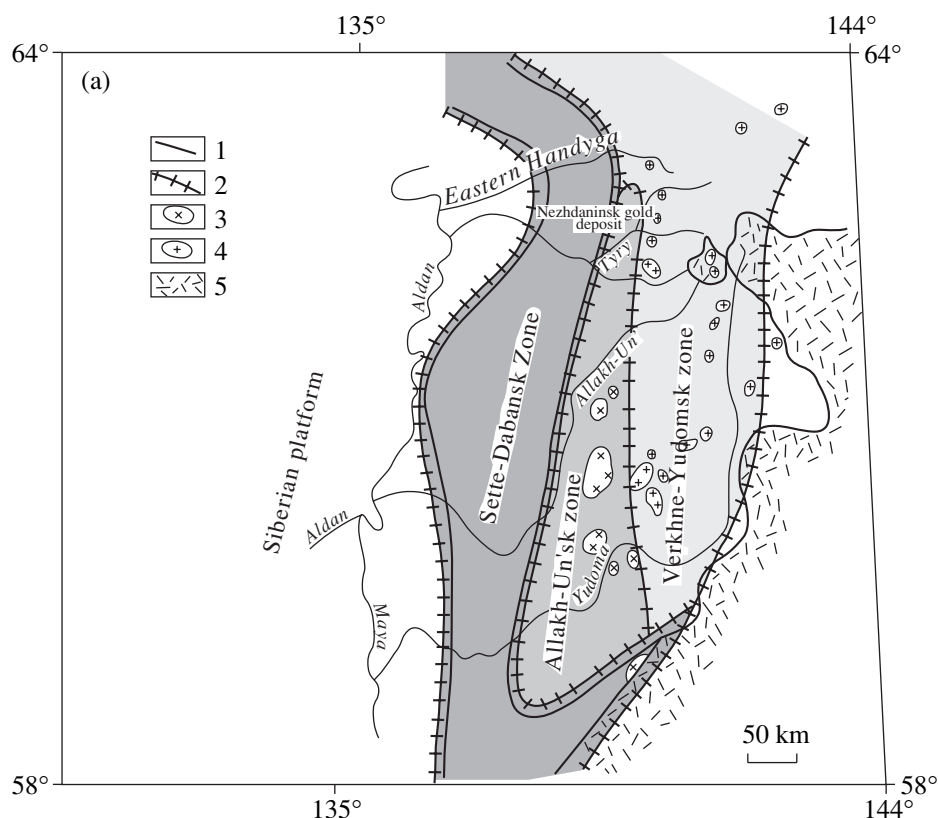


Fig. 1. (a) Metallogenic sketch of the South–Verkhoyansk synclinorium and (b) geological sketch of the Nezhdaninsk gold deposit. (a) (1) Boundaries of structural zones, (2) boundaries of metallogenic zones, (3) accretion amphybole–biotite granite massifs, (4) postaccretion granodiorite–granite and granite–leucocratic massifs, (5) volcanics of the Okhotsk–Chukotsk volcano–plutonic belt; (b) (1) siltstones and sandstones of the Menkechinsk suite, (2) siltstones and slates of the Dybinsk unit, (3) siltstones of the Djuptagansk suite, (4) sandstones, (5) spessartite dikes, (6) faults, (7) ore bodies.

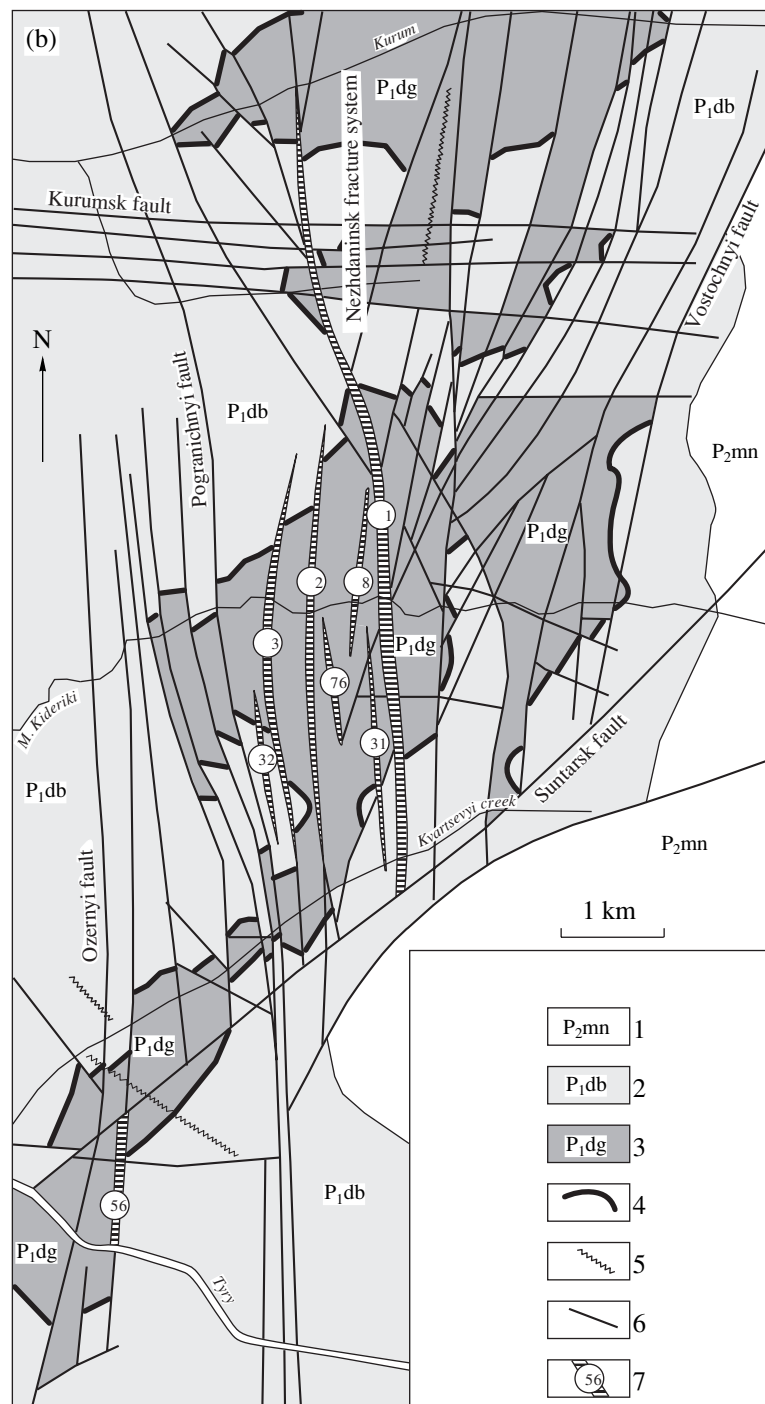


Fig. 1. (Contd.)

bonate matter. Siltstones of the upper Dybinsk suite contain organic matter and abundant disseminated marcasite. Low Permian rocks are overlain by sediments of the Upper Permian Menkechinsk suite. This suite consists of the interlayered sandstones and siltstones.

The terrigenous rocks were transformed due to regional metamorphism. The newly formed minerals constitute about 40–60% of the volume of primary

rocks. Chlorite + sericite and chlorite + carbonate mineral parageneses of the greenschist facies are of the typical metamorphic assemblages.

The Nezhdaninsk deposit is confined to a junction of the Kiderinsk longitudinal deep fault with the Suntarsk oblique fault. The Kiderinsk fault is traced up to 500 km along its strike. It subdivides the South Verkhoyansk synclinorium into two blocks; the west-

ern block is uplifted along this fault up to 1.5 to 2 km. Within the ore deposit area, the Kiderikinsk fault forms the Nezhdaninsk system of subparallel fractures, milonitization, and cataclasis sutures, which tend to join at the depth. The Tyrynsk fault, which extends from east to west for 100 km, is traced as the Kurumsk zone of intensive fracturing within the deposit. The Suntarsk fault is up to 0.5 km wide, and extends in the northeastern direction up to 300 km. An occurrence of lamprophyre dikes and fragments of garnet-bearing granite-gneisses indicates that the Suntarsk is a deep-seated fault and intersects the basement.

Different magmatic rocks are found in the ore deposit area. The Kurumsk massif is elongated along the north-south direction and is confined to the steep eastern limb of the Dybinsk anticline. It is surrounded by an aureole of hornfelses up to 2–3 km wide. It consists of amphibole-biotite granodiorite with subordinate biotite leucogranite. The Rb-Sr age of granodiorite is between 105 and 124 Ma. The Ar-Ar age of biotite granite is 97 Ma (data from P. Layer, University of Alaska). The massif is intersected by dikes of granite, aplite, and pegmatite.

The so-called Geldinsk stocks includes a number (13) of small intrusive bodies with sizes ranging from 50 × 100 m to 700 × 1200 m. They are situated on the southwestern flank of the ore deposit. They are ellipsoid or irregular in shape. These stocks are composed of diorite and quartz diorite. A high aluminum content and the predominance of Na over K are typical of these rocks. The Rb-Sr and Ar-Ar ages of quartz diorite are 92 to 106 Ma, and 92 Ma, respectively.

The dikes are widespread within the deposit. Dikes are confined to the submeridional, northeast, and northwest trending faults. The thickness of the dikes ranges from 0.5 to 0.8 m. They extend from a few hundreds meters to 10 km. Three types of dikes have been recognized: (1) Early Jurassic gabbro-diorite (K-Ar age is 140–154 Ma), (2) lamprophyre (K-Ar age of biotite is 131 Ma and the K-Ar age of amphibole is 94 Ma), and (3) diorite.

Geophysical data suggest the presence of a large massif of magmatic rocks that is located about 2 km from the deposit.

The multistage emplacement of magmatic rocks within the region of the Nezhdaninsk deposit is evident. Late Mesozoic magmatic rocks of the Late Jurassic gabbro-diorites are confined to the Setan'insk and Minorsk faults. They were formed synchronously with the Mesozoic folding, as well as the Kurumsk fracture zone. The Nezhdaninsk system was generated later. The Kurumsk and Dybinsk stocks are younger than the granite batholithes by about 35 Ma. These stocks are assumed to be genetically related to a magmatic event that was not linked with the Udsu subduction. An emplacement of the Kurumsk massif and Geldinsk stocks occurred at the period of formation of the Okhotsk-Chukotsk volcano-plutonic belt. These events

were related to the compression and formation of a new subduction zone along the eastern margin of the continent.

Three stages of mineral formation have been recognized in the ore deposit on the basis of spatial distribution, relationships among dikes and mineral aggregates, and minerals (Gamyagin, 1991). Numerous quartz veins and veinlets with poor sulfide and low-grade gold mineralization were formed during the early stage. Economic mineralized zones and lodes were formed during the second stage. They intersect barren quartz veins. Numerous quartz veinlets crosscutting the gold-bearing bodies were deposited during the third stage. The injection of dikes separates the mineral aggregates of the second and third stages.

The modern structure of the Nezhdaninsk deposit resulted from repetitive displacements in different directions along the transverse fractures linked with late folding, preintrusive longitudinal, postintrusive faults, and superimposed fractures. Most of these fractures host the ore mineralization.

More than 50 ore bodies located in the core of an anticline have been recognized within the Nezhdaninsk deposit. They are mostly confined to steeply dipping fractures, zones of milonitization, and cataclasis of the Nezhdaninsk system of a submeridional direction. Morphologically, the ore bodies are subdivided into three groups: (1) mineralized fracture zones consisting of quartz-sulfide veinlets, sulfide impregnations in the altered host rocks, and quartz-sulfide lenses (ore bodies Nos. 1, 3, 31, 56); (2) concordant veins (No. 8) and discordant plate-like and lenses-like veins (Nos. 5, 6, 14, 21); and (3) elongated stockwork-like veinlet zones (No. 76). The most important ore body—zone No. 1—is up to 10 km long, and extends to a depth of 1.9 km without evidence for its pinching out. Its thickness ranges from 2 to 40 m.

METASOMATIC ALTERATIONS

The host rocks underwent intensive metasomatic transformation within the ore deposit. Four types of wallrock alteration have been recognized. Two of them (propylites and sericite-carbonate rocks) are of regional distribution. Alterations of two other types are temporally and spatially related to the ore bodies. They are the berizites containing a syngenetic disseminated ore mineralization and sulfidization zones adjacent to the veins.

The berizites are mostly confined to the Nezhdaninsk fracture system. Thickness of the alteration halo is from 15 to 20 m, increasing up to 70–100 m in the places of the most intense fracturing.

Quartz-carbonate-sericite, quartz-carbonate-albite-sericite, and quartz-carbonate-albite facies of the berizites have been distinguished. Each facies is hosted by definite structures and displays a typical metaso-

matic zonation, and the typical mineral paragenetic assemblage of a back zone.

Berezites, which replace siltstone sequences containing few sandstone layers, are controlled by shear, foliation, and milonitization zones. Metasomatites of the quartz-carbonate-sericite facies are observed at upper parts of the sequence, while the beresites of the quartz-carbonate-albite-sericite facies are situated lower in the section. In essentially sandstone sequences, where the berezization is confined to the zones of intense fracturing and brecciation, metasomatites of the quartz-albite-sericite facies occur at the upper levels, while quartz-carbonate-albite metasomatites are formed at the deepest levels. The spatial distribution of the metasomatites of the different facies predetermines the vertical and lateral zonation pattern of the ore bodies.

The direction of the replacement reaction of albite by quartz and sericite defines the facies of the metasomatites. In the quartz-carbonate-sericite facies, albite is replaced by sericite and quartz within both outer and inner zones. As a result, the amount of sericite and quartz is increased. Within all zones of the metasomatites of the quartz-carbonate-albite facies, the replacement of sericite and quartz by albite and an increase of the albite amount are observed. An albitization is found in the outer zone of the metasomatites of quartz-carbonate-albite-sericite facies, while a sericitization occurs in the inner zone. A different direction of the replacements is caused by the thermal gradient in the system. The arsenopyrite geothermometer gave the temperatures of 370–380°C for the formation of the quartz-carbonate-albite metasomatites under a pressure of 1.5 kbar and the pyrite-pyrrothite buffer.

The formation of the metasomatites occurred with the input of arsenic and sulfur. This caused the replacement of ferruginous carbonate and sericite (fengite) by their varieties with lower iron contents and sulfide minerals, which are syngenetic to the berezites. The intensity of replacement reaction increased within the inner zones. The ferruginosity ($f = \text{Fe}/(\text{Fe} + \text{Mg})$) of the newly formed sericite and carbonate from the outer zones was equal to 0.20–0.40, while ferruginosity decreased to 0.05–0.20 for minerals from the inner zones. The fraction of carbonates of the magnesite-siderite series decreases when the intensity of the rock transformation is increased. An inverse correlation between the sulfide content in rocks and the iron content in minerals comprising these rocks was observed. The coefficients of correlation for carbonate and sericite are –0.94 and –0.92, respectively. A partial loss of Ca, Mg, and CO_2 occur during the replacement process.

The sulfide halos formed during the berezization display the zonation that coincides with the metasomatic zonation. Pyrite is a sulfide mineral in the outer metasomatic zones of the facies, while pyrrhotite is found in the quartz-carbonate-albite facies, as well.

The berezites and syngenetic disseminated ore mineralization are older in relation to the gold-bearing milky-white quartz with coarse-grained pyrite and arsenopyrite. This is because there is no increase in the intensity of the berezization and syngenetic sulfidization at the contact of these veins with host rocks, and one and the same vein intersects several zones of the alteration halo, including preore propylites and sericite-carbonate rocks. The formation of the thick and elongated ore bodies with disseminated sulfides occurred during the berezization stage. The average Au grade of the disseminated ores is 1 to 4 g/t. The ores of higher grade were locally formed as a result of the superposition of later veins and veinlets (see below).

The sulfidization of host rocks was observed around the quartz veins as arsenopyrite halos of small thickness (15 to 20 cm). The arsenopyrite metacrysts are shortprismatic in a shape. They were formed by the replacement of iron-containing minerals of host rocks. The synberezite arsenopyrite and pyrite are found within the sulfidization halos.

The process of gold deposition began during the alteration of host rocks along the fracture zones. Two arsenopyrite samples have been studied by instrumental neutron activation analysis (INAA), electron probe microanalysis, and secondary ion mass spectrometry (SIMS) (Genkin *et al.*, 1997). INAA showed that arsenopyrite contains 2.0 to 938.0 ppm of gold. SIMS found the range of the gold concentration to be from 1.17 to 386.64 ppm (average value is of 140.24 ppm). Pyrite contains up to 12.47 ppm Au (average value is of 2.45 ppm). Electron probe microanalysis showed that the gold content in arsenopyrite is up to 0.146 wt %. An extremely inhomogeneous distribution within the individual grains was revealed. Gold is concentrated in arsenopyrite depleted in arsenic (30.6 to 32.0 at. % As). The results obtained by different techniques disagree, because different grains have been studied. Mössbauer spectroscopy proved that gold was incorporated in the crystal structure of arsenopyrite. Therefore, chemically bound gold in sulfides is a major mode of gold occurrence in the older paragenetic assemblage.

PARAGENETIC ASSEMBLAGES AND CHEMISTRY OF ORE MINERALS

The earliest association (II–I) of the quartz veins and veinlets includes quartz, scheelite, coarse-grained arsenopyrite, and pyrite. The nearly coeval precipitation of these minerals is evident due to an absence of the cross-cutting and the replacement relationships among them. Sulfides are largely confined to vein selvages (Fig. 2), or overgrow the host rock fragments. The minerals fill in the intergranular space of milky-white vein quartz. Native gold in an association with arsenopyrite and pyrite aggregates is very rare. It overgrows or heals the tiny cracks in sulfide grains (Figs. 3a, 3b). Native gold occurs as thin grains of 0.16 to 0.25 mm in size. It overgrows the arsenopyrite aggregates or fills in the inter-

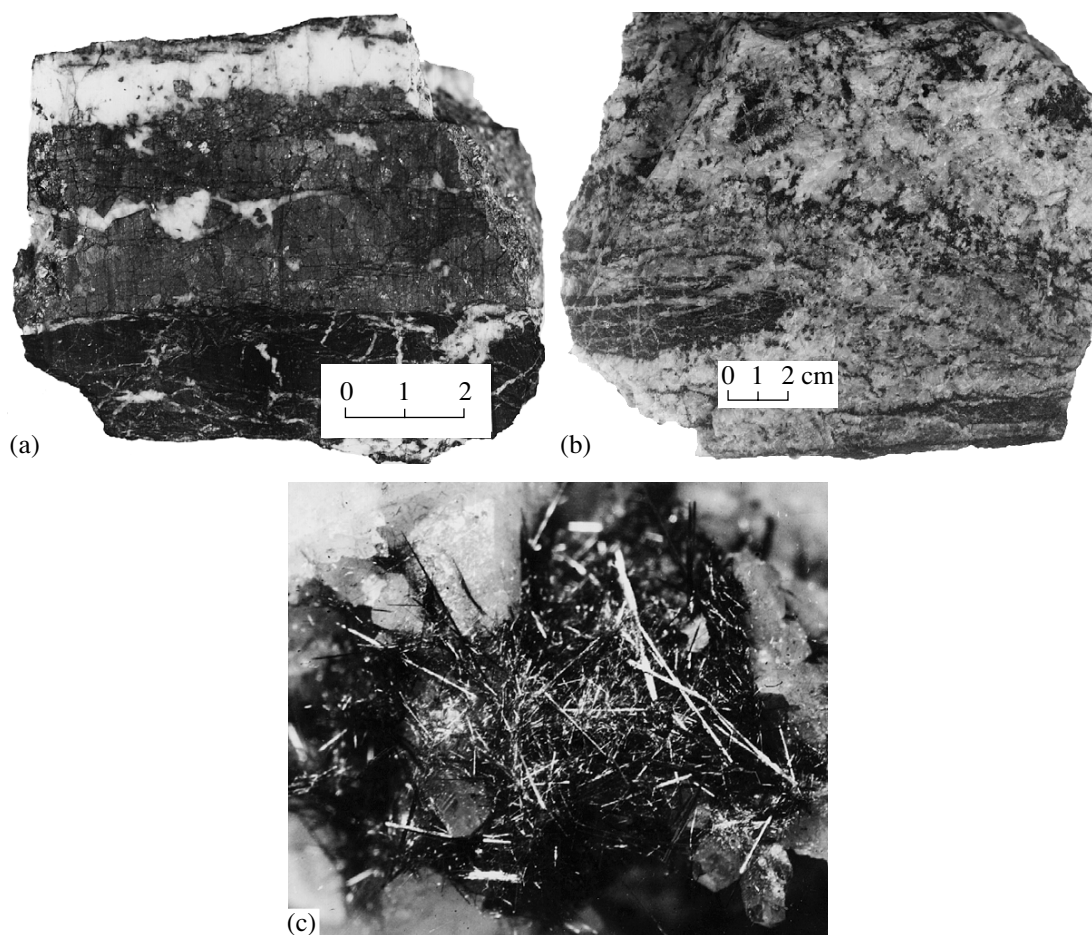


Fig. 2. Ore veins of the Nezhdaninsk deposit (polished samples). (a) Arsenopyrite bands at the selvage of a quartz vein, (b) patches and impregnations of gold-sulfosalt aggregates in quartz, (c) needle crustals of owyheeite in a cavity in quartz.

granular spaces of pyrite and arsenopyrite. However, a replacement of the sulfides by gold was not observed in many cases. Gold is suggested to have been partly coprecipitated with the minerals of the older paragenetic assemblage and to have completed its deposition.

A chemical composition of arsenopyrite from this assemblage is depleted in arsenic (from 31.2 to 32.2 at. %) in relation to the stoichiometric composition of FeAsS . The arsenopyrite grains commonly display a zonal distribution pattern of arsenic (Gamyanin, 1991). A gold admixture in arsenopyrite of 1.5 to 48 ppm (average is of 12.4 ± 9.43 ppm) was found (Bortnikov *et al.*, 1993). Fineness of native gold from this assemblage varies from 740 to 946‰.

The next paragenetic mineral assemblage (II–II) includes chalcopyrite, sphalerite, tetrahedrite, galena, and native gold (Figs. 2b, 3c, 3d). These minerals are younger, because they fill cracks in the fractured grains of earliest pyrite and arsenopyrite, and comprise veinlets in milky-white quartz. Mineral aggregates consisting of sulfides and quartz fill in cracks in older milky-white quartz and coarse-grained pyrite-arsenopyrite aggregates. Sphalerite is sporadically replaced by chal-

copyrite, galena, and tetrahedrite. However, mutual boundaries between native gold, chalcopyrite, galena, and sphalerite were observed, which suggests the coprecipitation of these minerals. Native gold occurs as rounded or ellipsoid inclusions intergrown with sphalerite, galena, and tetrahedrite (Figs. 3d, 3e). It rarely heals thin cracks in sulfides, and completes the deposition of this paragenetic assemblage.

Sphalerite from the Nezhdaninsk deposit contains a low amount of iron (0.8 mol % FeS). Tetrahedrite displays a variation of silver and arsenic from 1.4 to 25.6 wt %, and from 0.4 to 9.9 wt % (Gamyanin and Bortnikov, 1988). Older generations of tetrahedrite are richer in arsenic and prevail at depth. The fineness of native gold of this assemblages varies from 740 to 845‰.

These assemblages were followed by the formation of gold-sulfosalt paragenesis (II–III), which includes many sulfosalt minerals such as bournonite, geocronite, boulangerite, zinckenite, meneghenite, famatinite, chalcostibite, and berthierite, and fine-grained arsenopyrite and pyrite, as well. The latest deposition of these minerals is evident because they crosscut the

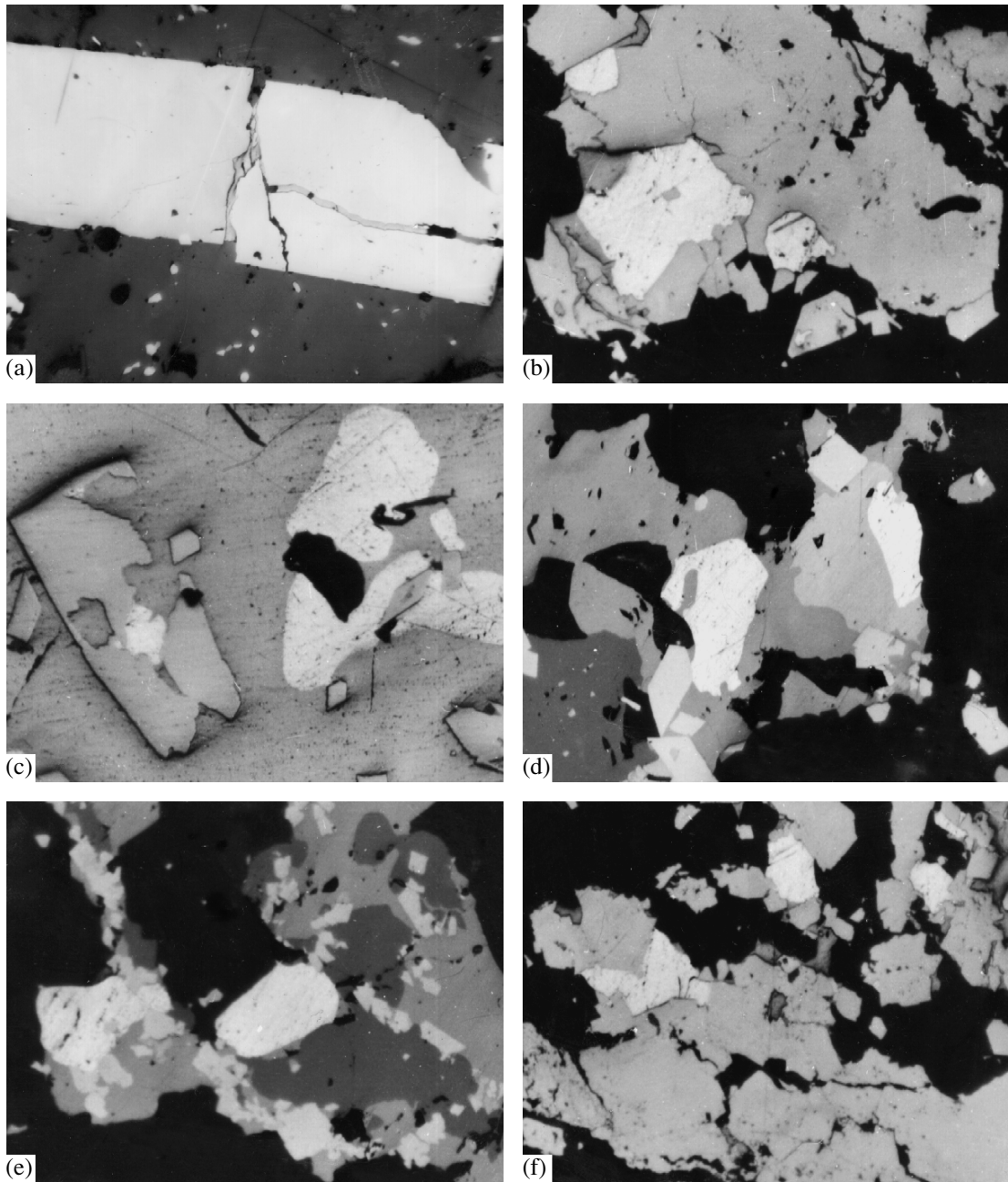


Fig. 3. Intergrowth textures of native gold and coarse-grained arsenopyrite. Reflected light, $\times 200$. (a) Native gold fills microcracks in arsenopyrite from the berezites, gray-quartz; (b) native gold overgrows arsenopyrite, which is crosscut by tiny veils of boulangerite; (c) an association of electrum (white) with galena (gray), early arsenopyrite (coarse relief grains) is replaced by galena and electrum, inclusions of idiomorphic fine-grained arsenopyrite in galena; (d) sphalerite (dark gray), fahlore (gray), and chalcopyrite (light gray), and electrum (white) fill intergranular spaces in quartz (dark), the absence of replacement relationships is evident; (e) sphalerite (dark gray)-fahlore (gray) veinlets with gold grains (white) and arsenopyrite (reliefic) in quartz (dark), arsenopyrite overgrows sphalerite; (f) native gold fills cracks in fine-grained arsenopyrite.

sphalerite aggregates, and replace galena. Metactysts of arsenopyrite and pyrite overgrow the sphalerite aggregates. Fine-grained pyrite and arsenopyrite and the sulfosalt aggregates fill in the same cracks. This indicates a co-precipitation of these minerals. Native gold was co-deposited with sulfosalt minerals in some cases, as the veinlets in arsenopyrite healed with native gold and

the sulfosalt veinlets suggest (Fig. 3b). However, some amount of native gold was deposited later than the fine-grained arsenopyrite (Fig. 3f).

Arsenopyrite is depleted in arsenic in relation to the stoichiometric composition of FeAsS . The arsenic content is 32.6 to 33.3 at. %. It contains a gold admixture of 10 to 35 ppm (average is 22.4 ± 5.79 ppm). An

arsenic admixture is detected in the lead sulfoantimonides. The As content in geocronite and in bournonite varies from 2.3 to 4.3 wt %, and from 1.0 to 6.1 wt %, respectively. The As content in meneghinite and jamesonite is up to 2.8 wt % and 0.4 wt %, respectively. The fineness of native gold from these assemblages is the lowest and ranges from 720 to 480‰.

The youngest paragenetic assemblages (II–IV) are chalcostibite-stibnite ones, because these minerals crosscut the earlier paragenetic assemblages.

A high silver content is a specific feature of the products of the latest stage of mineral formation. The process of regeneration of quartz, silver minerals, and sphalerite widely occurred during this stage (Gamyagin, 1991). The silver and lead sulfoantimonides such as diaphorite, andorite, and owyheeite are found. Owyheeite is widespread, and commonly fills in the cavities in quartz (Fig. 2c). The idiomorphic prismatic and needle inclusions of owyheeite and diaphorite occur within grains of Ag-bearing (up to 30 wt %) tetrahedrite. Owyheeite, diaphorite, and tetrahedrite are presumed to have been co-precipitated.

The admixtures of arsenic and bismuth were found in the lead and silver sulfantimonides. Owyheeite contains up to 1.1 wt % As and 0.6 wt % Bi (Moelo *et al.*, 1984). This mineral contains up to 2.4 wt % Cu, as well. The lead sulfoantimonides found in ores are pyrrargyrite, miargyrite, stephanite, polybasite, and billingsleyite. These minerals comprise variously oriented veinlets in quartz veins. They commonly fill in the cavities in quartz. Pyrrargyrite is the most widespread mineral. The silver sulfoantimonides are associated with boulangerite, owyheeite, and argentian tetrahedrite. Miargyrite, pyrrargyrite, and polybasite contain 0.5, 3.6, and 3.4 wt % As, respectively.

FLUID INCLUSIONS IN MINERALS

Fluid inclusions were studied in quartz from metasomatic rocks, milky-white vein and regenerated quartz, carbonates, and sphalerite. Thermometric measurements were carried out using doubly polished thin sections. Two hundreds samples collected from different ore bodies were examined. Fluid inclusions suitable for the measurements were found in 15 samples. In addition to a microthermometric study, separates of quartz were used to detect the anion and gaseous composition of a fluid.

Description of fluid inclusions. Fluid inclusions were subdivided as primary, pseudosecondary, and secondary. The primary inclusions studied in detail have irregular, rounded, or negative crystal shapes. These inclusions occur along the growth zones of the minerals, or within the grains. The sizes of fluid inclusions in samples studied are very small and range from 5 to 30 μm , mostly about 10–15 μm . The secondary inclusions are those confined to cracks or form the chain-like clusters intersecting the grain boundaries.

Two types of fluid inclusions were recognized and classified according to their phase composition (at room temperature):

Type I was H_2O -rich inclusions containing CO_2 . Two-phase inclusions consisting of liquid water and a vapor phase (which is mainly CO_2 and three-phase inclusions consisting of liquid water), liquid CO_2 , and gaseous CO_2 were found.

Type II was CO_2 -rich inclusions containing mainly a vapor phase. Two-phase inclusions consisting of liquid CO_2 and gaseous CO_2 were observed, as well.

Both types of fluid inclusions form irregular clusters which occur in the same growth zones of quartz and sphalerite grains. This suggests that the H_2O -rich and CO_2 inclusions have been simultaneously trapped. The volumetric proportions of vapor and liquid phases in the type I inclusions vary. A vapor bubble occupies 20 to 40 vol %. Variable liquid and gas phase proportions were observed within single sample. Type I inclusions prevail in most of the samples studied. However, there were samples with abundant inclusions of type II.

Microthermometric measurements. The type I inclusions were completely homogenized into a liquid phase at heating, and CO_2 was homogenized into a liquid phase or (rarely) into a vapor phase. Because of the considerable volume of CO_2 in the inclusions, the inclusions commonly decrepitated before the complete homogenization of the entrapped fluid. The homogenization temperatures vary in the wide range of 360 to 175°C (table). The homogenization temperatures of fluid inclusions in the metasomatic quartz of the II–I assemblage fall in a range 360 to 220°C, respectively. Fluid inclusions in milky-white quartz of the II–II assemblage and in quartz of the II–III assemblage were homogenized in the temperature range of 320 to 230°C, and of 315 to 175°C. The homogenization temperatures of fluid inclusions in the semitransparent quartz of the third stage are considerably higher, and vary from 340 to 260°C. Pseudosecondary fluid inclusions in cleiofane from the cavities in regenerated quartz homogenized at 175°C.

A homogenization of CO_2 in the type II fluid inclusions occurred at the temperatures of 14.2 to 25.5°C. Therefore, the density of CO_2 trapped in the inclusions ranges from 0.83 to 0.68 g/cm³.

The melting temperature of CO_2 in fluid inclusions in the minerals of all assemblages is lower than that of pure CO_2 (–56.6°C). This indicates that other volatiles may partly constitute the vapor phase.

The extremely small sizes of most of the fluid inclusions made it impossible to carry out many of the precise measurements at the freezing of the samples. The melting temperature of clathrate was found to be below 10°C. This demonstrates that the fluid contains soluble chlorine salts (Darling, 1991). A few measurements indicated the insufficient difference between the melting temperatures of clathrate (from +7.8 to +8.4°C) in fluid inclusions, and the minerals of different assem-

Results of Microthermometric Study of Fluid Inclusions in Minerals from the Nezhdaninsk Gold Deposit

Stages, assemblages	Mineral (Type of fluid Inclusion)	Homogenization temperature, °C		Melting temperature, °C		Salinity, wt % equiv NaCl	Density of CO ₂ , g/cm ³
		inclusions	CO ₂	CO ₂	clathrate		
II-I	Quartz (I)	360–220 (75)	+22.0 ... +25.5	–56.6 ... –57.0			0.75–0.83
	" (II)		+14.2 ... +22.6	–56.6 ... –57.3			0.83–0.75
II-II	Quartz (I)	320–230 (58)	+16.2 ... ±0.2	–56.8 ... –57.3	+7.8 ... +8.4	4.3–3.2	0.82
	" (II)		+16.2 ... +17.2				0.82
II-III	Quartz (I)	315–175 (104)	+22.0 ... +23.0	–56.8 ... –57.0	+7.9 ... +8.3	4.2–3.4	0.75–0.74
	" (II)		+25.4 ... +26.9				0.71–0.68
III-I	Sphalerite (I)	340–260 (91)	+17.2 ... ±0.3	–56.7 ... –57.1	+7.6 ... +8.1	4.1–3.8	0.81
	Quartz (I)						

blages. The salt content of the mineral-forming fluid estimated using Erockhin's (1993) equation is 3.2 to 4.3 wt % equiv. NaCl.

Ion and gas analyses. Concentrations of anions and gaseous molecules contained in the fluid inclusions were analyzed using chromatographic techniques (Mironova *et al.*, 1993; Savel'eva *et al.*, 1988). Samples of 20 to 30 g of quartz fragments crushed to –20 to +60 mesh were analyzed.

The highest contents of the Cl[–] ion of 49.3 g/kg H₂O were detected in the fluid inclusions in pegmatite quartz. The Cl[–] ion content in fluid inclusions trapped in gold-bearing quartz varies from 2 to 10.3 g/kg H₂O, or from 0.06 to 0.29 mol/kg H₂O. The fluid inclusions in barren quartz contain the Cl[–] ion concentration varying from 3.6 to 12 g/kg H₂O, or 0.1 to 0.36 mol/kg H₂O. The concentration of sulfate sulfur in fluid inclusions in gold-bearing quartz lies in a range of 0.1 to 5.9 g/kg H₂O, or 0.001 to 0.06 mol/kg H₂O. Fluid inclusions in barren quartz contain 1.6 to 3.5 g/kg H₂O, or 0.006 to 0.017 mol/kg H₂O. The F[–] ion has been detected in the fluid inclusions in a few samples. Its concentration lies within 0.07 and 0.3 g/kg H₂O. The content of NO₂⁺, NO₃[–], PO₄^{3–}, and Br[–] seem to be below the minimum detection limits of ion chromatography.

The volatile component contents in fluid inclusions trapped in gold-bearing quartz range widely. The CO₂ concentration varies from 201 to 1692 g/kg H₂O, or from 4.6 to 38.4 mol/kg H₂O. The CH₄ content lies within 0.19 to 2.60 g/kg H₂O, or 0.01 to 0.16 mol/kg H₂O. The N₂ concentration varies between 0.72 and 41.5 g/kg H₂O, or between 0.02 and 1.48 mol/kg H₂O. The fluid inclusions found in barren quartz and pegmatite quartz are depleted in CO₂ and CH₄. This study showed that the CO₂ content of the fluid, trapped in the fluid inclusions in the minerals, is of 3.3 to 33.5 mol %, the CH₄ content ranges of 0 to 0.6 mol %, and the N₂ content varies from 0.1 to 1.4 mol %. The fluid inclu-

sions found in the pegmatite and barren quartz display higher CH₄ concentrations.

The above results of the fluid inclusion study of the minerals from the Nezhdaninsk ore deposit show that their crystallization occurred in an ore-forming system where two fluids simultaneously circulated. One of the fluids was an essentially aqueous fluid with a low CO₂ content. The second fluid was dominantly gaseous and CO₂-rich. Differences in the vapor and liquid ratios in the H₂O-rich inclusions (type I), and in the CO₂-rich inclusions (type II), testify to the crystallization of minerals from the immiscible fluids, or from the fluids of a different composition simultaneously involved in the ore deposition. The ore-forming fluids contained chlorine, sulfate sulfur, fluorine, methane, and nitrogen.

STABLE ISOTOPES OF SULFUR, OXYGEN, AND CARBON IN MINERALS

Sulfur isotopes. A sulfur isotopic composition of pyrite and marcasite from host sedimentary rocks, accessory sulfides of various magmatic rocks, and sulfides from gold-bearing quartz veins and zones of a disseminated mineralization were studied.

A δ³⁴S value of marcasite from the sedimentary rocks varies from –1.9 to +10.3‰ (Fig. 4a). A sulfur isotopic composition of sulfides from the magmatic rocks fall in a range of –4.0 to +0.7‰ (Fig. 4b). Considerable variations of the δ³⁴S values of the sulfides from the auriferous zones were found (Figs. 4c, 4d). All the values obtained (155 analyses) range from –18.0 to +6.0‰. Most of the samples studied (125 data points) gave the values which fall within the relatively narrow range of –6.0 to +0.7‰ with a mode at 0–1‰. The variations of the δ³⁴S values of the individual minerals are compatible. The δ³⁴S values of pyrite lie between –11.0 and +6.0‰. Sulfur of other sulfides is slightly enriched in a light sulfur isotope. The δ³⁴S values of pyrrhotite, arsenopyrite, sphalerite, and galena lie between –6.7 and –1.6‰, –9.0 and +2.2‰, –6.2 and +0.7‰, and –8.2 and –1.4‰, respectively.

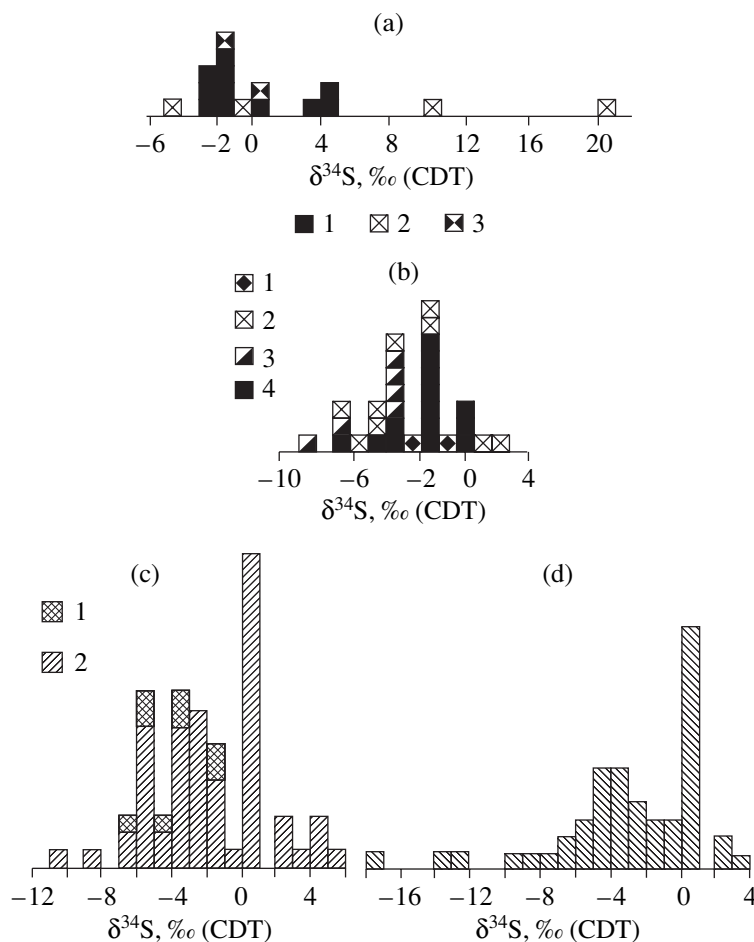


Fig. 4. Sulfur isotopic composition of sulfides from host rocks and intrusive rocks in (a) the Nezhdaninsk deposit and (b, c, d) from gold-bearing zones. (a) (1) Pyrite, (2) marcasite, (3) pyrrhotite; (b) (1) stibnite, (2) sulfosalts, (3) galena, (4) sphalerite; (c) (1) pyrrhotite, (2) pyrite; (d) arsenopyrite.

The data obtained suggests that the sulfur isotopic composition of the younger generation of sulfides is very homogenous and tends to be richer in a ^{34}S isotope. This suggestion is supported by the data on a sulfur isotope composition of minerals of the galena-sphalerite-sulfosalt assemblages, because their $\delta^{34}\text{S}$ values lie between -6.2 and $+0.7\text{‰}$.

Oxygen isotopes. Oxygen isotope compositions of minerals from the magmatic rocks, quartz, and carbonates from the gold-bearing and barren quartz veins and host and altered rocks were studied. One hundred and ten analyses were carried out.

An oxygen isotopic ratio of siltstones collected within the ore deposit lies within $+1.5$ to $+12.3\text{‰}$ (Fig. 5a). These data indicate that metasedimentary rocks are depleted in heavy ^{18}O isotope relative to the typical $\delta^{18}\text{O}$ values reported for shales and argillites, which range between $+13.0$ and $+20.0\text{‰}$ (Sheppard, 1986).

The minerals from the granitoid rocks display the considerable variations of the oxygen isotopic composition (Fig. 5a). The $\delta^{18}\text{O}$ values of potassium feldspars and plagioclase are $+1.1$ to $+11.8\text{‰}$, and $+5.7$ to $+13.1\text{‰}$, respectively. This value of quartz from the granitoid rocks varies from -1 to $+14.2\text{‰}$. Biotite and amphybole, given the $\delta^{18}\text{O}$ values of -6.2 to $+10.7\text{‰}$, and -6.2‰ , are depleted in the ^{18}O isotope. The isotopic composition of these minerals is different from values of $10.0 \pm 2\text{‰}$ found for the typical granite batholithes and granodiorites. Quartz is enriched in a heavy oxygen isotope in relation to other minerals. The values of $\Delta^{18}\text{O}_{\text{quartz-biotite}}$ and $\Delta^{18}\text{O}_{\text{feldspar-biotite}}$ vary from $+3.8$ to $+8.5\text{‰}$, and from $+3.0$ to $+5.9\text{‰}$, respectively. None of the minerals from the granitoids of the Nezhdaninsk ore deposit follow these trends. The isotopic composition of the magmatic rocks changes considerably due to the assimilation and contamination processes, and an interaction with the postmagmatic fluids. Assimilation and the contamination change the bulk composition of a melt. Thus, an isotopic exchange

among the minerals that crystallized from this melt would follow the same trends. The phenomena observed indicates an interaction of the magmatic rocks with a fluid depleted in ^{18}O isotope.

The ratio of oxygen isotopes of the altered rocks (Fig. 5b) differs from that of the metasedimentary rocks. A $\delta^{18}\text{O}$ of value of hofnfels is $+11.7\text{‰}$. A same oxygen isotope composition of $+11.7 \pm 0.2\text{‰}$ has the quartz-sericite metasomatites. The $\delta^{18}\text{O}$ values of the altered rocks of the carbonate facies range from $+2.3$ to $+1.7\text{‰}$. Most of samples are depleted in ^{18}O isotope. This indicates the isotopic exchange of rocks adjacent to the magmatic and hydrothermal ore bodies.

An oxygen isotopic ratio of quartz from the barren veins ranges from $+5.6$ to $+11.3\text{‰}$ (Fig. 5b).

The oxygen isotopic composition of quartz from the gold bearing ore bodies is extremely variable. Values obtained for all samples studied range from $+5.0$ to $+15.2\text{‰}$ (Fig. 5b). The $\delta^{18}\text{O}$ values of quartz from the most important mineralization zones vary within the same ranges. These values of quartz from ore zone 1 are from $+7.1$ to $+12.6\text{‰}$, those from ore zone 8 are from $+6.1$ to $+12.4\text{‰}$, and those from ore zone 56 vary from $+5.0$ to $+13.6\text{‰}$. Quartz from the central and near-surface parts of ore zone 1 is depleted in ^{18}O isotope ($\delta^{18}\text{O} = +10.0$ to $+7.1\text{‰}$), while the $\delta^{18}\text{O}$ values of quartz on the flanks and at the depth are higher than $+11.5\text{‰}$.

An oxygen isotopic composition of carbonates lies within the $\delta^{18}\text{O}$ values of $+10.1$ to $+22.5\text{‰}$ (Fig. 5b). The most significant differences (from $+10.1$ to $+22.5\text{‰}$) of the $\delta^{18}\text{O}$ values of carbonates from ore zone 1 were revealed. The carbonates from ore zone 56 have the relatively narrow isotopic variations of $+17.0$ to $+21.2\text{‰}$.

The frequency diagrams show two modes of the $\delta^{18}\text{O}$ values of quartz and carbonates, which are from $+8.0$ to $+9.0\text{‰}$ and $+12.0$ to $+13.0\text{‰}$, and $+15.0$ to $+16.0\text{‰}$ and $+20$ to $+21\text{‰}$, respectively.

The oxygen isotopic composition of the minerals from the mesothermal gold deposits was studied by many researches (Kerrick, 1990₂; Nesbitt, 1991; Goldfarb *et al.*, 1991). The narrow ranges of the $\delta^{18}\text{O}$ values of minerals were found in most deposits. The variations of the oxygen isotope ratio did not exceed 2.0 to 3.0‰ within a single deposit or, commonly, within a gold-bearing belt. The uniformity of the $\delta^{18}\text{O}$ values provides evidence for the formation of the mesothermal deposits as a result of large-scale hydrothermal activity within large territories. It is obvious that the Nezhdaninsk deposit, where the considerable variations of the oxygen isotopic composition have been revealed, differs from the previously described mesothermal deposits.

Carbon isotopes. The carbon isotopic composition of carbonates displays a considerable range of -14.7 to $+1.0\text{‰}$ (Fig. 6a). However, the majority of analyses (13 of 19) fall in the relatively narrow range of $\delta^{13}\text{C}$ values of -5.0 to -8.6‰ . The median value of this group is

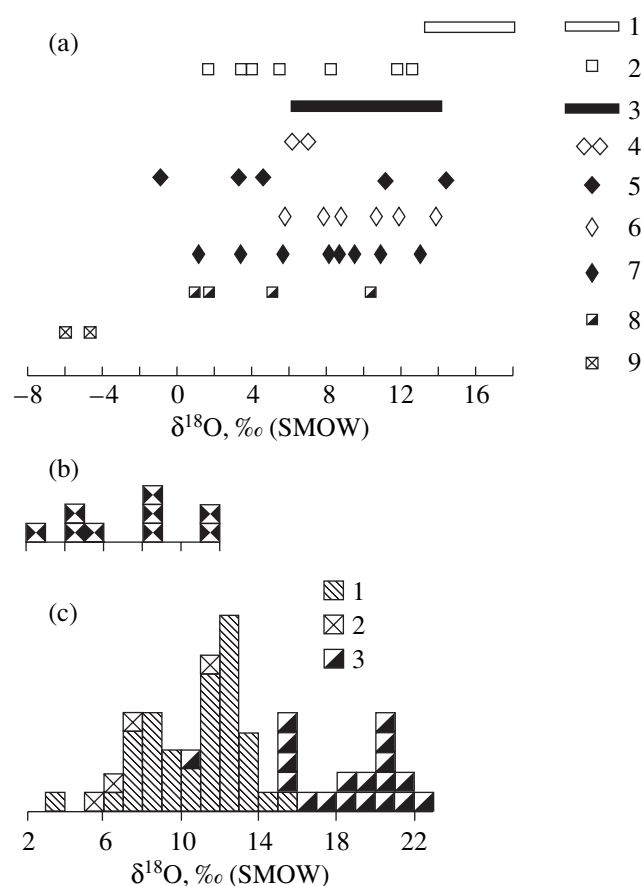


Fig. 5. Oxygen isotopic composition of rocks and minerals of the Nezhdaninsk ore deposit. (a) Host rocks and granite (bulk composition) and minerals of intrusive rocks: (1) slates, (2) siltstones, (3) granites, (4) garnet, (5) quartz, (6) plagioclase, (7) potassium feldspar, (8) biotite, (9) amphibole; (b) altered rocks (bulk composition); (c) minerals of quartz veins and mineralized zones: (1) quartz of gold veins, (2) barren quartz, (3) carbonates.

-7.5‰ . The second group of a few analyses (3) includes those ranging in the narrow field of -11.0 to -11.8‰ .

Two groups of carbonate minerals may be distinguished according to the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values. The first group includes the analyses of ankerite from the gold-bearing veins that show the linear relationships between the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values. The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values vary from -14.7 to $+1.0\text{‰}$, and from $+10.1$ to $+20.1\text{‰}$, respectively. The second group includes the analyses of ankerite from the altered rocks and magnesite, which have the relatively uniform $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values equal to $-8.7 \pm 2.5\text{‰}$ and $+20.9 \pm 1.2\text{‰}$, respectively.

The ratio of the carbon isotopes of carbonates from the mesothermal gold deposits located in the weakly metamorphosed siltstones, shales, graywackes, and sandstones was published in several papers. The considerable variations of the $\delta^{13}\text{C}$ values of -10.0 to

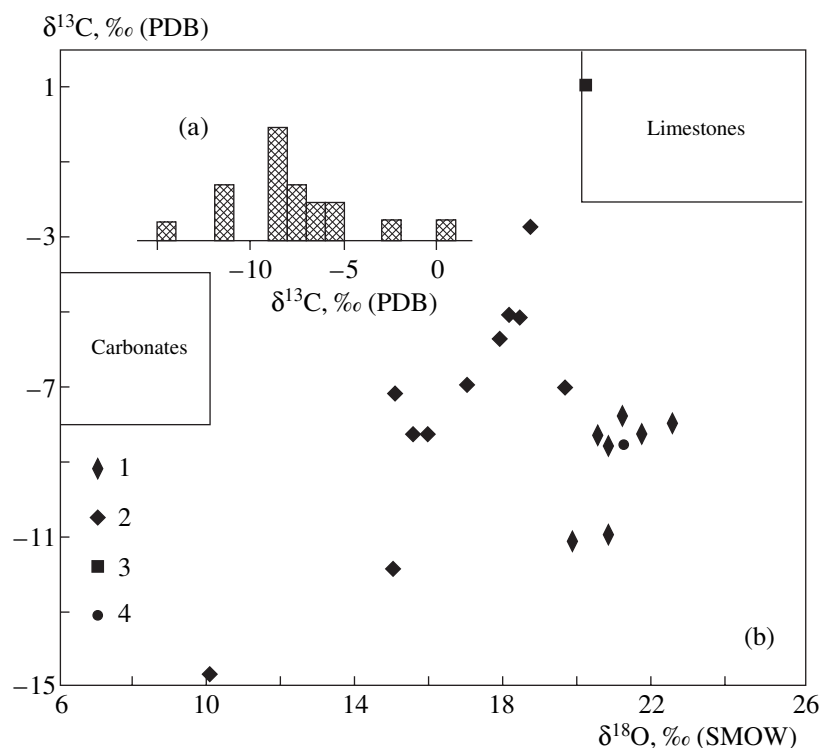


Fig. 6. Carbon and oxygen isotopic compositions of carbonates from the Nezhdaninsk deposit. (a) Frequency diagram of $\delta^{13}\text{C}$; (b) diagram of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values: (1) ankerite from metasomatites, (2) vein ankerite, (3) calcite, (4) magnesite.

+2.0‰ were found in the deposits of Canadian Cordilleras (Nesbitt, 1991). The $\delta^{13}\text{C}$ values of carbonates associated with gold in the Archaean mesothermal gold deposits range from -10.1 to +2.2‰ (Kerrick, 1990). Thus, the carbon isotopic composition of carbonates of the Nezhdaninsk gold deposit is similar to those found in other mesothermal deposits.

CONDITIONS OF THE DEPOSIT FORMATION

The above-mentioned data on the mineral composition of ores, intergrowths and chemistry of minerals, ratios of the stable isotopes, fluid inclusions, as well as thermodynamic calculations, were used to determine the physicochemical parameters of the ore-forming processes, and the composition and nature of a mineral-forming fluid.

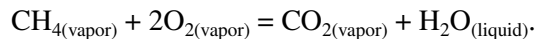
Temperature and pressure. The temperature of ore formation was estimated using data on fluid inclusions, arsenopyrite, and sphalerite–galena isotope geothermometers. As the fluid inclusions were entrapped whenever fluid immiscibility occurred, the corrections to the temperatures for pressure can be ignored. The homogenization temperatures of fluid inclusions are considered to be the true temperatures of mineral formation. Therefore, the ores were formed at 175 to 360°C. It should be noted that the mineral aggregates which are considered to be the paragenetic assemblages

of different ages were formed as a rule, from 220 to 360°C. The arsenopyrite geothermometers gave the higher values of 400 to 460°C (Bortnikov, 1993). Some of the isotopic temperatures (150 to 250°C) characterize the conditions of mineral formation, as well.

Pressure was calculated on the basis of the fluid inclusion study (Brown and Lamb, 1986). The values obtained are from 1.2 to 1.7 kbar. The presence of water in the vapor-dominated inclusion increases the fluid density. This shifts the isochore to the field of lower pressures. The errors for the calculated position of an isochore also proceed from the presence of CH_4 and N_2 in the vapor phase. The value of 1.2 kbar is considered to be the minimal possible figure, as such an internal pressure is needed to cause the decrepitation of the fluid inclusions of 12–13 μm in size (Leroy, 1979).

Volatile component behavior. The sulfur fugacity was estimated taking into account that the assemblage of pyrite and arsenopyrite is widespread in the deposit ores. The f_{S_2} value at the temperature range of 350 to 250°C, and pressure of 1 to 2 kbar, may vary between 10^{-8} and 10^{-12} bar. Evidence for this conclusion is the low iron content in sphalerite. The presence of pyrrhotite indicates that the sulfur fugacity could be decreased during some episodes of the ore formation.

The fugacities of oxygen, methane, and carbon dioxide are calculated at the temperatures of 250 to 350°C using the reaction:



The volatile component concentrations taken for calculations are based on this study of fluid inclusions. These parameters are $\sim 10^{-30.5}$ – $10^{-37.5}$ bar, $10^{2.56}$ to $10^{3.3}$ bar, and $10^{0.46}$ – $10^{1.7}$ bar, respectively. The parental mineral fluid was highly oxidized. Its reduction resulted from the interaction with the host rocks containing organic matter.

pH of fluid. The pH value of fluid is estimated from the assumption that it is buffered by its reactions with wall rocks. Metasomatic rocks that contain quartz + sericite + pyrite + ankerite are believed to be formed at 250–400°C and 0.5–3 kbar in equilibrium with fluid that have pH = 5.5–6.1 (Shcherban' *et al.*, 1990).

The ion- and gas chromatographic analyses show that the mineral-forming fluid contained about 40 mol % CO₂, 6.0 mol % N₂, and 1.5 mol % CH₄. The highest values were obtained for samples that contained the vapor-dominated fluid inclusions. Thus, a mole fraction of CO₂ in the fluid equal to 0.25–0.3 seems to be a reasonable value. The chlorine salt concentration did not exceed 4.5 wt % equiv. NaCl.

The S, O, and C isotope composition of mineral-forming fluids. The ratio of stable isotopes in the ore-forming fluid was calculated using conventional approaches and the known equations of isotope fractionation in the hydrothermal systems (Field and Fifarek, 1985; Ohmoto, 1986; Sheppard, 1986). The oxygen isotopic composition was calculated in the quartz–water and calcite–water systems. These calculations imply that the isotopic equilibrium between a fluid and precipitated minerals was attained at 300°C and preserved after mineral deposition. An equation of $\delta^{18}\text{O}_{\text{quartz}} - \delta^{18}\text{O}_{\text{H}_2\text{O}} = 3.34(10^6/T^2) - 3.31$ yields the

$\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values ranging from –1.0 to +8‰. Values of $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ equal to +2‰ and +6.0‰ correspond to the common analytical values of $\delta^{18}\text{O}_{\text{quartz}}$. An equation of $\delta^{18}\text{O}_{\text{calcite}} - \delta^{18}\text{O}_{\text{H}_2\text{O}} = 2.78(10^6/T^2) - 2.89$ gave the $\delta^{18}\text{O}$ values varying between +4.5‰ and +17.0‰. The most typical $\delta^{18}\text{O}$ values obtained for carbonate minerals correspond to the $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values equal to +10.4‰ and +11.4‰.

The values calculated using the various equations are different. The reason is as follows: quartz precipitate was equilibrated with the water of a hydrothermal fluid, while crystallized carbonates attained isotopic equilibrium with CO₂. The $\delta^{18}\text{O}_{\text{CO}_2}$ values are about 13‰ higher the $\delta^{18}\text{O}$ values of water in equilibrium. Therefore, the carbonate precipitate was enriched in the

^{18}O isotope in relation to quartz deposited from the same fluid.

The Nezhdaninsk deposit is different from many mesothermal ore deposits, for it was formed by fluid considerably depleted in ^{18}O isotope. The common $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ value of a fluid deposited in the gold ores, hosted by the metasedimentary rocks ranges from 5.0 to 10.0‰ (Nesbitt, 1991) or from 7.0 to 13.0‰ (Goldfarb *et al.*, 1991).

Carbon isotope composition of a mineral-forming fluid. CO₂ escape, as a result of the boiling or the loss of volatiles, causes a considerable change of the $\delta^{13}\text{C}$ of a residual solution, if the initial CO₂ concentration in the fluid was high (Kerrick, 1990). The fractionation between gaseous CO₂ and the soluble carbon species in a fluid is equal to –1.1 to +0.7‰ at 300°C. Therefore, we may assume that the $\delta^{13}\text{C}$ of the gaseous CO₂ that escaped from the fluid would be close to the carbon isotopic ratio of their ions and molecules soluble in a hydrothermal fluid. Methane is depleted in the heavy ^{13}C isotope by 25.0 to 29.0‰ relative to CO₂ at 300 and 250°C, respectively. The CH₄ content in the fluid is about 1 mol %. The CH₄ escapes to the vapor phase at 300°C and changes the $\delta^{13}\text{C}$ value of the residual fluid by +2.0‰. Assuming that the deposition of carbonates occurred in the field predominated by oxidized ions and

molecules of carbon, the $\delta^{13}\text{C}_{\text{H}_2\text{CO}_3}$ value equal to the $\delta^{13}\text{C}$ of a fluid may be used. Therefore, ankerite precipitated in equilibrium with a fluid at 300°C had to differ by 1.0‰ relative to the $\delta^{13}\text{C}$ of a hydrothermal solution. It seems to be evident that before fractionation, the initial fluid differed by about 3‰, relative to that of carbonates. The $\delta^{13}\text{C}_{\text{fluid}}$ values of the fluid that entered the site of the ore formation before the escape of the CO₂ + CH₄ mixture were between –17.0 to –2.0‰ (average value was about –10.0‰).

Values of $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ of fluid. The hydrogen sulfide was a dominated sulfur species soluble in a fluid as the ores were deposited under the sulfur and oxygen activities that were less than the values in equilibrium with the pyrite–pyrrhotite–magnetite triple point. The relation of the $\delta^{34}\text{S}_{\text{SS}}$ value of fluid $\sim \delta^{34}\text{S}_{\text{H}_2\text{S}}$ is true in such a case. Assuming an isotopic equilibrium between H₂S and sulfides, the enrichment of pyrite equilibrated with a fluid at 300°C in heavier isotope by 1.2‰ occurred. Thus, the $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ value of fluids at a site of deposition ranged from –12.0 to +5.0‰, and, in general, from 0 to –7.0‰.

Sources of constituents of ore-forming fluid. A special feature of the Nezhdaninsk ore deposit is the occurrence of magmatic and metasedimentary rocks depleted in a ^{18}O isotope. Such rocks could originate as a result of the crystallization of the magmatic rocks

with low content of the ^{18}O isotope, or the denudation of previously formed sedimentary rocks with low ^{18}O isotope content. The enrichment of the magmatic or metamorphosed rocks in a light oxygen isotopic occurs at the interaction and the isotope exchange of the rocks with fluids containing a low concentration of the ^{18}O isotope (Sidorenko and Bortschevskii, 1975). The hypothesis of the crystallization of the intrusive rocks from magma with low content of the ^{18}O isotope should be ruled out, because the relation of $\Delta_{\text{quartz-plagioclase}}$ and $\Delta_{\text{quartz-K-fsp}}$ differs from the values (1.4 to 2.2‰) typical for the magmatic minerals. The data observed indicate that the isotopic exchange between the minerals of magmatic rocks and the fluid of a different isotopic composition occurred. Moreover, the isotopic equilibrium between the minerals was not attained at the recrystallization of the rocks, because oxygen isotopes fractionated between potassium feldspar and plagioclase. This phenomena was not found in an experimental study at 500°C.

The fluid which changed the composition of metasedimentary rocks of the Nezhdaninsk ore could not originate as a result of the dehydration reactions. A fluid formed in such a way has the $\delta^{18}\text{O} = +4.0$ to $+15.0$ ‰. The interaction of this fluid with metapelitic rocks with $\delta^{18}\text{O}$ equal $+14.5$ ‰ causes a decrease of their $\delta^{18}\text{O}$ to $+12.5$ ‰ (Wickman and Taylor, 1985). The volume of water released at the dehydration of pelitic rocks is too small to cause such an isotopic shift.

The isotopic oxygen exchange of the heated sea water with the $\delta^{18}\text{O}$ equal to -1.0 ‰, and metapelites at the water-to-rock ratio of 0.3 to 0.5, shifts the $\delta^{18}\text{O}$ of the parent sea water and the metasedimentary rocks to 4.0 – 6.0 ‰ and about $+12.0$ ‰, respectively. This value is higher than those obtained for the metasedimentary rocks of the Nezhdaninsk ore deposit. Thus, only the interaction with the meteoric water could considerably modify the oxygen isotopic composition of the metasedimentary rocks. The mass balance during the infiltration of the heated meteoric water due to convection has been calculated to verify this assumption. The following fractionation coefficients of $1000\ln\alpha = 2.05(10^6/T^2) - 3.85$ and $1000\ln\alpha = 2.38(10^6/T^2) - 3.89$ were used for the calculations in the kaolinite–water system at 150 to 300°C and the muscovite–water system at 350 to 400°C, respectively.

The isotopic composition of a rock equilibrated at 200 to 250°C and 300 to 350°C with a fluid that originated from the meteoric water, which had the $\delta^{18}\text{O}$ of -15 ‰, falls between $+3.0$ to $+6.6$ ‰ and $+9.5$ and $+11.6$ ‰, respectively. Thus, the $\delta^{18}\text{O}$ values of the sedimentary rocks of the ore deposit indicate their modification due to the interaction with a large volume of the isotopically light fluids generated from the meteoric waters. The emplacement of large magmatic bodies is assumed to cause the formation of a convective cell that involved the ground waters. The ground waters were heated, and interacted with the host rocks. These stim-

ulated the exchange and incorporation of the constituents of the rocks, possibly, oxygen, sulfur, carbon, and methane. This process modified the ground waters in the mineral-forming fluid, which penetrated the fault zones and the site of ore deposition.

Data on the isotopic composition of the minerals suggests the isotope heterogeneity of the mineral-forming fluids. The calculated values of $\delta^{18}\text{O}_{\text{H}_2\text{O}}$, $\delta^{13}\text{C}_{\text{H}_2\text{CO}_3}$, and $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ are evidence for the mixing of the fluids of various isotopic ratio. This assumption proceeds from the studies of mineral parageneses and fluid inclusions that show that changes in the physico-chemical parameters could not considerably affect the isotopic ratio of the fluids and the minerals. The mixing of fluids took place in, or near, the site of the ore deposition, since no homogenization of the isotopic composition of sulfur, oxygen, and carbon occurred there.

The oxygen isotope composition of the mineral-forming fluid is depleted in the $\delta^{18}\text{O}$ relative to the primary magmatic waters ($\delta^{18}\text{O}$ is of $+5.5$ to 9.5 ‰) and lies in a range of values from $+3.0$ to $+14.0$ ‰, typical of the metamorphic waters (Sheppard, 1986).

Carbon can ascend from the deep sources, or be leached from sedimentary rocks, during regional metamorphism or at the interaction with fluids. The host rocks contain the organic matter and the marine carbonates as cement. The value of $\delta^{13}\text{C}_{\text{org}}$ lies within -10 to -35 ‰. CO_2 generated at the reduction and oxidation reactions, and the hydrolysis of the organic carbon, has $\delta^{13}\text{C}$ values less than -10.0 ‰, which are close to the isotopic composition of the organic carbon. The dissolution of carbonates, or the decarbonatization reactions, produce CO_2 with $\delta^{13}\text{C}$ of $+4.0$ to -8.0 ‰. Carbon with $\delta^{13}\text{C}$ ranging from -8.0 to -5.0 ‰ can be introduced from the granitoid magmatic chamber (Kyser, 1987). Mantle CO_2 possesses values close to -5.0 ‰.

The values of $\delta^{13}\text{C}_{\Sigma\text{C}}$ of the fluids indicate the influx of carbon from various reservoirs in the site of ore deposition. The isotopic composition of carbonates ranges generally from -5.0 to -9.0 ‰ and is evidence for the involvement of carbon from the deep sources in the process. The variations of $\delta^{13}\text{C}_{\Sigma\text{C}}$ from $+1.0$ to -15.0 ‰ suggest the incorporation of organic or marine carbon in the mineral formation.

The deep reservoirs and sulfides from the host rocks may be considered as possible sulfur sources. The $\delta^{34}\text{S}$ values typical of siltstones are of -30.0 to $+10.0$ ‰ (Nielsen, 1979). The $\delta^{34}\text{S}$ values of the hydrogen sulfide formed at the hydrolysis of marcasite lie in the range of -1.0 to $+13.0$ ‰. A fluid genetically related to the granite magma possesses the $\delta^{34}\text{S}$ values of -3.0 to $+7.0$ ‰. Data on the isotope compositions of the sulfides from the Nezhdaninsk deposit show that the sulfur remobilized from the host rocks, and was introduced from the magma participating in the mineral formation.

MODEL OF THE DEPOSIT FORMATION

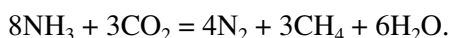
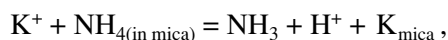
The model of formation proposed for the Nezhdaninsk ore deposit is based on the following general observations: (1) the deposit was formed as a result of the complicated and multistage geodynamic history of the region; (2) the deposit is located in a junction zone of several abyssal faults; (3) the gold mineralization is spatially and temporally associated with an emplacement of the silicic and intermediate intrusive rocks; (4) no rocks metamorphosed to amphibolite grade have been found in the district. The latter feature makes the Nezhdaninsk deposit different from the main giant mesothermal deposit. Most of the mesothermal deposits were formed as a result of the influx of a fluid that was a mixture of water, CO_2 , CH_4 , H_2S , and soluble chlorine salts, in a site of ore deposition. The CO_2 molar fraction of the fluid was up to 0.25. Its salinity was less than 4.5 wt % equiv NaCl. Such a fluid was equilibrated with rocks metamorphosed to the amphibolite grade. This feature of the fluid is the most important evidence for its generation due to dehydration and decarbonatization of the country rocks. As the metamorphosed rocks of the amphibolite facies are absent in the district, the genetic relation of the Nezhdaninsk deposit to the regional metamorphism is problematic.

In our opinion, the Nezhdaninsk deposit was formed during several stages. A generation of the Nezhdaninsk-Geldinsk ore-forming system was linked with the latest accretion event of the region. The formation of the magmatic chambers, due to the partial melting of the rocks of the lower crust in the subduction zones, and the greenstone metamorphism of the subducted sedimentary sequences, occurred during this event. A metamorphic transformation of the rocks caused their dehydration and decarbonatization. A mobilization of ore components from the rocks at their interaction with the metamorphic $\text{H}_2\text{O}+\text{CO}_2$ fluid accompanied these processes. As a result, the low-grade disseminated and stringer ore were formed in the Nezhdaninsk deposit. Economic ores of this type are known in the Allakh-Yun'sk zone (Buryak *et al.*, 1990).

The next stage in the history of the Nezhdaninsk-Geldinsk system was related to the postaccretion tectonomagmatic activation of the Early to Late Cretaceous ages. The Okhotsk subduction occurred in the Late Albian. A partial melting of moderately aluminous rocks of a contrast composition occurred at the depth of 30 to 35 km. The melts were removed along the tectonic zones. The composition of the accessory garnets shows that a magma of the granodiorite composition ascended up to the depths of 15 to 16 km (pressure from 5 to 6 kbar). The intermediate magma chambers were formed at these depths. The granodiorite magma was generated as a result of the intrachamber differentiation. The Kurumsk massif is a product of this process. The K-Ar and Rb-Sr ages ranging from 94 to 124 Ma indicate that the rocks of the Kurumsk massif and the Geldinsk stocks are younger by 35 Ma than the rocks of

the batholithe series. Their injection was associated with the accretion of the Okhotsk terrain to the Siberian craton, which completed in the Neokomian. A hypothesis on the crystallization of the Kurum massif rocks, under pressure of 1 kbar and at 760 to 770°C, and of the dioritic rocks of the Geldy stocks at 870°C, was proposed.

The remelting of early cumulates of the granodiorite magma at the lower parts of the intermediate chambers led to the generation of lamprophyre magma. Its derivatives are the early dikes that intruded after a crystallization of the rocks of the Kurumsk massif. The lamprophyre rocks and the Kurum massif are of nearly the same age. The gold quartz mineralization is spatially associated with the lamprophyre rocks. The age of the ores is estimated to range from 105 to 124 Ma. The pure soluble volatiles (H_2O , CO_2 , H_2S , HCl , and HF) were exsolved during the crystallization of magma, and the ore components from the magma were extracted. The fluid migrated along the faults, interacted with host rocks, and exsolved and incorporated their oxygen, carbon, and sulfur. In addition, an emplacement of granite led to the formation of hornfelses. The weakly metamorphosed siltstones and sandstones were dehydrated and decarbonatized. An additional volume of the fluid was generated and involved in the system. An emplacement of magmatic bodies and metamorphism created the thermal gradient, which caused a convection of the meteoric waters. The influx of a huge volume of meteoric waters, and their interaction with host sedimentary and magmatic rocks, caused considerable enrichment of these rocks in a light oxygen isotope. A reasonable assumption is that huge masses of these rocks were transformed under the influence of the heated meteoric waters during the thermal metamorphism. The ore components were mobilized from the rocks, and the meteoric and metamorphic waters were modified into a mineral-forming fluid. The evidence for an intensive exchange of the mineral-forming fluid with host rocks is the presence of noticeable concentrations of CH_4 and N_2 (up to 2 mol %), and an obvious prevalence of N_2 over CH_4 . The black shales contain up to 1100 ppm of NH_4^+ , which is mainly concentrated in micas (Bottrell and Miller, 1990). A release of the NH_4^+ -ions followed its oxidation to N_2 at the interaction with a fluid. This process was accompanied by a reduction of CO_2 to methane:



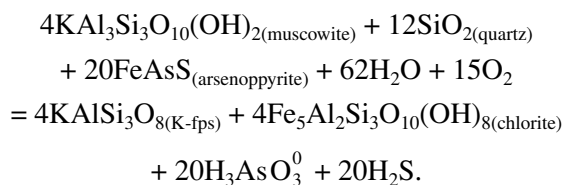
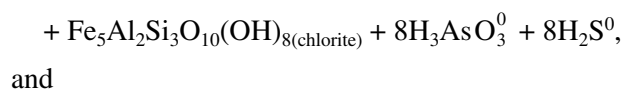
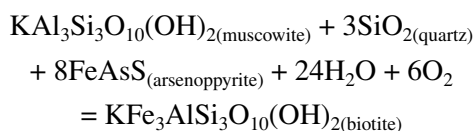
The mixing of all the fluids in the mineral-forming system took place near a zone of their discharge. This assumption is supported by the considerable variation of the isotopic composition of the minerals that explains the inhomogeneity of the mineral-forming system. A large gold deposit such as the Nezhdaninsk deposit may form as a result of the activity of a large-scale hydrothermal system. The localization of the

deposit within a zone of active magmatism, regional and contact metamorphism, and at the junction of regional faults, which were the feeder channels for melts and huge volumes of mineral-forming fluids, was an essential factor for the participation of large amounts of gold in the deposit formation.

The mineral-forming fluid was a mixture of $\text{H}_2\text{O} + \text{CO}_2 + \text{N}_2 + \text{H}_2\text{S} + \text{CH}_4$ with a low content of soluble chlorine salts (up to 4.5 wt % equiv NaCl). When it ascended to the depth of about 4–5 km, fluid immiscibility occurred as a result of the pressure drop and, possibly, a temperature decrease. Two fluids appeared as a result. The first was a liquid water with soluble CO_2 and chlorine salts, and the second was a vapor-dominated fluid consisting mainly of $\text{CO}_2 + \text{N}_2 + \text{H}_2\text{S} + \text{CH}_4$. The escape of the vapor phase dramatically changed the chemical composition, the oxidation state, and the pH values of the residual fluid (Bowers, 1991). The loss of CO_2 , H_2S , H_2 , and CH_4 led to an increase of the oxygen activity in the residual fluid, and the escape of CO_2 decreased the pH value, as well. The chlorine activity in the residual fluid increased, as well. Finally, a radical loss of the vapor phase caused a loss of heat in the mineral-forming system and, as a result, the residual solution cooled.

Most of researches believe that just this process is responsible for the deposition of the minerals and gold. It changes the isotopic composition of the fluid by increasing a fraction of the isotopically light oxygen and sulfur. We assume that fluid immiscibility in the mineral-forming system of the Nezhdaninsk gold deposit caused the fluid interaction with host rocks, because of disequilibrium between the fluid and host rocks. This led to the metasomatic alteration of host rocks and the formation of wide halos of berezites with disseminated sulfides.

The reactions are as follows:



These reactions resulted in the formation of arsenopyrite at the sulfidization of Fe^{2+} contained in silicates of host rocks. The deposition of large masses of sulfide minerals such as arsenopyrite and pyrite caused a radical decrease of the H_2S concentration in the mineral-forming solution. This phenomenon provoked instability in a bisulfide complex of gold. Gold was incorpo-

rated in arsenopyrite and pyrite as a solid solution. An absence of coeval native gold and arsenopyrite in the altered host rocks suggests that the mineral-forming fluid was undersaturated with gold.

Late native gold was deposited on surfaces of sulfide grains. The gold crystallization resulted from the electrochemical reactions that occurred at the contacts of sulfide grains and a solution. The importance of the electrochemical reactions were demonstrated in experimental studies (Sakharova *et al.*, 1975; Müller and Kerstein, 1994). Another mechanism which causes gold deposition is an oxidation of the sulfides accompanied with an appearance of polysulfide ions on their surface (Hayland and Bancroft, 1989). Microscopic studies revealed that native gold often overgrows the arsenopyrite and sphalerite grains, or fills cracks in these minerals. However, no evidence for the dissolution or replacement of these minerals by native gold was observed. In addition, native gold occurs as isolated grains in quartz, and fills the intergranular spaces or cracks. These observations suggest that, in addition to the electrochemical reactions, other mechanisms could be responsible for the gold precipitation from a solution. The instability of a bisulfide gold complex most probably results from a decrease of the H_2S concentration in a solution. This decrease occurred due to the deposition of a large amount of sulfide minerals according to the above reactions. Therefore, the most important mechanism responsible for the formation of large gold lodes is assumed to be the reactions between a fluid and host rocks. Fluid immiscibility triggered these reactions.

The next stage of the formation of the Nezhdaninsk gold deposit was linked with an emplacement of the Geldinsk stocks. They crystallized after the intrusion of the Kurumsk massif and dikes, which were the derivatives of the granodiorite magma chamber. We assume that an influx of the fluids enriched in silver and lead was related to this magmatic event. These fluids influenced the gold and produced quartz ores, the redeposition of quartz, and some sulfides.

Therefore, our investigations do not support hypotheses on only the metamorphogenic, meteoric, or magmatic nature of the fluids responsible for the formation of mesothermal gold deposits. The fluid which formed such a deposit was of a multiple origin and consisted of magmatic, metamorphic, and meteoric components. Variations of the chemical and isotopic composition of the mineral-forming fluid indicate the mixing of these components close to, or in, a site of ore deposition (Bortnikov *et al.*, 1996). The magmatic fluid dominated. Our model suggests that the mesothermal deposit formation is in close relation with the magmatic activity. This model includes a variety of mechanisms of fluid generation, such as direct input of fluids from the magma chamber, and mobilization of components during the dehydration and decarbonatization of host rocks due to the contact or contact-regional metamorphism.

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