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Geology of the Pyhäsalmi Ore Deposit, Finland

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Abstract

The Pyhäsalmi deposit is a stratabound massive pyrite-copper-zinc deposit in a Precambrian metamorphic schist belt mainly composed of metavolcanites. The host rocks of the ore are silicic fine-grained metavolcanites, and the stratum of ore is stratigraphically overlain by mafic metavolcanites. In the close environment of the ore the silicic metavolcanites, rhyolitic in composition, and the more mafic rocks have been chemically altered; hence, they are enriched in potassium, iron, and magnesium and depleted in sodium and calcium. The metamorphic derivatives of the altered volcanites are sericite schists and cordierite-anthophyllite rocks. The ore deposit, which is syngenetic with silicic volcanites, was folded during polyphase deformation. The orebody averages 0.85 percent Cu, 2.8 percent Zn, 37 percent S, 33 percent Fe, 0.2 g/ton Au, and 14 g/ton Ag. The main gangue minerals are quartz, barite, and local carbonates. The data presented indicate a fairly constant isotope composition of sulfur with an average value of ^{34}S of +7.5 per mil for pyrite. Isotopic dating of intrusive and supracrustal rocks yields ages of about 1,900 m.y. for the metavolcanites but a somewhat higher age (1,932 ± 1.5 m.y.) for the gneissic granite, east of the schist belt. The lead isotope compositions of the galena in the Pyhäsalmi ore and its silicic metavolcanite hosts lie on the same isochron within limits of experimental uncertainty. The lead isotopes applied to a plumbotectonic model yield an age of 1,970 m.y. and indicate a mantle origin.

Introduction

THE Pyhäsalmi mine is located in northern Finland in the commune of Pyhäjärvi, Oulu County, at latitude 63° 39′ 33″ N. longitude 25° 02′ 55″ E. The ore deposit was discovered in August 1958 when a local farmer was digging a well in his yard. Outokumpu Oy lost no time in implementing geological mapping and geophysical surveys in the area with the aid of magnetic, electromagnetic, and gravimetric methods. The electromagnetic and gravimetric surveys indicated the feasibility of drilling, and the first holes were bored in October of the same year. By the summer of 1959 the inventory drilling was concluded, and the sinking of the shaft commenced in late July. The development works for open-pit mining got underway at the same time, the first stage being the removal of overburden from the ore outcrop. The concentrator, which took 2½ years to build went into production in March 1962. After enlargement and automation of the concentrator in 1967, the annual rate gradually increased from 600,000 metric tons to the present 1,100,000 metric tons, which includes 200,000 metric tons of oxidized ore mined earlier and stored in a stock pile. Production started from the open pit at the same time as preparations were being made for underground mining. In 1976 the mining was shifted entirely underground.

The deposit is estimated to average 0.85 percent Cu, 2.8 percent Zn, 37 percent S, 33 percent Fe, 0.2 g/metric ton Au, and 14 g/metric ton Ag. Furthermore, the ore assays 4.9 percent BaO and 0.06 percent Pb with average of 15 ppm Se.

General Geology

The Pyhäslami ore deposit is located in the Vihanti ore zone (Kahma, 1973). This zone is thought to belong to the Svecofennian formations of the Svecokarelian orogeny. Nevertheless, as stated by Wilkman (1931) in his explanation to the geological map, some of the schists may be associated with inclusions and gneisses of the Presvecofennian basal complex. Migmatitic, granitized, and tectonized mica and partly also hornblende gneisses predominate, exhibiting no relicts that could give any hint of their origin. There are, however, some places with better preserved schists that, owing to their structure, can be recognized as metamorphosed acid or basic effusives and associated with them are tuffites, agglomerates, and pillow lavas. Some of the schists seem to be weakly sorted weathering products derived from the forementioned volcanics. However, occurrences of graphite-bearing graywacke schists suggest that volcanic activity was accompanied by deposition of clastic sediments (Huhtala, 1979). Essential constituents in the zone are pegmatite and diabase dikes and plutonic intrusives of various composition, structure, and age.

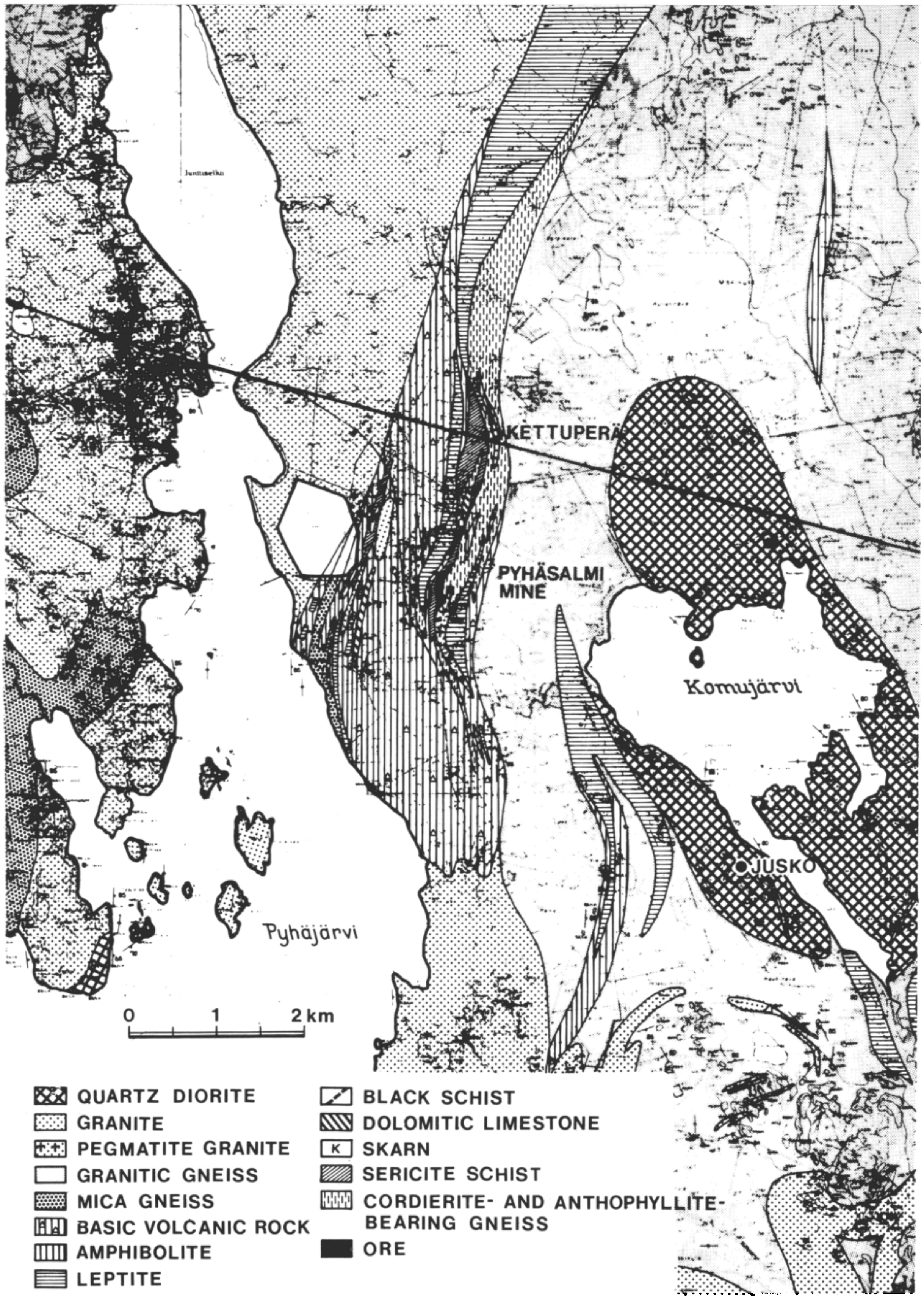

FIG. 1. Geology of the Ruotanen schist belt.

TABLE 1. Chemical Composition of

	1	2	3	4	5	6	7	8	9	10
SiO_2	79.9	80.1	79.3	75.3	81.6	72.8	70.1	71.4	74.0	52.4
TiO_2	0.17	0.17	0.17	0.26	0.09	0.21	0.29	0.22	0.02	0.76
Al_2O_3	11.2	10.7	11.5	13.0	10.1	12.74	12.41	12.04	14.5	14.4
FeO	2.33	1.52	2.50	2.29	1.52	3.14	3.91	2.93	1.46	11.91
MnO	0.05	0.07	0.06	0.05	0.05	0.04	0.04	0.03	0.02	0.23
MgO	0.57	0.19	0.39	0.83	0.32	0.95	0.68	0.46	0.11	6.4
CaO	1.35	2.02	0.59	1.41	2.01	2.28	1.90	2.07	0.30	8.5
Na_2O	3.00	3.19	4.58	3.81	2.71	3.71	4.56	5.91	4.38	2.93
K_2O	1.36	1.29	1.59	1.70	1.36	1.41	1.54	0.97	3.87	1.02
FeS_2	0.24	1.41	0.15	0.61	0.04	0.00	0.06	0.04	1.83	0.10
	100.17	100.66	100.83	99.26	99.80	97.88	95.49	96.04	100.49	98.65

1, Leptite; Pyhäsalmi, diamond drill hole (ddh) 244, depth 149.70 m.
2, Leptite, quartz-rich; Pyhäsalmi, ddh 244, depth 110.06 m.
3, Leptite, porphyrite; Pyhäsalmi, ddh 84, depth 67.85 m.
4, Leptite, inclusion in the massive ore; Pyhäsalmi, ddh 75, depth 47.95 m.
5, Leptite, quartz-rich; Pyhäsalmi, 500-level.
6, Leptite gneiss; Kettuperä ddh 12.
7, Granitic gneiss; Kettuperä, ddh 12.
8, Granitic gneiss; Kettuperä, ddh 18.
9, Pegmatite; Pyhäsalmi, ddh 238, depth 69.70 m.
10, Amphibolite, interbedded with leptite; Pyhäsalmi, ddh 84, depth 34.50 m.
11, Amphibolite; road cut, Pyhäsalmi.

The Ruotanen schist belt

The part of the schist area that includes the Pyhäsalmi ore deposit is called the Ruotanen schist belt (Fig. 1). It trends almost north-south and dips vertically or subvertically eastward. The fold axes and lineation plunge 40° to 60° south. In the east the schist zone is bordered by a domelike granite gneiss and in the west by a porphyry granite that is younger than the schists.

The Ruotanen schist belt is poorly exposed and hence the concept of it is largely based on geophysical survey and drilling data. The Pyhäsalmi ore deposit is located roughly in the middle of the schist belt, which is at least 30 km long. For lack of exposures the structure has not been established in detail. Stratigraphy of the schist belt has been dealt with by Helovuori (1977) and Huhtala (1979).

Lithologically the schist belt can be divided into two major associations: the amphibolite association and the leptite association. In the present context the term leptite refers, as it does in the Orijärvi area in southwestern Finland and the Bergslagen province in central Sweden, to fine-grained, acid, predominantly volcanic schists rich in quartz and feldspar (Eskola, 1914; Simonen, 1960; Geijer and Magnusson, 1944). The rocks of the amphibolite association occur in the western part of the schist belt and the leptite association in the eastern part. The ore deposit is associated with the leptites.

South of Pyhäsalmi the belt is composed predominantly of the amphibolites. Northward, however, the leptite association gradually becomes predominant over the amphibolites. The boundary between the two rock units is gradual as is revealed by the occurrence of amphibolitic portions in the leptite and of narrow leptite zones in the amphibolite in the transitional zone between the two lithologic complexes.

Amphibolite, the predominant rock type in the amphibolite complex, shows a banded structure due to the variation in abundance of the main minerals hornblende and plagioclase and of quartz and cummingtonite. Amphibolites with primary volcanic features are considered as mafic tuffs, agglomerates, or lavas in origin alternating with ophitic and banded amphibolites. The ancient mafic volcanites are particularly abundant in the amphibolite zone in the surroundings of the Pyhäsalmi mine. With the increase in micas and quartz the amphibolites grade into mica gneisses, some of which are sheared amphibolites and some intermediate tuffites in origin.

Plagioclase porphyrites frequently occur as intercalcations or crosscutting veins in agglomeratic and tuffitic amphibolites, showing occasionally anhydrite nodules. Uralite porphyrite has been encountered in a few places.

Impure dolomitic limestones and skarns occur as interlayers in the western margin of the amphibolite belt. Besides dolomite these rocks contain diopside, tremolite, serpentine pseudomorphs after olivine, and plagioclase as well as occasional garnet, quartz, chlorite, biotite, and titanite.

Also associated with carbonate rocks are graphite-bearing mica gneiss interlayers that often show moderate pyrrhotite dissemination. These "black

Rocks from the Pyhäsalmi Field

11	12	13	14	15	16	17	18	19	20	21	22
52.5	49.2	77.2	75.7	77.4	74.8	75.3	42.0	57.3	56.1	69.3	49.1
0.52	1.07	0.21	0.18	0.18	0.09	0.14	0.49	0.38	0.58	0.13	0.34
16.4	16.9	11.7	11.1	11.6	11.51	12.55	12.6	15.74	14.89	9.7	14.4
9.25	6.38	1.09	2.72	2.87	1.83	1.56	13.98	11.91	9.43	2.35	7.54
0.19	0.23	0.01	0.14	0.09	0.03	0.02	0.34	0.10	0.14	0.09	0.32
5.2	6.8	0.29	3.6	3.5	3.12	0.77	11.6	8.40	8.25	10.8	15.41
9.2	8.0	0.03	0.04	0.47	0.08	0.03	0.75	0.11	0.80	1.18	3.18
4.69	3.55	0.26	0.19	0.22	0.20	0.30	0.30	0.15	0.69	0.65	0.67
0.49	0.92	3.48	2.61	2.28	2.82	3.56	1.93	0.63	2.08	0.99	0.95
0.30	3.98	5.67	1.62	2.12	1.67	3.22	14.45	5.26	1.89	0.47	2.21
98.74	97.03	99.94	97.90	100.73	96.15	97.45	98.44	99.98	94.85	95.66	93.86

12, Plagioclase porphyrite; Pyhäsalmi.
13, Sericite schist; Pyhäsalmi, ddh 244, depth 10.40.
14, Cordierite-bearing sericite schist; Pyhäsalmi, ddh 244, depth 44.10.
15, Cordierite-bearing sericite schist; Pyhäsalmi, ddh 244, depth 63.60.
16, Sericite schist (lapilli tuff); Kettuperä.
17, Sericite schist; Kettuperä.
18, Cordierite-antophyllite rock; Pyhäsalmi, 85-level.
19, Cordierite-antophyllite rock; Kettuperä, ddh 18.
20, Cordierite-antophyllite rock; Kettuperä, ddh 17.
21, Cordierite-biotite rock; Pyhäsalmi, 125-level.
22, Cordierite-biotite rock; Kettuperä, ddh 18.

schists" produce intense geophysical anomalies in the western part of the schist belt. Mica gneisses free from graphite are also met with in the same area.

The main minerals in the leptites are quartz and plagioclase. The rock also contains variable amounts of potassium feldspar, biotite, muscovite, chlorite, and epidote. Leptites are often banded, and quartz occurs as aggregates or grain clusters, as in porphyry leptite. Leptites with euhedral corroded quartz grains have been encountered as inclusions in the ore. The structure of the porphyry leptites is a relict of the primary volcanic structure, whereas the banded or homogeneous leptites may be derived from clastic sediments whose material is predominantly volcanic.

As biotite increases, the leptites grade into mica gneisses that occur as intercalations in the leptite formation. In some places the mica gneisses are blastoporphyritic in texture, which points to a tuffitic origin.

The altered rocks include sericite schists that occur in the leptites adjacent to the ore deposit and in cordierite gneisses and cordierite-anthophyllite rocks east of the leptite zone. Quartz and sericite are the predominant minerals in sericite schist, but the rocks also contain variable amounts of plagioclase. Banded pyrite dissemination is characteristic of the rock; cordierite and sillimanite are occasional constituents. Sericite schists, which occur in association with the ore, show a fold structure at the southern end of the orebody which also conforms with the orebody.

North of the ore the sericite schist horizon continues persistently for at least 2 km, although the degree of sericitization gradually diminishes. In general the sericite schists are more intensely foliated than the unaltered leptites and also are more coarse grained and contain unaltered leptitic interlayers.

Cordierite gneisses and cordierite-anthophyllite rocks occur east of the ore deposit and the mineralized part of the leptites. The amphibolites and leptites again appear east of the cordierite-bearing rocks, close to the eastern contact of the Ruotanen schist belt. Although the cordierite gneisses vary in mineral composition, the most typical association is cordierite, quartz, biotite, and muscovite. Plagioclase and potash feldspar are usually lacking. Brown tourmaline, sillimanite, or garnet and spinel are encountered occasionally. These rocks often show a banded structure, but the primary volcanic or sedimentary structures are lacking.

The cordierite-anthophyllite rocks occur as nodules or lenses in the amphibolite that contains leptite interlayers. The contacts with the leptites are sharp. Laterally along the layering the cordierite-anthophyllite rocks often grade into anthophyllite-bearing amphibolites. The minerals in the rocks are cordierite, anthophyllite biotite, quartz, and plagioclase. Associated with the cordierite-anthophyllite rocks are elongated lenses of mica rocks with biotite, phlogopite, chlorite, talc, sericite, and quartz as mineral constituents. In places the mica rocks occur in intensely sheared zones, and they are also often encountered in the footwall contact of the ore. Leptites and sericite schists occasionally show a network of mica rock veins.

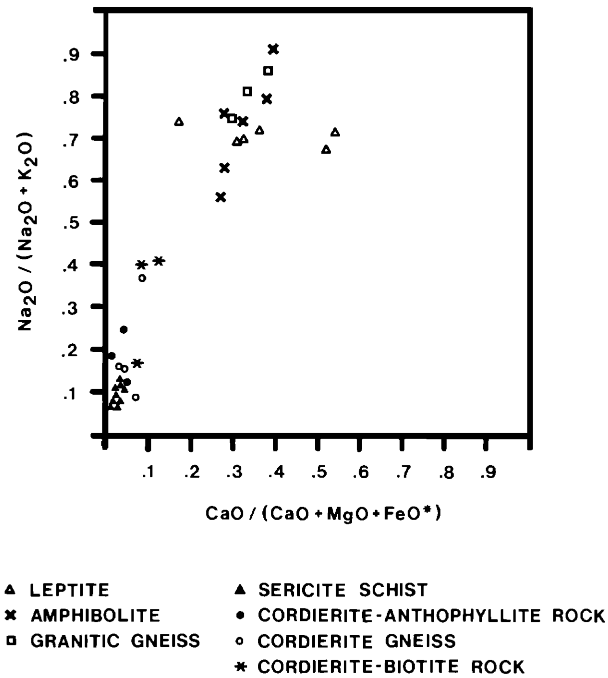

FIG. 2. A diagram showing the compositions of unaltered and altered rocks of the Pyhäsalmi area.

Plutonic rocks

In the east the schist belt is bordered by a granite gneiss body called the Kettuperä granite gneiss in this context. The contacts between the schist belt and the granite gneiss are not exposed, although at Kettuperä, north of Pyhäsalmi, the contact zone of the leptite has been intersected by drilling, showing that rather coarse-grained gneissose leptite occurs as conformable bands in the foliated granite gneiss. No clear intrusive structures have been noted. In the middle of the Kettuperä granite gneiss body, around Komujärvi, there is a less intensely foliated plutonic complex that is composed mainly of quartz diorite. This rock suite, held to be a synkinematic intrusion, is represented by the sample taken from the Jusko area and whose dating will be discussed in a later section.

The granite west of the schist belt is microcline granite that obviously crosscuts the belt. It is considered as postorogenic. Pyroxene diorite and granodiorite have been encountered as veins in the schist belt. The predominant vein rock is coarse-grained pink microcline pegmatite, which occurs as a sizable body, e.g., immediately east and south of the orebody in contact with it. Aplite and quartz tourmaline veins are also common, whereas only a few diabase dikes have been met with.

Geochemistry of the rock types

The chemical composition of some rock types in the Pyhäsalmi area are given in Table 1. The metavolcanites seem to differ only slightly from those reported by Huhtala (1979) from the nearby Pyhäsalmi-Pielavesi district. The amphibolites, often with their volcanic primary structures preserved, correspond to andesites or locally to basalts in composition. Likewise, the leptites resemble rhyolites or rhyodacites; they can be compared with the acid metavolcanites of the leptite belt in southwestern Finland described by Eskola (1914) and Latvalahti (1979). According to the classification by Irvine and Bragar (1971), the volcanic suite belongs to the calc-alkaline rock series.

In unaltered volcanogenic rocks the abundance of sodium always exceeds that of potassium. The same holds in the granite gneiss of Kettuperä, east of the Ruotanen schist belt. The medium-grained gneissic leptite of Kettuperä, which occurs at the eastern margin of the schist belt with the granite gneiss, does not differ notably from the granite gneiss in composition except for the slightly higher silica and aluminum contents in the gneissic leptite. These features probably indicate that the gneissic leptite derived from slightly weathered granite gneiss.

The difference in composition between the altered rocks adjacent to the ore, i.e., sericite schists and cordierite-bearing rocks on the one hand and leptites and amphibolites on the other, are clearly shown in a diagram (Fig. 2) that plots the alkali ratio as against the CaO/(CaO + MgO + FeO) ratio. The rocks close to the ore show a gain in potassium in

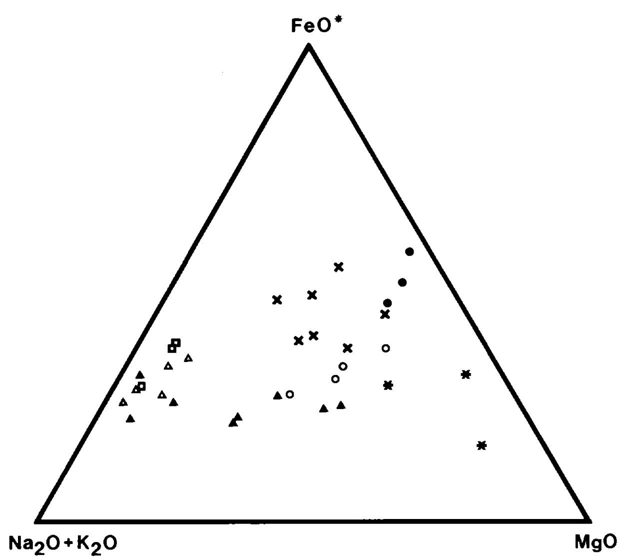

FIG. 3. AFM diagram showing the compositions of the rocks from the Pyhäsalmi area. Symbols as in Figure 2.

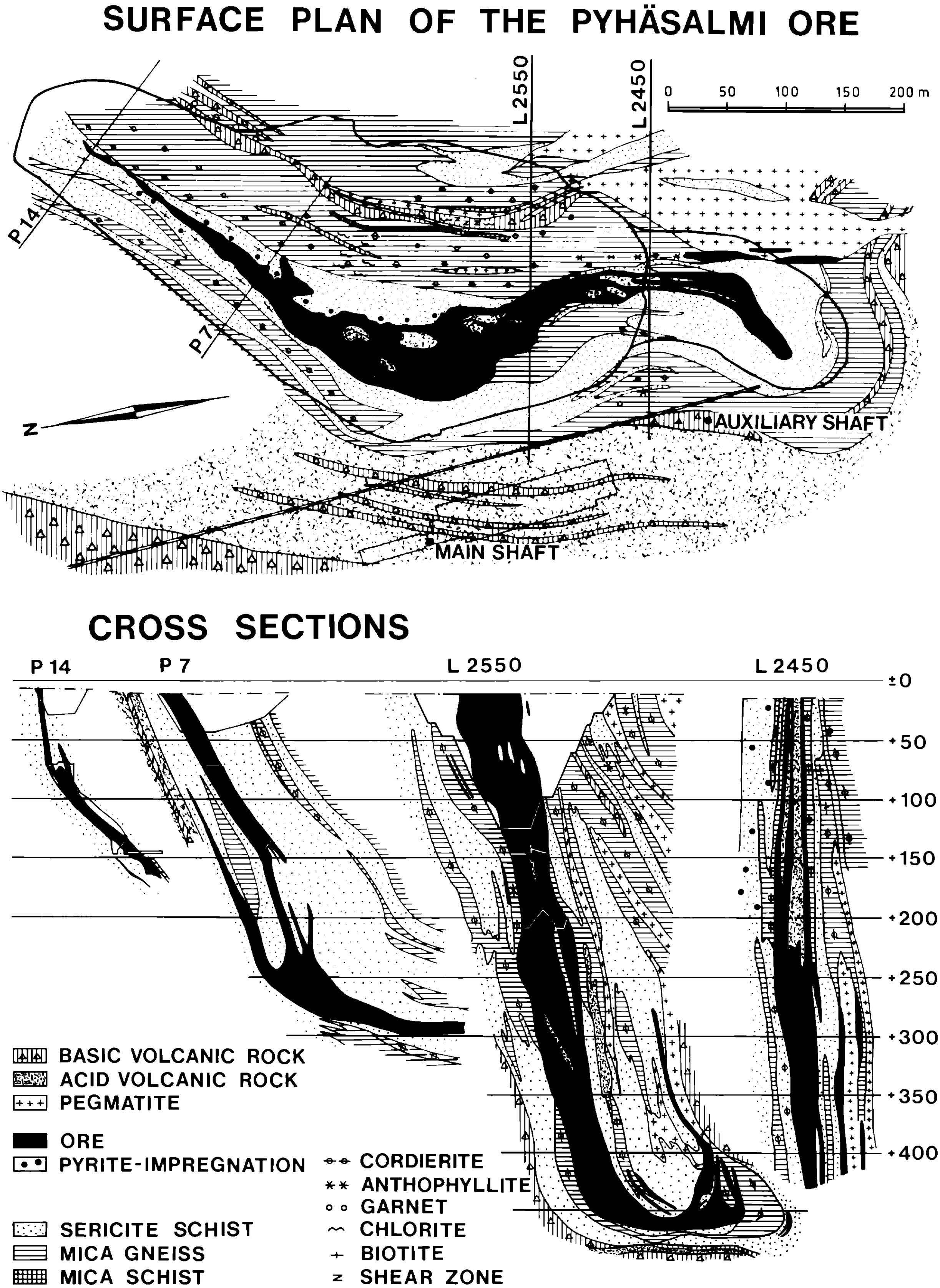

FIG. 4. Surface plan and cross sections of the Pyhäsalmi ore deposit.

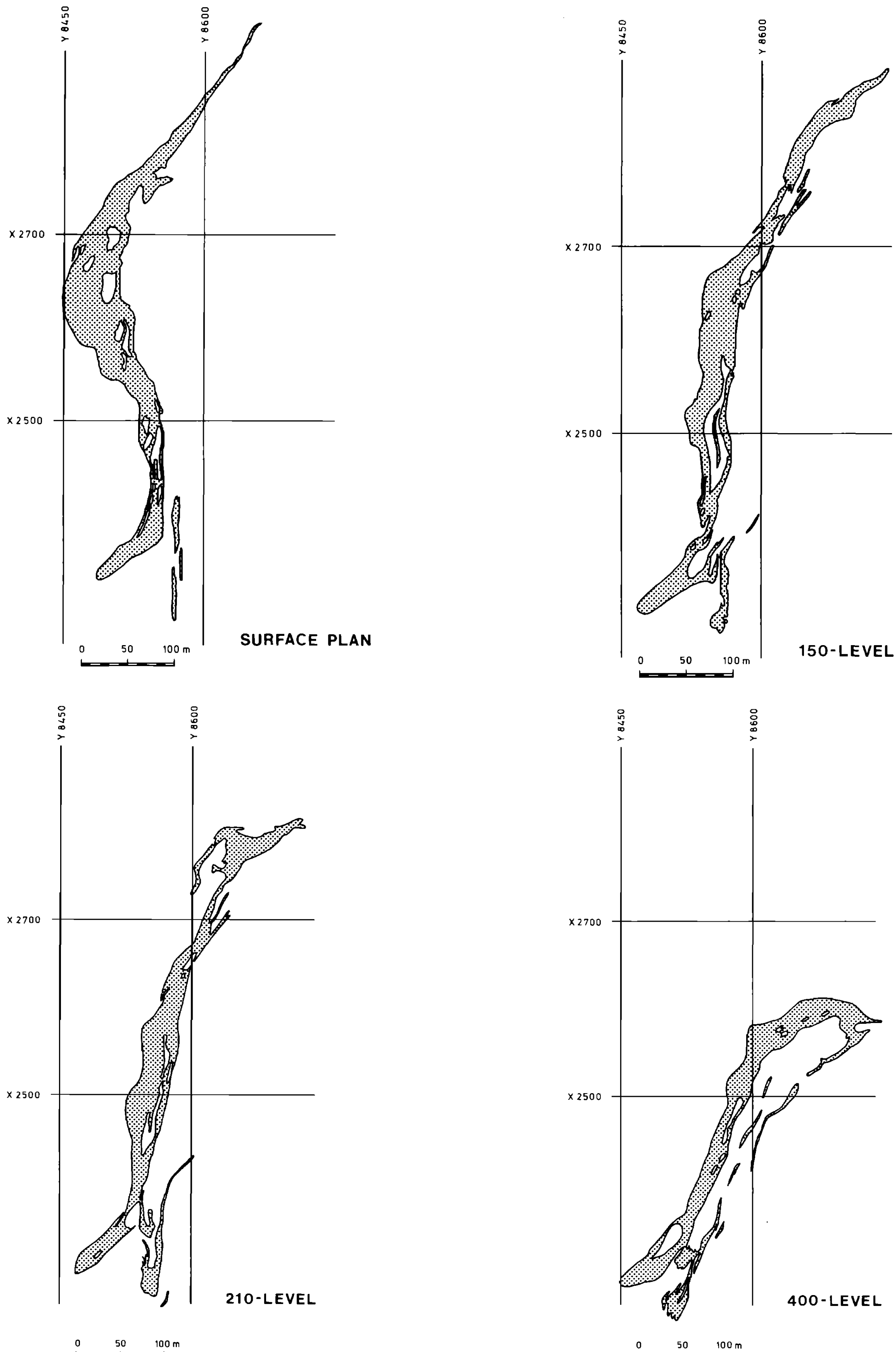

FIG. 5. Horizontal sections showing variations in the shape of the orebody.

relation to sodium accompanied by a loss in calcium in relation to iron and magnesium. The sericite schists were evidently derived from the leptites, and the cordierite-anthophyllite rocks from the amphibolites and biotite hornblende gneisses as a result of changes in chemical composition. The ternary diagram (Fig. 3) demonstrates that cordierite gneisses and cordierite-biotite rocks gained some MgO in comparison with FeO. In the cordierite-anthophyllite rocks, however, the MgO/FeO ratio is approximately the same as in amphibolites. The sericite schists also show a slight gain in SiO_2 compared with the leptites, although the change is more distinct in the alkali ratio and relative loss in calcium. The AFM diagram (Fig. 3) further suggests that, compared with the leptites, sericite schists have lost some alkalies and gained Fe and Mg.

The Ore Deposit

The Pyhäsalmi ore deposit is composed of massive pyrite ore that contains variable amounts of chalcopyrite and sphalerite, and some accessory sulfides and oxides. The main gangue minerals in the sphalerite-predominant ores are quartz and barite; in addition calcite is occasionally encountered (Papunen, 1967). On the surface plan the orebody is about 650 m long and up to 80 m wide. It extends down to a depth of over 600 m; hence the vertical dimension is the dominant one (Fig. 4).

The ore occurs in a sericite schist zone and is conformable with its environment, the structures of the wall rocks being reflected in the forms of the ore deposit. The sericite schist wall rock has been traced northward for several kilometers, and at Kettuperä, about 2 km north of the Pyhäsalmi ore deposit, the sericite schist shows a low-grade sphalerite occurrence. As shown in Figures 4 and 5 the northern end of the ore gradually pinches out above the +150 level, but deeper down it forms a syncline on whose western flank the bulk of the ore is located. The axis of the syncline plunges 40° to 50° south (A_1 in Fig. 13). In the south the ore ends at a fold whose axis plunge 70° to 90° south (A_2 in Fig. 13). The horizontal plan of the deeper levels displays the "eastern parallel orebody," which in vertical section appears to be on the eastern flank of the syncline. In cordierite-antophyllite rocks and in sericite schist this is only a weakly mineralized zone, but at the northern and southern ends it forms massive continuations of the main orebody. The eastern parallel body is characterized by pyrrhotite-rich mineral assemblages. The shape of the orebody is very complicated and evidently a result of polyphase deformation and local remobilization of the sulfides. In the south a coarse microcline pegmatite intersects the ore zone, which means that the southern end is quite varied in form.

As a rule the contacts of the massive ore are sharp. In places, however, the pyrite abundance gradually increases in sericite schist toward the massive ore, but even then the boundary against the massive ore is sharp. The sericite schist always contains pyrite impregnations and veins.

The pre- and postglacial weathering in the fracture zones and areas of intense jointing has resulted in supergene alterations in the subsurficial parts of the ore. Pyrrhotite has altered into melnikowitic pyrite and marcasite. Chalcopyrite has oxidized into chalcocite, covelline, and, occasionally, bornite. In places, pyrite has disintegrated into goethite. The sulfates are pisanite and brochantite.

The massive ore contains different types of wall-rock fragments from a few cubic centimeters to several thousands of cubic meters in size. Some of the fragments occur randomly in the ore, but some favor a certain zone parallel to the longitudinal axis of the ore. They can sometimes be recognized as layers with a distinct boudinage structure. Most of the fragments are acid volcanics, a number of which are well-preserved quartz-porphyry in structure. Some fragments are amphibolites and plagioclase-porphyry. Partly rounded or angular limestone fragments form a zone near the footwall contact, in a place where the zinc content of the ore is above average. Sericite quartzites and sericite schists seldom occur as fragments. Cordierite and tourmaline crystals occur in the ore as inclusions.

Ore types

On the basis of its structure and latest formation stage, the Pyhäsalmi ores can be classified into the following types:

Normal ores: They are massive, coarse pyrite varieties containing variable amounts of sphalerite and chalcopyrite. The accessories are galena, arsenopyrite, magnetite, and fahlores. Associated with the ore are barite and calcite that are, however, mainly confined to zinc-rich variants only. The ore often exhibits a layered structure due to the concentration of sphalerite into bands or layers and the parallelism of the relict sulfide bands in the mineralized sericite schist country rock is suggestive of primary layering. Another predominant structural feature besides banding is the breccia structure with abundant fragments of wall rocks mentioned earlier.

Disseminated ores: These occur in the transitional zone in the margins of the normal massive ores against the unmineralized country rock. They are replacement ores in which sulfides replace the minerals of the host rock.

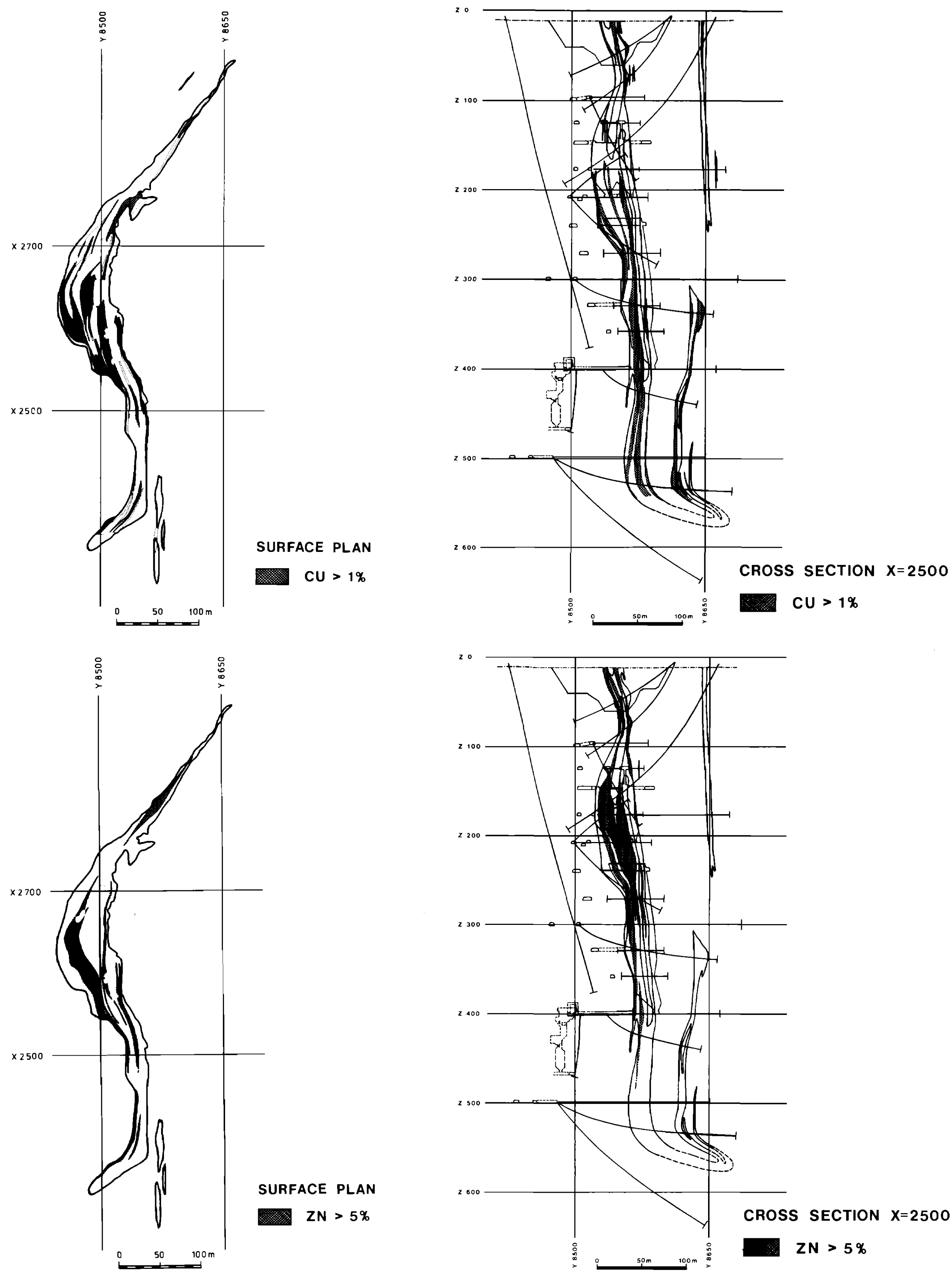

FIG. 6. Horizontal and vertical sections showing variations in the copper and zinc contents of the orebody.

Pyrrhotite-bearing ores: They are mainly restricted to the southern part of the deposit. Unlike the southern end, the occurrence of pyrrhotite in the central and northern parts of the deposit is random, varying from ubiquitous pyrrhotite veins in massive pyrite ores to pyrrhotite accumulations and disseminations. Mapping and microscopic observations show that the ore underwent intense fracturing and tectonic movement before deposition of the pyrrhotite phase.

Pyrrhotite selvage: This occurs in the massive ore in the eastern and southeastern parts of the massive orebody, where pegmatite is in contact with the massive ore. The contacts of the pyrrhotite ore with pegmatite on the side and normal pyrite-rich massive ore on the other are very sharp. The ore contains some chalcopyrite and random sphalerite as well as pyrrhotite, but pyrite occurs only locally as euhedral crystals. Contact relations indicate that the pyrrhotite ore was formed from normal massive pyrite ore by the emplacement of pegmatite into the orebody, presumably by the thermal effect of the pegmatite.

Pyrite: It occurs as rounded or fractured phenocrysts, 5 to 10 mm in diameter, in the porphyritic ore type. The fine-grained matrix consists of pyrite, sphalerite, and random chalcopyrite. The gangue is mainly quartz and oligoclase. There are also porphyritic ores in which the matrix is rich in chalcopyrite. The porphyritic ores were originally normal massive pyrite ores that obtained their present structure in movements following crystallization. These movements occurred close to the contact of the ore with the country rock, adjacent to the large country rock fragments in the ore and in the middle of the ore along a fault parallel to its major axis. In dimension these movement zones are narrow, no more than 10 to 50 cm in width. The porphyritic type is often rich in sphalerite.

Accumulations of ore minerals: These occur at the boundaries between ore and country rock and at the contacts of the country rock fragments with the ore. They differ markedly in composition from the ore types described above. The major constituents in these ores are minerals that in the massive pyrite ores are accessories, e.g., galena, arsenopyrite, various arsenosulfides, tellurides, molybdenite, and native silver and gold. These minerals have also been met with as stringers and accumulations in the country rock close to the contacts. The most common mineralizations of these types are chalcopyrite veins and accumulations that often contain euhedral tourmaline, scapolite, diopside, and tremolite. The occurrence of the veins and accumulations in relation to the normal compact pyrite ore shows that they were formed in the final stage of the ore-forming process. It can, however, also be demonstrated that some of the chalcopyrite mineralizations were produced by remobilization due to tectonic movements after the crystallization of the main ore.

TABLE 2. Chemical Composition of the Pyhäsalmi Ore (wt percent)

Fe	33.68	Mo	0.001	SiO_2	7.33
Zn	4.06	BaO	4.87	Ti	0.02
Cu	0.89	As	0.03	Al_2O_3	2.00
Pb	0.06	S	38.90	MgO	1.35
Cd	0.015	SO_4	2.55	Mn	0.05
Co	0.005	H_2O	1.27	CaO	1.41
Ni	0.003	CO_2	1.15	K_2O	0.24
Sn	0.001	P_2O_5	0.03	Na_2O	0.37

Sulfides in the country rocks: They occur predominantly as pyrite disseminations in sericite quartzites. The sericite quartzite and sericite schist fragments with pyrite dissemination in the massive ore, and the veins of compact pyrite that cut the sericite quartzite country rocks, indicate that the pyrite mineralization in sericite quartzites was well in advance of the formation of the massive pyrite ore.

Tourmaline-quartz veins: These are encountered in acid volcanics and sericite quartzites. Tourmaline has been recognized in these rocks and in massive pyrite ores and chalcopyrite accumulations. Accessory tourmaline appears to be characteristic of the whole environment of the ore deposit.

Geochemistry of the ore

The chemical composition of the Pyhäsalmi ore, indicated by a representative general sample collected at the exploration stage, is shown in Table 2.

According to the analytical data in Table 2 the mineral composition of the ore is:

Barite	7.42
Pyrite	63.48
Pyrrhotite	4.44
Sphalerite	7.04
Chalcopyrite	2.54
Total	84.92

The copper and zinc abundances vary greatly, both horizontally and vertically, the latter being the direction of plunge of the orebody (Fig. 6). The average precious metal contents in the ore are low: Au = 0.2 ppm and Ag = 14 ppm. As stated previously, the heterogeneous distribution of the precious metals is due to the occurrence of Au and Ag minerals at the contacts between the wall-rock inclusions and the ore. Likewise arsenic, lead, antimony, bismuth, and tellurium also show higher abundances around the inclusions. The selenium content in the ore is a fairly constant 0.001 percent. The barite and sphalerite abundances are positively correlated, hence, the highest Ba contents are generally encountered in the portions richest in zinc (Papunen, 1967). The mercury content is low in the ore. The mean annual copper concentrate assays 12 ppm Hg, the zinc con-

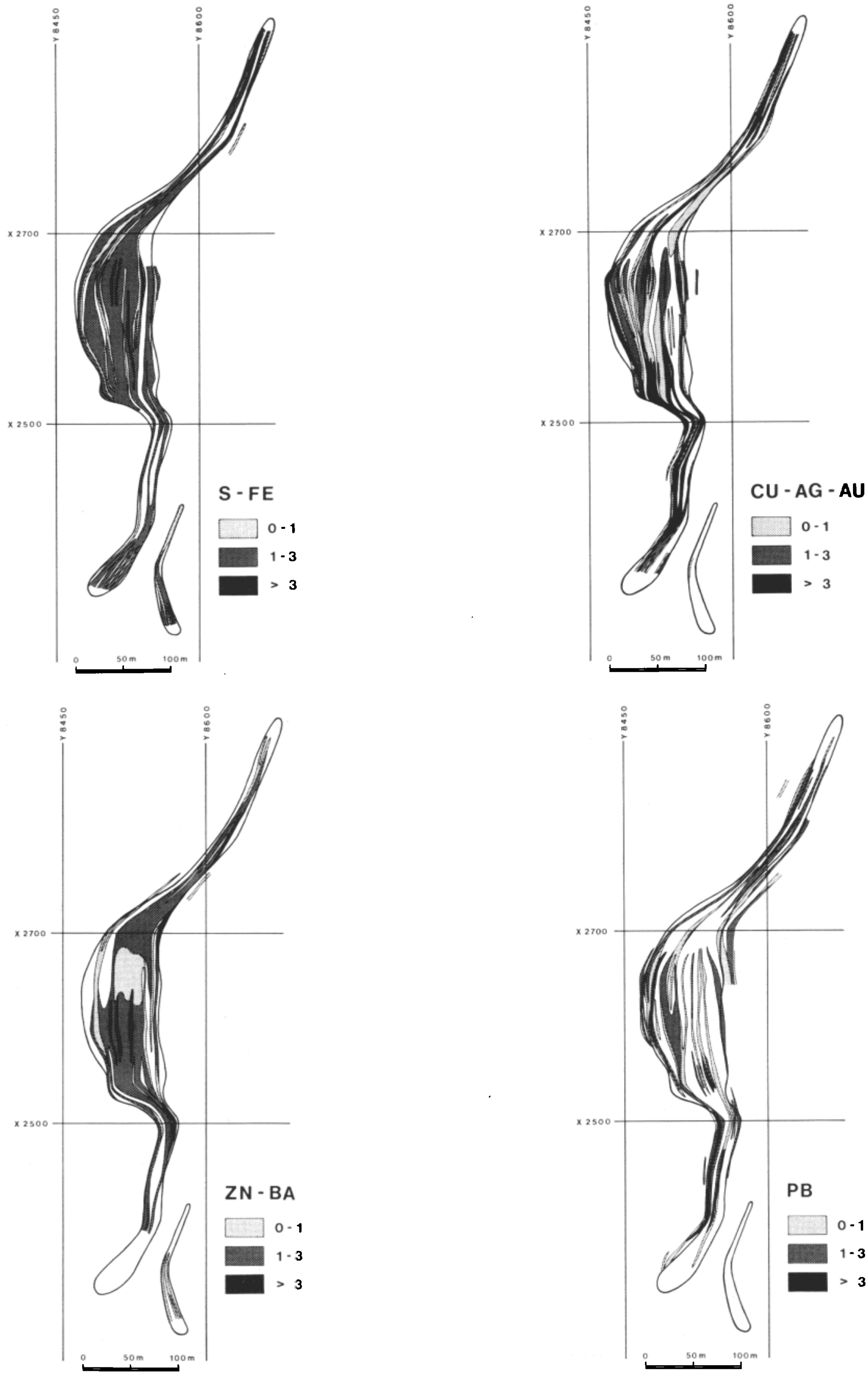

FIG. 7. Distribution of the four factor scores on the +75 level of the orebody.

TABLE 3. Correlation Coefficients (704 samples)

	log Cu	log Zn	log S	log Ba	log Pb	log Ag	log Au	log Fe
log Cu	1.00000	0.21196	0.32678	0.14417	0.06695	0.76700	0.31837	0.36787
log Zn	0.21196	1.00000	0.36745	0.74854	0.14442	0.22553	−0.10574	0.26893
log S	0.32678	0.36745	1.00000	0.34795	−0.29566	0.18404	−0.20922	0.95580
log Ba	0.14417	0.74854	0.34795	1.00000	0.07862	0.09719	−0.12113	0.27864
log Pb	0.06697	0.14442	−0.29566	0.07862	1.00000	0.42304	0.28691	−0.32607
log Ag	0.76700	0.22553	0.18404	0.09719	0.42304	1.00000	0.55849	0.18960
log Au	0.31837	−0.10574	−0.20922	−0.12113	0.28691	0.55849	1.00000	−0.17849
log Fe	0.36787	0.26893	0.95580	0.27864	−0.32607	0.18960	−0.17849	1.00000

centrate at 75 ppm Hg, and the pyrite concentrate at 4 ppm Hg.

Factor analysis

The factor analysis was based on analytical data obtained from core samples drilled in 1958 and 1959. The factor score maps were drawn with the aid of 13 drill sections 50 m apart. The samples, totaling 704, were assayed for Cu, Zn, S, Ba, Pb, Ag, Au, and Fe. The R-mode factor analysis was performed by an IBM 360/40 computer at the Computer Center of the Outokumpu Oy (Häkli, 1970).

Table 3 shows that copper correlates strongly with silver but rather weakly with gold. Gold shows a moderate correlation with silver. The correlation between zinc and barium is fairly high and that between sulfur and iron very high. Lead shows a rather low correlation with silver.

The factor analysis of the correlation matrix shows that four factors account for 88.8 percent of the variations in the eight response variables (Table 4).

Factor 1 (S-Fe factor) has high positive loadings of iron and sulfur and may thus be called the pyrite factor.

Factor 2 (Cu-Ag-Au factor) shows high positive loadings of copper, silver, and gold, which permits the factor to be called the precious metal or sulfosalt factor.

Factor 3 (Zn-Ba factor) has high loadings of barium and zinc and may thus be considered the paragenesis factor.

Factor 4 (Pb factor) shows a high positive loading of lead and a rather low positive loading of silver.

The factor scores of the 13 drill sections were projected onto the +75 level and contoured (Fig. 7). All the factor score maps show a weak zoning and a distinct banding, which could be interpreted as a stratabound structure. The same is also evident in the underground geology and analytical data of the deposit. The positive S-Fe factor scores are homogeneously distributed in the orebody and indicate the area occupied by the massive pyrite ore. The distribution of the Cu-Ag-Au factor scores demonstrates that chalcopyrite, sulfosalts, and precious metals are concentrated not only in the margins of the orebody but also in areas rich in wall-rock inclusions. Apart from at the southern end of the orebody the positive Zn-Ba factor scores are fairly homogeneously distributed throughout the whole orebody. The distribution of the positive Pb factor scores illustrates the preference for galena and some accessory minerals (bournonite, andorite, jordanite, native Ag) to be at the margins of the orebody and in those areas richest in wall-rock inclusions. The mode of distribution is analogous to the Cu-Ag-Au factor scores.

Sulfur isotopes

A set of ore and wall-rock samples was analyzed for sulfur isotopes by P. Hautala of Outokumpu Oy, Exploration Department (at the University of Utah, Salt Lake City). The samples originated from the open pit and from a drill core intersecting the ore zone. The results are presented as a frequency dia-

TABLE 4. Rotated Factors and Communalities (4 factors)

	1	2	3	4	Communalities
log Cu	0.38580	0.78780	−0.0997	0.04053	0.78111
log Zn	0.18287	0.06210	−0.90577	0.12167	0.87251
log S	0.93030	0.04907	−0.23402	−0.12573	0.93844
log Ba	0.13808	−0.00282	−0.93151	−0.00241	0.88680
log Pb	−0.25373	0.18919	−0.10009	0.92735	0.97016
log Ag	0.19932	0.85894	−0.08846	0.36920	0.92163
log Au	−0.34747	0.81431	0.08893	−0.01051	0.79187
log Fe	0.94370	0.08871	−0.13980	−0.15918	0.94333
Eigenvalues	2.86318	2.20932	1.43488	0.59855	
Cumulative percentage of eigenvalues	35.79	63.41	81.34	88.82	

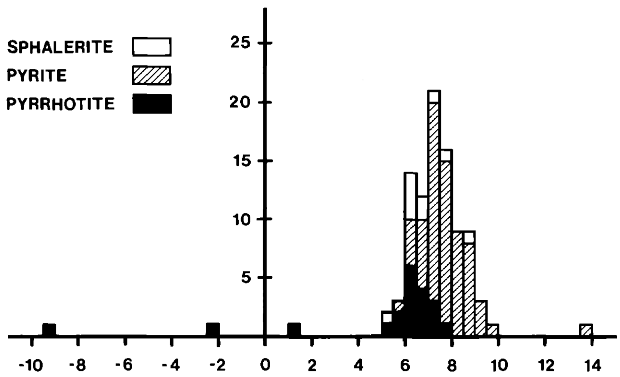

FIG. 8. Frequency distribution of the $\delta^{34}S$ values of different sulfide minerals.

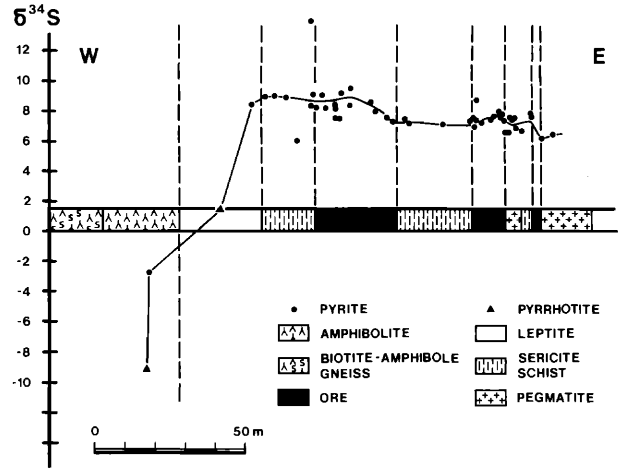

FIG. 9. A drill core section (+520 level, cross section X = 2,500) of the ore showing the variation of the $\delta^{34}S$ values of pyrite and pyrrhotite.

gram and drill core profile (Figs. 8 and 9). Because of some uncertainty in the results (P. Hautala, pers. commun.) the $\Delta\delta\ ^{34}S$ values between coexisting sulfide minerals were not used as temperature indicators.

The $\delta^{34}S$ values of pyrite are presented as a profile (Fig. 9) which indicates the overall, fairly constant isotope composition of the massive ore and the sericite schist wall rocks. The average value of the ore zone, measured from pyrite, is 7.5 per mil; the corresponding value for pyrrhotite is somewhat lower, possibly owing to the temperature factor in the fractionation of sulfur isotopes. The eastern parallel orebody shows slightly lower $\delta^{34}S$ values than the main orebody. The amphibolite west of the ore zone proper contains weak disseminations of pyrrhotite. The sulfur isotope composition of this pyrrhotite is negative, up to −9.5 per mil in two samples of amphibolite and thus differs considerably from that of the ore zone. This probably indicates a different history of sulfur in the amphibolite and in the ore zone proper.

Compared to other barite-bearing, stratiform sulfide deposits in the Svecokarelides, the ore zone of Pyhäsalmi seems to have quite constant $\delta^{34}S$ values of sulfur isotopes. For example the Åsen pyrite-barite ore in the Skellefteå field, Sweden, is isotopically very variable (Rickard et al., 1979). Also in the Vihanti ore zone the $\delta^{34}S$ values vary considerably across the ore formation and in the associated U/P horizon. Accordingly to Rehtijärvi et al. (1979), the variation depends strongly on the lithology of the host rocks and originally on the conditions during the deposition of the rock types. In the Vi-

TABLE 5. Analytical Data and Radiometric Ages for Minerals

Sample no.	Mineral	Fraction	U ppm	^{206}Pb radiogenic ppm	Isotopic abundance (^{206}Pb = 100)			Age (m.y.)		
					^{204}Pb	^{207}Pb	^{208}Pb	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$
A311A	zircon	d > 4.2	421.5	120.9	0.006725	11.709	7.191	1,889 ± 4	1,872 ± 2	1,856 ± 2
A486A	titanate	total	44.87	12.41	0.08278	12.363	10.243	1,828 ± 2	1,812 ± 1	1,798 ± 1
A500A	zircon	total	279.8	70.72	0.09227	12.732	10.090	1,867 ± 3	1,755 ± 1	1,662 ± 1
A620bA	zircon	total			0.04120	12.015	14.564	1,875		
A620bB	titanite	total	61.88	16.64	0.06330	12.237	17.902	1,851 ± 5	1,799 ± 2	1,755 ± 3
A751A	zircon	d > 4.6 m > 70	358.2	80.91	0.01495	11.845	11.384	1,893 ± 4	1,673 ± 1	1,504 ± 1
A751B	zircon	4.2 < d < 4.6	500.0	103.9	0.01342	11.736	9.918	1,879 ± 5	1,599 ± 2	1,395 ± 1
A751C	zircon	4.0 < d < 4.2	838.2	141.9	0.01794	11.535	9.143	1,837 ± 3	1,420 ± 1	1,159 ± 1
A834A	zircon	d > 4.6 m > 160	212.0	58.37	0.02070	11.854	13.002	1,882 ± 3	1,833 ± 0	1,791 ± 1
A834B	zircon	4.2 < d < 4.6 m > 70	364.7	95.55	0.02789	11.896	11.704	1,873 ± 4	1,788 ± 2	1,715 ± 2
A843C	zircon	4.0 < d < 4.2 m > 70	812.4	203.5	0.06132	12.290	11.227	1,863 ± 4	1,745 ± 1	1,649 ± 2
A834D	titanite	total	96.8	26.18	0.25876	14.977	25.954	1,859 ± 18	1,808 ± 7	1,764 ± 9

d = density in g·cm^{-3}.
n = mesh size in μm.

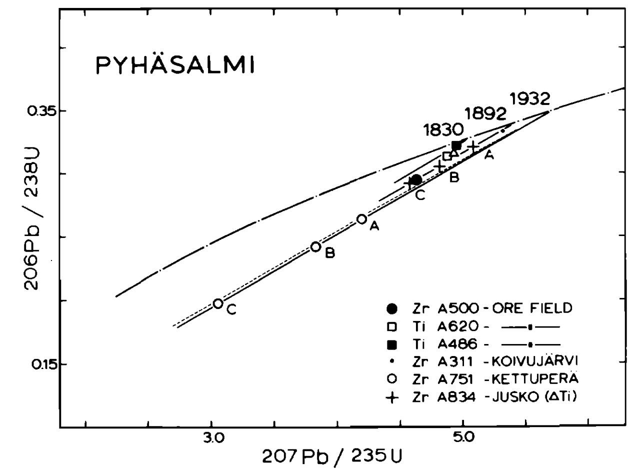

FIG. 10. Concordia diagram and U-Pb isotopic ratios for zircon and titanite samples from the Pyhäsalmi group volcanic rocks (A486, A500, and A620), Koivujärvi diorite (A311), Jusko quartz diorite (A834), and Kettuperä granite gneiss (A751). Each zircon fraction has a letter corresponding to the fractions listed in Table 5. The dashed line represents a continuous diffusion trajectory of Wasserburg (1963). Decay constants used are those of Jaffey et al. (1971).

hanti zinc ore proper, $\delta^{34}S$ values close +9 per mil were reported by Rouhunkoski (1968), but in the seven samples analyzed the variation was from +9.7 to −1.3 per mil.

Dating

Though the supracrustal rocks of the Vihanti-Pyhäsalmi-Pielavesi area have been regarded as Svecokarelidic, the stratigraphic correlation and age of the rocks have still not been established. Therefore several samples of intrusive and supracrustal rocks in the Pyhäsalmi area were selected for isotopic dating. The samples were taken from the Kettuperä granite gneiss, syntectonic quartz diorite from Jusko, and the metavolcanites of the Ruotanen schist belt. The samples were analyzed and the results interpretated by Dr. O. Kouvo at the Geological Survey of Finland.

U-Pb isotope analyses of minerals

Kettuperä granite gneiss: The three density fractions of zircons from the Kettuperä gneiss point to a simple chord, with an upper intersection age of 1,932 ± 1.5 m.y. and a lower intersection at 232 ± 1.6 m.y. according to York (1966) (Table 5 and Figure 10). The Kettuperä zircons are elongate, euhedral crystals characterized by a low common lead and medium uranium content. There is a general correlation between the degree of discordance and the uranium content, but, when compared with the samples A834 A-D from Jusko, the fractions are more discordant than those provided by the uranium content. This indicates that the rocks have a different geochemical history. It is noteworthy that the regression line lies on the lower side of the diffusion curve. The type of discordancy found in the Kettuperä gneiss may be related to zones of minor shearing at the sample site.

Jusko quartz diorite: The three zircon density fractions from the quartz diorite of Jusko define a chord that intersects concordia at 1,892 ± 1.5 m.y. and 312 ± 14 m.y. (York, 1966) and passes through the data points within better than analytical uncertainty. The data from the Koivujärvi gabbro-diorite (A311) 30 km south-southeast of Pyhäsalmi and from the silicic porphyry (A500) found within the sulfide orebody fit the same regression line and thus yield the same age. The continuous diffusion trajectory of 1,892 m.y. also closely fits the discordia. Depending partly on the isotopic composition of common lead used for correction, the calculated diffusion model ages range between 1,889 and 1,892 m.y. A lower age of 1,867 m.y. was calculated for titanite from the Jusko quartz diorite.

Pyhäsalmi plagioclase porphyrite (A620b): A $^{207}Pb/^{206}Pb$ age for the total zircon fraction from this rock shows a minimum age of 1,875 m.y. typical of the above-mentioned rocks as well. The diffusion model age for titanite from the same rock is 1,860 m.y.

It is not uncommon in the Finnish Precambrian terrain for younger titanites to be associated with older zircons within the same rock. This has been described, for example, from the Pihtipudas area 40 km southwest of Pyhäsalmi (Aho, 1979). The younger age, 1,800 to 1,830 m.y., has been reported for several granitoids, e.g., Vainospää in Lapland, Åva in southwestern Finland, and Hirvensalo in eastern Finland. This late metamorphic stage is indicated in Pyhäsalmi by titanite (A486) from an amphibolite in the ore field. The calculated diffusion model age for this mineral is 1,830 m.y.

From geological evidence already discussed the present material proves that the Kettuperä granite gneiss predates the syntectonic intrusions represented, for example, by Jusko quartz diorite. The numerous zircon concordia diagram data show that the mean age of syntectonic Svecokarelian intrusions lies within the age limits of 1,860 to 1,900 m.y. The age group slightly higher than 1,900 m.y. is represented by rocks such as Kalanti trondhjemite, some synorogenic volcanic rocks, and granulite metamorphics in Lapland. Further efforts are needed to establish a magmatic or metamorphic evolution for this Kettuperä rock unit, which is geologically and isotopically the oldest recognized crustal component in the Pyhäsalmi lithologic complex.

TABLE 6. Analytical Data for Total Rock Samples

Sample no.	Rock	Location and/or drill hole, depth	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
A484	Silicic volcanic rock	East of the ore PyS-84, 60–100 m	24.7719 ± 0.0190	16.3101 ± 0.0165	43.2369 ± 0.0492
A485	Silicic volcanic rock	Drift W of the ore	19.9563 ± 0.0161	15.8021 ± 0.0059	38.6648 ± 0.0192
A486	Amphibolite	Railroad cut, ore field	24.5270 ± 0.0480	16.2710 ± 0.0366	37.5247 ± 0.0879
A500	Quartz porphyry	Orebody	15.2146 ± 0.0072	15.1597 ± 0.0088	34.9430 ± 0.0202
A501	Altered silicic volcanic rock	Open pit	16.4900 ± 0.0117	15.2984 ± 0.0124	36.1119 ± 0.0317
A618	Amphibolite	PyS-4, 33.78–34.32 m	18.0581 ± 0.0151	15.4938 ± 0.0156	35.7224 ± 0.0369
A619A	Amphibolite	PyS-84, 30.43–30.93 m	18.0777 ± 0.0769	15.5066 ± 0.1081	37.0544 ± 0.2530
A619B	Quartz porphyry	PyS-84, 26.0–37.0 m	26.4786 ± 0.0593	16.4481 ± 0.0423	43.7025 ± 0.1167
A620A	Mafic volcanic rock	PyS-3, 63.74–64.30 m	20.8082 ± 0.0720	15.7902 ± 0.0890	35.4610 ± 0.2301
A620B	Plagioclase porphyrite	PyS-3, 128.79–129.28 m	22.6378 ± 0.0314	16.0272 ± 0.0294	39.7399 ± 0.0792
A620C	Altered mafic volcanic rock	PyS-3, 134.69–135.36 m	16.4070 ± 0.0129	15.3182 ± 0.0141	35.0442 ± 0.0456

Lead isotopic composition of some Pyhäsalmi group volcanic rocks

The isotopic abundances of lead have been used as a tracer to discriminate between different hypotheses for the origin of ore-forming solutions. Because the correlation between the isotopic composition of ore lead and the isotopic composition of the total lead of the associated volcanic rocks is of fundamental importance, 11 silicic to mafic volcanogenic rocks were selected for isotopic work. This preliminary approach consists of lead isotope analyses only.

The analytical results, with notes on sample character are listed in Table 6. The isotope ratios are plotted in Figure 11. The slope of the best-fit line through 10 points (sample A485 omitted) defines an age of 1,909 ± 27 m.y., if the conventional linear regression is used. The quoted uncertainty is ±2.

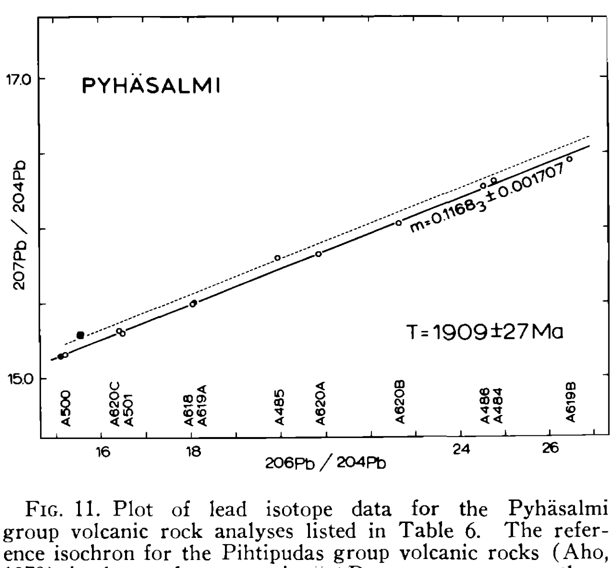

FIG. 11. Plot of lead isotope data for the Pyhäsalmi group volcanic rock analyses listed in Table 6. The reference isochron for the Pihtipudas group volcanic rocks (Aho, 1979) is shown, for comparison. Decay constants are those of Jaffey et al. (1971). ■ = Pihtipudas galena; ● = Pyhäsalmi galena.

If the results are weighted according to York (1966), the slope will define an age of 1,941 ± 19 m.y. These results are in agreement with the ages obtained on the Pihtipudas volcanic group (Aho, 1979) and the Pellinge metaandesites (Geol. Survey Finland, 1976), but they are, however, distinctively younger than the mafic Jatulian volcanism (e.g., Simonen et al., 1978).

Perhaps the most striking genetic feature of the data is that the point derived from the isotopic composition of galena lead in the Pyhäsalmi ore deposit lies on the same isochron within the limits of experimental uncertainty. This point represents the least radiogenic lead and is closest to the quartz porphyry sample A500 found within the ore.

Kouvo and Kulp (1961) have shown that central Finland can be divided into two units with different isotopic compositions of lead. This is demonstrated here by the two parallel isochrons of the Pyhäsalmi group and the Pihtipudas group (Aho, 1979) volcanic rocks with about the same lead/lead age but different primary leads. The values of the isotopic compositions of the sulfide leads used were those given by Stacey et al. (1977):

	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Pyhäsalmi	15.111	15.147	34.835
Pihtipudas	15.577	15.287	35.164

It may be purely accidental that the point A485, so far neglected, fits the Pihtipudas isochron of 1,898 m.y. age (Aho, 1979). This rock represents the silicic volcanic rocks west of the orebody near the westward amphibolite.

It is chronostratigraphically significant that a preliminary four-point whole-rock isochron of Koivujärvi, not included in this study, fits the Pyhäsalmi isochron and the Pyhäsalmi sulfide lead. The Koivujärvi region is 30 km south of Pyhäsalmi and the volcanic rocks analyzed include two amphibolites, one mafic agglomerate, and one silicic tuffite.

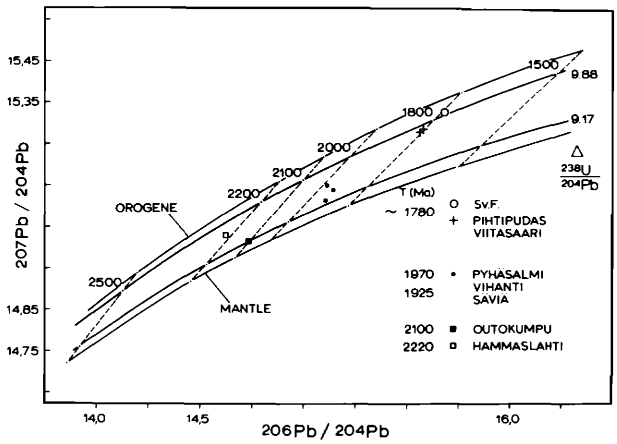

FIG. 12. Lead isotopic systematics and ages of some Finnish galenas including the Pyhäsalmi sulfide deposit (Simonen et al., 1978). ○ = Svecofennian in Southwestern Finland.

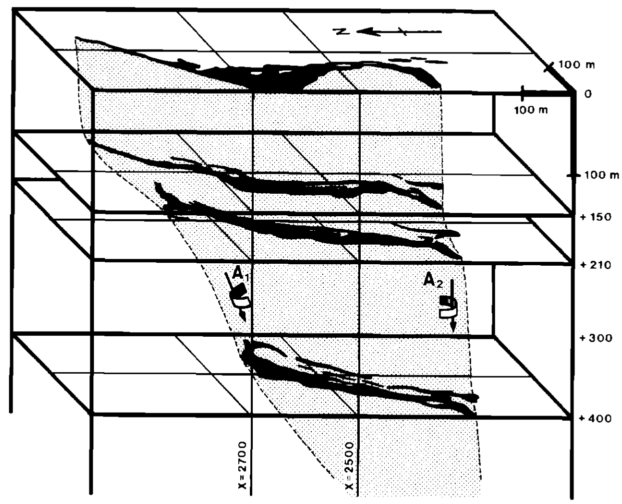

FIG. 13. A block diagram depicting the shape of the orebody. A_1 and A_2 are the fold axes measured from the northern and southern ends of the orebody.

Lead isotope abundance in galenas

In 1946 Holmes and Houtermans independently suggested the evolution model of lead isotope ratios in minerals (Holmes, 1946; Houtermans, 1946). The results gathered since then have given rise to the branch of isotope geology. In 1953 Collins, Russel, and Fahrquahr assumed that there is a unique average growth curve that covers most ordinary leads. The work by Stanton and Russel (1959) opened the discussion on the evolution of lead in submarine volcanic exhalative massive sulfide deposits, a subject that is of ever greater current interest.

Early in the work it was found that the differences along the isochrons were due to the difficulty of measuring the $^{206}Pb/^{204}Pb$ ratios. More recently temperature-controlled and well-calibrated lead runs have enhanced precision and allow observations to be made along isochrons as well as along growth curves. Right from the start it was found that the geologically heteregeneous Svecofennian sulfide deposits are isotopically homogeneous and that samples such as those from Korsnäs, Attu, Orijärvi, Pakila, and Metsämonttu fit the global curve. In contrast, the large massive sulfide deposits like Vihanti, Pyhäsalmi, Säviä, and Outokumpu fit the growth curves of a lower U/Pb value and show a higher age.

The isotopic composition of Pyhäsalmi sulfide lead was first reported in 1961 by Kouvo and Kulp. Revised values were published in 1973 (Geological Survey of Finland, Annual Report, 1973). The results obtained at the laboratories in Otaniemi (Geol. Society of Finland) and in Denver (U. S. Geol. Survey) are in good agreement:

	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Denver	15.111	15.147	34.835
Otaniemi	15.11	15.15	34.86

The lead isotope compositions of three Finnish sulfide occurrences, including the Pyhäsalmi ore deposit, were inserted in the article "Plumbotectonics IIA by Stacey et al. (1977). According to them the Stacey-Kramers two-stage model yields an age of 1,970 m.y. for the Pyhäsalmi lead, and a mantle origin is indicated by the lead isotope data. These authors also conclude that the somewhat higher $^{207}Pb/^{204}Pb$ value for the Pyhäsalmi deposit exhibits a more continental character than that for the Outokumpu deposit, which almost fits the mantle curve and has a model age of 2,100 m.y.

The evolution of sulfide lead during an interval of 1,800 to 2,200 m.y. in the Finnish Precambrian is shown in the diagram in Figure 12, compiled by Simonen et al. (1978). The curves and isochrons are taken from the plumbotectonics model by Doe and Zartman (in press).

This orogenic approach confirms the earlier observations presented almost two decades ago, that lead in the Finnish syngenetic sulfide deposits in Outokumpu and in the Vihanti ore zone (including the Pyhäsalmi deposit) does not fit the primary growth curve for terrestrial leads. Instead their isotopic composition becomes more radiogenic to the west of the Prekarelian craton in the manner shown in the diagram.

Discussion

The Pyhäsalmi ore is a Precambrian syngenetic massive sulfide ore in a volcanic rock sequence. It is associated with the isoclinally folded Ruotanen schist belt as shown by the observations made in and around the mine. The Pyhäsalmi orebody is a deflated sack in shape that narrows downward (Fig.

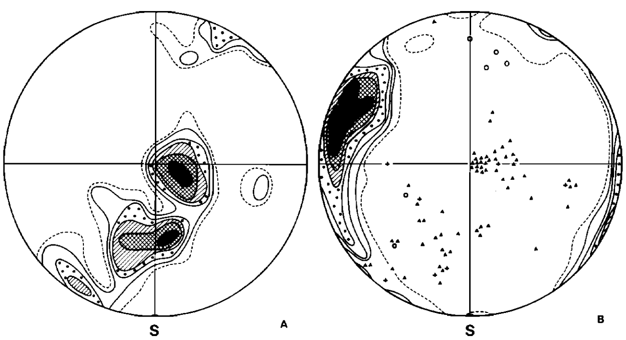

FIG. 14. A. Lineations and fold axes (173 observations) measured in the mine above the +250 level; contours <2%, 3%, 4%, 5–6%, 7–10%, and >10%. B. Lineations (▲), fold axes (○), and schistosity (contoured) (408 observations) measured in the open pit.

13). The ore conforms roughly with the sericite schist zone. The complicated structures and forms in the ore proper and its environment can be attributed to polyphase deformation as follows:

The first phase of folding (F_1) was isoclinal, the fold axis being subhorizontal and trending north-south with the subvertical axial plane. The deeper parts of the ore syncline (Fig. 4) were turned eastward as a result of a second phase of deformation (F_2) in which the axial plane of F_1 was folded, but the axis of the second folding was about the same as in the first phase of deformation. The third deformation phase (F_3) produced deep depressions and culminations of isoclinally folded strata with the axis of F_3 trending about N 60° W. The Pyhäsalmi ore is located in a narrow depression of the third phase of deformation. The two maxima of fold axes in the diagram of structural elements (Fig. 14) are caused by F_3 folding of F_1 axes; both maxima represent the northern and southern limbs of a synform trending N 60° W (F_3). The deformation resulted in folding in the ore deposit and brecciation of the siliceous and carbonate rocks that originally constituted interlayers in the massive ore. The sulfides were remobilized and brecciated the wall rocks but only on a small scale. Microcline pegmatite granite intersected the southern part of the ore zone along the axial plane of the F_1 folds. The main peak of regional metamorphism of the amphibolite facies probably preceeded the pegmatite intrusion.

The fluid inclusions in the ore minerals were studied by I. Haapala and K. Kinnunen at the Geological Survey of Finland. Sphalerite is without primary fluid inclusions (I. Haapala, pers. commun.); whereas tourmaline from the chalcopyrite-pyrrhotite assemblage (+150-m level, beyond the main orebody) contains primary inclusions with ca. 75 percent aqueous solution by volume and ca. 25 percent vapor by volume, a small pyrrhotite crystal and an isotropic daughter mineral. The filling temperature (minimum temperature of crystallization) varied from 292° to 315°C, the anisotropic phase dissolved at ca. 290°C, but the pyrrhotite crystal did not dissolve even at 400°C. The freezing point depression (−2.0° to −5.1°C) indicates a salinity equivalent to 4.0 to 8.2 wt percent NaCl. The complex behavior during freezing suggests the presence of carbon compounds in the inclusions. The tourmaline studied evidently originates from the recrystallization period and hence the filling temperature indicates only the metamorphic conditions. Pressure corrections in filling temperature should therefore be made.

The absence of fluid inclusions from sphalerite is surprising because weakly metamorphosed volcanic-exhalative deposits, e.g., the Kuroko type of ores, display abundant fluid inclusions in sphalerite. An explanation may be the "dry" conditions that prevailed during metamorphism, at least in the massive orebody proper.

The Kettuperä granite gneiss, the oldest rock unit in the area, may well be considered as the basement of the volcanic formation. Hence, stratigraphically the lowest rocks in the Ruotanen schist belt are at the eastern margin and get younger westward.

The sequence begins with mixed clastic metasediments and intermediate to basic metavolcanites and volcanoclastics. These are overlain by acidic metavolcanites (leptites) and these in turn by a thick pile of basic metavolcanites (amphibolite complex).

The ore deposit is associated with altered rocks and is located in the acid metavolcanites. Of the altered rocks the sericite schists constitutes a long belt on either side of the orebody, whereas the cordierite-bearing rocks—altered mafic volcanites—occur east of the ore zone, stratigraphically in the footwall of the ore. The zone of altered rocks is up to 200 m thick and it can be traced for several kilometers northward. Thus, the alteration zone, which is of the "blanchet" type, extends far north of the orebody. The coarse volcanoclastic material around the orebody is scarce in leptites, but farther north, in Kettuperä, the lapilli tuffs and agglomerates abound in the leptite belt. Therefore, during the extrusion of acidic material, the volcanic center was located north of the orebody, and the deposit may be classified as of the distal type. The sulfur isotopes seem to be more constant in the ore zone proper than in other stratabound barite-bearing pyrite deposits. This might indicate that conditions remained stable during the deposition of sulfides.

The alteration phenomena in the leptites and underlying basic volcanites are typical of massive stratabound deposits elsewhere. The depletion of Na and

Ca and the gain in K, Mg, and Fe are the result of replacement of plagioclase by sheet silicates such as sericite, chlorite, and biotite during the circulation of volcanic thermal waters. Regional metamorphism caused the recrystallization of altered rocks into sericite schists and cordierite-bearing gneisses, cordierite-anthophyllite rocks, and biotite rocks.

The evolution of the Svecokarelidic sulfide lead as described above suggests that the formation of Pyhäsalmi, Vihanti, and Säviä, and evidently also of the other deposits in the area mentioned by Huhtala (1979), took place 1,900 m.y. ago; thus they differ in age from the older Jatulian Karelidic rocks (2,000 to 2,200 m.y.) and the Svecofennidic sulfide deposits in southwestern and central Finland (1,800 m.y.). The orogenic approach of the plumbotectonic model indicates that the lead of the Finnish syngenetic sulfide deposits in Outokumpu and in the Vihanti-Pyhäsalmi district does not fit the global growth curve but is located below it, whereas the lead of Svecofennidic sulfides is located on the growth curve.

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