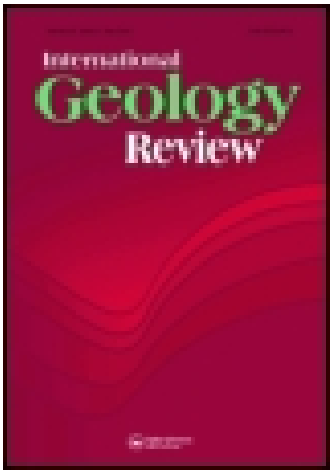

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Abstract

The Natalka lode gold deposit, also known as the Matrosov mine, is located in the Magadan region of northeastern Russia at 61° 39′ N, 147° 50′ E. The deposit was discovered in 1943 and production started in 1945. The mine has produced more than 75 metric tons of gold, with an average grade 4 g/metric ton (mt), and has reserves of about 450 mt.

The Natalka deposit occurs along the southwestern flank of the Yana-Kolyma metallogenic belt and is confined to the major, NW-trending Tenka fault. The deposit is hosted by Upper Permian carbonaceous sediments, subjected to greenschist metamorphism. The ore zones occur along a Z-shaped, strike-slip fault zone that extends for about 12 to 13 km. In plan view, the ore zones are about 5 km long and 100 to 200 m wide in the northwest portion, 350 to 400 m wide in the central portion, and 600 m wide in the southeast portion of the deposit.

The main ore minerals are arsenopyrite and pyrite, which comprise about 95% of the sulfides, along with subordinate pyrrhotite, Co-Ni sulfarsenides, sphalerite, chalcopyrite, galena, native gold, ilmenite, and rutile. Scheelite, tetrahedrite, bournonite, boulangerite, and stibnite occur locally. The major gangue mineral is quartz, with subordinate carbonates, feldspars, chlorite, sericite, kaolinite, montmorillonite, and barite. The total sulfide content of the ore zones ranges from 1 to 3%, and in places up to 5%. Native gold occurs as large individual grains ranging from 0.1 to 2.0 mm in diameter, or as fine disseminations in arsenopyrite. The average gold fineness is 750 to 790.

Fluid inclusion studies reveal homogenization temperatures of 150° to 360° C, with mainly liquid and as much as 5% vapor. Two temperature peaks of 280° to 320° C and 180° to 240° C occur in many samples. The $\delta^{34}S$ composition of sulfides in orebodies ranges from -6.3 to -2.4 per mil and approximates that of sedimentary rock-hosted pyrite. The $\delta^{34}S$ values of the ore solutions are interpreted as having been close to that of the sulfide minerals. The $\delta^{18}O$ composition of ore quartz ranges from 13.9 to 14.1 per mil. The calculated $\delta^{18}O$ composition for the ore fluid ranges from 7.1 to 7.3 per mil at 300° C. The $\delta^{18}O$ values of oxygen indicate a quite homogeneous fluid of metamorphic origin.

The sulfur, arsenic, and gold in the ore deposit were mobilized during metamorphism that included transformation of pyrite to pyrrhotite. The PT conditions for this reaction are estimated at about 400° C and 2.5 kbar, approximately at the biotite isograd. Associated decarbonatization and dehydration reactions produced much of the ore fluid. The interaction of ore-fluid sulfur with Fe-bearing silicate and oxide minerals probably caused deposition of sulfide minerals and gold.

Introduction

THE NATALKA LODE GOLD deposit is located approximately 390 km northwest of Magadan, at 61° 39′ N. Lat. and 147° 50′ E. Long. It was discovered in 1943 by E. P. Mashko during exploration for bedrock sources of placer gold deposits that initially were mined in the period from 1939 to 1941. In 1945, the Natalka deposit began production. The nearby Omchak and Pavlik lode gold deposits were discovered at the same time. These three lode deposits, together with gold placers in streams that cut through the orebodies, comprise the Omchak mining district.

The discovery of the Natalka deposit and initiation of lode gold mining occurred simul-

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TABLE 1. Past Production of Major Placer Gold Deposits of the Omchak Mining District

Placer deposit	Past gold production (metric tons)
Omchak River	97.0
Geologicheskiy Creek	1.0
Natalka Creek	32.1
Glukhar Creek	9.0
Vanin Creek	1.5
Pavlik Creek	20.0

taneously with a decrease in placer gold production throughout northeastern Russia; the decline in placer output began during World War II and continued through the late 1950s. Prior to this time, the Yana-Kolyma metallogenic belt was a major placer gold province. From 1929 to the end of the war, about 700 metric tons of gold were produced from rich placer districts. Placer gold mining in the Omchak mining district, for example, has produced more than 160 metric tons of gold (Table 1).

Lode gold deposits in the region were developed to compensate for a decrease in placer gold production. Beginning in the 1940s, more than ten lode gold deposits—including Natalka, Igumen, Rodion, Utinka, Degdekan, Yugler, Maldyak, and Shturm—were developed in a short time (Table 2). The total production from these deposits, excluding Natalka, was about 30 metric tons of gold. The Natalka deposit, now continuously mined for 50 years, has produced more than 75 metric tons of gold and 22 metric tons of silver. Reserves at Natalka are 450 metric tons of gold and 130 metric tons of silver.[1] Including the Pavlik and Omchak lode deposits, reserves for the mineral district total more than 520 metric tons of gold.

The Natalka deposit is mined mainly by underground workings, totalling more than 350 km in length. Mining now is occurring at the 600-m level, at an elevation of 1000 to 1100 m above sea level. Drilling has intersected ore-grade mineralization at the 450-m level, and one deep drill hole intersected ore-grade mineralization at the 50-m level. Recently, a small test open-pit was started above part of the underground workings. Gold is recovered at a processing plant located 6 km east of the mine. The gold recovery procedure includes gravity concentration, gravity concentrate amalgamation, flotation, gravitation of flotation tailings, and sorption leaching of flotation concentrate and amalgamation tailings. Recovery rates range upward to 75% of contained gold, with losses resulting from fine-grained gold hosted by sulfides and a high carbon content. The annual capacity of the dressing plant is 660,000 metric tons of ore, and the average ore grade is 4.0 g/metric ton gold. The tailings, totalling 5 million mt with a weighted average gold content of about 1 g/metric ton, are stored in two tailing dumps.

[1]The inferred gold resources of Natalka exceed 1000 metric tons, ranking it among world-class deposits.

The geological setting, structure, and mineralogy of the Natalka deposit were described by E. P. Mashko and G. A. Topunova (pers. commun., 1949); P. I. Skornyakov and E. P. Gromova (pers. commun., 1955); and, in particular, V. D. Volodin, A. N. Balushev, A. I. Barkan, and I. K. Eremenko (pers. commun., 1956). Various genetic aspects of gold mineralization were studied by Shilo (1960), Eremin and Osipov (1974), Ivanyuk et al. (1983), M. P. Krutous (pers. commun., 1984), and Voroshin et al. (1989). Because this information heretofore has not been available to foreign geologists, this paper presents these older descriptions of the Natalka deposit as well as the results of more recent research.

Regional Geologic Setting

The Omchak mining district is a part of the Yana-Kolyma metallogenic belt. This major metallogenic belt, containing rich placer-gold mining districts and numerous mesothermal gold-quartz vein deposits and prospects, trends northwestward for more than 1000 km (Fig. 1). The Yana-Kolyma belt is located in a complex collisional zone between the North Asian craton margin and an assemblage of cratonal, passive-continental-margin, island-arc, and accretionary-wedge terranes of the Kolyma-Omolon superterrane, overlapped by post-accretionary sedimentary and volcanic assemblages (Parfenov et al., 1993). The metallogenic belt occurs mainly in the Kular-Nera terrane, a turbidite sequence that consists of a

TABLE 2. Past Production and Reserves of the Major Lode Gold Deposits of the Yana-Kolyma Metallogenic Belt

Deposit	Past production		Gold reserves
	Years	Gold (metric tons)	(metric tons)
Natalka	1944–1993	75	450
Pavlik	-	-	57.6
Shkolnoye	1985–1993	2.8	23
Omchak	-	-	16.6
Vetren	-	-	13.3
Shturm	1942	0.192	10.3
Utinka	1945–1956	6.4	7.2
Igumen	1945–1966	11.5	5.8
Degdekan	1946–1949	0.166	5.6
Burkhala	-	-	4.6
Verkhne-Khakchan	-	-	4.1
Svetloye	-	-	3.5
Yugler	1947–1955	1.098	3.0
Rodion	1947–1956; 1965–1970	4.1	1.5
Maldyak	1946–1949	1.782	0.9

thick assemblage of argillite, siltstone, and minor sandstone of Late Permian, Triassic, and Early Jurassic age, and in Middle to Late Jurassic turbidite deposits (known as the Verkhoyansk Complex) of the overlapping Inyali-Debin Foredeep.

The collisional deformation in the Yana-Kolyma metallogenic belt consists of zones of linear (longitudinal and diagonal) and domal folding, a network of generally NW-striking faults, belts of granitoid (including "batholith") plutons and dike swarms, and low-grade greenschist facies regional metamorphism. The distribution of lode gold deposits within the Yana-Kolyma belt is structurally controlled by the major NW-trending Mesozoic faults. During the tectonic evolution of the metallogenic belt, the orientation of the regional compressive stresses changed at least three times since the main period of Late Jurassic folding (Shakhtyrov, 1985). During the complex structural history, gold-bearing vein formation spanned pre- to post-batholith events. K-Ar age of pre- and post-ore igneous rocks indicates that gold-quartz mineralization of the Yana-Kolyma metallogenic belt ranged from 135 to 100 Ma (Firsov, 1985). Nevertheless, relationships with igneous rocks indicate different relative ages of Au-bearing quartz veins (S. V. Voroshin, pers. commun., 1992).

Stratigraphy

The sedimentary rocks of the Omchak mining district are part of the Verkhoyansk Complex and occur in four conformable units: the Pioner, Atkan, Omchak, and Staratel suites (Byakov and Vedernikov, 1990). To the southeast, the Pioner suite conformably overlies the Rodionov suite of the Lower Permian, and, to the west, the Staratel suite is conformably overlain by Triassic sedimentary rocks. The ore zones of the Natalka deposit are hosted by the Pioner, Atkan, and Omchak suites. The Pavlik deposit occurs in the Atkan and Omchak suites, and the Omchak deposit occurs in the Pioner and Atkan suites (Fig. 2).

Pioner suite

The Pioner suite, 2300 to 2600 m in thickness, is composed mainly of nonbedded, poorly bedded, and sometimes obscure-spotted, dark-grey to black silty argillites with disseminated sandstones. Beds of grey calcareous quartz-feldspar sandstones and sandy siltstones 10 to 50 cm thick are common. The upper part of the suite consists of dark-grey silty argillites with uniformly thin horizontal bedding. The Pioner suite was formed under relatively deep-sea conditions (Byakov and Vedernikov, 1990).

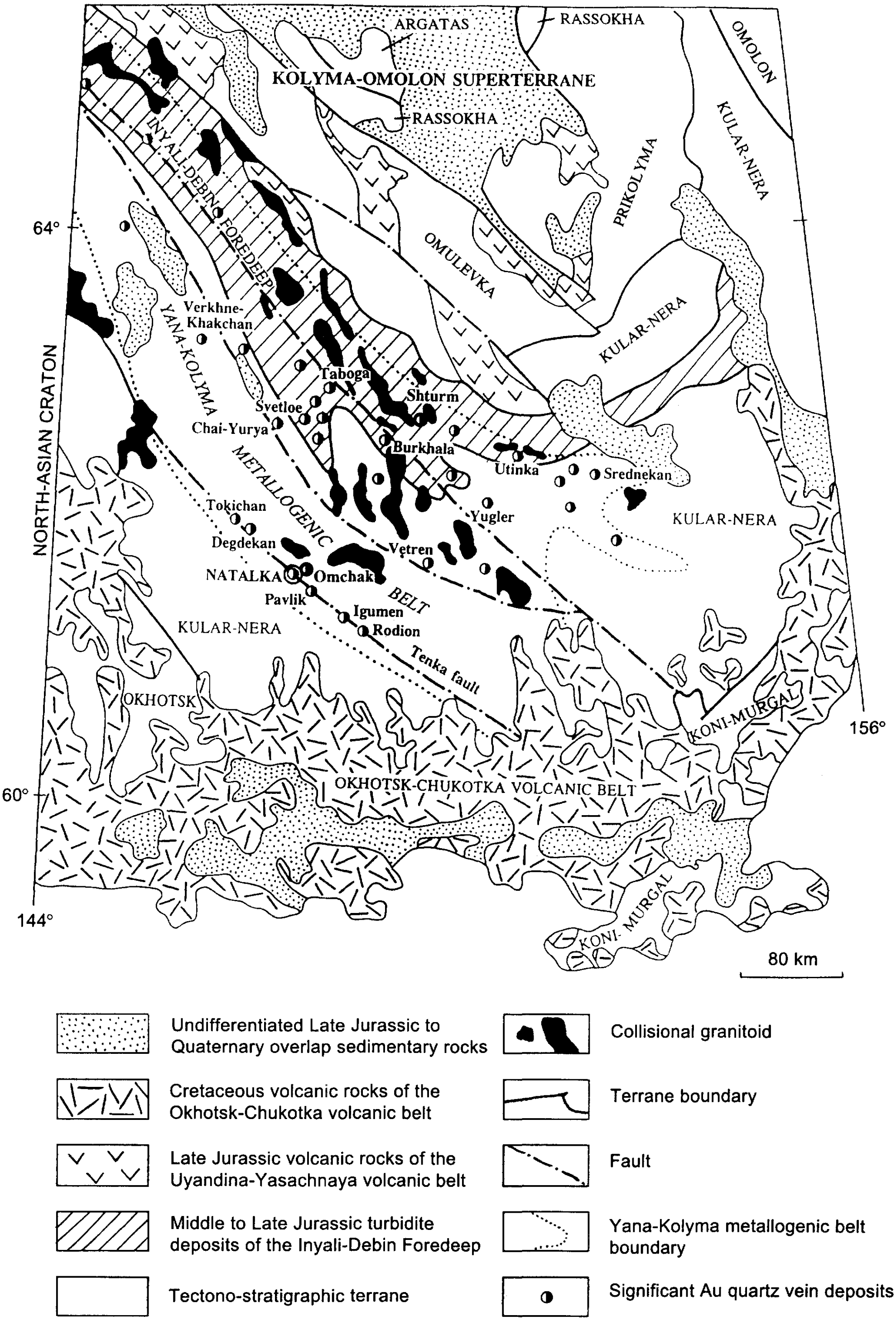

FIG. 1. Tectono-stratigraphic terrane map of the western and central part of the Magadan region, showing bedrock geology, major faults, and significant Au quartz deposits.

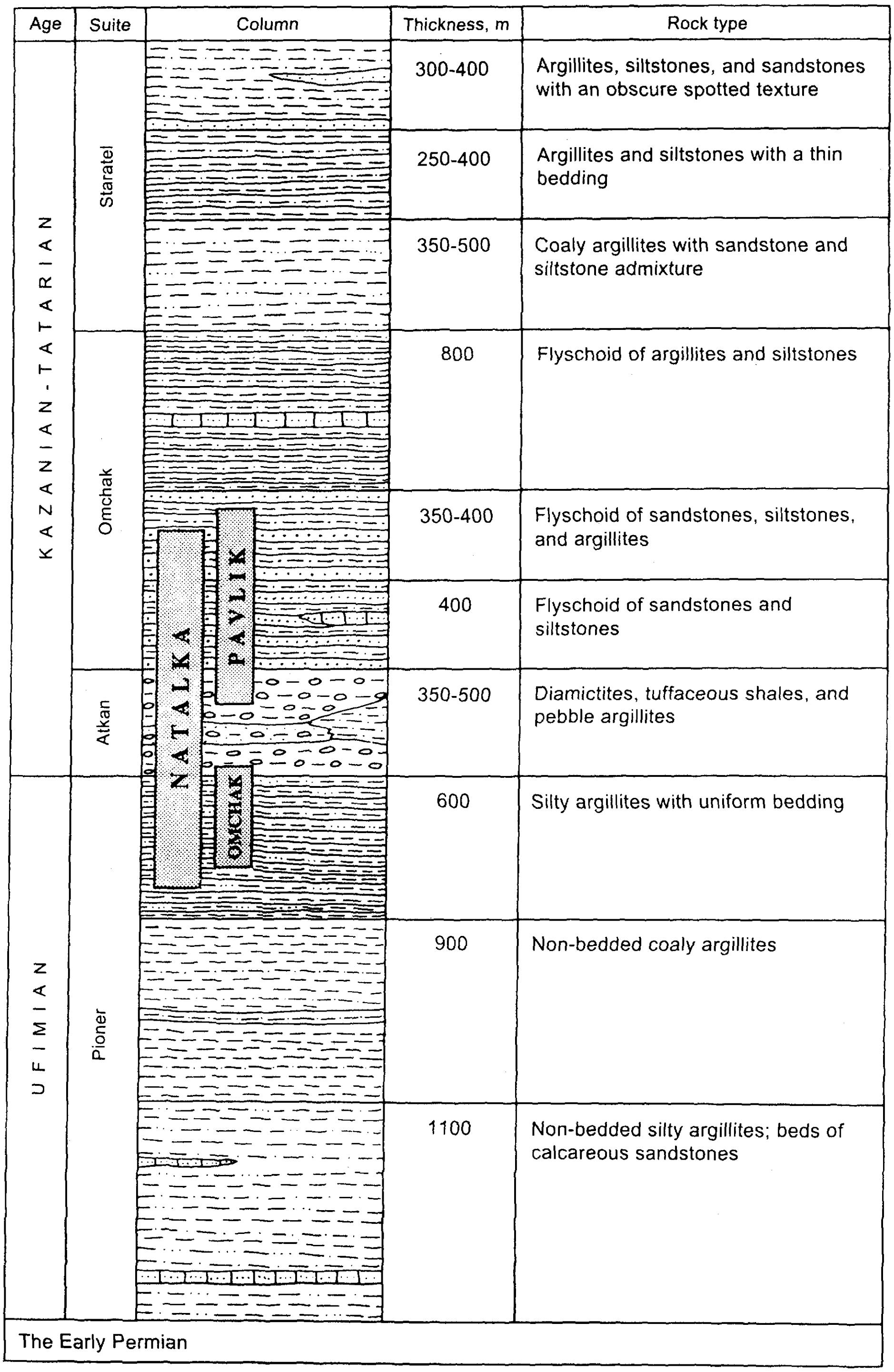

FIG. 2. Generalized stratigraphic column of the Omchak mining-district sedimentary rocks, showing occurrences of the lode gold deposits (modified from Byakov and Vedernikov, 1990).

Atkan suite

The distinctive Atkan suite—composed of 350 to 500 m of tuffaceous shales, pebble argillites, and diamictites—is a regional marker for Permian-age sedimentary rocks. The suite contains distinctive gravelstones or diamictites composed of sand-gravel or, less frequently, pebble-size clastic material irregularly disseminated in an argillite matrix. The amount of coarse clastic material ranges up to 10% of the rock volume. The clasts mostly are highly altered felsic, intermediate, and mafic volcanic rocks, and rarely are argillites, calcareous sandstones, granitoids, and quartz. The beds of argillites, siltstones, sandstones, and gravelstones are discontinuous along strike. Calcareous concretions as much as 20 cm in diameter occur in the upper portion of the suite. The formation of diamictites is interpreted as having been coeval with a significant cooling of the regional climate and formation of widespread sea drift ice (Epshtein, 1972). According to Byakov and Vedernikov (1990), diamictites formed during both volcanic activity and rock sliding.

Omchak suite

The Omchak suite, with a total thickness of about 1600 m, consists of argillites, siltstones, and rare sandstones with a typical flyschoid structure. Gravelstones, conglomerates, and diamictites are rare. The suite has pronounced horizontal bedding; cross- and graded bedding also occur locally. The upper part of the suite contains rare calcareous sandstone beds. Most of the Omchak suite formed in relatively deep continental slope conditions.

Staratel suite

The Staratel suite (total thickness of 900 to 1300 m) consists of dark-grey sandstones, siltstones, and silty argillites with poorly developed bedding. Local beds of light-grey, fine-grained sandstones also occur. This suite is characterized by faint spotted textures and a disturbed density differentiation of clastic material. The middle portion of the suite contains abundant thin intercalated argillites and siltstones and rare sandstones in packets with thicknesses as high as several meters. The Staratel suite was deposited under more shallow conditions than the Omchak suite, as indicated by a complete absence of flysch deposits. Deposition is interpreted as having been similar to a modern outer continental shelf environment (Byakov and Vedernikov, 1990).

Chemical composition of sediments

Whole-rock chemical analyses indicate that the SiO_2 content of all these suites is close to granodiorite. Al_2O_3 content is about 16%; the K_2O/Na_2O ratio is usually around one. The CO_2 content usually is not higher than 2%, but may range up to 20% or more in high-calcareous beds and concretions. The C_{org} content ranges from a few tenths of a percent to 2%, but sometimes reaches 4 to 5%. All the rocks contain irregularly disseminated sulfides, including pyrite and pyrrhotite. The total sulfur content usually is less than a few tenths of a percent (S. V. Voroshin, pers. commun., 1992).

Metamorphism

Two prograde metamorphic events have occurred in the Omchak mining district since the Late Jurassic: greenschist facies regional metamorphism and contact metamorphism (Fig. 3) (Gelman et al., 1975; Gelman, 1976; Krutous, 1991). Schistosity, porphyroblasts, and zoning of metamorphic minerals permit the identification of local metamorphic facies. The greenschist facies of regional metamorphism exhibits no obvious spatial relationship to igneous rocks. It includes chlorite-sericite and stilpnomelane subfacies with a boundary defined by the first occurrence of stilpnomelane (Table 3). Pyrite is the main sulfide mineral disseminated in greenschist facies rocks. The occurrence of cubic pyrite, pyrite aggregates, and very scarce framboidal pyrite are the first indicators of metamorphic recrystallization of sedimentary sulfides.

Along the NW-striking faults are hornfels zones that range from biotite-grade rocks with sphene and ilmenite to cordierite-andalusite hornfels and garnet-pyroxene skarns. Granitoid bodies crop out adjacent to or occur immediately below these contact metamorphic aureoles. Contact metamorphism above inferred deep plutons (as at the Natalka deposit) is characterized by ilmenite porphyroblasts (usually replaced by sphene and leucoxene), incip-

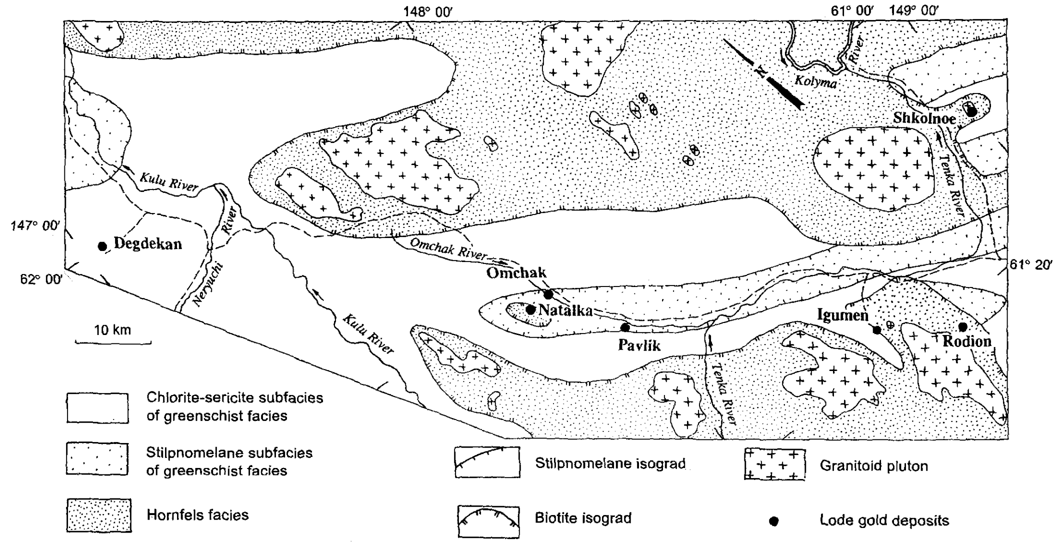

FIG. 3. Simplified map of regional prograde metamorphism and occurrences of lode gold deposits (modified from M. L. Gelman, M. P. Krutous, A. U. Filippov, and O. G. Epshtein, 1976 unpublished map).

ient cordierite and/or andalusite, and local nodules of clayey and carbonaceous material of equant or elongated form that occur only in carbonaceous-clayey rocks. Metamorphic pyrrhotite, derived from primary sedimentary pyrite, is the main contact metamorphic sulfide mineral. Pyrite relicts are typical of outer zones (biotite subfacies), whereas complete pyrite-to-pyrrhotite alteration is characteristic of inner zones (Voroshin et al., 1993).

Gold-bearing quartz veins occur within all the above metamorphic zones (Table 3; Fig. 3). Orebodies of the Igumen and Rodion deposits, within hornfels facies, also were subjected to subsequent metamorphism (Firsov, 1957, 1958; Tyukova, 1989).

Igneous Rocks

The igneous rocks in the Omchak mining district are Late Jurassic to Late Cretaceous medium to small granitoid bodies and dikes of variable composition. They are interpreted as having formed during and subsequent to accretion of the Kolyma-Omolom superterrane to the North Asian craton. More than five magmatic complexes are distinguished by age and chemical composition, each exhibiting several mafic-to-felsic intrusive phases (Anorov et al., 1991).

Within the Natalka deposit area, only mafic, intermediate, and felsic dikes occur, with thicknesses of as much as 15 m and lengths ranging from tens to hundreds of meters. Mafic dikes are more persistent, thicker, and older than felsic dikes. According to Firsov (1964), the K-Ar age of spessartite in the dikes is 150 Ma. The dikes contain abundant fragments of pre-ore vein quartz. Ore formation postdates emplacement of both mafic and felsic dikes. Milky chlorite-carbonate-quartz veins are cut by dikes, and both dikes and milky quartz veins pre-date formation of the ore-stage gold mineralization.

The orebodies of the Pavlik deposit are cut by rhyolite breccias, which are contemporaneous with the rhyolite breccia stock near the mouth of Vanin Creek (Fig. 4). The rhyolite breccias of the Vanin stock are cut by Late Cretaceous hornblende quartz diorite (Voroshin et al., 1989). Thus, gold mineralization in the Omchak district formed subsequent to the ~150 Ma intrusion of mafic-felsic dikes and prior to the Late Cretaceous granitoid intrusions.

Structure

The Natalka and associated deposits are confined to the major, NW-trending Tenka fault

TABLE 3. Metamorphic Facies and Location of the Lode Gold Deposits

Metamorphic facies	Associated minerals	Index minerals	Au quartz deposits
Chlorite-sericite subfacies of greenschist facies	Quartz, feldspars, sericite, chlorite, carbonates, rutile, pyrite, carbonaceous material	Pyrite	Degdekan
Stilpnomelane subfacies of greenschist facies	Quartz, feldspars, sericite, stilpnomelane, chlorite, carbonates, rutile, ilmenite, pyrite, carbonaceous material	Stilpnomelane, pyrite	Pavlik, Omchak, Natalka
Biotite subfacies of hornfels facies	Quartz, feldspars, muscovite, biotite, carbonates, rutile, ilmenite, sphene, pyrite, pyrrhotite, carbonaceous material	Biotite, pyrrhotite	Natalka, Igumen, Rodion
Cordierite or pyroxene-garnet subfacies of hornfels facies	Quartz, feldspars, muscovite, biotite, cordierite, andalusite, amphibole, pyroxene, garnet, rutile, sphene, pyrrhotite, graphite	Cordierite, garnet, pyrrhotite	Igumen, Rodion

(Fig. 1). According to geophysical data (Vaschilov, 1970), the Tenka fault is defined by symmetrical gravity anomalies and is interpreted as a horst block of folded basement. The thickness of the sedimentary rock units is 4 km in the apical portion of this horst and up to 5 to 6 km or more on both sides. The horst block undulates along strike and can be discriminated as smaller individual blocks. Major domal folds, which comprise the regional Tenka anticline, correspond to basement blocks in the upper parts of the Tenka fault.

Local domal folds along the Tenka fault differ from folds elsewhere in the Kular-Nera terrane. The Kular-Nera terrane generally contains conjugate en-echelon anticlines and synclines that indicate right-lateral horizontal compression. In contrast, domal folds along the Tenka fault zone are not conjugate with synclines (Voroshin et al., 1989), and their formation is interpreted to have been the result mainly of vertical movements that probably were related to collisional deformation and magmatic diapirism (Shakhtyrov and Eremin, 1983). The low-amplitude domal folds are superimposed on the major folded structures of the Kular-Nera terrane and thus represent an independent, younger structure.

The Omchak mining district occurs in the southwestern part of the Omchak domal fold of the Tenka fault zone. This domal fold trends NNW and extends for nearly 90 km from the Kulu River in the north to the mouth of the Omchak River in the south. Dike swarms are widespread in the middle portion of the anticline, and small to intermediate-size bodies, ranging from gabbro and diorite to leucocratic granite, occur along its periphery. Two ore-hosting systems of en-echelon faults occur as horst-anticlines. They trend to the west from the Omchak anticline axis, indicating a left-lateral movement. The Natalka deposit occurs in the northern fault system, whereas the southern system hosts the Pavlik deposit (Fig. 4).

The Natalka deposit occurs in the Z-shaped fault zone (which is as long as 13 km), with evidence for strike-slip faulting. A block of rocks of ca. 6 to 8 km in width—a raised portion of the Omchak graben-syncline—occurs in the fault zone. Ore-hosting faults trend 320 to 340° within most of the deposit. These faults dip 60° northward in the southeastern part of the deposit, become nearly vertical near the middle of the deposit, and dip southward in the northwestern end.

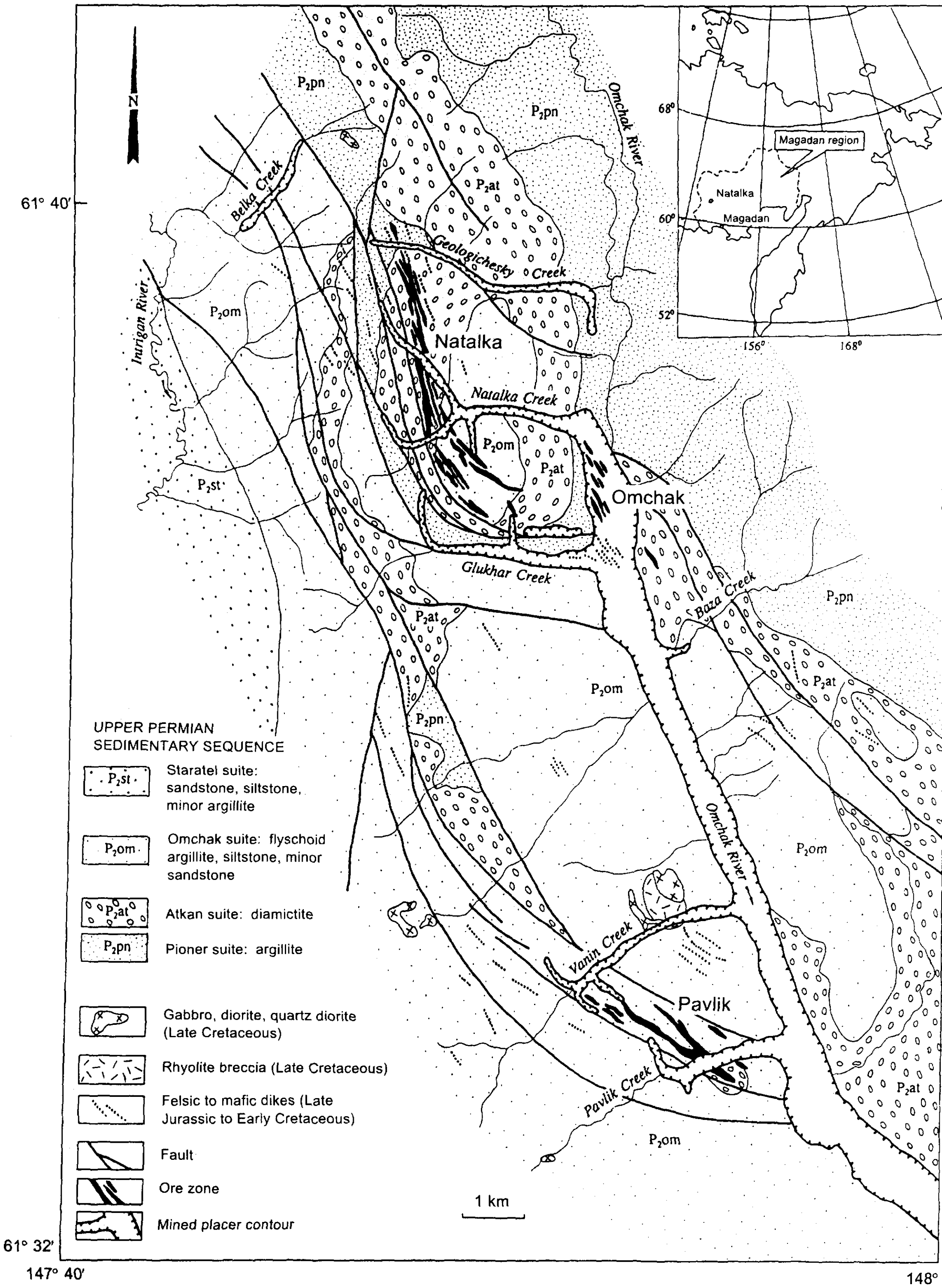

FIG. 4. Geologic map of the Omchak mining district.

Most of the Natalka deposit occurs within a system of faults that forms a horsetail structure off the Main fault (Figs. 5 and 6) and (in combination with the latter) merges with the Omchak fault zone to the southeast. The Main fault generally dips eastward. To the northeast, orebodies are bounded by the Northeast fault, which gradually trends southeast away from the set of ore zones related to the Main fault and dips northeast at a somewhat lower angle than the Main fault.

The Main fault and other ore-controlling faults exhibit complex compressive and strike-slip characteristics. Generally, a longitudinal zoning of the Z-shaped set of ore-hosting faults occurs at the Natalka deposit. The northwestern part of the deposit contains only relatively minor faults that developed under strike-slip and compression. Faults in the middle portion exhibit a combination of mainly compressive and lesser strike-slip movement, and those in the southeastern portion of the deposit exhibit extension-related dip-slip deformation. The main ore zones are confined to the transition zone between extension and compressional structures.

The structural development of the Omchak mining district was characterized by periodic changes of tectonic stress fields that resulted in both left-lateral and right-lateral fissure systems. The pre-ore dikes and the ore-bearing veins both were emplaced during compressive and left-lateral movement. Subsequent horizontal compression led to post-ore folding and right-lateral movement. Some post-ore dikes intruded rocks of the Omchak mining district during a later renewal of left-lateral movement, and then granitoid bodies intruded during right-lateral movement.

Ore Zones

Ninety-one ore zones are recognized and delineated in the Natalka deposit, with 8 major ore zones containing more than 60% of Au reserves (Table 4; Figs. 5 and 6). In plan view, the ore zones are about 5 km long and 100 to 200 m wide in the northwest, 350 to 400 m wide in the center, and 600 m wide in the southeast.

Ore zones consist of quartz veins, stringers, and altered wall rock. Subparallel stringers occur in longitudinal tension cracks that usually lie at moderate angles to schistosity and represent the main ore reserve. Transverse joints are variably oriented and less mineralized.

Quartz veins

Quartz veins have an axial occurrence within some ore zones (such as Nos. 3 and 39) and exhibit brecciated, brecciform, banded, or massive structures. Sulfides are mostly confined to the host rock clasts in brecciated and banded veins, but also may occur in the quartz matrix. Individual veins are as thick as several meters and as long as hundreds of meters.

Quartz stringers

Subparallel quartz stringers along steep joints represent the most widespread type of ore mineralization. The thickness of the stringers ranges from less than one mm to as much as several cm. The number of stringers varies from 5 to 50 per m in the ore zones. Stringers usually constitute not more than 10% of the wall rock, but may reach 30 to 40% in places. Sulfides are present both in quartz stringers and wall rocks.

Thin and variably oriented quartz, sulfide-quartz, and sulfide stringers also form ore zones. The number of stringers, as in the above subparallel sets, also is quite variable. Hydrothermally altered (silicification, sulfidization, carbonatization) sedimentary rocks and dikes occur in association with the mineralization. Sulfide-rich, intensely contorted, and crushed ore zones occur in faults.

Ore-zone morphology

Usually, ore zones include several morphologic and structural features. A transverse zoning sometimes occurs, consisting of a quartz vein surrounded by quartz stringers, and a peripheral area of hydrothermal alteration in host rocks. An irregular pattern of several morphologic types in an orebody is more typical. Subparallel quartz stringers after steep joints are widespread in the northwestern part of the Natalka deposit, whereas thin and variably oriented quartz, sulfide-quartz, and sulfide stringers and altered host rocks are more common in the southeastern part. The distribution of gold ore is irregular. Vertically elongated bonanza zones regularly alternate with low-

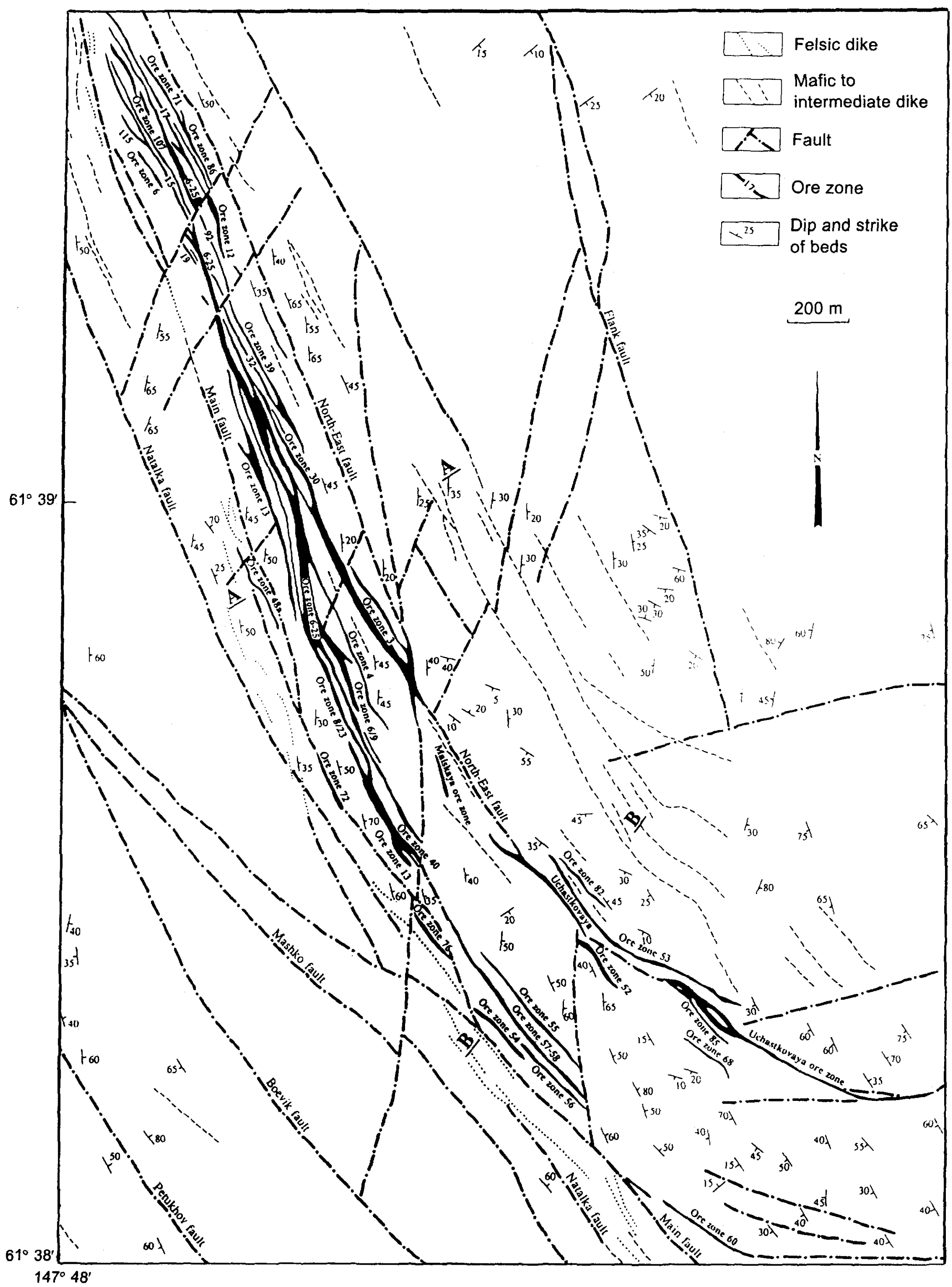

FIG. 5. Structural map of the Natalka deposit (modified from A. I. Kalinin, V. K. Kanischev, and A. G. Orlov, 1991, unpublished map).

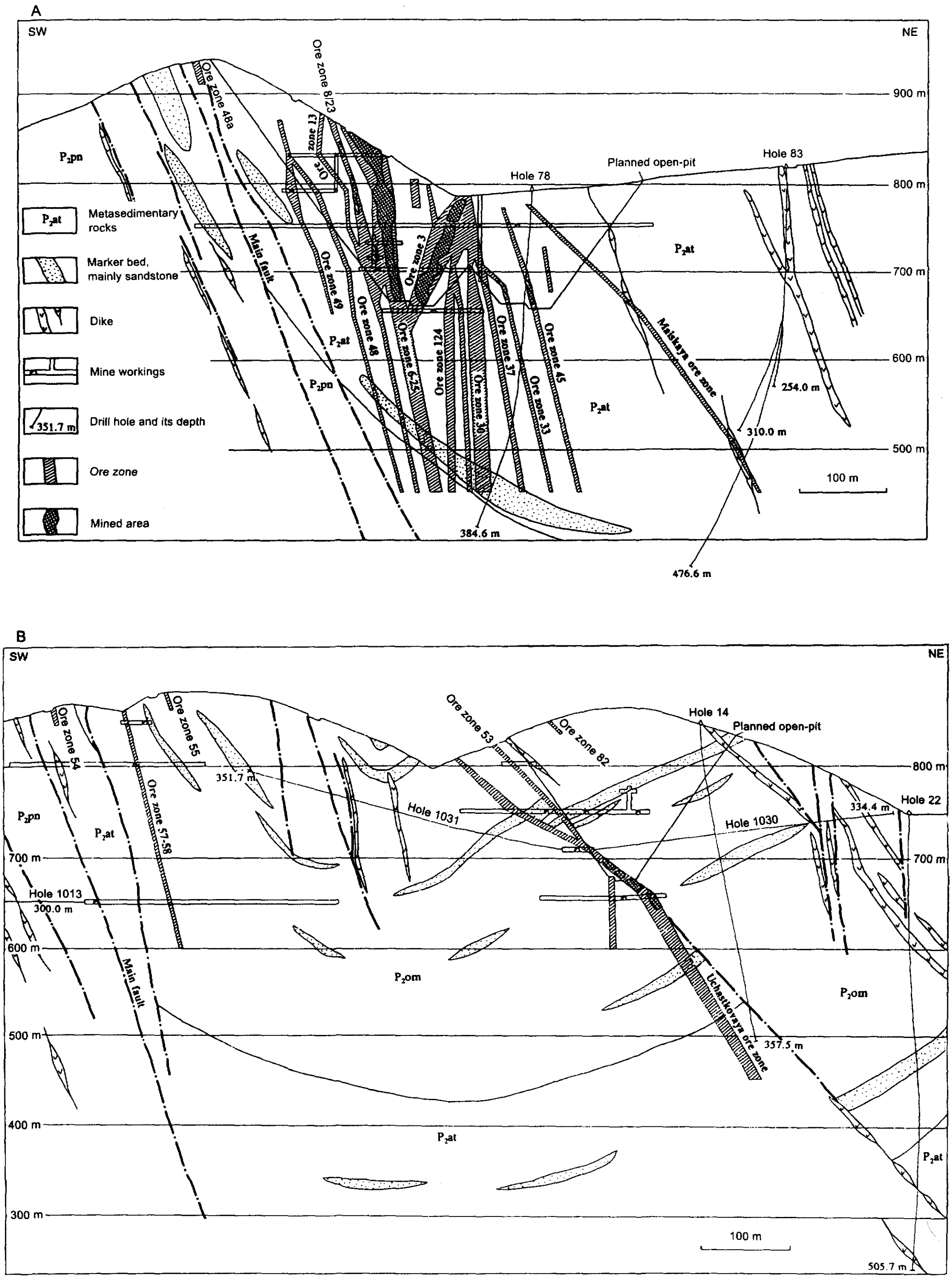

FIG. 6. A. Simplified SW–NE cross-section (A–A′) of the central part of the Natalka deposit. B. Simplified SW–NE cross-section (B–B′) of the southeastern part of the Natalka deposit (see Fig. 5).

TABLE 4. Characteristics of the Major Ore Zones of the Natalka Deposit

Ore zone no. or name	Strike	Dip angle	Dip direction	Length (m)	Thickness (m)		Au grade (g/t)		Share of total reserves, %
					Range	Average	Range	Average	
6-25	330-340	70-85	NE	2100	2.0-37.0	8.0	0.2-28.9	3.64	10.2
17	310-340	75-80	NE	1500	2.0-16.5	5.0	1.6-27.7	3.81	6.0
30	320-340	85	NE	460	1.5-20.5	6.6	1.2-16.3	3.30	4.9
33	305-340	80	NE	570	2.0-12.0	5.4	1.7-34.2	3.96	6.6
3	335-345	60-70	SW	600	2.0-32.0	13.4	1.3-22.7	4.69	7.5
8/23	335-345	70-85	NE	2300	2.0-26.0	5.0	0.2-14.6	3.89	8.4
Mayskaya	335-345	65-75	NE	1500	2.0-20.0	4.8	0.2-25.7	4.12	5.9
Uchastkovaya	330-85	30-50	NE	2300	1.5-26.5	6.0	0.5-28.0	4.47	13.9

grade areas within an orebody. High grades are confined to the areas of quartz veining and intense sulfidization.

Post-ore tectonic events are widespread, but displacement of orebodies usually is insignificant. Displacements usually are not more than 0.5 to 1.5 m, and in some cases are as great as 10 to 15 m along gentle and steep faults.

Wall-Rock Alteration

Alteration of wall rocks occurs in broad aureoles, extending to several hundred meters, around all orebodies. Alteration is most intense within an ore zone, but also may occur in intervals between ore zones. Typical alteration includes silicification, carbonatization, sulfidization, albitization, K-feldspathization, sericitization, or scapolitization. Usually, several types of wall-rock alteration are superimposed. The distribution of alteration minerals depends on host rocks. Alteration of silicate minerals is insignificant in argillites, but sulfidization is widespread. Siltstones, sandstones, and diamictites are silicified, carbonatized, and sulfidized. Mafic dikes exhibit carbonatization and sulfidization and lesser silicification, albitization, and scapolitization. Felsic dikes are intensely altered to quartz, albite, and sericite.

A main feature of the alteration is intense silicification and carbonatization. Quartz replaces silicate minerals. Organic matter usually is preserved in the silicified sedimentary rocks as grey material locally forming a relict banded structure. Carbonate minerals occur between large grains of quartz and feldspar, and replace cement; feldspar and sometimes quartz and other silicate minerals also are replaced by carbonates, which include ankerite, dolomite, and calcite. Arsenopyrite and pyrite are the main sulfide minerals in the alteration zones. Arsenopyrite is the predominant sulfide in altered wall rocks, and usually replaces sedimentary pyrite or Fe-bearing silicates and oxides. Pyrite also occurs as a hydrothermal alteration mineral, as indicated by euhedral crystals occurring in mineralized shear zones or dikes. Albite and potassium feldspar form well-developed crystals adjacent to quartz veins and stringers, especially where hosted by dikes. Sericite alteration usually replaces feldspars. Scapolite replacing plagioclase was discovered by P. I. Skornyakov (pers. commun., 1952) in altered mafic dikes.

The chemical composition of the altered wall rocks indicates addition of CaO, CO_2, S, As, and Na_2O, and depletion in SiO_2 and K_2O (D. N. Safronov, V. I. Nayborodin, and V. I. Goncharov, pers. commun., 1978). The addition of CaO, CO_2, S, and As is indicated by intense carbonatization and sulfidization of wall rocks. Silica in altered rocks averages 4 to 5% less than in unaltered rocks. Areas of lowest SiO_2 coincide with zones of intense carbonatization. Thus, SiO_2 is redistributed as quartz veins within ore zones and is replaced by abundant carbonates. The behavior of alkalies is similar. The high Na_2O content of albitized wall rocks reflects replacement of K_2O; the K_2O content of quartz veins is high because of the presence of hydrothermal sericite and K-feldspar. The Fe contents of altered and unaltered wall rocks are similar, indicating that the iron required for sulfide formation mostly was taken from host rocks.

Ore Mineralogy and Paragenesis

Roughly 40 ore and gangue minerals have been identified in the Natalka deposit. The major gangue constituent is quartz. Subordinate gangue phases include carbonates (calcite, magnesite, dolomite, and ankerite), feldspars (albite and K-feldspar), chlorite, sericite, kaolinite, montmorillonite, barite, and apatite. The main sulfide minerals are arsenopyrite and pyrite, with arsenopyrite dominant, and they constitute more than 95% of the sulfides. Small amounts of pyrrhotite, Co-Ni sulfarsenides, sphalerite, chalcopyrite, galena, native gold, ilmenite, and rutile are widespread. Scheelite, millerite, tetrahedrite, bournonite, boulangerite, and stibnite occur locally. The total sulfide content of the ore zones is 1 to 3%, and locally reaches 5%.

Ore minerals formed approximately simultaneously with the precipitation of gangue quartz. The crystallization sequence of the ore minerals was: (1) pyrite + arsenopyrite + pyrrhotite + Co-Ni sulfarsenides; (2) sphalerite + chalcopyrite + galena + native gold; (3) tetrahedrite + bournonite + native gold; (4) boulangerite; (5) stibnite.

The pyrite + arsenopyrite + pyrrhotite + Co-Ni sulfarsenides assemblage included replacement of disseminated pyrite in the wall rocks by arsenopyrite. The abundance of arsenopyrite relative to pyrite increased from the host rocks toward the veins, with arsenopyrite occurring exclusively within some veins. Sporadically occurring scheelite probably is contemporaneous with this assemblage. Pyrite forms idiomorphic cubic and pentagonal-dodecahedral crystals, or complex aggregates in altered metasedimentary rocks, dikes, and quartz veins. The pyrite in ore zones always contains arsenic, ranging up to 2.5%. Arsenopyrite occurs as isometric or flattened crystals in quartz veins and in adjacent altered metasedimentary rocks, and as elongated prismatic crystals in mineralized dikes. Arsenopyrite composition is constant for all ore zones, and the S/As atomic ratio is about 1 (Voroshin et al., 1989). Gold and silver grades of arsenopyrite grains are as high as hundreds of grams per metric ton. Pyrrhotite usually occurs as small isometric inclusions in arsenopyrite and pyrite, and also occurs with pyrite as a primary mineral in host rocks. Cobaltite (alloclasite (?)) occurs as scarce small (0.001–0.01 mm) inclusions in pyrite in altered wall rocks and is variable in composition (Voroshin et al., 1989).

The sphalerite + chalcopyrite + galena + native gold assemblage, formed subsequent to the arsenopyrite and pyrite, is quite subordinate, but is ubiquitous. Sphalerite forms small, isometric, anhedral, light-brown (usually individual) grains, with an FeS content of less than 10 mol% (Voroshin et al., 1989). The Cd and Mn content usually is less than 0.1 wt%. Chalcopyrite occurs as emulsion disseminations in sphalerite and in small inclusions in pyrite and arsenopyrite. Galena occurs with sphalerite, is usually intergrown with native gold, and also occurs in fractures and microvugs of arsenopyrite. Very low contents of Ag, Sb, and Bi are characteristic of galena (Voroshin et al., 1989). Native gold occurs as individual grains as large as 0.1 to 2.0 mm in diameter, or as fine disseminations (0.001 to 0.1 mm) in arsenopyrite. According to S. V. Yablokova and V. P. Pluteshko (pers. commun., 1979), gold grains more than 0.1 mm in diameter make up 37 to 40% of the ore. Gold has an irregular shape, often forms intergrowths with other minerals, and sometimes may have a granular inner structure. On the whole, gold is uniform in composition and is independent of ore type, host rocks, and spatial distribution. Gold fineness averages 750 to 790 with extreme values of 550 and 850.

The tetrahedrite + bournonite + native gold assemblage occurs mainly in mineralized dikes in the southwestern part of the deposit. Tetrahedrite composition varies at different sites, whereas bournonite exhibits a constant composition (Voroshin et al., 1989). The sulfides form intergrowths with native gold, sphalerite, chalcopyrite, galena, arsenopyrite, and pyrite; some samples reveal successive replacement of galena by bournonite and boulangerite.

Boulangerite occurs in mineralized spessartite dikes in the southwestern part of the deposit, and forms needle- and sheaf-like aggregates as large as 2 to 3 mm in feldspar-quartz veinlets. Boulangerite replaces all of the above mineral assemblages. In studies of other gold deposits of the Yana-Kolyma metallogenic belt, including the Shkolnoye and Kontrandya deposits (R. A. Eremin and N. E. Savva, pers. commun., 1983; Voroshin and Eremin, 1995), boulangerite also formed later than native gold.

However, no direct boulangerite-gold ingrowths are known to exist in the Natalka deposit.

Stibnite occurs only in quartz veinlets and as hydrothermal disseminations in altered rhyolite dikes in the southeastern part of the deposit. In veinlets, stibnite occurs as continuous large crystal aggregates; if quartz is abundant, stibnite occurs in interstices between quartz crystals. Stibnite is associated spatially only with arsenopyrite and pyrite; small crystals of the latter are cemented by stibnite and cut by thin stibnite veinlets.

Ore Geochemistry

Methods

About 7400 rock samples, including 3000 from underground mine workings, were collected in the Omchak mining district for geochemical analysis. All accessible crosscut workings were chip-channel sampled at 2-m intervals. Surface sampling was conducted on a 500 × 50 m grid over an area of ca. 120 km^2. Quantitative emission spectrography was employed to determine the contents of chromium, manganese, cobalt, nickel, copper, zinc, arsenic, molybdenum, silver, tin, antimony, barium, tungsten, lead, and bismuth; gold content was determined by atomic absorption.[2]

Results

The local geochemical background was determined for all analyzed elements within the sampled area (Table 5). The average harmonic content of each element was calculated by dividing the total number of samples into the sum of the inverse contents of an element in order to to assess the geochemical background and to avoid inclusion of anomalously high values in background calculations.

Three assemblages of elements were identified by combined correlation and factor analyses (Table 6). The first assemblage consists of gold, arsenic, tungsten, bismuth, and antimony and exhibits its highest concentrations within ore zones. In the most anomalous rocks, the amount of arsenic is 200 times, gold 100 times, and tungsten 3 times higher than that of the geochemical background. Farther from the ore zones, but still within the Omchak mining district, the concentrations of these elements are lower than the assessed geochemical background (Fig. 7).

TABLE 5. Local Geochemical Backgrounds, Omchak Mining District

Element	Local geochemical background (ppm)	Clarke of shale (Vinogradov, 1962)
Cr	45	100
Mn	1450	670
Co	20	20
Ni	18	95
Cu	20	57
Zn	125	80
As	40	6.6
Mo	1.35	2
Ag	0.35	0.1
Sn	5	10
Sb	1.9	2
Ba	100	800
W	1.5	2
Au[1]	0.1	0.001
Pb	21	20
Bi	1.7	0.01

[1]Au local geochemical background only for the Natalka deposit.

The second assemblage consists of silver, lead, zinc, and tin. The local geochemical backgrounds for these elements are close to the clarke values. This assemblage is typical of rocks collected from the upper portion of the deposit. The elements of the assemblage are not found in the central portion of the deposit because shallow parts of the orebodies have been eroded away there. The Pavlik deposit, having undergone much less erosion than the Natalka deposit, contains numerous rock samples that are enriched in this element assemblage (Fig. 7).

The third element assemblage consists of chromium, nickel, cobalt, barium, and (less consistently) copper and molybdenum. The geochemical backgrounds for each element—except cobalt—are lower than the clarke values. The orebodies do not contain anomalous amounts of these elements and ore samples often are depleted in these elements (Fig. 7).

[2]All analyses were performed at the Northeast Interdisciplinary Research Institute, Russian Academy of Sciences, Far East Branch, Magadan. Analytical data were processed on the ES-1066 computer with the "Leader" software package.

TABLE 6. Factor Weights of Elements in the Omchak Mining District

Assemblage	Element	Factor		
		1	2	3
I	Au	-0.87	0.42	0.10
	As	-0.88	0.40	0.08
	W	-0.95	0.24	0.09
	Bi	-0.76	-0.51	-0.38
	Sb	-0.57	-0.72	-0.06
II	Ag	0.21	0.94	-0.11
	Pb	0.51	0.82	-0.19
	Zn	0.56	0.74	-0.33
	Sn	0.39	0.88	0.21
III	Cr	0.97	0.11	0.09
	Ni	0.98	0.10	0.02
	Co	0.96	0.20	0.01
	Ba	0.92	-0.21	0.30
	Cu	0.81	-0.52	-0.09
	Mo	0.82	0.43	0.34

The geochemistry of the ore zones generally is consistent across the entire mining district. The element assemblages described above do not exhibit distinct spatial geochemical zonation because of the structural complexity of the ore zones. The latter are defined by anomalously high concentrations of gold, arsenic, tungsten, bismuth, and antimony, and also are characterized by increased levels of silver, lead, zinc, and tin at shallow depths. Barium, molybdenum, and nickel tend to decrease at depth, whereas cobalt increases at depth. Still, in both cases, these differences in element contents generally are statistically insignificant, which indicates that little geochemical change has taken place over a considerable vertical interval.

Fluid-Inclusion Studies

Fluid inclusions—mainly in vein quartz and also in vein calcite and scheelite—were studied by D. N. Safronov, V. I. Nayborodin, and V. I. Goncharov (pers. commun., 1978); R. A. Eremin, V. V. Vortsepnev, and V. I. Goncharov (pers. commun., 1981); B. O. Ivanyuk (1981); and N. A. Gibsher (pers. commun., 1983). The ore-bearing quartz exhibits a dull aspect and is milky because of abundant fluid inclusions ranging from 1 to 20 microns in diameter, with a predominance between 3 and 10 microns. Inclusions with diameters of less than 5 microns predominantly are rounded and vacuole-like in form, whereas larger inclusions are irregular. Usually groups of inclusions include up to several dozens of vacuoles. The majority of inclusions (>80%) are not confined to linear structures or zones of growth, but have a random occurrence.

Several types of fluid inclusions occur in the ore-bearing quartz: single-phase aqueous inclusions (3–5% of the total), single-phase liquid CO_2 inclusions (5%), two-phase liquid-vapor inclusions (the predominant inclusion, 60–80%), two-phase vapor-dominated inclusions (5–10%), and three-phase liquid H_2O–liquid CO_2–vapor inclusions (20–30%). Often the same grain contains vacuoles having different phase relations. The occurrence in quartz of coeval primary inclusions with variable phase relations indicates trapping of heterogeneous (boiling) solutions (N. A. Gibsher, pers. commun., 1983).

Final homogenization temperatures of the inclusions trended mainly to liquid; no more than 5% homogenized to vapor. The temperature interval of homogenization ranges from 150° to 360° C. Two temperature peaks—of 280 to 320° C and 180 to 240° C—occur in many samples of the ore-bearing quartz. Often decrepitation of the fluid inclusion took place

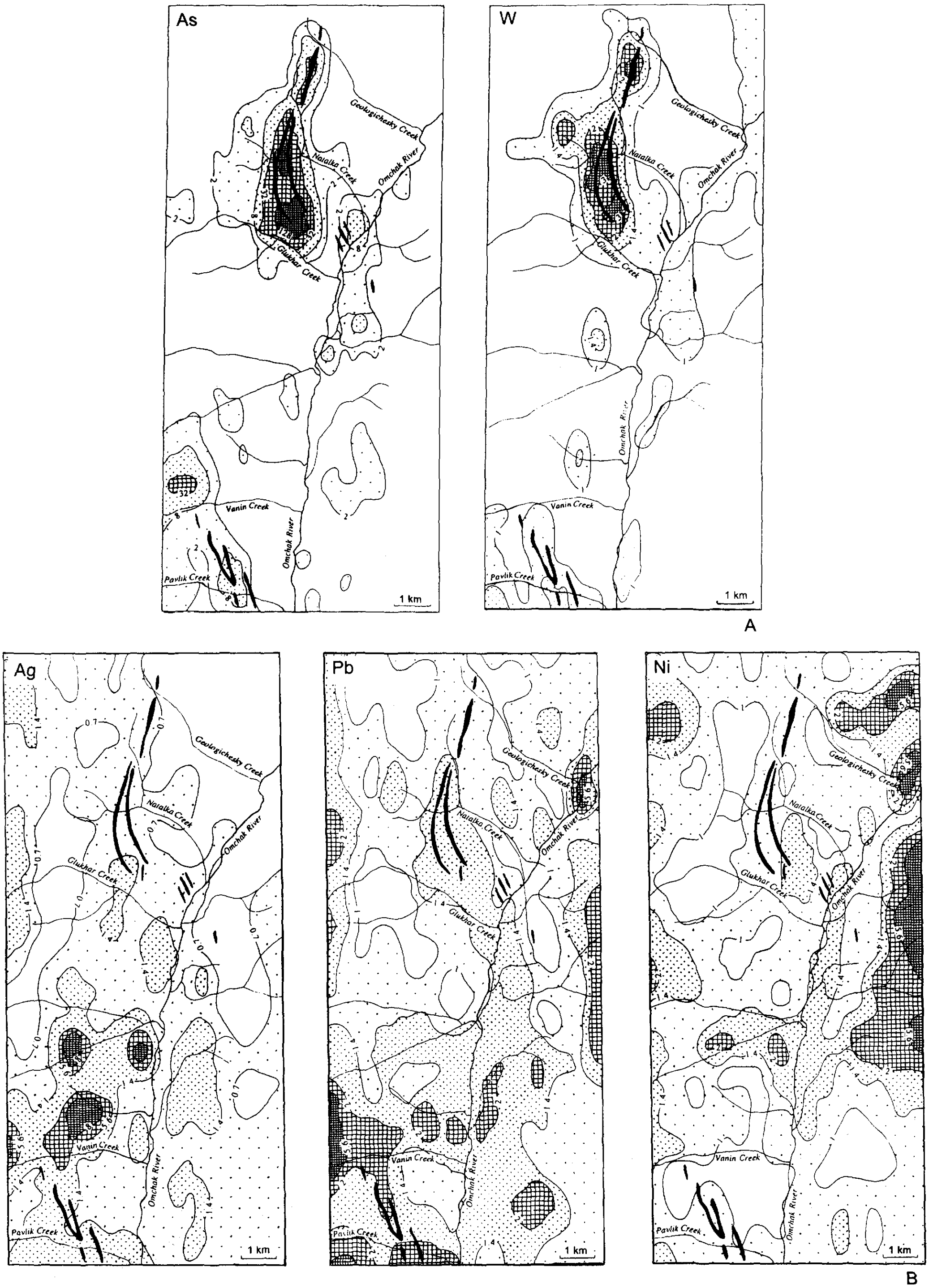

FIG. 7. A. Distributions of arsenic and tungsten in surface bedrock over the Omchak mining district. B. Distributions of silver, lead, and nickel in surface bedrock over the Omchak mining district. All values are multiples of the backgrounds.

at 260 to 300° C during heating experiments. A trend of fluid-inclusion homogenization temperatures with depth does not occur. Two similar temperature peaks of 260 to 340° C and 180 to 220° C occur in pre-ore vein quartz. The homogenization temperatures for scheelite range between 240 and 280° C and for calcite between 80 and 230° C (R. A. Eremin, V. V. Vortsepnev, and V. I. Goncharov, pers. commun., 1981). The density of CO_2 was determined for single-phase liquid CO_2 inclusions in ore-bearing quartz and ranges from approximately 0.5 to 0.65 g/ml to 0.75 g/ml (D. N. Safronov, V. I. Nayborodin, and V. I. Goncharov, pers. commun., 1978), or 0.88 to 0.9 g/ml (N. A. Gibsher, pers. commun., 1983). Estimates of trapping pressure range from 400 to 1100 bar, and, according to N. A. Gibsher (pers. commun., 1983), may be as great as 1500 to 2000 bar. She suggests that such a high pressure was episodic and that large pressure fluctuations were associated with fracturing of vein quartz. Freezing-point depression measurements for the two-phase liquid-vapor inclusions in ore-bearing quartz vary from 0° to -2.5° C, implying a salinity of less than 5 equiv wt% NaCl. CO_2-melting temperatures of -56.6° C indicate that the non-aqueous volatile phase was pure CO_2 (N. A. Gibsher, pers. commun., 1983). Leachate analyses indicate the predominance of Na among cations (Na:K = 2:1; Na:Ca = 3:1) (R. A. Eremin, V. V. Vortsepnev, and V. I. Goncharov, pers. commun., 1981).

Stable-Isotope Studies

Analytical procedures

Sulfide and quartz separates were prepared by crushing samples and hand picking minerals using a binocular microscope. Sulfur- and oxygen-isotope values were obtained from the Far East Institute of Geology, Russian Academy of Sciences,[3] Vladivostok. Sulfur-isotope compositions were analyzed by combustion with excess CuO to produce SO_2 at 800° C using a MI-1201v mass spectrometer. The $\delta^{34}S$ values are reported relative to the Canyon Diablo troilite (CDT) standard. Analyses are considered accurate to ±0.2 per mil. Before mass spectrometric analysis the oxygen was extracted from quartz with bromine pentafluoride and converted to CO_2. The $\delta^{18}O$ data are quoted relative to SMOW. Analyses are considered accurate to ±0.2 per mil.

Results

The results of 41 $\delta^{34}S$ analyses of sulfides from wall rocks and ores are presented in Table 7 and Figure 8.[4] The $\delta^{34}S$ composition of pyrite hosted by non-mineralized metasedimentary rocks is less than 0 per mil, and $\delta^{34}S$ composition of pyrite hosted by magmatic rocks is greater than 0 per mil, irrespective of the age of the host rocks. The isotopic composition of pyrite in orebodies, irrespective of the host rocks, is somewhat negative and is very close to that of sedimentary rock–hosted pyrite. The $\delta^{34}S$ of pyrite in the dike-hosted ores differs from that of pyrite hosted by non-mineralized dikes, and probably indicates that sulfur in mineralized dikes is derived mostly from the surrounding sedimentary rocks. The $\delta^{34}S$ of pyrite in pre-ore chlorite-carbonate-quartz veins also is very close to that of sedimentary rock–hosted pyrite. The isotopic composition of ore-hosted arsenopyrite is the same as that of pyrite; fractionation between these two minerals is insignificant. The relatively lower $\delta^{34}S$ values for bournonite, boulangerite, and stibnite, in comparison with pyrite and arsenopyrite, probably indicates an increase of the oxidized/reduced sulfur-species ratio in solution during the later stages of mineralization (Rye and Ohmoto, 1974). Nevertheless, reduced sulfur dominated solutions during mineralization, as indicated by the presence of a C/CO_2 buffer in host rocks and the presence of ilmenite but not magnetite. Thus, the $\delta^{34}S$ values of the ore solutions were close to those measured for the sulfides.

The temperatures of sulfide formation could be determined by the sphalerite-galena pair, but unfortunately, we were unable to extract a sufficient amount of pure galena from the Natalka deposit. At the nearby Degdekan deposit, the temperature estimates obtained by sphalerite-galena pairs are 174° C and 202° C (Voroshin and Eremin, 1995). The same temperatures probably are typical of the Natalka deposit, because both deposits are characterized by

[3]L. V. Borovik, analyst.

[4]Pyrrhotite hosted in metasedimentary rocks was not sampled because of its small size.

TABLE 7. Values of δ^{34} for Sulfides (per mil)

Sample no.	Location	Py	Asp	Sp	Bnt	Bul	Stb
V-3947	Metasediments of	-2.7					
V-4006	the Omchak suite	-3.5					
V-4008		-1.2					
V-4009		-5.0					
V-3989	Metasediments of	-5.0					
V-3995	the Atkan suite	-3.0					
V-3954/1	Metasediments of	-3.8					
V-3954/2	the Pioner suite	-4.5					
V-3955		-0.6					
V-3959		-1.1					
V-3960		-0.9					
V-2379		-2.4					
V-2693	Post-ore rhyolite	0.1					
S-637	breccia	1.0					
V-3958	Pre-ore spessartite dike	2.6					
V-3949	Pre-ore chlorite-	-2.9					
V-3956	carbonate-quartz	-0.4					
S-619	veins	-1.5					
V-2372	Metasedimentary	-4.4	-3.7				
V-2596	rock-hosted ore	-4.1	-3.7				
V-2597		-4.4	-3.5				
V-2600			-3.6				
V-2605			-4.1				
V-2614			-3.0				
V-2615				-2.4			
V-2606	Dike-hosted ore		-4.1				
S-675		-4.5	-4.2				
S-677		-4.6					-5.5
S-696		-3.4	-4.2				
V-2662		-3.7	-4.7	-3.8		-5.7	
V-2349					-5.1		
V-2317							-6.3

Abbreviations: Py = pyrite; Asp = arsenopyrite; Sp = sphalerite; Bnt = bournonite; Bul = boulangerite; Stb = stibnite.

similar isotopic compositions and mineral paragenesis.

The results of six isotopic analyses of oxygen in quartz are presented in Table 8. The $\delta^{18}O$ values for the four samples of ore quartz are consistent between 13.9 and 14.1 per mil and are independent of sampling site. The two samples from chlorite-carbonate-quartz veins exhibit slightly lower $\delta^{18}O$ values. The data obtained by Clayton et al. (1972) were used to calculate the $\delta^{18}O$ values for the fluid that was in equilibrium with ore-bearing quartz at 300° C temperatures, corresponding to the highest peak from fluid-inclusion homogenization measurements. The fluid $\delta^{18}O$ values range from 7.1 to 7.3 per mil and are consistent with very homogeneous metamorphic fluid. The $\delta^{18}O$ values of the fluid that was in equilibrium with pre-ore quartz at 300° C range from 5.9 to 6.0 per mil. The small difference between $\delta\ ^{18}O$ values of ore and pre-ore quartz-forming fluids indicates that both types of vein formed from metamorphic fluid or fluid that was in equilibrium with metasedimentary rocks.

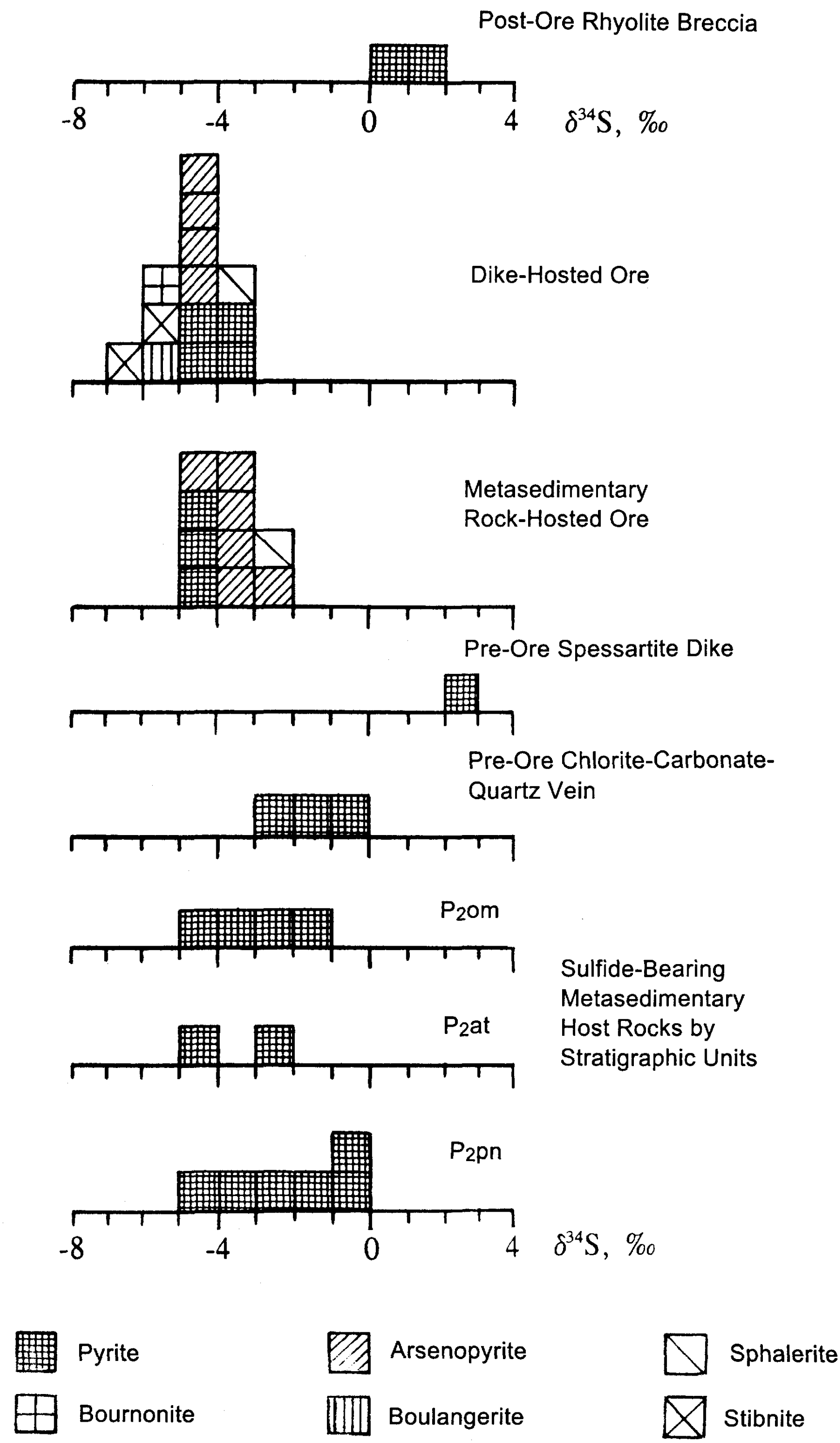

FIG. 8. Sulfur-isotope composition of sulfides from the Omchak mining district.

TABLE 8. Oxygen-Isotope Composition of Quartz

Sample no.	$\delta^{18}O$ (per mil)	Location
S-647	13.9	Omchak deposit, metasedimentary rock host
V-2596	14.1	Natalka deposit, ore zone 45, 650-m level, metasedimentary rock host
V-2606	13.9	Natalka deposit, Uchastkovaya ore zone, 650-m level, spessartite dike rock host
V-4004	13.9	Natalka deposit, ore zone 3, surface, metasedimentary rock host
V-3949	12.8	Pre-ore chlorite-carbonate-quartz vein, metasedimentary rock host
V-3966	12.7	Pre-ore chlorite-carbonate-quartz vein, metasedimentary rock host

Genetic Model

Geologic control

Different viewpoints exist concerning the genesis of the mesothermal gold-quartz deposits of the Yana-Kolyma metallogenic belt. Some of the earliest ideas suggested that gold was related genetically to granite batholiths (Bilibin, 1961a). Later, Bilibin (1961b) indicated that gold mineralization and small pre-batholith diorite intrusions may have formed from the same magma chamber. Other workers interpreted gold-vein formation as a part of a paragenetic sequence of small intrusive events (Apeltsyn, 1956; Matveyenko and Shatalov, 1958), whereas others interpreted gold mineralization to be an exclusively granitoid-related process (Rozhkov et al., 1971; Gamyanin, 1974; Gavrikov, 1976; Sobolev, 1989). In addition, some workers suggested a metamorphogenic character of gold deposits (Serebryakov, 1968; Firsov, 1985), whereas others interpreted gold mineralization to be related to deep mantle degassing (Anikeyev et al., 1976; Gelman, 1976).

Our genetic model is based on the observation that mesothermal Au-quartz deposits occur in thick terrigenous sequences that were subjected to greenschist metamorphism. Orebodies occur mainly at and above the biotite isograd, and also at and above the pyrite-to-pyrrhotite transformation boundary in metasedimentary rocks. The formation of metamorphic biotite and development of pyrrhotite after pyrite are coeval with the early stages of granitoid magmatism. Mesothermal Au-bearing quartz vein deposits of the Yana-Kolyma metallogenic belt pre-date granitoid intrusions (Skornyakov, 1949a; Firsov, 1985) or are post-granitoid, such as the Dorozhnoye and Shkolnoye deposits. These contrasting relationships easily can be explained by the fact that granitoid magmatism and metamorphism were multistage and long-lived (P. N. Anorov, G. M. Yudina, V. G. Shakhtyrov, and S. V. Voroshin, pers. commun., 1991). This interpretation also explains the observations that gold mineralization was not related to any single magmatic complex, either in space or in time, and that the deposits have different relative ages. Some Au-bearing quartz veins have undergone contact metamorphism that was connected with post-accretion granitoids (Skornyakov, 1949b; Firsov, 1957, 1958; Tyukova, 1989).

Mineralization in the Omchak mining district occurred from the Late Jurassic to Late Cretaceous. Ore-bearing veins cut Late Jurassic spessartite dikes, and are cut by rhyolite breccias that are cut by Late Cretaceous granitoids. However, in some gold deposits in adjacent districts (Degdekan and Goltsovskoye), similar spessartite dikes are coeval with ore-bearing veins (Voroshin, 1988). This relation indicates that some mineralization may be Late Jurassic in nature.

Sources of ore and solutions

Terrigenous sedimentary rocks of the Verkhoyansk Complex are the most favorable source for sulfur and arsenic in the orebodies. The sulfur content of these rocks ranges from hundredths to tenths of a percent, and sometimes up to several percent. Arsenic content is tens of g/mt. Sulfur occurs predominantly as iron sulfides (pyrite and pyrrhotite) and an isomorphic arsenic admixture occurs in pyrite. These components were mobilized during metamorphism as a consequence of phase transformations of iron sulfides. The conversion of pyrite to pyrrhotite during metamorphism is well-documented in the Verkhoyansk Complex (Skornyakov, 1947; Izmaylov, 1976; Voroshin et al., 1993).

Pyrite was the main sulfide mineral in metasedimentary rocks below the biotite isograd of greenschist facies metamorphism. At the same time as metamorphic biotite formed, pyrite was altered to pyrrhotite, as indicated by pyrrhotite pseudomorphs of cubic habit. The pyrite-to-pyrrhotite transformation is not simply the incongruent decomposition of pyrite at 743° C (Kullerud and Yoder, 1959). The C/CO_2 buffer controls this reaction at temperatures above about 200° C (Hall, 1986). In the metasedimentary rocks of the Verkhoyansk Complex this reaction occurred at about 400° C and 2.5 kbar, as indicated by the concentration of cobalt and nickel in cogenetic pyrite and pyrrhotite and by the compositions of arsenopyrite and sphalerite that are cogenetic with pyrite and pyrrhotite (Voroshin et al., 1993). These P-T conditions approximately correspond to the biotite isograd (Ferry, 1981, 1984). Simultaneous with the pyrrhotitization of pyrite, decarbonatization and dehydration reactions occurred (Ferry, 1984). The main ore-fluid components—H_2O, CO_2, and H_2S—as well as gold and arsenic were mobilized during these metamorphic events.

Because the pyrite-to-pyrrhotite alteration is important for the release of gold-mobilizing species, quantitative balance computations were made for samples obtained from the Shturm mining district within the Yana-Kolyma metallogenic belt, where both pyrite and pyrrhotite are present (Voroshin et al., 1993). The following data were used for these calculations: 0.1% sulfur in the form of pyrite in sedimentary rocks; 53.5% sulfur, 0.33% arsenic, and 0.18 g/mt Au in the pyrite; 38.7% sulfur, 0.003% arsenic, and 0.04 g/mt Au in the pyrrhotite; 1% sulfur, 2% arsenic, and 10 g/mt Au in the ore. To produce 1 metric ton of ore with these grades, leaching of the following rock volumes are required: 37 metric tons for sulfur, 3000 metric tons for arsenic, and 38,500 metric tons for gold. These calculations assume 100% efficiency for deposition of the S, As, and Au. Different source volumes of terrigenous rocks would be required for mobilization of the three ore components during the reaction of pyrite

Transportation of ore components

As discussed above, the boundary physicochemical parameters for ore formation were: a maximum temperature of 400° C; a maximum pressure of 2.5 kbar, and initial buffering by the pyrite-to-pyrrhotite transformation. The ore solutions evolved to slightly lower temperature and pressure and a higher acidity and oxidation state, as indicated by depletion of the pyrite-pyrrhotite buffer with precipitation of pyrite and chlorite. Under these conditions, the sulfur in solution probably was in the form of H_2S^0 and HS^-; arsenic probably was in the form of $H_3AsO_3^0$ (Heinrich and Eadington, 1986; Ballantyne and Moore, 1988); and gold probably was in the form of thiocomplexes $Au(HS)_2^-$ and $Au_2S(HS)_2^{2-}$ (Seward, 1973). The role of arsenic in the formation of gold complexes is uncertain. The speculative compound $H_2(AuAs)S_3^0$ may have existed (Akhmedzhanova et al., 1988).

Deposition of ore components

The interaction of sulfur with iron-bearing silicates and oxides was the most probable mechanism for sulfide deposition. This sulfidization is most notable as disseminated mineralization in altered wall rocks. The deposition of arsenic, leading to the formation of arsenopyrite, also was part of the sulfidization process. The lower limit of arsenopyrite stability is about 250° C (Kozlov et al., 1986; Tyukova and Kozlov, 1986), which accounts for the absence of arsenopyrite in sedimentary rocks that were altered by lower-temperature metamorphic fluids. In these lower-grade areas, arsenic occurs mainly as an isomorphic admixture in pyrite. Arsenic occurs in the form of arsenopyrite only at temperatures higher than 250° C, and thus provides a minimum T estimate for the beginning of lode gold deposition in the Yana-Kolyma metallogenic belt.

Gold deposition was associated closely with the formation of sulfides, as indicated by paragenetic observations. Possibly, the first precipitated sulfides (including arsenopyrite and pyrite) formed simultaneously with metastable gold-bearing iron sulfide, with subsequent decomposition into pyrite and native gold sure fluctuations, associated with fracturing of vein quartz, also resulted in gold precipitation. These processes were accompanied by loss of gas components and, probably, destruction of the gold complexes (Newberry and Brew, 1988).

Relative geochronology at the Natalka deposit

The geochronology of gold mineralization at the Natalka deposit is interpreted as follows:

1. Deposition of marine sedimentary strata, including ubiquitous syn-sedimentary pyrite and metalliferous organic carbon.
2. Regional greenschist facies metamorphism and deformation of sedimentary rocks during: recrystallization of iron sulfides, with formation of coarse pyrite crystals containing admixtures of As and Au; formation of carbonate concretions; partial oxidation of organic carbon compounds; and formation of metamorphic chlorite-carbonate-quartz veins.
3. Metamorphism related to doming and granitoid magmatism; injection of spessartite and rhyolite dikes; formation of metamorphic biotite and pyrite-to-pyrrhotite alteration in hornfels; mobilization of main ore components; and simultaneous dehydration and decarbonatization of metasedimentary rocks.
4. Migration of ore fluid and deposition of ore from near the biotite isograd during metamorphic events. A lengthy migration was not necessary. In some cases, the front of contact metamorphism, connected with post-accretion granitoids, overlapped ores previously formed during earlier metamorphic events, resulting in multiple episodes of gold deposition in this region.

Similar deposit types

Our model can be applied to many other gold deposits hosted in terrigenous strata. In the Muruntau deposit in Uzbekistan, the main orebodies are associated with pyritized host rocks and the zone of pyrrhotite (after pyrite) is located in sub-ore-grade source areas (Kremenetsky et al., 1990). The transportation of gold-bearing ore fluids to beyond the granitoid hornfels aureoles and the related biotite isograd occurs in the Taymyr Peninsula in Siberia (Zabiyaka et al., 1990). Gold deposits in western Victoria in Australia, hosted in Paleozoic terrigenous rocks, are quite similar to gold deposits in the Yana-Kolyma metallogenic belt. The geologic setting and mineralogy are identical, and the genetic model presented here is applicable to them as well (Phillips and Hughes, 1995). Gold-bearing quartz veins of the Alaska-Juneau gold belt in southeastern Alaska occur in greenschist facies metasedimentary rocks near the greenschist-amphibolite boundary and suggest similar ore genesis (Newberry and Brew, 1988; Goldfarb et al., 1989). The ores and host rocks of the Meguma terrane in Nova Scotia (Sangster, 1990) also suggest that similar ore-forming processes occurred in this region.

Conclusions

The Natalka deposit is a major lode gold mine in Russia, with known reserves and good prospects for the future. The potential of the deposit will be greater, however, if the reported data about high grades of platinum-group elements are true and if these elements can be extracted (Sidorov et al., 1994). The genetic model discussed in this article assumes higher-grade contact metamorphism of host rocks at depth and, probably, ore metamorphism as well. If so, mineralization at depth could be similar to the Igumen lode gold deposit, where gold is redistributed as a result of contact metamorphism and bonanza occurrences with coarse gold are present. Recently, several high-grade ore zones with large native gold grains have been discovered at the 600-m level of the Natalka deposit, which may indicate redistribution and concentration of ore at the Natalka deposit. These high-grade ores seem to be concentrated at deep levels of this deposit. However, the lack of well-defined zonation of mineralization attests to the considerable vertical interval of this deposit.

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