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Tectonics and stratigraphy of the vicinity of Outokumpu, North Karelia, Finland

Including a structural analysis of the Outokumpu ore deposit

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TECTONICS AND STRATIGRAPHY OF THE VICINITY OF OUTOKUMPU, NORTH KARELIA, FINLAND

INCLUDING A STRUCTURAL ANALYSIS OF THE OUTOKUMPU ORE DEPOSIT

BY

GABOR GAÁL, TAPIO KOISTINEN and ESA MATTILA

WITH 39 FIGURES, AND TWO TABLES IN THE TEXT, ONE PLATE AND SIX MAPS

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The Precambrian rock types are described in three stratigraphic units: 1) Geosynclinal rocks with mica schists, black schists and the Outokumpu association comprising ophiolitic serpentinite, dolomite, skarn, cherty quartzite and the ores, 2) epicontinental rocks and 3) basement gneiss.

The cover rocks show the effects of polyphase deformation in four phases. The first deformational phase produced 045° trending large recumbent folds (F_1) presumable with SE vergency. The second phase refolded the F_1 structures coaxially into open antiformal and synformal folds (F_2) with vertical axial planes. The third phase threw the previous folds into 015° to N—S trending upright folds (F_3). At this stage the Sotkuma dome was formed. The folds of the fourth phase (F_4) followed immediately F_3 or developed simultaneously. These 155° trending isoclinal folds with axial planes dipping 30—90° WSW represent the deformation of the eastern margin of the Karelidic belt.

Metamorphism reached the conditions of the amphibolite facies during the first deformational phase. During the third and fourth deformational phases there has been a repeated cycle of regional metamorphism under the conditions of the greenschist facies.

The ore predates s_1 and it is enveloped by a recumbent F_1 fold, laying parallel to the fold limbs and the axial plane foliation of the wall rocks. The s_1 parallel lithological layering is probably stratification, but this cannot be verified by the structural analysis. The subsequent deformational phases caused moderate refolding, brecciation of the ore and block movements. Sulphide mobilization can be correlated with the deformational phases.

Gabor Gaál, Outokumpu Company, Exploration, P.O. Box 27, SF-02101 Espoo 10, Finland

Tapio Koistinen, Outokumpu Company, Keretti mine, SF-83500 Outokumpu, Finland

Esa Mattila, Outokumpu Company, Vihanti mine, SF-86440 Lampinsaari, Finland

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INTRODUCTION

This study presents the results of a detailed investigation of an area of 420 km^2 in North Karelia, which includes the high grade copper-cobalt deposits of Outokumpu and Vuonos and the low grade nickel deposit of Vuonos. The NE—SW elongation of the area, which is called in this paper the vicinity of Outokumpu, reflects the regional strike of the lithological units (Fig. 1).

Owing to its valuable ore bodies and its economic potential, the vicinity of Outokumpu and its environs is one of the most thoroughly studied parts of the Finnish Precambrian since the discovery of the copper-cobalt ore in 1910. Many geologists of renown have contributed to our knowledge of its geology with summary of its exploration history given by Vähätalo (1953) and references therein. Here only those works are mentioned that immediately concern the area and which have a direct influence on the following text. The first comprehensive geological account was given by Frosterus and Wilkman (1920). Their Joensuu map sheet on a scale of 1 : 400 000 provides basic reconnaissance information of the eastern part of the country where large areas still await remapping. Väyrynen (1933, 1939) probably had the greatest influence on the shaping of ideas regarding the stratigraphy and tectonic history of the area. After Wegmann (1928) had brought the tectonic experience of Alpine geologists to Finland, Väyrynen interpreted the Karelian orogeny on the basis of the Alpine model.

The characteristic rock of the Outokumpu association, serpentinite, has been exhaustively described by Haapala (1936). Petrological description of the Outokumpu ore and the host rocks was given by Vähätalo (1953) while Disler (1953) concentrated on the structural problems of the ore. Borchert (1954) used the data of Vähätalo and Disler as a basis for suggesting that the ore had a volcanic-exhalative origin, an interpretation supported by the sulphur isotope study by Mäkelä (1974). Saksela (1957) explained the genesis of the ore as the result of tectonic processes, metamorphism and sulphide remobilization from black schist, while Peltola (1960, 1968) gave a detailed petrological and geochemical description of the black schists of the Outokumpu region.

In 1952 the Exploration Department of the Outokumpu Company started an extensive and thorough exploration of the Outokumpu region. This led to the discovery of both the copper-cobalt ore and the low grade nickel ore at Vuonos

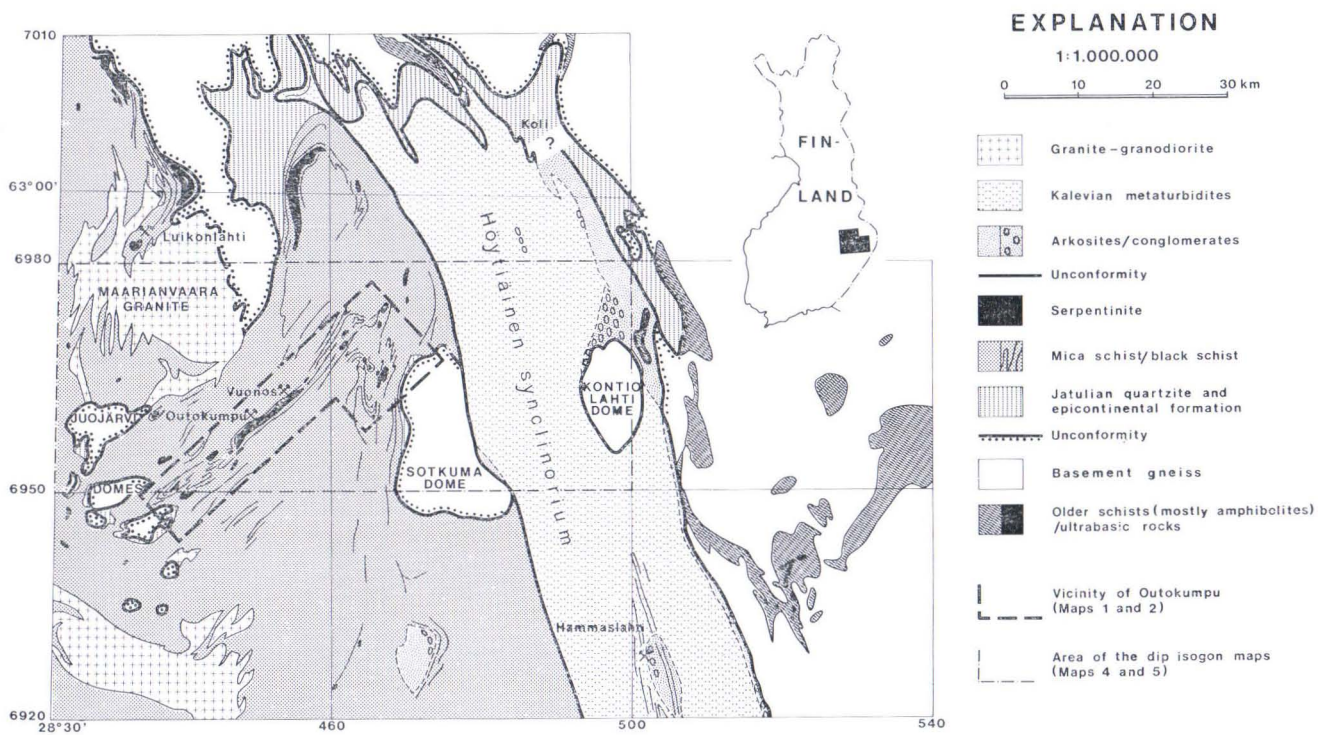


FIG. 1. The areas under investigation in the regional geological framework of North Karelia. Map modified after Frosterus and Wilkman (1920), Gaál (1964), Huhma (1971) and Nykänen (1971).

as well as some other areas of minor copper-cobalt mineralization. The results of the work are given by A. Huhma & M. Huhma (1970) and in three 1 : 100 000 map sheets by A. Huhma (1971).

In spite of the great amount of work and new data, some unsolved problems still remained particularly in relation to the tectonic history of the area and the ore deposits. To find solutions the present study was initiated in 1967 by the Exploration Department of the Outokumpu Company. For this purpose the vicinity of Outokumpu was remapped on a scale of 1 : 10 000 and in some minor parts 1 : 2 000. The field work was carried out under the direction of G. Gaál during the summers of 1967 and 1968 by field parties lead by G. Gaál, E. Mattila and P. Karhu. The field data were subsequently compiled by G. Gaál in 1967 and 1969. It was not until 1971 that the time seemed to be ripe for the results to be finalized. However, the task had grown too big to be a one-man project and a team was formed to compile a detailed geological description of the vicinity of Outokumpu, the main emphasis being on the tectonics and stratigraphy of the area. The team consisted of G. Gaál (exploration geologist), T. Koistinen (mining geologist) and E. Mattila (exploration geologist). G. Gaál carried out the regional structural analysis and the compiling work. T. Koistinen dealt with the tectonics and petrology of the Outokumpu ore and its surroundings. E. Mattila described the petrology of the epicontinental rocks.

STRATIGRAPHY AND PETROLOGY

Geosynclinal rocks

Mica schists

The dominating rocks of the area under study are mica schists with a remarkably uniform mineralogical composition containing an average of 30 % quartz, 35 % plagioclase and 27 % other minerals, mainly biotite. As the most abundant type Frosterus and Wilkman (1920) called these rocks phyllites adjacent to the Sotkuma dome and mica schist in the western part of the area. Indeed, there is a slight increase in grain size from east to west that leads to a more complete destruction of the primary structures westwards. But there are exceptions, as in an exposure southeast of Salmilampi (Fig. 19) where graded bedding is preserved very much as in the mica schists above the Sotkuma dome.

Rock types

Under the microscope the rocks exhibit granoblastic to lepidoblastic texture, being fairly equigranular with only mica and garnet forming larger porphyroblasts. The plagioclase (An_{20-35}) is usually devoid of twinning. It is more or less altered,

sericitized or saussuritized. In grain size it is similar to quartz, being an average of 0.15 mm. The grain size of biotite averages 0.5 mm, varying between 0.1 and 2 mm. Biotite is often chloritized and in narrow secondary shear zones it may be entirely altered into chlorite. Minor constituents are almandine, retrograde chlorite, potassium feldspar and occasional graphite. In some thin sections almandine is fairly abundant. Accessories are apatite, tourmaline, zircon, epidote, magnetite, sulphides and graphite. In a drill hole in the Mökkivaara section SE of Outokumpu, fine sillimanite needles (fibrolite) and kyanite, occurring separately, were found in the mica schist.

According to their texture and composition the monotonous mica schist group can be subdivided into different types, which are mappable in well-exposed areas. The following have been distinguished: bedded mica schist, homogeneous mica schist, mylonitic mica schist and veined gneiss.

Bedded mica schist

The bedded mica schists are of sedimentary origin with interbedded dark layers rich in mica and light ones rich in plagioclase and quartz (Fig. 2). The mica-rich layers often contain appreciable amounts of graphite, and the plagioclase-quartz-rich layers calc-silicate intercalations from 0.5 to 2 cm thick. Graded bedding has been observed as a primary structure. The cycles are generally from 5 to 10 cm thick, but in exceptional cases they can attain several metres. A light quartz-plagioclase-rich material, that was originally graywacke, predominates at the bottom of the cycle. It grades upwards into mica-rich, primarily pelitic material in considerably lesser amounts. The grain size does not change with the mineral composition. Slumping structures (Fig. 3) have been observed on the island of Kulkevainen on the lake Viinijärvi.

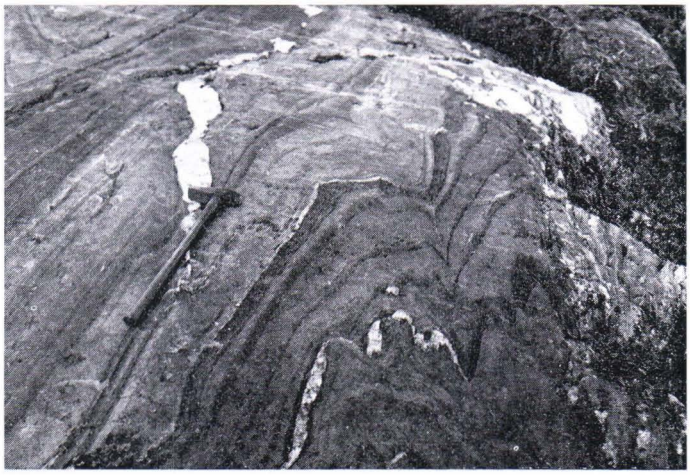


Fig. 2. Bedded mica schist, west of Rukkajärvi.

FIG. 3. Slumping structures in graphite-bearing mica schist, island Kulkevainen, Lake Viinijärvi.

Characteristic of the bedded type is a laminated mica schist, which was mapped as a distinct, coherent layer above the Sotkuma dome. The cycles in this type are much thinner, being from 1 to 5 cm thick.

The graphite-bearing mica schist is another variety of the bedded type, the only difference being a marked graphite content. It is several tens of metres thick close to the black schists with which it exhibits transitional contacts. In contrast to the black schist, the graphite-bearing mica schist contains fewer sulphides and does not become rusty on weathering. It is characteristically dark grey or bluish to black in colour owing to the finely distributed graphite. It usually contains light coloured, continuous or boudinated calc-silicate layers and concretions.

Homogeneous mica schist

In the homogeneous mica schist the mineral components are distributed evenly throughout the rock and generally without perceptible bedding. The mineral composition varies greatly. Some types have a low mica content and should preferably be called mica-quartzite or arcosite. In another type, mica is the predominant mineral. As a rule the homogeneous type contains light to dark green calc-silicate layers and concretions. In fold hinge zones these layers are usually sheared parallel to the axial plane and disrupted, forming necklace-structures in sections oblique or perpendicular to the fold axis. The homogeneous mica schist exhibits poor foliation and can be almost isotropic (Fig. 4). This gives it a strange appearance reminiscent of a fine-grained intrusive rock. Another textural type of the homogeneous mica schist shows very poor foliation but distinct lineation.

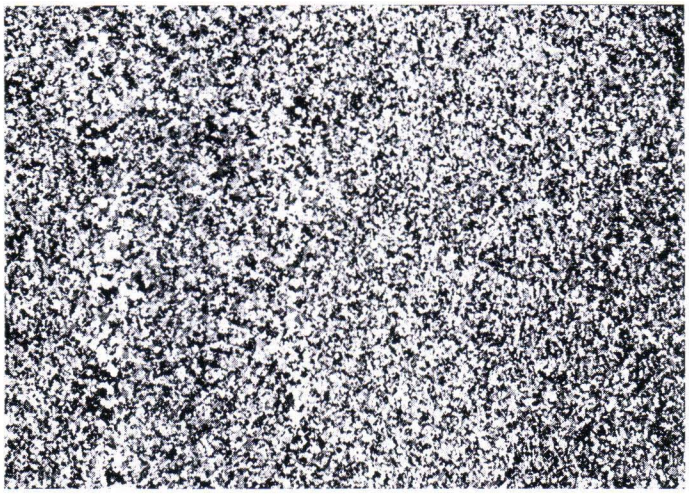


FIG. 4. Homogeneous mica schist, crossed nicols, 3 x magnification.

Cataclastic mica schist

In zones of mylonitization and cataclastic shearing repeated deformation has given rise to the development of a coarse-grained flaky biotite schist out of the different types of mica schist. The original silica content of the mylonitized rock recrystallized in quartz veins and lenses situated in zones of minimal pressure.

Two distinct layers of sulphide-bearing mylonitic mica schist of the Sotkuma dome were mapped (Map 1). The rock is a rusty weathering cataclastic mica schist consisting solely of biotite and muscovite. Numerous quartz veins and lenses containing also some coarse-grained plagioclase are encountered in the schist. The rusty weathering is due to fine-grained pyrrhotite impregnation.

Veined gneiss

Along the coast and on the islands of the lake Juojärvi in the western part of the study area, there are occasional occurrences of quartzo-feldspathic veins in mica gneiss. Minor layers several metres thick may be called veined gneiss.

Origin of the mica schist

The rhytmic alternation of mica-rich and quartz-plagioclase-rich material in graded bedding is the most typical primary structure of the mica schists in well-preserved parts. The characteristic feature of the mineralogical composition is the

high plagioclase content. The graded bedding was probably formed in a marine geosynclinal environment; the high feldspar content points to the rapid sedimentation of poorly weathered material whose average composition was probably that of graywacke with some argillaceous portion. It is difficult to estimate the thickness of these rocks because of the intensive folding, especially the recumbent folding of the first deformational phase. However, in the vicinity of Outokumpu the total thickness is estimated to be at least 3 000 m (Fig. 26), with the upper limit being unknown.

Black schists

In the Outokumpu region Peltola (1960) has classified black schists as those that contain more than 1 % of both graphite and sulphur. They form coherent horizons of considerable lateral extent (Map 1) in the mica schist. The thickness of the lithological units varies on the whole from a few metres to about one hundred metres. Dependent on the graphite content, the colour varies from black to bluish grey with the presence of graphite-pigmented amphibole conspicuous in the calcareous type. The presence of sulphides causing rusty weathering is characteristic.

Under the microscope the silicates of the black schists are mostly strongly pigmented by graphite and sulphides. The sulphides are mainly pyrite and pyrrhotite of varying relative amounts. The sulphide content increases parallel to that of carbon (Peltola, 1968). They both commonly range from a few per cent to 10—15 per cent and within the same limits.

The silicates are very poor in iron, which, in addition to the lower content of alumina and silica, conspicuously distinguishes the silicates of the black schists from those of the carbon-free mica schists. Unlike mica schist, microcline is common in black schist, and the anorthite content in plagioclase is higher, ranging generally between 25 and 70 %. In calcareous types quartz and mica are present in small amounts and tremolitic amphibole is abundant. Sphene is a constant accessory with rutile and ilmenite as inclusions in some places. In addition, tourmaline, apatite, »thucholite», magnetite, chalcopyrite, sphalerite and carbonate occur. Porphyroblasts of andalusite have been encountered locally in argillaceous black schist and diopside in highly metamorphosed calcareous black schists (Peltola, 1960). The texture of the black schist is lepidoblastic or nematoblastic. Porphyroblastic minerals that are frequently helicitic in texture occur.

The contents of calcium and magnesium in all the black schists of the Outokumpu region exceed those in the mica gneiss. Vanadium, molybdenum and uranium are met with in amounts characteristic of sapropelic sediments.

According to Peltola (1960, 1968), the black schists are metamorphic derivatives of sapropelic sediments that deposited in a marine environment.

The Outokumpu association

A characteristic feature in the vicinity of Outokumpu is the regular occurrence of quartzites, skarns and carbonate rocks around serpentinite bodies. This whole rock group is called the Outokumpu association in the present paper. The complete sequence of the enveloping rocks from the serpentinite core outwards is: carbonate-bearing serpentinite — carbonate rock — tremolite skarn — diopside skarn — quartzite. The outer zone adjacent to the surrounding mica schist is commonly occupied by black schist.

Serpentinites

The serpentinite bodies are elongated lenses ranging from a few metres to several hundred metres thick. On map 1 they have been established by mapping or drilling. Those near the eastern shore of Viinijärvi were drawn on the basis of characteristic geophysical anomalies and structures.

The serpentinites of the area have been interpreted previously as ophiolites in a eugeosynclinal environment. The structural analysis and petrological observations support this view. Their mode of occurrence strongly suggests tectonic control. Analysis of the folding indicates that they prefer emplacement in pressure-minima zones, above all in the hinge zones of isoclinal folds. No significant effect of contact metamorphism can be seen and all the metamorphic phenomena around the serpentinites are typical of regional metamorphism. The conclusion may be drawn that the serpentinites were transported tectonically during the folding and metamorphism as bodies already solidified. They are not magma intrusions in their present position but tectonic inclusions.

Haapala (1936) divided the serpentinites of the Outokumpu region into two different zones according to the predominant serpentine mineral. In the western zone the major serpentine mineral is chrysotile and in the eastern zone it is antigorite. The boundary between the zones is in a N—S direction west of Horsmanaho.

The serpentinites of the western zone

The western zone comprises the areas around Vuonos, Outokumpu and Juojärvi. Because the regional fold axis in this area is subhorizontal, the sections of the serpentinite bodies on the map are elongate; around Outokumpu and Vuonos some are even thousands of metres long (Map 3). The major lenses are composed of many smaller lenses with signs of slipping movements between them. Except on these sheared zones, the serpentinites of the western zone are massive and without any dimensional orientation of mineral constituents.

Haapala (1936) distinguishes three different serpentine rock types in Outokumpu.

The dunitic type is considered to have been originally mainly dunitic in composition. The rock was almost monomineralic and granular texture is still frequently visible. Olivine can now be found only in negligible amounts as relics. Scattered grains of chromite, occasional magnetite and in places serpentinuous pseudomorphs after amphibole are met with. Chlorite, talc and carbonate occur as secondary minerals. The colour varies from greenish to black.

The saxonitic type is now largely made up of chrysotile. In addition it may contain olivine, enstatite, anthophyllite, tremolite, bastite, chlorite, kaemmererite, talc, carbonate, chromite, magnetite and pyrrhotite. Olivine was the major component of the pre-metamorphic rock although in exceptional cases enstatite was also dominant. Primary minerals are rare. Nevertheless their relics are often easily discernible. This rock type is mottled in appearance.

The porphyritic type contains elongate crystals of olivine (Fig. 5) or their serpentinuous pseudomorphs, thus giving the rock a porphyritic appearance. In addition to chrysotile and olivine, it may contain tremolite, anthophyllite, dolomite, talc, chlorite, chromite, magnetite and sulphides. Olivine may be conspicuously well preserved.

The distribution of the various types is fairly complicated. The parts most altered are regularly to be found in the interior of the bodies and the well-preserved olivine in the contact zones. The former consist mostly of dark green to pale green serpentinite varieties, while magnetite dust has coloured the contact zones black. Sulphides are slightly more abundant near the contacts.

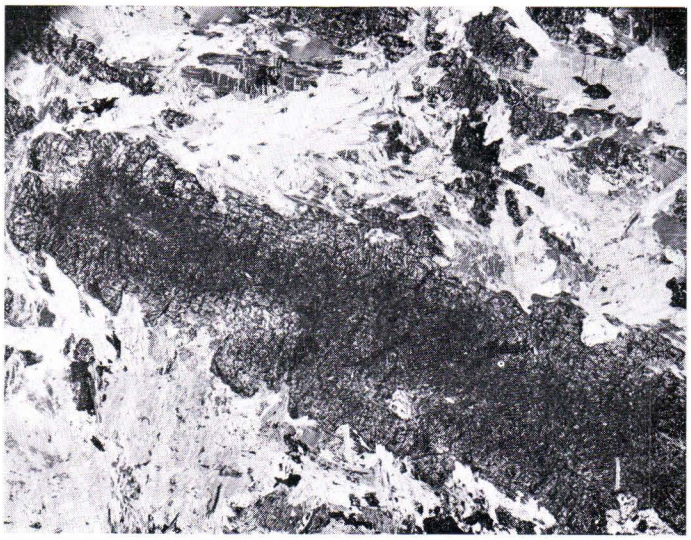


FIG. 5. Large pigmented grain of olivine, mostly serpentinized, in the porphyritic type of serpentinite, crossed nicols, 3.5 x magnification. Outokumpu, drill hole Oku 574, 152 m.

Serpentinites of the eastern zone

The tectonic and metamorphic history of the serpentinites in the eastern zone is more complicated than those of the western zone, evidently as a result of the strong influence of the third deformational phase, which only slightly affected the western zone. Two of the occurrences, Horsmanaho and Sola, were analyzed tectonically (Figs. 24 and 25). At Sola, where the largest outcrops of serpentinite occur, the rock exhibits distinct foliation and refolding of the lenses. It is strongly tectonized especially along the contacts and is surrounded by an admixture of talc-magnesite rock, tremolite skarn, zoisite-tremolite-skarn and chlorite schist.

As previously mentioned, the predominant serpentine mineral of the eastern bodies is antigorite. The composition of these bodies ranges from serpentinite or serpentine-bearing talc-magnesite rock to pure talc-magnesite rock.

The most detailed petrological information on the rock is available from Sola. The serpentinite here is black in foliated horizons and encloses small areas of pale green varieties with gradual transition. The pale green portions consist of colourless spherulitic aggregates of antigorite blades and black spots of chrysotile with a mesh texture in which relics of idiomorphic olivine (Fo_{90}), upto 10 mm in diameter, occur. The olivine is altered to bowlingite at its contacts and along fissures. In the antigorite portions patches of magnesite from 1 to 5 mm in diameter occur partly replaced by talc. According to Haapala (1936), the sequence of crystallization is: olivine — chrysotile and magnesite — magnesite with talc — antigorite. The antigorite was repeatedly attacked by carbonatization and by the formation of talc. The resulting serpentine-bearing magnesite-talc rock is a pale or greenish, soft and massive variant with relics of antigorite.

Quartzite, skarn and carbonate rocks

As Huhma & Huhma (1970) have already pointed out, the quartzites, skarns and carbonate rocks associated with the serpentinite differ from those of the epicontinental facies. They are ascribed to a genetically close relationship with the serpentinites on the basis of their mode of occurrence, structure, mineralogical and chemical composition. This problem will be discussed in connection with the serpentinization.

Quartzite, skarn and carbonate rocks are marked on map 1 using a common print. The relative amounts of these rock components vary from place to place, their total quantity being in direct relation to the size of the serpentinite bodies.

Quartzites

The mode of occurrence of the quartzite is illustrated on map 3. On this scale, narrow horizons of skarn and carbonate rocks are not differentiated within the quartzite. With the exception of some secondary sheared contacts, the quartzite is not in direct contact with the serpentinite, being separated by skarn and carbonate rocks.

In its purest form the quartzite is an almost monomineralic quartz rock. It is usually pale grey in colour, but sulphide dust and graphite darken it and chromite dust gives it a brownish hue. Occurrences of fuchsite and tawmawite near the contacts with the black schists may tint the rock green. Parallel to the foliation, the quartzite commonly contains green layers of chrome tremolite and chrome diopside. Uvarovite crystals are sometimes encountered, and rarely chromian tourmaline, phlogopite and chlorite. Feldspar is absent, except in some thin intercalations near the contact with the black schists and occasionally adjacent to the cordierite antophyllite rock. There are occasional carbonate grains. Under the microscope no detrital heavy minerals have been seen. The quartzite contains an average of 0.10—0.15 % nickel in pentlandite.

The Outokumpu quartzite is distinguished not only by its unique mineralogical and chemical composition, but also by its characteristic texture. It usually has a very distinct foliation due to the flattening of the quartz grains, which is accentuated by the subparallel calc-silicate layers ranging in thickness from a few millimetres to several centimetres. However, there are also weakly foliated or massive granoblastic varieties in certain tectonic positions. The foliated quartzite has without exception a well-developed lineation on the s-plane emphasized by the orientation of tremolite needles. The compositional layering of the quartzite may resemble fine sedimentary layering or lamination (Fig. 6). Under the microscope, however, the texture can be called granulitic (Figs. 8 and 10). Alternating layers of fine-grained and coarser-grained quartz with undulating extinction are bounded by sharp and straight contacts that are obviously traces of shear planes. These are accompanied by fine-grained sulphides and graphite (Fig. 9). Mobile sulphides within the ore are adjusted along these lines (Fig. 8). The banding is most regular on the bc-plane of the F_1 subphase

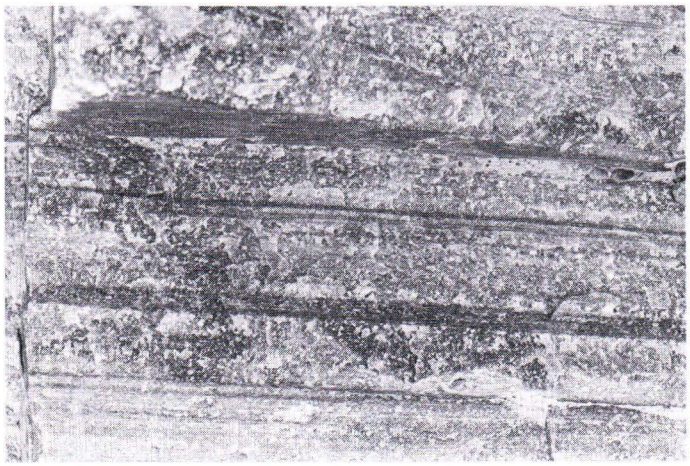


FIG. 6. Layered structure of the Outokumpu quartzite in the bc-section,in natural size 0,5 × 0.75 m. Vuonos open pit.

FIG. 7. The structure of Fig. 6, under the microscope: typical granulite texture of the Outokumpu quartzite in the bc-section. Crossed nicols, 3.5 x magnification. Outokumpu mine, Keretti section.

FIG. 8. Microscopic texture of the layered low grade ore: pyrrhotite and chalcopyrite in layers parallel to s_1. Note the undulatory extinction of the quartz grains caused by s_1. Crossed nicols, 3.2 x magnification. Outokumpu mine, stope KL4.

FIG. 9. Stylolite in the Outokumpu quartzite. Black in the stylolite: carbonaceous material with sulphide dust. Parallel nicols, 4 x magnification.

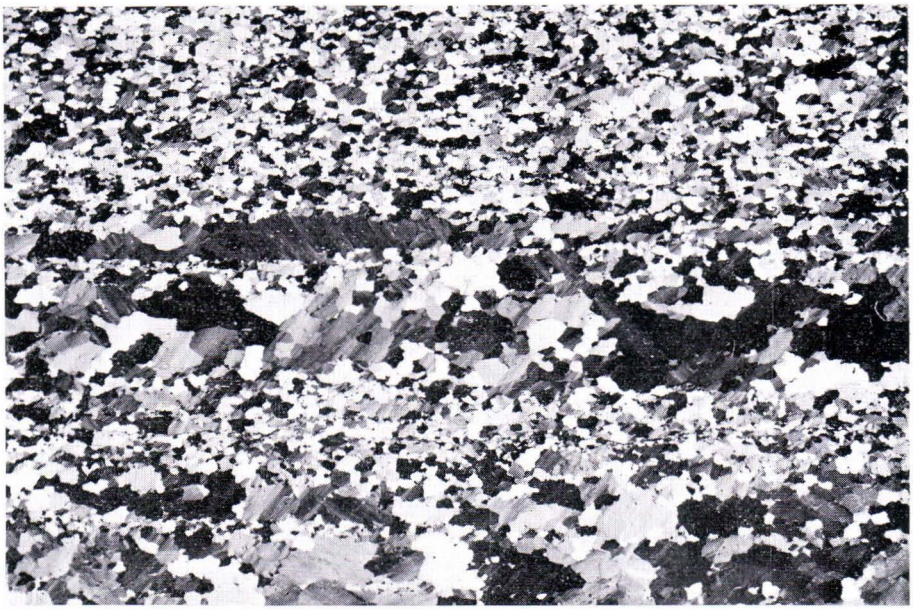


FIG. 10. Granulitic texture of the Outokumpu quartzite in the bc-section. Note the undulatory extinction of the quartz in conjugate directions. Crossed nicols, 6× magnification. Outokumpu mine, Keretti section.

of deformation, while on the ac-plane the different bands are more irregular and interlocking. In the quartzites of Sola, banding and schistosity are less continuous with evidence of deformation of the simple pattern resulting from effects associated with the F_3 phase of deformation in this particular area.

Stylolites are interesting and not uncommon minor structures in the quartzites (Fig. 9). Two separate stylolites have been observed to cross.

Petrofabrics of the quartzite

Nine samples of the Outokumpu quartzite from the Keretti section of the Outokumpu mine were studied for quartz orientation. The sample sites together with the orientation diagrams are shown on Fig. 11. The quartzite of the section can be regarded as a single layer from a few metres to 120 m thick that has been effected by strong similar folding and thrown into tight chevron folds.

The diagrams of the section can be divided into two groups. On the northwestern side, which shows the effect of strong penetrative deformation and an average dip of 130°/60° of the F_1 axial planes, two-girdle diagrams (QD 1, 2, 3, 5) of orthorhombic symmetry predominate. Diagrams QD1 and QD2 show strong maxima parallel to the a-axis. Diagram QD4 is an untypical one-girdle figure. On the southeastern side of the section, the average dip of the F_1 axialplanes is 115°/25° and the

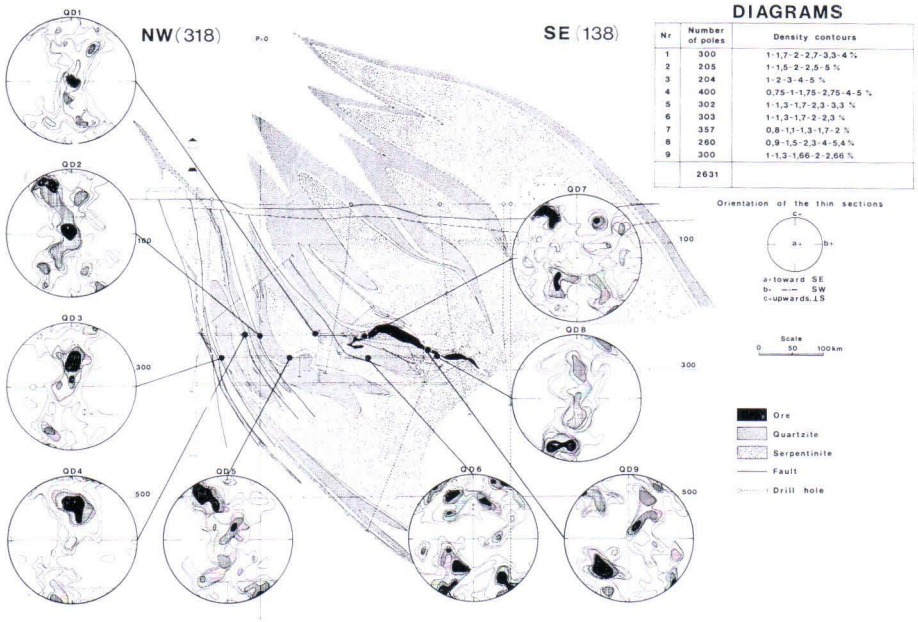


FIG. 11. Quartz orientation in the Outokumpu quartzite. Keretti section 71—72, Outokumpu mine.

quartzite is evidently less affected by penetrative deformation. The characteristic geometry of the orientation diagrams in this case seems to be of the »granulite type from Saxony» (QD6 and 7). The orientation in QD9 is somewhat unclear but it is reminiscent of the granulite type. QD8 is a one-girdle diagram.

Skarn

Skarns occur as a rock type linking quartzite and carbonate rock. A skarn formation may be only a seam some centimetres thick, but it may be as much as several metres thick. The thickest formations are encountered where quartzite-tongues pinch into the serpentinite (Map 3).

Diopside skarn occurs next to the contact of quartzite. Owing to the chromium content, the diopside is green and often zonal. The rock is almost monomineralic or tremolite-bearing, usually with some carbonate. A common accessory is pyrrhotite. The grain size ranges from less than one millimetre to several centimetres and occasionally even up to several decimetres. The rock is either weakly or non-foliated. Tremolite (-actinolite)skarn occurs between diopside skarn and dolomite or dolomite-bearing serpentinite. The colour of the tremolite varies from pale greenish to

deep green with varying chromium content. The rock ranges from diopside-bearing tremolite skarn to a monomineralic tremolite rock, frequently with a slight carbonate content. Near the carbonate rocks it changes into tremolite-carbonate rock or talcose rock with tremolite and locally into chlorite-schist with tremolite. The grain size varies from fine- to very coarse-grained. The rock is often foliated, but it may also be quite massive. Numerous chrome minerals have occasionally been met with in skarn in addition to those mentioned above, e.g. uvarovite, zincian chromite (Thayer, 1964), fuchsite, kaemmererite, rarely eskolaite (Kouvo & Vuorelainen, 1958) and its pseudomorphs (Vuorelainen, Häkli & Kataja, 1958), tawmawite and in tremolite skarn chromian tourmaline (Peltola, Vuorelainen & Häkli, 1968).

According to Huhma & Huhma (1970), these skarns differ distinctly from those of the epicontinental series described later in this text. The absence of feldspar and biotite is conspicuous as is the chrome and nickel content, which is about twenty times higher than that in the epicontinental skarns.

Carbonate rocks

Varying abundances of carbonate rocks occur between tremolite skarn and serpentinite being anything from negligible to several metres thick. They gradually grade into serpentinite. Carbonate-bearing talcose rocks or a pure talc-rock are sometimes present instead of carbonate rocks. Chlorite may be abundant together with talc.

Two types of carbonate rocks occur. In the western zone of the serpentinites, the carbonate is dolomite; in the eastern zone it is magnesite. Pure carbonate rocks are pale grey to grey in colour, homogeneous and equigranular. However, they are seldom pure. They are dotted with serpentine near the serpentinite. In places they exhibit a texture of ophitic appearance (Vähätalo, 1953) or a more irregular association between carbonate rock and serpentinite. The tremolite content gradually increases near the skarn, and talc and chlorite abound locally. Pale brown mica and kaemmererite may occur occasionally. Chromite, pyrrhotite and pentlandite are common acessories.

Serpentinization

Serpentinization must have been the dominant factor in the petrological processes producing the rocks of the Outokumpu association in their present form. A. Huhma (1970) came to the conclusion that the Outokumpu quartzite was originally a chemical colloidal silica precipitate formed by the serpentinization and carbonatization of the intruding ultrabasic magma. In agreement with Haapala (1936), he considers that the carbonate rocks are products of metasomatism associated with the serpentinization. Thus, skarns can be regarded as an indirect consequence of serpentinization, because they were formed as a result of the metamorphic reaction between quartz

and carbonate rocks (A. Huhma, 1970). The petrology of serpentinization has been dealt with by Marttila (1972) and it is discussed here as it relates to the stratigraphical and structural aspects of the problem.

During the serpentinization of olivine, SiO_2 and MgO are freed in accordance with the equation given by Turner and Verhoogen (1960):

$$5\ Mg_2SiO_4 + 4\ H_2O \rightarrow 2\ H_4Mg_3SiO_2O_9 + 4\ MgO + SiO_2 \qquad (1)$$

700 g, 219 cc 72 g 552 g, 220 cc 160 g 60 g

According to Thayer (1966), an addition of 10 % fayalite in natural olivine changes the equation to:

$$5\ (Mg_9Fe)_2SiO_4 + 4.15\ H_2O \rightarrow$$

735 g, 221 cc

$$1.87\ (H_4Mg_3Si_2O_9) + 33^1/_3Fe_3O_4 + 3.4\ MgO + 1.3\ SiO_2 + 0.4\ H_2 \qquad (2)$$

517 g, 206 cc 77 g, 15 cc 136 g 78 g 8, 960 cc

The amount of silica to be removed is greater than for pure forsterite.

The corresponding reactions of enstatite are:

$$7\ MgSiO_3 + 4\ H_2O \rightarrow 2H_4Mg_3Si_2O_9 + MgO + 3\ SiO_2 \qquad (3)$$

700 g, 219 cc 72 g 552 g, 220 cc 40 g 180 g

$$7\ (Mg_9Fe)\ SiO_3 + 4\ H_2O \rightarrow$$

722 g, 219 cc

$$1.9\ (H_4Mg_3Si_2O_9) + 0.23\ Fe_3O_4 + 0.6\ MgO + 3.2\ SiO_2 + 0.2\ H_2 \qquad (4)$$

524 g, 209 cc 54 g, 10 cc 24 g 192 g 4.480 cc

According to equations *1—4*, a comparatively large amount of magnesium and silica has to be removed from the serpentinized ultramafic rocks. Thayer (1966) also plotted the SiO_2 percentage against that of H_2O. The diagram shows the tendency of H_2O to increase and that of the SiO_2 to decrease.

The ultramafic magma was emplaced during the initial stage of the geosynclinal development along a certain stratigraphic horizon near the surface or on the surface. Serpentinization also preceded the first deformation phase, since the quartzite exhibits F_1 structures. Taking into account the fact that no supracrustal rock is encountered between the serpentinite and the quartzite-carbonate rock association, the conclusion must be drawn that serpentinization took place on the bottom of the geosynclinal sea. During serpentinization first SiO_2 and then MgO were freed in

fluids that precipitated around the margins of the ultrabasic masses. The components precipitated on or near the serpentinite bodies. A shell was formed around the bodies in the outer portion of quartzitic and in the inner portion of dolomitic composition. The subsequent recumbent folding and metamorphism of the first deformational phase tectonically disrupted this irregular lithological unit. Its further development is described in the structural analysis.

Cordierite-anthophyllite rock

The cordierite-anthophyllite rock is an uncommon rock member of the Outokumpu formation occurring in quartzite as comparatively sharp, intervening layers. It is seldom more than one metre thick, and its layers are irregular in form. It occurs in an irregular horizon in quartzite-tongues pinching into the serpentinite above the northwestern limit of the ore. This anomalous horizon seems to continue for practically the whole length of the ore.

The cordierite-anthophyllite rock may consist of the two main minerals, or of these and cummingtonite. Pale brown phlogopite, almandine and staurolite are frequent components; small amounts of quartz are common and plagioclase occurs occasionally. The oxides are rutile, ilmenite, chromite and zincian spinel. The most common sulphide is pyrrhotite; in places it is found together with chalcopyrite. The rock is often altered: cordierite into greenish or brownish pinite and phlogopite into chlorite. In several cases almost all the minerals mentioned occur in a single specimen.

The silicate mineralogy of the enclosing quartzite is often exceptional within the Outokumpu formation. Diopside is absent, and an amphibole identified as cummingtonite (abundant twinning in fine-grained needles) replaces tremolite. There is apparently an anomalous nickel content in the horizon with cordierite-anthophyllite rocks. Some copper-cobalt anomalies are also met with as local concentrations of cobalt pentlandite and chalcopyrite both in quartzite and cordierite-anthophyllite rocks. A corresponding horizon with an analogous position vis à vis the copper-cobalt ore has been noted at Vuonos.

Phlogopitic mica schist

Near the northwestern edge of the Outokumpu ore, an uncommon type of mica schist occurs in association with the ore (Plate I/1 and 2). It is lighter in colour than the normal mica schist owing to phlogopitic mica. Besides phlogopitic mica, it also contains quartz, microcline, locally abundant plagioclase with a low anorthite content (down to albite) and accessories. The occurrences are only associated with the Outokumpu association and stratigraphically they form an independent unit of considerable lateral extent.

Ores

The ore mineralizations within the Outokumpu association can be divided into two groups: the copper-cobalt ores of the Outokumpu type and the nickel mineralizations (Huhma & Huhma, 1970). The copper-cobalt ores are found in quartzites, skarns, carbonate rocks extending into black schist or mica schist. The nickel mineralizations have been encountered in quartzites, skarns and in black schist at the contact of the serpentinite. There are no concentrations of nickel in the serpentinites.

The Outokumpu copper-cobalt ore

The ore is an elongated and flat body located in the F_1 fold of Outokumpu within the rock types of the Outokumpu association (Map 3). It is about 4 000 metres long. Its maximum width is 400 metres. It is for the most part less than 10 metres thick but it can be occasionally as much as 40 metres thick. Originally a single body, it has been cut by faults into blocks known as the Kaasila, Kumpu and Lietukka ore bodies (Map 6). Below the main ore there are two small satellite ore bodies, the Poikanen ore and the Turunen ore. The ore crops out in two places that were originally covered by gravel.

The ore is a polymetallic sulphide deposit containing an average of 3.5—4 % copper, c. 28 % iron, slightly less than 25 % sulphur and about 1 % zinc. The nickel content varies from 0.02 to 0.50 %, and the cobalt content mostly from 0.20 to 0.40 %. There are also small amounts of silver, gold and selenium. The original ore reserves were about 28 mill. metric tons.

The host rock of the ore is quartzite. The sulphide-rich parts are actually an admixture of sulphides and quartz or fragments of quartzite. Diopside and tremolite may occur as minor components in the host rock quartzite. The skarn is occasionally mineralized at the contacts and frequently at the southeastern edge of the body. At the northwestern edge, in the main ore and the Poikanen and Turunen ore bodies, there are occurrences of ore layers associated with black schist and phlogopitic mica schist. The hanging wall and the footwall rocks of the ore are shown on Fig. 35.

Two ore types have been distinguished petrographically: the pyrite-predominating and the pyrrhotite-predominating ore. The pyrite-predominating type, which forms the central part of the ore, is enveloped by the pyrrhotite-predominating type. The thickness of this envelope varies from zero to several metres. Both ends of the ore are composed entirely of the pyrrhotite-predominating type (e.g. Fig. 34) as is also the crest at the southeastern edge of the ore. In skarn the mineralization is always pyrrhotite-predominant.

The two petrographic types grade into one another over a short distance that varies from a couple of centimetres to several decimetres in length.

It is important to distinguish between the primary pyrrhotite-predominating ore type and the concentrates of the remobilized sulphides, which may attain the composition of the pyrrhotite-predominating ore. The distinction is illustrated in Fig. 39.

Although the minerals of the two petrographic types are essentially the same, the difference between the types is distinct. This is largely a result of the difference in the relative amounts of pyrite and pyrrhotite, as is implied by their names. A more exact distinction between the types can be made on the basis of the mode of occurrence and the texture of the pyrite.

The pyrite in the pyrite-predominating ore type occurs as euhedral grains or grain clusters, the grain size varying from 0.01 to several millimetres. In the massive ore the abundance of the mineral may approach 40 %. In the pyrrhotite-predominating type pyrite may be absent, but it is frequently encountered at the southeastern edge of the ore. In this ore type pyrite occurs as separate, porphyroblastic cubes of varying size ranging in diameter from minute to 30 cm. The crystals are cataclastic, commonly penetrated by microveins of chalcopyrite and pyrrhotite, and sometimes by sphalerite and quartz. Occasionally the pyrite cubes are surrounded by a halo of sphalerite or chalcopyrite. Pyrite contains cobalt in amounts ranging from a few tenths of a percent to several percent.

Pyrrhotite occurs as grains ranging in diameter from minute to a couple of millimetres, or as groups of grains. In places it exhibits deformation lamellae. Its quantity in the pyrite-predominating type is commonly between 10 and 20 per cent; in the massive pyrrhotite-predominating ore type the quantity may exceed 50 %.

Chalcopyrite occurs much like pyrrhotite, with which it may be intimately associated in places.

Sphalerite is commonly associated with the chalcopyrite, and infrequently as exsolution stars in the chalcopyrite. Locally fairly abundant cubanite has been noticed in only a few places at the southeastern edge of the ore body in the chalcopyrite-rich pyrrhotite-predominating ore and very rarely in the pyrite-predominating ore.

Mackinavite seems to be common together with the cubanite-occurrences (Kouvo, Vuorelainen & Long, 1963) and it replaces pentlandite.

Cobalt pentlandite is a constant accessory associated with pyrrhotite. It occurs both as unmixing lamellae and grains.

Argentian pentlandite is apparently the main carrier of the silver in the ore, and occurs in close association with the chalcopyrite (Vuorelainen, Häkli & Papunen, 1972).

Magnetite is encountered here and there as a minor component in both of the petrographic ore types.

The accessories are numerous and include stannite, galena and gold. In some rare cases molybdenite seems to be associated with the mineralization in the skarn.

The petrographic ore types exhibit a great variety of structural features. The various structural types have been classified by Vähätalo (1953) and Disler (1953). Vähätalo distinguished: 1) disseminated ore type, 2) structural normal ore and 3) brecciated ore. The structural types by Disler (1953) are: 1) layered type, 2) massive type and 3) type with quartzitic inclusions.

In connection with the structural analysis the foregoing classifications have been modified somewhat and the following structural types recognized: 1) layered type, 2) massive type, 3) brecciated type.

The layered type is distinguished by compositional layering that is more or less regular and parallel to the foliation (Figs. 8 and 38; Plate I/1 and 2). This layering is not necessarily primary bedding. Sub-types within this group are layered low grade ore, in which the quartzitic host rock predominates, and layered high grade ore, which is a kind of foliated compact ore with layers of varying quartz content. The layered low grade ore has gradations into an unoriented disseminated type of ore. On the other hand the layered high grade ore exhibits transitions into the massive ore type.

The layered low grade type occurs mostly at the footwall, less frequently at the hangingwall contacts (Figs. 34 and 37) and is typical of the Poikanen ore body (Fig. 31). The layered high grade type is encountered at or near the contacts (Fig. 38) or across the ore, where the ore body is thin or the amount of sulphides is only moderate. The layered structure is more typical of the pyrite-predominant high-grade ore type than of the pyrrhotite-predominant one. The layered low-grade ore is mostly pyrrhotite-predominant at the contacts.

The massive type in its most characteristic form is a homogeneous and unorientated massive mixture of sulphides and quartz, the relative amounts of the various sulphides varying within wide limits. The amount of sulphides commonly exceeds 50 %. The massive pyrite-predominant type is encountered farther from the contact zones, mostly inside the thick parts of the ore. The sulphide-rich pyrrhotite-predominant ore is commonly massive.

According to the structural analysis the brecciated ore type is of tectonic origin. The fragments of the breccia are of the layered and massive ore types. All gradations from folded layered or slightly disturbed massive types to breccia type occur. The matrix is either sulphide crushed and deformed in situ or a mass of remobilized sulphides (Fig. 39).

As stated above, the structural and petrographic classifications overlap, since all the structural types are encountered in both of the petrographic types.

The Vuonos copper-cobalt ore

It is not long since mining operations started at Vuonos and hence a detailed description of the ores must await further results. However, it is clear that the Outokumpu and Vuonos copper-cobalt ores are analogous: they both lie along the same F_1 fold axis within the same structure; they contain the same metals, with individual differences in their contents; their ore types are the same, although in Vuonos the pyrite-predominating type is only a minor component. Both ore bodies are about the same length, but the other dimensions of the Vuonos ore are smaller.

The Vuonos nickel ore

At present a low-grade nickel ore is mined by open pit and underground method in connection with the underground mining of the Cu-Co ore of Vuonos. This nickel ore is unique among the traditional sulphide nickel ores, since it is almost entirely outside the ultramafic bodies in the associated quartzites, skarns and other derivative rocks of the Outokumpu association. No concentrations of nickel occur in the ultramafics themselves. The nickel percentages in the associated rocks are of little more than marginal grade, being from less than 0.20 per cent to some tenths of a percent in the exploitable ore.

The nickel ore is located above the copper cobalt ore of Vuonos, most of it being underneath large serpentinite masses. Nickel concentrations occur chiefly near the lower contact of these masses in quartzite tongues and in skarn-chlorite schist halos around them; they continue upwards along the dip and outcrop along the strike.

The chief nickel mineral is pentlandite as grains and as exsolution lamellae in pyrrhotite. Although pyrite does occur, it is untypical of the parts with the highest grade, where there are zones with anomalous copper and cobalt contents, and where at the same time the pentlandite contains varying amounts of cobalt. Copper occurs in chalcopyrite, while cobalt is chiefly included in cobalt pentlandite. A magnesium-rich, calcium-deficient zone resembling the cordierite-anthophyllite rock zone of Outokumpu has also been observed; it contains e.g. cummingtonite and almandine. The simultaneous concentration of magnesium and nickel is characteristic of the nickel mineralization (Peltola, Kupias, Voutilainen, 1971).

Epicontinental rocks

The mica schists have not been deposited directly on the basement gneiss anywhere along the western margin of the Sotkuma dome being separated from the basement by epicontinental rocks which vary in thickness from 1 metre to 130 metres. These rocks are quartzites, calc-silicate quartzites, diopside skarns, chlorite-biotite schist, various weathering breccias and conglomerates. Between the basement gneiss the sediments there are occurrences of satrolite of varying thickness representing weathering products of the basement gneiss.

The parts of the epicontinental group best exposed have been mapped in detail (see Figs. 12 and 13).

West of Sotkuma (Fig. 12) there are fairly large outcrops where the basement gneiss grades into satrolite and from satrolite into breccia with coarse-grained diopside skarn matrix and fragments of the satrolite basement gneiss and fine grained schists rich in graphite and sulphides. The amount of satrolite rock gradually decreases until it occurs as individual fragments surrounded by diopside skarn. The rock obviously represents a rubble-like weathering breccia of basement gneiss with calcium-rich

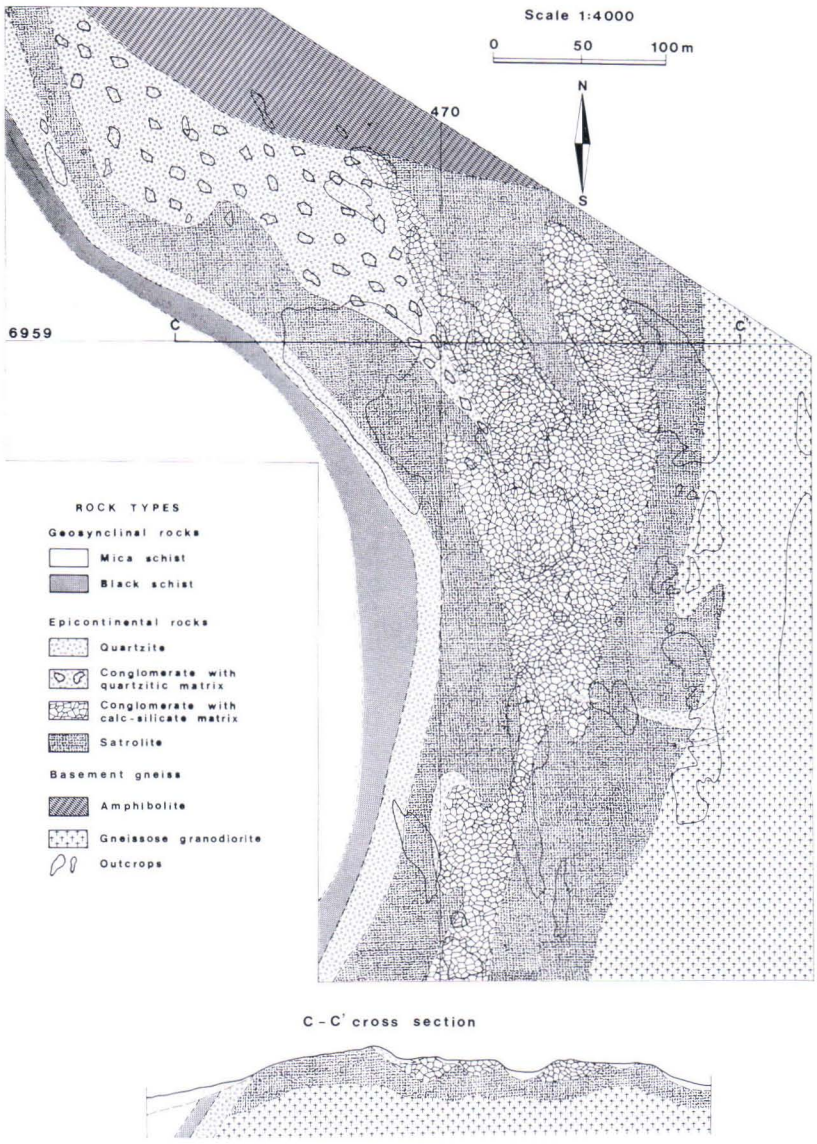


FIG. 12. Detailed geological map of the epicontinental rocks west of Sotkuma.

matter deposited in the interstices between the fragments. In places the skarn matrix grades into clastic quartzite and the breccia with the quartzitic matrix seems to overlie the breccia with the skarn cement. Thus the sedimentation on top of the satrolite began with a calcium-rich matter and this was followed by the deposition of quartzite and black schist.

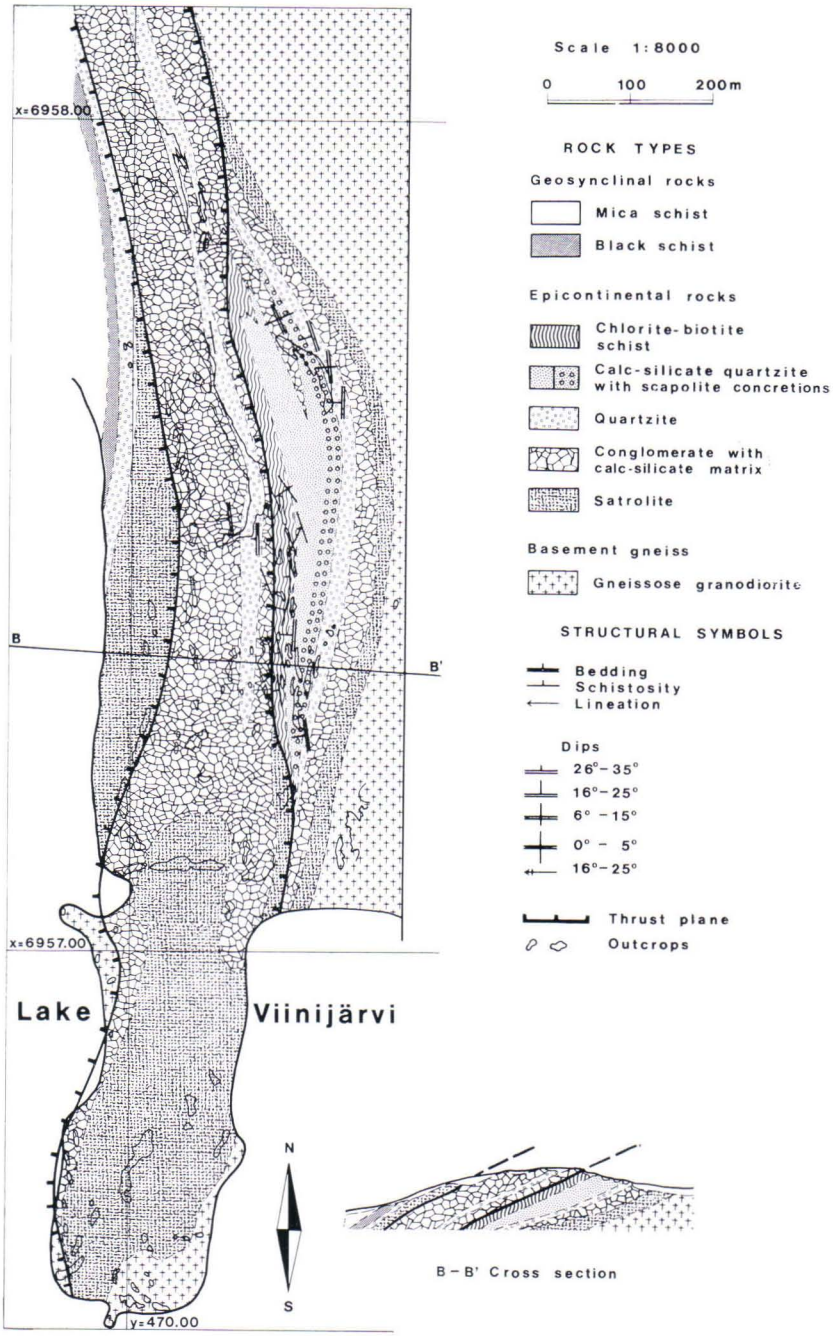


FIG. 13. Detailed geological map of the epicontinental rocks of Pöhönniemi on the shore of lake Viinijärvi.

At Pöhönniemi (Fig. 13) the sedimentation also began with calcium-rich matter. The epicontinental group starts here with satrolite weathering breccia (in the southern part of Pöhönniemi) in which the joints of the satrolite fragments are filled with coarse-grained diopside skarn. Throughout the whole area the bedding dips gently westwards and imbrication has caused disturbances in the original stratigraphic sequence. The rock types met in outcrops from east to west are (cross section B—B′, Fig. 13) quartzite, scapolite-bearing calc-silicate quartzite, diopside skarn, chlorite-biotite schist, conglomerate with a diopside skarn cement, quartzite with conglomerate portions, conglomerate with a diopside skarn cement, satrolite and finally a heterogeneous quartzitic schist rich in graphite and sulphides.

Satrolite

Analysis 4, Table 1.

According to Sederholm (1931) satrolite is a rock type that formed from the material of the ancient continent weathered in situ.

TABLE 1. The elemental compositions of the epicontinental rocks. Analyzed in the laboratories of the Outokumpu Company in Pori.

	1	2	3	4
SiO_2	60.1	71.0	57.6	75.4
TiO_2	0.19	0.41	0.16	0.15
Al_2O_3	7.4	11.0	6.7	14.2
Tot Fe	0.73	1.34	2.10	0.62
MgO	1.3	2.4	11.8	0.4
CaO	16.0	8.8	14.6	1.5
Na_2O	0.35	0.41	1.2	3.5
K_2O	2.9	0.5	2.2	2.7
P_2O_5	0.04	0.05	0.10	0.03
H_2O	1.4	1.7	1.4	1.2
CO_2	9.5	2.0	1.2	0.13
S	0.10	0.37	0.10	0.10
	100.01	99.98	99.16	99.83
Mn %	0.08	0.05	0.08	0.015
Cu	0.0007	0.003	0.0005	0.0007
Ni	0.001	0.005	0.003	0.002
Zn	—	0.002	—	0.002
Co	0.0005	0.003	0.001	0.001
V	0.005	0.015	0.01	0.003
Cr	0.01	0.015	0.005	0.003
B	—	0.001	0.001	0.001
Ga	0.003	0.007	0.005	0.01
Rb	0.008	0.004	0.008	0.009
Li	—	—	—	—

1. Calc-silicate quartzite rich in calcite, Pöhönniemi, Polvijärvi
2. Calc-silicate quartzite, northern shore of Rukkajärvi, Polvijärvi
3. Diopside skarn, Pöhönniemi, Polvijärvi
4. Satrolite, northern shore of Rukkajärvi, Polvijärvi

Macroscopically, satrolite resembles the pale basement gneiss into which it grades. Features indicating incipient satrolitization are the decrease in the abundance or complete absence of biotite. The intensity of the weathering, both chemical and physical, determined whether the basement gneiss produced loose sand or whether it partly retained its compact rocky consistency.

The mineral composition of satrolite depends primarily on the composition of the rock that had weathered. In addition to residuals from the basement gneiss, quartz and feldspar, satrolite contains abundant authigenic sedimentary material. The predominant authigenic minerals are potassium feldspar and plagioclase, secondary muscovite, chlorite, biotite, tremolite, carbonate, diopside, sphene, apatite, zoisite, prehnite and opaques occur in varying amounts. Of these, sphene is occasionally fairly abundant, obviously as a result of the Ti-bearing solutions liberated during the weathering of biotite. In places, apatite is also abundant. Prehnite occurs in cross-cutting veins.

Potassium feldspar of two generations occurs. The residual potassium feldspar is usually turbid as a result of alteration products. The grains are allotriomorphic, fairly large and optically coherent. Exsolved albite is common in this feldspar. The authigenic potassium feldspar is almost unaltered, hatched microline occupying the interstices between the other minerals and sometimes enveloping the residual grains. Potassium feldspar is often encountered as irregular aggregations of small crystals. The optical orientation of individual grains in these aggregates does not differ much from grain to grain and shows almost simultaneous extinction even though the boundaries of the grains are fairly defined. It is obvious that the recrystallization of potassium feldspar started on numerous small residual grains, acting as crystallization centres. The grains grew until they met each other, thus giving rise to partial arrangement in optical orientation.

Plagioclase is also of two generations. The residual generation is strongly altered, although sometimes the twinning is still vaguely visible. It approaches oligoclase in composition. Potassium feldspar exsolution bodies of irregular shape are common as is the replacement of the plagioclase by potassium feldspar. The two feldspars often form a very complicated intergrowth between the aggregates of quartz grains. The residual plagioclase is less common than the residual potassium feldspar. The authigenic plagioclase is unaltered, untwinned and often distinctly zoned. The grains are smallish, equigranular, partly idiomorphic and occur in accumulations. The anorthite content is invariably higher than in the residual type, although it shows some regional variation (An_{20-40}).

Quartz usually occurs as aggregates, and hence it has spotty appearance resembling clastic texture. The grain size of quartz somewhat exceeds that of the other minerals in the rock. The mineral exhibits strongly undulose extinction.

Satrolite is in general a non-foliated rock. In the originally less decomposed type, the texture of the basement gneiss is partly preserved and feldspars occur as large coherent grains. The intensively decomposed satrolite has a clastic texture: the space

between the clastic quartz grains or aggregates is occupied by a fine-grained cement composed of granulated and authigenic minerals such as feldspar, carbonate, epidote, sphene and muscovite. Exsolution, replacement and recrystallization phenomena are common in this matrix and consequently the texture of the rock is locally very complicated.

Calc-silicate quartzites

Analyses 1 and 2, Table 1.

Calc-silicate quartzites have been encountered on the northern shore of Rukkajärvi and at Pöhönniemi. Their composition is reminiscent of arkose, but their feldspars are clearly authigenic. The rock is light-coloured or greenish grey. The bedding is distinct and the pure quartz-feldspar layers alternate with layers rich in micas or carbonates. The feldspars also exhibit bedded variation, the plagioclase being almost completely absent from certain layers rich in potassium feldspar and vice versa.

The main minerals in the calc-silicate-quartzite are quartz, plagioclase, potassium feldspar and occasional muscovite and biotite. Sphene is the most typical accessory and it is fairly abundant in the rock. Other accessories are chlorite, carbonate, tremolite, sphene, apatite, zoisite, zircon and sulphides.

Quartz varies in grain size from layer to layer the average being from 0.05 to 0.1 mm. Clastic quartz grains are rounded and the undulose extinction is weak but recognizable.

Authigenic plagioclase (An_{65}) occurs as grains with round edges. These are smaller than those of the other minerals and occupy the interstices between the quartz grains. Twinning can be observed although it is not common. It is little altered, and muscovite has only been encountered in it occasionally.

Residual plagioclase (An_{10-25}) has been observed as an accessory. It differs markedly from the secondary plagioclase by the lower An content and more intensive alteration.

Potassium feldspar is clear and fresh cross-twinned microcline filling the spaces between the quartz grains. As in the satrolite, the potassium feldspar seems to have started to recrystallize from numerous centres, and this led to the partial joining and reorientation of the optics of several grains. The result of this process is that large potassium feldspar poikiloblasts have formed in the calc-silicate quartzite around the quartz grains.

Micas are encountered in varying amounts. When present in abundance, their orientation is distinct, which gives rise to well defined foliation. Colourless muscovite is the predominant mica although it is often accompanied by a weakly pleochroic pale biotite that has altered sporadically into greenish chlorite. Zircon inclusions with the pleochroic halos so characteristic of biotite in general are rare.

Calc-silicate quartzite seems to have formed through the action of almost isochemical metamorphism from those sediments in which the quartz grains were embedded in a carbonate-bearing and aluminium-rich clay groundmass. The original sedimentary features are clearly reflected in the clastic texture and in the variation in the compositional layering. During metamorphism K_2O, derived from clay formed muscovite and potassium feldspar. The same effect was produced by the potassium-bearing solutions liberated through the granitization processes. The presence of carbonate was the prerequisite for the formation of An-rich plagioclase. If the amount of Al_2O_3 available was less than that of CaO, some of the CaO remained in carbonate.

The scapolite-bearing calc-silicate quartzite of Pöhönniemi

The special feature of the calc-silicate quartzite of Pöhönniemi is its scapolite content. Scapolite occurs as porphyroblasts averaging from 4 to 7 cm in diameter (Fig. 14). These are ellipsoidal and slightly flattened parallel to the schistosity. They often show a radial structure under the microscope and they are mostly zoned.

The porphyroblasts do not exhibit any distinct variation from one layer to the next and they are encountered throughout the calc-silicate quartzite bed. However, their abundance would seem to be at its lowest in the portions richest in carbonate. They are poikiloblastic with abundant quartz, carbonate, muscovite and titanite inclusion. Placioclase and potassium feldspar do not occur as inclusions.

The composition of scapolite was determined with an X-ray diffraction method as a function of the difference between the numerical values 2 θ for (400) and (112) in accordance with the diagram by Burley, Freeman and Shaw (1961). This diagram gives the composition of scapolite expressed in abundances of pure end members — marialite ($3NaAlSi_3O_8 \cdot NaCl$) and meionite ($3CaAl_2Si_2O_8 \cdot CaCO_3$).

FIG. 14. Scapolite porphyroblasts in the calc-silicate quartzite of Pöhönniemi.

Determinations on several porphyroblasts showed that the abundance of the meionite component varies between 50 % and 67 %. Fifteen point analyses were made from a single porphyroblast. They revealed a distinct zonal structure, the composition in the centre being homogeneous with 65 % meionite, whereas that in the margins varied between 57 % and 67 %.

Scapolite is generally considered as a metasomatic-pneumatolytic mineral that had obtained the necessary chlorine for the formation of marialite from a magmatic source. This mineral is often described as having been formed by the action of chlorine-bearing pegmatite intruding into limestone or plagioclase-bearing rock, thus giving rise to the scapolitization of plagioclase. According to Hietanen (1967), scapolite can be produced by an isochemical process and does not require elements from external sources, the chlorine being primarily present in the pore water of the sediments. However, since scapolite is rare in ordinary marine sediments it is obvious that certain conditions must prevail to maintain the chlorine abundance at a level high enough to warrant the crystallization of scapolite. It seems that the chlorite-biotite schist overlying the calc-silicate quartzite at Pöhönniemi as a primarily clay sediment prevented the escape of chlorine from the sand layer. The metamorphic reactions were initiated as the sediments were buried deeper and the temperature started to rise. Scapolite began to crystallize from the solution or salts occupying the spaces between the sand grains. The porphyroblasts enclosed residual minerals as well as those that had crystallized earlier. Slight variations in the compositions of the solutions gave rise to the zonality observed in scapolite porphyroblasts.

Diopside skarns, conglomerates and weathering breccias with skarn matrix

Analysis 3, Table 1.

Diopside skarn occurs in the epicontinental group either as independent layers or as a groundmass in conglomerates and weathering breccias (Fig. 15). Its major minerals are coarse-grained diopside, potassium feldspar, plagioclase and quartz as well as occasional carbonate and tremolite. Titanite, apatite, epidote, chlorite, muscovite and opaques are encountered as accessories. Diopside skarn is pale grey or slightly brownish in colour. The diopside is very coarse-grained and in places the grains may be several tens of centimetres long. In general, the rock is non-foliated except where it is somewhat richer in tremolite.

Diopside is macroscopically rather pale and microscopically colourless. $cA\gamma$ averages 43°, which, according to Winchell (1956), suggests a composition of 43 % $CaFeSi_2O_6$. Diopside occurs as large poikiloblastic grains containing other minerals as inclusions.

Part of the potassium feldspar is probably residual although most of it is authigenic and fairly fresh microcline, occurring either as large poikiloblasts or irregularly shaped schlieren between the other minerals. The amounts of alteration products vary considerably; albite exsolution bodies are common.

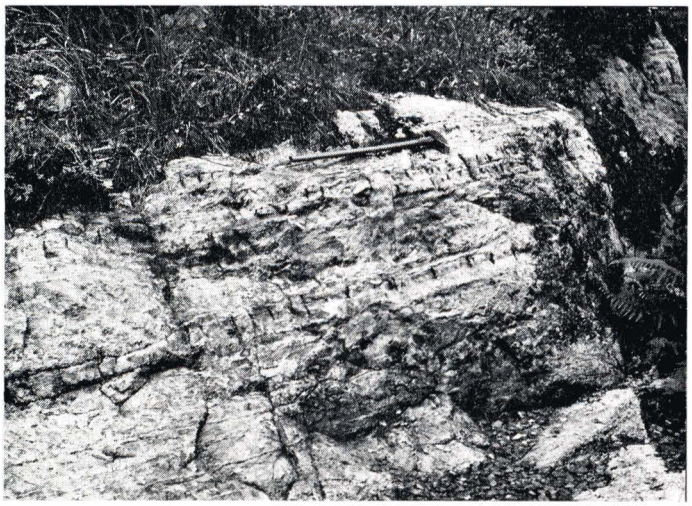


FIG. 15. Conglomerate with diopside skarn matrix. Thin diopside skarn intercalates with cross joints. Pöhönniemi.

Plagioclase (An_{12-36}) forms small equidimensional, angular and partly idiomorphic grains, which frequently occur in aggregates. Twinning is unusual but zonality common. The marginal parts of the grains are almost invariably richer in anorthite than is the centre.

Tremolite is only slightly greenish under the microscope, the grains often being idiomorphic. It is commonly associated with the cross-cutting joints.

Quartz occurs as grains of varying size or as grain accumulations within other minerals. In general, the undulatory extinction is fairly strongly developed. The mineral seems to be residual, originating from the basement gneiss.

Quartzite

The quartzite underlies the scapolite-bearing calc-silicate quartzite at Pöhönniemi. A quartzite layer is also included in the epicontinental group on the south-eastern shore of Rukkajärvi. Further, quartzitic material is encountered as cement in the weathering breccia in the exposures west of Sotkuma.

The quartzite is bluish grey in colour, It has no perceptible foliation or bedding. Its grain size varies from 0.2 to 0.5 mm. In places clastic texture is preserved, but generally it has been destroyed by recrystallization. In addition to quartz, which exhibits strongly undulose extinction, the rock contains residual potassium feldspar. There are also varying proportions (5—30 %) of small satrolite fragments stained by fine-grained opaques and composed of plagioclase and potassium feldspar as well as of their alteration products (Fig. 16). In composition and structure, these fragments resemble the heterogeneous cement occupying the spaces between the quartz grains in the satrolites. In places, quartzite also contains sporadically larger (2—20 cm)

FIG. 16. Bluish-gray quartzite containing abundant satrolite fragments. Crossed nicols. Pöhönniemi.

satrolite fragments. Accessories are muscovite, carbonate, apatite, titanite, rutile, zircon, epidote, chlorite, biotite and ore minerals of which the most common is idiomorphic pyrite.

The quartzite of the epicontinental group differs completely from the quartzite associated with the serpentinites of the Outokumpu association owing to its clastic texture and the presence of feldspars and heavy minerals. The quartzites of the epicontinental group must be considered as metamorphosed inpure terrigeneous sands.

Chlorite-biotite schist

Fine-grained, dark grey chlorite-biotite schist occurs in the epicontinental group at Pöhönniemi. Its chief minerals are potassium feldspar, chlorite, biotite, quartz, muscovite and albitic plagioclace. Graphite, titanite, apatite, tourmaline and opaques are accessories. Zeolites are occasionally met with in veins. The chlorite-biotite schist has calc-silicate quartzitic as well as carbonate- and tremolite-bearing intercalates. Potassium feldspar is the predominant light mineral in the rock. It occurs either as equigranular allotriomorphic grains (⌀ 0.1 mm) or as larger (⌀ 0.5—0.7 mm) porphyroblasts stained by graphite dust and filled with drop-shaped albite exsolution

bodies and quartz inclusions. Biotite shows strong pleochroism. It is largely altered into chlorite with anomalous violet birefringence colour so much that chlorite is often the sole dark mineral. The schist contains minor proportions of plagioclase. The mineral is untwinned and stained by alteration products.

Basement gneiss

The description of the basement gneiss is restricted in the present study to the close vicinity of the epicontinental sediments in the area between Lahnajärvi and Viinijärvi. Here the basement gneiss is a medium-grained, pale grey or slightly pinkish rock. Quartz, potassium feldspar, plagioclase and biotite are the major minerals. Muscovite, chlorite, apatite, zircon, titanite and opaques are the accessories. Close to the epicontinental group the rock is cataclastic and muscovite is more abundant. Foliation is parallel to the contact with the sedimentary rocks. The texture of the basement gneiss further away from the epicontinental contact is granoblastic.

The potassium feldspar occurs either in grains clouded by alteration products or as fresh, cross-twinned microcline. The less altered grains represent a younger generation produced by potassium metasomatism.

The plagioclase (An_{12-20}) exhibits commonly albite twinning as well as antiperthite. It is frequently replaced by potassium feldspar. This has given rise to complicated textures in which the rounded and strongly altered plagioclase relics are surrounded by potassium feldspar. The rock has undergone cataclastic deformation as if frequently indicated by distorted and broken plagioclase grains.

The amount of biotite varies somewhat; generally it is between 5 and 20 %. It occurs as small individual scales or bundles of scales. Pleochroism is strong and the alteration into chlorite common. Zircon inclusions with pleochroic halos are frequent.

According to the age determinations by Kouvo (1958), the zircon from the basement gneiss at Sotkuma gives 2 530 million years for Pb^{207}/Pb^{206} age. The rubidium-strontium dating suggests 1 804 million years for the same rock.

Metamorphic history

The mineral paragenesis of the predominating rock type, mica schist, indicates that the progressive metamorphism has reached in the vicinity of Outokumpu a metamorphic grade corresponding to the amphibolite facies. The mineral paragenesis of the black schists (Peltola, 1960) as well as the calc silicate rocks above the Sotkuma dome fit into the same scheme. The results of the structural analysis given later imply that the peak of primary progressive metamorphism was attained during the first deformational phase.

Several minerals with diagnostic value have been encountered, e.g. andalusite in the argillaceous black schist, kyanite and sillimanite in the mica schist, and the rich mineral association of the cordierite-anthophyllite rock, including almandine and staurolite. Although kyanite and sillimanite have not been observed in Outo-

kumpu in coexisting phase, only nearby, it is worthy to note, that according to the experimental studies by Althaus (1967) and Richardson et al. (1969) the triple point andalusite — sillimanite — kyanite indicate PT-conditions of 600° C and 6 kbar. The mode of formation of the cordierite-anthophyllite rock in Outokumpu is uncertain. In metapelites cordierite and almandine are known to be common in low- and medium-pressure regional metamorphic terranes, while staurolite forms a wide range of rock pressure (Miyashiro, 1973). Although the conditions in the Outokumpu district seem to have been complex, the type of metamorphism probably corresponds to the low-pressure regional metamorphism with comparatively high temperature.

Strong alteration and a new mineral growth in the eastern parts of the vicinity of Outokumpu, e.g. in the serpentinite of Sola, indicate, that there has been a repeated cycle of regional metamorphism under the conditions of the greenschist facies. According to the structural analysis in Sola the peak of this metamorphism was attained during the third deformational phase. In the western part of the vicinity of Outokumpu the alteration is comparatively slight, and when restricted to narrow movement belts, it can be attributed to the prolonged recrystallization down from either of the peaks of metamorphism. Such phenomena are common in the cordierite anthophyllite rock. In the mica schist the plagioclase is sericitized, in some cases epidotized, and contains fine-grained potash feldspar exsolution bodies. The biotite exhibits chloritization and the simultaneous formation of potash feldspar (Peltola, 1960).

The study of sulphur isotopes by Mäkelä (1974) indicates temperatures around 350° for reaching the isotopic equilibrium in the Outokumpu ore. This value can be accepted as a statistical value in concordance with the determinations, but may be indirectly considered to be representative of temperatures associated with the regional metamorphism. The above temperature corresponds to the greenschist facies.

East of the vicinity of Outokumpu, the Höytiäinen synclinorium and the Koli area were affected by a progressive metamorphism up to the greenschist facies (Gaál, 1964, Pelkonen, 1966). There is so far no proof of any other peak of metamorphism in these areas.

GEOPHYSICAL DATA

In the poorly exposed terrain of the vicinity of Outokumpu geophysical maps are of primary importance for the understanding of the geological structures. Geophysically the area is one of the best studied parts of Finland. There is a wide selection of geophysical maps: the 1 : 20 000 aeromagnetic map of the Geological Survey and results of an extensive ground geophysical survey by the Exploration Department of the Outokumpu Company compiled on electric, magnetic and gravimetric maps on scales of 1 : 10 000 and 1 : 20 000. The slingram map, real component, combined with the outcrop map of the vicinity of Outokumpu (Map 2) gives an idea of the area surveyed and the electric anomaly patterns.

A general characterization of the geophysical properties of the main rock types is necessary before the geophysical maps can be understood. The dominant rock type of the anomaly maps is black schist, which, due to its graphite and sulphide content, gives distinct electric anomalies. The anomalies caused by the black schist units become weaker or disappear under a Quaternary cover of more than 20 metres. Black schists also yield magnetic anomalies if they contain pyrrhotite. Some horizons do not show up on the magnetic maps and it can be assumed that they are mainly pyrite-bearing. Black schists are the most important geophysical leading horizons as units in the interpretation of geophysical data.

Serpentinite is usually magnetic because of unevenly distributed magnetite. It can be localized with some probability on the magnetic anomaly maps. On the Bouguer anomaly maps, the chrysotile serpentinites of Outokumpu and Vuonos give negative anomalies but the antigorite serpentinites of the eastern zone of the area distinct positive anomalies. Thus the serpentinite bodies can be generally outlined on the gravimetric maps. Skarns give positive gravity anomalies owing to their highest density among the rocks of the Outokumpu association. If they contain pyrrhotite, even if only locally, they also yield electric and magnetic anomalies. With the exception of some carbon and pyrrhotite-bearing sites, quartzite and dolomite do not regularly produce anomalies. The mica schists form the anomaly-free background of the geophysical maps.

STRUCTURAL ANALYSIS

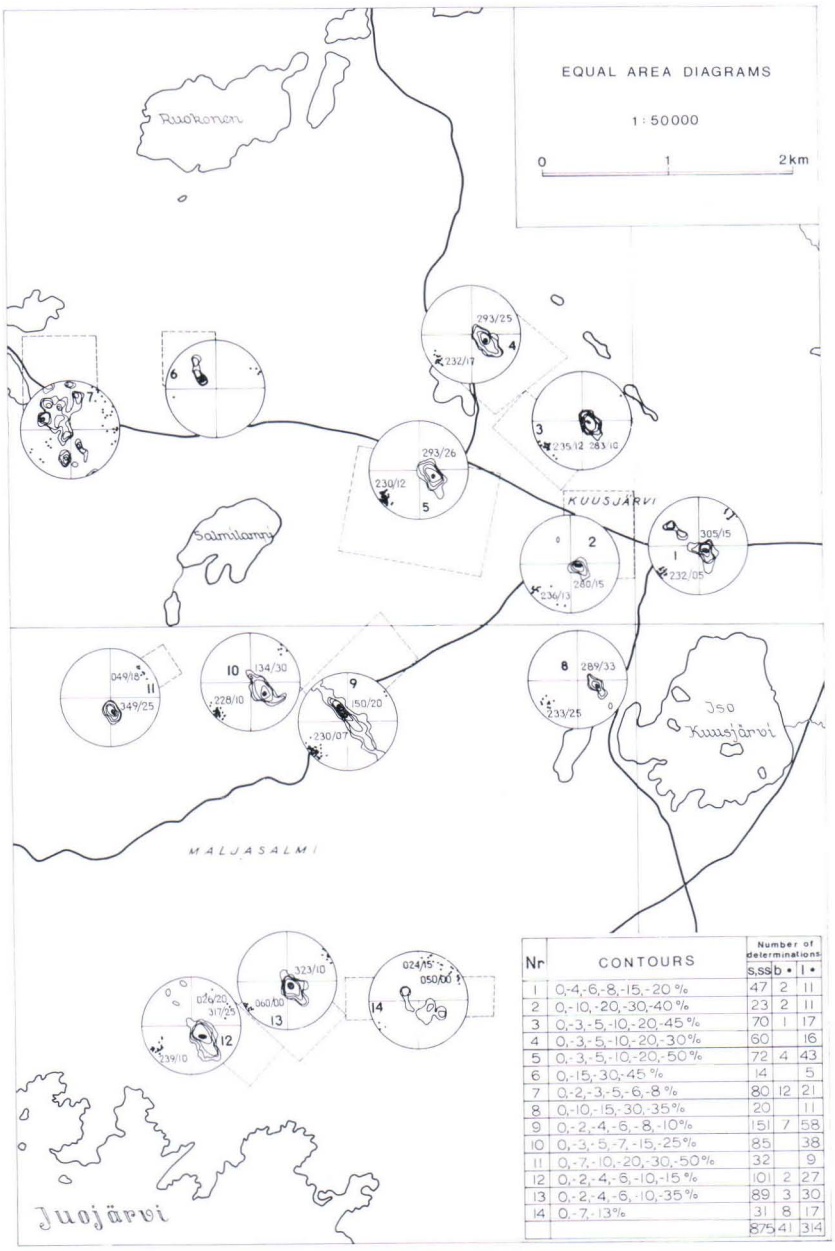
The geological map of North Karelia (Fig. 1) demonstrates the basic problem of the tectonics of the area, viz. the NE strike of the Outokumpu—Vuonos belt differs conspicuously from the NNW strike of the Karelidic belt. To explain this abrupt change, not only the immediate vicinity of Outokumpu but also its broader surroundings must be considered.

General

Methods

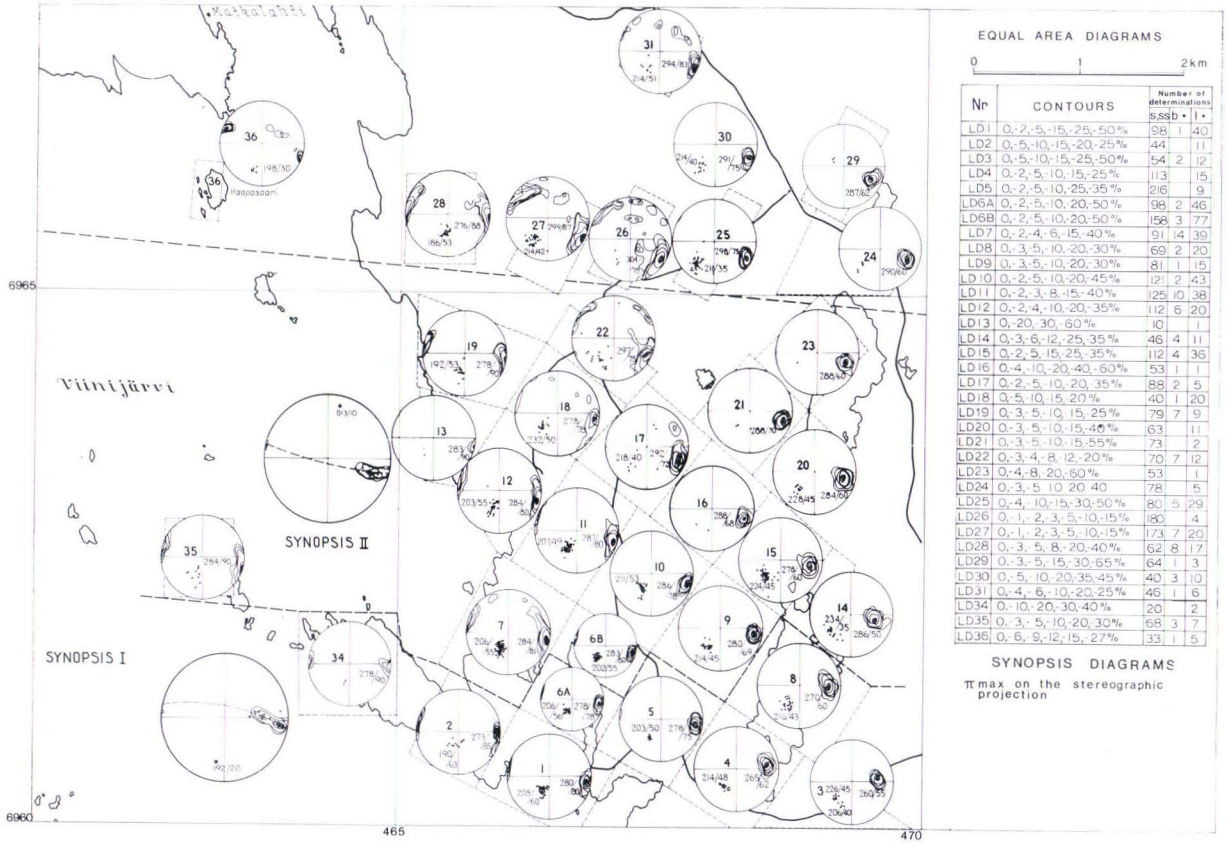
A vast amount of structural data was registered in the field books during the regional mapping. The graphic representation of all data was possible on conventional tectonic maps on a scale bigger than 1 : 10 000. This method is not suitable for printing. Hence we had to choose a method that allows the data to be condensed onto small scale maps without loosing the significance of single elements. The method of Elliot (1965, 1968) was applied to compile on isogon maps the tectonic data of the map sheets of Outokumpu — 4222 and Kontiolahti — 4224, making also use of the data collected outside the vicinity of Outokumpu (A. Huhma, 1971). The isogon maps were constructed on a scale of 1 : 50 000 and reduced to 1 : 200 000 for printing (Maps 4. and 5.). These maps give a sound basis to the structural interpretation of the central part of the Karelidic belt.

The tectonic data of the vicinity of Outokumpu were also plotted on the equal area projection and some of the results are shown in Figures 17 and 18.



Nr	CONTOURS	Number of determinations S,SS	b •	l •
1	0,-4,-6,-8,-15,-20 %	47	2	11
2	0,-10,-20,-30,-40 %	23	2	11
3	0,-3,-5,-10,-20,-45 %	70	1	17
4	0,-3,-5,-10,-20,-30 %	60		16
5	0,-3,-5,-10,-20,-50 %	72	4	43
6	0,-15,-30,-45 %	14		5
7	0,-2,-3,-5,-6,-8 %	80	12	21
8	0,-10,-15,-30,-35 %	20		11
9	0,-2,-4,-6,-8,-10 %	151	7	58
10	0,-3,-5,-7,-15,-25 %	85		38
11	0,-7,-10,-20,-30,-50 %	32		9
12	0,-2,-4,-6,-10,-15 %	101	2	27
13	0,-2,-4,-6,-10,-35 %	89	3	30
14	0,-7,-13 %	31	8	17
		875	41	314

FIG. 17. Structural element west of Outokumpu plotted on the equal area net. Bedding- and s_1 planes contoured, fold axes and lineations drawn as heavy dots. Since the area is characterized by slightly refolded secumbent F_1 folds, the maxima of the s_1 -planes coincide with the flat pi-maxima. Fold axes and lineations are subparallel.



Nr	CONTOURS	Number of determinations s,ssb	•	I •
LD1	0,-2,-5,-15,-25,-50%	98	1	40
LD2	0,-5,-10,-15,-20,-25%	44		11
LD3	0,-5,-10,-15,-25,-50%	54	2	12
LD4	0,-2,-5,-10,-15,-25%	113		15
LD5	0,-2,-5,-10,-25,-35%	216		9
LD6A	0,-2,-5,-10,-20,-50%	98	2	46
LD6B	0,-2,-5,-10,-20,-50%	158	3	77
LD7	0,-2,-4,-6,-15,-40%	91	14	39
LD8	0,-3,-5,-10,-20,-30%	69	2	20
LD9	0,-3,-5,-10,-20,-30%	81	1	15
LD10	0,-2,-5,-10,-20,-45%	121	2	43
LD11	0,-2,-3,-8,-15,-40%	125	10	38
LD12	0,-2,-4,-10,-20,-35%	112	6	20
LD13	0,-20,-30,-60%	10		1
LD14	0,-3,-6,-12,-25,-35%	46	4	11
LD15	0,-2,-5,-15,-25,-35%	112	4	36
LD16	0,-4,-10,-20,-40,-60%	53	1	1
LD17	0,-2,-5,-10,-20,-35%	88	2	5
LD18	0,-5,-10,-15,-20%	40	1	20
LD19	0,-3,-5,-10,-15,-25%	79	7	9
LD20	0,-3,-5,-10,-15,-40%	63		11
LD21	0,-3,-5,-10,-15,-55%	73		2
LD22	0,-3,-4,-8,-12,-20%	70	7	12
LD23	0,-4,-8,-20,-60%	53		1
LD24	0,-3,-5,-10,-20,-40	78		5
LD25	0,-4,-10,-15,-30,-50%	80	5	29
LD26	0,-1,-2,-3,-5,-10,-15%	180		4
LD27	0,-1,-2,-3,-5,-10,-15%	173	7	20
LD28	0,-3,-5,-8,-20,-40%	62	8	17
LD29	0,-3,-5,-15,-30,-65%	64	1	3
LD30	0,-5,-10,-20,-35,-45%	40	3	10
LD31	0,-4,-6,-10,-20,-25%	46	1	6
LD34	0,-10,-20,-30,-40%	20		2
LD35	0,-3,-5,-10,-20,-30%	68	3	7
LD36	0,-6,-9,-12,-15,-27%	33	1	5

FIG. 18. Structural elements west of the Sotkuma dome plotted on the equal area net. Bedding- and s planes contoured, fold axes and lineations drawn as heavy dots. SW-plunging F_1 folds are strongly refolded by the third deformational phase. The subhorizontal axes of the F_3 folds are perceptible in the synopsis diagrams I and II, in which the pi-maxima of the equal area diagrams are plotted on the stereographic projection.

During the field work one of the main aims was to link the minor structures seen in the field to the major structures that govern the pattern of the geological map. The goal was to assemble series of observations that would produce a logical and unambiguous geometric picture with a minimum of hypothetical assumptions. The key to the problem was found in the analysis of minor folds in order to work out the general trend of a formation and its top and bottom. In the vicinity of Outokumpu, effects of polyphase deformation complicate the application of the method but even so the results are positive. The majority of the fold structures of the bedding plane can be designated as either right-hand or left-hand folds. In the case of one phase folding these structures allow the structural bottom of the formation to be determined. In many places observations were made where a subsequent generation of folds with the same or opposite direction of relative movement was superimposed over earlier left- or right-hand folds.

Tectonic style and history of the Outokumpu region

The tectonic patterns of the area are results of polyphase deformation that produced characteristic fold interference patterns such as described by Ramsay (1962).

The first deformational phase produced 045° trending large recumbent folds (F_1) presumable with SE vergency. The trend of this fold set is easy to recognize in the western part of the plunge isogon map (Map 4). In the second deformational phase F_1 folds were refolded into coaxial open antiformal and synformal folds (F_2) with vertical axial planes. F_1 and F_2 folds form interference patterns of type 3 as defined by Ramsay (1962). The third deformational phase threw the previous folds into 015° to N—S trending folds (F_3), which are seen on Map 4 as culminations and depressions in the west and a zone of steep plunge associated with the Sotkuma dome. F_1 and F_3 form characteristic fold interference patterns of type 2 as defined by Ramsay (1962). The folds of the fourth deformational phase (F_4) came immediately after the third deformational phase or developed simultaneously. These 155° trending isoclinal folds with axial planes dipping 30°—90° WSW represent the deformation of the eastern margin of the Karelidic belt east of the Sotkuma dome.

The structural analysis given later indicates that the F_1 and F_2 folds result from the action of a different stress field that acted before the N—NNW trend of the Karelidic belt was formed.

Geometric analysis

First deformational phase (F_1)

Minor folds

The minor folds visible in the exposures of the vicinity of Outokumpu are in majority F_1 folds. They are best preserved in the Outokumpu and Vuonos mines

and SW of Outokumpu where, in some places, the subsequent deformation was relatively weak and their axial planes are still in the original subhorizontal position. In the area of the lake Viinijärvi and east of it, the F_1 folds are superposed by the F_3 folds and are exposed in various oblique sections. They have vertical to moderate dipping axial planes with a S—SW plunging fold axis and were plotted on Map 1 as right- and left-hand folds.

The nature on the F_1 minor folds was studied in several places by mapping the fold structures in detail and analyzing them. The results are summarized in Table 2 and the structures are shown in Figures 19—21.

Major folds

Since in the vicinity of Outokumpu the F_1 folds were refolded during the second and the third phase of deformation, the deciphering of the major F_1 structures is the most difficult problem in the structural analysis. However, as they play a dominant role in the localization of the ore, understanding of their development and subsequent deformation is important. The conclusion drawn from the structural analysis is that the F_1 folds were formed as very large SE-vergent recumbent folds with subhorizontal 045° trending axes.

The study of the minor structures indicates strong thinning of their limbs and thickening in their hinge zones. The occurrence of rodding lineation parallel to the F_1 fold axis indicates elongation along the B axis (Fig. 22). It can be concluded from the anomaly patterns of the geophysical maps that the individual folds were flat-lying chevron folds with extremely large amplitudes and short wave lengths. Folds of this type must have been formed at considerable depths where the PT conditions of the amphibolite facies prevailed. Shear must have been important in the formation mechanism of these folds as is attested by the rock types of the Outokumpu association, which are sheared out in the limb zones but gathered in the hinge zones.

The best known example of a hinge zone of a major F_1 fold structure is the area of the Outokumpu and Vuonos mines. In this study it is called the F_1 fold of Outokumpu. Before the second deformational phase this fold was a recumbent anticline facing SE with a 050° trending axis. During the second and third deformational phases it has been refolded and it can be traced over a distance of 36 km from Juojärvi in the southwest to Polvijärvi in the northeast. Between Keretti and Polvijärvi the axial plane of the fold plunges towards SE at 30° to 60°. This plunging anticline forms a synform, which is illustrated in the vertical cross section in Fig. 26. The limbs and hinge zone of the Outokumpu fold consist of several parasitic folds. In the hinge zones of these parasitic folds, the rocks of the Outokumpu association are preserved as strongly folded and sheared »tectonic inclusions». The lower limb is thrown into NW-vergent minor folds, the most significant of which contain the Outokumpu and Vuonos ore bodies. The upper limb is folded into SE plunging minor folds with SE-vergency as can be seen in the example of Horsmanaho (Fig. 24).

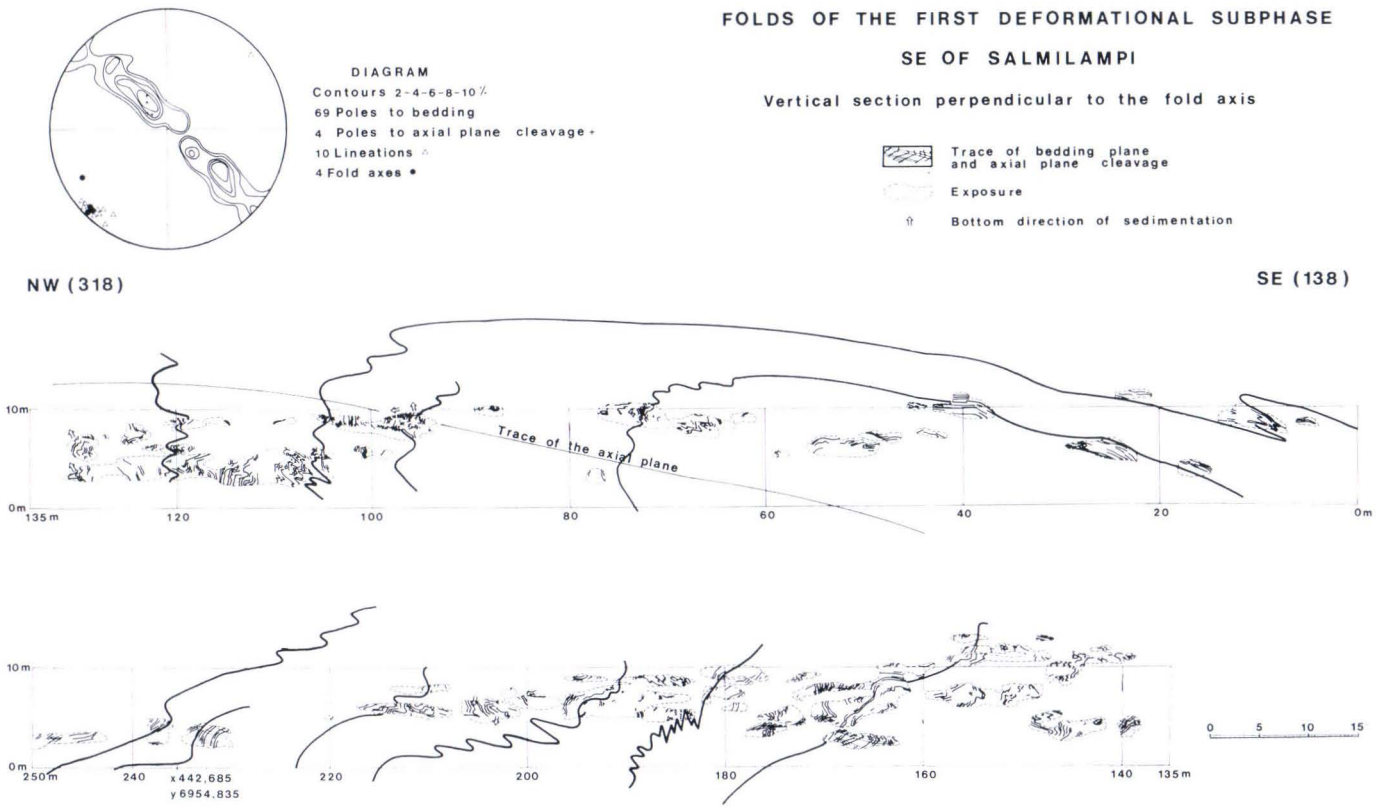


FIG. 19. Folds of the first deformational phase southeast of Salmilampi (x = 6954.83; y = 442.68). Vertical section perpendicular to the fold axis. According to top and bottom determinations, the beds are inverted and thrown into recumbent folds of NW-vergency. Note the strong thickening of the layers in the hinge zones and subparallel axial plane cleavage.

FIG. 20. Folding in the mica schist of Sola (x = 6963.7; y = 466.11). Note the left-hand folded bedding plane (heavy line), F_1, with subparallel to slightly fanned axial plane cleavage which is marked by thin continuous quartz veins and is refolded into right-handed F_3 folds (thin line).

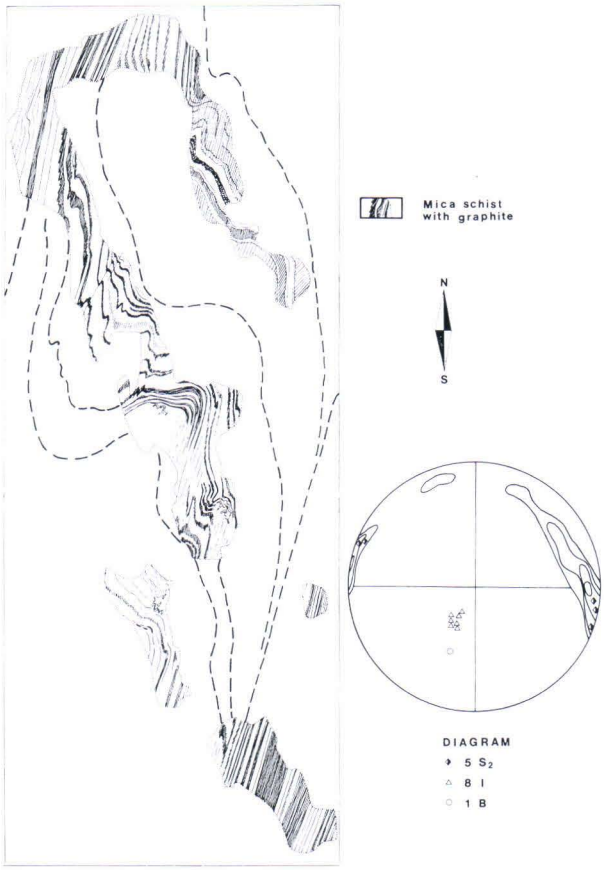


FIG. 21. Left-hand F_1 fold in the graphite bearing mica schist of Haninniemi (x = 6961.42; y = 465.79). This is a flexural flow with tightly subparallel axial plane cleavage. On the equal area diagram: contoured poles to bedding. Scale of map 1:50.

Southwest of Keretti the F_1 fold of Outokumpu bends around the N 30° E striking axis of the Kuusjärvi culmination, and at Maljasalmi its axial plane dips 20°—30° NW. The culmination occupies a zone about 5 km long in the F_1 fold of Outokumpu; it is interpreted as an open fold of the third deformational phase. West of the Kuusjärvi culmination, the recumbent F_1 fold of Outokumpu is folded into the Maljasalmi synform of the second deformational phase. Northwest of Maljasalmi its axis comes out of the ground. Southwest of the Juojärvi depression it rises above the present erosion level with the emergence of the basement in the Juojärvi domes. Here the distance between its lower limb and the basement complex is no more than a few hundred metres.

TABLE 2. Characteristic features of F_1 minor folds

Location	Geometric classification	Attitude of the folds		Mechanism of folding	Interpretation	Fig. in text
		Fold axis	Axial plane			
South of the lake Salmilampi, x = 6954.2 y = 442.8	Similar type of folds, layer thickness ratio: limb/hinge zone ∽ 1 : 7	231°/05°	145°/10°—30°	Flexural flow folds with subparallel axial plane cleavage in the hinge zones	According to top and bottom determinations, the beds are inverted and thrown into recumbent folds with NW-vergency.	19
Poikanen ore body, stope KH8	Similar type of folds	213°/02°	213°/00°—25°	Shear folds	See text	31
Sola, x = 6963.7 y = 466.11	Similar type of fold	180°/60°	270°/90°	Flexural flow folds with subparallel to slightly fanned axial plane cleavage.	The bottom of sedimentation is toward E and the bedding plane is left-hand folded and the axial plane cleavage, marked by thin quartz veins, is refolded into right-handed F_3 folds.	20
Haninniemi, x = 6961.42 y = 465.79	Similar type of fold	205°/65°	285°/85°	Flexural flow fold with tightly subparallel axial plane cleavage	Single left-hand fold, no superposition	21

FIG. 22. Well developed l_1 lineation in the Outokumpu quartzite. Vuonos open pit.

East of Viinijärvi, above the dome of Sotkuma, top and bottom determinations together with the fold analysis allow the traces of the axial planes of the major F_1 folds to be localized (Map 1).

Second deformational phase (F_2)

The first deformational phase was followed immediately by the second phase, which buckled the recumbent F_1 folds into open antiforms and synforms. The axes of the F_2 folds are subparallel to the F_1 fold axes, their axial planes being steep to vertical. These folds are restricted to the southwestern part of the area (Map 1), where two large synforms, the Maljasalmi and the Salmilampi synforms, separated by an antiform, can be distinguished. Southeast of the Maljasalmi synform an antiform with basement gneiss in its core, can be recognized.

Third deformational phase (F_3)

Minor folds

As shown in Fig. 27, the originally recumbent F_1 folds form normal and overturned sets after the third phase of deformation. In the overturned sets, e.g. those between the Polvijärvi antiform and the Sola synform, anticlines form synforms and synclines antiforms. In overturned sets the minor F_3 folds superpose the F_1 folds in a reverse manner: left-hand F_3 folds are superposed on right-hand F_1 folds and vice versa. A case like this is described in a left-hand F_1 fold at Sola (Fig. 20), where the axial plane is folded right-handed. A further characteristic case was studied in detail and is presented in Fig. 23. Here, the bedding plane, which is marked by graphite-bearing mica schist and calc-silicate intercalations, is tightly folded into an isoclinal left-hand fold. This fold, which is a similar fold with axial plane cleavage, is interpreted as an F_1 structure. The axial plane of F_1 which is a zone of strong crushing with segregation of numerous quartz veins, was refolded into a right-hand F_3 fold.

Major folds

Eastwards, the N—S to 020° striking F_3 folds affected the area with increasing intensity. The major F_3 structures are from west to east (Fig. 27): Juojärvi depression, Kuusjärvi culmination, Outokumpu depression, Vuonos depression, Viinijärvi synform, Polvijärvi antiform, Sola synform and Sotkuma dome. The Juojärvi depression and the Kuusjärvi culmination are wide open folds with vertical axial planes striking 010°—030°. The dominant F_3 structure in the vicinity of Outokumpu is the Viinijärvi synform, which is a fairly tight fold striking N—S; its western limb dips SE and its eastern limb W. This synform was originally described as the Outokumpu syncline, later as the Outokumpu »Deckensynklinale» by Wegmann (1928) and still

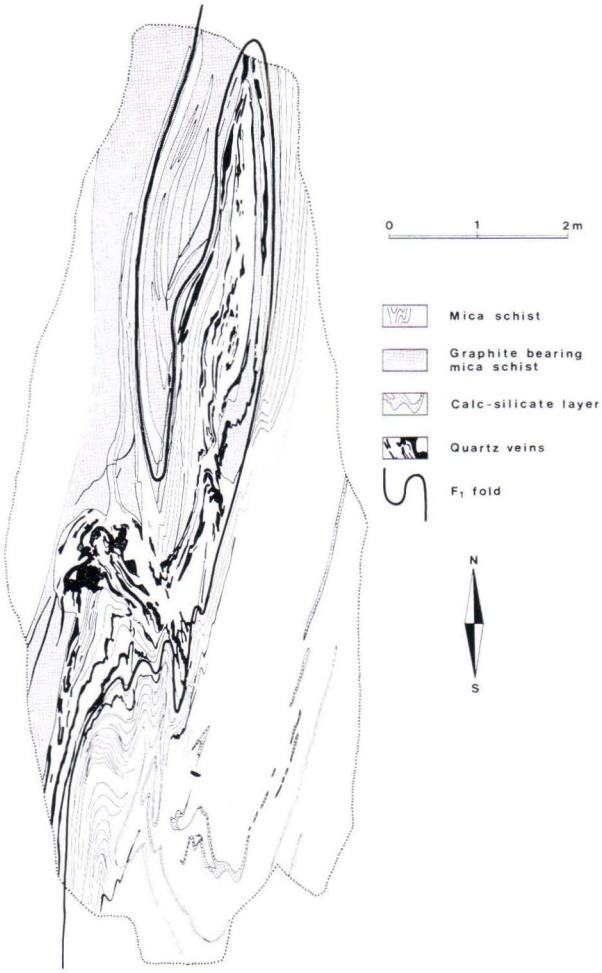


FIG. 23. Left-hand F_1 fold (heavy line) superposed by right-hand F_3 fold, marked by quartz veins, in the graphite bearing mica schist of Sola (x = 6963.45; y = 465.96).

later as the Outokumpu synclinorium by Vähätalo (1953). The next two F_3 folds eastwards, the Polvijärvi antiform and the Sola synform are tight isoclinal folds of a similar type with vertical axial planes striking 010°—020°. The Sotkuma dome rose during the third phase of deformation. This is demonstrated on the diagram map of Fig. 18. The axial planes of the isoclinal F_1 folds, represented by π max in diagrams 1—36, are bent over the dome of Sotkuma around a subhorizontal N—E striking axis. This is illustrated on the synopsis diagrams I and II, which also indicate an E—W striking culmination along Haninniemi.

Fourth deformational phase (F_4)

F_4 folds have not been detected in the vicinity of Outokumpu, except some minor culminations and depressions in the Outokumpu ore body. The domain of the F_4 folds is the Höytiäinen synclinorium, where the Kalevian metaturbidites exhibit 020°—030° strikes (Map 5) and are, according to top and bottom criteria (A. Huhma, 1970), isoclinally folded. Below the lake Höytiäinen the axial planes of the F_4 folds are vertical, but east of the lake they turn over towards ENE. The Höytiäinen synclinorium is cut by several reverse faults parallel to the F_4 axial planes. These faults belong to the same generation that caused the imbricate structure of the Koli area (Gaál 1964, Piirainen 1968). The eastern contact of the Sotkuma dome is also a plane striking NNW and dipping 60°—90° westwards and it is interpreted as a reverse fault belonging to the fourth deformational phase.

The F_3 and F_4 folds meet in a 5 to 10 km wide zone which trends N—S through the eastern half of the Sotkuma dome (Map 4). This zone, which would be a key area for the structural analysis, is outside the vicinity of Outokumpu and its polyphase deformation has not been studied for age relations. Thus there is no conclusive evidence for the age difference between the third and fourth deformational phases.

Evidently F_3 and F_4 structures have been recorded north of the Kontiolahti dome where two phases of deformation have been distinguished (Gaál, 1964). The older s_1 structures, which correspond to s_3, strike N—S. The younger s_2 planes, which would be here s_4, cut s_1 in NW—SE direction. However, farther to the north of the Kontiolahti dome the two phases overlap and can no longer be distinguished.

The vertical cross sections

One of the aims but probably the most difficult task of a geometric analysis is the construction of the geological profiles. There are many obstacles in areas of polyphase deformation and in the most difficult cases, they can make the construction of vertical sections impossible. However, in the vicinity of Outokumpu our thorough knowledge of the geometric relations enabled the profiles to be drawn on various scales. For the ore-critical areas of Sola and Horsmanaho block diagrams (Figs. 24 and 25) have been drawn using vertical sections perpendicular to the direction of the dominant fold axis and projecting surface and drill hole data along the plunge. This method works well in homoaxial areas a few square kilometres in size where rock types and structures continue almost unchanged along the dominating linear structures. The block diagrams of Horsmanaho and Sola give reliable three dimensional pictures in which surface and drill hole data match reasonably well.

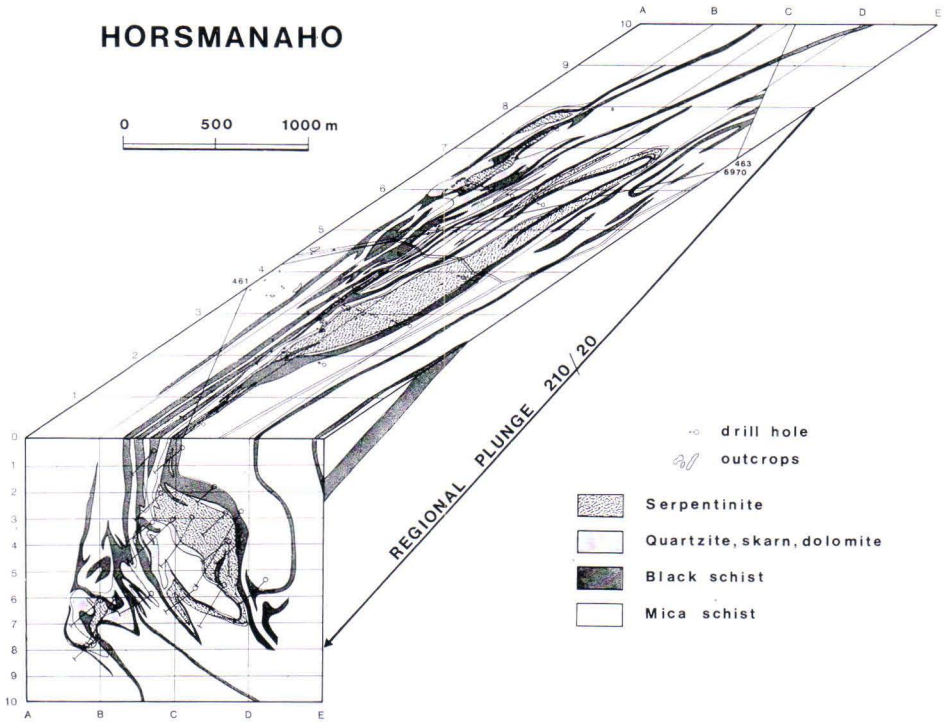


FIG. 24. Block diagram of the surroundings of Horsmanaho.

Only the pinch and swell structures of the serpentinite bodies cause irregularities in the picture.

The main geological profile A—A′ through Sola and Horsmanaho in a WNW—ESE direction (Fig. 26) is more problematic. The profile line was chosen perpendicular to the regional plunges (Map 1) but principal difficulty is related to the refolding of the F_1 folds by F_3. With the associated rise of the Sotkuma dome, the F_1 linear elements were steepened and twisted in a N—S direction (see Map 4) which causes distortion in the picture. The other weak point in the construction is that the lithological units and the structures certainly do not continue unchanged over the distance of projection. The profile must be regarded merely as a generalized representation that is limited by the method employed.

The most remarkable feature of the profile is the great depth of the Viinijärvi synform between the Sotkuma dome and the domal structure of Maarianvaara. Farther to the south the basement is still deeper. Map 4 suggests that its deepest level, between the Sotkuma and Juojärvi domes below Viinijärvi is about 15 km.

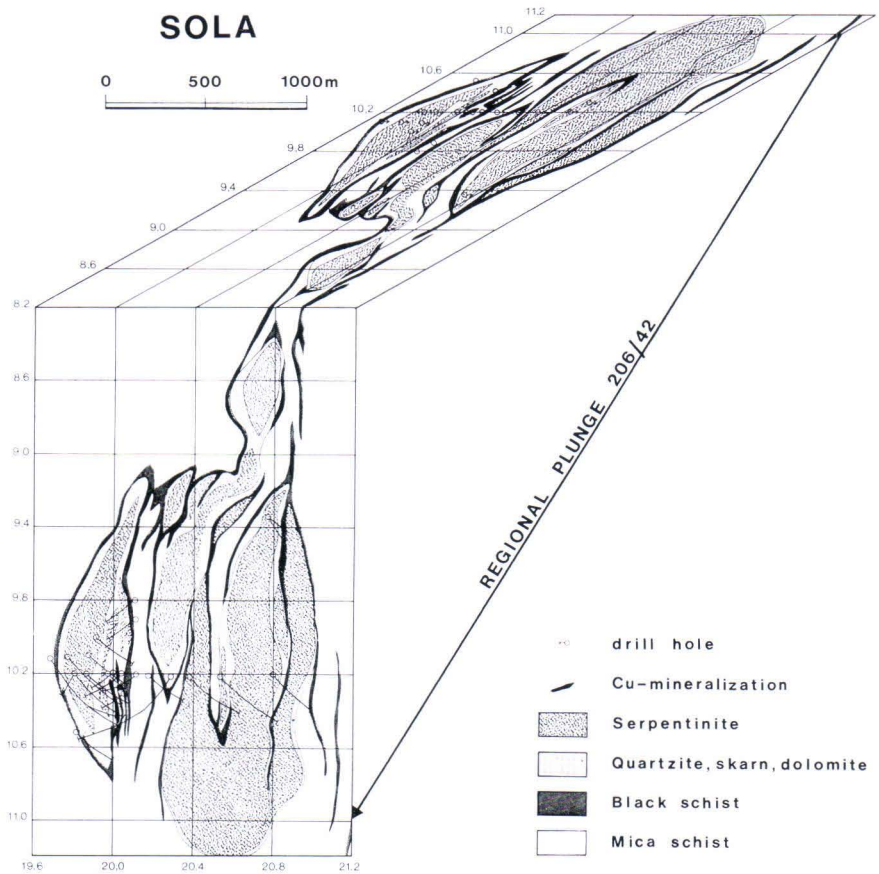


FIG. 25. Block diagram of the surroundings of Sola.

This area is characterized by steep to vertical dips of foliation as seen on Map 5. Such an enormous piling up of geosynclinal sediments could be explained by refolding during F_3.

A seismic profile in a NW—SE direction from Sotkuma across Sola and Horsmanaho was measured by E. Penttilä in 1967. Results indicate a relatively even reflection surface at a depth of approximately 2 km between the Sotkuma dome and Horsmanaho. This has been interpreted as the upper contact of the basement gneiss. According to the structural analysis it is improbable, that the basement would be so shallow in an area with steep structural elements. The reflection surface is more probably caused by some incontinuity within the geosynclinal unit.

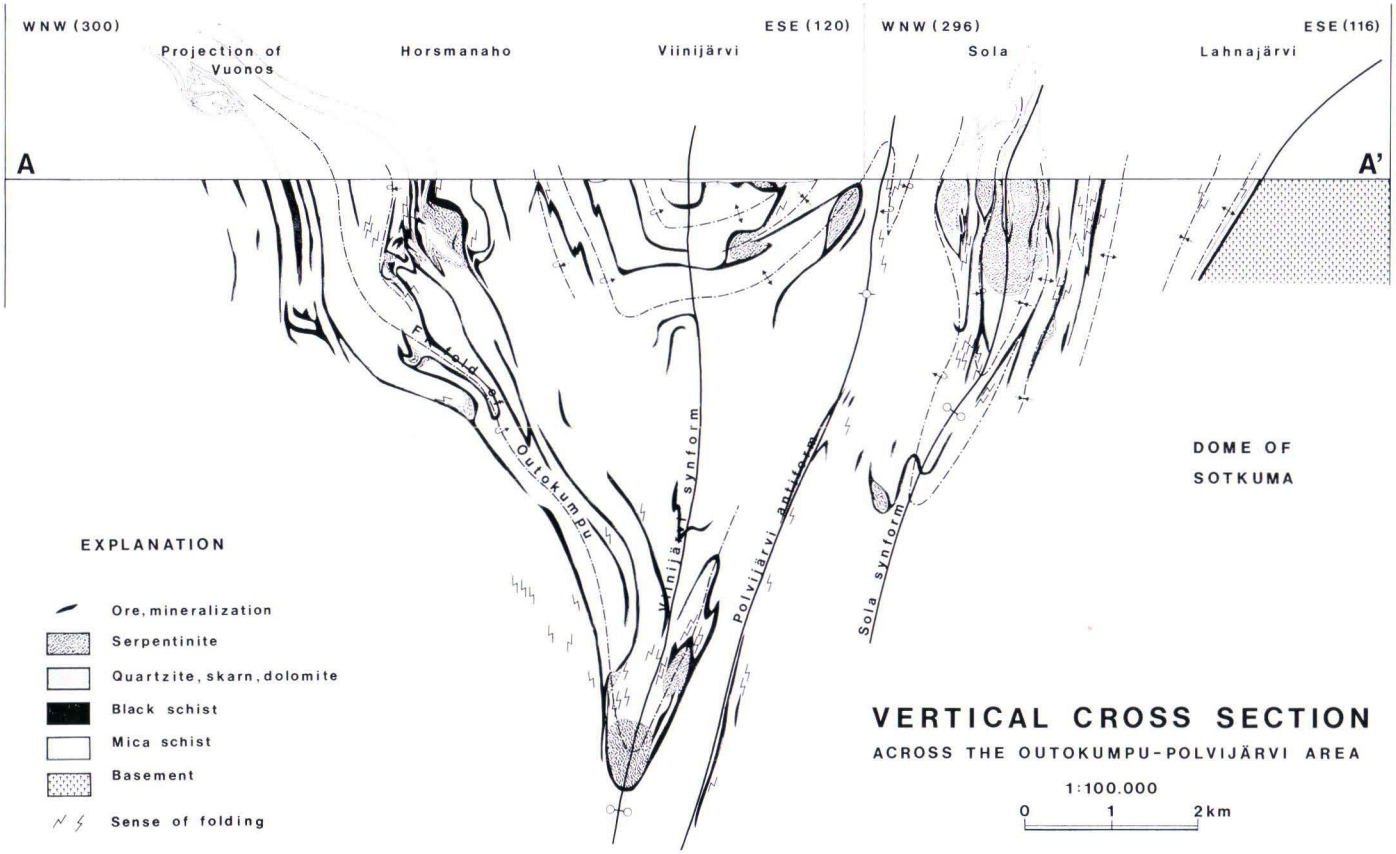


FIG. 26 Vertical cross section across the vicinity of Outokumpu. Data of the geological map projected along the regional plunge on a vertical section perpendicular to the fold axis.

Deformation history of the Outokumpu ore

This chapter deals only with the Outokumpu ore body. The Vuonos ore body, which is also tectonically analogous to the Outokumpu ore, awaits further research, the mining operations being still at an initial stage.

According to the results of the regional study, the ore body is associated with the large F_1 fold of Outokumpu with axial plane dipping 30°—60° to the southeast. The ore is located in a parasitic fold with NW vergency in the lower limb of this structure (Fig. 26). The deformational phases observed regionally have been recognized in the ore body, with the tectonic history of the ore traced back as far as the generation of the F_1 structures. The abundance of structural information from the mine has made it possible to study the nature of the successive deformations in greater detail than in the regional analysis.

The ore and wall rocks were mapped in key areas, special attention being paid to the age relations in the polyphase deformation (Figs. 31 and 34). All available geological data from drilling, surface and underground mapping in the Outokumpu and Vuonos mine areas have been compiled on Map 3. Additionally, a map of the morphology of the footwall and hangingwall contacts of the ore has been constructed (Map 6) and used for the tectonic analysis.

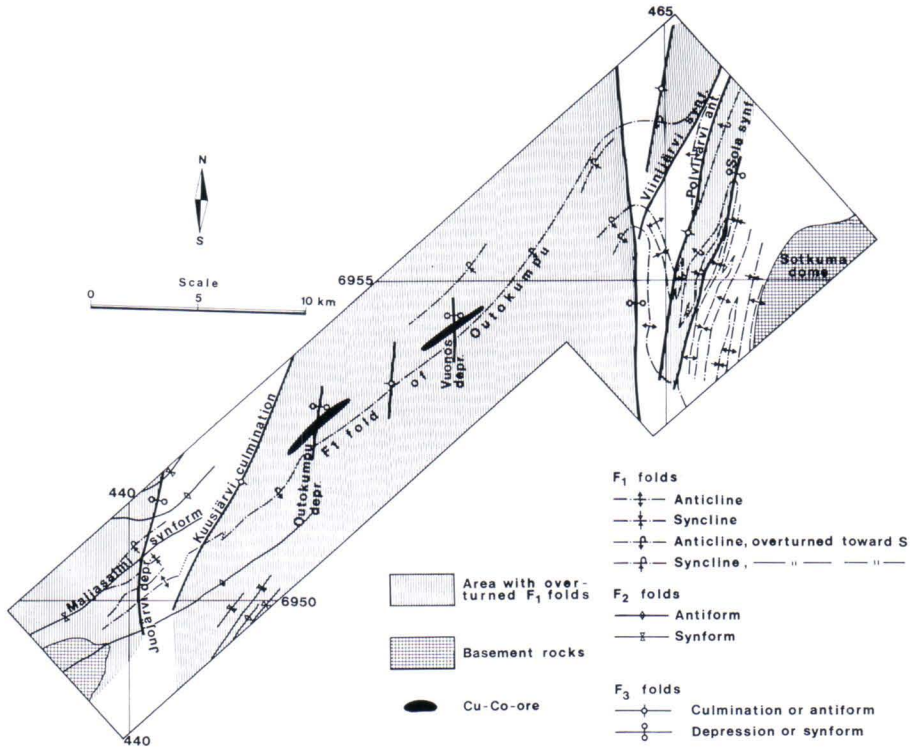


FIG. 27. The structural interpretation of the vicinity of Outokumpu.

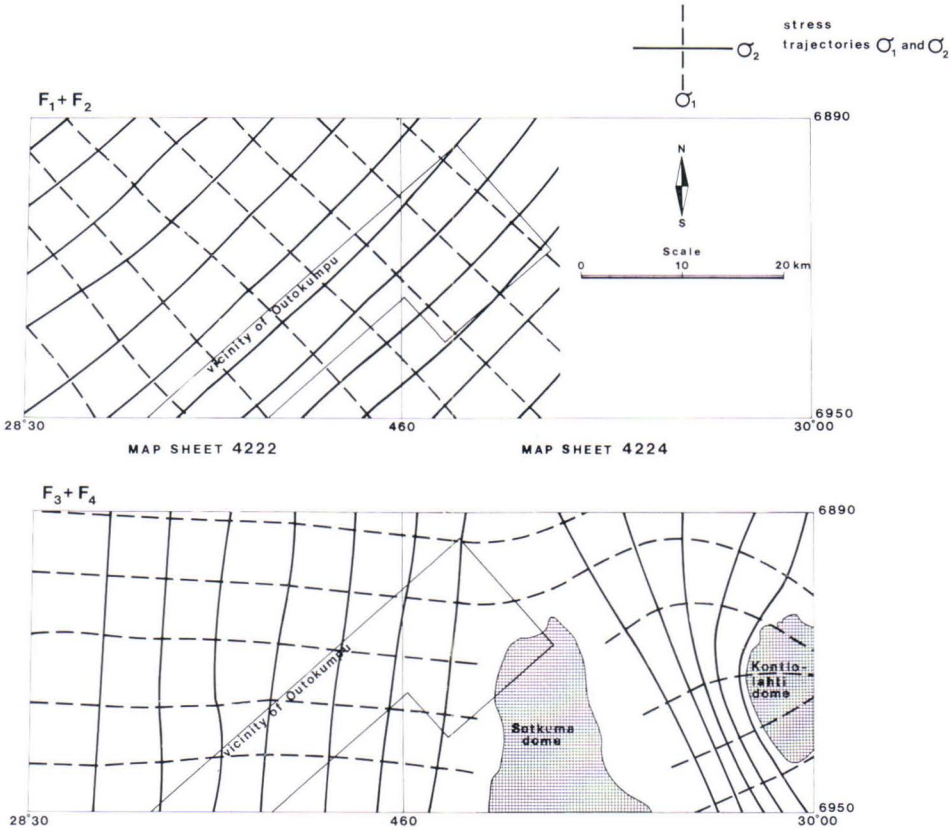


FIG. 28. Inferred stress trajectories during F_1+F_2 and F_3+F_4 phases.

First deformational phase (F_1)

The F_1 structures are well preserved in most parts of the ore and wall rocks, subsequent deformation having been either weak or only locally significant.

Folding

F_1 folding was extremely intense, and F_1 similar folds are the predominant features in the geological profiles on Map 3. These were originally flat-lying chevron folds, later tilted by F_2. In the Poikanen ore body (Fig. 34), where the structures are distinct owing to the lithological differences, (Fig. 31 and Plate I) detailed mapping of s planes and folds disclosed two fold generations: F_{1a} and F_{1b}. F_{1a} represents the deformation determining the general structure. F_{1b} is locally superposed on F_{1a} folds by heterogeneous shear resulting in similar type of folds.

The structures of the first deformational phase are closely associated with the ore body. After the completion of the F_{1a} folding the ore body lay enclosed in an isoclinal F_{1a} fold (Fig. 30). The position of the Poikanen ore body below the upper edge of the main ore body can be explained by the F_{1b} deformation. The F_{1b} folding concentrated on the upper edge of the ore plate and one segment, the Poikanen ore body, moved to a position below the main ore body. The schist horizon associated with the upper edge of the main ore body (Fig. 35), can be followed to the Poikanen ore body (Fig. 34).

Lineation

Lineation (l_1) is a distinct feature of the wall rocks. The host rock of the layered low-grade ore is also lineated as is the layered high-grade ore, especially near the contacts. In the massive ore type the lineation is absent or weakly developed.

The lineation in both the ore and the wall rock is subhorizontal with 045° trend (Fig. 29). The fairly large dispersion in its attitude is mainly due to the effects of later deformation. The traces of l_1 have been drawn as dashed lines on Map 6. These run almost parallel to the elongation of the ore body. However, a slight deviation is visible in the southwestern part, which means that the shape of the ore is not strictly controlled by the B-axis of F_1.

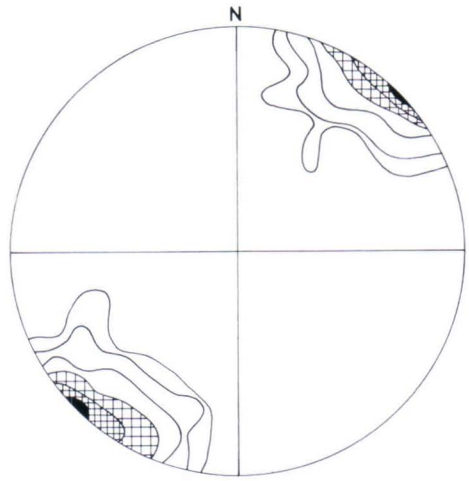


FIG. 29. 260 lineations around the Outokumpu ore body on the equal area projection. Contours: 1—2—5—10—15—23 %. Maximum: 225°/02°.

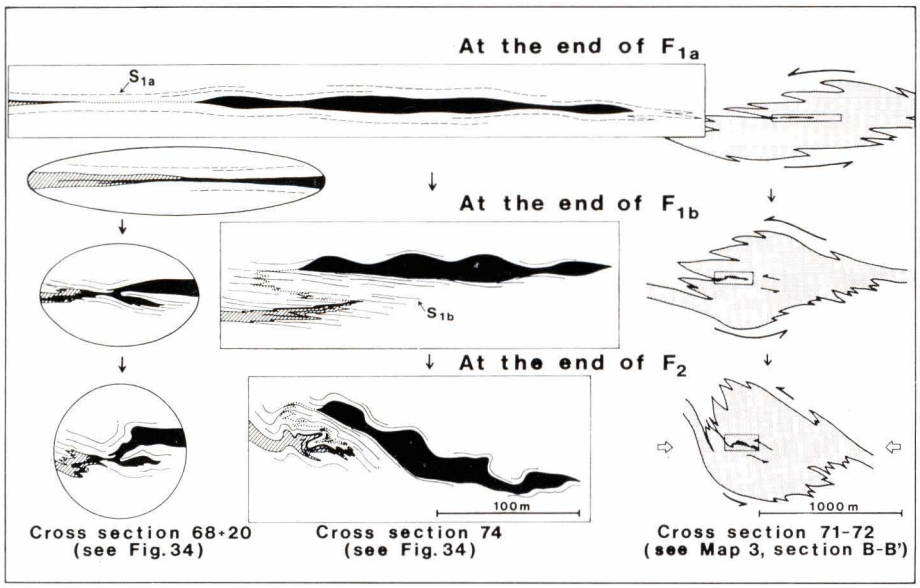


FIG. 30. Schematic picture illustrating the deformation history of the Outokumpu ore during F_1 and F_2. Black: ore hatching: black schist + mica schist.

Foliation

Foliation (s_1) is well developed in the wall rocks, as well as in the layered type of ore. In the latter the sulphide layers are controlled by s_1 (Fig. 8). In the massive ore type s_1 is practically absent.

Two sets of s planes associated with the F_1 similar folds were revealed. The older (s_{1a}) represents the predominant foliation. The younger (s_{1b}) is concentrated in local horizons. Wherever younger lineation (l_{1b}) is present there is only a very slight dispersion of the maximum of the lineation poles because l_{1a} and l_{1b} are homoaxial within the limits of error in the measurements.

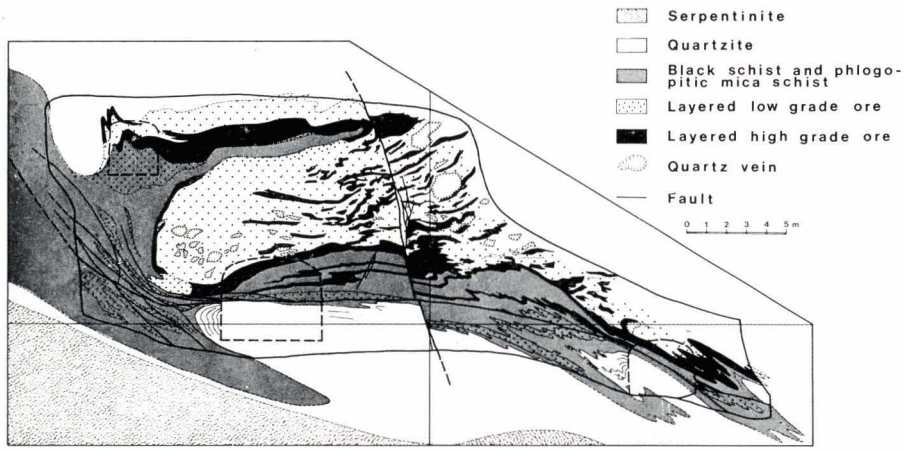


FIG. 31. Detailed map of stope KH8. Poikanen ore body, vertical section 74.

Layering

s_{1a} layering is the predominant planar structure in the ore. This observation is supported by microscopic evidence (Fig. 8). The F_{1b} folding, even though it was deformation of decreased intensity and penetration, was capable of producing its own, secondary s_{1b} layering superposed on s_{1a} layering (Fig. 32). Also in the main ore body some irregularities in the layering seem to be a combination of s_{1a} and s_{1b}.

In his study of sulphur isotopes, Mäkelä (1974) came to the conclusion that the layering in the Poikanen ore body is of sedimentary origin. In the structural analysis primary stratification in the ore cannot be verified, because the earliest history is concealed behind the very strong F_1 deformation. The primary stratification is strongly transposed by F_1 and the layering parallels metamorphic foliation. On the structural basis the layering could be interpreted as either metamorphic segregation parallel to s_{1a} or relics of stratification plus s_{1a}.

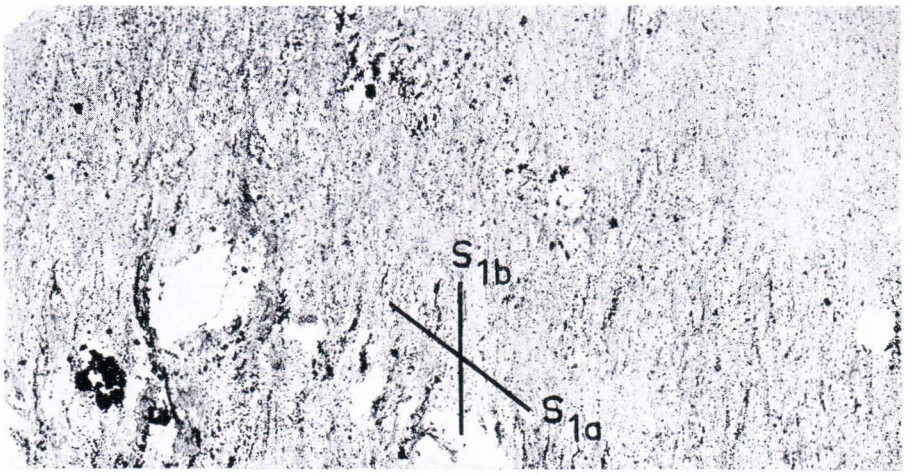


FIG. 32. s_{1b} superposed on s_{1a}, which is defined by vein quartz. Quartzite with low-grade sulphide and graphite content. Parallel nicols, 2.7 x magnification. Outokumpu mine, Turunen ore body, 215 p 2 yp 6.

Mobilization and metamorphism

Mäkelä (1974) states a volcanic origin of the bulk of the sulphur in the Outokumpu ore deposit. The structural analysis throws light on the subsequent history of the sulphides. The tectonic position and characteristics of the ore indicate that the sulphide material was highly mobile during the first deformational phase and the concentration into the present site was controlled by F_{1a} folding. According to Mikkola and Väisänen (1972), the sulphides were remobilized during metamorphism

and deformation and the structures in the ore were greatly changed and many epigenetic features were produced. Among the types of remobilization the metahydrothermal veins containing quartz and sulphides indicate considerable metamorphism and suggest connection with F_1. Thin quartz veins are common in the layered ore as well as in the wall rocks conformable to s_{1a}.

At the end of the first deformational phase the distribution of the two petrographic ore types corresponded in general to the present state (Fig. 34), because F_1 structures are met in both of the petrographic ore types. It could not be concluded whether the occurrence of the two types is a result of an early separation from a common source, the more mobile phase flowing farthest into the fold hinge, or is the pyrrhotite-predominant type a transformation product from the pyrite-predominant type.

Mineralogical and textural investigations do not as yet suffice for accurate conclusions to be drawn concerning the peak of the metamorphism and the peak of the deformation. The regional prograde metamorphism must have reached the conditions of the amphibolite facies during F_1 with mineral growth still during the F_{1b} (Fig. 33).

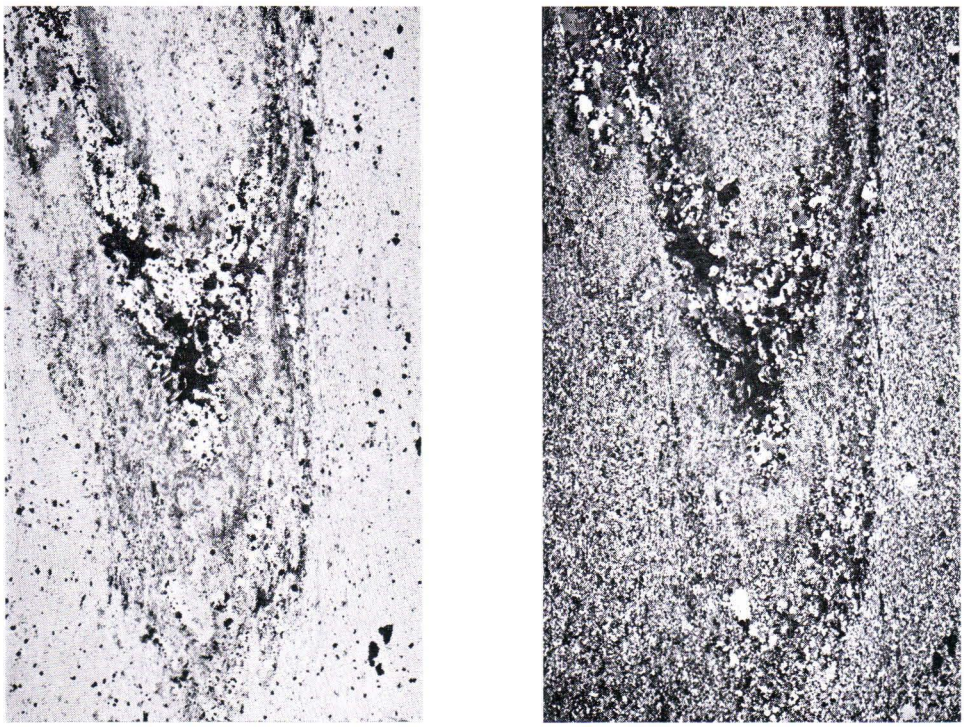


FIG. 33. F_{1b} fold with s_{1b} axial plane cleavage; photomicrograph of a fold shown on Plate I. Parallel and crossed nicols, 2.4 x magnification.

Second deformational phase (F_2)

According to the regional structural analysis, the axial planes of the F_2 folds are steep to vertical and the fold axes subparallel to those of the F_1 folds. The influence of F_2 in the ore and the wall rocks is apparent in three ways. Firstly, the ore became tilted towards the southeast together with the enclosing rocks (Fig. 30). Secondly, the subhorizontal compression caused more or less disharmonic buckling (Fig. 37). Thirdly, the continuous thrusting produced NW-vergent folds that may indicate a close relationship between F_1 and F_2 (sections 88 and 89, Fig. 34). The deformation during F_2 was still highly plastic with some competence difference between the rock components. A striking feature of F_2 is that it concentrated very unevenly, producing

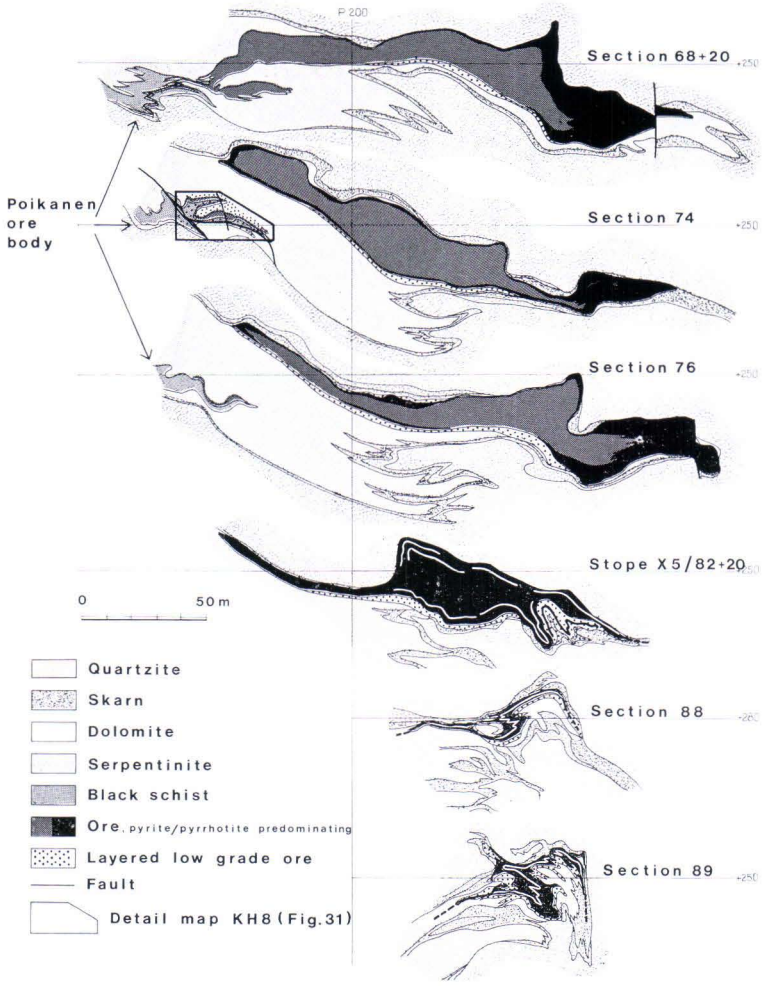


FIG. 34. Cross sections illustrating the deformation of the Lietukka ore body.

intense buckling in some places and leaving others completely unaffected. The contacts of the ore have bizarre shapes where the downward buckles interfere with the upward buckles as in the middle of the Lietukka ore body (Map 6).

The fold axes of F_2 in the southwestern part of the Lietukka ore body (sections 82—88) trend between 238° and 252°, thus slightly deviating from the trend of l_1. Here the attitude of tremolite needles parallels l_1 whereas F_2 deforms the needles (Fig. 36).

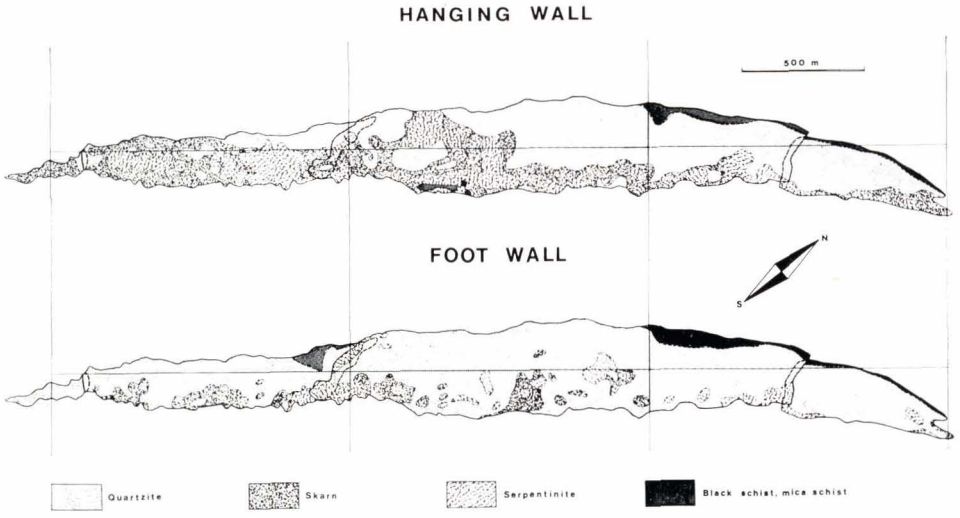


FIG. 35. The hanging wall and the footwall rocks of the Outokumpu ore body.

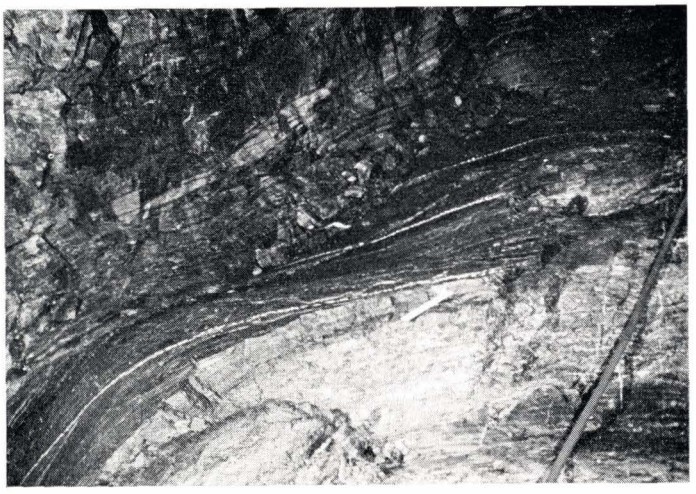


FIG. 36. l_1 deformed by F_2 in the footwall quartzite. Outokumpu mine, stope X5.

ponents. Some of the breccias are composed of pyrite-predominant ore fragments in a remobilized pyrrhotite-predominant sulphide matrix (Fig. 39). The plastic flow type of remobilization of sulphides as described by Mikkola and Väisänen (1972) largely conforms with the deformation typical of F_2. The diffusion of the sulphide material into cutting fractures, as described by the above authors, must be post F_2.

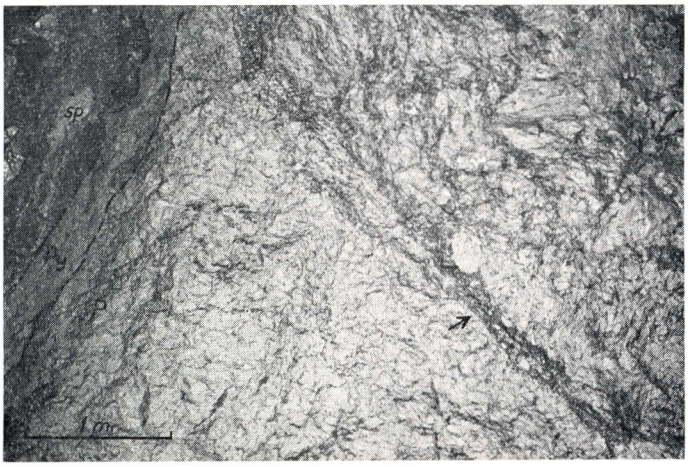


FIG. 39. Wall-rock serpentinite (sp), pyrrhotite-predominating ore (Py) and massive pyrite-predominating ore (P). In the right upper part of the picture the pyrite predominating ore is brecciated and the interstices between the fragments are filled by remobilized pyrrhotite predominating sulphides. Outokumpu mine, stope S9.

Third and fourth deformational phase (F_3—F_4)

The regional structural analysis indicates that the effects of the third deformational phase were fairly weak in the western part of the vicinity of Outokumpu. It produced open folds, culminations and depressions trending 010°—030°. One of the depressions crosses the Outokumpu ore, another the Vuonos ore. The NE-end of the Vuonos ore is again gently plunging NE. In the culmination between them the structural horizon of the ores is above the erosional surface (Map 3). The southwestern end of the Outokumpu ore body and both ends of the Vuonos ore body terminate underground.

There are several smaller culminations and depressions in addition to the larger ones mentioned. They have been drawn on Map 6 on the basis of the attitude of l_1 and the morphologic contours of the ore. Some cross structures are not true culminations and depressions, but macroboudinage structures at the contact of the serpentinite.

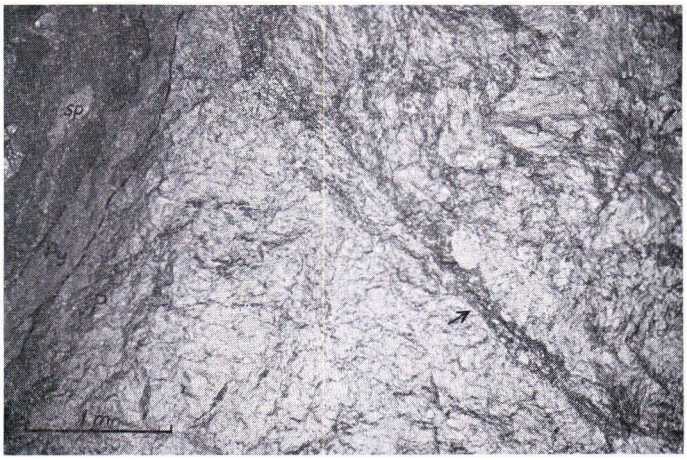
The large depression of Outokumpu is interpreted as an F_3 structure. The general directions of the local culminations and depressions in the ore body would suggest also the existence of F_4 structures.

FIG. 37. s_1 layered low-grade ore deformed by F_2. Massive ore of the footwall present in upper left-hand corner.

During F_2 deformation the mobile sulphides have concentrated in pressure minima. The ore was also boudinated (Fig. 38) during F_2 while in places competence difference between different parts of the ore, e.g. low-grade and high-grade ore, led to the formation of tectonic breccias. They must be post F_1, since the fragments contain F_1 structures and textures. The breccias are characteristic in places where F_2 deformation was especially strong. In these zones brecciation occurred also within the massive ore. In the massive pyrrhotite-predominant ore type the brecciated structure cannot be always distinctly seen owing to the likeness of the breccia com-

FIG. 38. Boudinage in layered high-grade pyrite-predominating ore. Remobilized chalcopyrite and pyrrhotite concentrated in the necks. Outokumpu mine, KCp2.

ponents. Some of the breccias are composed of pyrite-predominant ore fragments in a remobilized pyrrhotite-predominant sulphide matrix (Fig. 39). The plastic flow type of remobilization of sulphides as described by Mikkola and Väisänen (1972) largely conforms with the deformation typical of F_2. The diffusion of the sulphide material into cutting fractures, as described by the above authors, must be post F_2.



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Kinematic analysis

Before reviewing the material of the geometric analysis in terms of movement, the works of Wegmann (1928, 1961) should be cited. Wegmann's ideas had a strong impact on the concepts of the tectonics of the Karelidic belt. He created a synthesis of the geological development of this region applying the alpine model of geosyncline and orogeny. Wegmann's tectonic approach is basically kinematic. His colourful »Bewegungsbild» explains the structures of the area as a product of a single but complicated act of deformation forming nappes, overfoldings, invaginations and virgations in a movement with E vergency. Wegmann compares the mica schists of the Outokumpu region with the flysch of the Alpine orogeny and regards the serpentinites as ophiolites of the Alpine type that were emplaced along the thrust planes of the nappes.

The weak point in the interpretation of Wegmann has been recognized by Väyrynen (1939): Wegmann's cross sections of North Karelia »are obviously drawn on the assumption that the fold axis in the whole of this area pitches due S. Instead already partly at the hills of Juuanvaara and Timonvaara, but especially to the west of them up to the village of Säyneinen, another direction of the axis prevails exclusively, plunging in mean 25° due S 50° W . . ., thus almost at right angles to the axial direction in the Nunnanlahti district and the one found in the whole Höytiäinen basin, the mean for which is 30° S 20° E.» Väyrynen realized that these two areas of different axial directions in the Outokumpu region »are bounded so sharply against each as has been observed on the Juuanvaara—Timovaara line . . .» and it is obvious that »both areas have apparently a kinematic development of their own.» In the present study these phenomena are explained by polyphase deformation. Consequently each deformational phase is assumed to have its own kinematic history.

The structures of the first deformational phase, the large recumbent F_1 folds with a subhorizontal axis trending 045° imply a tangential movement in a NW—SE direction. The movement occurred mainly by folding and the strain was more or less evenly distributed throughout the sedimentary cover. It is difficult to point out any zone that could have been the thrustplane of a nappe during the first deformation phase. The tangential movement could not have been purely translative forming nappes of the »Abscherungsdecke» type of the Eastern Alps with little deformation within the nappe unit. Thrusting occurred by overfolding as in the deformations of the »schistes lustrés» in the Penninikum of Graubünden, Switzerland. Therefore, the vergency of the movement should be determined by fold analysis in areas where the F_1 folds retained their original subhorizontal position. The only place where a complete analysis of this kind could be carried out was the F_1 fold south of Salmilampi (Fig. 19). Here the folds exhibit a NW vergency in inverted mica schist beds. This combination could be caused by movement of overall SE vergency. Assuming the SE vergency and taking into account the evidence of the isogon maps (Maps 4 and 5), the F_1 folds in the vicinity of Outokumpu could be considered to have

originated somewhere northwest of the Maarianvaara granite. However, overthrusting took place before the intrusion of the granite and the formation of the domal structure of Maarianvaara.

The folds of the second deformational phase have no vergency, since their axial plane is vertical. Although they are coaxial with the F_1 folds, they have to be separated geometrically into two different phases, because the second phase deforms the F_1 structures. The F_2 folds developed immediately after the first deformational phase, as it is indicated in the structural analysis of the Outokumpu ore body (Fig. 34, sections 88 and 89). A possible explanation is, that F_2 folds developed at large F_1 fold noses, where a great deal of rock hardening caused buckling of the rocks. Elsewhere F_2 movement just continued by transportation of s_1. Signs of this phenomenon are visible on Fig. 19, in the profile between 200 and 180 metres, where folds with vertical axial plane developed in the hinge zone of the large F_1 fold.

The folds of the third deformational phase were caused by an E—W compression and have no vergency. This is also the case with the structures of the fourth deformational phase below the lake Höytiäinen, where the axial planes are vertical. The F_4 folds show the direction of tectonic movement at the eastern margin of the Karelidic belt, where the Kalevian metaturbidites are pushed along low to high angle thrust planes onto the quartzites and conglomerates of the Koli area toward ENE.

Dynamic analysis

Theoretical and experimental studies during the past ten years have removed much of the obscurity surrounding the stress history of folding (e.g. Dietrich and Carter, 1969). It has been established that the main principal stress σ_1 acts perpendicular to the fold axis and the intermediate principal stress σ_2 parallel to it in large-scale regional folds. By applying this principle the dip isogon map of the linear elements (Map 4) can be used directly for the construction of stress trajectory maps of the ancient stress fields (Fig. 28).

During the first and second deformational phases a fairly uniform stress field was active with σ_1 in NW—SE and σ_2 in NE—SW direction. No information can be obtained for lake Höytiäinen and the eastern margin of the Karelidic belt.

It is tentatively assumed, that the third and fourth deformational phases developed simultaneously within one complex stress field. σ_1 acted in a N 75° W to N 65° E direction as shown on the trajectory map of Fig. 28. The inconsistency of σ_1 can probably be attributed to the effect of the dome structures that were formed during this structural event. Neither the Sotkuma nor the Kontiolahti dome represent merely simple fold interference patterns, rather they are both active tectonic units that acted intrusively on their environment. The problem has been discussed by Eskola (1949) and Preston (1954). The forceful intrusion of the domes into the overlying sedimentary cover produced local concentric stress fields, which were superposed on the regional stress field.

CONCLUSIONS

The changes in the stress fields indicate, that the Outokumpu region has a long tectonic history. This view is supported by the few age determinations, carried out in the Outokumpu ore. Kouvo and Kulp (1961) determined the isotopic compositions of two galena samples and got from the isochron plot Pb^{207}/Pb^{204} and Pb^{205}/Pb^{204} a model age of 2300 m.y. A crosscutting pegmatite dyke in the ore, obviously an apophyse of the Maarianvaara granite, yielded by the Rb—Sr method a much younger age of 1845 m.y. (Kouvo, 1958). The high model age of the galena compared with the common 1800 m.y. age of plutonic rocks intruding the Karelian belt suggests according to Kouvo and Kulp (1961), that unless the leads had an unusual crustal history, the Karelian rocks may have undergone more than one orogenic episode.

The deformational phases can be linked with the results of the age determinations. The ore is controlled by the F_1 folding with the 045° axial trend, which is a strange direction in the NNW striking Karelidic zone. The dated galena must have been crystallized during the first or the second deformational phase and that would give to the F_1 folding a minimal age of 2300 m.y. On the other hand the Maarianvaara granite intruded during the third deformational phase and that was about 1800 m.y. ago. Consequently a minimal age difference of 500 m.y. must be assumed between the first and the third deformational phase. This time span could be enough to comprise the development even of a new orogeny and it is suggested, that the turbidites of the Höytiäinen synclinorium (Fig. 1) and the molasse like sediments of the Koli area became deposited and folded during this time interval.

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The authors emphasize that they take full responsibility for the contents and opinions put forward in the text.

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PLATE I: 1. Folding in the Poikanen ore body: layered type of ore in quartzite and black schist interbedded with phlogopitic mica schist. Compositional layering parallel to S_{1a} which is folded by F_{1b}. Outokumpu mine, stope KH8.

Plate II

PLATE I: 2. An enlargement of the lower left-hand corner of Plate I: 1.

G. G. Gaál, T. Koistinen and E. Mattila:

Tectonics and stratigraphy, vicinity of Outokumpu

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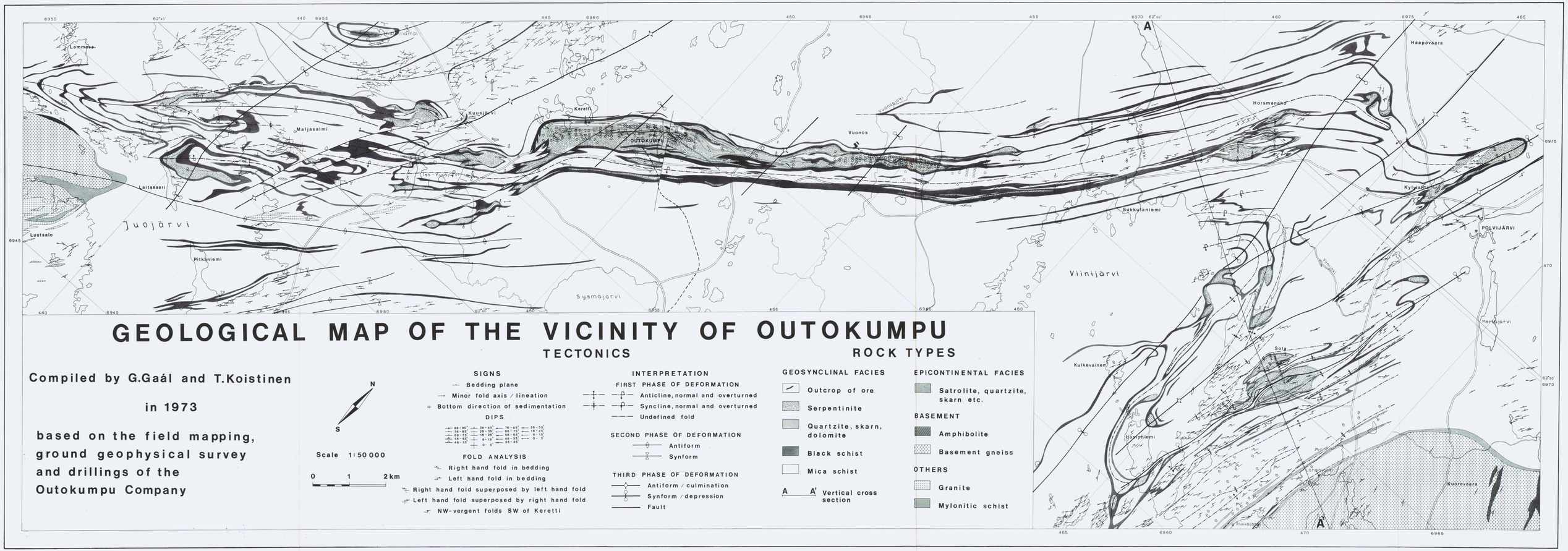

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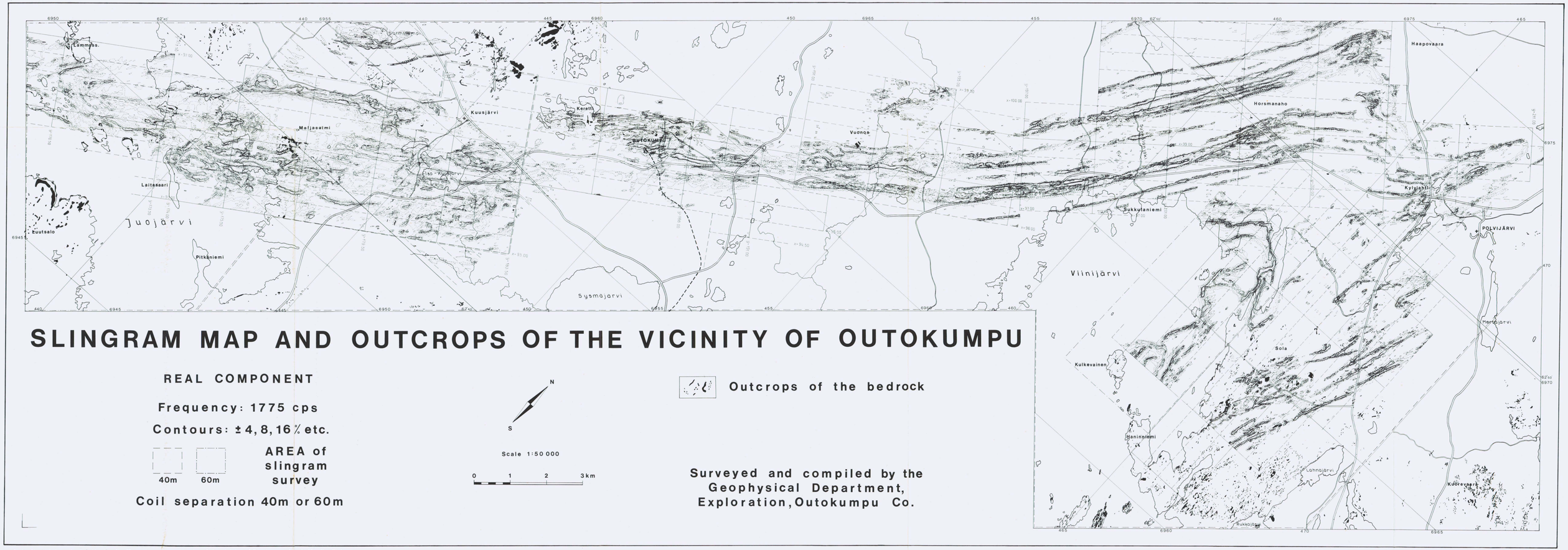

SLINGRAM MAP AND OUTCROPS OF THE VICINITY OF OUTOKUMPU

REAL COMPONENT

Frequency: 1775 cps

Contours: ± 4, 8, 16 % etc.

40m 60m — AREA of slingram survey

Coil separation 40m or 60m

N
S

Scale 1:50 000

0 1 2 3 km

Outcrops of the bedrock

Surveyed and compiled by the Geophysical Department, Exploration, Outokumpu Co.

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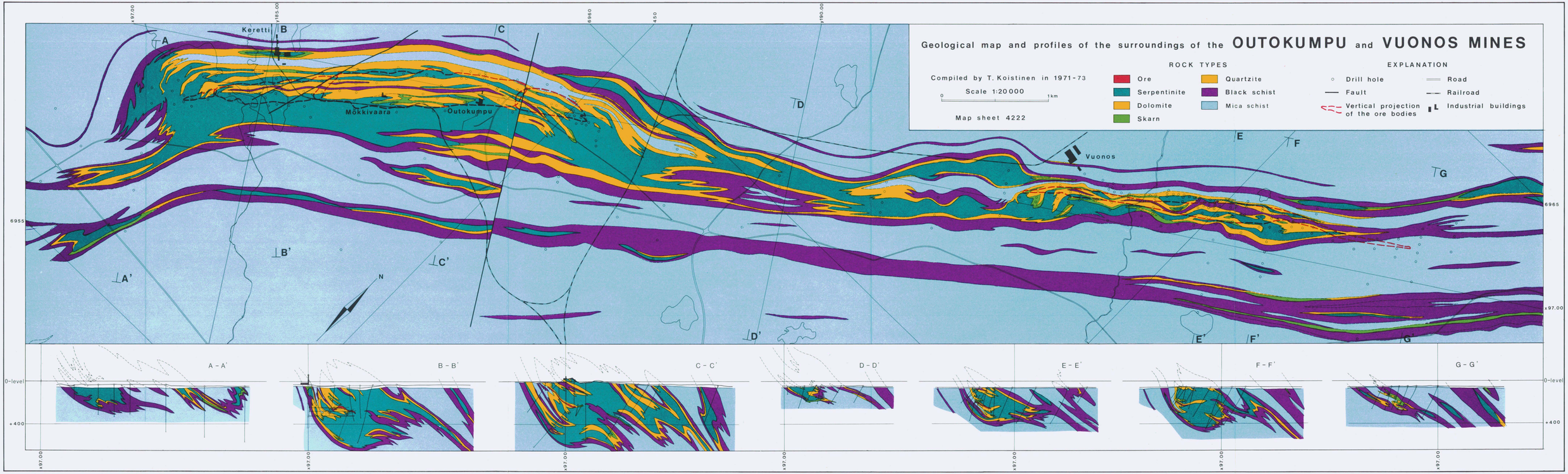
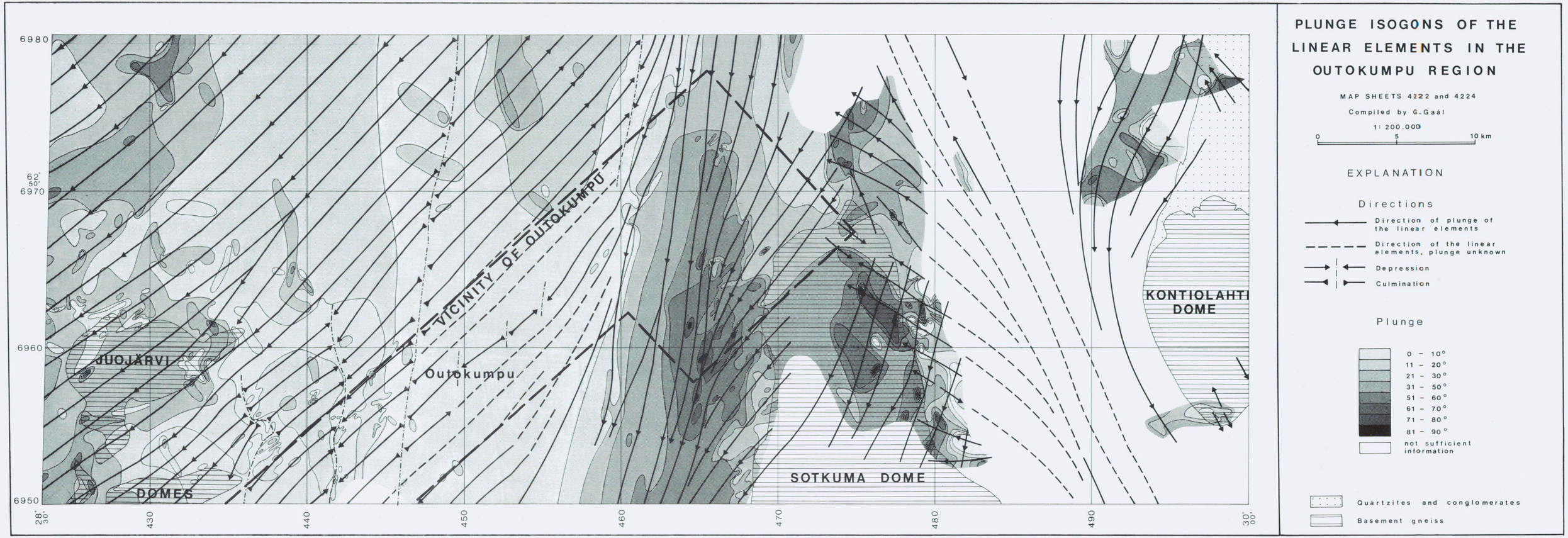


MAANMITTAUSHALLITUKSEN KARTTAPAINO HELSINKI 1975



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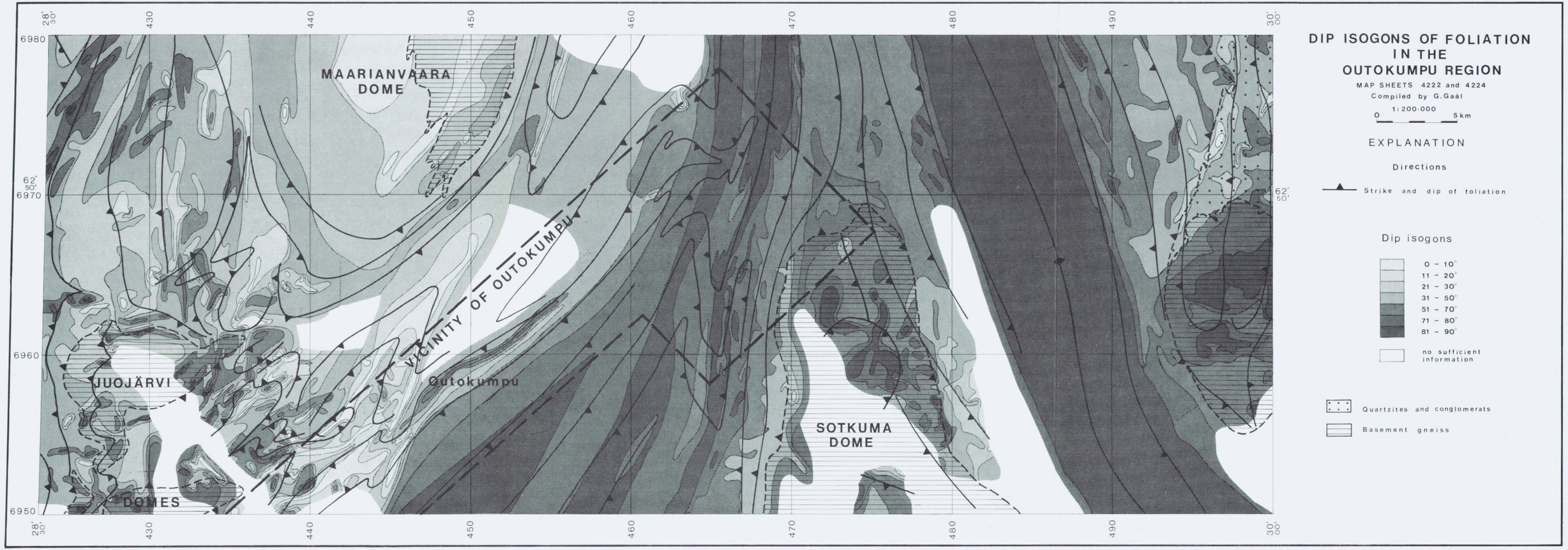

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MORPHOLOGY OF THE FOOT WALL AND THE HANGING WALL OF THE OUTOKUMPU ORE BODY

By T. Koistinen

1:10000

0 100 200 300 400 500m

HANGING WALL

LIETUKKA ORE BODY

KUMPU ORE BODY

KAASILA ORE BODY

Poikanen ore body

X=97.00

Y=185.00

89 88 X5 76 74 68+20

FOOT WALL

Poikanen ore body

X=97.00

Y=185.00

89 88 X5 76 74 68+20

Y=1500 X=550

Y=2000 X=500

Cross sections illustrating the deformation of the Lietukka ore body (Fig. 34)

Explanation

N

S

Outline and outcrop of the ore body

Contour lines at 5m intervals

Trace of I_1

Culmination

Depression

Moderate buckling (F_2), gentle slope

Strong buckling (F_2), steep slope

Thrust fault, dip direction

Strike slip fault