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The Karelian formations and their depositional basement in the Kiihtelysvaara — Värtsilä area, East Finland

by L. J. Pekkarinen

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THE KARELIAN FORMATIONS AND THEIR DEPOSITIONAL BASEMENT IN THE KIIHTELYSVAARA— VÄRTSILÄ AREA, EAST FINLAND

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L. J. PEKKARINEN

with 69 figures and 19 tables in the text and 2 appendices

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The study describes the general sequence of the Karelides in the Kiihtelysvaara—Värtsilä area, and the petrology, mineralogy and geochemistry of each unit. The aim is an accurate and generally applicable history of the evolution of the Karelian formations.

In the study area Middle Precambrian supracrustal rocks of the Karelian Sequence were deposited unconformably on an Early Precambrian Basement presently 2 600—2 800 Ma old. Furthermore, early sedimentary sequences appear to be cratonic graben or half-graben-type deposits (Prejatulian) followed by cratonic shallow-water and partly marine deposits (Jatulian). About 2 000—2 100 Ma ago they were all extensively intruded two or three times by diabase. Prejatulian (Sariolan) and Jatulian deposition were both followed by periods of weathering and erosion.

A second major period of supracrustal rock formation occurred in the Kiihtelysvaara—Värtsilä area beginning with miogeosynclinal flysch-type sediments and related rocks (Kalevian). In the Kiihtelysvaara-Värtsilä area this probably began about 2 000 Ma ago, but the onset of Kalevian sedimentation is still not well established.

Sedimentation and volcanism in the area were followed by deformation and metamorphism during the Svecokarelidic orogeny approximately 1 800—1 900 Ma ago.

Key words: stratigraphy, supracrustal rocks, basement, sedimentation, unconformity, Precambrian, eastern Finland.

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INTRODUCTION

Early in the 1960s the Geological Survey of Finland started to remap the Karelidic schist belt in the Finnish Baltic Shield. The mapping was concentrated on areas known to play a major part in clarifying the geology of the Karelidic schists. One of these key areas was the schist area in North Karelia. Later, in 1973—1976, explorational surveys carried out by the Exploration Department of the Geological Survey showed that the stratigraphy and structure were more complicated than had been anticipated from the mapping data. The problem is in fact so involved that in the present context only the formations in the contact zone between the Prekarelidic basement complex and the Karelidic schist are treated in detail.

Location and physiography of the area

The Kiihtelysvaara—Värtsilä area is situated in North Karelia, eastern Finland (Fig. 1). It is about 60 km long and 7 to 15 km wide (Fig. 2).

The terrain in the western and southwestern parts of the area is gently sloping and characterized by swamps, moors and eskers. The elevation varies from 80 to 110 m. Eastwards, the terrain becomes more hilly, the differences in elevation being from 40 to 80 metres or even more. Starting from the village of Mönni in the northwestern corner of the study area (Fig. 2), the boundary of the hilly country marked by narrow quartzite ridges runs via Kiihtelysvaara to Saario near the Soviet frontier. The hills rise to elevations of slightly less or slightly more than 200 metres above sea level.

In the western lowlands, vast tracts are without outcrops and are covered by Quaternary deposits. In the eastern parts, the flanks of the hills are well or moderately exposed. The contact zones of the rock units tend to be poorly exposed and covered by thick Quaternary deposits.

Historical outline and previous studies

In his studies on the Precambrian area of eastern Finland published back in 1874, Wiik came to the conclusion that the granite gneiss formation was the oldest formation (Fig. 1), and that it was overlain by fine-grained mica and chlorite schists. The youngest formations were the clastic quartz-rich sediments. Tigerstedt (1892) assumed that the conglomerates, quartzites, phyllites and limestones in Karelia formed a sequence deposited discordantly on

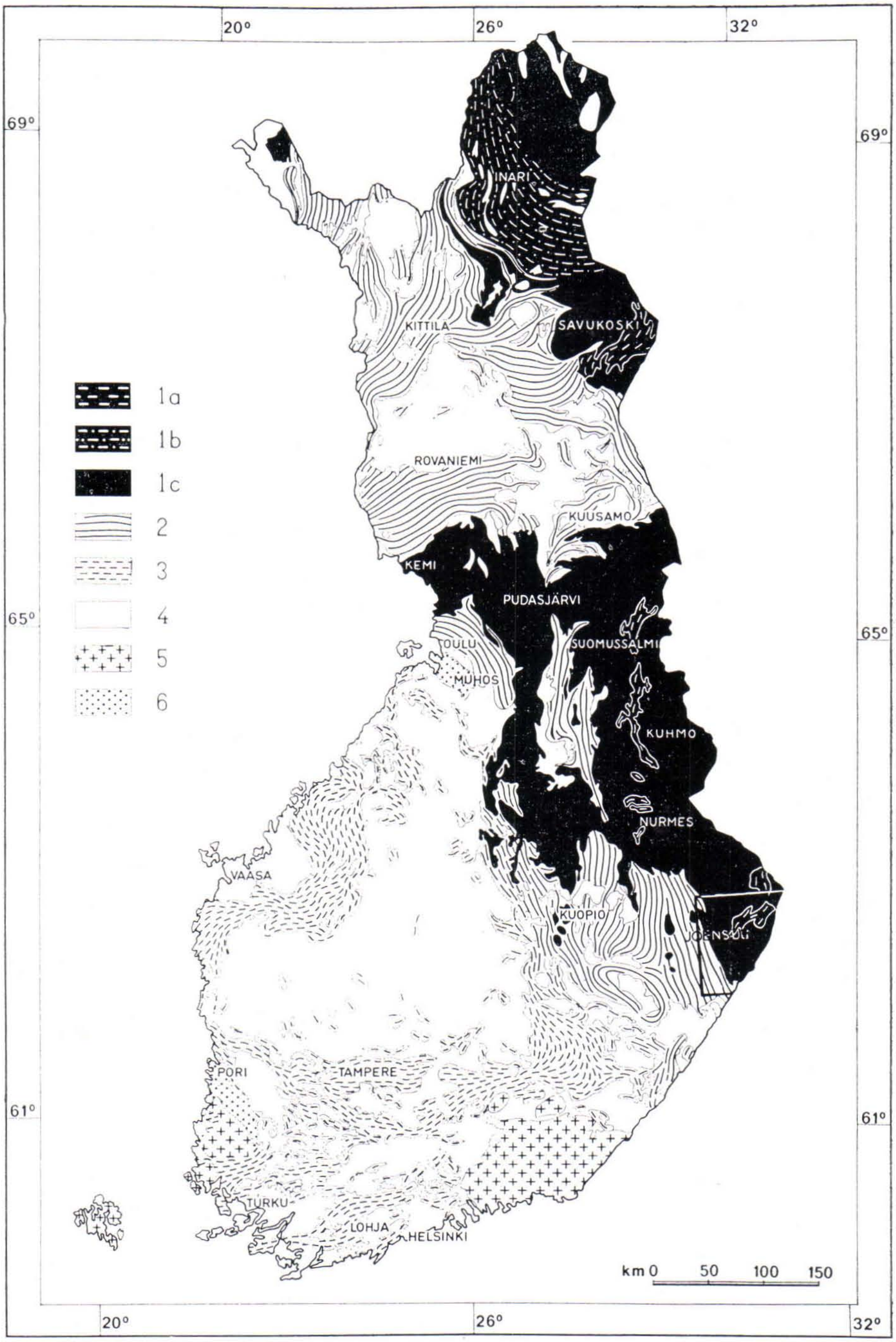

Fig. 1. Outline of the Precambrian of Finland, mainly according to Simonen (1971, p. 1411). *Presvecokarelidic* (*Presvecokarelian*): 1 a = schist and paragneiss; 1 b = granulite; 1 c = orthogneiss. *Svecokarelidic* (*Svecokarelian*): 2 = Karelidic schist belt; 3 = Svecofennidic schist belt; 4 = orogenic plutonic rocks. *Later Precambrian*, anorogenic: 5 = rapakivi intrusions; 6 = Jotnian sediments. The index map area in Fig. 2 is marked by a rectangle.

the sea floor on the granite gneiss basement. Sederholm (1893, 1897) allied the schist formations of Karelia with the Huronian series of North America. He called the quartzites *Jatulian* and the phyllite-mica schist formation *Ladogian*. He considered that the Ladogian schists were older that the Jatulian quartzites because the Ladogian schists were penetrated by granites and were more intensely metamorphosed than the Jatulian quartzites.

At the beginning of this century, however, phyllites and mica schists not penetrated by granites were found in the Karelidic schist belt. Frosterus (1902) and Ramsay (1902) called these schists *Kalevian* and held them to be younger than the Ladogian schists penetrated by granites. In their opinion, the Kalevian schists as well were older than the Jatulian. Later, during geological mapping in the Tohmajärvi area Frosterus and Wilkman (1920) found that also the Kalevian schists were penetrated by granites. Thus, the border between the Kalevian and Ladogian schist was still not clearly defined.

Eskola (1925) combined the three formations' the Ladogian, the Kalevian and the Jatulian, into the single *Karelian* formation, which underwent folding and metamorphism during one orogeny alone, the Karelidic orogeny. He named the Jatulian polymictic basal conglomerates Sariolan (Eskola 1919) conglomerates after a special type of polymictic basal conglomerate that he was the first to find near Tshobine in Soviet Karelia.

Wegmann (1928, 1929 a) maintained that the differences between the three series depended upon their position in the geosyncline: the Jatulian was composed of epicontinental sediments deposited on the foreland, the Kalevian of flysch sediments of the littoral sea, and the Ladogian of central geosynclinal sediments. The old granite gneiss area was a resistance area against which the Karelidic schists were pushed from the west.

In his classic works, Väyrynen (1928, 1933, 1954) gave the following standard section of the Karelian formations in eastern Finland:

Kalevian Group	Phyllites and mica schists
------- Erosion, Karelian metadiabases -------	
	Postjatulian folding
Jatulian Group	Marine Jatulian carbonaceous phyllites and dolomites Kainuan quartzites Sariolan arkosites and conglomerates
——— GREAT DISCORDANCE ———	
Granite Gneiss Basement	

According to Väyrynen, the Jatulian was older than the Kalevian. Although of the opinion that the Kalevian phyllites at Tohmajärvi were analogous to the Ladogian staurolite phyllites on the Soviet side of the border, Väyrynen (1954), like Hackman (1933), maintained that there was no distinct boundary between the Kalevian and Ladogian schists. The stratigraphic schemes of Väyrynen (1933, 1954) have since been revised by several authors (Härme 1949; A. Mikkola 1949; Nykänen 1968, 1971 b, 1971 c; Piirainen 1968, 1975; Silvennoinen 1972; Laajoki 1973, 1975 a, 1975 b; Mäkelä 1976; Laajoki and Saikkonen 1977).

The Karelidic schist belt was long thought to be younger than the more strongly metamorphosed Svecofennidic schist belt; after long deliberations, however, T. Mikkola (1953 and 1959), Metzger (1959) and Simonen (1960 a and b) suggested that both belong to the same orogenic cycle. This conclusion was confirmed by radiometric datings made by Kouvo (1958), and by Gerling and Polkanov (1958). Simonen (1960a and b) and T. Mikkola (1961) therefore proposed that Svecofennides and Karelides probably only represent different sedimentary facies of the same orogeny. T. Mikkola (1961) considered that the Jatulian was a shelf facies and the Kalevian a miogeosynclinal facies in the Svecofenno-Karelidic (or Fennidic) orogenic cycle. Later, Simonen (1971, p. 1419) gave the name *Svecokarelidic* to this orogeny.

Numerous interesting targets, e.g. the conglomerates, were described in the explanation to the Joensuu general geological map (1 : 400 000) by Frosterus and Wilkman (1920) and to a certain extent in the explanation to the Savonlinna geological map by Hackman (1933). A description of the geology of the area is given by Nykänen in his explanation to the 1 : 100 000 geological map sheets of Tohmajärvi (Nykänen 1968) and Kiihtelysvaara (Nykänen 1971b), as well as in his paper dealing with the Karelides (Nykänen 1971c). T. Mikkola (1955) and Ojakangas (1965) have studied the sedimentation of the Jatulian quartzites. The rocks of the area have also been discussed in two unpublished theses: one by Carlson (1967), who described the stratigraphy of the Jatulian formation in Southwest Kiihtelysvaara, and one by Kallio (1976), who studied the metadiabases at Kiihtelysvaara. Some explorational findings have been recorded in a few unpublished working reports (Hyvärinen and Siikarla 1971; Pekkarinen 1974; Räisänen 1975).

Outline of the study

To present a meaningful general picture, this study describes first the general sequence of the Karelides in the study area, then the petrology, mineralogy and geochemistry of each unit in turn. The aim is an accurate and generally applicable history of the evolution of the Karelian formations.

GENERAL GEOLOGICAL SETTING

The Precambrian metamorphic rocks in Finland are classified into three main groups: the oldest is the Early Precambrian (Archean) Presvecokarelidic basement complex in eastern Finland (Fig. 1). It is largely composed of granitoidic orthogneisses that include some zones of paragneisses. Westwards the complex is bordered by the Karelidic schist belt. The Karelides are an ancient mountain range that rose during the Middle Precambrian (Proterozoic) era. [1]) Today only their deeply eroded roots are visible. These constitute the Karelidic schist belt, which can be traced from Lake Ladoga to Lapland. The Karelidic schist belt is characterized by the sequence: basal conglomerate, arkose, quartzite, dolomite, phyllite and mica schist. The corresponding stratigraphic position in southern and western Finland is occupied by the pelitic and psammitic schists of the Svecofennidic belt.

The Kiihtelysvaara—Värtsilä area is that part of the Karelidic schist belt (Figs. 1 and 2) known as the schist zone of North Karelia. To define it more accurately, the study area is located on the eastern margin of the Höytiäinen syncline basin (Väyrynen 1954), where it constitutes the central part of the contact zone between the Presvecokarelidic basement complex and the Karelian formations.

[1]) In the present paper the boundary between the Early Precambrian and Middle Precambrian is put at 2 600 Ma in accordance with the Geological Time Table compiled by F.W.B. Van Eysinga/3rd Edition, Elsevier, Amsterdam, 1975.

The Presvecokarelidic Basement Complex

The Presvecokarelidic Basement Complex, called here briefly the Prekarelidic basement, represents an ancient craton that extends eastwards far beyond the Finnish border. Aeromagnetic maps (Fig. 3) show anomaly zones that stand out clearly from their background and are due to metavolcanics and metasediments, and the rocks of iron formation associated with both of them. The magnetically comparatively steady areas between the anomaly zones are composed of quartz- and granodiorites and granites.

The metavolcanics are predominantly amphibolites derived from lavas and tuffs, and amphibole and chlorite schists, although silicic volcanics, agglomerates and ultramafic rocks are encountered as well (Nykänen 1968, 1971b; Lavikainen 1973, 1975). The metasediments are largely mica schists, mica gneisses, and quartz-feldspar schists. Some of the latter, however, may be volcanogenic in origin. Black schist occurs as intercalations a few metres thick in the volcanics and metasediments.

In the western part of the area (Fig. 2) amphibolites and amphibole schists dominate mica schists and mica gneisses, but in the eastern part the converse is true.

According to Väyrynen (1954) and Nykänen (1971b), the amphibolites and amphibole schists are the oldest rocks in the area even though their depositional base has not been encountered. The plutonic rocks are younger still; on the basis of crosscutting relations they can be classified from oldest to youngest as gabbro and diorite — quartz- and granodiorite — oligoclase granite (leucogranodiorite) — potassium granite (Nykänen 1971b). The zircon ages of these rocks (Wetherill *et al.* 1962; Kouvo and Tilton 1966; Fig. 26) indicate that the rocks were emplaced about 2 600 to 2 800 Ma ago.

Karelidic rocks

The Karelian formations of the Karelian Sequence [2]) of the Karelidic schist belt of North Karelia are generally divided into two stratigraphic groups: the Jatulian and the Kalevian, which represent two different sedimentation facies (Wegmann 1928, 1929a; Väyrynen 1954; T. Mikkola 1953, 1959, 1961; Nykänen 1968, 1971b). Stratigraphically the lowest, the Jatulian metasediments have been considered as epicontinental sediments deposited on the margins of the continental blocks. In origin they are weathering gravel, arkose sands, quartz sands and carbonate- and carbon-bearing clay sediments. The Jatulian group also contains metabasaltic lavas, pyroclastites and hypabyssal rocks. Geosynclinal sediments of flysch type have deposited on the sediments of the Jatulian group (Wegmann 1928; Väyrynen 1954; Nykänen 1968, 1971b). The sediments of the Kalevian group were originally sandy clays that were deposited through the action of turbulent currents.

Lowest in the Karelian Sequence in the Kiihtelysvaara—Värtsilä area is a deposit of slightly sheared arkose, or rather arkosite (slightly metamorphosed), of variable thickness and associated with some local conglomerates. This arkosite conglomerate formation rests on the Prekarelidic basement with an angular unconformity in between (Fig. 4). In some places underneath it relics of weathering gravel or

[2]) Terminology: the Svecokarelidic orogeny; the Karelides are mountain range that arose during the orogeny. All that now remains of that range is the Karelidic schist belt. The Karelian Sequence consists of sedimentary-volcanogenic rocks deposited on the Prekarelidic basement.

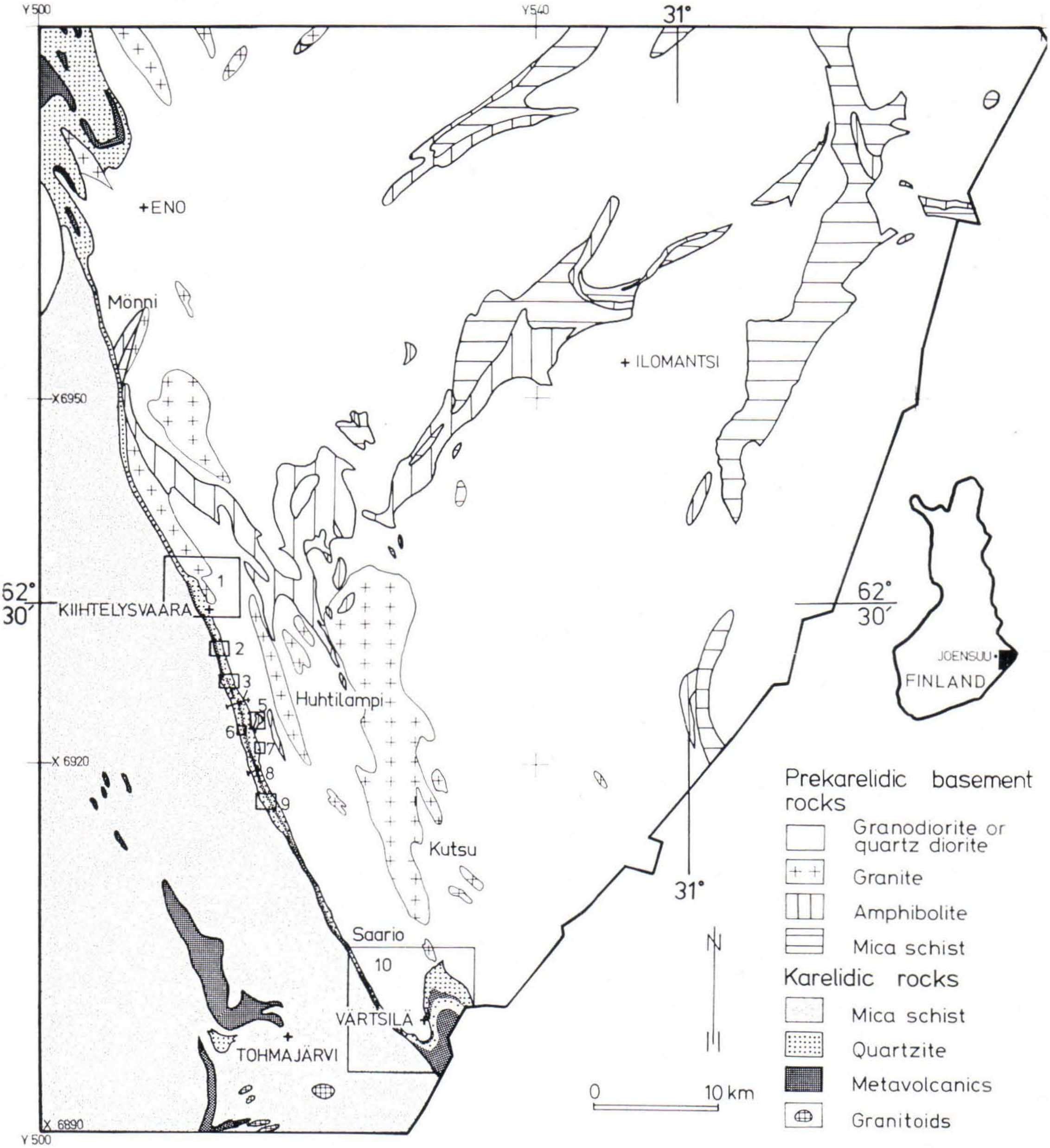

Fig. 2. Lithology of southern North Karelia modified from the general geological map of Finland on a scale of 1: 400 000 (Frosterus and Wilkman 1917, sheet D 3) and from geological maps on a scale 1: 100 000 (Nykänen 1967 and 1971 a; Lavikainen 1973 and 1975). Location of the areas investigated and mapped in detail: 1 = Karsikkojärvi; 2 = Kortevaara; 3 = Hyypiä; 4 = Viistola; 5 = Särkilampi; 6 = Kalkunmäki; 7 = Viesimo; 8 = Valkeavaara; 9 = Haluksenlammit; 10 = Värtsilä.

weathering crust of the ancient continent have been preserved; in this case the contact is gradual. The Arkosite formation is also known as the Sariolan formation (Eskola 1963; Nykänen 1968, 1971c) or the Sariolan facies (Väyrynen 1933, 1954).

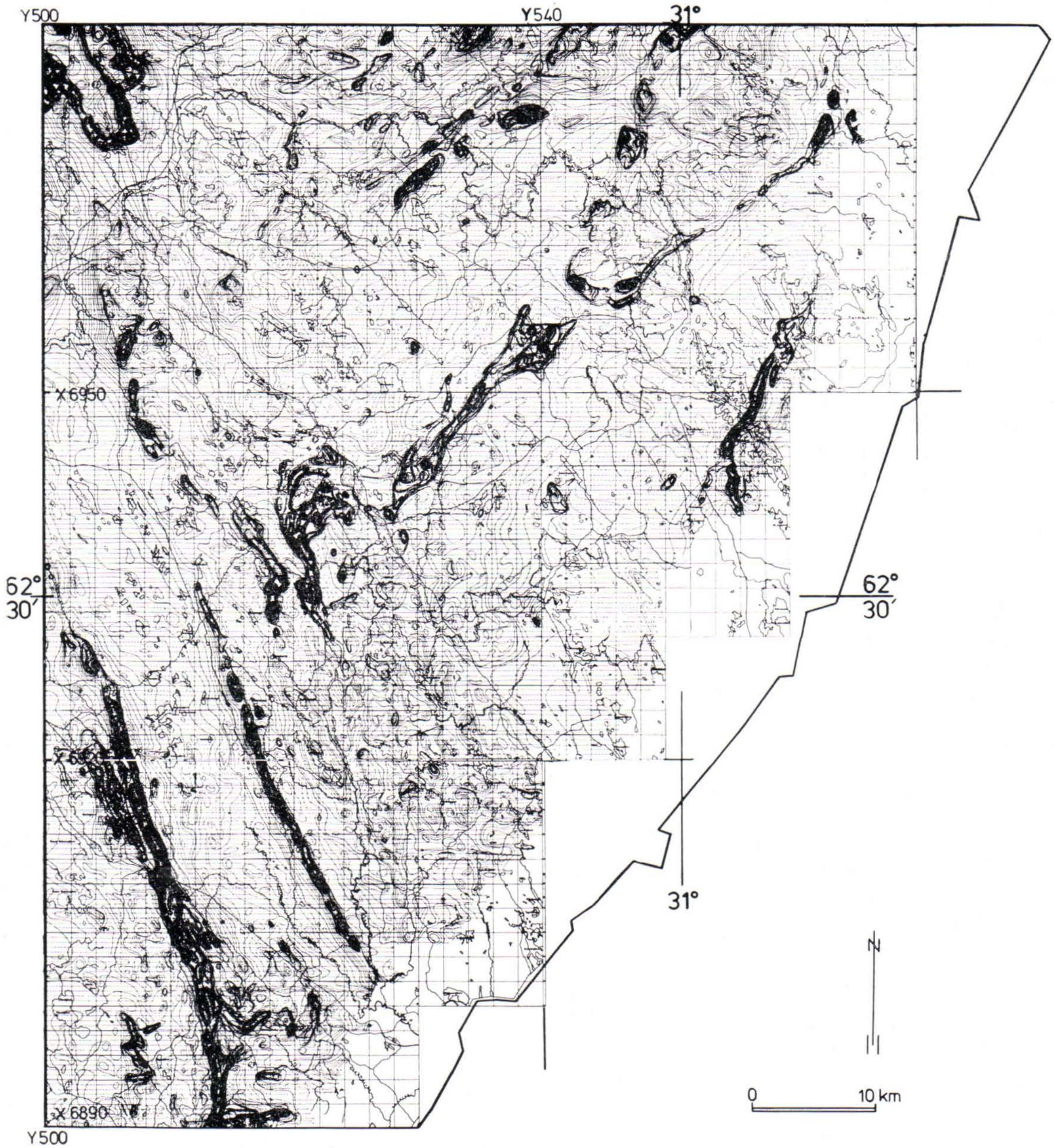

Fig. 3. Aeromagnetic anomaly map of southern North Karelia (same area as in Fig. 2). Geophysics Department, Geological Survey of Finland.

Indications abound suggesting that the quartzite beds overlying the arkosite formations were formed from the latter by the action of prolonged weathering and reworking. These Kainuan quartzites are highly resistant to weathering and constitute the main, readily traceable horizon (Väyrynen 1933, 1954; Nykänen 1971b). In the study area, however, the

Fig. 4. Unconformity of breccia-conglomerate over folded Prekarelidic biotite-chlorite schist. Length of tag 12 cm. Särkilampi, Kiihtelysvaara. Photo by E. Halme.

quartzites do not form a continuous sequence; they are composed of two discrete formations separated by the 50 to 80 m thick Volcanite Formation. The thickest of the two formations (150 to 550 m) is the Lower quartzite formation, which is mainly composed of orthoquartzite with some sericite quartzite layers. The composition of the Upper quartzite formation is more diverse and its thickness is only half that of the lower formation.

Next in sequence to the quartzites is a series of dolomites, carbonaceous slates and volcanites of variable thickness. This Marine Jatulian association (Väyrynen 1933, 1954) is more common in the study area than is implied in some geological maps, e.g. those compiled by Nykänen (1967, 1971a). As a rule the Marine Jatulian begins with clayey dolomite followed by carbonaceous slates alternating with dolomite layers. Volcanic rocks also occur. In larger deposits, however (Viistola section and Haluksenlammit), the dolomite—carbonaceous slate—volcanite suite is composed of two parts separated by a rather thin hematite rock—quartzite layer (the Hematite rock—Quartzite Formation): the lower Dolomite—Volcanite Formation and the upper Dolomite—Carbonaceous slate—Volcanite Formation.

Between the Jatulian and Kalevian formations there is a fairly continuous conglomerate bed that represents an unconformity between these formations. The conglomerate bed contains as pebbles and cobbles various Jatulian rocks including volcanics. The conglomerate is overlain by a dark graded-bedded quartzite (the Turbidite conglomerate—quartzite Formation), which grades upwards into mica schist (the Mica schist Formation).

Crosscutting the Prekarelidic basement and the formations overlain by Kalevian formations are a number of mafic, diabase-like rocks that were folded later together with the wall rocks and were metamorphosed during the Svecokarelidic orogeny; they are therefore often referred to as metadiabases. They, together with the mafic lavas and tuffs, obviously belong to the same intrusion series; the metadiabases represent the magma consolidated in the eruption channel, and the metalavas the rocks discharged on the surface. It would seem, however, that there was more than one eruption during the Jatulian stage. In the study area the eruptions started about 2 100 Ma ago; it has not been possible to date the later eruptions (Sakko, personal communication). It is noteworthy that in the upper Jatulian beds, i.e. the carbonaceous slates—phyllites, the emplacement was partly conformable. In the thicker sills the rocks are coarse grained and gabbroic in structure.

No granitoids have been encountered in association with the Karelides in the contact zone of the Karelian formations and the Prekarelidic basement. Farther south, however, in the Tohmajärvi and Kitee areas, also these rocks are present; the best known are the lateorogenic (age group 1 800—1 850 Ma, Annual Report of the Geological Survey 1974) Petrovaara trondhjemite and the Kitee granite (Nykänen 1968, 1975; Kouvo 1976, p. 9).

Tectonics

In the Prekarelidic basement complex the dominant direction of the structures varies between NE and NNE (Fig. 2), although fold structures are frequently encountered whose direction fluctuates between NW and NNW (Nykänen 1971b). The fold structures in the Prekarelidic basement are Presvecokarelidic (2 600—2 800 Ma) in origin. The cataclastic structure so common in the rocks of the Prekarelidic basement is presumably due to movements that took place during the Svecokarelidic orogeny. These movements were not the cause of plastic folding in the basement but rather breaking and blocking. The basement seems to have been broken during the movements into blocks that trend mainly NW or NNW; further, the same fracture, shear and fault zones seem to have been active several times. The older fracture and shear zones have also acted as channels along which the mafic magma, e.g. diabases, intruded. The NNW trend of Svecokarelidic schistosity is locally recognizable as transversal schistosity. Even today, the ancient fracture and shear zones are not without influence on the topography.

In the Pyhäselkä and Tohmajärvi areas the Karelidic schist zone of North Karelia trends NNW roughly concordantly with the contact of the Prekarelidic basement. Separated by an anticlinal ridge, the Karelidic schist zone is divided into two extensive synclinal basins largely occupied by Kalevian phyllites and mica schists: the Höytiäinen basin in the east and the Pyhäselkä basin in the west (Väyrynen 1954). The anticlinorium ridge extends from Hammaslahti via Onkamo to Tohmajärvi (Nykänen 1968, 1971b). It is characterized by volcanites, black slates, dolomites, conglomerates, and impure quartzites and arkosites whose exact location in the sequence is, however, not known for sure (Väyrynen 1954; Nykänen 1968; Pekkarinen 1974). At the eastern margin of the Höytiäinen synclinorium the Prekarelidic basement is bordered by a zone, about one kilometre wide, of Jatulian rocks that dip about 60° west and rest on the Prekarelidic basement (Fig. 2). In the Värtsilä area they bend slightly eastwards and near Sääperijärvi they form a small syncline that deepens southeastwards (Fig. 21).

In the Kiihtelysvaara—Tohmajärvi area the Karelidic schists of the Höytiäinen basin occur in a large axial depression in which, according to Wegmann (1928) and Nykänen (1971b and c), the regional fold axis plunges very gently southeastwards. The Karelian sequence was folded during the Svecokarelidic orogeny, about 1 800—1950 Ma ago, against the Prekarelidic craton trough, a compression that acted roughly from west to east. Thus, the axial planes of the folds generated dip at varying angles westwards (Väyrynen 1954; Nykänen 1971b). This regional macrotectonic folding seems to have been gently sloping, although more intense minor folding also occurs. Especially the mica schists and phyllites exhibit intense transverse cleavage, which trends roughly NNW (Fig. 20). Like the axial planes of the folds, the cleavage dips steeply westwards.

South of Tohmajärvi the rocks later underwent transversal deformation, and thus, in places, the trend of schistosity has turned towards E—W (Nykänen 1968). In the mica schists near the contact with the Jatulian formations this transversal deformation has given rise to kink-band structure and minor folding.

Metamorphism

The Prekarelidic basement and the Karelidic rocks in North Karelia were metamorphosed during the Svecokarelidic orogeny 1 800—1 950 Ma ago. The way the formations deformed depended on their tectonic position in relation to folding.

In the eastern part, ie. in the Prekarelidic basement, metamorphism seems to have been comparatively mild. Westwards, however, roughly from the contact of the Prekarelidic basement, the degree of metamorphism clearly increases. On the basis of the mineral assemblages of the Jatulian sediments and volcanics, it has been concluded that they were metamorphosed under the conditions of greenschist facies. The Jatulian conglomerates, sandstones, carbonate rocks and carbonaceous shales were metamorphosed into metaconglomerates, arkosites, quartzites, dolomites and carbonaceous slates. The volcanites and diabases associated with the Jatulian sedimentary rocks were metamorphosed into metavolcanites and metadiabases. Westwards, roughly at the contact between the Jatulian and Kalevian formations, the conditions grade into those of amphibolite facies. The Kalevian greywacky sandy clays metamorphosed into mica schists that frequently contain staurolite, andalusite and almandine as porphyroblasts. The boundary between the metamorphic facies, however, cuts the boundary between the sedimentary facies at a low angle owing to the fact that the degree of metamorphism increases slightly southwards as well as westwards.

Nykänen (1971c) has suggested that the Kitee Late-Karelidic granite gave rise to progressive contact metamorphism in the Karelidic schist belt at Tohmajärvi, and that this produced, among other things, the andalusite and staurolite porphyroblasts. Another possibility worth serious consideration is that during folding the Karelidic formations were forced deeper into the crust of the Earth, and that there they were metamorphosed under different pressure and temperature conditions (cf. Korsman 1977).

In the course of this study, however, it became evident that the facies relations in the area are so complicated that to solve them is a problem in itself. Therefore, problems pertinent to the facies have been treated only in so far as they have bearing on this investigation.

THE CONTACT ZONE BETWEEN THE KARELIAN FORMATIONS AND THE PRE-KARELIDIC BASEMENT

The origin of the Karelian sedimentary rocks and the associated volcanics can be studied in successive stratigraphic units whose depositional environment and temporal variation in volcanism are identified. In the Kiihtelysvaara—Värtsilä area there are several places for which the contact relations of the Karelian formations and the Prekarelidic basement are comparatively well established. These targets are described from north to south, excluding the adjacent detail targets at Viistola, Särkilampi and Kalkunmäki.

Geologic sections of subareas

The Karsikkojärvi area

About 2 to 3 km north of Kiihtelysvaara there is a small lake, Karsikkojärvi. In the present paper, the environment of this lake is called the Karsikkojärvi area (Figs. 2 and 5).

Depositional basement of the Karelian formations

The contact of the Prekarelidic basement with the Karelidic formations is exposed in a cut on the road to Ilomantsi. The contact is very sinuous, suggesting an uneven depositional basement. The rock of the basement is of grey gneissose grano-quartz diorite with lenses of pink granite. Close to the contact the rock is intensely cataclastic and slightly altered; the contact seam itself though is fairly sharp.

The Arkosite Formation

The Karelian sequence begins with the Arkosite Formation, whose lowest beds, the so-called basal arkosite, rest on the Prekarelidic basement with an angular unconformity in between. The basal arkosite is rather heterogeneous and slightly layered. Upwards in the sequence it grades into arkosite that is more distinctly layered. The grains in the lower arkosite are of quartz, plagioclase and potassium feldspar, interspersed here and there with larger pebbles of quartz and feldspar. The matrix is rich in sericite, and also contains quartz, biotite and carbonate.

The Lower quartzite Formation

The arkosites are succeeded by a greenish sericite quartzite that, higher in the sequence, grades into an almost white or slightly pinkish orthoquartzite. Together they constitute the Lower quartzite Formation.

The Volcanite Formation

In the southern part of the area the Lower quartzite Formation is overlain by a formation, 50 to 80 m thick, composed of lavas, tuffs and tuffites (Volcanite Formation); in the northern part they are, however, lacking. A sharp change is also visible on the aeromagnetic map (Fig. 3). Anomalies due to volcanites come to an end near Karsikkojärvi, which indicates that volcanites may not occur farther north. The same feature is also evident on the map of the bedrock of Kiihtelysvaara compiled by Nykänen (1971a). It can, of course, be assumed that farther north there was a complete absence of volcanic activity; nevertheless, metadiabase dykes that cut the quartzites and which have been interpreted as discharge channels for

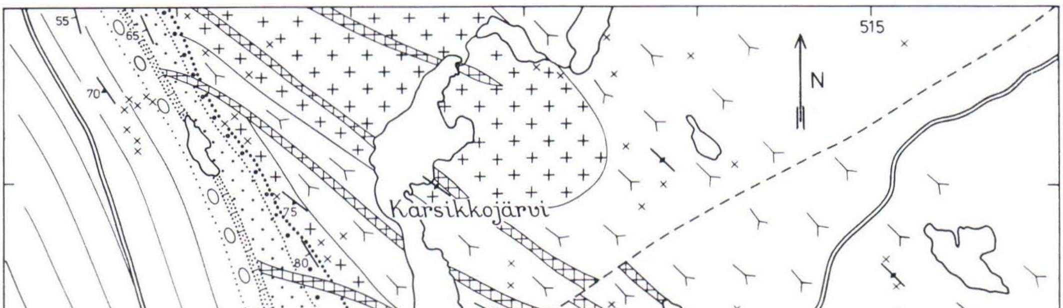

Fig. 5. Geological map of the Karsikkojärvi area. *Kalevian:* 1 = mica schist; 2 = turbidite conglomerate and quartzite. *Jatulian:* 3 = carbonate-bearing quartzite; 4 = metavolcanite; 5 = metadiabase; 6 = quartzite; 7 = quartz conglomerate interbeds. *Prejatulian:* 8 = arkosite; 9 = arkosic conglomerate. *Prekarelidic:* 10 = granite; 11 = granodiorite and quartzdiorite. Other symbols: 12 = bedding, foliation and lineation; 13 = major unconformity; 14 = unconformity; 15 = fracture or fault; 16 = outcrop.

effusive volcanic matter occur in both the northern and southern parts of the area. Hence, it is more likely that there were volcanic eruptions in the northern part of the area but that the volcanic beds were eroded soon after eruption. A further indication of intense erosion is the thickness of the Lower quartzite Formation in the northern part: it is only half of that in the south.

The Upper quartzite Formation

The volcanites are succeeded by a bed of calcareous quartzite (Upper quartzite Formation). However, owing to the sparcity of outcrops, information on this rock is not very accurate.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The Kalevian mica schists with staurolite porphyroblasts begin very near the quartzite outcrops of the Upper quartzite Formation. The numerous thin quartzite interlayers, which increase in frequency downwards in the mica schist, suggest, however, that there is more quartzite under the mica schists (Turbidite conglomerate—quartzite Formation).

Sequence of the Karelian formations

From top to bottom, the sequence in the Karsikkojärvi area is as follows (estimated thickness in brackets):

Southern part	*Northern part*
Mica schist Formation; staurolite-bearing mica schist, in its lower part thin quartzite interlayers	
Turbidite conglomerate—quartzite Formation?	
---------- unconformity ----------	
Upper quartzite Formation (0—30 m); carbonate-bearing quartzite	Upper quartzite Formation (lacking or very thin)
---------- unconformity ----------	
Volcanite Formation (70 m);	(lacking)
Lower quartzite Formation (400 m); upper part ortho-quartzite, lower part sericite quartzite	Lower quartzite Formation (200 m)
---------- unconformity ----------	
Arkosite Formation (100—150 m); arkosite, basal arkosite and local conglomerate	Arkosite Formation (100 m); arkosite and basal arkosite
---------- major unconformity ----------	
Prekarelidic basement (gneissose grano- and quartz diorites and granites).	

The Kortevaara area

Kortevaaıa is a geologically interesting area located about two kilometres south of Kiihtelysvaara (Figs. 2 and 6 A—B). The origin of the chalcopyrite dissemination encountered in the Lower quartzite Formation in the area was established by detailed mapping, and magnetic and electromagnetic survey; the results were checked by diamond drilling.

The Arkosite Formation

The Karelian Sequence begins with the Arkosite Formation, which, separated by an angular unconformity, rests on the Prekarelidic gneissose grano-quartz diorite and has a mantle-like bed of basal arkosite. Upwards the basal arkosite grades into a more mature arkosite composed mainly of quartz, potassium feldspar and sericite. The upper contact of the arkosite is not exposed; it was, however, intersected by diamond drilling, which showed that, in the arkosite close to the contact, feldspar with its shape still recognisable was completely sericitized. Locally the rock is brecciated. The brecciated quartzite some 500 m southeast of the Kortevaara farm is possibly part of this zone.

The Lower quartzite Formation

The Lower quartzite Formation begins with greenish sericite quartzite that, separated by an unconformity, rests on arkosite. It contains fairly abundant quartz conglomerate interbeds, 2 to 20 cm thick, and local thin slate intercalations. Higher in the sequence the abundance of sericite decreases and the rock grades into a pinkish orthoquartzite. Although only faintly visible, cross bedding is common in both types of quartzite.

A small Cu mineralization was encountered in the Lower quartzite near the Kortevaara dyke (Fig. 6A). Chalcopyrite has impregnated the porous quartzite layers and the hair fractures in it. Nevertheless, the Cu content is low (0.1 to 0.2 % Cu) throughout the zone; only in a few nests is it somewhat higher (about 0.5 % Cu). Large pyrite crystals occur here and there beyond the chalcopyrite-bearing zone. The sulphide mineralization seems to be genetically related to the metadiabase dyke.

The Volcanite Formation

The Lower quartzite Formation is overlain by the 80 m thick Volcanite Formation. Its lavas and tuffs are dark-green fine-grained rocks that mainly consist of plagioclase (albite to oligoclase) and chlorite. They also contain minor biotite, quartz, titanite and magnetite.

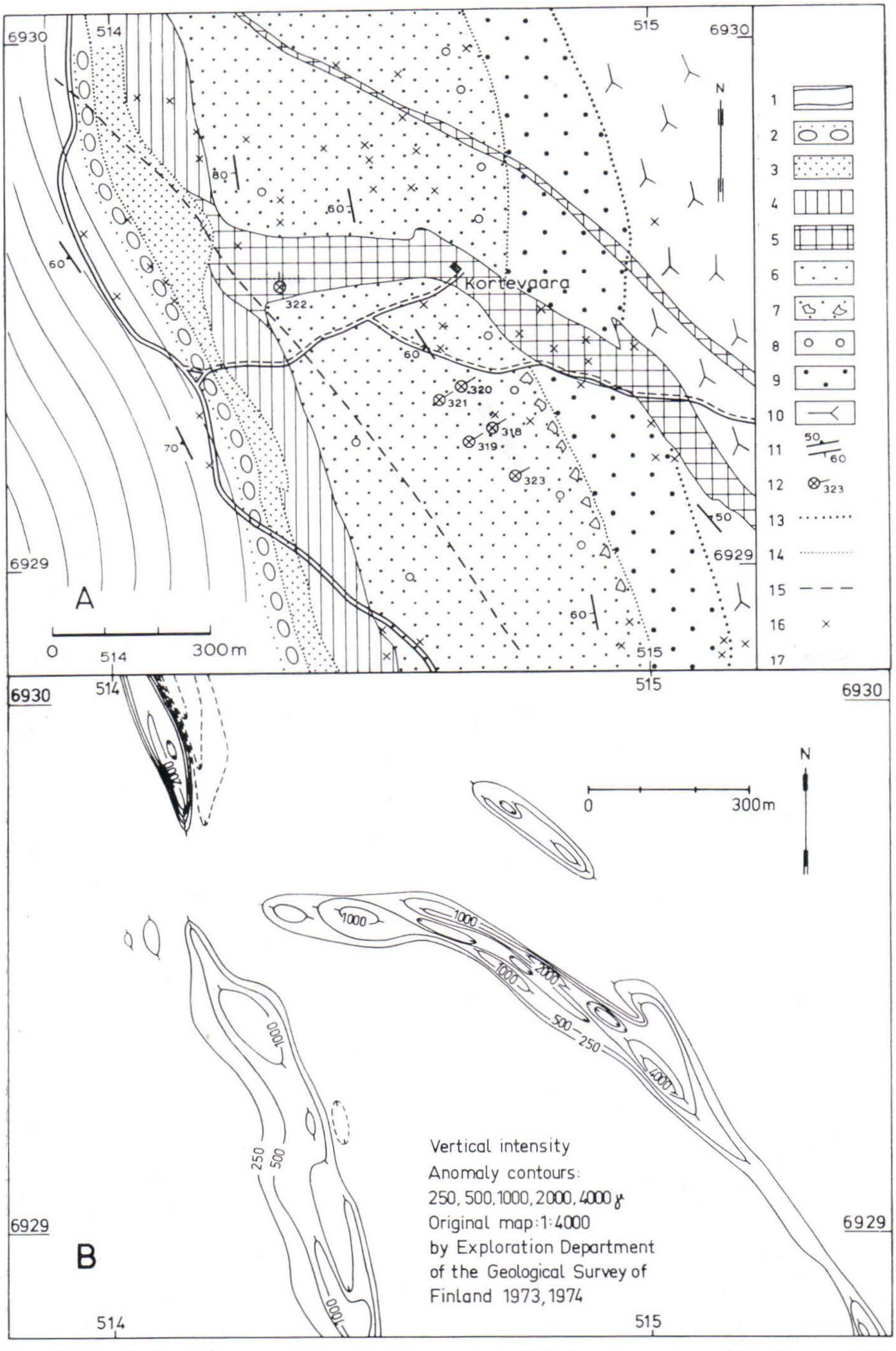

Fig. 6. A) Geological map of the Kortevaara area. *Kalevian:* 1 = mica schist; 2 = turbidite conglomerate and quartzite. *Jatulian:* 3 = upper quartzite; 4 = metavolcanite; 5 = metadiabase; 6 = lower quartzite; 7 = brecciated quartzite; 8 = quartz conglomerate interbeds. *Prejatulian:* 9 = arkosite. *Prekarelidic:* 10 = granodiorite and quartz diorite. Other symbols: 11 = bedding and foliation; 12 = drill hole; 13 = major unconformity; 14 = unconformity; 15 = fracture; 16 = outcrop; 17 = chalcopyrite.

B) Magnetic vertical intensity map of the Kortevaara area.

Typically, magnetite is fairly homogeneously distributed in the rock, and thus the volcanite beds are readily detected by magnetic survey (cf. Figs. 6A and 6B).

The Upper quartzite Formation

The Volcanite Formation is succeeded by the Upper quartzite Formation, a comparatively thin bed of impure carbonate-bearing quartzite.

Metadiabase dykes

Numerous metadiabase dykes that cut both the Prekarelidic basement and the arkosite and quartzite beds have been encountered in the area. The most significant is the dyke, about 100 m wide, that occurs in the centre of the map area (Fig. 6A) and is known as the Kortevaara dyke. Field observations, geophysical survey and drilling data suggest that this dyke represents the ancient discharge channel for the effusives that erupted on the Lower quartzite Formation (Figs. 6A and 6B). A peculiar feature is that the dyke appears on the map as an arch that bends gently westwards, possibly because of bending and tilting in the arkosite and quartzite wall rocks during folding.

The metadiabase in the Kortevaara dyke is a dark, slightly greenish and predominantly blasto-ophitic, medium-grained rock. It is composed of cloudy plagioclase (oligoclase), and varying amounts of chlorite, hornblende, biotite, quartz, titanite and magnetite. The abundance of hornblende is highest at the eastern end of the dyke; westwards it alters completely into chlorite. The middle of the dyke is paler than the margins; it contains more plagioclase and quartz, and in places shows pyrite and chalcopyrite dissemination. On both sides of the pale middle portion of the dyke there is a »spotty» variant in which the spots are composed of biotite-magnetite aggregates and which may be the cause of the slight bifurcation in the magnetic anomaly pattern (Fig. 6B).

Some internal dykes of fine-grained metadiabase from 20 cm to 4 m wide have been found within and almost parallel to the Kortevaara dyke (Fig. 64). The rock contains some plagioclase (andesine) and hornblende phenocrysts (Fig. 67E). The groundmass is composed mainly of plagioclase, hornblende, biotite and epidote.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The contact of the Upper quartzite Formation with the Turbidite conglomerate—quartzite Formation is not exposed. However, on the basis of adjacent outcrops, it has been assumed that the lowest parts of the Turbidite conglomerate—quartzite Formation consist of a 15 m thick conglomerate bed. This Kortevaara or Kiihtelysvaara conglomerate (Frosterus and Wilkman 1920; Nykänen 1971b) has been interpreted as a Kalevian basal conglomerate (Fig. 7). The pebbles, cobbles and boulders in the conglomerate represent divers Karelidic rocks such as quartzite, volcanites, carbonaceous slates and quartz. Their diameter usually varies between 5 and 50 m and they are distinctly rounded (Fig. 7). Nevertheless, also found in the deposit are some metadiabase boulders whose diameter is almost 4 m. The matrix in the conglomerate is composed predominantly of quartz, micas and carbonate. The conglomerate is overlain by a 15 m thich bed of darkgrey quartzite with phyllite interbands which upwards passes gradually into staurolite-bearing mica schist.

The Sequence of the Karelian formations

From top to bottom, the sequence in the Kortevaara area (Fig. 6A) is as follows (estimated thickness in brackets):

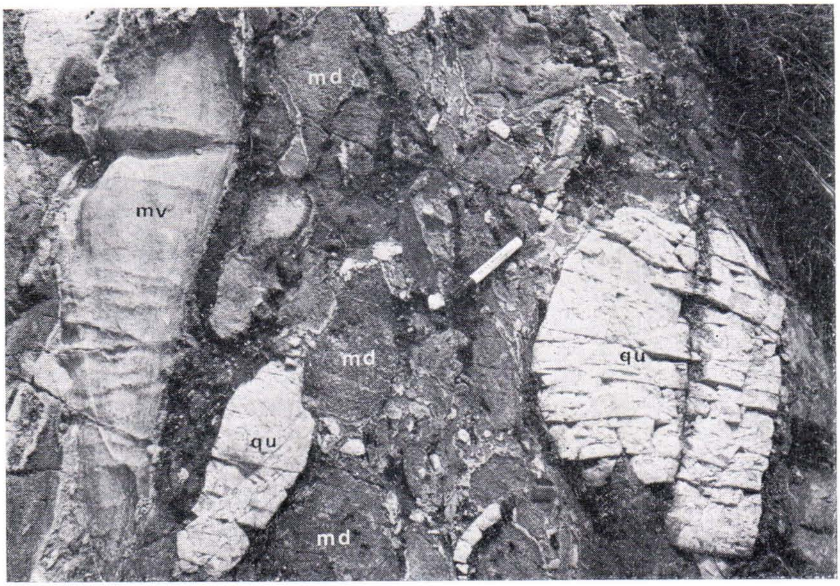

Fig. 7. Turbidite conglomerate; pebbles, cobbles and boulders mainly quartzite (qu), metadiabase (md) and metavolcanite (mv). Length of pen 14 cm. Kortevaara, Kiihtelysvaara. Photo by E. Räisänen.

Mica schist Formation; staurolite-bearing mica schist

Turbidite conglomerate—quartzite Formation; dark quartzite (15 m) underlain by conglomerate (15 m)

---------------------- unconformity ----------------------

Upper quartzite Formation (40 m); carbonaceous quartzite

---------------------- unconformity ----------------------

Volcanite Formation (80 m)

Lower quartzite Formation (370 m); upper part orthoquartzite, the lower sericite quartzite with thin quartz conglomerate and clay slate intercalations

---------------------- unconformity ----------------------

Arkosite Formation (100—150 m); arkosite and basal arkosite

-------------------- major unconformity --------------------

Prekarelidic basement (gneissose grano- and quartz diorite).

From the depositional basement to the contact with the Turbidite conglomerate—quartzite Formation, this sequence is about 640 m thick.

The Hyypiä area

A ridge of Jatulian quartzites passes SSE through the village of Kiihtelysvaara. Some 5 to 7 km southeast of the village church there are a few hills that rise above the environment; one of these is Hyypiänvaara (its highest point is 202 m above sea level), and the surrounding area is generally called Hyypiä (Figs. 2 and 8).

Depositional basement of the Karelian formations

The contact of the Prekarelidic basement with the Karelian formations is exposed in two localities. The rock in the basement is a grey, gneissose grano-quartz diorite with some small portions of oligoclase granite. The rock is cataclastic and slightly altered close to the contact; feldspar, for example, has turned into sericite.

The Arkosite Formation

The Karelian Sequence begins with the Arkosite Formation, which, separated by an angular unconformity, rests on the rocks of the Prekarelidic basement. The lowest member is coarse basal arkosite with associated small conglomerate lenses. In one outcrop the thickness of the arkositic conglomerate varies from 1 to 2 m. The basal conglomerate and basal arkosite grade upwards for a short distance

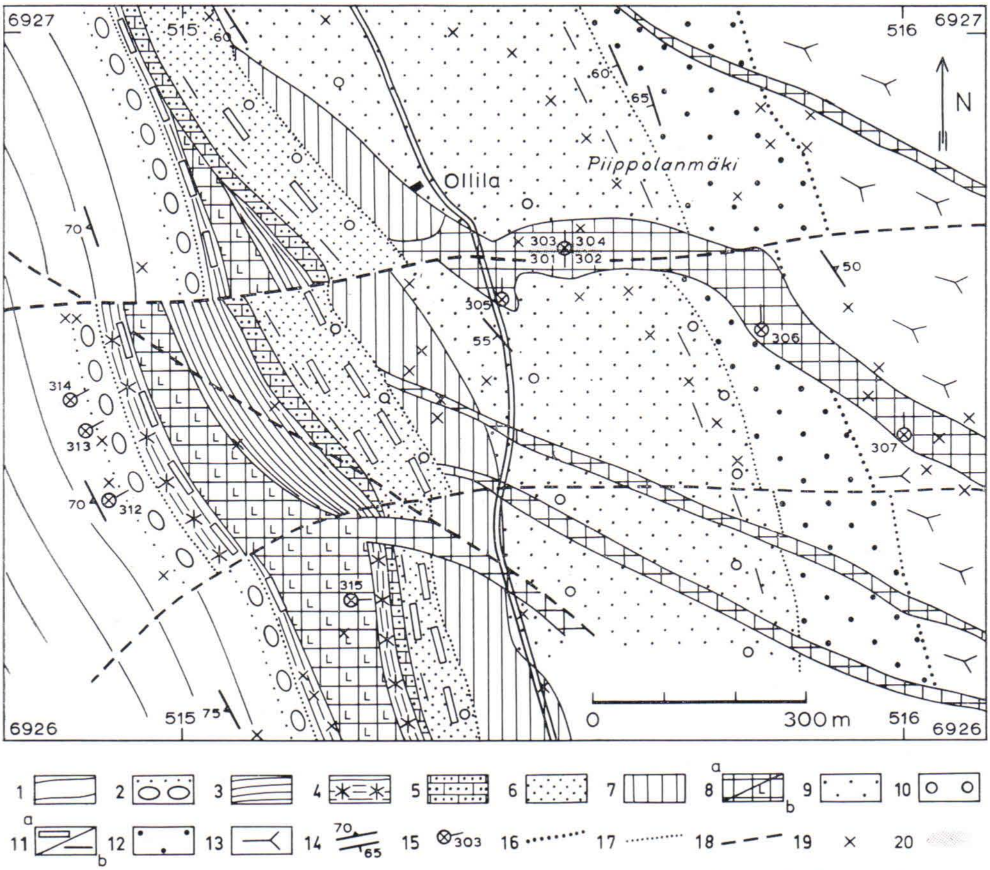

Fig. 8. Geological map of the Hyypiä area. *Kalevian:* 1 = mica schist; 2 = turbidite conglomerate and quartzite. *Jatulian:* 3 = carbonaceous slate; 4 = graphite-rich layers; 5 = quartz-bearing dolomite; 6 = upper quartzite; 7 = metavolcanite; 8 a = metadiabase; 8 b = gabbroic metadiabase; 9 = lower quartzite; 10 = quartz conglomerate interbeds; 11 a = volcanic intercalations; 11 b = pelitic intercalations. *Prejatulian:* 12 = arkosite. *Prekarelidic:* 13 = granodiorite and quartz diorite. Other symbols: 14 = bedding and foliation; 15 = drill hole; 16 = major unconformity; 17 = unconformity; 18 = fracture or fault; 19 = outcrop; 20 = chalcopyrite.

into a cross-bedded and more distinctly layered arkosite (Fig. 9).

In the upper portion of the Arkosite Formation drilling has revealed a 10 to 20 m thick alteration zone in which the feldspars have been replaced by sericite. The topmost layer is composed of a sericite-rich schist a few metres thick.

The Lower quartzite Formation

The Lower quartzite Formation begins with a greenish sericite quartzite that has fairly abundant thin quartz conglomerate interbeds and, here and there, some very thin clay slate intercalations. Upwards in the sequence the abundance of sericite decreases and the rock grades into an almost white or pinkish orthoquartzite. In the upper portion of the sericite quartzite there are locally several successive layers that exhibit ripple marks of various size (Fig. 46).

The Volcanite Formation

The Lower quartzite Formation is overlain by the 60 m thick Volcanite formation (Fig. 8). The volcanites include lava, fragmentary

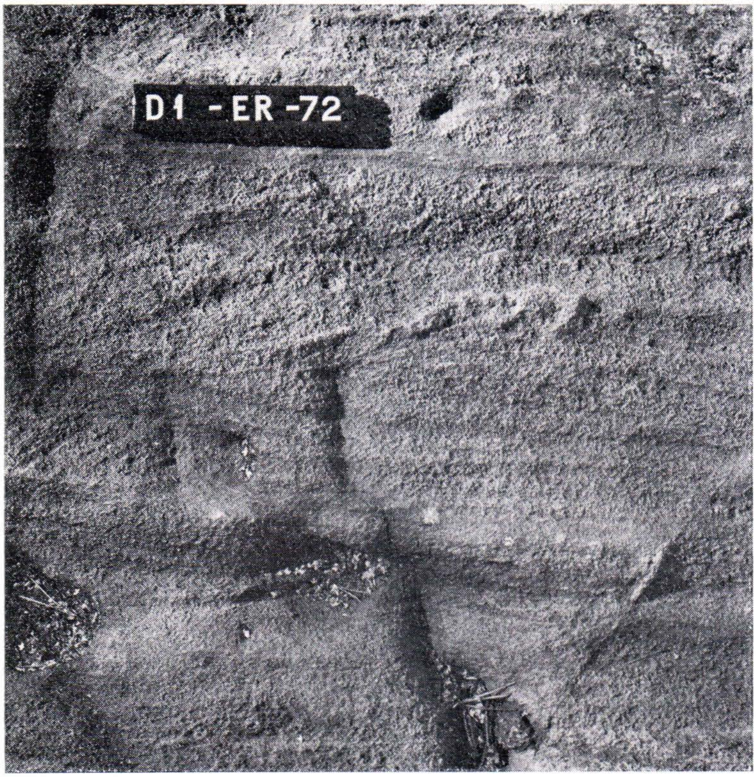

Fig. 9. Lower arkosite showing cross-bedding. Top of sequence upwards. Length of tag 12 cm. Hyypiä, Kiihtelysvaara. Photo by E. Halme.

lava (Fig. 63) and amygdaloid; highest in the sequence, tuffs and tuffites are also encountered. The lava often contains sparse plagioclase (albite-oligoclase) phenocrysts. The aphanitic groundmass is composed of plagioclase, chlorite, biotite, quartz, titanite and magnetite. Amygdules filled with quartz or calcite are also common (Fig. 66); the largest of them vary in diameter from 2 to 30 mm. The fragments in the fragmentary lava (Fig. 63) are composed of the same material as the adjacent lava except that the matrix also contains appreciable epidote. Fractures and cavities filled with quartz are also common. The tuffs and tuffites are heterogeneous rocks composed mainly of quartz, chlorite and biotite.

The Upper quartzite Formation

The Volcanite Formation is succeeded by the impure quartzites of the Upper quartzite Formation, which, owing to volcanic weathering products, chiefly chlorite and magnetite, are greyish in colour. They also contain some tuffitic interbeds, 1 to 15 m thick, indicating that subdued volcanic activity continued after the main phase.

The Dolomite—Carbonaceous slate—Volcanite Formation

The Upper quartzite Formation is overlain by alternate layers of dolomite and carbonaceous slate (Dolomite—Carbonaceous slate—Volcanite Formation). A sill about 100 m wide of coarse, gabbroic metadiabase has intruded between the uppermost beds. In the upper portion of the sill there are thin lava beds that contain hornblende.

The metadiabase dykes and sills

In the Hyypiä area the map is characterized by metadiabase dykes that, starting in the Prekarelidic basement, cut the arkosites and quartzites and extend as far as the Volcanite Formation (Fig. 8). There are, however, some dykes that are obviously younger; they reach higher levels in the sequence up to the dolomite and carbonaceous slate layers. Their zircon ages have, however, not yet been determined. The metadiabase dykes are generally 5 to 20 m wide, except for the Piippolanmäki or »Hyypiä» dyke (Nykänen 1971b, p. 40), which is nearly 100 m wide (Fig. 8). This dyke represents the channel for the lava flow that discharged onto the Lower quartzite Formation.

At the eastern end of the Piippolanmäki dyke, where the wall rock is Prekarelidic grano-quartz diorite, the predominant dark mineral in metadiabase is hornblende. Farther west, where the dyke cuts arkosites, the hornblende has partly altered into chlorite. Still farther west, at the location of the quartzite layers, the hornblende is almost completely chloritized. The dyke also shows differentiation: the margins are darker, containing plagioclase (oligoclase) and variable amounts of hornblende, chlorite and biotite, and some minor quartz, epidote, titanite, carbonate and magnetite. The middle part of the dyke is paler than the margins because of the higher plagioclase and quartz abundances. The Piippolanmäki metadiabase dyke also contains dark, fine-grained metadiabase that constitutes 1 to 5 cm wide internal dykes approximately parallel to the host dyke.

Scattered portions of chalcopyrite-bearing rock are encountered in the middle part of the Piippolanmäki metadiabase dyke. The chalcopyrite occurs mainly as fine-grained dissemination, but partly also as richer nests in association with quartz. The average copper content in the rock is, however, low, hardly exceeding 0.1 to 0.3 %. The mineralization is thus without economic significance.

West of Hyypiänvaara is the Hyypiä sill that intruded between the dolomite and carbonaceous slates. The discharge channels of the sill seem to have been so narrow that they can no longer be recognised for sure. Brecciated by volcanic material, however, quartzite that might be part of the contact of the ancient discharge channel has been encountered in the Upper quartzite Formation near Hyypiä. The sill is predominantly a coarse-grained gabbroic metadiabase that is composed mainly of green hornblende and plagioclase (oligoclase to andesine). It also contains biotite, pale amphibole, epidote, chlorite, titanite and leucoxene.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The contact of the Dolomite—Carbonaceous slate—Volcanite Formation with the Turbidite conglomerate—quartzite Formation was intersected by three drill holes (Fig. 8). It was noted that the latter begins with a bed of conglomerate about 4 m thick. The subrounded cobbles and pebbles of this conglomerate are Karelian rocks such as quartzite, volcanite, carbonaceous phyllite and vein quartz (Fig. 59D). The matrix is predominantly quartz and micas. The portions richest in micas display almandine porphyroblasts, which may be as much as 2 cm in diameter. The conglomerate is succeeded by a dark graded-bedded quartzite in which each phase includes a quartzitic lower and a phyllitic upper part. Upwards in the sequence the quartzitic portion rapidly becomes thinner as the phyllitic portion becomes thicker, and the quartzite passes into a staurolite-bearing mica schists.

The Sequence of the Karelian formations

From top to bottom, the sequence in Hyypiä (Fig. 8) is as follows (estimated thickness in brackets):

Mica Schist Formation; staurolite-bearing mica schist

Turbidite conglomerate—quartzite Formation; dark quartzite (50 m) underlain by polymictic conglomerate (4 m)

---------------------- unconformity ----------------------

Dolomite—Carbonaceous slate—Volcanite Formation (165 m)

— carbonaceous slate, thin dolomite and volcanite intercalations (10 m)
— metavolcanite (10 m)
— gabbroic metadiabase sill (80 m)
— carbonaceous slate, thin dolomite intercalations (70 m)
— dolomite, quartz-bearing (5 m)

Upper quartzite Formation (100 m); impure quartzite with a few volcanite and clay slate intercalations and thin conglomerate interbeds in the lowest part

---------------------- unconformity ----------------------

Volcanite Formation (70 m)

Lower quartzite Formation (300 m); upper part orthoquartzite with very thin quartz conglomerate interbeds in the uppermost part; lower part sericite quartzite with quartz conglomerate interbeds in the lowest part

---------------------- unconformity ----------------------

Arkosite Formation (120 m); arkosite, basal arkosite and local conglomerate

------------------- major unconformity -------------------

Prekarelidic basement (gneissose grano- and quartz diorite and oligoclase granite).

Measured from the depositional basement to the contact with the Turbidite conglomerate—quartzite Formation, this sequence is about 770 m thick.

The Särkilampi area

In this context the Särkilampi area (Figs. 2 and 10) comprises the environment of the lake of that name. This area has long been known for its thick conglomerate beds, particularly the Särkilampi (Pölkkylampi) conglomerate (Frosterus and Wilkman 1920; Carlson 1967; Nykänen 1971b). The interest of geologists was once more aroused in the Särkilampi area in the course of exploration activities put in motion by the Geological Survey after Cu-Mo-Au-bearing outcrops were found in the Prekarelidic basement. The area was submitted to detailed mapping and diamond drilling. The mineralized zone is located near the contact with the Karelian Sequence, and hence, attention was also paid to the contact relations.

Depositional basement of the Karelian formations

Surrounded and cut by Prekarelidic gneissose grano-quartz diorite and tourmaline granite, the area is composed of amphibolites, mica schists, leptites and small iron-formation lenses in amphibolite. The area is intensely tectonized. The strike and dip of schistosity and the trend of the fold axes vary considerably (Fig. 10). The gneissose grano-quartz diorite is fairly conformable to the schists; the tourmaline granite, however, cuts them.

The amphibolite is a fine-grained, schistose and tuffitic rock composed of hornblende, plagioclase, epidote, biotite and quartz. It also contains minor titanite, carbonate and tourmaline. In some places bands rich in magnetite are encountered. The mica schists contain biotite, chlorite, quartz and plagioclase. Occasionally there is abundant tourmaline, particularly around the veins of tourmaline granite. The rock also contains epidote, apatite and carbonate. The leptite is composed of potassium feldspar, quartz, plagioclase, sericite and biotite. The rock often exhibits large feldspar phenocrysts embedded in a fine-grained matrix; in structure it resembles volcanites.

Rocks of iron formation occur in the transitional zone between the amphibolite and mica schist as small lenses that are 1.5 m thick at the most and some tens of metres long (Fig. 10). Their content of total iron is low, about 20 % (Niiniskorpi 1975). The iron formation is quartz-predominant and exhibits quartz bands that vary in thickness from a few mm to several cm. In addition to magnetite, the magnetite-rich bands contain amphibole, biotite, epidote, plagioclase, chlorite, apatite and carbonate. The quartz bands are generally fairly pure, but they may have the above minerals as accessories. Locally the amphibolite contains abundant magnetite, usually as bands from 1 to 15 cm thick.

Skarn rock with a low-grade sulphide mineralization (Fig. 10) is encountered in the contact

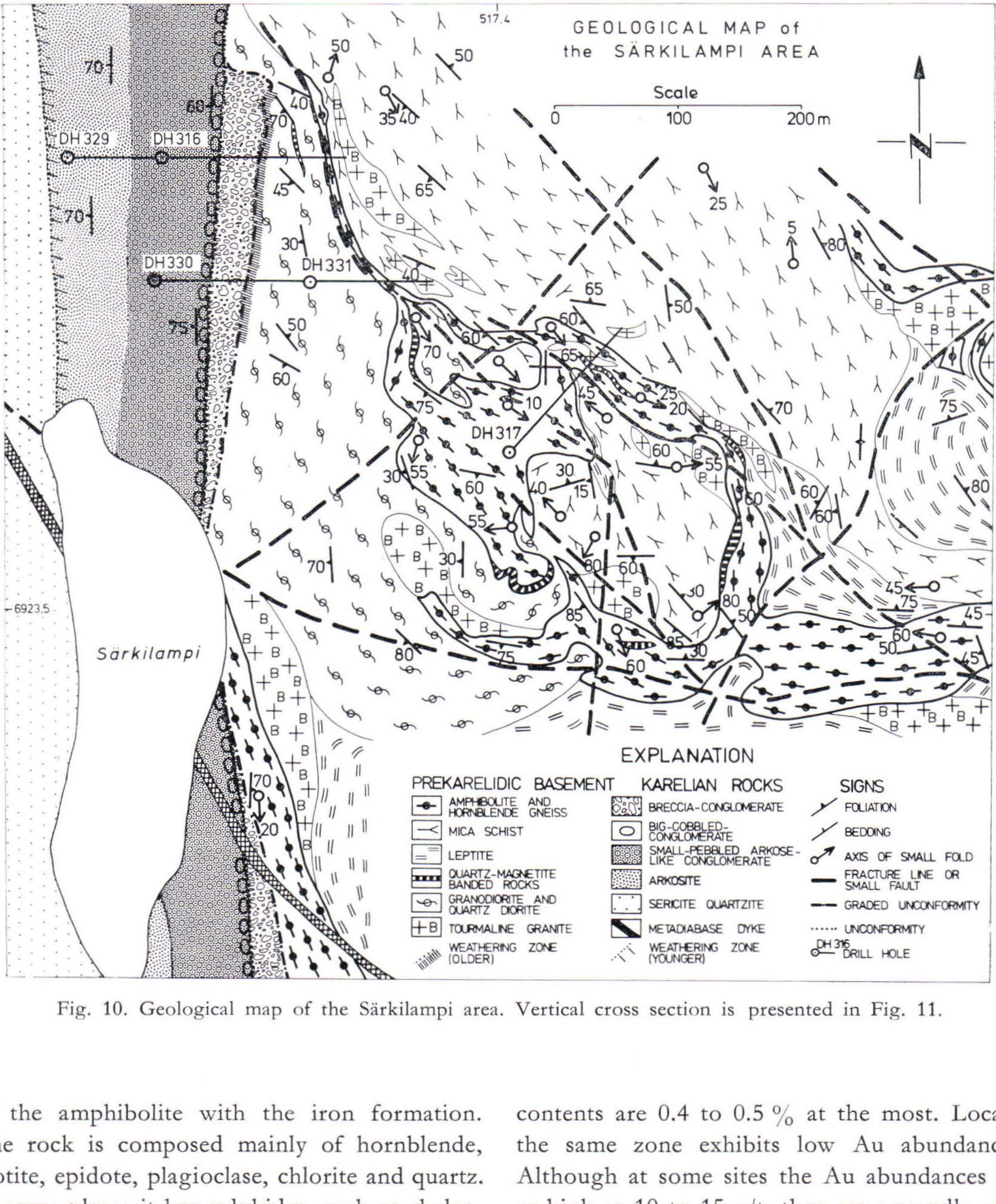

Fig. 10. Geological map of the Särkilampi area. Vertical cross section is presented in Fig. 11.

of the amphibolite with the iron formation. The rock is composed mainly of hornblende, biotite, epidote, plagioclase, chlorite and quartz. In some places it has sulphides, such as chalcopyrite, pyrrhotite and molybdenite, as dissemination. Also encountered in the area are crosscutting fractures filled with pyrite. Economically, however, the mineralization is worthless, for the thickness of the mineralized layers seldom attains even one metre and the Cu-Mo contents are 0.4 to 0.5 % at the most. Locally the same zone exhibits low Au abundances. Although at some sites the Au abundances are as high as 10 to 15 g/t, they are generally only a fraction of a gram per ton and the occurrence is highly erratic.

The main minerals in the gneissose granoquartz dioritic rock are plagioclase (oligoclase-andesine), quartz, biotite and hornblende. In addition there are carbonate, epidote, chlorite,

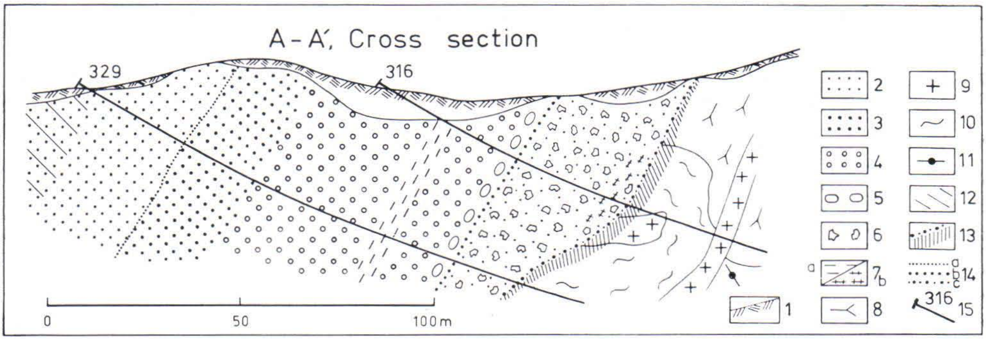

Fig. 11. Vertical cross section across the Arkosite Formation in the Särkilampi (Pölkkylampi) area in Fig. 10. Symbols: 1 = overburden; 2 = upper arkosite; 3 = lower arkosite; 4 = small-pebbled conglomerate; 5 = big-cobbled conglomerate; 6 = breccia-conglomerate; 7 a = argillaceous siltstone; 7 b = broken mudstone lenses; 8 = Prekarelidic granodiorite and quartz diorite; 9 = tourmaline granite; 10 = mica schist; 11 = amphibolite and hornblende gneiss; 12 = younger (Early Jatulian) weathering crust; 13 = older (Early Karelian) weathering crust; 14 a = unconformity; 14 b = major unconformity; 14 c = graded unconformity; 15 = drill hole.

apatite, titanite, tourmaline and zircon. The major minerals in the tourmaline granite are plagioclase (albite-oligoclase), potassium feldspar, quartz, tourmaline, biotite and sericite.

A few tens of metres from the contact with the Karelian formations there is a marked increase in fracturing in the rocks of the Prekarelidic basement. Close to the contact there is a zone 1 to 5 m wide in which the rock is more intensely fractured; along the margins of the fragments it is also altered. It may well be that this is a relic of an ancient weathering crust in which, as a result of recrystallization, the weathering products occur as sericite, carbonate and chlorite. A rock that has undergone particularly intense alteration has been encountered underneath the breccia-like conglomerate.

The Breccia-conglomerate Formation

The Karelian Sequence usually begins with a bed of heterogeneous conglomerate of varying thickness. In the Särkilampi area it rests unconformably on the rocks of the Prekarelidic basement (Fig. 11). On account of the angular fragments (Fig. 32), it is called breccia-conglomerate in the present context. Locally its contact with the basement is gradual.

The breccia-conglomerate occurrence at Särkilampi is about 1 km long and pocket-like; it is thickest south of Pölkkylampi, where its maximum width has been estimated to be 40 m.

The Särkilampi breccia-conglomerate contains fragments of divers size mixed up with each other, which upwards tend to decrease in size. The largest fragments close to the bottom are over one metre in diameter. The subangular fragments include various rocks of the Prekarelidic basement such as grano-quartz diorite, tourmaline granite, amphibolite, mica schist, leptite and iron-formation rocks (Fig. 12). A few of the fragments have preserved, as documents of their past, minor folds, quartz and pegmatite veins, old joints and magnetite bands. Some, like the tourmaline granite fragments, are well preserved; others, like the amphibolite fragments, in which the hornblende has completely altered into chlorite, exhibit intense alteration. Recognizable among the fragments are the same rock types that are encountered in the adjacent outcrops; hence the fragments were transported for only a very short distance. The finest fraction is composed mainly of quartz, sericite, chlorite, biotite and carbonate. The breccia-conglomerate reveals sporadic broken layers of mudstone a

Fig. 12. Unconformity of breccia-conglomerate over Prekarelidic quartz diorite. Length of tag 12 cm. Särkilampi, Kiihtelysvaara. Photo by E. Halme.

few cm thick. In the basal part of the formation there are carbonate-rich lenses a few cm or dm thick.

The Arkosite Formation

In the Särkilampi area the Arkosite Formation begins with big-cobbled conglomerate (Big-cobbled conglomerate Member), which often rests unconformably on breccia-conglomerate, but in some places directly overlies the Prekarelidic basement (Fig. 10). Unlike the fragments in the breccia-conglomerate, the pebbles in the Särkilampi big-cobbled conglomerate (Fig. 36) are distinctly rounded and represent only a few rock types. The diameter of the pebbles and cobbles varies from 2 to 25 cm. The predominant rock types are pale oligoclase granite and tourmaline granite. Quartz diorite and vein quartz are less common. Schist fragments are very rare, and thus the conglomerate is pale grey or pinkish in colour. The matrix is arkositic, containing quartz, sericite, feldspars, carbonate, biotite and chlorite. It is evident that the material is not so local as in the breccia-conglomerate.

For a short distance the big-cobbled conglomerate passes upwards into small-pebbled conglomerate (Small-pebbled conglomerate Member). It was deposited cyclically, each phase consisting of a thicker conglomeratic lower part and a thinner clay slate upper part that repeat themselves every 0.2 to 1.2 m (Fig. 33). Within each phase the colours vary in such a way that the conglomeratic portion is pinkish and the slaty portion greenish. Further variation to this colour scheme is provided by the pink carbonate lenses and nodules a few cm or dm thick that are particularly abundant below the clay slate portion (Fig. 33). Between the conglomerate layers, however, there is a thicker (4 m) clay slate layer with thin silt intercalations (Fig. 11). The diameter of the pebbles in the conglomerate varies from 4 to 16 mm and they are slightly rounded. The pebbles include quartz, feldspars, oligoclase granite, tourmaline granite and pegmatite; other types are rare. The matrix is arkositic and contains sericite, carbonate, biotite and chlorite in addition to quartz and feldspar. The major components of the clay slate are quartz micas and carbonate. The abundance of quartz is higher in the silty layers than in the clay slate.

Upwards in the sequence the small-pebbled conglomerate is succeeded by a layer about 20 m thick of horizontally bedded arkosite (Lower arkosite Member), greenish or pinkish in colour. The rock is composed of granular quartz, plagioclase and potassium feldspar grains embedded in a matrix rich in sericite and carbonate.

The Lower arkosite is overlain by the greenish and somewhat more coarse arkosite that shows cross-bedding (Upper arkosite Member). It contains sporadic vein quartz pebbles, 1 to 4 cm in diameter, and more rarely granite or leptite pebbles (Fig. 39). The granular grains are quartz and potassium feldspar. The interstices between the larger grains are filled with matrix composed mainly of sericite and quartz. Upwards in the sequence the feldspar of the Upper arkosite is intensely altered (Fig. 40F), so that in the topmost layer it is completely replaced by sericite. The top layer of the arkosite deposit presumably underwent intense chemical weathering in situ (Fig. 42). The rocks of the Lower quartzite Formation deposited unconformably on the intensely weathered rock.

The sequence

In the Särkilampi area (Fig. 11) the sequence of the lower Karelian formations is as follows (estimated thickness in brackets):

(Lower quartzite Formation, 550 m)

---------------------- unconformity ----------------------

Arkosite Formation (135 m)

Upper arkosite Member (60 m); greenish or pinkish, cross-bedded arkosite; upper part intensely altered, lower part fairly fresh

Lower arkosite Member (20 m); greenish or grey, horizontally bedded arkosite

Conglomerate Member (55 m);

— pinkish-green, horizontally bedded, small-pebbled conglomerate (34 m)
— clay slate, with thin silt layers and lenses (4 m).
— pinkish-green, horizontally bedded, small-pebbled conglomerate (11 m)
— big-cobbled conglomerate (6 m)

-------------------- major unconformity --------------------

Breccia-conglomerate Formation (max. 40 m)

-------------------- graded unconformity --------------------

Prekarelidic basement (diverse rock types).

From the depositional basement to the contact with the Lower quartzite Formation, the sequence is roughly 175 m thick.

The Kalkunmäki area

The Kalkumäki area, which is composed of the hill of that name and its environment (Figs. 2 and 13), constitutes the SE extension of the Särkilampi area. On this hill, both the Lower and Upper quartzite Formations as well as the intervening Volcanite Formation are well exposed. Furthermore, in the western part of the area, a quartz-bearing dolomite crops out that, in the sequence, overlies the Upper quartzite. On the same slope there is a fracture zone a few metres wide that, at a low angle, cuts the layering of the upper quartzite and the overlying quartz-bearing dolomite. This is not the first time that the Mn-bearing minerals occurring in this fracture have aroused the interest of investigators (cf. Hyvärinen and Siikarla 1971; Nykänen 1971b).

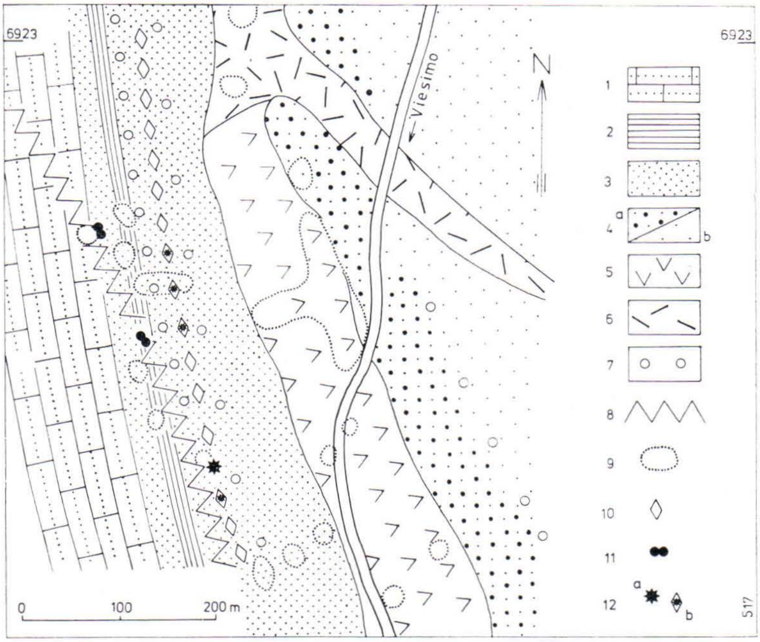

Fig. 13. Detailed geolocigal map of the Kalkunmäki area. Symbols: 1 = carbonate-bearing quartzite; 2 = clay slate—phyllite; 3 = upper quartzite; 4a = feldspathic quartzite; 4 b = orthoquartzite; 5 = metavolcanite; 6 = metadiabase; 7 = conglomerate interbeds; 8 = breccia zone; 9 = outcrop; 10 = andalusite; 11 = Mn-bearing minerals; 12 a = lazulite; 12 b = phosphate-bearing minerals.

The Lower quartzite Formation

In this area, as in the Hyypiä area, the Lower quartzite Formation resting on the Arkosite Formation begins with sericite quartzite that upwards grades into a more pure orthoquartzite. Unlike in the northern areas, the orthoquartzite is overlain by a layer of feldspathic quartzite, a few metres thick, succeeded by the Volcanite Formation. In addition to granular quartz grains, the feldspathic quartzite also contains rounded and fairly fresh potassium feldspar grains that impart to the rock a pinkish hue. The matrix between the quartz and potassium feldspar grains consists mainly of quartz and sericite. Small and pinkish andalusite porphyroblasts are sometimes visible.

The Volcanite Formation

In the Kalkunmäki area, the Volcanite Formation is about 80 m thick and rests on the Lower quartzite Formation (Fig. 13). This Volcanite Formation includes lava sheets of slightly varying composition, amygdaloids and, topmost, tuffite layers. Plagioclase (albite-oligoclase) phenocrysts are visible here and there in the lava rocks. The finer groundmass contains plagioclase, chlorite, biotite, quartz, titanite and magnetite. Some layers have amygdules filled with quartz or calcite. In the upper part of the Volcanite Formation there is a layer whose phenocrysts are of pale hornblende and the matrix predominantly of hornblende and plagioclase. The topmost beds in the Volcanite Formation are heterogeneous biotite- and chlorite-rich rocks, obviously tuffs and tuffites in origin.

The Upper quartzite Formation

The colour of the quartzites in the Upper quartzite Formation overlying the Volcanite Formation varies from grey to pink. This is because the matrix between the granular quartz grains contains, apart from quartz and sericite, variable amounts of chlorite, magnetite and fine-grained hematite. Roughly 50 m from the volcanites there are some layers a few dm

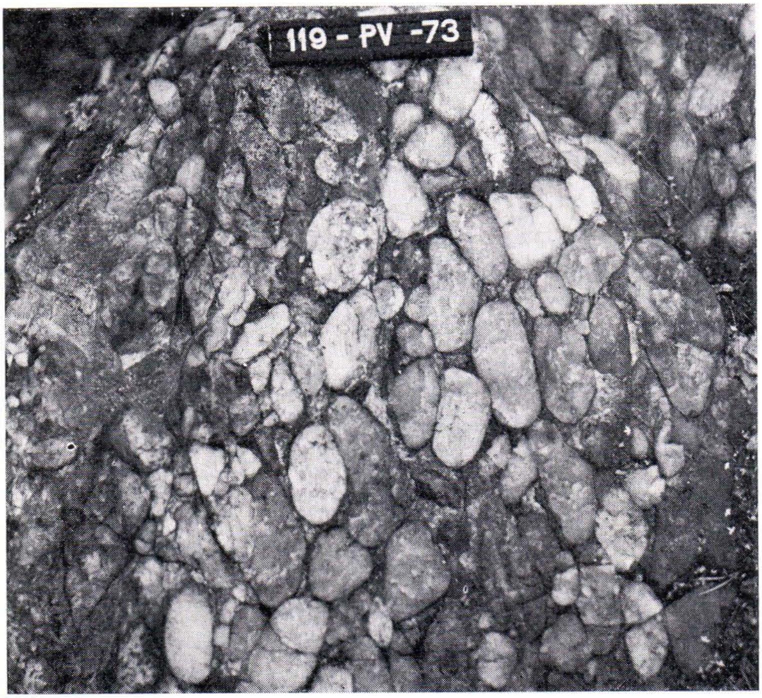

Fig. 14. Conglomerate with quartz pebbles and cobbles in hematite-rich matrix from the middle part of the Upper quartzite Formation. Length of tag 12 cm. Kalkunmäki, Kiihtelysvaara. Photo by E. Halme.

thick that are densely spotted with pink andalusite porphyroblasts that contain inclusions of lazulite and rare Sr- and Ba-bearing sulphate and phosphate minerals (p. 92). In the topmost part of the quartzite formation the quartz-sericite matrix is replaced by carbonate.

In the lower and middle parts of the Upper quartzite Formation there are abundant conglomerate layers from a few dm to a couple of metres thick. Their pebbles and cobbles are of bluish, white or pink vein quartz but they also include volcanites (Fig. 14). The quartz sandy matrix is rich in sericite and often tinged pink by fine-grained hematite. In some places hematite may be the predominant matrix mineral.

Clay slate and phyllite intercalations a few dm or metres thick occur sporadically in the upper part of the quartzite deposit.

The Kalkunmäki breccia

On the western slope of Kalkunmäki there is a breccia zone a few metres wide that cuts diagonally the bedding in the Upper quartzite (Fig. 13). In several places the rock has been broken into pieces the size of the palm of a hand. Wherever the breccia zone cuts the 4 m thick clay slate—phyllite bed, the rock is unmistakably pink owing mainly to Mn-bearing mica, although other Mn minerals, such as piedmontite, Mn-bearing tourmaline and possibly hollandite are also present (Hyvärinen and Siikarla 1971; Nykänen 1971b). Northwestwards in the zone, even the carbonate-bearing quartzite contains Mn-bearing mica and piedmontite. The Mn-bearing minerals seem to be genetically correlated with the formation of the breccia zone. The manganese was presumably mobilized from the layers with Mn nodules in the carbonate-bearing quartzite (p. 92).

The Viistola section

Throughout the study area the sediments succeeding the Volcanite Formation are sur-

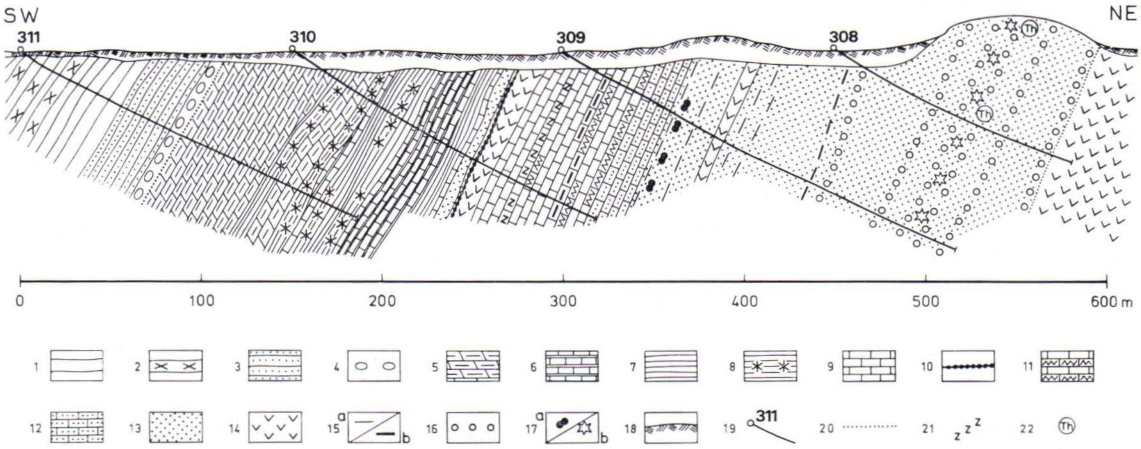

Fig. 15. A section through the middle and upper parts of the Jatulian sequence and the lower part of the Kalevian sequence adjacent to the Viistola farm in Kiihtelysvaara. Symbols: *Kalevian:* 1 = mica schist; 2 = staurolite-bearing mica schist; 3 = quartzite; 4 = conglomerate. *Jatulian:* 5 = tremolite-bearing dolomite; 6 = laminated dolomite; 7 = carbonaceous slate; 8 = carbon-rich layers; 9 = dolomite; 10 = hematite-rich layer; 11 = chert nodules in dolomite; 12 = quartz-bearing dolomite; 13 = upper quartzite; 14 = metavolcanite; 15 a = pelitic intercalations; 15 b = volcanic intercalations; 16 = quartz conglomerate interbeds. Other symbols: 17 a = manganese nodules; 17 b =l azulite-bearing nodules; 18 = overburden; 19 = drill hole; 20 = unconformity; 21 = breccia; 22 = thorium-bearing minerals.

prisingly poorly exposed, particularly with regard to dolomite and carbonaceous slate beds, and it has not been easy to determine the depositional sequence. Hence, near Viistola, a farm northwest of Särkilampi, the formations from the Volcanite Formation upwards were intersected by diamond drilling. The section (Figs. 2 and 15), compiled on the basis of drilling data from four drill holes, has in fact turned out to be the key to the stratigraphy of the study area.

The Upper quartzite Formation

Near Viistola the Volcanite Formation is overlain by a formation roughly 200 m thick of heterogeneous Upper quartzites that are grey or pink in colour. The grey hue is due to detrital volcanic weathering material, such as chlorite and magnetite, which is present in the matrix between the quartz grains. The pinkish hue derives from hematite particles in the matrix. In the topmost part of the quartzite formation, the quartz and sericite matrix between the quartz grains is gradually replaced by carbonate. These layers also contain some Mn-rich nodule-like aggregates a few cm long (p. 92).

The lower and middle parts of the Upper quartzite Formation have abundant conglomerate beds a few cm or dm thick. Since these beds tend to be very regular, it has been possible to trace some of them for as much as 200 m. The pebbles and cobbles are mainly of bluish, white or pink vein quartz, although volcanites have also been encountered. In general, the diameters of the fragments in the conglomerate measure from 2 to 6 cm although the largest may be as much as 12 cm. The pinkish and sandy matrix is composed chiefly of quartz, sericite, chlorite and fine-grained hematite. Detrital magnetite and tourmaline are concentrated in some layers. Some of the conglomerate beds are slightly radio-active owing to Th-bearing minerals such as brockite and thorite. The quartzite between these beds displays sporadic pebble-like aggregates a few mm to a couple of cm in diameter. They are

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Fig. 16. Impure quartzite with tuffitic interbeds from the Upper quartzite Formation. Length of tag 12 cm. Viistola, Kiihtelysvaara. Photo by E. Halme.

enveloped by a reddish-brown shell; the core is composed mainly of quartz and a bluish mineral that has been indentified as lazulite (p. 89).

Here and there in the middle and upper parts of the Upper quartzite Formation there are laminated clay slate and silt intercalations whose thickness varies from a few mm to several metres (Fig. 45D). Their predominant minerals are chlorite, sericite and quartz. Some tuffitic intercalations have also been met with that fluctuate in thickness from a few cm to a few metres (Fig. 16); they are mainly composed of chlorite, biotite, quartz, sericite, albite and magnetite.

The Dolomite—Volcanite Formation

The Upper quartzite Formation is succeeded in the sequence by pinkish, unfoliated or slightly layered carbonate rocks predominantly dolomitic in composition (Fig. 53A). Thin quartzite and clay slate intercalations are met with in some places in the lower part of this Dolomite Member. Higher up, the intercalations grow less frequent and the rock passes into a slightly layered and more massive variant. The carbonate rock in the lower part is impure, containing such minerals as sericite, chlorite and quartz; fewer impurities have been noted in the rock in the upper part. Most of the carbonate is dolomite, although calcitic veins and nodules occur regularly. Abundant stylolite seams are encountered in association with the calcitic portions (Fig. 53A). The carbonate rock deposit is also characterized by some layers, a few dm to a couple of metres thick, that contain chert nodules. The predominant type has small overlapping nodules very close to each other (Fig. 53B), although larger bodies and lenses have also been encountered. The chert layers seem to have brecciated readily

(Fig. 53C). The nodules are composed of fine-grained, pale or greyish quartz, and the intervening space is filled with a pink, mainly calcitic matrix.

The dolomite layers are overlain by a metavolcanite layer (Volcanite Member) roughly 10 m thick. This greenish, mottled rock is composed chiefly of chlorite, biotite, plagioclase, quartz and magnetite. The mottles are small amygdules filled with carbonates, quartz and hematite. The rock is obviously altered lava in origin.

The Hematite rock—Quartzite Formation

The metavolcanite bed is succeeded by a thin layer of brecciated chert overlain by a layer about 1 m thick of almost compact hematite. This layer is composed of massive hematite and small hematite laths between which there are small amounts of chlorite, biotite and quartz (Fig. 56C). Larger hematite crystals are encountered here and there. The hematite rock layer is overlain first by a layer of clay slate stained red by hematite (ferriferous slate), and then by a layer of quartzite about 5 m thick. The rounded quartz grains in this quartzite are embedded in a matrix composed of fine-grained quartz, sericite, biotite, chlorite and hematite.

Dolomite—Carbonaceous slate—Volcanite Formation

Above the quartzite layer there is a thin layer of calcitic carbonate rock, carmine red in colour, overlain by a deposit of fine-grained and laminated dolomite. The alternating pink and green laminae are due to the alternation of clayey (micas, chlorite) and carbonate layers. Small amounts of quartz are almost invariably present.

Upwards in the sequence the laminated dolomite grades into greenish clay slate that for a distance of a few metres again grades into dark-grey fine-grained graphite-bearing carbonaceous slate. At its most typical, this carbonaceous slate is a fine-grained rock composed of sericite, chlorite, biotite, quartz and fine graphite. Westwards the rock grades into a more intensely metamorphosed and phyllite type. Slickensides and breccias lined with graphite are encountered in the portions richest in graphite (Fig. 54B). Owing to secondary processes, the graphite scales on the slickensides have increased in size. Narrow joints filled with calcite are common. On the basis of Slingram anomalies, it has been concluded that the graphite-rich layers extend for roughly 2 km.

Upwards the carbonaceous slate layers grade into banded dolomite (Fig. 15) in which wider greenish bands alternate with narrower grey bands (Fig. 54C). In addition to dolomite, the greenish dolomitic bands contain tremolite and minor talc and micas. In some layers tremolite occurs as radial bunches. The grey bands are phyllite that contains fine-grained graphite as well as micas, chlorite and quartz. The banded rock was presumably formed from clayey carbonates in the course of metamorphism. The banded dolomite also contains some wider phyllite intercalations that may include volcanic material.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The contact between the Dolomite—Carbonaceous slate—Volcanite Formation and the Turbidite conglomerate-quartzite Formation was intersected by a drill hole (Fig. 15, dh 311). It was noticed that the Turbidite conglomerate—quartzite Formation rests unconformably on banded dolomite, beginning with a conglomerate layer about 20 cm thick (Fig. 59C). The conglomerate grades into a dark, graded-bedded quartzite in which each phase includes a quartzitic lower part and a phyllitic upper part. Except for small amounts of sericite, the quartzitic portion is composed almost entirely of quartz (Fig. 61A) with minor tourmaline and potassium feldspar. Upwards in the sequence, the graded-bedded quartzite grades for a short distance into a graded-bedded mica schist (Mica schist Formation). Some tens of metres from the contact with the Jatulian formations the mica-rich bands of mica schist exhibit abundant staurolite and, less frequently, almandine porphyroblasts (Fig. 61B).

The sequence

The succession in the Viistola section (Fig. 15) is from the youngest downwards (estimated thickness in brackets):

Mica schist Formation; staurolite-bearing mica schist

Turbidite conglomerate—quartzite Formation; dark quartzite (37 m) underlain by conglomerate (0.20 m)

------------ unconformity ------------

Dolomite—Carbonaceous slate—Volcanite Formation (145 m)

- greenish-grey banded tremolite-bearing dolomite alternating with carbonaceous slate (75 m)
- dark-grey carbonaceous slate; dolomite intercalations in upper portion (33 m)
- pink-green laminated fine-grained dolomite; thin clay slate intercalations (35 m)
- carmine red, massive calcitic dolomite (1.7 m)

Hematite rock—Quartzite Formation (8.2 m)

- pinkish, impure quartzite (5 m)
- pink-green laminated hematite-bearing clay slate (2.3 m)
- hematite rock layer (0.9 m)

Dolomite—Volcanite Formation (84 m)

- metavolcanite, topmost a thin layer of brecciated cherty dolomite (10 m)
- pale or pinkish dolomite with calcitic lenses and veins, and thin clay slate and quartzite intercalations in lower part (74 m)

Upper quartzite Formation (205); impure quartzite; in upper part a few clay slate and volcanite intercalations and in lower part thin conglomerate interbeds.

Measured up to the basal contact of the Turbidite conglomerate—quartzite Formation, the sequence that rests on the Volcanite Formation is about 440 m thick.

The Viesimo area

Southeast of Raatevaara and north of the river Viesimojoki there is a hill known as Härkäkallio. It abounds in conglomerate outcrops (Figs. 2 and 17), which have been described by several authors, e.g. Frosterus and Wilkman (1920), Carlson (1967) and Nykänen (1971b).

The Arkosite Formation

In the Viesimo area the Karelian Sequence begins with an exceptionally thick deposit of conglomerate—arkosite whose thickness, although difficult to estimate because its contact with the Prekarelidic basement is not exposed, may be some 400 m. Erratic blocks of breccia-conglomerate imply that the lowest layer is breccia-conglomerate. Most of the outcrops, however, are of big-cobbled conglomerate like that at Särkilampi (ø = 2—40 cm), even though the proportion of matrix is higher. In some places, small-pebbled conglomerate (ø = 4—20 mm) occurs as lenses. Most of the pebbles, cobbles and boulders in the conglomerate are Prekarelidic granitoids such as oligoclase granite, tourmaline granite and grano-quartz diorite. Large (ø = 5—10 cm), subangular quartz fragments are also encountered. Fragments of Prekarelidic schists are very rare; the schist pebbles encountered are either mica schist or porphyritic amphibolite whose hornblende has completely altered into chlorite. The arkosic matrix of the conglomerate is composed mainly of quartz, feldspars, micas, chlorite and carbonate, with some small amounts of tourmaline, titanite, epidote and apatite. In the lower part of the deposit, carbonate lenses are met with that contain tremolite and chlorite in addition to the minerals given above. Thin clay slate intercalations also occur.

Exposed on the conglomerate is heterogeneous arkosite (Lower arkosite Member) a few metres thick that contains appreciable

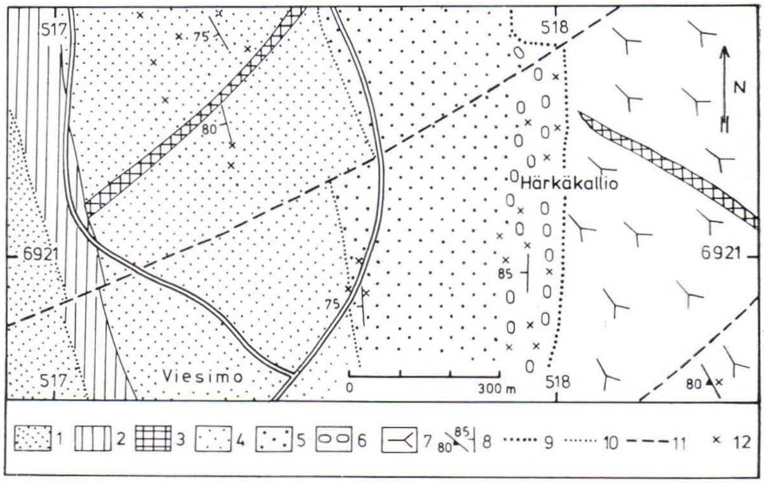


Fig. 17. Geological map of the Viesimo area. *Jatulian:* 1 = upper quartzite; 2 = metavolcanite; 3 = metadiabase; 4 = lower quartzite. *Prejatulian:* 5 = arkosite; 6 = Viesimo conglomerate. *Prekarelidic:* 7 = granodiorite and quartz diorite. Other symbols: 8 = bedding and foliation; 9 = major unconformity; 10 = unconformity; 11 = fracture or fault; 12 = outcrop.

granular plagioclase and quartz (Table 7, No. 26; p. 73). Westwards, a wide zone devoid of outcrops, and about which little is known, extends for about 200 m.

The first outcrops beyond the exposure-free zone are fairly intensely altered arkosite (Upper arkosite Member). The feldspar in the rock is almost completely sericitized. In places the sericite-rich matrix is so soft that quartz pebbles and cobbles can be loosened from the rock by hand. The arkosite exhibits bag-like forms, 10 to 50 cm long, that are stained red by iron oxide and are obviously precipitation structures.

The contact of the Arkosite Formation with the Lower quartzite Formation

Although the contact of the Arkosite Formation with the Lower quartzite Formation is not exposed, the alterations (weathering crust) in the Upper arkosite Member suggest that the contact is not far from these outcrops. The trends of the lower quartzite and arkosite conglomerate layers differ from each other, the latter (roughly N—S) deviating from the former by roughly 20 ° northwards. This might mean that, separated by a subangular unconformity, the Lower quartzite Formation rests on the Arkosite Formation (Fig. 17).

The Valkeavaara section

The quartzite ridge that extends from Kiihtelysvaara to Raatevaara is interrupted at the Viesimojoki. Farther south, it continues at practically the same height as before. On the side of the road to Valkeavaara there are some outcrops of metavolcanite and metadiabase of the Dolomite—Carbonaceous slate—Volcanite Formation showing fine-grained disseminated chalcopyrite. The volcanite layers and the overlying dolomite and carbonaceous slate layers were intersected by diamond drilling (Figs. 2 and 18) to check the magnetic and electromagnetic anomalies.

The Volcanite Formation and the Upper quartzite Formation

The metavolcanite (Volcanite Formation) overlying the Lower quartzite Formation consists of slightly blastoporphyric lavas, amygdaloids and, highest in the sequence, tuffs and tuffites. In mineral composition, these are analogous to the metavolcanites in the Kalkunmäki area.

The Volcanite Formation is succeeded by impure quartzites of the Upper quartzite For-

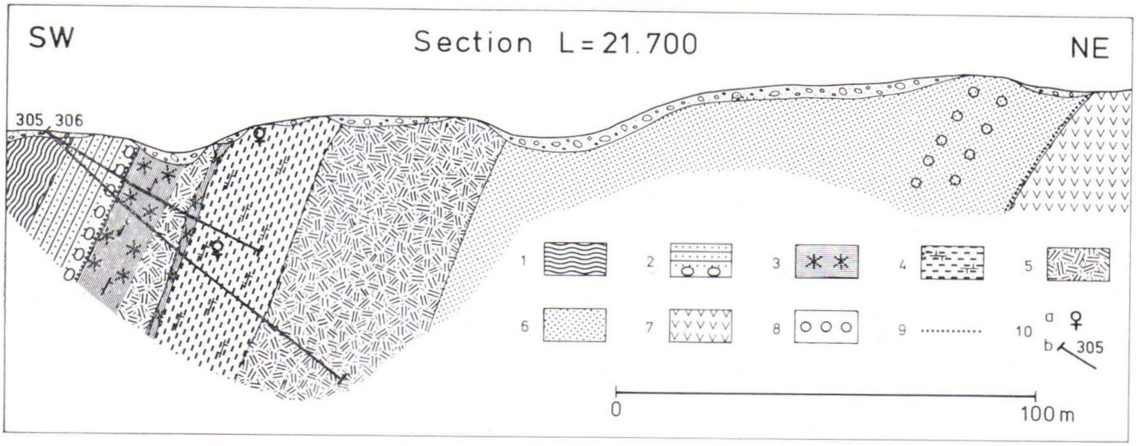


Fig. 18. A section through the middle and upper parts of the Jatulian sequence and the lower part of the Kalevian sequence. Valkeavaara, Kiihtelysvaara. Symbols: *Kalevian:* 1 = mica schist; 2 = turbidite conglomerate and quartzite. *Jatulian:* 3 = graphite-bearing phyllite; 4 = alternating dolomite and volcanic layers; 5 = gabbroic metadiabase; 6 = upper quartzite; 7 = metavolcanite, mainly lava; 8 = conglomeratic interbeds. Other symbols: 9 = unconformity; 10 a = chalcopyrite; 10 b = drill hole.

mation. The conglomerate interbeds that occur in the lower part of this quartzite formation are exceptionally thick, being as much as 2 to 4 m.

The Dolomite—Carbonaceous slate—Volcanite Formation

The Upper quartzite Formation is overlain by alternating thinner dolomite layers and thicker volcanite layers. Upwards, these grade into alternating layers of dolomite and carbonaceous phyllite. The dolomitic layers are heterogeneous, containing tremolite bands, talc and chlorite. The volcanite layers are predominantly banded tuffs and tuffites, but they also include lavas. The chief minerals in the volcanites are plagioclase (oligoclase), biotite, pale amphibole, quartz and chlorite. The lavas exhibit scattered pale-green amphibole phenocrysts. Some of the volcanite layers contain fine-grained chalcopyrite dissemination, which is, however, without economic significance, the copper content being below 0.1 %. The carbonaceous phyllite is composed of biotite, sericite, quartz and fine-grained graphite. Abundant pyrite is occasionally encountered adjacent to the metadiabase sills.

The metadiabase sills

Gabbroic metadiabase has intruded as sills between the dolomite—volcanite—carbonaceous phyllite layers (Fig. 18). The widest of them, about 50 m thick, has intruded the upper contact of the Upper quartzite. The chief minerals in the gabbroic metadiabase are plagioclase (andesine), hornblende and biotite. The minor constituents are quartz, epidote, chlorite and titanite.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The contact of the Dolomite—Carbonaceous slate—Volcanite Formation with the Turbidite conglomerate—quartzite Formation was intersected by two drill holes (Fig. 18), which showed that the Turbidite conglomerate—quartzite Formation begins with a conglomerate layer about half a metre thick (Fig. 59B) that rests unconformably on carbonaceous phyllite. The conglomerate is succeeded by a deposit of dark graded-bedded quartzite about 15 m thick (Fig. 60). This grades for a short distance into mica schist with staurolite and andalusite porphyroblasts.

The sequence of the Karelian formations

From top to bottom, the sequence in the Valkeavaara area (Fig. 18) is as follows (estimated thickness in brackets):

Mica schist Formation; mica schist with staurolite and andalusite porphyroblasts

Turbidite conglomerate—quartzite Formation; dark quartzite (15 m) underlain by conglomerate (0.50 m)

---------------------- unconformity ----------------------

The Dolomite—Carbonaceous slate—Volcanite Formation (78 m)

— dark grey, carbonaceous phyllite with thin dolomite intercalations (11 m)

— a sill of medium-grained metadiabase (5.5 m)

— carbonaceous phyllite (1.5 m)

— a bed of alternating dolomite, carbonaceous phyllite and metavolcanite layers (20 m)

— a sill of medium- to coarse-grained gabbroic metadiabase (40 m)

Upper quartzite Formation (110 m); impure quartzite with abundant conglomerate interbeds in lowest part

---------------------- unconformity ----------------------

Volcanite Formation (70 m)

Lower quartzite Formation (350 m); upper part orthoquartzite, lower part sericite quartzite with thin quartz conglomerate interbeds in lower part

---------------------- unconformity ----------------------

(Arkosite Formation, 100 m?)

-------------------- major unconformity --------------------

The Prekarelidic basement (gneissose grano-quartz diorite).

Measured from the depositional basement to the contact with the Turbidite conglomerate—quartzite Formation, this sequence is about 700 m thick.

The Haluksenlammit area

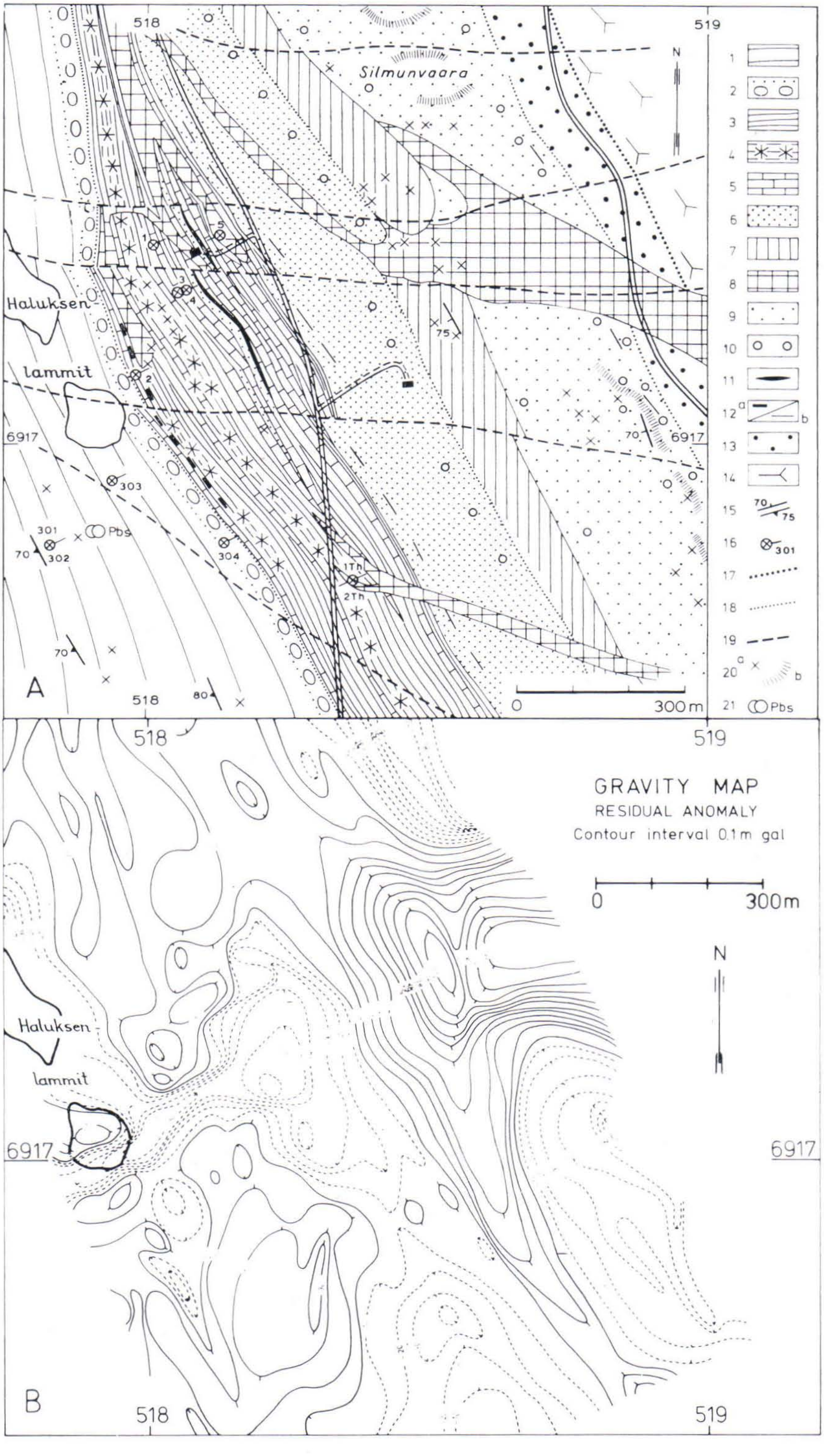
About two kilometres south of Valkeavaara, there are three ponds that together are known as Haluksenlammit (Figs. 2 and 19A). Eastwards, the terrain is traversed by a quartzite ridge that, every 200 to 500 m, is cut diagonally by ravines and gorges that were formed by metadiabase dykes eroded deeper than the environment. A hematite showing encouraged the Geological Survey to conduct detailed mapping, geophysical survey and diamond drilling in the area in 1961 (Hyvärinen and Siikarla 1971). In 1974, these were followed up by supplementary studies triggered by a Pb-Zn-Cu showing. Both investigations provided a number of interesting details concerning the stratigraphy of the area; these are briefly described in the text.

The Arkosite Formation and the Lower quartzite Formation

The contact of the Prekarelidic basement with the Karelian formations is not exposed in the Haluksenlammit area. Erratic blocks, however, suggest that the Karelian sequence begins with arkosite (Arkosite Formation). The outcrops on the eastern slope of the quartzite ridge are sericite quartzite that westwards grades into orthoquartzite (Lower quartzite Formation).

The Volcanite Formation

The Lower quartzite Formation is overlain by the 50 to 80 m thick Volcanite Formation (Fig. 19A), which is composed of basic lavas, amygdaloids and, topmost, pyroclastites. Predominant are slightly blastoporphyric lavas, which, here and there, exhibit plagioclase (albite-oligoclase) and hornblende phenocrysts in an aphanitic matrix of plagioclase, biotite, hornblende, chlorite, epidote, titanite and magnetite. Some beds contain amygdules, 2 to 30 cm in diameter, filled with quartz or calcite. The uppermost beds in the Volcanite Formation are heterogeneous rocks rich in biotite and chlorite and are presumably tuffs and tuffites in origin.


518
519
Silmunvaara
N
Haluksen
lammit
6917
303
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Pbs
304
1Th
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18
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20 a
b
21 Pbs
GRAVITY MAP
RESIDUAL ANOMALY
Contour interval 0.1 m gal
0
300m
B

The Upper quartzite Formation

The Volcanite Formation is overlain by rocks of the Upper quartzite Formation. The lower part of this formation is characterized by quartz conglomerate interbeds, and the upper part by clay slate and phyllite intercalations.

The Dolomite—Volcanite Formation

As in the Viistola section, the Upper quartzite Formation is succeeded by a deposit of weakly layered, dolomitic carbonate rocks (Fig. 19A). In some places in the lower part of the sequence there are quartzite and clay slate—phyllite intercalations that upwards become less frequent as the rock grades into a more massive dolomite. The carbonate rocks are similar to those in the Viistola section, although, judging by the tremolite bands, they were somewhat more intensely metamorphosed. The carbonate is usually dolomite, but lenses and veins of calcite are also encountered. There are also some chert nodules, but not so commonly as in the Viistola section.

The Hematite rock—Quartzite Formation

The dolomite layers are succeeded fairly sharply by a layer of almost compact hematite 1 to 1.5 m thick (Hyvärinen and Siikarla 1971). This layer is composed partly of massive hematite and partly of fine hematite laths interspersed with small amounts of micas, chlorite and quartz. Larger hematite crystals are encountered here and there. The hematite layer is overlain by a thin layer of clay slate stained red by hematite and overlain by a quartzite layer 4 to 5 m thick.

The Dolomite—Carbonaceous slate—Volcanite Formation

The quartzite is succeeded by a pinkish, laminated dolomite that contains quartz, chlorite, tremolite and some talc as impurities. According to chemical analysis (p. 104), the carbonate is dolomite-predominant.

Dark grey carbonaceous slate, or rather graphite-bearing phyllite, has deposited on the dolomites and contains scattered tremolite-bearing dolomite intercalations. Although seldom very high, the abundance of graphite in the phyllites is sufficient to tinge the rock dark grey. The deposit does, however, contain some layers, a metre or so thick, that are rich in graphite, as well as some layers, comparatively poor in graphite, that exhibit small staurolite porphyroblasts. The graphite-bearing phyllite layers and tremolite-bearing dolomite layers alternate in the uppermost part of the formation. Occasional volcanite intercalations are also encountered.

Tynni (1971) has recognised from the carbonate rock beds grain-shaped microfossils, about 30 μ in diameter, that may be algae in origin. Similar microfossils have also been encountered in carbonate rock xenoliths in metadiabases.

The metadiabase dykes and sills

The map of the Haluksenlammit area is characterized by metadiabase dykes (Fig. 19A) that, starting from the Prekarelidic basement, cut the quartzite beds. The wider metadiabase dykes have been traced not only owing to the magnetic anomalies associated with them but also with the aid of a gravimetric map (Figs.

Fig. 19. A) Geological map of the Haluksenlammit area. *Kalevian:* 1 = mica schist; 2 = turbidite conglomerate and quartzite. *Jatulian:* 3 = carbonaceous slate; 4 = carbon-rich layers; 5 = dolomite; 6 = upper quartzite; 7 = metavolcanite; 8 = metadiabase; 9 = lower quartzite; 10 = conglomerate interbeds; 11 = hematite-rich layer; 12 a = volcanic intercalations; 12 b = pelitic intercalations. *Prejatulian:* 13 = arkosite. *Prekarelidic:* 14 = granodiorite and quartz diorite. Other symbols: 15 = bedding and foliation; 16 = drill hole; 17 = major unconformity; 18 = unconformity; 19 = fracture of fault; 20 a = small outcrop; 20 b = large outcrop; 21 = galenite. B) Simplified residual anomaly gravity map of the Haluksenlammit area. Original map 1: 10 000 by Geophysics Department of the Geological Survey of Finland.

19A—B). Most of the metadiabase dykes are from 5 to 20 m wide; an exception is the Silmunvaara dyke, which is over 100 m wide. This dyke represents the ancient discharge channel of a lava flow that poured over the Lower quartzite, and was active on at least more than one occassion. The metadiabase dykes are usually medium-grained rocks composed of plagioclase (oligoclase-andesine) and variable amounts of hornblende, biotite, chlorite and quartz.

Diamond drilling has demonstrated that some of the metadiabase dykes also cut the Upper quartzite and as sills have penetrated the dolomite and carbonaceous phyllite layers in the higher horizons. The dykes and sills are mainly composed of plagioclase (oligoclase—andesine) and green hornblende. In addition to hornblende, they often contain pale fibrous amphibole, with biotite, chlorite, quartz, carbonate and leucoxene as other minor constituents.

The Turbidite conglomerate—quartzite Formation and the Mica schist Formation

The contact of the Dolomite—Carbonaceous slate—Volcanite Formation with the Turbidite conglomerate—quartzite Formation was intersected by drilling at two sites. It was established that, beginning with a conglomerate bed 2 to 4 m thick, the Turbidite conglomerate—quartzite Formation rests unconformably on either the dolomite or carbonaceous phyllite, depending on the location. The conglomerate exhibits rounded or angular pebbles and cobbles that include a variety of Karelian rocks such as quartzites, volcanites, metadiabases, graphite phyllites and vein quartz. The matrix is mainly quartz and micas, but there is also some local pyrite. The conglomerate is succeeded by a dark, graded-bedded quartzite that upwards grades into staurolite-bearing mica schist.

Somewhat west of the contact of the above quartzite there are phyllite or mica schist folds with well-preserved graded-bedding (Fig. 20). Along the axial plane the rock contains quartz veins that vary from a few cm to several metres in thickness. Occasionally the quartz veins contain nests of galena, sphalerite and chalcopyrite. The degree of mineralization is, however, low, and the occurrences incoherent and without economic value.

The sequence of the Karelian formations

In the Haluksenlammit area (Fig. 19A) the sequence, from top to bottom, is as follows (estimated thickness in brackets):

Mica schist Formation; staurolite-bearing mica schist
Turbidite conglomerate—quartzite Formation; dark quartzite (20 m) and polymictic conglomerate (4 m)

------------------------ unconformity ------------------------

Dolomite—Carbonaceous slate—Volcanite Formation (173 m)

- greenish-grey banded, tremolite-bearing dolomite alternating with carbonaceous phyllite (8 m)
- dark grey carbonaceous phyllite with volcanite intercalations (20 m)
- medium-grained metadiabase sill (20 m)
- dark grey carbonaceous phyllite, dolomite intercalations in upper part, graphite-rich layers in the middle (90 m)
- banded dolomite, thin phyllite intercalations in upper part, tremolite bands and layers with chert nodules in lower part (35 m)

Hematite rock—Quartzite Formation (5.7 m)

- pinkish, impure quartzite (4 m)
- red-green laminated, hematite-bearing clay slate (0.1—0.5 m)
- hematite rock layer (1—1.2 m)

Dolomite—Volcanite Formation (74 m)

- pale or pinkish, non-foliated dolomite with calcitic veins, tremolite bands and chert nodule layers (40 m)
- red-green banded clay slate (8 m)
- medium-grained metadiabase sill (13 m)
- pinkish dolomite with thin clay slate and quartzite intercalations (13 m)

Fig. 20. Folded, graded-bedded phyllite from the Mica schist Formation showing axial plane cleavage. Length of tag 12 cm. Haluksenlammit, Tohmajärvi. Photo by E. Halme.

Upper quartzite Formation (170 m); impure quartzite, occasional clay slate-phyllite intercalations in upper parts, thin conglomerate interbeds in lower parts

---------------------- unconformity ----------------------

Volcanite Formation (80 m)

Lower quartzite Formation (350 m); upper part orthoquartzite with thin quartz conglomerate interbeds in uppermost portions; lower part sericite quartzite with thin conglomerate interbeds in lowest portions

---------------------- unconformity ----------------------

(Arkosite Formation, 100 m?)

------------------- major unconformity -------------------

The Prekarelidic basement (gneissose grano-quartz diorite).

Measured from the depositional basement to the contact with the Turbidite conglomerate—quartzite Formation, this sequence is about 760 m thick.

The Värtsilä area

The quartzites of the Lower quartzite Formation can be readily traced south-southeast from the Haluksenlammit area as far as Saario (Figs. 2 and 21); for lack of exposures, however, the layers overlying the quartzites, such as volcanites, are difficult to follow. The poorly exposed area extends from Saario to Värtsilä. According to Nykänen (1967), the Jatulian formations pinch out between Saario and Värtsilä, although erratic blocks, aerial photos, ground geophysical surveys and airborne geophysics suggest that the quartzites and volcanites continue as far as Värtsilä (Fig. 21). There, through a narrow anticlinal ridge, they join the Sääperi syncline.

Around the lake Sääperi (Fig. 21) the Karelian formations occur in a small syncline whose fold axis plunges roughly southwards at an angle of 15° to 20° (Nykänen 1968; Räisänen

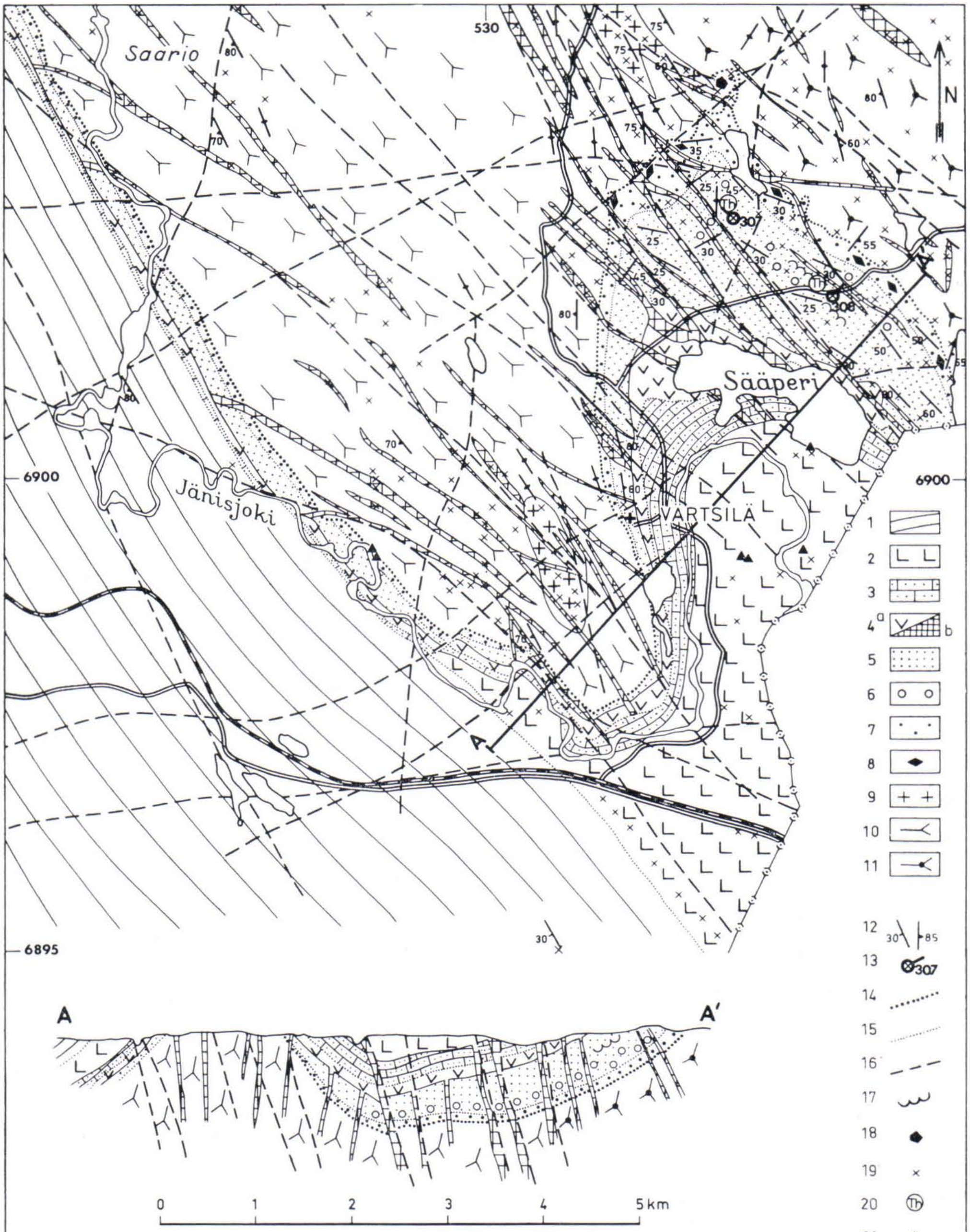

Fig. 21. Geological map of the Värtsilä area. *Kalevian:* 1 = mica schist with quartzitic layers at base. *Jatulian:* = 2 gabbroic metadiabase; 3 = quartz-bearing dolomite; 4 a = metavolcanite; 4 b = metadiabase; 5 = quartzite; 6 = quartz conglomerate interbeds. *Prejatulian:* 7 = arkosite; 8 = carbonate lenses. *Prekarelidic:* 9 = granite; 10 = granodiorite and quartz diorite; 11 = oligoclase granite (trondhjemite). Other symbols: 12 = bedding and foliation; 13 = drill hole; 14 = major unconformity; 15 = unconformity; 16 = fracture or fault; 17 = ripple marks; 18 = mud cracks; 19 = outcrop; 20 = thorium-bearing minerals; 21 = erratic block of dolomite.

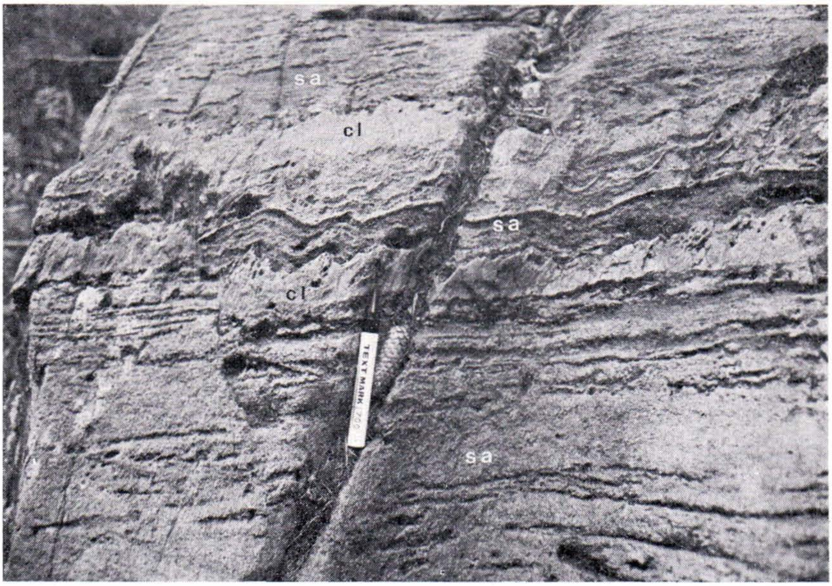

Fig. 22. Thicker sandy layers (sa) alternating with relatively thin finer-grained layers(cl) showing horizontal bedding and transverse cleavage in Lower arkosite. The pen, length 14 cm, is adjusted to show vertical direction. Sääperi, Värtsilä. Photo by E. Räisänen.

1975). The syncline is bordered by quartzite ridges that rise about 50 to 60 m above the surface of the lake. Outcrops abound on both sides of the ridges whereas the middle of the syncline is less exposed. On the eastern margin of the syncline the bedding is gentle, dipping 20 ° to 40 ° NW; on the western margin, though, it is steeper and eastwards. At the northern end of the syncline the metasediments exhibit vertical axial plane foliation that trends roughly N—S (Fig. 21).

Depositional basement of the Karelian formations

The contact of the Prekarelidic basement with the Karelian Sequence is visible in a number of outcrops. The sinuous nature of the contact indicates that the depositional basement was uneven. The basement is of either Prekarelidic grano-quartz diorite, oligoclase granite or granite. In some places beneath the contact there are relics of an ancient in-situ weathering breccia where the rock is still intensely fragmented a few metres from the contact (Fig. 29A). The margins of the fragments are intensely altered, the alteration products including sericite, chlorite and carbonate.

The Arkosite Formation

The Karelian Sequence begins with the Arkosite Formation, which, separated by an angular unconformity, rests on the Prekarelidic basement. Owing to the roughness of the depositional basement and the subhorizontal bedding of the arkosite, some small isolated plates of arkosite are to be found in the basement as well (Fig. 21). Basal arkosite is the lowest member in the series; it contains local fine-grained carbonate-rich layers (Table 7, No. 15) or tremolite-bearing carbonate lenses. Upwards the basal arkosite grades into a more distinctly layered arkosite.

The granular grains in this Lower arkosite are quartz and feldspars; the rare pebbles are quartz, feldspar or oligoclase granite. The matrix includes fine-grained quartz, feldspars, micas, chlorite and carbonate. Locally sandy and clayey siltstones alternate in the Lower arkosite (Fig. 22). Occasionally the clayey siltstone layers exhibit mudcracks filled with arkositic matter (Fig. 35A—B).

Upwards the Lower arkosite passes into greenish or pinkish cross-bedded Upper arkosite. This rock contains granular quartz and potassium feldspar in a matrix mainly composed of sericite and quartz. Rounded quartz pebbles and cobbles interspersed with rare granite pebbles are encountered. The intensely altered upper part of the Upper arkosite overlain by gently dipping quartzite beds was intersected by diamond drilling (Fig. 21, dh 307). The a.teration zone is 20 to 30 m thick and consists of a pinkish and porous rock whose feldspars are replaced by sericite and to a lesser extent by carbonate and hematite pigment (Fig. 41). The porosity may be attributed to the partial dissolving of the carbonate. This zone seems to be a relic of the ancient weathering crust.

The Lower quartzite Formation

The Lower quartzite Formation begins with a greenish sericite quartzite resting unconformably on arkosites. It has numerous quartz conglomerate interbeds 2 to 20 cm thick (Fig. 45A) and some thin clay slate intercalations (Fig. 45B). Upwards in the sequence the abundance of sericite decreases, the conglomeratic interbeds disappear and the rock passes into pinkish orthoquartzite. Although poorly visible, cross-bedding is common in both quartzite types. Symmetric ripple marks have been encountered in several places in the upper part of the sericite quartzite.

Narrow as they are, the conglomerate interbeds in the sericite quartzite are so regular that it has been possible to trace some of them for as much as 200 m. The elongated pebbles in the conglomerate are from 1 to 3 cm in diameter; they are composed of pale or bluish vein quartz, fine-grained quartz and, rarely, fuchsite-bearing Prekarelidic quartzite. The sandy matrix is often stained red by hematite. Some of the conglomerate beds are slightly radioactive on account of Th-bearing minerals, e.g. brockite and thorite (Räisänen 1975); their abundances, however, are without economic significance.

The Volcanite Formation

The Lower quartzite Formation is overlain by a deposit of volcanites about 50 to 80 m thick (Fig. 21) that dips gently towards the middle of the syncline. This Volcanite Formation is composed of lavas, amygdaloids and, uppermost, tuffs and tuffites. The predominant lava type shows rare plagioclase (oligoclase) phenocrysts in an aphanitic groundmass of plagioclase, hornblende, biotite, chlorite, quartz and magnetite. East of the lake Sääperi a variolitic rock occurs that has abundant varioles 2 to 15 mm in diameter (Fig. 65). The varioles are composed predominantly of albitic plagioclase, with some quartz and chlorite. The fine-grained mass that envelops the varioles is mainly chlorite. Amygdules filled with quartz also occur. Highest in the sequence are heterogeneous chlorite- and quartz-rich rocks that are probably tuffites. Coarse-grained pyroclastic rocks, which Nykänen (1968) has interpreted as agglomerates, occur northwest of Lake Sääperi.

The metadiabase dykes

The arkosite and quartzites in the Sääperi area are cut by numerous metadiabase dykes, which frequently trend NW—SE, or thereabouts. In width they vary from 0.5 to 50 m (Fig. 23). Some of them terminate in the Volcanite Formation but some seem to extend upwards in the sequence to the formations overlying the volcanite. In many places the dykes penetrate the quartzite layers almost perpendicularly. The metadiabases are usually medium-grained rocks composed of somewhat cloudy plagioclase (oligoclase-andesine) and variable amounts of hornblende, biotite, chlorite, quartz, epidote, titanite and magnetite. The narrower dykes are fine-grained rock, but the mineral composition is the same. Some dykes contain a low-grade chalcopyrite and pyrite dissemination.

Fig. 23. A sinuous metadiabase dyke cutting the sericite quartzite. Presumably an apophyse from a wider dyke. Length of compass 12 cm. Sääperi, Värtsilä. Photo by E. Halme.

The caprocks of the Volcanite Formation

As suggested by the topography and the erratic blocks (Fig. 21), the calcareous quartzite, quartz-bearing dolomite, dolomite breccia and dolomite with carbonaceous slate bands probably overlie the Volcanite Formation in the Sääperi area. For want of outcrops, however, it has not been possible to verify this. Hausen (1930) and Hackman (1933) both considered it possible that dolomites occur in this area. Their assumption is corroborated by the presence of dolomite layers not so far away in the Pieni Jänisjärvi area in Soviet Karelia.

South of the lake Sääperi there is a subhorizontal platy metadiabase sill (Fig. 21) that presumably rests on dolomite beds. As a result of folding, the western margin of the plate has tilted steeply westwards. The rock in the sill predominantly medium-grained and gabbroic. The major constituents are plagioclase (andesine), hornblende and biotite, with some small amounts of quartz, chlorite, epidote, titanite, leucoxene and magnetite. Right in the middle of the syncline a coarse-grained and hornblende-rich rock is encountered that largely resembles the Hyypiä sill (p. 25).

On the western margin of the sill the rock has undergone intense alteration and contains abundant chlorite and some epidote, sericite and carbonate. The caprocks proper of the sill are not exposed, although remnants of effusive rocks are encountered in the upper portion of the sill.

In the Sääperi area the outcrops of the gabbroic metadiabase extend westwards almost to the contact with the Mica schist Formation. Owing to the lack of exposures, however, the rock types in the contact zone have not been established.

The sequence of the Karelian formations

In the Sääperi area (Fig. 21), the sequence, from top to bottom, is probably as follows (estimated thickness in brackets):

Mica schist Formation; staurolite-bearing mica schist (Turbidite conglomerate—quartzite Formation?)

---------------------- unconformity ----------------------

Dolomite—Carbonaceous slate—Volcanite Formation and Upper quartzite Formation (350 m?)

— gabbroic metadiabase sill, remnants of effusive rock in the roof (200 m)

— carbonaceous slate, dolomite, quartzite? (150 m)

---------------------- unconformity ----------------------

Volcanite Formation (80 m)

Lower quartzite Formation (400 m); upper portion orthoquartzite, lower portion sericite quartzite with thin quartz conglomerate interbeds and clay slate intercalations

---------------------- unconformity ----------------------

Arkosite Formation (W portion a few metres, E portion 100—150 m); cross-bedded arkosite whose uppermost part is intensely altered, then horizontally layered arkosite, basal arkosite and, locally, arkosic conglomerate

------------------- major unconformity -------------------

The Prekarelidic basement (gneissose grano-quartz diorite, oligoclase granite and granite).

Measured from the depositional basement to the contact with the Mica schist Formation, this sequence is some 750 m thick.

Type section of the Kiihtelysvaara—Värtsilä area

The data presented in the foregoing chapters reveal features that occur in the example areas throughout the Karelian formations. To illustrate this, seven stratigraphic columnar sections have been compiled that show the lithostratigraphic units in the contact zone of the Karelian formations with the Prekarelidic basement between Kiihtelysvaara and Värtsilä (Fig. 24). The column at the extreme left represents the southernmost section and the column at the extreme right but one the northernmost section; the intervening sections are in their appropriate order. To correlate the lithostratigraphic units in the different sections, lines have been drawn between them. A dotted line is used whenever a boundary also denotes a marked unconformity, and a dashed line whenever the boundaries are not well established. The formations vary considerably in thickness from one section to another; certain formations are even totally absent from some of the sections. The dissimilarites are most marked between hte southern main area and the northernmost subarea. The latter presumably lacks not only the volcanites that usually rest on the lower quartzites, but also the upper quartzites, dolomites and carbonaceous slates.

On the basis of the above data, a generalized columnar section was compiled to illustrate the stratigraphy of the Karelian formations in the study area; it is on the extreme right in the diagram (Fig. 24). This generalized columnar section, or the type section of the Karelian Sequence in the Kiihtelysvaara—Värtsilä area, shows ten lithostratigraphic units or formations, which, listed from top to bottom and named in allusion to their predominant rock types, are as follows (the estimated range of thickness in brackets):

	(metres)
— Mica schist Formation	(>1 000)
— Turbidite conglomerate—quartzite Formation	(15 — 55)
— Dolomite—Carbonaceous slate—Volcanite Fm	(N part missing; S part 10 — 350?)
— Hematite rock—Quartzite Formation	(0 — 9)
— Dolomite—Volcanite Formation	(N part missing; S part 5 — 85)
— Upper quartzite Formation	(N part missing; S part 30 — 205)
— Volcanite Formation	(N part missing; S part 50 — 80)
— Lower quartzite Formation	(150 — 550)
— Arkosite Formation	(?10 — 400)
— Breccia—conglomerate Formation	(0 — 40)

The arguments that have lead to this subdivision will be discussed in the next chapters.

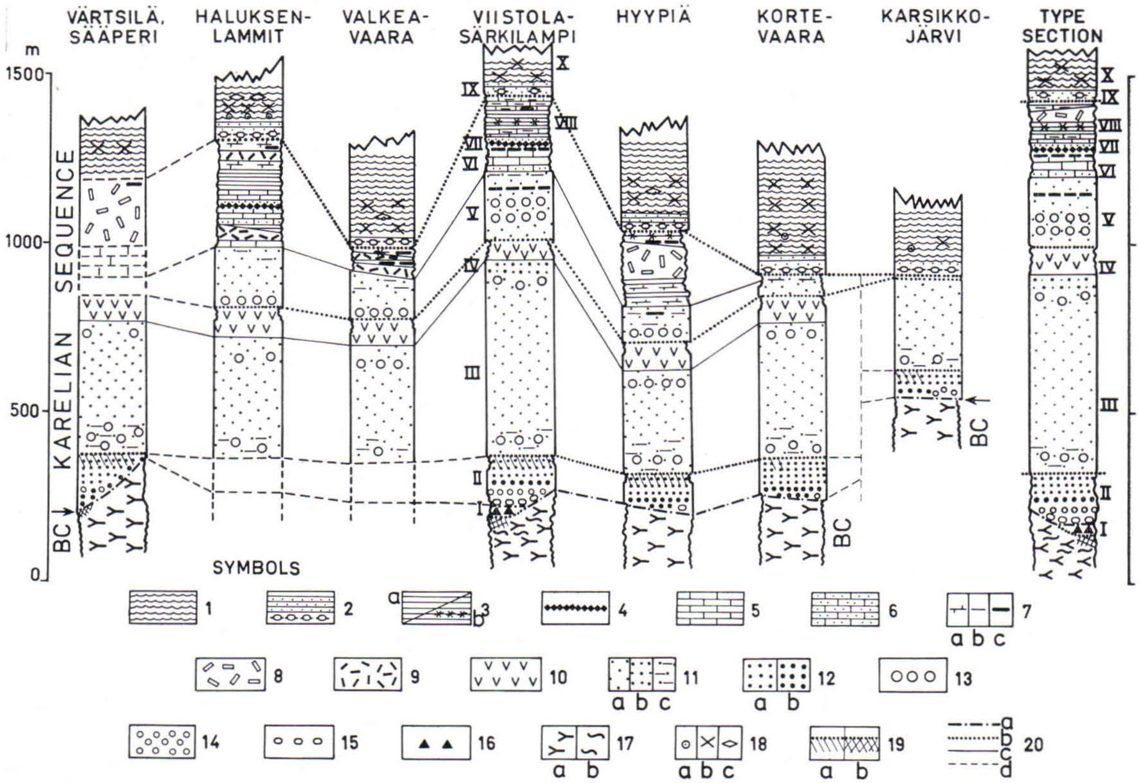

Fig. 24. Stratigraphic columns of the subareas of the Kiihtelysvaara—Värtsilä area and the Type Section. BC = Prekarelidic Basement Complex. Karelian lithostratigraphic units: I = Breccia-conglomerate Formation; II = Arkosite Formation; III = Lower quartzite Formation; IV = Volcanite Formation; V = Upper quartzite Formation; VI = Dolomite—Volcanite Formation; VII = Hematite rock—Quartzite Formation; VIII = Dolomite—Carbonaceous slate—Volcanite Formation; IX = Turbidite conglomerate—quartzite Formation; X = Mica schist Formation. Symbols: 1 = mica schist; 2 = turbidite conglomerate and quartzite; 3 a = carbonaceous slate 3 b = carbon-rich layers; 4 = hematite-rich layer; 5 = dolomite; 6 = quartz-bearing dolomite; 7 a = dolomite intercalations; 7 b = pelitic intercalations; 7 c = volcanic intercalations; 8 = gabbroic metadiabase; 9 = metadiabase; 10 = metavolcanite; 11 a = orthoquartzite; 11 b = feldspathic quartzite; 11 c = sericite quartzite; 12 a = upper arkosite; 12 b = lower arkosite; 13 = conglomerate interbeds; 14 = small-pebbled conglomerate; 15 = big-cobbled conglomerate; 16 = breccia-conglomerate; 17 a = Prekarelidic plutonic rocks; 17 b = Prekarelidic schists; 18 a, b, c = almandine, staurolite, or andalusite porphyroblasts; 19 a = younger (Early Jatulian) weathering crust; 19 b = older (Early Karelian) weathering crust; 20 a = major unconformity; 20 b = unconformity; 20 c = contact of formation; 20 d = inferred.

THE PREKARELIDIC BASEMENT COMPLEX AS THE DEPOSITIONAL BASEMENT OF THE KARELIAN SEDIMENTS

Numerous field observations indicate that the lowest units in the Karelian Sequence were deposited unconformably on the Prekarelidic basement. Many factors suggest that they consist of residual sediments of varying degrees of maturity derived from Prekarelidic rocks.

Characteristics of the rocks of the Prekarelidic basement complex

The following describes the distribution, mode of occurrence, mineral and chemical compositions and age relations of the rocks of the Prekarelidic basement complex. The aim is to specify the depositional basement and primary source of the Karelian sedimentary rocks.

The frequency of the rock types

The frequency of the various rock types in the Prekarelidic basement in North Karelia was determined planimetrically for an area of about 2 300 km². The determinations were based on four 1 : 100 000 geological maps, i.e. the map sheets of Tohmajärvi (Nykänen 1967), Kiihtelysvaara (Nykänen 1971a), Ilomantsi (Lavikainen 1973) and Oskajärvi (Lavikainen 1975). Only the main rock types were counted; hence, the minor gabbro bodies and ultramafics were omitted, as were the Karelian metadiabase dykes. The Prekarelidic basement consists largely of the following rock types:

— granites	7.5 %
— oligoclase granite (leucogranite, trondhjemite)	3.0 %
— granodiorites and quartz diorites	74.8 %
— leptites	0.2 %
— mica schists and mica gneisses	8.7 %
— hornblende gneisses and amphibolites	5.8 %
	100.0 %

Prekarelidic plutonic rocks

The plutonites of the Prekarelidic basement play a key role in the identification of source material. Mafic plutonites, such as ultramafics and gabbros, are, however, comparatively rare (Nykänen 1971b). The predominant plutonite in the basement is a cataclastic, gneissose granitoid that varies in composition from quartz diorite to granodiorite and further to oligoclase granite (trondhjemite). Its major minerals are plagioclase (oligoclase-andesine) in short laths, quartz, biotite and hornblende (Table 1, Nos. 1 to 3). Minor potassium feldspar and accessories also occur. Tourmaline granite is notably more rare but easily recognisable; it is often plagioclase-predominant (Table 1, No. 4), but also includes portions rich in potassium feldspar. The granites in the basement are pinkish and often fairly coarse-grained rocks in which the abundances of plagioclase (albite-oligoclase), cross-hatched potassium feldspar and quartz are roughly equal. They also contain small amounts of micas and accessories (Table 1, No. 5). Some examples of the chemical compositions of the plutonites are given in Table 2 (Nos. 1 to 4).

Prekarelidic supracrustal rocks

The schist association of the Prekarelidic basement consists of amphibolites, hornblende gneisses, rocks of iron formation, mica schists, mica gneisses (Fig. 25). Leptites, arkose gneisses and some less frequent rock types (Nykänen 1968, 1971b; Lavikainen 1973, 1975, 1977 and Niiniskorpi 1975). Examples of the mineral compositions of these rocks are given in Table 1 (Nos. 6 to 8) and of their chemical compositions in Table 2 (Nos. 5 to 7).

Pegmatite and quartz dykes and veins

The pegmatite and quartz dykes and veins and small intrusions in the Prekarelidic basement are another significant source of coarser clastic material such as pebbles in conglomerates and granular grains in arkosites. On the basis of the mode of occurrence and age relations these rocks can be subdivided roughly into three groups.

Table 1

Mineral compositions of various rocks in the Prekarelidic basement complex. Determined by point-counting method.

Minerals	1	2	3	4	5	6	7	8
Quartz	20.1	24.8	26.6	48.2	28.7	+	18.0	22.6
Plagioclase	54.6	50.9	61.3	46.8	33.6	55.5	21.0	47.3
(An)	(15—32)	(15—30)	(14—22)	(18—22)	(10—20)	(24—30)	[b]	(22)
Potassium feldspar ..	2.9	10.3	6.4	+	31.0		+	
Hornblende	3.4	1.3				33.8	40.3	
Biotite	14.2	8.5	2.0	0.5	2.7	4.2	0.6	27.8
Sericite[a])	0.2	1.2	2.0	2.7	2.2	+	+	2.3
Chlorite	0.5	0.9	1.3	+	1.4	0.2	1.2	+
Carbonate	+	0.1			+	+		
Apatite	0.3	0.2	0.1		0.1	+		+
Titanite	0.6	0.5	0.1		0.2	2.1	0.5	
Zircon	+	+	+	+	+	+		+
Tourmaline			+	1.8				
Epidote	3.1	1.1	0.2		0.1	1.2	18.4	
Opaque	0.1	0.2	+		+	3.0	+	+
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

[a]) partly derived by decomposition of feldspar
[b]) exact determination not possible
\+ detected

1. Average of 13 quartz diorites
2. Average of 13 granodiorites
3. Average of 9 oligoclase granites (trondhjemites)
4. Tourmaline granite, Särkilampi, Kiihtelysvaara
5. Average of 14 granites
6. Amphibolite, Särkilampi, Kiihtelysvaara
7. Hornblende gneiss, Särkilampi, Kiihtelysvaara
8. Mica gneiss, Särkilampi, Kiihtelysvaara

a) Most of them, i.e. the intensely deformed and folded types, belong to the same group as the Prekarelidic granitoids. These occurrences are characterized by pale, greyish or slightly pinkish quartz that exhibits rather marked undulatory extinction in thin section.

b) Occurrences that differ conspicuously from those just described have been met with in association with some fracture zones trending N 10°—15°W. One of these zones passes through Mustalampi—Eronlampi, where abundant irregular quartz-rich veins and nodules are encountered in several places in the transition zone between amphibolite and mica schist (Fig. 25); typical of these is bluish or greyish quartz that in thin section is either slightly or not at all undulatory (The Mustalampi type). Here and there the rock exhibits large potassium feldspar phenocrysts due to which it resembles augen gneiss. In places the quartz veins and nodules are enveloped by a mica-rich seam that contains uranium- and thorium-bearing minerals, such as brockite and thorite, and also apatite and molybdenite. These veins and bodies give the impression that they were formed metasomatically from primary rock through replacement and impregnation. The age of the uraninite, 2340 Ma (Fig. 28), indicates that these occurrences are younger than the rocks of the Prekarelidic basement proper (Fig. 26).

c) Even in the basement there is an abundance of quartz veins crosscutting all the other rocks with sharp contacts. They do not, however, belong to the same group as the above; the quartz in these veins is clear or pale and, in thin section, slightly undulatory. The galena encountered in them occasionally gives a lead age 1800 to 1900 Ma (Kouvo personal communication). It is noteworthy that similar and contemporaneous quartz veins occur farther west, where they cut Karelian formations, e.g. mica schists (p. 42), and are thus younger than the formations in question.

Table 2
Chemical composition of various rocks in the Prekarelidic basement complex (weight per cent).

	1	2	3	4	5	6	7	8	9
SiO_2	67.70	71.81	74.58	73.74	51.20	69.8	72.42	69.1	70.63
TiO_2	0.45	0.17	0.53	0.26	0.97	0.5	0.30	0.4	0.24
Al_2O_3	15.00	15.35	13.04	14.20	12.77	12.6	14.55	14.7	15.12
Fe_2O_3	1.54	0.87	0.46	0.43	3.25	} 3.4 a)	0.61	1.4	1.47
FeO	2.00	0.49	0.95	0.89	7.99		1.05	1.7	1.17
MnO	0.07	0.03	0.04	0.01	0.33		0.05	0.1	0.04
MgO	2.48	0.48	2.09	0.69	7.07	1.9	1.76	1.8	0.83
CaO	2.30	1.22	0.97	0.25	10.85	0.3	1.81	2.0	1.56
Na_2O	4.90	4.42	5.34	3.76	2.90	3.1	2.90	4.4	3.83
K_2O	2.13	3.88	0.88	5.08	0.60	3.7	3.20	3.0	3.45
P_2O_5	0.14	0.03	0.12	0.06	0.16		0.13	0.1	0.01
CO_2	0.00	0.60	0.00	0.00	0.00		0.00		
H_2O^+	0.68	1.02	1.26	0.56	1.74	} 2.0 b)	0.95	1.0 c)	0.91
H_2O^-	0.05	0.04	0.04	0.04	0.14		0.09		0.23
S	nd	nd	nd	nd	0.04		0.00		
Total	99.44	100.41	100.30	99.97	100.01		99.82	99.7	99.49
	XRF			XRF	XRF		XRF		
Rb_2O	0.01			0.02	0.008		0.010		
SrO	0.06			0.01	0.015		0.012		
ZrO_2	0.04			0.05	0.00		0.012		
BaO	0.04			0.05	0.02		0.05		

a) total iron; b) estimated; c) total water; nd = not determined

1. Gneissose quartz diorite, Uskali, Kiihtelysvaara. Analyst: A. Heikkinen; XRF analyses by V. Hoffrén (Nykänen 1968, p. 11).
2. Gneissose granodiorite, Hyypiä, Kiihtelysvaara. Analyst: P. Ojanperä (Nykänen 1971b, p. 19).
3. Slightly weathered oligoclase granite (leucogranodiorite) sample taken 3 m below the unconformity, Hyypiä, Kiihtelysvaara. Analyst: P. Ojanperä (Nykänen 1971b, p. 35).
4. Granite, Kutsu, Tohmajärvi. Analyst: A. Heikkinen; XRF analysis by V. Hoffrén (Nykänen 1968, p. 11).
5. Stratified hornblende schist (tuffite), Keskijärvi, Kiihtelysvaara. Analyst: P. Ojanperä; XRF analyses by V. Hoffrén (Nykänen 1971b, p. 19).
6. Mica gneiss, Särkilampi, Kiihtelysvaara. Semiquantitative multichannel emission spectrometer analysis by G. Wansén. Iron determination by A. Nurmi.
7. Graded-bedded quartz feldspar schist, Linnasuo, Tuupovaara. Analyst: P. Ojanperä, XRF analyses by V. Hoffrén (Nykänen 1971b, p. 19).
8. Rough weighted average of Prekarelidic basement rocks.
9. Average of 173 Prekarelidic gneissose granites in Soviet Karelia (W. Z. Negrutsa 1974).

The average chemical composition of the Prekarelidic rocks

For use in a later chapter (p. 78), an approximate mean chemical composition was calculated for the rocks of the Prekarelidic basement. The calculation was based on the chemical data listed in Table 2 weighted by the relative frequencies of the rocks given on page 50 (the frequencies of the grano- and quartz diorites were assumed to be equal). The result, the rough weighted mean (Table 2, No. 8), suggests a granodioritic composition close to the average composition of 173 gneiss granites from Soviet Karelia (Table 2, No. 9) reported by Negrutsa (1974).

The average abundances of copper, nickel, zinc, lead, cobalt, phosphorus and sulphur in Prekarelidic rocks are given in a later chapter (Fig. 43).

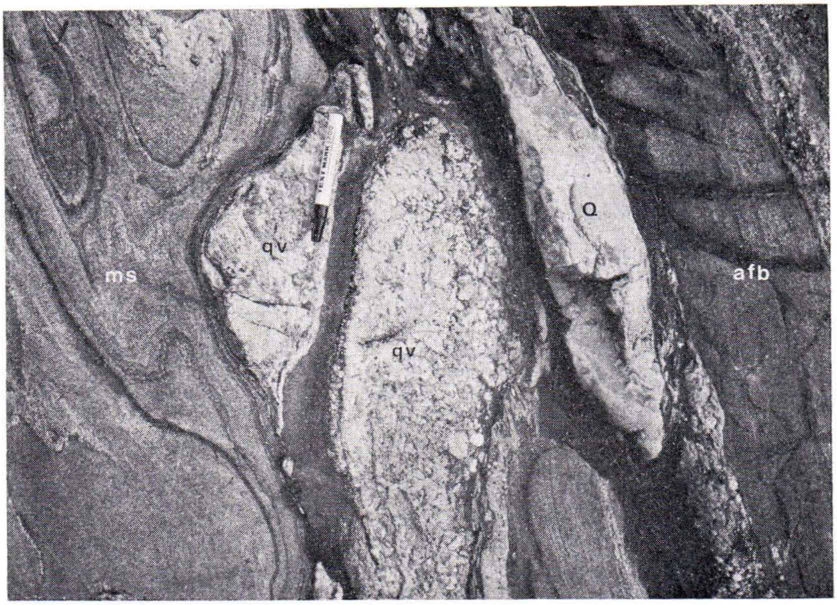

Fig. 25. Quartz-rich veins (qv) whose margins are rich in mica and uraninite in the transitional zone between the Prekarelidic amphibolite (afb) and mica schists (ms). A younger fissure filled with quartz (Q) slightly to right of centre. Length of pen 14 cm. Mustalampi, Kiihtelysvaara.

The radiometric age of the Prekarelidic rocks

The Prekarelidic plutonites

The ages of some plutonites in North Karelia have been determined. The U and Pb of the zircons and one titanite from some representative samples have been plotted in a diagram (Fig. 26), which shows that these rocks belong to the age group of 2765 Ma. The Rb-Sr age of the muscovite in the tourmaline granite that cuts the Prekarelidic schists and quartz diorite at Särkilampi is as high as 2630 Ma (Kouvo and Tilton 1966); hence this granite is indisputably Prekarelidic. It is noteworthy that both K/Ar and Rb/Sr methods show that the biotites of the samples shown in the diagram are significantly younger, i.e. 1750 to 2000 Ma (Kouvo and Tilton 1966; Nykänen 1968). It is plausible, however, that this age merely reflects the age of regional metamorphism, about 1800 Ma ago (Kouvo and Tilton 1966). The Prekarelidic ages given by the zircons in the basement rocks in North Karelia (2600 to 2800 Ma) and the younger ages of their biotites reveal that the basement was only slightly metamorphosed during the Svecokarelidic orogeny.

The Prekarelidic supracrustal rocks

Cutting relations show that the Prekarelidic supracrustal schists are older than the plutonites. Datings are still rare and subject to controversy; nevertheless, zircon dating indicates that some volcanites in the Ilomantsi schist zone are markedly older than the plutonites (Geological Survey of Finland 1973, 1974; Lavikainen 1977). On the basis of the zircon age of an orthogneiss clast from the conglomerate in the lower schist association of Ilomantsi, Lavikainen (1977) has estimated that these schists are younger than 2900 Ma.

The 2340 Ma age group

Because contact relations suggest that they are younger than the Prekarelidic basement proper, special attention was directed to the

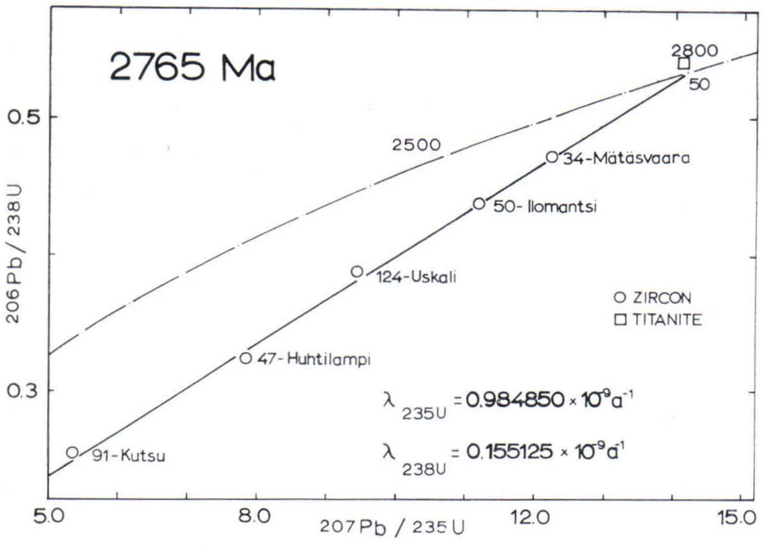

Fig. 26. Summary plot of zircon and titanite U-Pb analyses showing an age of 2765 Ma for the Prekarelidic basement complex in North Karelia (ref. 34, 47, Wetherill *et al.* 1962; 50, Kouvo and Tilton 1966; Tilton and Grünefelder 1968; 91, 124, Nykänen 1968).

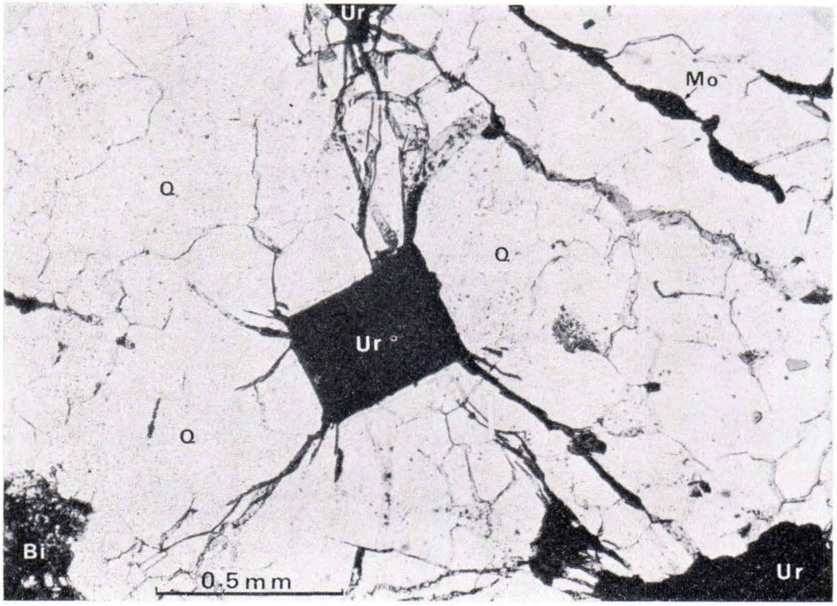

Fig. 27. Microphotograph of uraninite cubes in a quartz vein with a few molybdenite and biotite flakes. Notice microfractures in quartz around uraninite crystals. Q = quartz, Ur = uraninite, Mo = molybdenite, Bi = biotite. Thin section No. Ku 00633. Crossed nicols. Photo by E. Halme.

U-Th-bearing quartz veins and bodies that, in this study area, are controlled by fracture-displacement lineaments in the Prekarelidic basement. The veins were dated with the aid of the uraninite in the quartz veins of Mustalampi type (Fig. 27). The age obtained, 2340 Ma, showed them to be younger than the Prekarelidic basement proper (Table 3 and Fig. 28).

The significance of the age obtained, 2340 Ma, is corroborated by other known ages of the same age group: zircon ages, 2300 to 2600, Ma of the mantled domes (Kouvo and Tilton 1966); zircon age, 2360 Ma, of the potassium granite in the Lappish granulite complex (Meriläinen 1976); and zircon age, 2300 Ma, of the detrital zircons in the lower greywacke beds

Table 3

U-Pb analytical data for Mustalampi uraninite. Isotope analyses by O. Kouvo at the Geological Survey of Finland.

Sample	U content (in %)	Radiogenic Pb (in %) 206	207	208	Apparent ages in Ma $^{207}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$
A677—GSF 76	36.66	12.78	1.90	0.9387	2 322 ±15	2 255 ±28	2 183 ±53

$\lambda 238 = 0.155125 \times 10^{-9} a^{-1}$,
$\lambda 235 = 0.984850 \times 10^{-9} a^{-1}$,
$^{238}u/^{235}u = 137.88$

A 677—GSF 76. — Uraninite from an epigenetic U-Th mineralization in a migmatitic Prekarelidic mica gneiss. Sample taken from about 0.5 km southwest of Mustalampi, Kiihtelysvaara (Thin section No. Ku 00633, Appendix 1). The mineralization occurs in a fault and shear zone trending 345°, which has been traced for some 30 km. Here and there in the silicified sample there are potassium feldspar porphyroblasts among the bluish quartz; also present are uraninite, thorite, apatite, molybdenite, magnetite and galena in association with biotite scales (Anal. U = 1.8 %, Th = 0.45 %, Mo = 0.36 %, Pb = 0.07 %, Zn = 0.04 %, S = 0.62 %). Uraninite occurs mainly as subeuhedral cubic crystals (edge 0.2 — 1.2 mm) often surrounded by molybdenite scales.

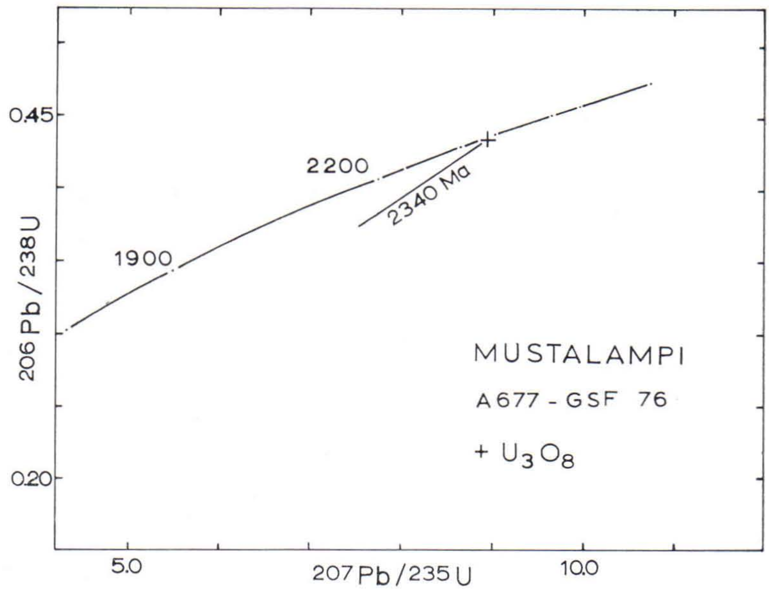

Fig. 28. Concordia plot for uraninite from Mustalampi, Kiihtelysvaara.

in the Tampere area (Kouvo and Tilton 1966). It is not known whether there is any connection between these ages, or whether they are merely a coincidence; what is important, though, is the indications in the gneiss domes and mantled gneiss domes of palingenetic recrystallization or granitization (Eskola 1949; Härme 1949; Preston 1954). Eskola believed that this was due to metasomatic granitization and palingenesis associated with the rising of the domes, possibly during the Svecokarelidic orogeny about 1800 Ma ago. It is possible, however, that there is some connection between the formation of a fracture-fault system, the onset of dome formation, and granitization. There is every reason to suppose that at that time, about 2340 Ma ago, the Prekarelidic basement underwent restricted granitization (cf. Lobach-Zhuchenko *et al.* 1974, pp. 48 and 186). Further, block movements may have triggered the process that led to the development of sedimentation basins.

The older or Early Karelian weathering crust and breccia-conglomerate

Few geological documents have been preserved to indicate what happened during the denudation of the mountain ranges that rose during Prekarelidic time, which, to accentuate its importance, could be called the Early Karelian period. This may be because the weathering products then formed were often reworked, and relics of the weathering crusts, breccias and gravel buried under the younger sediments were only rarely preserved. In addition to the targets now to be described, observations have also been conducted on satrolites, weathering crusts, weathering breccias and breccia-conglomerates outside the present study area, e.g. in eastern Finland (Sederholm 1931, p. 77; Ojakangas 1965, p. 65; Piirainen 1968, pp. 17—18; Silvennoinen 1972, pp. 12—14; Gaál, *et. al.* 1975, pp. 28—30), and in Soviet Karelia (Metzger 1924; Sokolov *et al.* 1970; Sokolov *et al.* 1972). It is, however, difficult to compare the descriptions because of the ambiguity in the use of such terms as »satrolite».

Observations on the occurrence of the Early Karelian weathering crust and breccia-conglomerate in the study area

Intensely eroded weathering crust

Towards the contact with the Karelian formations the rocks of the Prekarelidic basement often show changes in mineral composition and an increase in cataclastic structure. Cataclastic structure is best developed in the plutonites, which are broken up into feldspar-quartz aggregates, 1 to 5 mm in diameter, embedded in finely ground quartz, feldspar and micas. The changes in mineral composition are mainly due to the increase in sericite in the fines. Often, however, even in the immediate vicinity of the contact with the Karelian formations, the rocks are only mildly altered, possibly because the weathering crust covering the surface was almost completely eroded during the initial stages of Karelian sedimentation.

Remnants of weathering breccias

In some places below the contact, e.g. under the Sääperi basal arkosite (Figs. 29A—B), there are occurrences a few metres thick of somewhat broken and altered, breccia-like rock. The alteration was most intense along the boundaries of the blocks; the alteration products are sericite, carbonate and chlorite (Table 4). This is, however, more likely to be a somewhat eroded section than the very top of the ancient weathering crust.

It was not always easy to delineate the contact

Table 4

Examples of mineral compositions of Prekarelidic plutonic rocks in the proximity of the major unconformity. Determined by point-counting method.

Minerals	9	10	11
Quartz	48.6	34.6	42.3
Plagioclase	21.2	33.6	37.3
(An)	b)	(20)	(23)
Microcline	14.3	+	+
Biotite	1.0	2.3	1.1
Sericite a)	10.6	24.9	18.6
Chlorite	+	1.1	0.5
Carbonate	4.3	1.7	+
Apatite	+	+	+
Titanite	+		
Zircon	+	+	+
Opaque	+	+	+
	100.0 c)	100.0 c)	100.0 c)

\+ detected minor constituent
a) mainly derived by decomposition of feldspar
b) exact determination not possible
c) after addition of all minor constituents

9. Slightly weathered Prekarelidic granodiorite, about 2 m below the major unconformity (Specimen No. R306/4241/149.45), Hyypiä, Kiihtelysvaara.
10. Slightly weathered Prekarelidic oligoclase granite (trondhjemite), about 3 m below the major unconformity (Specimen No. 84/EOR—74), Sääperi, Värtsilä.
11. Slightly weathered Prekarelidic oligoclase granite (trondhjemite), about 1 m below the major unconformity (Specimen No. 22B/EOR—72), Hyypiä, Kiihtelysvaara.

Fig. 29. A) Unconformity of Prejatulian basal arkosite above slightly weathered Prekarelidic oligoclase granite. Notice the isolated angular fragments of oligoclase granite near the base of the basal arkosite. Length of pen 14 cm. Kukkolampi. Värtsilä.

B) Prekarelidic oligoclase granite showing weakly developed spherical weathering in a well-jointed rock just below the unconformity. Length of pen 14 cm. Kukkolampi, Värtsilä. Photo by E. Räisänen.

of the Prekarelidic basement with the Karelian formations, particularly when the rock in the depositional basement is of Prekarelidic oligoclase granite overlain by slightly recrystallized basal arkosite. Thus, in the opinion of the present author, the »sapropelitic (or sathrolithic)» basal breccia schists described from the Sääperi area by Nykänen (1968, pp. 19 and 20) are in fact weakly layered arkosites.

The weathering crust and Breccia-conglomerate Formation in the Särkilampi area

Only few remnants have survived of the surface layer of the mantle of rock-waste formed in situ as a result of weathering. A crust of intensely altered rock 3 to 5 m thick that was encountered in a drill hole below the Särkilampi breccia-conglomerate may be considered as a remnant of the sathrolith (Fig. 30; Table 5).

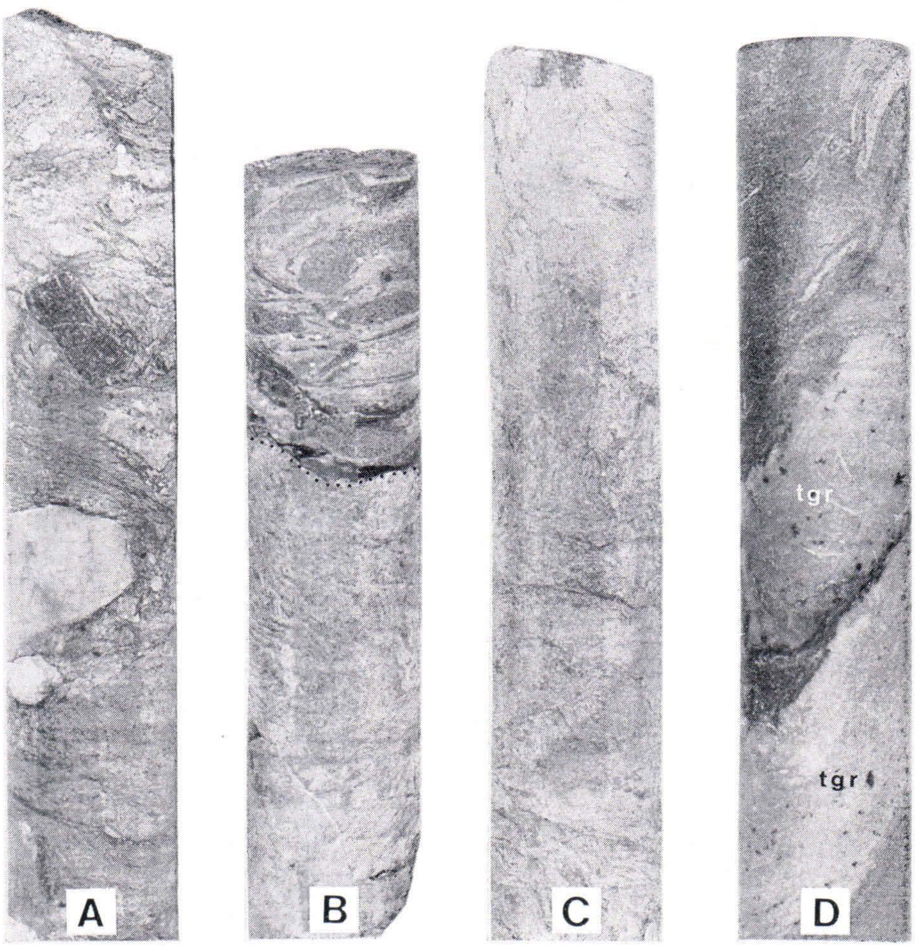

Fig. 30. Drill core samples from various depths in the older (Early Karelian) weathering crust below the Särkilampi breccia-conglomerate. A = breccia-conglomerate some 2 m above the contact; B = a contact between the breccia-conglomerate and the weathering crust (dotted line); C = the weathering crust about 1 m below the contact; D = a fairly fresh biotite-chlorite schist and a crosscutting tourmaline granite vein (tgr) some 9 m below the contact. Diameter of cores about 32 mm. Särkilampi, Kiihtelysvaara. Photo by E. Halme.

On the basis of structure and variation in mineral composition, the following zones (from top to bottom) have been established (Fig. 31):

Zone A: the breccia-conglomerate above the unconformity; distinctly separate, angular fragments; degree of weathering as in zones C and D; fines between fragments rich in carbonates; heavy minerals, such as tourmaline and magnetite, enriched in lowest part, and even a small gold nugget has been found near the contact.

Zone B: contact usually graded (graded unconformity).

Zone C: from 0 to 1.5 m downwards from the contact; rock broken into very small pieces; plagioclase and biotite replaced by sericite, chlorite and carbonate;

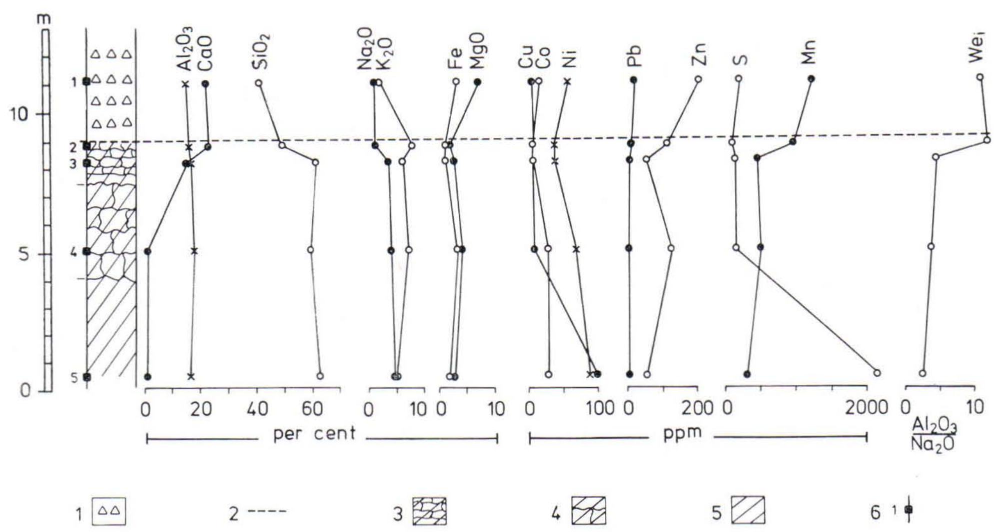

Fig. 31. The variation in chemical composition at various depths in the older (Early Karelian) weathering crust (Table 5) below the Särkilampi breccia-conglomerate. 1 = A zone (breccia-conglomerate); 2 = B graded unconformity; 3 = C zone; 4 = D zone; 5 = E zone; 6 = analysed sample and its number.

Table 5

Variation in the chemical composition of the older (Early Karelian) weathering crust at various depths below the Särkilampi breccia-conglomerate. SiO_2, Al_2O_3, MgO, CaO, Na_2O and K_2O were analysed by emission spectroscopy using an ARL 31 000 Quantometer (by G. Wansèn). Total iron, Mn, Cu, Co, Ni, Pb and Zn were determined by AAS, and sulphur by a Leco M621 sulphur analyser (A. Nurmi). The sample numbers (1—5) and the symbols for the zones are the same as in Fig. 31.

Specimen No.	Zone	per cent							ppm						
		$SiO_2$¹)	$Al_2O_3$¹)	Fe	MgO¹)	CaO¹)	Na_2O¹)	K_2O¹)	Mn	Cu	Co	Ni	Pb	Zn	S
1	A	41.4	15.4	3.1	7.1	22.7	1.4	2.2	1 260	10	20	60	30	210	220
2	B	49.3	16.8	1.6	2.2	23.6	1.4	8.1	990	10	10	40	20	120	120
3	C	61.4	17.2	1.5	2.9	15.7	3.6	6.1	460	10	10	40	10	60	160
4	D	59.5	17.3	3.3	4.2	1.8	4.3	7.3	520	10	30	70	10	130	160
5	E	63.0	16.2	2.5	2.9	1.0	5.1	5.4	310	100	30	90	10	60	2 160

¹) semiquantitative

sericite and chlorite scales show secondary orientation; tourmaline grains heavily fractured.

Zone D: from 1.5 to 5 m downwards from the contact; rock intensely broken; plagioclase cloudy owing to alteration products, abundant sericite in margins and cleavages of grains; minor carbonate between grains; tourmaline grains show cleavage.

Zone E: over 5 m downwards from the contact; rock fractured but only slightly altered consisting of biotite, quartz, plagioclase, chlorite and tourmaline.

In composition the deposit overlying the weathering crust resembles weathering gravel. The present author has called it breccia-conglomerate since its fragments are fairly angular and unlike those in arkositic conglomerates (Table 6 and Fig. 32). The deposit often

Table 6

Compositional characteristics of various Karelian basal conglomerates in the Kiihtelysvaara—Värtsilä area.

Characteristics	Breccia-conglomerate	Big-cobbled arkosic conglomerate	Small-pebbled arkosic conglomerate
Gravel-sized fragments of Prekarelidic			
— quartz-granodiorite	locally common	common	common
— oligoclase granite (trondhjemite)	locally common	abundant	abundant
— granite	locally common	common	common
— basic volcanics	not observed	rare	not observed
— amphibolite and hornblende gneiss	locally common and intensely altered	rare	not observed
— mica schist and gneiss	locally common	rare	rare
— leptite	rare	rare	not observed
— vein quartz and fine-grained quartz	present	present	common
— feldspars	present	present	common
— tourmaline	locally common	present	present
Intensely weathered (sericitized) fragments	abundant	present	present
Mustalampi-type vein quartz	not observed	not observed	present
Matrix	rock and mineral debris of all sized, rich in sericite and carbonate	arkosic, commonly rich in sericite and carbonate	arkosic, commonly rich in sericite and carbonate
Typical heavy minerals	tourmaline, magnetite, zircon, titanite	tourmaline, zircon	tourmaline, zircon

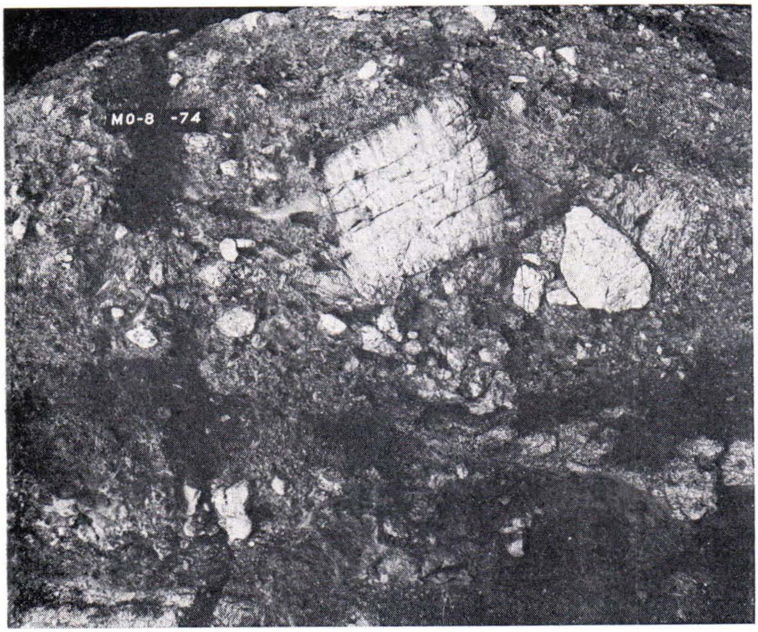

Fig. 32. Särkilampi breccia-conglomerate with angular fragments of Prekarelidic rocks. Length of tag 12 cm. Särkilampi, Kiihtelysvaara. Photo by E. Halme.

seems to rest on its basement through a graded unconformity (Fig. 12) although in places the contact is fairly sharp (Fig. 4). Thus, the material appears to be local and is clearly derived from the underlying bedrock.

Although they differ from each other in chemical composition, the zones of the weathering crust (Table 5 and Fig. 31) show some degree of regularity. Comparison of the surface layers with the deeper zones shows that the abundance of Al_2O_3 in the rocks varies only slightly, which is consistent with the general behaviour of Al_2O_3. The most conspicuous loss is in the abundances of Na_2O and SiO_2; a somewhat less marked loss is apparent in the contents of total iron and MgO. The most distinct gain is in the CaO content, followed by that in K_2O and Mn. Of the minor constituents, Cu, Co, Ni and S are depleted, and Pb and Zn enriched. The Al_2O_3/Na_2O ratio, which is used here as a weathering index, indicates that the change was most marked in the uppermost part of the weathering zone. In chemical composition the breccia-conglomerate resembles the uppermost zones of the weathering crust, although its total iron, MgO and trace element abundances are slightly higher (Table 5, and Fig. 31).

Conclusions concerning the conditions of formation of the Early Karelian weathering crust and gravel

To begin with, there is a conspicuous abundance of sericite in the weathering crust that covered the Prekarelidic basement. Nevertheless, sericite itself is not a primary alteration product, but clearly an authigenic mineral. It also seems obvious that the primary constituents of the sericite were derived from feldspars that underwent disintegration during kaolin-type weathering, although it cannot be established for sure whether the weathering product really was kaolinite or some other clay mineral; it is possible that the potassium necessary for the formation of sericite was initially absorbed in clay minerals. The occurrence of kaolin-type weathering demonstrates that the paleoclimate was not wholly dry.

Another conspicuous feature is the high abundance of carbonate, particularly in the uppermost zone of the weathering crust and in breccia-conglomerate. Its occurrence as both matrix and isolated aggregates and lenses refers to calcium-rich caliche deposits, which are formed in certain semiarid or arid regions (Allen 1965).

The internal structure of the breccia-conglomerate is poorly developed. Nevertheless, some broken mudstone lenses with mudcracks indicate a certain variation in the rate of deposition.

Thus the conditions under which the Early Karelian weathering crust and the closely related breccia-conglomerate were formed on the Prekarelidic basement can be characterized as follows:

- Relics of the weathering crust, weathering breccias and breccia-conglomerates have survived as documents of the early evolutionary history of the Karelian formations.
- The physical disintegration of the rocks was intense, perhaps because of substantial variation in temperature and rather high relief.
- The chemical weathering was incomplete, which led to the formation of a residue rich in quartz but with significant amounts of clay minerals.
- The weathering of feldspars and the enrichment in the upper part of the weathering crust in the carbonates that formed during weathering suggest a semi-arid rather than an arid paleoclimate.
- The breccia-conglomerate probably derives from weathering gravel that rolled down into ravines and down to the bases of hills (talus); its origin can probably be synchronized with that of the weathering crust.

THE KIIHTELYSVAARA ARKOSIC SUITE

The denudation of the Prekarelidic mountain ranges during the Middle Precambrian period was probably a protracted and complex event. In its early stages it was presumably accompanied by the formation of graben and horsts (p. 55) trending roughly NNW. The resulting changes in relief accelerated weathering, erosion and sedimentation. In the Kiihtelysvaara—Värtsilä area the lower division of the Karelian formations consists of a sequence of slightly metamorphosed arkoses and arkosic conglomerates (the Kiihtelysvaara Arkosic Suite).

The associations and their mode of occurrence

In the present context the Kiihtelysvaara Arkosic Suite, which contains arkoses and conglomerates, is called the Arkosite Formation; the lowest stratigraphic unit is a mildly reworked arkose, which is known as Lower arkosite (the Lower arkosite Member) owing to the slight metamorphism. The lowermost arkosite blanket, which rests on the Prekarelidic basement with an angular unconformity in between, is called basal arkosite (Fig. 29A). In some areas, e.g. Sääperi, minor lenses of conglomerate with a few pebbles are associated with the basal arkosite. On the northern flank of the Sääperi syncline the arkositic layers alternate with narrow clay and siltstone layers (Fig. 22). The occurrences of Lower arkosite consist of lens- or wedge-like bodies, laterally connected with each other, whose thickness varies from a few metres to several tens of metres.

Here and there the basal arkosite passes laterally into a coarser arkosic conglomerate (the Arkosic conglomerate Member). The conglomerate occurrences are laterally restricted bodies, lenses or wedges in shape, which obviously fill the depressions in the ancient Prekarelidic basement. They are from a few metres to over 150 m thick. Their basal parts are occupied by coarse, big-cobbled conglomerates, which upwards grade into small-pebbled conglomerates (the Särkilampi conglomerate) or alternate with them (the Viesimo conglomerate). A feature peculiar to the Särkilampi conglomerate is that it rests partly on a breccia-conglomerate and partly directly on the Prekarelidic basement with an erosion surface as an unconformity in between (Fig. 10). The erratic blocks suggest that the same may be true for the Viesimo conglomerate. Upwards the conglomerate beds grade into mildly reworked Lower arkosite.

The Lower arkosite is followed upwards in the sequence by the more intensely reworked Upper arkosite (the Upper arkosite Member). This is composed of laterally overlapping lens- and sheet-shaped forms deposited on the Lower arkosite. The thickness of the occurrences varies from a few metres to almost 100 m. The Upper arkosite is overlain by the Lower quartzite Formation and separated from it by an unconformity denoted by weathering crust and erosion surface.

The total thickness of the Arkosite Formation varies considerably from one area to the next, depending on the thicknesses of the primary beds and on how deeply the layers were eroded at each site before the quartz sands (Lower quartzite Formation) deposited. The Arkosite Formation seems to be thickest (170 to 400 m) in the area between Hyypiä and Viesimo, from where northwestwards and southeastwards it pinches out conspicuously. On the eastern and northern flank of the Sääperi syncline it is some hundred metres thick but it becomes rapidly thinner westwards until at the anticlinal ridge it is only a few metres thick.

Primary sedimentary structures and textures

Bedding

Big-cobbled conglomerate

The coarsest variants of the Arkosic conglomerate Member (Fig. 36), called big-cobbled conglomerates in the field, are crudely bedded and exhibit large-scale horizontal or inclined bedding. In larger occurrences, as in Viesimo, the big-cobbled conglomerate beds alternate with beds of a less coarse variant.

Small-pebbled conglomerate

The finer-grained variants of the arkosic conglomerate, or small-pebbled conglomerates, are also horizontally bedded. In the Särkilampi area the bedding in the small-pebbled conglomerate is somewhat reminiscent of graded-bedding, because each phase consists of a coarser and thicker conglomeratic lower part and a thinner and finer-grained clay-siltstone upper part (Fig. 33). Measurements on drill cores indicate that the thickness of the beds fluctuates from 0.2 to 1.2 m (128 beds). The variation in bed thickness is consistent with lognormal distribution (Fig. 34), which is due to the cyclic arrival of material at the depositional site.

Lower arkosite

In the lower part of the Lower arkosite Member, that is, in the basal arkosite (Fig. 29A), the bedding is poorly developed, but upwards the rock is succeeded by a distinctly bedded arkosite. This Lower arkosite is often horizontally bedded (the Särkilampi and Sääperi areas). In the Lower arkosite in the Sääperi area the thicker sandy layers alternate with thinner argillaceous silt layers (Fig. 22). The

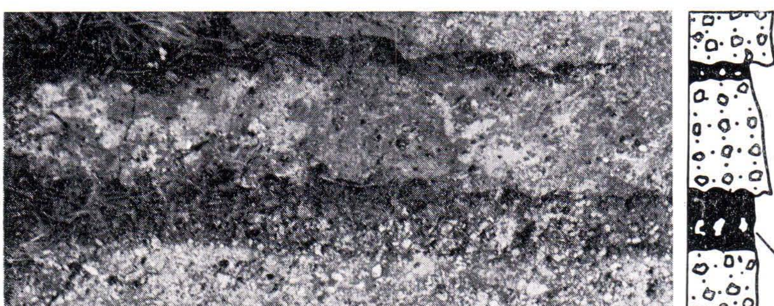

Fig. 33. Small-pebbled portion in the Särkilampi conglomerate showing crudely developed horizontal type bedding. Top of sequence upwards. Length of pen 14 cm. Särkilampi, Kiihtelysvaara.

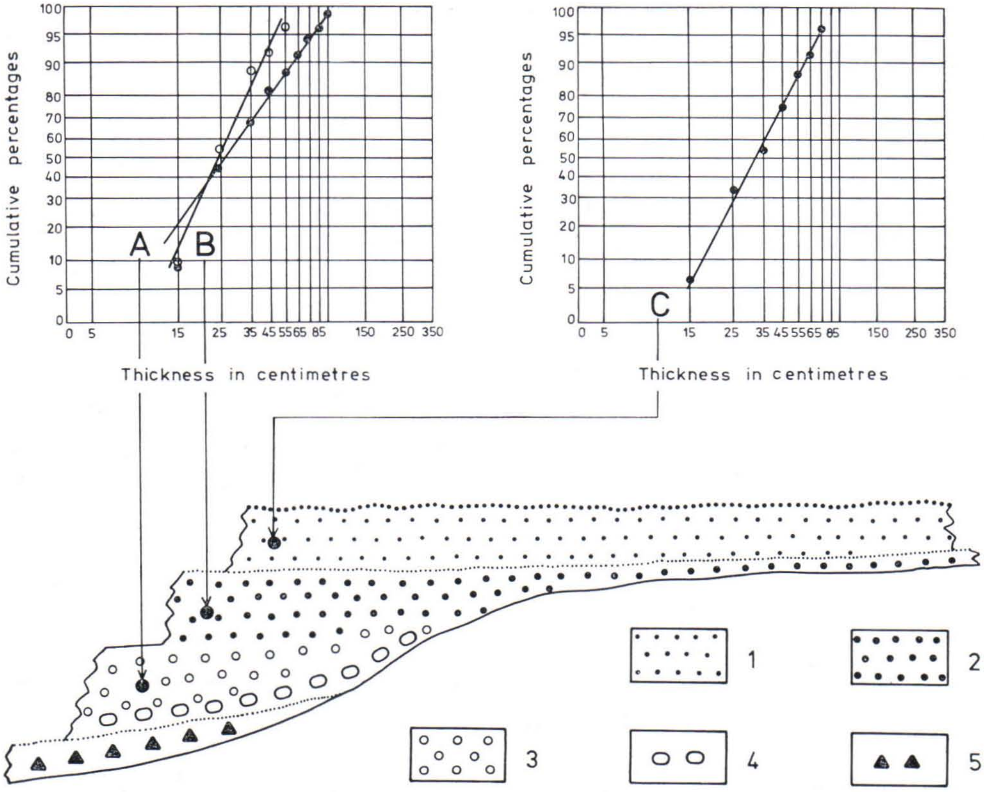

Fig. 34. Schematic representation of bedding thickness of various arkosites plotted on logarithmic probability paper. A) Horizontally bedded arkosic conglomerate. B) Horizontally bedded arkosite. C) Cross-bedded arkosite.

Symbols: 1 = upper arkosite; 2 = lower arkosite; 3 = small-pebbled arkosic conglomerate; 4 = big-cobbled arkosic conglomerate; 5 = breccia-conglomerate.

Lower arkosite in the Särkilampi area presumably deposited cyclically as did the underlying small-pebbled conglomerate, since the variation in thickness in its beds is also consistent with lognormal distribution (Fig. 34). The graph, however, differs from that of the small-pebbled conglomerates, and thus suggests a different rate of deposition. In some places, e.g. the Hyypiä area, the Lower arkosite also shows cross-bedding (Fig. 9).

Upper arkosite

The Upper arkosite typically exhibits large-scale planar cross-bedding. Here too the variation in bed thickness is lognormal (Fig. 34), but the graph differs notably from that of the Lower arkosite and thus indicates a different mode of sedimentation.

Mud cracks

In the Sääperi area well-preserved mud cracks have been found in the lower part of the Lower arkosite, where arkosic beds alternate with laminar argillaceous siltstone beds. The mud cracks, injected with sandy matter, exhibit polygonal or forking patterns (Fig. 35A—B). Such mud cracks, which often form during the drying and compaction of a water-saturated muddy sediment, indicate alternating dry and wet periods.

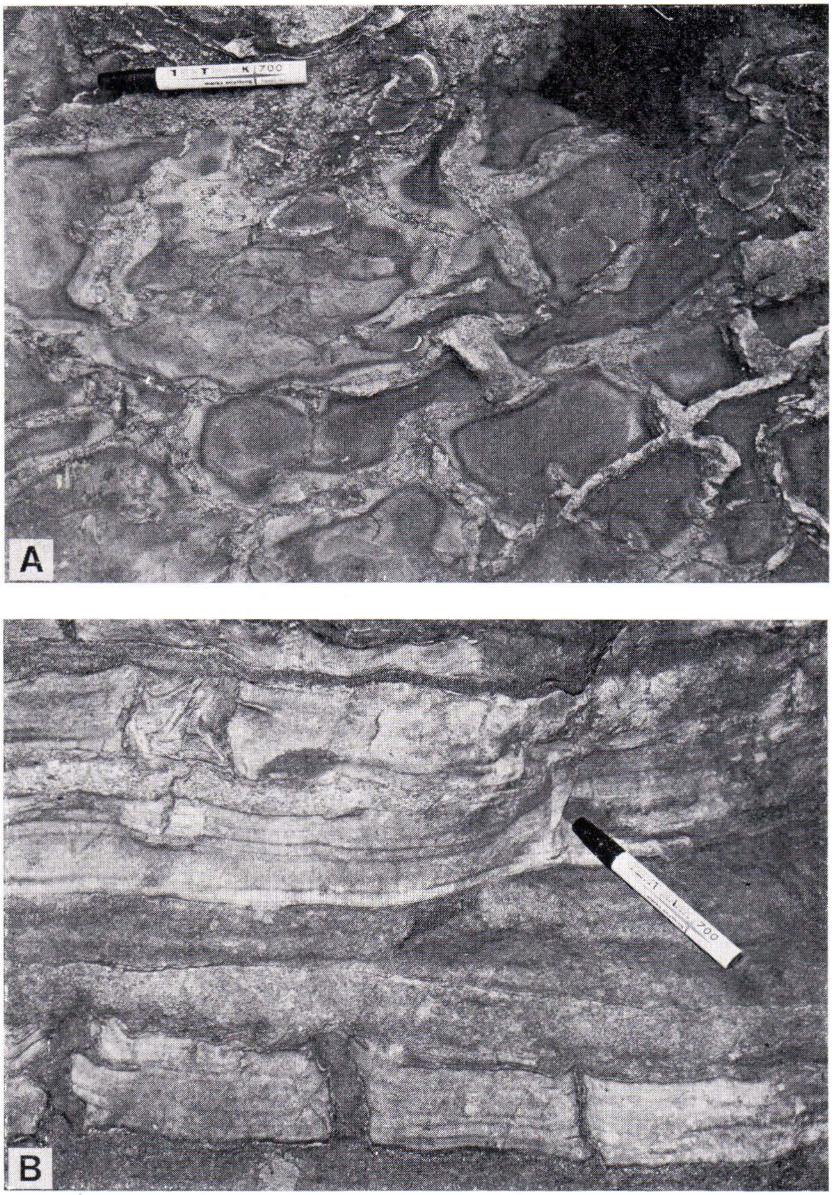

Fig. 35. Mud-cracked argillaceous siltstone layers alternating with arkosic layers in the Lower arkosite. A) Plan.

B) Vertical section. Length of pen 14 cm. Sääperi, Värtsilä. Photo by E. Räisänen.

Grain size and grain-size distribution

Arkosic conglomerates

The clasts in the big-cobbled conglomerate are mostly of cobble size although some of pebble size are also met with; fragments of boulder size are rare (Fig. 36). The most common size is from 6 to 12 cm. In this conglomerate, however, the size of the clasts is reduced rapidly upwards and the conglomerate grades into the small-pebbled variety.

The clasts in the small-pebbled conglomerate are mainly of the size of coarse sand or pebbles, although some larger clasts have been encountered (Fig. 33) as well. The average size fluctuates between 4 and 16 mm. In the cyc-

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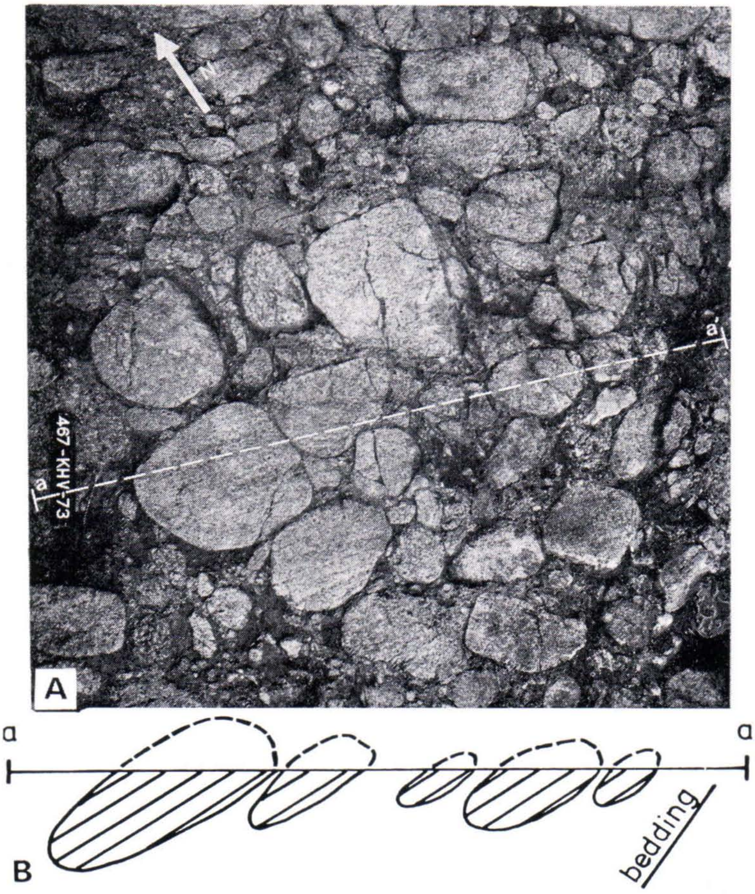


Fig. 36. A) Big-cobbled portion in the Särkilampi conglomerate. B) a—á hypothetic cross section. Length of tag 12 cm. Särkilampi, Kiihtelysvaara. Photo by E. Halme.

lically deposited small-pebbled Särkilampi conglomerate each phase consists of a lower part with the coarseness of conglomerate and an upper part with the coarseness of silt and clay. The thicker the bed, the coarser is the lower part. The material in the small-pebbled conglomerate is slightly sorted.

Lower and Upper arkosite

Approximate grain-size distributions in the Lower and Upper arkosites were determined under the microscope with the aid of an ocular micrometre by counting the number of grains that, in thin sections, belong to the different categories. Both histograms obtained (Figs. 37A and B) are based on a population of 400 grains. The results are, however, only indicative for the smaller grain sizes, recrystallization having blurred the original boundaries of the smaller grains. The histogram of the Lower arkosite (A) shows polymodal grain size distribution with a weak main mode in category 0—0.5 phi. The histogram of the Upper arkosite (B) also exhibits polymodal distribution, but with a more distinct main mode in category 0—0.5 phi. The quartiles Q_1 (25 %), Q_2 (50 %) and Q_3 (75 %) were determined for both arkosites on the basis of the cumulative distribution curves. The median and coefficients

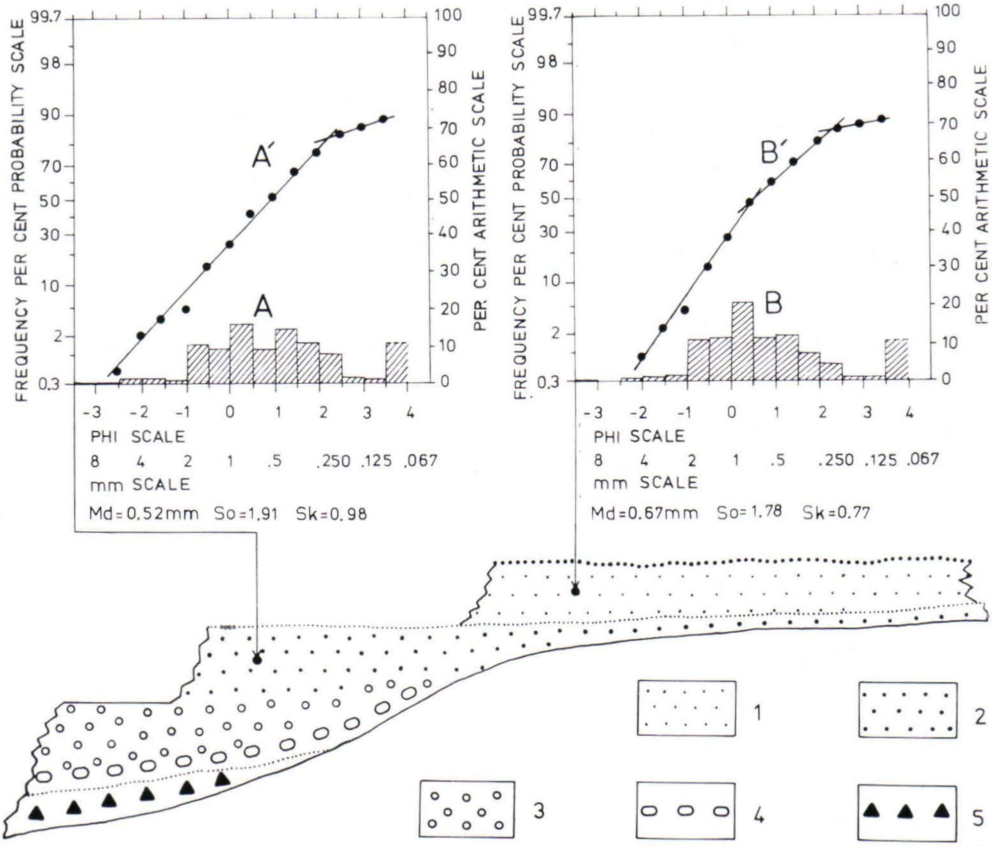


Fig. 37. Schematic diagram showing the grain-size distribution of two arkosites. A histogram and Á, log-probability curve for the Lower arkosite. B histogram and B′ log-probability curve for the Upper arkosite. Median (Md), sorting (So) and skewness (Sk) are also marked. Symbols: 1 = upper arkosite; 2 = lower arkosite; 3 = small-pebbled arkosic conglomerate; 4 = big-cobbled arkosic conglomerate; 5 = breccia-conglomerate.

of sorting and skewness given in Fig. 37 were calculated from the quartiles. The sorting index thus calculated indicates that the Lower arkosite ($S_0 = 1.91$) is moderately sorted and that the Upper arkosite is still more sorted and coarser.

If the grain size distribution is plotted on log probability paper, rectilinear segments appear in the distribution curve. Each segment represents the subpopulations referring to suspension, saltation and rolling, which permit conclusions to be drawn concerning the sedimentation environment (Visher 1969). The distribution curve of the Lower arkosite, presented as cumulative percentages (Fig. 37A′), shows truncation between 2 and 2.5 phi. The truncation has been interpreted as the demarcation between the saltation and suspension populations. The distribution curve of the Upper arkosite (B′) is similar, except that it also exhibits weak truncation at 1—0.5 phi, which may represent the boundary between the saltation and rolling populations. If these diagrams are compared with those given by Visher (1969), it is noticed that the former exhibit some features pertinent to fluvial sediments.

Shape and roundness

The pebbles and cobbles in the big-cobbled conglomerate are slightly elongated and bladed

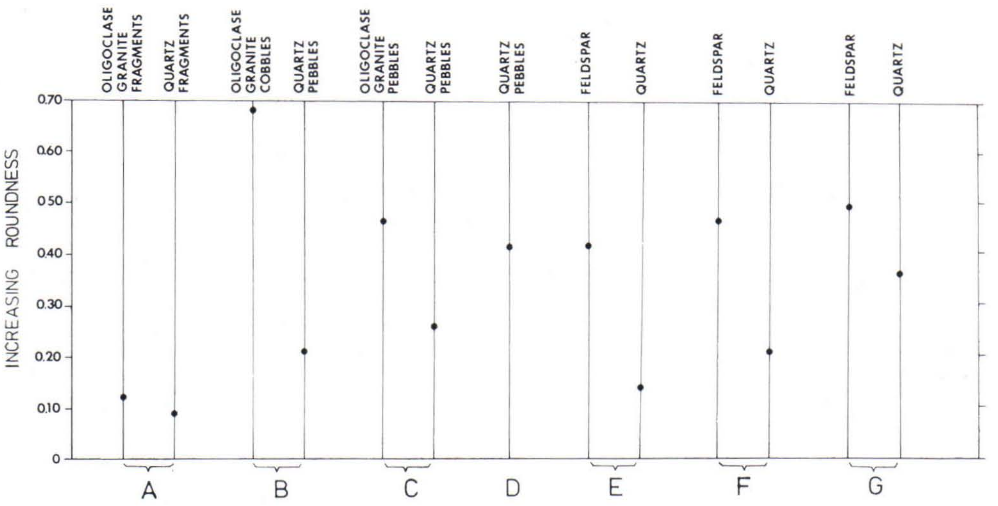


Fig. 38. Comparison of (average) roundness of clasts in breccia-conglomerate, arkosic conglomerate and arkosites. A = breccia-conglomerate; B = big-cobbled conglomerate; C = small-pebbled conglomerate; D = isolated quartz pebbles in the upper arkosite; E = basal arkosite; F = lower arkosite; G = upper arkosite.

Fig. 39. Upper arkosite with isolated pebbles of quartz (Q) and oligoclase granite (trd). Length of pen 14 cm. Särkilampi, Kiihtelysvaara. Photo by E. Räisänen.

in cross section. They are so well preserved that their flatness cannot be attributed to subsequent deformation; it must be a primary feature (Fig. 36).

The roundness was determined for some pebbles (by cutting them with a diamond saw) as suggested by Krumbein (1941). The fragments of oligoclase granite and quartz were distinctly more rounded than those of breccia-conglomerate (Fig. 38). If the oligoclase granite pebbles in the big-cobbled conglomerate at Särkilampi derived from around Saario in the southeast

(p. 69), as suggested by the present author, they were transported for a distance of about 20 km; hence, the well-developed roundness.

The degree of roundness seems to depend on the coarseness of the material: the constituents in the small-pebbled conglomerate are clearly less rounded than those in the big-cobbled conglomerate (Fig. 38).

Thin sections show that the granular quartz and feldspar grains in arkosites are almost invariably elongated (Fig. 40A—E); the quartz grains in particular tend to be ovoidal in shape. Roundness also was measured in the quartz and feldspar grains in arkosites (Fig. 38). The grains to be measured were first photographed in thin sections by a polaroid camera attached to a microscope; the roundness was then determined as described above. The feldspar grains in the basal arkosite are already fairly well rounded, whereas the quartz grains are still rather angular. In the upper part of the Lower arkosite, both the feldspar and quartz grains are slightly more rounded, whereas, not until the Upper arkosite are also the quartz grains distinctly rounded. The relationship between roundness and grain size is further demonstrated by the fact that individual quartz and oligoclase granite pebbles in the Upper arkosite are significantly more rounded than sand grains (Figs. 38 and 39). It is obvious that the constituents of the basal arkosites are not quite local either, and that the material in the subsequent arkosite beds derived from a greater distance.

Fabric of conglomerates

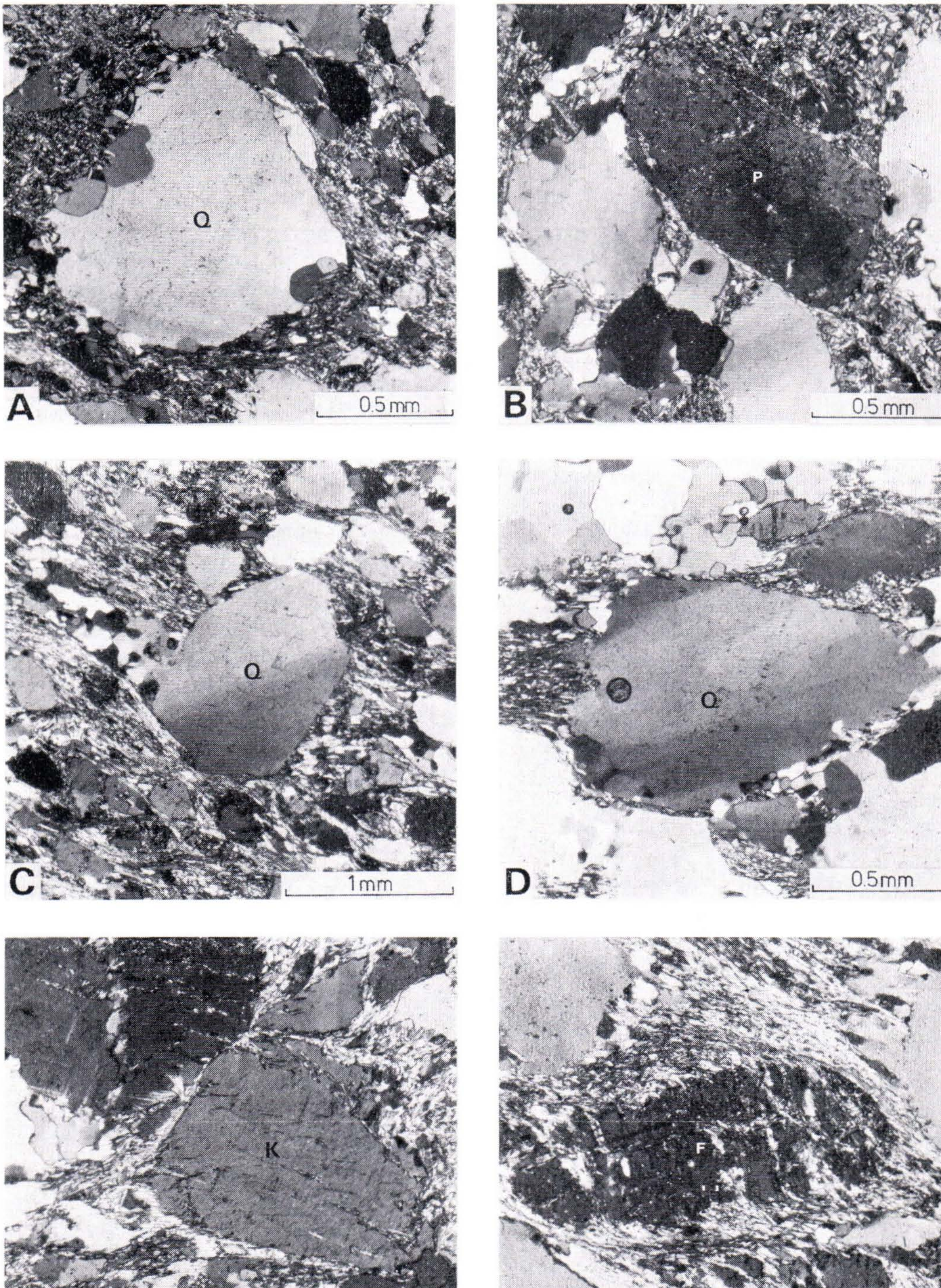
A faint WNW orientation has been established for the cobbles in the big-cobbled Särkilampi conglomerate; their plunge deviates by about 20 ° from the plane of bedding (Fig. 36). The coarseness of the constituents and the angle of plunge of the cobbles indicate a fluvial environment with a comparatively high gradient, in which the cobbles tend to orientate parallel to the flow. If this is so, the direction of the paleoflow was roughly from SE to NW. This concept is corroborated by the high abundance of oligoclase granite constituents that could well have derived from the oligoclase granite area at Saario (Nykänen 1968, p. 14), 15 to 20 km southeast of Särkilampi.

The pebbles in the small-pebbled conglomerate are not so distinctly elongated as those in the big-cobbled conglomerate; hence, they are without distinct orientation.

The mineralogical character of arkosites and related rocks

Big-cobbled conglomerate

The cobbles of the arkosic big-cobbled conglomerate are mainly from plutonites of the Prekarelidic basement, more rarely from schists (Fig. 36 and Table 6). Cobbles of oligoclase granite (or trondhjemite) abound in the Särkilampi and Viesimo conglomerates. Lithologically they are very similar to the Saario oligoclase granite in the basement (Table 7, No 13; cf. Nykänen 1968, p. 14). The degree of weathering in the clasts of the conglomerate varies at random suggesting that both fresh and somewhat weathered primary material were mixed together during deposition. In most places the cobbles are packed so densely that they touch one another (Fig. 36); in the Viesimo conglomerate, however, there are portions in which the matrix predominates. The matrix between the cobbles is arkosic and commonly rich in sericite and carbonate.



A
Q
0.5 mm
B
P
0.5 mm
C
Q
1 mm
D
Q
0.5mm
E
K
0.5mm
F
F
0.5 mm

Small-pebbled conglomerate

The pebbles of the small-pebbled arkosic conglomerate are more variegated, and consist of mineral fragments of quartz, feldspar and tourmaline (Fig. 33 and Table 6) as well as rock fragments. In the plagioclase fragments the degree of weathering varies from fresh to weathered, whereas the potassium feldspar is fairly fresh. In the small-pebbled conglomerate at Särkilampi the clay—silstone upper part of each phase is composed mainly of sericite, quartz and carbonate. Moreover, irregular carbonate nodules and lenses occur almost invariably in the clay—siltstone portion (Fig. 33). The white or pink carbonate in them is predominantly calcite. The mode of occurrence resembles that of the lime-rich caliche deposits that form in certain semi-arid regions (cf. p. 61).

Lower arkosite

The predominant granular constituents in the Lower arkosite are quartz and feldspars (Figs. 40A—C and Table 7, Nos. 16—20). Tourmaline and rock fragments occur sporadically. Plagioclase is the predominant feldspar (Nos. 16 and 17) in the lower beds, but the most sparse in the upper beds (Nos. 18—20). In internal structure the quartz grains show a slightly undulatory lamination (Fig. 40A); less often they are polycrystalline. In the lower beds the plagioclase grains are comparatively fresh (Fig. 40B), but in the upper beds intensely sericitized. The potassium feldspar grains are fairly fresh. The interstices between the granular grains are filled with a matrix composed mainly of fine-grained quartz, sericite, carbonate, biotite and chlorite. In some beds most of the cement is carbonate (No. 20). Owing to alterations in the plagioclase, the total abundance of sericite in arkosite is quite high, particularly in the upper portion of the Lower arkosite (Nos. 18 and 19). The primary feldspar abundance has been estimated at between 10 and 40 %. The primary mineral composition suggests that the lower arkosites are subarkoses and arkoses.

The argillaceous siltstone interbeds in the Lower arkosite are composed mainly of micas, chlorite and quartz, except in the basal part of the deposit where they also contain abundant carbonate (Table 7, Nos. 14 and 15). Carbonate lenses and nodules are also common. Fairly pure tremolite has been found in intensely metamorphosed carbonate lenses, e.g. in the basal arkosite in the Sääperi area (Hytönen and Ojanperä 1976).

Upper arkosite

The predominant granular grains in the lower and middle beds in the Upper arkosite Member are quartz and potassium feldspar (Table 7, Nos. 21—24). The quartz grains show undulatory lamellar texture (Fig. 40D), less often

Fig. 40. Photomicrographs of typical sand-sized grains in various arkosites.

A) Large angular quartz (Q) grain and some smaller quartz grains in a matrix consisting mainly of quartz, micas and carbonate. Basal arkosite (Thin section No. Ku 00062). Crossed nicols. B) Large rounded plagioclase (P) grain and some smaller quartz grains in a matrix consisting mainly of quartz, micas and carbonate. Basal arkosite (Thin section No. Ku 00062). Crossed nicols. C) Large poorly rounded quartz (Q) grain and some smaller quartz grains in a matrix consisting mainly of quartz and sericite. Lower arkosite (Thin section No. Ku 00285). Crossed nicols. D) Large moderately rounded quartz (Q) grain and some smaller quartz grains in a matrix consisting mainly of quartz and sericite. Upper arkosite (Thin section No. Ku 00283). Crossed nicols. E) Large rounded potassium feldspar (K) grain and some smaller quartz grains in a matrix consisting mainly of quartz and sericite. Upper arkosite (Thin section No. Ku 00283). Crossed nicols. F) Large strongly sericitized feldspar (F) grain and smaller quartz grains in a matrix consisting mainly of sericite and quartz. Upper arkosite (Thin section No. Ku 00325). Crossed nicols. Photo by E. Halme.

Table 7
Mineral compositions of various rocks in the Kiihtelysvaara Arkosic Suite. Determined by point-counting method.

Minerals	12	13	14	15	16	17	18
Quartz	29.9	33.3	33.1	55.6	56.1	40.0	61.6
Plagioclase	51.9	60.8	+	+	15.5 a)	41.6 a)	+
(An %)	(27—30)	(14—20)			(10—18)	(15—20)	
Potassium feldspar	+	0.2	+	+	+	+	5.9 a)
Biotite	10.4	1.4	22.6	19.9	5.8	5.7	+
Sericite	4.2 b)	2.2	19.1	1.2	19.7	11.8	32.0 b)
Chlorite	2.6	1.4	+	2.6	+	+	+
Carbonate	+	+	24.6	18.8	2.8	0.8	0.4
Apatite	+	+	+	+			
Titanite	+	+		+	+	+	
Zircon	+	+	+	+	+		+
Tourmaline	0.9	+	+	+	+	+	
Epidote				1.8		+	
Opaque	+	+				+	+
Rock fragments					+	+	
Total d)	100.0	100.0	100.0	100.0	100.0	100.0	100.0

+) detected minor constituent; a) including the feldspar not identified; b) partly derived from decomposition of

12. Quartz dioritic clast (ø 10 cm) from the Sääperi arkosic conglomerate (Specimen No. 83/EOR—74), Värtsilä.
13. Oligoclase granite (trondhjemitic) clast (ø 20 cm) from the Särkilampi arkosic conglomerate (Specimen No. 71A/EOR—74), Kiihtelysvaara.
14. Calcareous siltstone (Specimen No. 22/EOR—72), Hyypiä, Kiihtelysvaara.
15. Calcareous siltstone (Specimen No. 55/EOR—74), Sääperi, Värtsilä.
16. Basal arkosite (Specimen No. 28/LJP—75), Karsikkojärvi, Kiihtelysvaara.
17. Lower arkosite (Specimen No. 105/JJK—74), Viesimo, Kiihtelysvaara.
18. Lower arkosite (drill core No. R329/4241/46.50), Särkilampi, Kiihtelysvaara.
19. Lower arkosite (drill core No. R329/4241/39.75), Särkilampi, Kiihtelysvaara.
20. Lower arkosite (drill core No. R329/4241/31.70), Särkilampi, Kiihtelysvaara.

polygonal texture. The potassium feldspar grains exhibit mild sericitization along the cleavage and margins (Fig. 40E). Substantially altered plagioclase grains have been encountered very rarely. The interstices between the granular grains are filled with a fine-grained matrix composed mainly of quartz and sericite but also with small amounts of carbonate, chlorite, tourmaline, zircon and rutile. It has been estimated that the primary feldspar abundance in the Upper arkosite is 10 to 40 %. On the basis of primary composition, the upper arkosites, like the lower arkosites, are subarkoses and arkoses.

Discrete rounded quartz pebbles and cobbles are typical of the Upper arkosite; the largest of them attain almost 20 cm in diameter. Some pebbles of oligoclase granite and granite have also been encountered (Fig. 39). Preliminary studies (K. Kinnunen, personal communication) show that the quartz in some of the quartz pebbles is bluish and slightly undulatory in thin section; according to the fluid inclusion, these are very similar to the quartz of the Mustalampi type (p. 51).

Younger or Early Jatulian weathering crust

It is noteworthy that the potassium feldspar in the upper beds of the Upper arkosite Member is intensely altered, as is manifested by the rapid increase in sericite abundance (Table 7, Nos. 25—29). In the Sääperi area, drilling intersected a 20 to 30 m thick ancient weathering zone overlain by piles of quartzites. The poorly consolidated and pinkish rock of this younger or Early Jatulian weathering crust (Fig. 41) shows feldspar grain and rare oligoclase granite pebbles that are still recognizable. The feldspars, however, have been completely replaced by sericite aggregates (Fig. 40E).

19	20	21	22	23	24	25	26	27	28	29
53.0	47.3	60.1	55.2	57.0	55.5	60.3	69.3	70.8	66.6	59.9
+	+	+	+	+						
7.6 a)	17.4 a)	25.0 a)	11.6 a)	10.6 a)	7.0 a)	2.0 a)	1.0 a)	+	+	+
0.3	3.0	+								
22.5 b)	11.2	20.2	19.3	31.2 c)	37.3	37.5 c)	29.6 c)	29.1 c)	33.1 c)	39.9 c)
+	+								+	
16.5	21.1	+	+	+						
			+		+	+				+
+	+	+	+	+	+	+	+	+	+	+
			+		+	+			+	+
+	+	+	+	+	+		+	+	+	
100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

plagioclase; c) partly derived from decomposition of potassium feldspar; d) after addition of all minor constituents

21. Upper arkosite (drill core No. R329/4241/25.75), Särkilampi, Kiihtelysvaara.
22. Upper arkosite (drill core No. R329/4241/1.55), Särkilampi, Kiihtelysvaara.
23. Upper arkosite (Specimen No. 1S/MS—74), Särkilampi, Kiihtelysvaara.
24. Upper arkosite (Specimen No. 2S/MS—74), Särkilampi, Kiihtelysvaara.
25. Upper arkosite (Specimen No. 20A/EOR—72), Hyypiä, Kiihtelysvaara.
26. Upper arkosite (Specimen No. 103/JJK—74), Viesimo, Kiihtelysvaara.
27. Upper arkosite (drill core No. R307/4232/167.75), Sääperi, Värtsilä.
28. Upper arkosite (drill core No. R307/4232/163.35), Sääperi, Värtsilä.
29. Upper arkosite (drill core No. R307/4241/93.45), Hyypiä, Kiihtelysvaara.

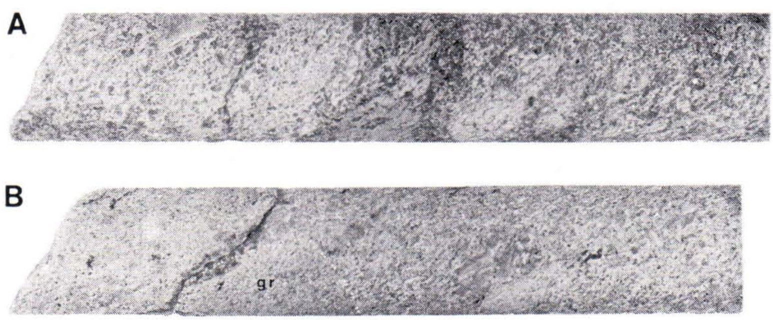


Fig. 41. Drill-core samples from the upper part of the younger (Early-Jatulian) weathering crust, Sääperi, Värtsilä. A) Intensely sericitized Upper arkosite, the hanging wall of the Arkosite Formation (No. R307/4232/162.95). B) Intensely altered Upper arkosite showing intensely altered granite pebble (gr), the hanging wall of the Arkosite Formation (No. R307/4232/171.55). Drill cores about 32 mm in diameter. Photo by E. Halme.

The chemical character of arkosites and related rocks

The Särkilampi reference series

Since the mineral composition in arkosites varies, it is to be expected that the same is true of their chemical composition. A rough idea of the variation in chemical composition across the Arkosite formation is given by the set of samples collected from the various rock units at Särkilampi (Fig. 10). The drill- core

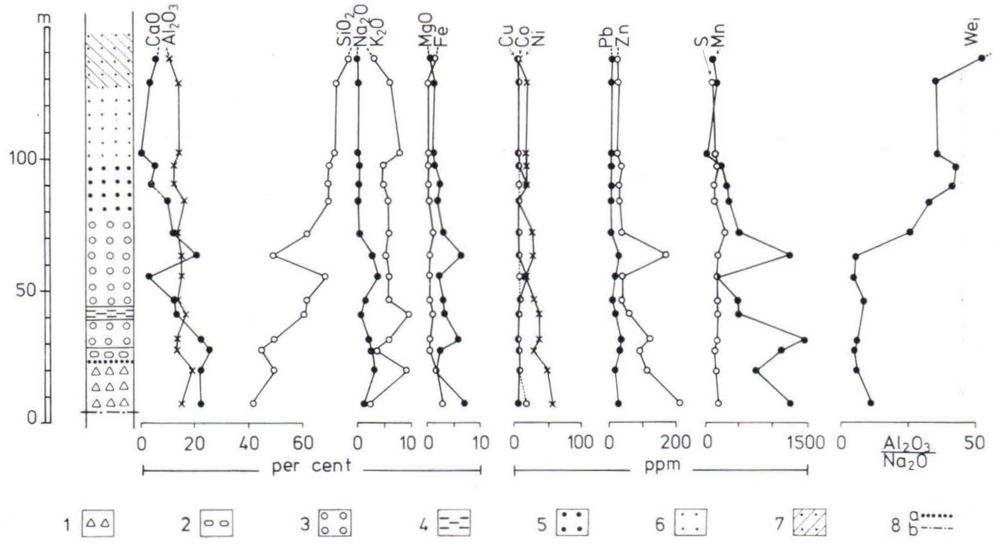


Fig. 42. Schematic diagram showing variation in chemical composition across the Särkilampi Arkosite Formation. Symbols: 1 = breccia-conglomerate; 2 = big-cobbled conglomerate; 3 = small-pebbled conglomerate; 4 = argillaceous siltstone; 5 = lower arkosite; 6 = upper arkosite; 7 = younger weathering crust; 8 a = unconformity; 8 b = graded unconformity.

samples, about 30 cm long, were analysed by emission spectrophotometer (G. Wansén) for SiO_2, Al_2O_3, MgO, CaO, Na_2O and K_2O (semiquantitative analyses). Total iron, Mn, Cu, Ni, Pb and Zn were assayed by AAS, and sulphur by a Leco M 621 sulphur analyser (A. Nurmi). A certain regularity was observed in the variation in chemical composition (Fig. 42):

— SiO_2 increases upwards in the sequence, except in beds rich in carbonate, owing to the rise in the abundance of quartz.

— Al_2O_3 decreases slowly upwards.

— Total iron and MnO decrease distinctly upwards, excluding the small increase in the upper portion of the Upper arkosite owing to (secondary) Fe oxide; the curve clearly reflects the gradual decrease in the abundance of iron-magnesium minerals.

— CaO is high in arkosic conglomerate and in the Lower arkosite; it declines rapidly in the Upper arkosite, although it shows a slight increase in the upper portion of the latter; the curve illustrates clearly the variation in carbonate abundance.

— Na_2O declines rather rapidly upwards, perhaps because of the expulsion of Na liberated in the decomposition of plagioclase.

— K_2O is fairly high throughout, but varies in accordance with the mica and potassium feldspar abundances; in the Upper arkosite the rapid increase resulting from the high abundance of potassium feldspar is succeeded by a decline due to the decomposition of potassium feldspar.

— Mn, Cu, Co, Ni, Pb, Zn and sulphur are distinctly depleted upwards, whereas Mn and Zn and, less markedly, Pb are enriched in the beds rich in carbonate.

— The Al_2O_3/Na_2O ratio, which has been used here as a weathering index, shows the greatest changes in the upper portions of the Lower and Upper arkosites; the curve illustrates well the variation in the degree of alteration in the feldspars.

Table 8
Chemical composition of two arkosites in the Arkosite Formation (weight per cent).

	10	11
SiO_2	73.06	89.91
TiO_2	0.17	0.11
Al_2O_3	13.87	4.11
Fe_2O_3	0.97	0.29
FeO	0.45	0.34
MnO	0.04	0.05
MgO	3.32	1.60
CaO	0.20	0.46
Na_2O	0.19	0.59
K_2O	4.83	1.68
P_2O_5	0.06	0.03
CO_2	0.00	0.00
H_2O^+	2.68	0.76
H_2O^-	0.12	0.02
	99.96	99.95

10. Lower arkosite, Hyypiä, Kiihtelysvaara. Analyst: P. Ojanperä. (Nykänen 1971b, p. 35).

11. Sericitized upper arkosite, Hyypiä, Kiihtelysvaara. Analyst: P. Ojanperä (Nykänen 1971b, 35).

The whole-rock analyses

As is to be expected, the chemical composition of the arkosites (Table 8, Nos. 10 and 11) correlates rather well with their mineral composition. The first analysis (No. 10), representing the composition of the carbonate-poor Lower arkosite, is not unlike typical analyses of arkosites reported by Pettijohn (1975, p. 216); the rather high MgO content in this arkosite is due to the comparatively high biotite and chlorite abundances. The second analysis (No. 11), referring to the composition of the Upper arkosite, represents a mature arkose or subarkose.

The average contents of the minor constituents

The distribution of the trace elements Cu, Ni, Zn, Pb and Co, and of the phosphorus and sulphur in the conglomerates and arkosites of the arkosic association was studied statistically. Representative samples, mainly drill-core samples, were collected throughout the study area and analysed for the above-mentioned elements. For the sake of comparison, analyses typical of rocks in the Prekarelidic basement were also included. By means of an ADP program, the analytical data were compiled into an extended scatter diagram, which shows the arithmetic means, the geometric mean, the median and the standard deviation (Fig. 43). The Ni, Co and Zn contents in the conglomerates and arkosites are conspicuously lower than in the rocks of the basement; the changes are not so clear for the other elements. The limited number of samples does not warrant the drawing of more far-reaching conclusions.

Conclusions concerning the depositional environment and origin of the material

It can be assumed that when the relief of the Prekarelidic basement was levelled and the bedrock weathered, various residual sediments, e.g. conglomerates and arkoses, were formed, depending on the distance and mode of transport. The Lower arkosite and Arkosic conglomerates represent less »washed» resistates formed through reworking, whereas the Upper arkosite refers to better washed resistates that were transported a longer distance. There is, however, no reason to preclude the existence of various intermediate types.

In the light of what was stated above about Arkosic conglomerates, their depositional environment and constituents can be characterized as follows:

— The material shows moderate mechanical reworking, sorting and rounding; thus it was transported for at least a short distance.

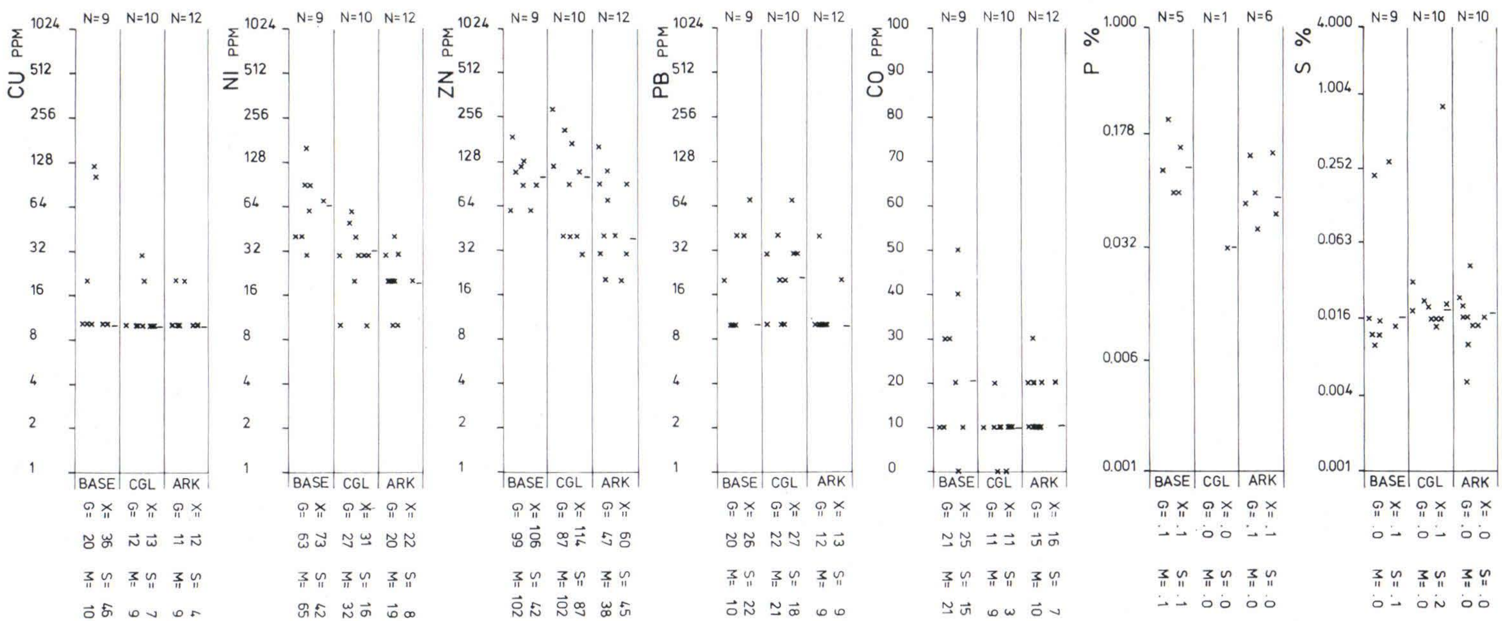


Fig. 43. Scatter diagram showing distribution of copper, nickel, zinc, lead, cobalt, phosphorus and sulphur in the rocks of the Kiihtelysvaara Arkosic Suite compared with those in the rocks of the Prekarelidic basement complex. Cu, Ni, Zn, Pb and Co were analysed by AAS, and sulphur by a Leco M 621 sulphur analyser (A. Nurmi). Phosphorus was assayed at the ore laboratory of the Chemistry department. Scales of columns logarithmic. Symbols: BASE = Prekarelidic rocks; CGL = conglomerates; ARK = arkosites; X = arithmetic mean; M = median; S = standard deviation; N = number of observations. The medians marked on the column by short horizontal lines.

— The mode of occurrence and the bed structures indicate that the constituents were deposited cyclically after transport by strong currents; the characteristic features resemble those of the alluvial-fan type of deposits described by Bull (1962), which Ojakangas (1965) has suggested are of fluvial origin.
— The abundance of carbonates as cement and as nodules and lenses within the beds demonstrates that, owing to carbonaceous solutions, agglutination might have taken place rather rapidly; hence, it can be assumed that the material was deposited under arid or semi-arid paleoclimatic conditions.
— The degree of weathering in the granitoid and feldspar clasts varies at random; clasts are encountered that derived either from redeposited weathering crust or from fresh granitoids; hence, the sources of the material were areas of rather uneven relief in the Prekarelidic basement, e.g. slopes on which both weathered and fresh material were available.

The depositional environment and the constituents of the Lower arkosite can be further characterized as follows:

— Excluding the thin mantle of basal arkosite, the material underwent moderate mechanical reworking and sorting (S_0=1.91), and its grains became distinctly rounded; in other words, the material is not strictly of local origin.
— The mode of occurrence and the sedimentary structures, e.g. bedding, refer to fluvial deposition. The arkosic material probably deposited in an area of lower relief and farther away from the source than did the conglomerates; even so, the phases seem to be related to each other.
— The mud-cracks in the argillaceous silt beds in the Lower arkosite (Fig. 35) demonstrate alternate dry and wet seasons. On the other hand, the abundance of carbonates refers to a desert-like depositional environment (Glennie 1970, p. 144) and to semi-arid paleoclimatic conditions, as do the conglomerates.
— The Lower arkosite was presumably formed, mainly through mechanical reworking, from the weathering products of rocks in the Prekarelidic basement, but also to some extent from a chemically altered short-ranged feldspar-bearing resistate; hence, the material exhibits a certain relationship to the adjacent bedrock.

The depositional environment and the constituents of the Upper arkosite are characterized as follows:

— The material is mechanically more reworked, coarser and more sorted (Fig. 37) than the Lower arkosite, and the individual grains are significantly more rounded (Fig. 38); in other words, the material was transported for a considerable distance.
— The mode of occurrence, the coarseness and the large-scale cross-bedding with disseminated quartz and granite pebbles in the lower portion of the foreset-laminae (Fig. 39) indicate that the constituents were transported by strong currents; it is therefore a fluvial sediment.
— The disappearance of carbonate suggests that the paleoclimate became wetter. This would presumably mean the formation of a more regular river system and the passing of sediments into braided river deposits. It is not impossible that the »sheet sands» of the Upper arkosites were formed as a result of the continuous migration of river beds (Schumm 1968).
— The Upper arkosite may have formed, mainly through intense mechanical reworking, from the weathering products of rocks in the Prekarelidic basement; it may, however, also include a chemically weathered, feldspar-bearing resistate transported over a long distance, whose constituents are distinctly better washed than those in the Lower arkosite.

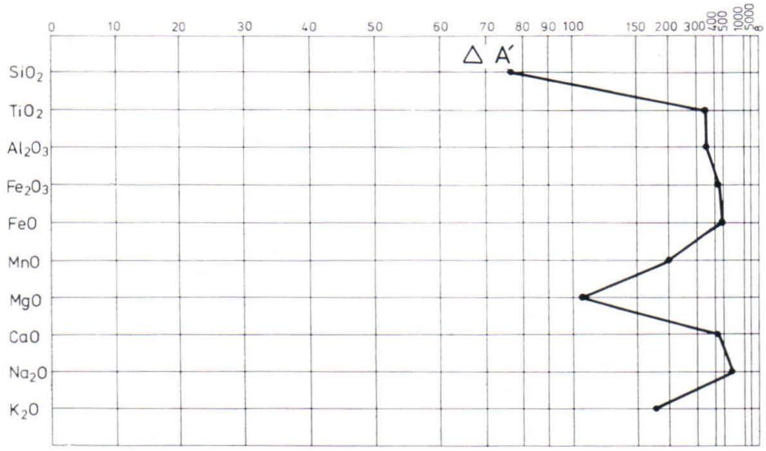


Fig. 44. Relative losses and gains of oxide constituents (△ Á) in the reworking of the source material with the assumed composition of (A) which gives rise to arkosite Á as an intermediate product.

Attempts were made to clarify the type of reworking process by determining relative gains and losses in oxidic constituents. The starting point was that the material of the Upper arkosite was assumed to derive from such a great distance that it had time to mix with material from all the most common rocks in the Prekarelidic basement. If we assume that the weighted average composition of the rocks in the Prekarelidic basement given in Table 2 (No. 8) represents the composition of the source material (A), and that the composition of the Upper arkosite given in Table 8 (No. 11) refers to the composition of a residue (A′), then the relative gains and losses in chemical constituents (Fig. 44) can be estimated with the aid of a »gain and loss» diagram (see Pettijohn 1957, pp. 501 and 502). One way of interpreting this diagram is to presume that 77 g of Upper arkosite contain as much SiO_2 as do 100 g of starting material, or that 357 g of arkosite are needed to obtain as much Al_2O_3 as do 100 g of starting material. It is obvious that Al_2O_3 was depleted in relation to SiO_2; thus the following order of loss is obtained for the oxide constituents: Na_2O, FeO, Fe_2O_3, CaO, Al_2O_3, TiO_2, K_2O, MnO, MgO and SiO_2.

This order of loss for oxides differs markedly from the average order (cf. Goldich 1938), particularly for Fe_2O_3 and Al_2O_3, which, depending on the climate, generally tend to become enriched during the weathering process. This, in accordance with the facts stated previously, suggests that mechanical reworking predominated over chemical weathering. Fines, such as micas and clay minerals, were probably transported farther from the source areas by wind and water; this would explain the losses in some of the constituents, e.g. FeO, Fe_2O_3 and Al_2O_3.

THE KIIHTELYSVAARA ORTHOQUARTZITE SUITE

The period of weathering and denudation that followed the deposition of arkose sands presumably came to an end in the Kiihtelysvaara—Värtsilä area when the sea invaded the mainland. The transgression of the sea introduced a new stage in Karelian sedimentation, and the arkose sand deposits that had avoided erosion were gradually covered by quartz sand, giving rise to orthoquartzitic beds characteristic of the Jatulian formations that in this context are called the Kiihtelysvaara Orthoquartzite Suite. The deposition of quartz sands in the study area did not, however, take place without interruption; volcanic activity and a period of denudation intervened. Eventually the volcanic beds still in existence were covered by sandy material transported from the mainland, and the volcanic detritus was mixed with the depositing sand. Later the shallow littoral depositional environment passed into a somewhat deeper marine environment and the sandy sediments graded into carbonate sediments. The quartz sands were gradually consolidated and eventually, during the Svecokarelidic orogeny (about 1800 to 1950 Ma), they were metamorphosed into quartzites.

The associations and their mode of occurrence

In the Kiihtelysvaara—Värtsilä area the quartzite horizon that rests on the Arkosite Formation and is separated from it by an unconformity marked by a denudation surface is known as the Lower quartzite Formation. This formation begins with sericite quartzite (the Sericite quartzite Member) and grades into a more pure quartzite (the Orthoquartzite Member). In places it is overlain by a bed of feldspathic quartzite (the Feldspathic quartzite Member).

The occurrences of the Sericite quartzite Member typically show lens- or sheet-like shapes that pinch out westwards and are laterally associated with each other constituting the lower section of the Lower quartzite Formation. The lower part of the Sericite quartzite Member is characterized by numerous quartz conglomerate interbeds and narrow clay slate interlayers. The number of interbeds and the abundance of sericite gradually diminishes upwards and the rock grades into orthoquartzite.

The Orthoquartzite Member occurs in lens- or wedge-like shapes and constitutes the middle and upper sections of the Lower Quartzite Formation. Except in the uppermost layers of the Member, where the abundance of sericite increases and interbeds of quartz conglomerate 5 to 10 cm thick are met with locally, the occurrences are largely sediments composed of comparatively pure quartz sand.

The occurrences of the Feldspathic quartzite Member are fairly thin lens-like shapes that have not been encountered anywhere except in the Kalkunmäki area, in the upper part of the Lower quartzite Formation immediately below the Volcanite Formation.

The total thickness of the Lower quartzite Formation varies considerably, the greatest thicknesses, between 300 and 550 m, occurring between Karsikkojärvi and Haluksenlammit and in the Sääperi syncline. The maximum thickness is attained at Särkilampi. Between Haluksenlammit and Sääperi the total thickness is somewhat less, being from 200 to 300 m. In the northernmost part, northwest of Karsikkojärvi, the thickness of the deposit suddenly drops to almost half and continues to decrease slightly northwestwards.

The Lower quartzite Formation is overlain by fairly continuous metavolcanite beds that extend from the Sääperi area northwards to Karsikkojärvi in Kiihtelysvaara (Fig. 5); here they suddenly come to an end. The thickness of this Volcanite Formation varies from 30 to 80 m. Volcanic rocks will be discussed in a later chapter (p. 114).

In the present context, the quartzite horizon that rests on volcanites, with a denudation unconformity in between, is called the Upper quartzite Formation. Typical of the lower beds of this formation are the fairly coarse grain size, the volcanic detritus, and the conglomerate interbeds that are repeated every few metres. The material in the upper part is more fine-grained, containing numerous clay slate, tuff and tuffite intercalations and abundant carbonates in the topmost part.

The total thickness of the Upper quartzite Formation varies considerably from one area to another, being thickest, from 100 to 200 m, between Hyypiä and Haluksenlammit; it is noticeably less thick, perhaps only a few tens of metres, south of Haluksenlammit as far as Sääperi. North of Hyypiä the thickness of the quartzite formation is rapidly reduced until at Kortevaara it is little more than ten metres or so.

Primary sedimentary structures and textures

Bedding

Sericite quartzite and Orthoquartzite

The Sericite quartzite Member often shows large-scale cross-bedding, the dip of the laminae varying from 15 ° to 20 °. Sandwiched between sandy cross-bedded layers, quartz conglomerate interbeds 2 to 25 cm thick are rhythmically repeated in the lower part of the sericite quartzites (Fig. 45A). Slightly laminated clay slate intercalations 2 to 10 cm thick are met with here and there (Fig. 45B), indicating that most of the material was transported and deposited by water. Upwards in the sequence the intercalations become more sparse until they gradually disappear altogether.

In some places in the Orthoquartzite Member, large-scale cross-bedding similar to that in the Sericite quartzite has been encountered, but owing to the monotonic mineral composition, the bedding is often difficult to see.

Some preliminary determinations of the direction of the paleocurrent were obtained from the cross-bedding in the quartzites; according to these, the principal direction was towards NW and the secondary direction towards SW. The determinations are not without ambiguity, because the quartzite beds were tilted at an angle of 60 ° during folding and the bedding is rarely so well developed that measuring is feasible. The determinations of the direction of the paleocurrent conducted by T. Mikkola (1955) and Ojakangas (1965) on the quartzites north of Kiihtelysvaara were based on more comprehensive data. Mikkola (1955, pp. 28—29) concluded that the prevailing direction was towards NE and the secondary direction towards NW. The prevailing direction by Ojakangas (1965, pp. 22—23) was towards NW; his other significant direction was towards SW, whereas the NE direction was less important. The incompatibility of the directions does not necessarily mean that there is any discrepancy between them; in a littoral zone material may well be transported to and fro.

Upper quartzite

The fairly coarse quartzite in the lower and middle parts of the Upper quartzite Formation contains volcanic weathering products (Fig. 45C) and often exhibits large-scale cross-bedding. Mantle-like quartz conglomerate interbeds 0.2 to 2 m thick occur at fairly large intervals between the cross-bedded quartz sand layers (Fig. 14). Also present are some rare tuffitic intercalations a few cm or dm thick (Fig. 16). Roughly in the middle of the quartzite deposit the conglomerate interbeds come to an abrupt end. The quartzite is more fine-grained in the upper part of the Upper quartzite Formation than in the lower layers and contains abundant clay slate intercalations a few mm or cm thick as well as rare volcanic

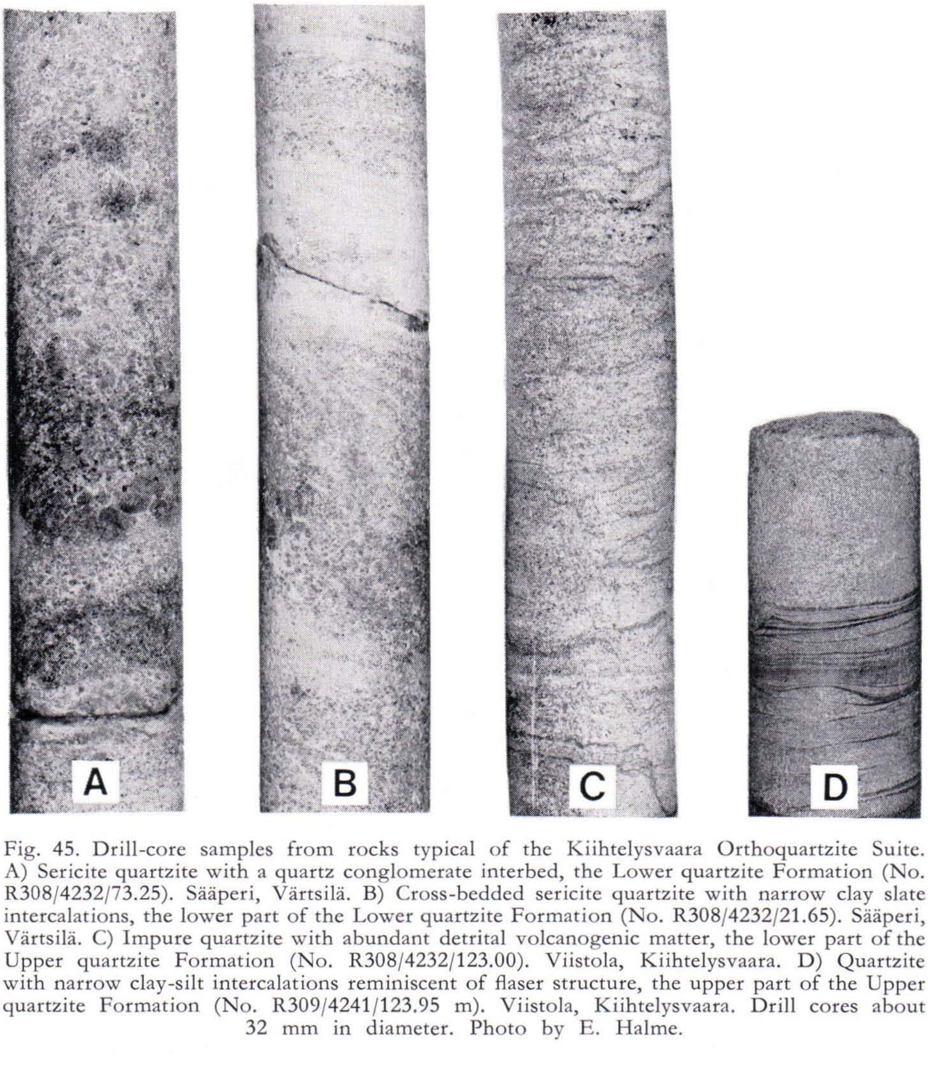


Fig. 45. Drill-core samples from rocks typical of the Kiihtelysvaara Orthoquartzite Suite. A) Sericite quartzite with a quartz conglomerate interbed, the Lower quartzite Formation (No. R308/4232/73.25). Sääperi, Värtsilä. B) Cross-bedded sericite quartzite with narrow clay slate intercalations, the lower part of the Lower quartzite Formation (No. R308/4232/21.65). Sääperi, Värtsilä. C) Impure quartzite with abundant detrital volcanogenic matter, the lower part of the Upper quartzite Formation (No. R308/4232/123.00). Viistola, Kiihtelysvaara. D) Quartzite with narrow clay-silt intercalations reminiscent of flaser structure, the upper part of the Upper quartzite Formation (No. R309/4241/123.95 m). Viistola, Kiihtelysvaara. Drill cores about 32 mm in diameter. Photo by E. Halme.

interbeds some metres thick. In the topmost carbonate-bearing part, which is also richer in clayey and silty material, flaser bedding has been encountered locally (Fig. 45D) as have clay slate intercalations a few dm or m thick with thinly interlayered silt/clay bedding.

Ripple marks

Well-developed ripple marks have been met with in the transitional zone between the Sericite quartzite and the Orthoquartzite, e.g. in the Hyypiä (Fig. 46) and Sääperi areas. Small ripple marks have also been observed in the upper layers of the Orthoquartzite Member at Raatevaara. All of these ripple marks are fairly symmetric; in some a secondary crest of low magnitude occurs on the bottom of the groove parallel to the axis. Typically there are several planes, one on top of the other, whose ripple marks differ in size. The symmetry in the ripple marks, the internal chevron type of structure and the »ripple index», L/H = 6.8 to 11.3, indicate that the ripples derive from the oscillation of shallow littoral water.

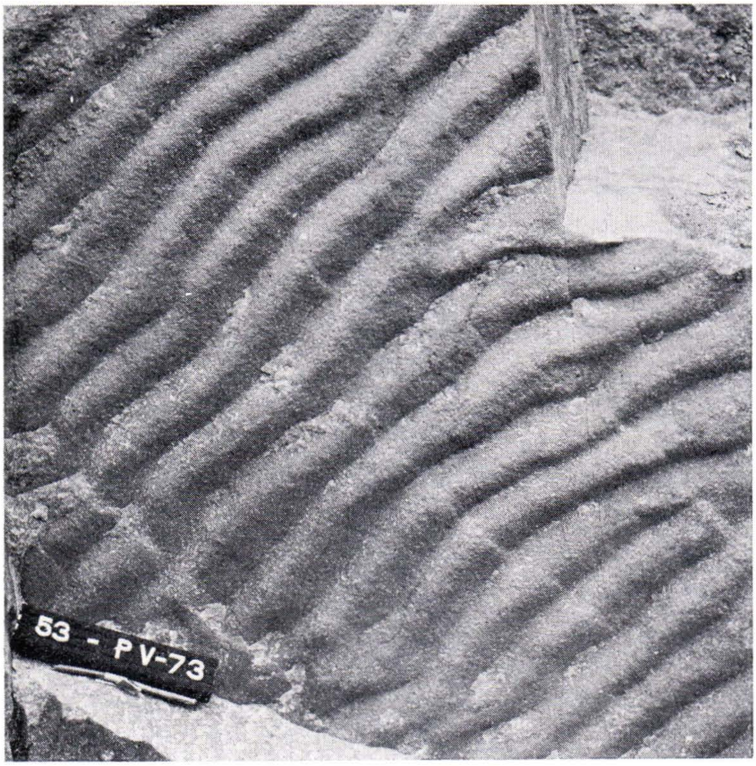


Fig. 46. Ripple-marked slab of sericite quartzite from the Lower quartzite Formation folded into a steep position; the lower side towards the observer. Length of tag 12 cm. Hyypiä, Kiihtelysvaara. Photo by E. Halme.

Grain size and grain-size distribution

Approximate grain-size distributions were determined for Sericite quartzite, Orthoquartzite and Feldspathic quartzite as described on page 66. The histograms of the Sericite quartzite and Orthoquartzite (Fig. 47A and B) were based on a population of 400 grains, and that of Feldspathic quartzite (C) on a population of 100 grains. On account of slight recrystallization in the rocks measured, the results for the finest fraction are only indicative. The histograms of the two former (A and B) show a slight polymodal distribution and, compared with the arkosites (Fig. 37), a more sharp main mode at the class 0—0.5 phi. The distribution curves of the quartzites demonstrate that, compared with the arkosites, the fraction coarser than 0 phi and the finest fraction are clearly depleted. The third distribution, that of the Feldspathic quartzite (C), is less distinct and does not show a clear main mode.

The Sericite quartzite and Orthoquartzite (Fig. 47) are distinctly better sorted and tend to be more fine-grained than the arkosites (Fig. 37). The Feldspathic quartzite is sorted like the Sericite quartzite but is somewhat more fine-grained.

The grain-size distributions of the Sericite quartzite, Orthoquartzite and Feldspathic quartzite are plotted as cumulative percentages on log probability paper (Fig. 47A′, B′, C′). The distribution curves of the Sericite quartzite and Orthoquartzite are very alike. The curve of the Feldspathic quartzite differs from the former; it is, however, based only a small population (100 grains). The most marked truncation points in the graphs occur at 0.5—1.0 phi and 2.0—3.0 phi; indistinct truncation occurs at 1.0—1.5 phi. The first truncation might denote the boundary between the rolling and saltation populations, the second one the boundary between the saltation and suspension populations, and the last one the fact that the saltation

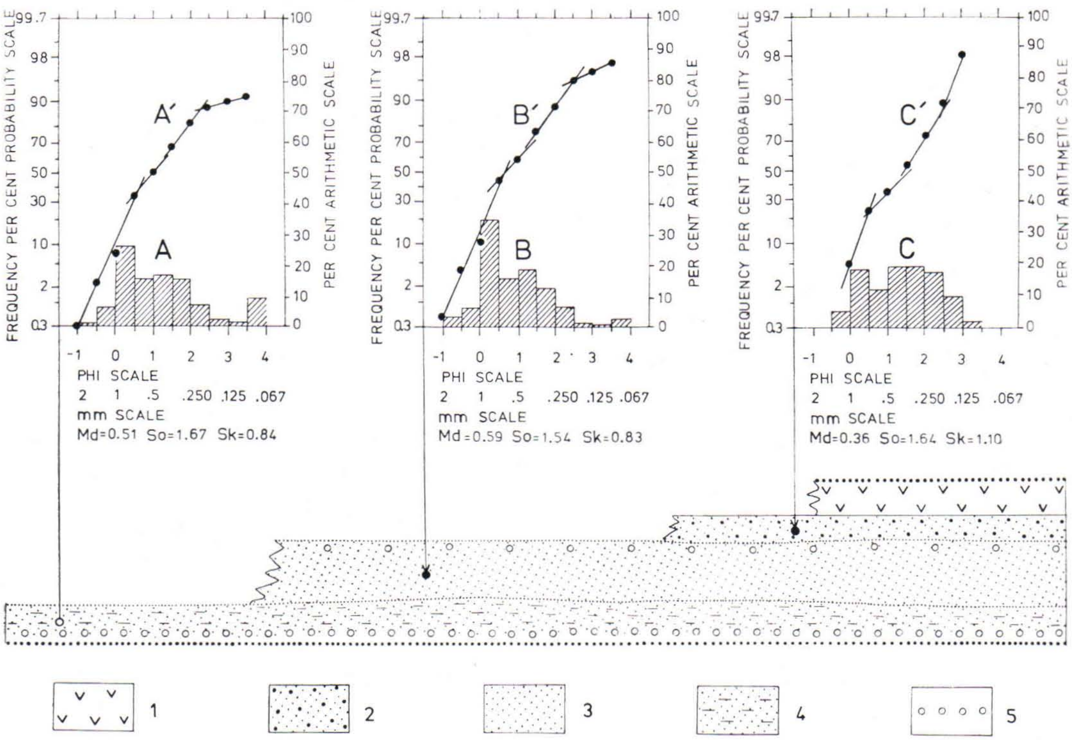


Fig. 47. Schematic diagram showing grain-size distribution of various quartzites in the Lower quartzite Formation. A histogram and A′ log probability curve for Sericite quartzite. B histogram and B′ log probability curve for Orthoquartzite. C histogram and C′ log probability curve for Feldspathic quartzite. Median (Md), sorting (So) and skewness (Sk) are also marked. Symbols: 1 = metavolcanite; 2 = feldspathic quartzite; 3 = orthoquartzite; 4 = sericite quartzite; 5 = quartz conglomerate interbeds.

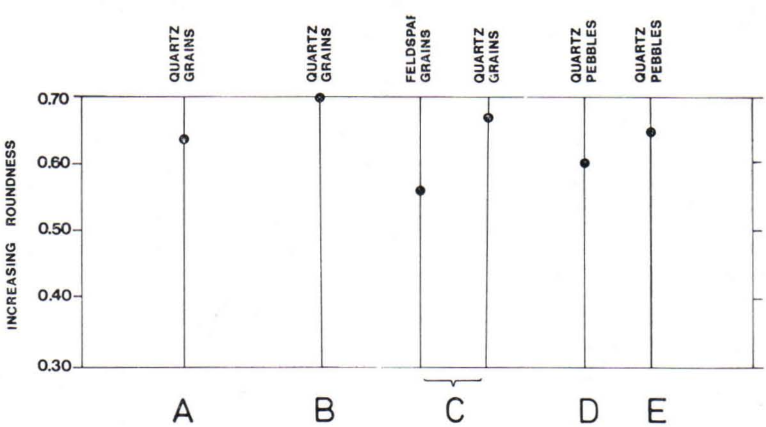
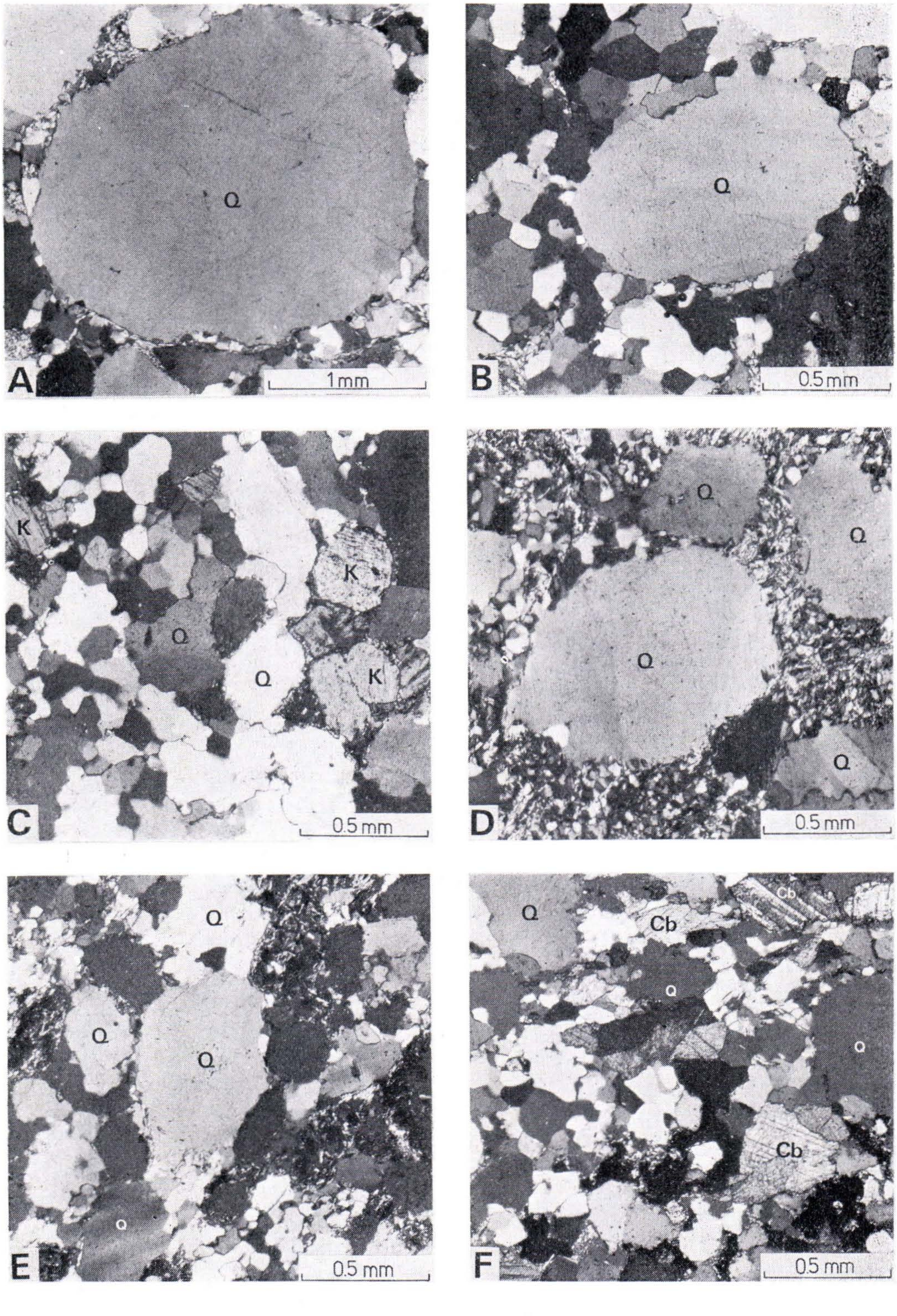


Fig. 48. Comparison of the average roundness of granular quartz and feldspar grains and quartz pebbles in the conglomerate interlayers in various quartzites of the Kiihtelysvaara Orthoquartzite Suite. A = sericite quartzite; B = orthoquartzite; C = feldspathic quartzite; D = conglomerate layer in Sericite quartzite; E = conglomerate layer in the lower part of the Upper quartzite.

population is composed of two subpopulations. Compared with the graphs given by Visher (1969), the shape of the curves indicates a littoral depositional environment.

Attempts to treat the rocks of the Upper quartzite Formation in the same way failed owing to the heterogeneity and more intense recrystallization of these rocks.


A
Q
1 mm
B
Q
0.5 mm
C
K
K
K
Q
Q
0.5 mm
D
Q
Q
Q
Q
0.5 mm
E
Q
Q
Q
Q
0.5 mm
F
Q
Cb
Cb
Q
Q
Cb
0.5 mm

Shape and roundness

Thin-sections show that the granular quartz grains in the Sericite quartzite, Orthoquartzite and Feldspathic quartzite are almost invariably elongated and ovoidal in shape. Also somewhat elongated are the feldspar grains of the Feldspathic quartzite. The granular quartz grains in the Upper quartzite, whose primary boundaries it was possible to determine despite recrystallization, are similar in roundness to those in the Sericite quartzite and Orthoquartzite.

Compared with the quartz grains in the arkosites (Fig. 38), the roundness of the granular grains in the Sericite quartzite, Orthoquartzite and Feldspathic quartzite (Fig. 48, A—C) exhibits a sudden jump. This is also seen in the pebbles, the individual quartz pebbles in the Upper arkosite (Fig. 38) being classified as rounded, and the quartz pebbles in the conglomerate interbeds in the Sericite quartzite as well-rounded (Fig. 48D). It is worth noticing that, in addition to well-rounded quartz grains, the feldspathic quartzite contains rounded potassium feldspar grains (Fig. 48C) and some only slightly rounded quartz grains. Hence, it seems that, during transport, »fresher» feldspar-bearing material was mixed with the quartz sands immediately before the volcanic eruptions.

The partial recrystallization of the granular quartz grains in the Upper quartzite prevented their roundness from being determined reliably, although, visually, they seem to be as rounded as those in Sericite quartzite or Orthoquartzite. The roundness of the quartz pebbles in the conglomerate interbeds of the Upper quartzite was successfully determined (Fig. 48E) as well-rounded.

The mineralogical character of orthoquartzites and related rocks

Sericite quartzite, orthoquartzite and feldspathic quartzite

The term sericite quartzite has been applied in a descriptive sense whenever the abundance of sericite exceeds 10 % although, according to the classification proposed by Pettijohn (1957), most of these rocks fall into the orthoquartzite group. The granular grains in the sericite quartzite are almost exclusively quartz (Table 9, Nos. 30—33). According to the terminology by Young (1976, p. 598), the quartz grains are undulatory lamellar or slightly polygonal, less often polycrystalline. Potassium feldspar occurs, but is very rare. The space between the granular quartz grains is filled with a matrix composed of authigenic sericite and quartz (Fig. 49A). Heavy minerals, such as tourmaline, zircon and rutile, are encountered in minor amounts.

Fig. 49. Photomicrographs of typical sand-sized grains in various quartzites.

A) Large rounded quartz (Q) grain and some smaller quartz grains cemented by quartz and sericite. Dusty borders mark outlines of original quartz grains. Sericite quartzite (Thin section No. 19618). Crossed nicols. B) Large quartz (Q) grain and some smaller quartz grains cemented mainly by quartz. Dusty borders mark outlines of original quartz grains. Orthoquartzite (Thin section No. 19257). Crossed nicols. C) Rounded quartz (Q) and potassium feldspar (K) grains cemented by quartz in crystallographic continuity with detrital quartz grains. Dusty borders mark outlines of original quartz and feldspar grains. Feldspathic quartzite (Thin section No. Ku 00302). Crossed nicols. D) Larger quartz grains (Q) in a matrix consisting mainly of quartz, sericite and chlorite. Upper quartzite (Thin section No. 17877). Crossed nicols. E) Impure, tuffaceous quartzite consisting mainly of quartz grains (Q) in fine-grained chloritic and sericitic matrix. Upper quartzite (Thin section No. 17874). Crossed nicols. F) Larger quartz grains (Q) in a matrix consisting mainly of carbonate (Cb), quartz and sericite. Upper quartzite (Thin section No. 17869). Crossed nicols. Photo by E. Halme.

Table 9
Mineral compositions of various quartzites in the Kiihtelysvaara Orthoquartzite Suite. Determined by point-counting

Minerals	30	31	32	33	34	35
Quartz	82.2	88.4	89.0	89.3	95.6	91.4
Plagioclase						
Potassium feldspar						
Biotite						
Sericite and chlorite	17.7 a)	11.5 a)	10.9 a)	10.6 a)	4.4 a)	8.6 a)
Carbonate						
Apatite	+					
Titanite or rutile			+	+		
Zircon	+	+	+	+	+	+
Tourmaline	+	+	+	+		
Andalusite						
Opaque	+		+			
Rock fragments						
Total[b]	100.0	100.0	100.0	100.0	100.0	100.0

+) detected minor constituent; a) mainly sericite; b) after addition of all minor constituents.

30. Sericite quartzite (drill core No. R307/4232/146.20), Sääperi, Värtsilä.
31. Sericite quartzite (Specimen No. 30C/LJP—75), Kangasvaara, Tohmajärvi.
32. Sericite quartzite (drill core No. R321/4241/23.45), Kortevaara, Kiihtelysvaara.
33. Sericite quartzite (Specimen No. 3S/MS—74), Särkilampi, Kiihtelysvaara.
34. Orthoquartzite (Specimen No. 4S/MS—74), Särkilampi, Kiihtelysvaara.
35. Orthoquartzite (Specimen No. 5S/MS—74), Särkilampi, Kiihtelysvaara.
36. Orthoquartzite (Specimen No. 6S/MS—74), Särkilampi, Kiihtelysvaara.
37. Orthoquartzite (Specimen No. 7S/MS—74), Särkilampi, Kiihtelysvaara.
38. Orthoquartzite (Specimen No. 8S/MS—74), Särkilampi, Kiihtelysvaara.
39. Feldspathic quartzite (Specimen No. 115/PV—73), Raatevaara, Kiihtelysvaara.

Upwards in the sequence the amount of sericite in the matrix between the granular quartz grains decreases gradually. Nevertheless, even the purest types contain some sericite (Table 9, Nos. 34—37), in which case the granular quartz grains are mainly embedded in authigenic quartz (Fig. 49B) and the rocks are fairly typical orthoquartzites. As in the Sericite quartzite, the granular quartz grains in the Orthoquartzite are undulatory lamellar or mildly polygonal and represent, according to Young (1976), very slightly metamorphosed quartz. Zircon is the predominant heavy mineral; tourmaline and rutile are less common. In the upper part of this quartzite bed the abundance of sericite begins to rise once more (Table 9, No. 38).

At Raatevaara, the predominant granular grains in the Feldspathic quartzite are quartz and potassium feldspar (Table 9, No. 39 and Fig. 49C). Most of the quartz grains are similar to those in the Orthoquartzite, but some are less rounded. The grains of potassium feldspar are comparatively fresh. The spaces between the granular grains are mainly filled with quartz matrix. A sign of the increase in the degree of metamorphism westwards are the small andalusite porphyroblasts occasionally encountered.

Upper quartzite

The rocks that belong to the Upper quartzite Formation are not pure quartz sands in origin but are mixed with volcanic weathering products and detritus (Table 9, Nos. 40—44; Fig. 49E). The granular grains are almost exclusively quartz; potassium feldspar plagioclase and volcanic fragments are rare. In general, the quartz method.

36	37	38	39	40	41	42	43	44
98.0	94.2	87.0	88.8	70.5	76.8	78.1	87.1	61.3
			+	+	+	+	+	+
			10.8					+
				+	+	+	+	+
2.0 a)	5.8 a)	12.9 a)	0.2 a)	25.2	23.0	21.8	12.7	25.7
							+	12.9
				+				
			+	+	+	+		
+	+	+	+	+	+	+	+	
			+	4.1	+	+	+	+
			+			+		
			+	+	+	+	+	+
				+	+			
100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

40. Upper quartzite, about 10 m above the Volcanite Formation (drill core No. R308/4241/123.00), Viistola, Kiihtelysvaara.
41. Upper quartzite, about 20 m above the Volcanite Formation (drill core No. R308/4241/110.50), Viistola, Kiihtelysvaara.
42. Upper quartzite, about 80 m above the Volcanite Formation (drill core No. R309/4241/186.20), Viistola, Kiihtelysvaara.
43. Upper quartzite, about 190 m above the Volcanite Formation (drill core No. R309/4241/77.45), Viistola, Kiihtelysvaara.
44. Carbonate-bearing upper quartzite, about 220 m above the Volcanite Formation (drill core No. R309/4241/53.20), Viistola, Kiihtelysvaara.

grains are undulatory lamellar or polygonal, less often polycrystalline (Fig. 49D). They are clearly more intensely deformed than the quartz grains in Sericite quartzite or Orthoquartzite. The interstices between the granular quartz grains are filled with a matrix of authigenic chlorite, sericite and quartz (Nos. 40—43). The uppermost layers also contain carbonate (No. 44; Fig. 49E). Heavy minerals, such as tourmaline and detrital magnetite and pyrite from volcanic weathering products, have accumulated in some layers (No. 40).

Quartz conglomerate interbeds

In the study area quartz conglomerates occur on three quartzite horizons: a) as interbeds in the lower part of Sericite quartzite; b) as interbeds in the upper part of Orthoquartzite; and c) as interbeds in the lower part of Upper quartzite. They are much alike in mineral composition but differ in detail.

Most of the pebbles in the conglomerate interbeds in the Sericite quartzite are of pale or pinkish polygonal quartz although bluish, mildly lamellar quartz is also abundant. Pale, polycrystalline quartz and sparse fuchsite-bearing quartzite are also present. A bluish, Mustalampi type of quartz (p. 51) has also been encountered. The quartz-sandy matrix is rich in sericite and often tinged red by hematite pigment. Small amounts of heavy minerals, such as tourmaline, rutile and zircon, have been concentrated in some layers. In the Sääperi area some of the layers are radioactive owing to the Th-U minerals thorite and brockite and their metamictic decomposition products.

The minor conglomerate interbeds in the upper part of the Orthoquartzite are largely

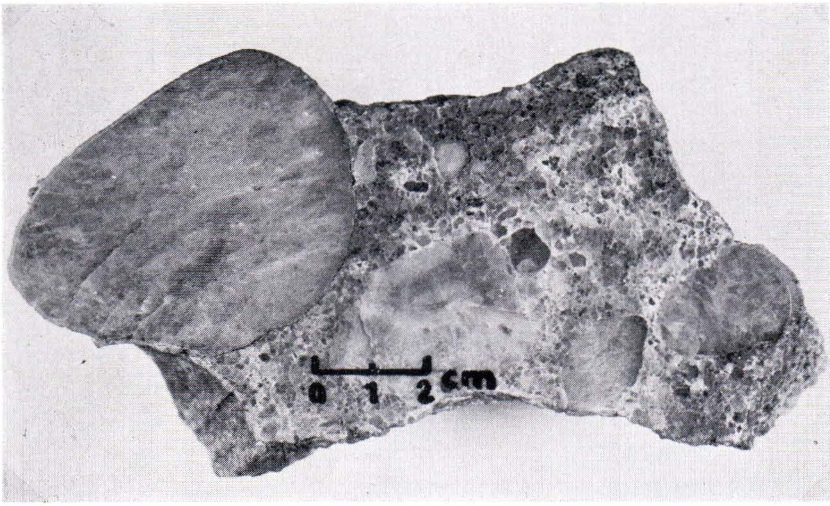


Fig. 50. Well-rounded pebbles and cobbles of quartz from a quartz conglomerate interbed, Upper quartzite Formation. Raatevaara, Kiihtelysvaara. Photo by E. Halme.

composed of the above constituents except that the U-Th minerals are rare and occur only sporadically.

Most of the pebbles and cobbles in the conglomerate interbeds in the Upper quartzite (Figs. 14 and 50) are pinkish, greyish or bluish quartz as in the interbeds in the Sericite quartzite, except that the quartz is more intensely deformed in the Upper quartzite. Also present are pebbles of white, polycrystalline quartz, fine-grained sinter-like quartz and fragments from the Volcanite Formation. The quartz sand matrix is fairly rich in sericite and chlorite and is often pinkish in colour owing to the hematite pigment. Some layers contain small lenses of hematite, which is sporadically the predominant constituent in the matrix between the quartz pebbles (Fig. 14). The matrix of the conglomerate is often enriched in heavy minerals. These seem to form two populations: the older detrital population consisting of tourmaline, rutile and zircon, and the younger detrital population characterized by magnetite, pyrite, titanite and apatite. The latter may derive from disintegrated rocks of the Volcanite Formation. A few radioactive layers encountered in the Upper quartzite in the Viistola section contain small amounts of brockite and thorite and their decomposition products.

Clay slate and tuffite interbeds

The clay slate interbeds in the Sericite quartzite (Fig. 45B) are so fine-grained that it is not easy to identify the mineral species. Nevertheless, the major minerals are sericite and quartz; tourmaline is also fairly common.

In contrast, the clay slate interbeds in the Upper quartzite (Fig. 45D) are more distinctly laminated, containing silt lamellae and volcanic weathering products. Their major minerals include sericite, chlorite, quartz and tourmaline. The silt lamellae contain the same minerals but more quartz.

The quartzites in the Upper quartzite Formation contain local tuffitic interlayers (Fig. 16). The proportions of volcanic material and of the concurrently deposited quartz sand vary and hence the mineral composition also varies. Tuffitic interlayers are composed mainly of chlorite, biotite, quartz, sericite and magnetite; intensely altered albite grains have also been encountered.

Phosphate-bearing mottles and pebbles, and Mn-rich nodules

At Raatevaara, some layers in the Upper quartzite show abundant pinkish mottles 1 to

3 mm in diameter (p. 32). They are andalusite grains surrounded by a seam tinted pink by hematite. The core of the andalusite consists of fine-grained matter in which lazulite and piedmontite, to name but two minerals, have been identified.

Here and there in the Upper quartzite at Viistola (Fig. 15) phosphate-bearing »pebbles» enveloped by a brownish-red hematite seam are found in a corresponding position. They often occur immediately above the quartz conglomerate interbeds. The core of the nodule-like »pebbles», 1 to 2 cm in diameter, is composed of rounded quartz grains and lazulite matrix. According to point-counting analysis, quartz assays 72.6 %, lazulite 27.3 %, and the rest (tourmaline, rutile, hematite) 0.1 %. The blue lazulite is strongly pleochroic (X = colourless, Z = azure-blue), optically negative, often twinned with $\alpha = 1.623$ and $\gamma = 1.660$.

Dark-brown Mn-rich nodules irregular in shape occur here and there throughout the transitional zone between the Upper quartzite and dolomite. According to X-ray determination (P. Kallio) and electron microprobe analysis (T. Paasivirta), the nodules are composed of manganite and minor hematite.

The chemical character of orthoquartzites and related rocks

The Särkilampi—Viistola reference series

Analytical data on samples collected from the rock units across the Kiihtelysvaara Orthoquartzite Suite in the Särkilampi—Viistola area give a general idea of the variation in chemical composition. The material, which consists of drill-core samples and hand specimens from exposures, was analysed by emission spectrophotometer (G. Wansén) for SiO_2, Al_2O_3, MgO, CaO, Na_2O, K_2O, Ti and V (semiquantitatively). The total iron, Mn, Cu, Co, Ni, Pb and Zn were determined by AAS and sulphur by a Leco M621 sulphur analyser (A. Nurmi). Phosphorus was analysed in the ore laboratory. Trace elements, and sulphur and phosphorus abundances typical of quartzites are given later in this paper (Fig. 58). The variation in chemical composition exhibits a certain regularity (Fig. 51):

- SiO_2 is invariably high in the Lower quartzite, but highest in the Orthoquartzite; the volcanites cause the curve to decline; thereafter the Upper quartzite causes it to rise but so that the abundances are clearly below the level of the Sericite quartzite and Orthoquartzite.
- Al_2O_3 responds fairly well to the variation in mica abundances and partly also to that in chlorite (± feldspar) in the different layers; the abundances are distinctly higher in the Sericite quartzite and Upper quartzite than in the Orthoquartzite.
- CaO indicates more or less directly the carbonate abundance; the curve shows that the Upper quartzite grades into rocks of the Dolomite—Volcanite Formation.
- MgO largely follows CaO as a result of the dolomitic carbonate; the biotite and chlorite abundances are also reflected in the shape of the curve.
- The total iron, calculated to Fe_2O_3, is low throughout, but rises at the volcanic beds (in places also in the quartz conglomerate beds, p. 88).
- Na_2O is persistently rather low, but shows a significant rise at the volcanic beds.

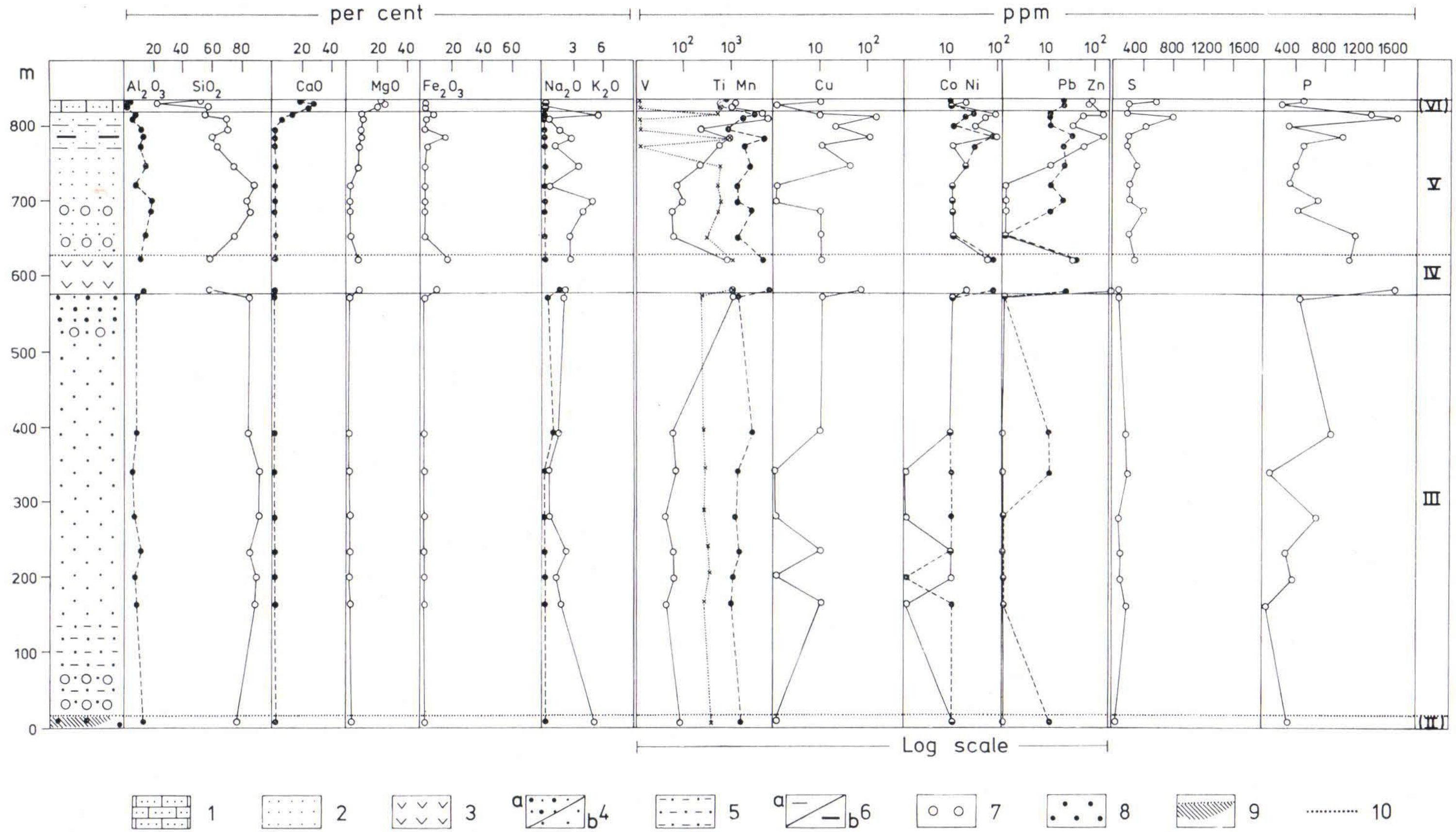


Fig. 51. Schematic diagram showing variation in chemical composition across the Kiihtelysvaara Orthoquartzite Suite in the Särkilampi—Viistola area. Lithostratigraphic units: II = Arkosite Formation; III = Lower quartzite Formation; IV = Volcanite Formation; V = Upper quartzite Formation; VI = Dolomite—Volcanite Formation. Symbols: 1 = quartz-bearing dolomite; 2 = upper quartzite; 3 = metavolcanite; 4 a = feldspathic quartzite; 4 b = orthoquartzite; 5 = sericite quartzite; 6 a = clay slate intercalations; 6 b = volcanic intercalations; 7 = quartz conglomerate interbeds; 8 = upper arkosite; 9 = younger (Early Jatulian) weathering crust; 10 = unconformity.

— K_2O is largely sympathetic with the Al_2O_3 curve and mainly reflects the variation in the abundance of micas.
— The shape of the Mn abundance curve is governed to a great extent by CaO and MgO; the sharp rise at the end of the curve is due to the manganese nodule-bearing layers in the Upper quartzite.
— Ti and V are distinctly higher in volcanic and clay slate layers.
— Cu, Co and Ni curves are similar in shape but differ in detail; the highest abundances are attributed to the volcanic beds.
— Pb and Zn curves resemble each other; the highest abundances coincide with the volcanic beds.
— The sulphur abundance tends to be persistently low but rises clearly at the volcanic beds.
— The level of phosphorus is low throughout but shows some fluctuation at the quartz conglomerate beds; the highest abundances are encountered in volcanic and clay slate beds.

Two whole-rock analyses

Two chemical analyses of the quartzites from the Lower quartzite Formation (Table 10, Nos. 12 and 13) reflect rather well the mineral composition of these rocks. The first analysis (No. 12) refers to sericite quartzite, the second (No. 13) to orthoquartzite, even though the abundance of sericite in the latter is not much below that in the former. If these data are compared with those given by Pettijohn (1975, p. 233), it is clear that both analyses represent rather typical orthoquartzites. The analyses are also very similar to those of Lower Jatulian quartzites from Soviet Karelia (Sokolov *et al.* 1970, pp. 109—114). Owing to their sericite content, the quartzites in the study area typically contain significant amounts of Al_2O_3 and K_2O in addition to the main consituent SiO_2; the abundances of the other constituents are, however, low.

Th-(U)-bearing quartz conglomerate interbeds

Radioactive, Th-(U)-bearing quartz conglomerate interbeds occur in the study area as a) interbeds in Sericite quartzite, and b) interbeds in Upper quartzite. This radioactivity is due to Th-U minerals, e.g. detrital thorite and authigenic brockite, and their metamictic decomposition products. Both these minerals are predominantly Th minerals, although they also contain small amounts of uranium. Gamma-spectral analyses of the 15 samples collected from the conglomerate interbeds in the Sericite-quartzite in the Sääperi area average Th = 1960 ppm and U = 123 ppm (Räisänen 1975). The corresponding average for the four samples from the conglomerate interbeds of the Upper quartzite at Viistola are Th = 450 ppm and U = 2 ppm. Although the rock contains brockite, a phosphate mineral, its abundance

Table 10
Chemical composition of two quartzites in the Lower quartzite Formation (weight per cent).

	12	13
SiO_2	95.72	96.30
TiO_2	0.04	0.05
Al_2O_3	2.88	2.34
Fe_2O_3	0.02	0.12
FeO	0.05	0.01
MnO	0.00	0.00
MgO	0.07	0.03
CaO	0.11	0.15
Na_2O	0.01	0.08
K_2O	0.86	0.71
P_2O_5	0.04	0.01
CO_2	0.00	0.00
H_2O^+	0.31	0.32
H_2O^-	0.01	0.02
	100.12	100.14

12. Sericite quartzite, Patsola, Värtsilä. Analyst: A Heikkinen (Nykänen 1968, p. 43).
13. Orthoquartzite, Hyypiä, Kiihtelysvaara. Analyst: P. Ojanperä (Nykänen 1971b, p. 35).

is so low that it hardly affects the total phosphorus content in the rock (P = 0.1 %). These beds also exhibit a slightly elevated iron content (about 1 % Fe), which in some beds has developed into a hematite matrix (Fig. 14).

Phosphate-bearing mottles and pebbles, and Mn-rich nodules

The layers in the Upper quartzite at Raatevaara contain Al-rich nodule-like andalusite porphyroblasts; chemical analysis shows abnormal Sr, Ba, P and S abundances (Table 11, No. 14). Andalusite concentrate separated from the rock demonstrates that especially strontium (3.23 %) and phosphorus (1.72 %) (Table 11, No. 15) are enriched. A small portion of the phosphorus is incorporated in lazulite, but the bulk into other phosphate minerals. Electron microprobe analysis (by T. Paasivirta) showed that Sr is incorporated in a complex Sr-phosphate-sulphate, possibly svanbergite, and Ba possibly in barite and celestite. In a fine-grained material, however, these minerals are not easy to identify with certainty.

The nodule-like quartz-lazulite pebbles that occur in a corresponding stratigraphic position in the Upper quartzite at Viistola also contain appreciable phosphorus. Analysis of such a pebble (Table 11, No. 16) shows that it contains 3.57 % P, but low abundances of Sr, Ba and S. A semiquantitative electron microprobe analysis of a lazulite from the same aggregate (T. Paasivirta) indicates $Al_2O_3 = 33.0$, $FeO = 2.8$, $MgO = 11.9$ and $P_2O_5 = 40$ wt%.

Abnormal Mn abundances have been noted in the zone where the Upper quartzite grades into dolomites. The manganese is not, however, evenly distributed but is concentrated into the manganese-rich nodules that occur here and there. Analysis of such a nodule (Table 11, No. 17) demonstrates that it is rich in manganese (Mn 44 %) but comparatively poor in iron (Fe 1.83 %). The bulk of the Mn is apparently incorporated in manganite [MnO (OH)], and most of the Fe in hematite.

Table 11

Partial chemical analyses of »nodular» quartzite and phosphate- and manganese-bearing nodules in the Upper quartzite Formation (weight per cent). Analysts: V. Hoffrén, P. Ojanperä and P. Väänänen.

	14	15	16	17
Fe	0.60	0.58		1.83
Mn				44.00
V				0.01
Sr	0.67	3.23		0.007
Ba	0.15	0.22		0.35
P	0.36	1.72	3.57	0.0
S	0.27	0.75		0.07

14. »Nodular» quartzite, about 80 m above the Volcanite Formation, Raatevaara, Kiihtelysvaara (Specimen No. 123/PV—73).
15. Phosphate-bearing »nodules» from nodular quartzite, about 80 m above the Volcanite Formation, Raatevaara, Kiihtelysvaara (Specimen No. 31/OA—61).
16. Phosphate »pebble» from a tuffaceous quartzite, about 50 m above the Volcanite Formation, Viistola, Kiihtelysvaara (Specimen No. 23/JJK—74).
17. Manganese nodule from a carbonate-cemented quartzite, Viistola, Kiihtelysvaara (drill core No. R309/4241/75.60).

Conclusions concerning the depositional environment and origin of the material

The transgression of the sea far onto the mainland from the southwest or west presumably brought a turn in the Jatulian sedimentation. Derived from arkose sands by the action of chemical weathering, weathering products rich in quartz and clay minerals were transported, reworked, sorted and deposited at the margins of the basin thus formed. Bedding structures suggest that initially the material arrived cyclically and deposited in shallow littoral waters. Mainly cross-bedded sands (Sericite quartzite Member) containing minor clay minerals were

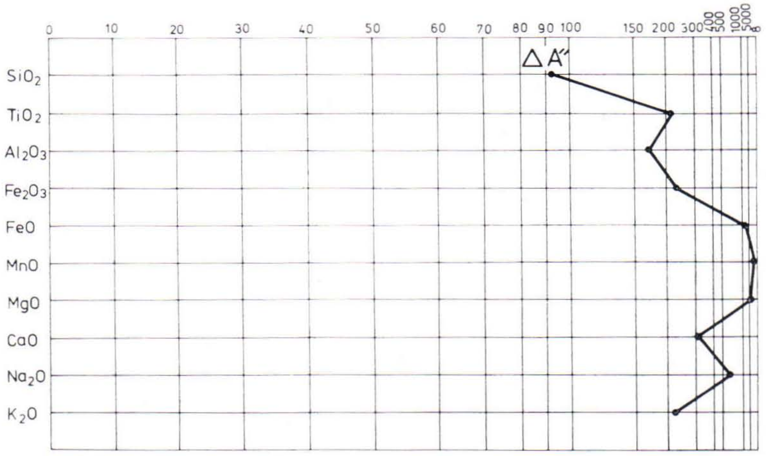


Fig. 52. Relative losses and gains of oxide constituents (△ A″) in the working of an arkosite with composition A′ resulting in the formation of orthoquartzite A″.

deposited near the shore; now and then quartz pebbles were also deposited and formed their own conglomeratic beds. At the end of the cycle the occasional thin clayey layer was deposited on the sandy layer. Eventually the sandy matter transported to the littoral zone came to contain less and less clay and quartz pebbles, and the formation of intercalations ceased.

The quartz sands that were deposited later were more washed and better sorted (Orthoquartzite Member). Despite the absence of intercalations, other factors are present suggesting that the sands were deposited in shallow littoral waters and not on the mainland, e.g. ripple marks, subhorizontal cross-bed laminae (15 ° to 20 °), larger grain size and less developed sorting of the material than in dune sands (Visher 1969, Füchtbauer and Müller 1970), and grain-size distribution. Quartz conglomerate interbeds and feldspathic layers (Feldspathic quartzite Member) in the upper part of the Orthoquartzite deposit suggest that at some later date transgression turned into regression.

The roundness of the granular quartz grains in the Sericite quartzite and Orthoquartzite (Fig. 48) changes more suddenly than in the arkosites (Fig. 38); whether the roundness was due to wind or water can no longer be deduced. It is likely that the transportation included aeolian episodes, whose importance for the rounding of the mineral grains has been stressed by several investigators. In some pink quartzite beds a hematite crust envelops the quartz grains, a feature that some authors maintain suggests desert conditions. A film like this, however, tends to form eventually under the most diverse conditions and hence, it is not a valid criterion of type of climate (Folk 1976, p. 606).

In internal structure, the quartz in the Sericite quartzite and Orthoquartzite (p. 85) is similar to that in the arkosites (pp. 71 and 72), indicating that they probably derived from different resistates of the same primary source. If we assume that the composition of the Upper arkosite given in Table 8 (No. 11) represents the composition of a resistate (A′) and that the composition of the Orthoquartzite listed in Table 10 (No. 13) illustrates the composition of an intensely reworked resistate (A″), then the relative losses and gains (△ A″) of the chemical constituents can be determined on the basis of a »gain and loss» diagram (Fig. 52). In relation to SiO_2, the following order of loss is obtained: MnO, MgO, FeO, Na_2O, CaO, Fe_2O_3, K_2O, TiO_2 and Al_2O_3. This is consistent with the facts obtained from the sediments in terms of their increasing degree of maturity. Chemical weathering, due to which all minerals except quartz were decomposed, indicates that the paleoclimate was not completely arid.

Later on, the deposition of the quartz sands was interrupted by a period of volcanic activity. During marine regression the quartz sand piles that were already somewhat consolidated rose partly above sea level, and the lavas discharged mainly on top of the quartz sand piles on the mainland (p. 125). The lava beds that were above sea level were submitted to erosion and seem to have been substantially denuded.

Eventually the marine regression switched once more to transgression, and the volcanite beds were gradually covered by quartz sands transported from the mainland (Upper quartzite Formation). The quartz sands that first deposited were rather coarse, containing material from decomposed volcanites (Fig. 16). The tuffite interbeds in the quartzite suggest that slight volcanic activity occurred from time to time. The coarse cross-bedding and numerous quartz conglomerate interbeds in the psammitic sediments of the lower part of the Upper quartzite Formation show that the material was transported to the littoral zone by strong currents.

The finer grain size of the quartz sands that deposited later, the gradual increase in silt intercalations in them and the content of carbonate in the upper part all indicate that deposition took place in continuously deepening water. The Mn-rich nodules in the upper layers also refer to deepening water; according to Strakhov (1970, p. 207), their formation requires not only comparatively deep water but also fairly high pH (over 8) and suitable redox potential (Eh).

The regularity of the quartz conglomerate beds and the heavy-mineral accumulations in both the Lower and Upper quartzite Formations suggest an ancient littoral gravel. At least two alternative modes of formation can be suggested for the hematite that occurs in the matrix between quartz pebbles: a) precipitation from iron-bearing solutions that circulated in the pores of the rock; b) primary precipitates. The quartz pebbles entirely enveloped by hematite in some layers favour the latter alternative. The precipitation of iron seems to have affected, at least indirectly, the formation of Th-U phosphates (cf. Piirainen 1968). The phosphate-bearing nodules and »pebbles» (pp. 88 and 89) that are locally associated with quartz conglomerate interbeds in the Upper quartzite were probably originally Al-hydroxide aggregates rounded by waves that had absorbed Sr, Ba, phosphates, sulphates and iron hydroxide (nodules) and to which quartz grains (pebbles) had become attached.

THE KIIHTELYSVAARA CARBONATE—BLACK SLATE SUITE

During the »more marine» stage that succeeded the deposition of quartz sands, the rate of deposition of sands in the study area slowed down significantly or even came to an end and, conditions changed to favour the formation of carbonates. Thereafter, the configuration of the margin of the basin appears to have altered and clastic sediments deposited from time to time. Later, when conditions changed yet once again, the principal constituents to deposit were carbonaceous clays and some carbonates. The alternations in the configuration of the margins of the basin reflected movements deeper down in the crust, which were occasionally followed by volcanic activity. Later, some of the partly consolidated piles of sediments and associated volcanites rose from the sea and had time to undergo substantial erosion before the flysch sediments deposited. Only much later, during the Svecokarelidic orogeny (about 1800 to 1950 Ma), were the rocks of this association (Marine Jatulian, Väyrynen 1933, p. 64), which in this content are called the Kiihtelysvaara Carbonate—Black Slate Suite, metamorphosed into dolomites, carbonaceous slates or graphite-bearing phyllites. The associated volcanics altered into metavolcanites and metadiabases.

The associations and their mode of occurrence

In the study area the quartzites of the Upper quartzite Formation are fairly regularly succeeded by a deposit of varying thickness of dolomitic carbonate rocks, carbonaceous slates and volcanics. This deposit is often rather thin, although in some areas, e.g. the Viistola section and Haluksenlammit, the associated formations show a manyfold increase in thickness.

In the fully developed Jatulian sequence the rocks of the Upper quartzite Formation are succeeded by a bed of dolomitic carbonate sediments and volcanic rocks; in the present context, this bed is called the Dolomite—Volcanite Formation. The carbonate rocks in the lower part of this formation are rather heterogeneous containing abundant thin quartzite and clay slate intercalations and some tuff interbeds. The rocks in the upper part are more homogeneous; even so, chert nodules have been observed in some beds. In the Viistola section (Fig. 15), the dolomite layers are overlain by a volcanite bed (the Volcanite Member) about 10 m thick that has not been encountered in the Haluksenlammit area, where the dolomite layers are cut by numerous metadiabase dykes.

In the Viistola section (Fig. 15) and the Haluksenlammit area (Fig. 19A) the Dolomite—Volcanite Formation is succeeded by the rather thin Hematite rock—Quartzite Formation. In the Viistola section a thin layer of brecciated cherty dolomite rests on the volcanite bed and is succeeded by an almost compact hematite layer about one metre thick. In the Haluksenlammit area the hematite layer is from one to one and a half metres in thickness and rests either directly on the dolomite or on a thin layer of ferriferous slate. In both places the hematite layer is overlain by a ferriferous slate whose thickness varies from a few centimetres to two metres. It is succeeded regularly by a quartzite layer 3 to 5 metres thick. These restricted lens-like occurrences extend for a distance of a half to one kilometre.

In the Viistola section and the Haluksenlammit area the rocks of the Hematite rock—Quartzite Formation are succeeded by a carbonaceous slate deposit whose lower and upper parts contain abundant dolomite layers and the topmost part also volcanics. This rock association is called the Dolomite—Carbonaceous slate—Volcanite Formation. Lowermost in the fully developed formation is the Dolomite Member, which consists of layered dolomites. It is succeeded by the Carbonaceous slate Member composed of carbonaceous slates or graphite-bearing phyllites; this in turn is overlain by the Alternating dolomite—carbonaceous slate Member. Closely associated with the latter are mafic dykes and sills and minor extrusive beds (the Volcanite Member). The largest hypabyssal instrusions are the Hyypiä and Sääperi sills, the former being almost 100 m and the latter some 200 m wide. Erratic blocks (Fig. 21) indicate that dolomites and carbonaceous slates occur in the Sääperi area as well, although, for lack of outcrops, the occurrences are not known in detail.

The total thickness of the Kiihtelysvaara Carbonate—Black slate Suite varies greatly from one area to the next, depending on the primary depositional thickness and on the depth to which these rocks were eroded before the flysch sediments deposited. The Dolomite—Volcanite

Formation and the Dolomite—Carbonaceous slate—Volcanite Formation are seldom as well developed as in the Viistola, Haluksenlammit and Hyypiä areas, where the Dolomite—Volcanite Formation is locally as much as 70 metres and the Dolomite—Carbonaceous slate—Volcanite Formation from 150 to 170 m thick. In many places these formations have been reduced into a deposit of alternating dolomite and carbonaceous slate layers with some minor volcanics and whose thickness is no more than a few metres, or ten metres at the most. Northwest of Hyypiä in particular, the total thickness of this deposit declines abruptly, and north of Kortevaara it possibly vanishes altogether.

Primary and other sedimentary structures

Bedding

Owing to recrystallization, primary bedding structures are seldom visible in the carbonate rocks (Fig. 53A) of the Dolomite—Volcanite Formation. Close to the quartzite intercalations in the lower part of the formation, however, slightly developed cross-bedding is discernible. The cross-bedding is best developed in the quartzite interbeds. The clay slate intercalations exhibit indistinct lamination.

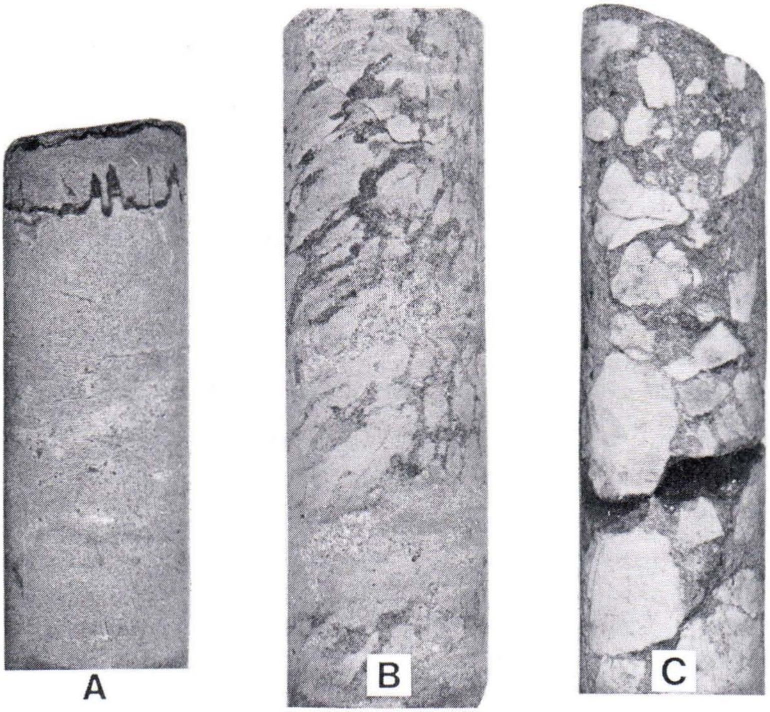


Fig. 53. Drill-core samples from rocks typical of the lower section of the Kiihtelysvaara Carbonate—Black slate Suite. A) Massive dolomite from the lower part of the Dolomite—Volcanite Formation; note the stylolite seam at the upper end of the core (drill core No. R310/4241/189.70). B) Calcareous nodular chert from the lower part of the Dolomite—Volcanite Formation; the elongated chert nodules 1 to 3 cm long constitute a three-dimensional network in a calcitic matrix (drill core No. R310/4241/186.25). C) Brecciated calcareous nodular chert; the matrix between the chert fragments predominantly calcite (drill core No. R310/4241/148.85). All the samples from the Viistola section at Kiihtelysvaara. Drill cores about 32 mm in diameter. Photo by E. Halme.

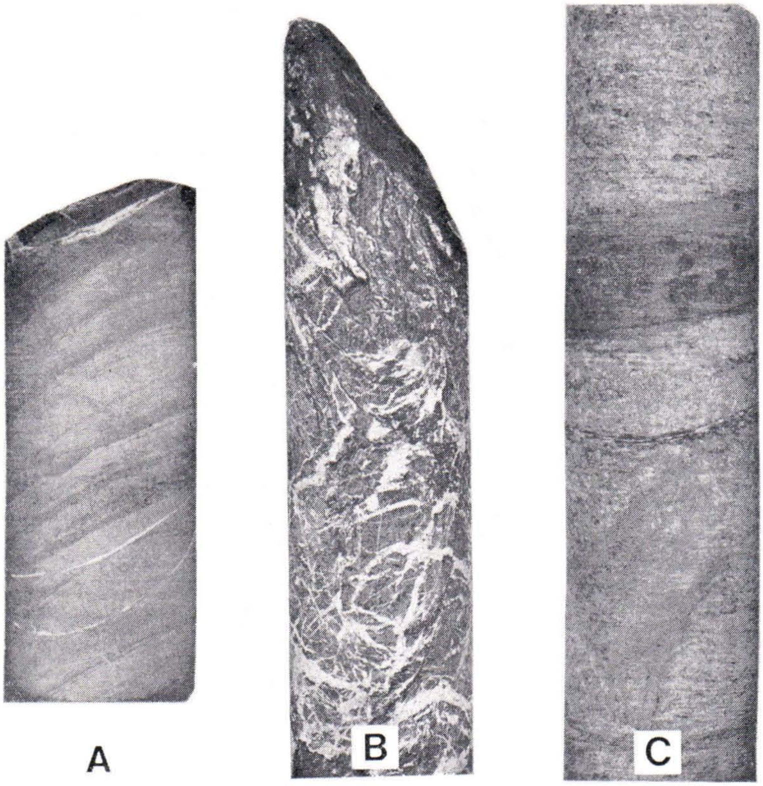


Fig. 54. Drill-core samples from rocks typical of the upper section of the Kiihtelysvaara Carbonate—Black slate Suite. A) Fine-grained lameller dolomite from the lower part of the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R310/4241/99.00). B) Sheared carbonaceous slate with abundant fissures filled with calcite from the middle part of the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R310/4241/68.20). C) Alternating tremolite-bearing dolomite layers (grey) and carbonaceous slate layers (dark grey) from the upper part of the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R310/4241/147.55). All the samples from the Viistola section at Kiihtelysvaara. Drill cores about 32 mm in diameter. Photo by E. Halme.

The hematite layer from one metre to one and a half metres thick in the Hematite rock—Quartzite Formation shows vague bedding on account of the variation in abundance of the argillaceous matter. Red and greenish laminae alternate in a ferriferous slate associated with the hematite fayer, whereas the quartzite that succeeds the ferriferous slate exhibits cross-bedding.

In the lower part of the Dolomite—Carbonaceous slate—Volcanite Formation, dolomitic layers alternate with very thin clay slate layers (Dolomite Member; Fig. 54A). Both dolomite and clay slate show thin alternating laminae: in the dolomite pinkish and greenish and in the clay slate greyish and greenish. The carbonaceous slate (Carbonaceous slate Member) that succeeds the bedded dolomite also exhibits laminar structure; it is, however, seldom visible owing to the dark hue of the rock and tectonization (Fig. 54B).

In the upper part of the Dolomite—Carbonaceous slate—Volcanite Formation the thicker greenish dolomite layers alternate with thinner layers of carbonaceous slate (Alternating dolomite—carbonaceous slate Member; Fig. 54C). They include some random and very thin arkosite layers.

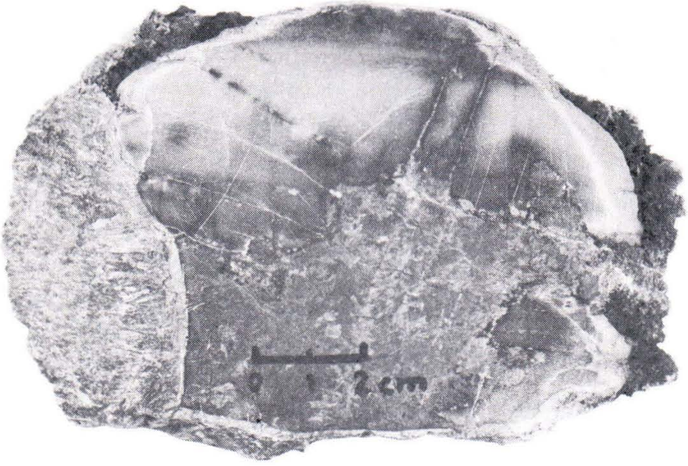


Fig. 55. A chert nodule in tremolite-bearing dolomite. Erratic block near Silmunvaara, Kiihtelysvaara. Photo by E. Halme.

Cherty layers

Chert nodules occur here and there in the carbonate rocks of the Dolomite—Volcanite Formation and in the rocks of the lower part of the Dolomite—Carbonaceous slate—Volcanite Formation (Fig. 15). The elongated nodules are 1 to 3 cm long and constitute a rather dense network (Fig. 53B) whose orientation may be close to the primary bedding. Larger individual nodules (Fig. 55) have been met with in places, e.g. in the Haluksenlammit area. The layers with chert nodules seem to have been readily brecciated, and hence, a structure reminiscent of conglomerate has formed (Fig. 53C).

Stylolite seams

Tooth-like stylolite seams have been encountered at intervals of 10 to 50 cm (Fig. 53A) in many places in the purest carbonate rocks of the Dolomite—Volcanite Formation. The amplitude of the stylolites varies from one mm to 2 cm. Stylolite seams have also been met with, even though not so often, in the dolomites resting on the hematite-rock-quartzite bed. It appears that the stylolite seams occur in those parts of the carbonate layers that were most intensely recrystallized. The mode of occurrence refers to dissolution structures.

Grain-like microfossils

Tynni (1971) has identified grain-like microfossils about 30 μ in diameter from slides of carbonate rocks in the Haluksenlammit area. These microfossils, which he has interpreted as algae, have been found in carbonate rocks both above and below the hematite rock—quartzite layer as well as in carbonate rock xenoliths in metadiabases. Preliminary studies indicate that they also occur in an analogous stratigraphic position in carbonate rocks in the Viistola section (Tynni, personal communication). Stromatolitic bedding, which is attributed to algae and which has been found in Jatulian dolomites e.g. in the Kemi area (Härme and Perttunen 1963), has not been encountered. In the Sääperi area, however, some dolomite erratic blocks have been encountered that slightly resemble stromatolite bedding in structure (cf. Hausen 1930, p. 77).

The mineralogical character of carbonate sediments, carbonaceous slates and related rocks

Rocks of the Dolomite—Volcanite Formation

The quartzites grade into dolomites; hence, the lowest layers in the Dolomite—Volcanite Formation contain small amounts of detrital quartz and minor sericite and chlorite in addition to carbonate (Table 12, No. 45; Fig. 56A). The rocks in the upper part of the formation are somewhat purer carbonate rocks (Nos. 46 and 47; Fig. 56B). Electron microprobe analyses (T. Paasivirta) show that the carbonates are predominantly dolomite, although calcitic bodies and veins are also common. In the upper layers of the deposit the quartz occurs as tiny mottles that structurally resemble chert nodules. These nodules, which are most abundant in the middle of the deposit, exhibit a net-like structure in places (Fig. 53B), although they also occur locally as individual nodules (Fig. 55). The hue of the nodules varies from white to grey, the core being paler than the margins. In the net-like types the spaces between the nodules are filled with calcitic matrix, the bulk of the rock consisting of nodules (No. 51). The nodules are composed of very fine-grained quartz, although locally more coarse-grained and more intensely recrystallized portions and small carbonate mottles occur. In the more intensely metamorphosed rocks, e.g. those in the Haluksenlammit area, the dolomite layers contain some tremolite and talc.

The quartzite interbeds in the lower part of the Dolomite—Volcanite Formation are predominantly composed of quartz and carbonate grains and carbonate matrix (calcarenacous orthoquartzite). Potassium feldspar has also been encountered occasionally. Micaceous clay minerals, sericite, chlorite and quartz have been identified in the clay slate intercalations in the dolomite.

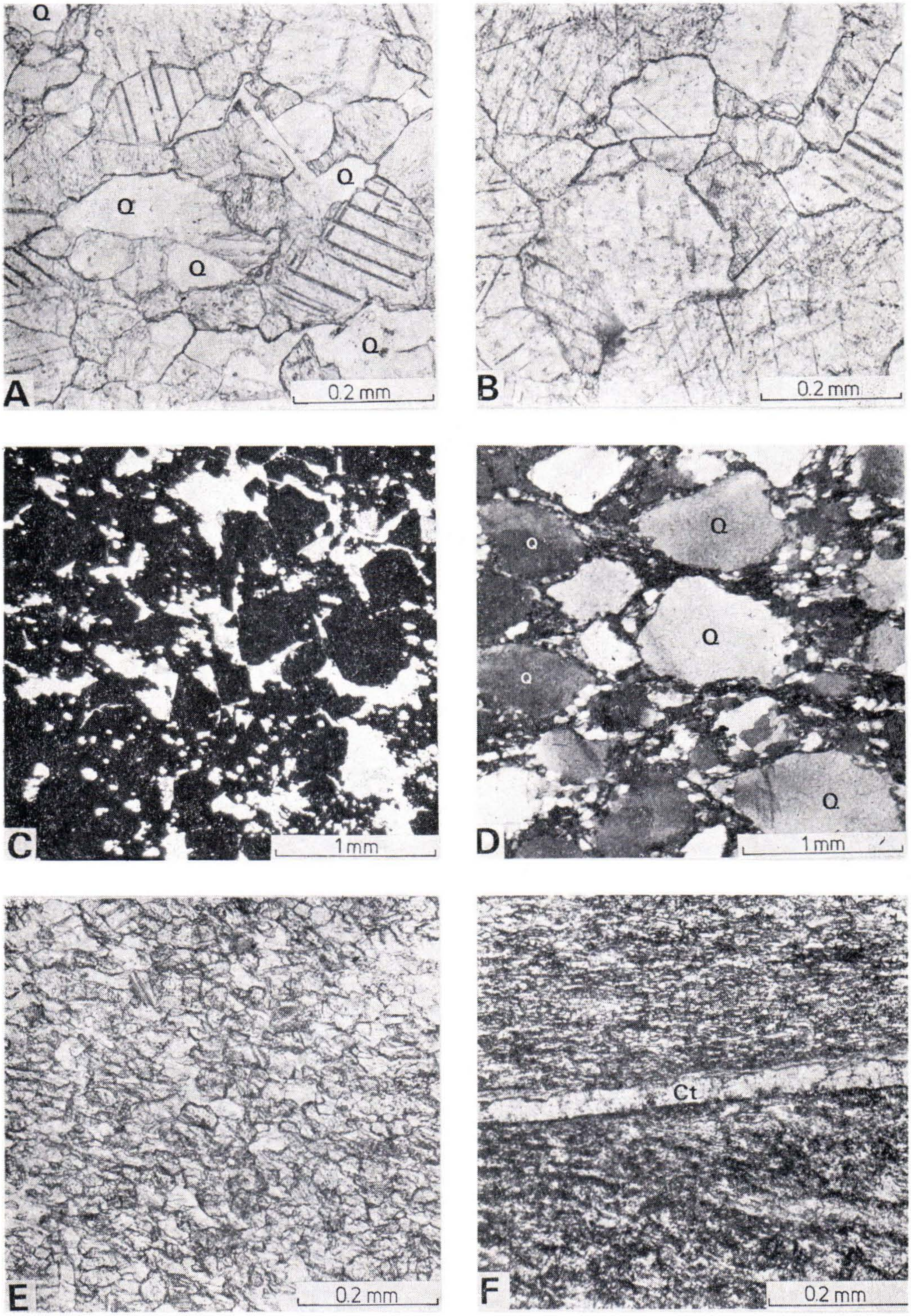
Rocks of the Hematite rock—Quartzite Formation

The hematite rock layer that belongs to this formation is composed mainly of small hematite scales and massive hematite. Here and there, the hematite rock at Haluksenlammit also contains octahedral crystals 2 to 3 mm in diameter that are presumably martite pseudomorphs after magnetite. Similar but smaller crystals have also been found in the hematite rock in the Viistola section (Fig. 56C). In addition to hematite, the rock contains layers with varying abundances of micaceous clay minerals, micas, chlorite and quartz.

Micaceous clay minerals, chlorite, quartz, tourmaline and rutile as well as hematite have been identified in the ferriferous slate resting on the hematite layer. The most fine-grained laminae in it contain less hematite and are greenish in colour, whereas the coarsest laminae are richer in hematite and reddish brown. In the quartzite succeeding the ferriferous slate, the interstices between the rounded quartz grains are filled with fine-grained matrix composed mainly of micas, quartz, chlorite and fine-grained hematite (Table 12, No. 52; Fig. 56D).

Rocks of the Dolomite—Carbonaceous slate—Volcanite Formation

The Dolomite Member consists of fine-grained laminated dolomite layers whose lamellar structure is partly due to variation in grain size and partly to variation in the abundances of argillaceous material such as micas and chlorite (Fig. 56E). Even the purest dolomite layers contain small amounts of quartz (Table 12, Nos. 48 and 49). Microprobe analyses show


A
Q
Q
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Q
Q
0.2 mm
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0.2 mm
C
1 mm
D
Q
Q
Q
Q
Q
1 mm
E
0.2 mm
F
Ct
0.2 mm

Table 12
Mineral compositions of various rocks in the Kiihtelysvaara Carbonate—Black slate Suite. Determined by point-counting method.

Minerals	45	46	47	48	49	50	51	52
Quartz	8.0	+	+	17.8	16.2	+	+	90.3
Biotite						15.8	+	4.6
Sericite	1.2	0.7	1.6	+	0.6	+	0.5	5.0 a)
Chlorite						+	+	
Carbonate	90.7	99.2	98.3	82.0	82.9	49.4	12.9	
Tremolite						34.5		
Talc						+		
Chert							86.5	
Others	0.1	0.1	0.1	0.2	0.3	0.3	0.1	0.1
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

+) detected minor constituent; a) including chlorite.

45. Light-coloured quartz-bearing dolomite from about 55 m below the hematite layer in the Dolomite—Volcanite Formation (drill core No. 309/4241/37.40), Viistola, Kiihtelysvaara.
46. Reddish dolomite from about 40 m below the hematite layer in the Dolomite—Volcanite Formation (drill core No. R310/4241/164.75), Viistola, Kiihtelysvaara.
47. Reddish dolomite from about 15 m below the hematite layer in the Dolomite—Volcanite Formation (drill core No. R310/4241/137.40), Viistola, Kiihtelysvaara.
48. Red-coloured, quartz-bearing dolomite from about 9 m above the hematite layer in the Dolomite—Carbonaceous slate-Volcanite Formation (drill core No. R310/4241/109.70), Viistola, Kiihtelysvaara.
49. Laminated quartz-bearing dolomite from about 30 m above the hematite layer in the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R310/4241/87.65), Viistola, Kiihtelysvaara.
50. Greenish tremolite-bearing dolomite from about 140 m above the hematite layer in the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R311/4241/113.60), Viistola, Kiihtelysvaara.
51. Calcareous nodular chert from about 65 m below the hematite layer in the Dolomite—Volcanite Formation (drill core No R310/4241/186.65), Viistola, Kiihtelysvaara.
52. Quartzite from the Hematite rock-Quartzite Formation (drill core No. R310/4241/116.20), Viistola, Kiihtelysvaara.

that the carbonate is dolomite (T. Paasivirta), although small calcitic bodies have also been encountered; the more intensely metamorphosed rocks contain tremolite and talc as well.

The Carbonaceous slate Member comprises fine-grained rocks that are tinted dark grey by dust-like amorphous carbon or graphite and contain micas, chlorite, quartz, tourmaline, rutile and pyrite (Fig. 56F). Some graphite-rich layers have been encountered in the deposits (Fig. 15), although as a rule the graphite content is not very high. In some places the carbon-rich layers exhibit shearing, which has produced secondary growth in the graphite scales (Fig. 54B). Graphite also occurs as small scales in the more intensely metamorphosed rocks. The other constituents in these graphite-bearing phyllites are biotite, sericite, chlorite

Fig. 56. Photomicrographs of rocks typical of the Kiihtelysvaara Carbonate—Black slate Suite.

A) Massive dolomite consisting of anhedral mosaic of dolomite with numerous small detrital quartz (Q) grains. Lower part of Dolomite—Volcanite Formation (Thin section No. 17863). Ordinary light. B) Massive dolomite consisting mainly of anhedral mosaic of dolomite grains. Middle part of Dolomite—Volcanite Formation (Thin section No. 17854). Ordinary light. C) Hematite rock consisting mainly of anhedral mosaic of hematite (black) in a quartz and mica matrix with a few euhedral grains of hematite pseudomorphs after magnetite. Hematite rock—Quartzite Formation (Thin section No. 17851) Ordinary light. D) Quartzite consisting of quartz grains (Q) in a matrix rich in quartz, micas and hematite. Hematite rock—Quartzite Formation (Thin section No. 17860). Crossed nicols. E) Laminated dolomite consisting of fine-grained anhedral mosaic of dolomite with numerous sericite grains and some detrital quartz. Lower part of Dolomite—Carbonaceous slate—Volcanite Formation (Thin section No. 17838) Ordinary light. F) Slightly laminated carbonaceous slate. The darker laminae contain more fine-grained (amorphic) carbon and less quartz, micas and chlorite than the lighter laminae. A few feather joints filled with calcite (Ct). The middle part of the Dolomite—Carbonaceous slate—Volcanite Formation (Thin section No. 17861). Ordinary light. Photo by E. Halme.

and quartz. Unlike black slates in general, the carbonaceous slates in the study area contain very little pyrite. Exceptions are the carbonaceous slate layers in association with volcanite beds in the Valkeavaara area, in which the mode of occurrence of pyrite refers to contamination from a volcanic source.

Apart from the dolomitic carbonate, the dolomitic layers of the Alternating dolomite—carbonaceous slate Member contain divers amounts of tremolite, micas and talc (Table 12, No. 50). The carbonaceous slate layers consist of micas, chlorite and quartz, and variable amounts of fine-grained graphite. It is noteworthy that graphite is not exclusively restricted to the carbonaceous slate layers but that some dolomite layers also contain a very fine-grained graphite dust.

The metavolcanites, metadiabases and gabbroic metadiabases of the Dolomite—Volcanite Formation and Dolomite—Carbonaceous slate—Volcanite Formation will be discussed in a later chapter (p. 114).

The chemical character of the carbonate sediments, carbonaceous slates and related rocks

The Viistola reference series

The analytical data on the samples collected from the different rock units of the Viistola section illustrate well the variation in chemical composition across the Kiihtelysvaara Carbonate—Black slate Suite (Fig. 57). Some samples from the Kalevian Kiihtelysvaara Turbidite Suite have been included for comparison. The drill core samples were assayed by emission spectrophotometer (G. Wansén) for SiO_2, Al_2O_3, MgO, CaO, Na_2O, K_2O, Ti and V (semiquantitatively). Total iron, Mn, Cu, Co, Ni, Pb and Zn were analysed by AAS and sulphur by a Leco M 621 sulphur analyser (A. Nurmi). The phosphorus analyses were conducted in the ore laboratory. A certain regularity was observed in the variation in chemical composition (Fig. 57):

- SiO_2 is highest in the quartzite layers; it decreases somewhat in the volcanites, clay slate and carbonaceous slate layers, being lowest in the pure dolomite layers; the broad peak at the upper end of the curve is due to (Kalevian) quartzites of the Turbidite conglomerate quartzite Formation.
- The Al_2O_3 curve shows high values at the clay slates and carbonaceous slates mainly on account of the abundant micas and chlorite; the lowest values are at the dolomite and quartzite layers; the rise at the upper end of the curve is due to (Kalevian) mica schists of the Mica schist Formation.
- The CaO curve is fairly straightforward and reflects the variation in carbonate abundance; hence, the grading of the upper quartzites into dolomites is clearly visible at the lower end of the curve.
- The MgO curve is largely analogous to that of CaO and indicates that most of the MgO is incorporated in dolomitic carbonate, although the abundances of mica and chlorite (± tremolite, ± talc) are also reflected in the shape of the curve.
- Total iron as Fe_2O_3 is persistently low, even though it rises slightly at the volcanite—carbonaceous slate layers; the high single peak in the first third of the curve is due to a hematite layer and ferriferous slate; the mica schists show higher iron than the carbonaceous slates.
- The Na_2O abundance is low with only one small peak caused by a feldspathic layer in the middle of the curve.
- The K_2O curve is very similar to that of Al_2O_3 and reflects the variation in the abundance of micas (± feldspar); the abundance rises in the mica schists.

— Manganese largely follows CaO and MgO, which indicates that this element is incorporated in the carbonate layers as well as in the clay slate and carbonaceous slate layers.
— Titanium is closely related to Al_2O_3, which suggests analogous chemical behaviour; the abundances are clearly higher in the mica schists.
— The vanadium curve exhibits low values in dolomites but a relative enrichment in clay slate and carbonaceous slate layers; the highest peak is at the hematite layer; the abundance of vanadium in the (Kalevian) quartzites and mica schists is roughly equal to that in the carbonaceous slates.
— The Cu, Co and Ni curves are alike in shape but not in detail; the highest abundances are due to volcanites, clay slate, carbonaceous slate and hematite rock layers; note the low abundances in the (Kalevian) quartzites and the rise in level in the mica schists.
— The Pb and Zn curves resemble each other, although not in detail; with some minor exceptions, the abundances in the dolomite, volcanite, clay slate and carbonaceous slate layers are almost equal: note the low abundances in the (Kalevian) quartzites and the rise in level in the mica schists.
— The sulphur abundance is rather low throughout but shows a slight rise in the clay slate and carbonaceous slate layers; the level rises distinctly in the mica schists.
— The phosphorus curve fluctuates intensely, although the abundances, even the highest, are fairly low, and no significant enrichment has been noted in any layers.

Chemical composition of the carbonate sediments

Three samples of carbonate sediments in the study area were analysed (Table 13). Analysis 18 represents a non-foliated carbonate rock

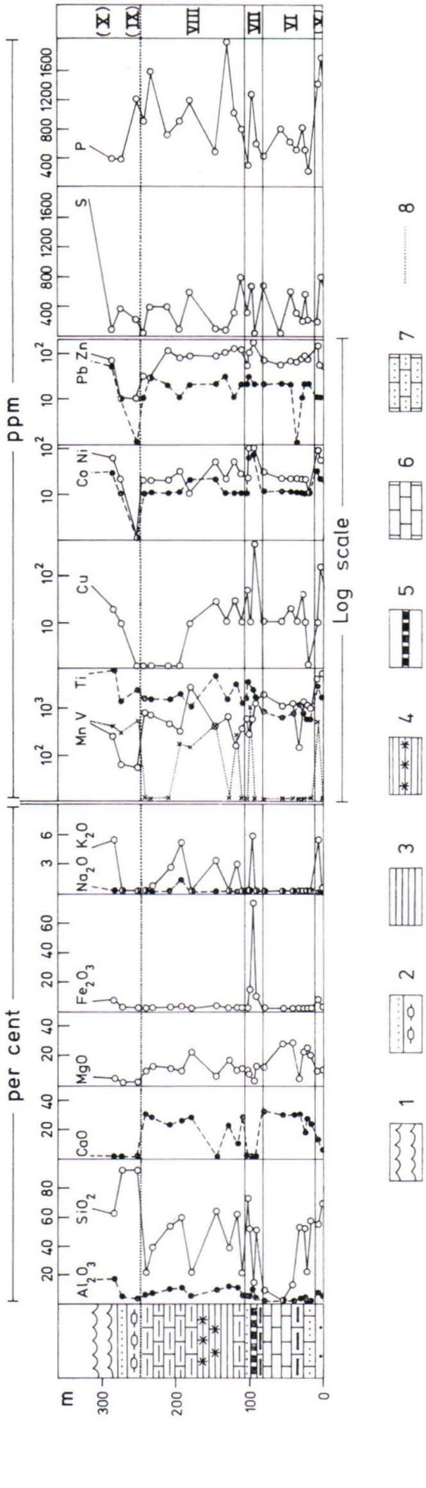

Fig. 57. Schematic diagram showing variation in chemical composition across the Kiihtelysvaara Carbonate—Black slate Suite in the Viistola area. Lithostratigraphic units: V = Upper quartzite Formation; VI = Dolomite—Volcanite Formation; VII = Hematite rock—Quartzite Formation. VIII = Dolomite—Carbonaceous slate—Volcanite Formation; IX = Turbidite conglomerate—quartzite Formation; X = Mica schist Formation. Symbols: 1 = mica schist; 2 = turbidite conglomerate and quartzite; 3 = carbonaceous slate; 4 = carbon-rich layers; 5 = hematite layer; 6 = dolomite; 7 = quartz-bearing dolomite; 8 = unconformity.

with some micas and quartz from the Dolomite—Volcanite Formation. Analyses 19 and 20 represent fine-grained, bedded carbonate rocks from the Dolomite—Carbonaceous slate—Volcanite Formation; the former contains quartz and micas, whereas the latter is a rather pure carbonate rock. Chemical analyses show that the carbonate in the rock is largely dolomite. Electron microprobe analyses (T. Paasivirta), however, demonstrated that calcitic bodies and veins are also encountered, particularly in the coarse-grained portions at the middle and top of the Dolomite—Volcanite Formation.

Table 13

Chemical composition of three dolomites in the Kiihtelysvaara Carbonate—Black slate Suite. Analyst: K. Karhunen. Emission spectrographic analyses by A. Löfgren.

	18	19	20
SiO_2	3.86 [a])	13.56 [a])	1.40 [a])
Al_2O_3[b])	3.52	2.52	0.78
FeO	1.84	0.81	0.24
MnO	0.23	0.13	0.11
MgO	20.44	17.67	21.33
CaO	27.80	26.41	29.52
CO_2	42.80	37.80	45.60
Total	100.49	98.90	98.98

	Emiss. Spec	Emiss. Spec	Emiss. Spec
Ba	0.01 [c])	0.01 [c])	0.01 [c])
Sr	0.04	0.03	0.03
Ti	0.21	0.038	0.006
V	0.003	0.003	0.003

a) »Residue (mostly silica)»; b) Including Fe_2O_3; c) Constituent does not exceed the figure given.

18. Reddish dolomite from about 70 m below the hematite layer in the Dolomite—Volcanite Formation (drill core No. R 1/ Khv-61/145.85), Haluksenlammit, Kiihtelysvaara.
19. Reddish quartz-bearing dolomite from about 10 m above the hematite layer in the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R 1/Khv-61/52.65), Haluksenlammit, Kiihtelysvaara.
20. Light coloured fine-grained dolomite from about 35 m above the hematite layer in the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R 2/Khv—61/187.00), Haluksenlammit, Kiihtelysvaara.

The analyses reveal that the dolomites contain small amounts of ferrous iron incorporated mainly in micas and chlorite, siderite being practically non-existent. Moreover, the dolomites contain Mn, Sr and Ba as trace elements. According to Pettijohn (1975, p. 362) and Bathurst (1975, pp. 261—264), the abundances of these elements are typical of dolomites.

Chemical composition of the iron-rich sediments

The hematite layer and its associated ferriferous slates form an iron-rich unit that differs chemically from the environment. In the hematite layer (Table 14) the iron content varies from 37.4 to 57.2 % (weighted average for drill cores 41.3 %); in the ferriferous slate it varies from 11.0 to 15.4 % (weighted average for drill cores 14.2 %). Whole rock analysis of the hematite rock (No. 21) shows that the iron occurs as ferric oxide, i.e. as hematite rather than magnetite. Hence, the hematite rock resembles the hematite layers in the dolomite deposit at Suojärvi (Metzger 1924, p. 53) rather than the magnetite-predominant iron formations at Väyrylänkylä, Puolanka (Laajoki 1975, p. 99; Laajoki and Saikkonen 1977) or at Tuomivaara, Sotkamo (Mäkelä 1976). The hematite rock is clearly richer in total iron, Al_2O_3, K_2O and Ti, but poorer in SiO_2, MgO, CaO, P and S than are the rocks of iron formation at Väyrylänkylä.

Chemical composition of the carbonaceous slates

The carbonaceous slates and graphite schists are tinted dark grey by amorphous carbon or graphite, even though their carbon content is often low, being no more than a few percentages. However, this deposit also contains some layers 1 to 10 m thick in which the carbon content varies from 20 to 50 % (average of five samples

Table 14

Chemical composition of iron-rich sediments in the Kiihtelysvaara Hematite rock—Quartzite Formation (weight per cent). Whole-rock analysis (No. 21) by A. Heikkinen. Analyses, Nos. 22—28 performed at Ore Laboratory.

	21	Constituent	22	23	24	25	26	27	28
SiO_2	15.97	$Iron_{HCl}$	57.2	37.4	38.6	42.2	38.5	43.1	9.97
TiO_2	0.70	$Iron_{tot}$	57.2						11.0
Al_2O_3	6.90	Mn	0.02						0.02
Fe_2O_3	67.40	Ti	0.26						0.66
FeO	0.86	V	0.16						0.07
MnO	0.02	P	0.13						0.08
MgO	1.63	S	0.07						0.09
CaO	1.26								
Na_2O	0.05								
K_2O	2.70								
P_2O_5	0.90								
CO_2	0.00								
H_2O^+	0.72								
H_2O^-	0.18								
V_2O_5	0.06								
S	0.01								
Total	99.36								

21. Hematite rock (drill core No. R5/Khv—61/93.80—94.30, Haluksenlammit, Kiihtelysvaara.
22. Hematite rock (drill core No. R310/4241/118.95—119.00), Viistola, Kiihtelysvaara.
23. Hematite layer (drill core No. R1/Khv—61/65.01—66.34; actual thickness of layer 120 cm), Haluksenlammit, Kiihtelysvaara.
24. Hematite layer (drill core No. R2/Khv—61/218.94—220.44; actual thickness of layer 140 cm), Haluksenlammit, Kiihtelysvaara.
25. Hematite layer (drill core No. R3/Khv—61/41.72—43.02; actual thickness of layer 127 cm), Haluksenlammit, Kiihtelysvaara.
26. Hematite layer (drill core No. R4/Khv—61/24.13—26.00; actual thickness of layer 145 cm), Haluksenlammit, Kiihtelysvaara.
27. Hematite layer (drill core No. R5/Khv—61/93.30—94.85; actual thickness of layer 70 cm), Haluksenlammit, Kiihtelysvaara.
28. Ferriferous slate, drill core No. R310/4241/118.15—118.25, Viistola, Kiihtelysvaara.

30 % C). Unlike black slates in general, the carbonaceous slates in the study area have low sulphur, e.g. 0.1 % in the Viistola section. Exceptions are the graphite slates at Valkeavaara, which assay 1 to 3 % sulphur (p. 38).

The average contents of the minor constituents

The distribution of trace elements Cu, Ni, Zn, Pb and Co and of sulphur and phosphorus in the dolomites and carbonaceous slates in the study area was examined statistically by comparing them with the corresponding values for the Lower and Upper quartzites. The analytical data were elaborated into graphical form by ADP by applying extended scatter diagram that also show the arithmetic mean, geometric mean, median and standard deviation (Fig. 58).

On account of the limited observational data, the comparison was based on the medians typical of each rock type rather than on the arithmetic means. The distributions of the elements exhibit a certain regularity (Fig. 58):

— The copper content is low (10 ppm) in the Lower quartzite and dolomites but rises two-fold in the Upper quartzite and three-fold in the carbonaceous phyllites.

— The nickel content is low (10 ppm) in both quartzites but rises two-fold in the dolomites and eight-fold in the carbonaceous slates.

— The zinc content is lowest (about 35 ppm) in both quartzites but rises seven-fold in the dolomites and eight-fold in the carbonaceous slates.

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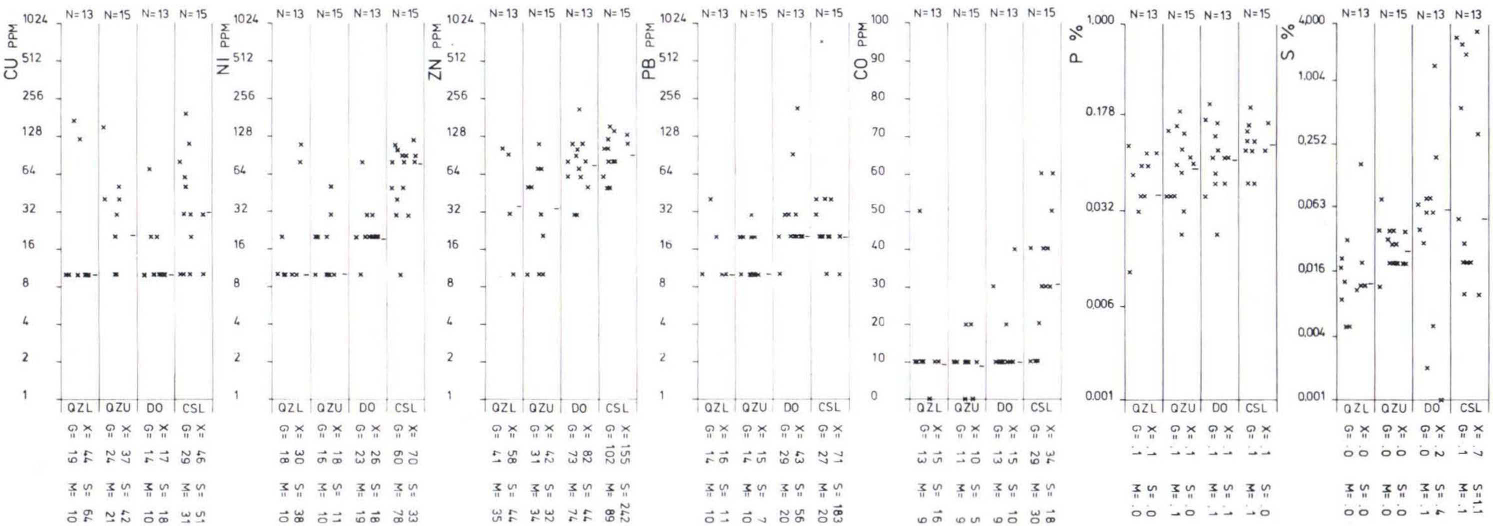

Fig. 58. Scatter diagram showing the distribution of copper, nickel, zinc, lead, cobalt, phosphorus and sulphur in the rocks of the Kiihtelysvaara Carbonate—Black slate Suite compared with the abundances of these elements in the rocks of the Orthoquartzite Suite. Cu, Ni, Zn, Pb and Co were assayed by AAS, and sulphur by a Leco M 621 sulphur analyser (A. Nurmi). Phosphorus was analysed at the ore laboratory of the Chemistry department. Scales of columns logarithmic. Symbols: QZL = lower quartzite; QZU = upper quartzite; DO = dolomite; CSL = carbonaceous slate; X = arithmetic mean; M = median; S = standard deviation; N = number of observations. The medians marked on the column by short horizontal lines.

- The cobalt content is low (9 ppm) in both quartzites and in dolomite but rises three-fold in the carbonaceous slates.
- The sulphur content is very low (0.01 %) in the Lower quartzite but rises three-fold in the Upper quartzite and five-fold in the dolomites and carbonaceous slates.
- The phosphorus content is lowest (0.04 to 0.06 %) in both quartzites but rises about two-fold in the dolomites and carbonaceous slates.

Conclusions concerning the depositional environment and origin of the material

Recrystallization has almost completely obliterated primary structures from the carbonate rocks of the Dolomite—Volcanite Formation, and the rock is composed of a medium- to coarse-grained crystalline mosaic (Fig. 53A). Nevertheless, in some thin sections, there are ghost-like figures that may indicate the outlines of the original clastic grains. These, as well as the presence of detrital quartz grains and cross-bedding, suggest that the carbonate rocks in at least the lower part of the deposit are allochthonous, i.e. they were deposited mechanically and not formed in situ.

It is not easy to say whether the carbonates were formed originally by chemical or biochemical precipitation. They do, however, contain microfossils resembling blue-green algae (Tynni 1971), and hence, the contribution of micro-organisms is quite possible. The carbonates were formed at a rather high rate, which indicates an arid or semi-arid paleoclimate. According to Neuvonen (1974, 1975), the paleolatitude of the Baltic Shield was fairly far south (22 °N) during the Sariolan/Jatulian epoch, a fact that corroborates the presumption made above.

Dolomite carbonate is the principal constituent of these rocks although calcite also occurs as nests and veins. Numbers of investigators, pondering over the dolomite dilemma, have suggested several alternative explanations: some assume that limestones received extra magnesium (dolomitization) either during diagenesis or later epigenetically; others maintain that the constituents of the dolomites were formed primarily under conditions unlike those prevailing at present. In the case at hand, the replacement structures in the rock suggest that the dolomite is not primary but a product of dolomitization. The coarse calcitic nests and veins may be due to subsequent calcitization.

The abundant chert nodules and lenses in the carbonate rocks of the Dolomite—Volcanite Formation are an interesting feature. Whether they are replacement structures or primary precipitates is not easily established. Their occurrence in the parts of the carbonate rocks that contain abundant stylolitic dissolution structures (Fig. 53A), and the three-dimensional network that the nodules form within the rock refer to replacement structures rather than to primary origin. The source of the silica is more difficult to explain; nevertheless, the numerous tiny microcrystalline quartz blebs in the carbonate rocks suggest that when the carbonate was forming small amounts of silica were precipitated from the sea water either by chemical or biochemical processes. The silica that was needed to form the nodules probably dissolved from the host rock and reprecipitated as nodules by means of diagenetic differentiation (Ramberg 1952, p. 222). The origin of the larger chert lenses is, however, not readily explained this way. Perhaps in those ancient days silica was precipitated directly from sea water (through volcanic actions?).

Later the formation of carbonates seems to have been temporarily suspended owing to brief regression of the sea, a stage that included volcanic activity (p. 125). This may have been the cause of the significant changes that then took place over large areas of the marine sedimentation environment, particularly when the pH of the water shifted towards higher acidity. The outcome was the formation of the modest hematite layers that occur not only in the Kiihtelysvaara—Tohmajärvi area but also in many places in Soviet Karelia, e.g. Soanlahti, Suojärvi and Tulomajärvi (Metzger 1924, Hausen 1930, Kharitonov 1966, Sokolov *et. al.* 1970). It is typical of them that nowhere do they form larger deposits. Obviously, iron was precipitated under the oxidizing conditions of shallow water (resembling lake ores). It is noteworthy that in the study area hematitic matter is associated with argillaceous matter and, owing to its non-cherty character, differs conspicuously from the chertic iron formation rocks in Väyrylänkylä, Puolanka (Laajoki 1975; Laajoki and Saikkonen 1975). The regression of the sea lasted so long, however, that in the study area the hematite and ferriferous slate layers had time to be covered by sandy sediments, as in Soviet Karelia at Tulomajärvi and Suojärvi (Metzger 1924; Sokolov *et al.* 1970, pp. 55—56).

With the beginning of the new transgression, carbonates started to form once more. The carbonate rocks that formed as the first units in the Dolomite—Carbonaceous schist—Volcanite Formation are composed of massive dolomite similar to that overlain by the hematite layer except that the former is succeeded by a fine-grained lamellar carbonate rock. Its laminar structure is due to variation in the amount of rather minor clayey matter. A pale and thicker layer was deposited when the rate of carbonate formation was at its highest, but a darker and thinner layer rich in argillaceous matter when it was at its lowest (Fig. 54A). The well-preserved layered structures suggest that these layers were formed in situ, i.e. they are autochthonous.

The carbonaceous matter is composed of practically pure dolomite (p. 101), indicating that the dolomite is either primary or that the dolomitization took place very soon after deposition. It has been suggested that in ancient times and under certain conditions (temperature, salinity, pH) dolomitic carbonate was formed primarily on a larger scale than at present. The role played by biochemical precipitation is also unclear, since, in this deposit as well, microfossils resembling blue-green algae have been found (Tynni 1971). Clear biogenic structures, however, such as the stromatolite structures so common in the dolomites at Tervola (Härme and Perttunen 1963), have not been identified, although some observations on the erratic blocks of dolomite in the Sääperi area suggest the occurrence of stromatolites (cf. Sokolov 1959).

Conditions obviously changed later; thus, the formation of carbonate was inhibited and mainly clay was deposited. This gave rise to the laminated clay slate-carbonaceous slate layers that succeed the bedded dolomite. The red-greenish laminated clay slate only had time to deposit as a thin layer, however, before an abrupt change in sedimentation conditions caused the depositing clay to become carbon-bearing. The carbonaceous slates and phyllites thus formed are darkened by a fine-grained amorphous carbon or graphite dust. They too exhibit vague laminar structure. Similar carbonaceous rocks, known as »shungite-bearing» rocks or »shungites», are both more abundant and better preserved in Soviet Karelia in the areas of Ääninen, Suojärvi, Tulomajärvi and Pieni Jänisjärvi (Metzger 1924; Hausen 1930; Kharitonov 1966; Galdobina and Gorlov 1975; Sokolov and Kalinin 1975).

The source of the carbon in the carbonaceous slates has still not been established. The local occurrence of carbon-bearing slates in close association with volcanites refers to an endogenic origin for these slates (Galdobina and Gorlov 1975); in general, however, it has been assumed that the carbon derives from remnants of tiny micro-organisms (Sidorenko and Sidorenko 1971). In any case, the restricted size of the occurrences indicates that the carbonaceous slates were formed only under specific conditions. It is possible that, in isolated off-shore marine bays, conditions changed and became lethal for micro-organisms. This led to their extermination and the subsequent mixing of their organic residue with the depositing clay.

The dolomite layers in the carbonaceous slates and phyllites demonstrate that, from time to time during sedimentation, the water could flow more freely to the basins; hence, under the appropriate conditions, small amounts of carbonate were precipitated. As a result, layers of dolomite and carbonaceous slate were formed alternately in the upper part of the Dolomite—Carbonaceous slate—Volcanite Formation. Since this association also contains scattered sandy interbeds, deposition presumably took place in rather shallow water.

THE KIIHTELYSVAARA TURBIDITE SUITE

As the sea withdrew from the land carbonate- and carbon-bearing matter ceased to deposit. The partly consolidated sedimentary piles and associated volcanites rose above sea level in many places and came under the influence of erosion. Before long, a new sedimentation stage set in under these conditions. First, the debris produced by erosion was washed out to sea while turbulent streams transported the coarse clastic material to the margins of the basin and the finer substances farther out from the shore. The normal greywacke-turbidite type of sediments were then deposited. This suite corresponds to the basal part of the Kalevian flysch formation described by Väyrynen (1933, 1954) and is called the Kiihtelysvaara Turbidite Suite in the present context. It is noteworthy that volcanic rocks have not been encountered in this association. Later, during the Svecokarelidic orogeny (about 1800—1950 Ma), the rocks of this suite were metamorphosed into conglomerates, quartzites and mica schists with staurolite and andalusite porphyroblasts.

The associations and their mode of occurrence

In the Kiihtelysvaara—Värtsilä area the conglomerate—quartzite horizon rests either directly on quartzites or on dolomites, carbonaceous slates and volcanites with an unconformity marked by an erosion surface in between. Known as the Turbidite conglomerate—quartzite Formation, it begins with a fairly coherent conglomerate layer 0.2—15 m thick (the Turbidite conglomerate Member) that grades upwards into graded-bedded quartzite (the Graded-bedded quartzite Member). The conglomerate—quartzite occurrences consist typically of plate-like forms, which taper westwards and may reach as much as 50 m in thickness.

Upwards the graded-bedded quartzite grades into either graded-bedded phyllites or mica schist, depending on the degree of metamorphism. In this context the deposit is called the Mica schist Formation. Since the area occupied by the mica schists extends westwards far beyond the study area, the geology of the Mica schist Formation has been treated only in so far as it has bearing on the present study.

Primary sedimentary structures and textures

The Turbidite conglomerate at Kiihtelysvaara (Figs. 7 and 59) shows large-scale grading structure, and for rather a short distance, grades upwards into a graded-bedded quartzite (Fig. 60).

The largest, most angular and most plate-like fragments have been found in the lower part of the conglomerate layer. Upwards, the fragments become smaller and can be divided into two groups according to their degree of roundness: rounded and well-rounded fragments composed of vein quartz, quartzite and less often of volcanites (Fig. 7); subangular and subrounded fragments composed of carbonaceous slate and various volcanites.

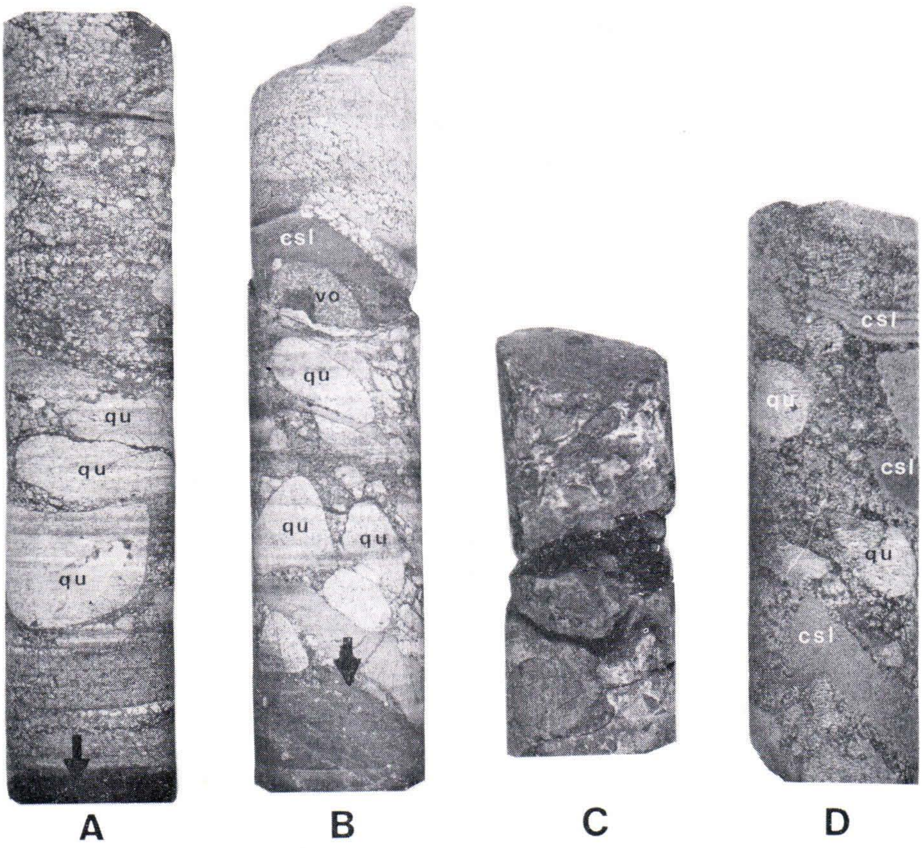

Fig. 59. Drill-core samples from conglomerate at the base of the Kiihtelysvaara Turbidite Suite, i.e. from Kalevian basal conglomerate. A) A conglomerate sample from the basal part of the turbidite conglomerate; the pebbles are predominantly quartzite (qu); the contact of the conglomerate with the carbonaceous slate (arrow) is visible at the lower end of the core (drill core No. R303/4232/123.35). Haluksenlammit, Tohmajärvi. B) A conglomerate sample from the basal part of the turbidite conglomerate; pebbles predominantly quartzite (qu), carbonaceous slate (csl) and volcanite (vo); the contact of the conglomerate with the carbonaceous slate (arrow) is visible at the lower end of the core (drill core No. R306/4232/23.05). Valkeavaara, Kiihtelysvaara. C) A brecciated basal part of the turbidite conglomerate (drill core No. R311/4241/99.80). Viistola, Kiihtelysvaara. D) A conglomerate sample from the basal part of the turbidite conglomerate; pebbles predominantly quartzite (qu), carbonaceous slate (csl); the matrix rich in debris from effusives and diabases (drill core No. R314/4241/79.80). Hyypiä, Kiihtelysvaara. Drill cores 32 mm in diameter. Photo by E. Halme.

Fig. 60. Graded-bedded quartzite (AE-turbidite) from the Turbidite conglomerate—quartzite Formation. Each unit shows an increase in pelitic material towards the top (left). Length of pen 14 cm. Valkeavaara, Kiihtelysvaara.

The pebbles of vein quartz are as well rounded (P = 0.64) as the pebbles in the quartz conglomerate interbeds in the lower quartzite horizons from which they probably derive. The quartzite fragments (Figs. 7 and 59) vary from a few millimetres to half a metre in size and are rounded (P = 0.58). This suggests that the quartz sandstone was still rather poorly con-

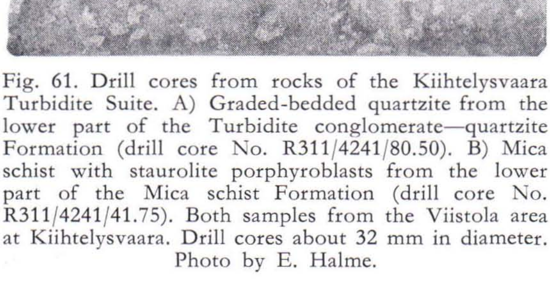

Fig. 61. Drill cores from rocks of the Kiihtelysvaara Turbidite Suite. A) Graded-bedded quartzite from the lower part of the Turbidite conglomerate—quartzite Formation (drill core No. R311/4241/80.50). B) Mica schist with staurolite porphyroblasts from the lower part of the Mica schist Formation (drill core No. R311/4241/41.75). Both samples from the Viistola area at Kiihtelysvaara. Drill cores about 32 mm in diameter. Photo by E. Halme.

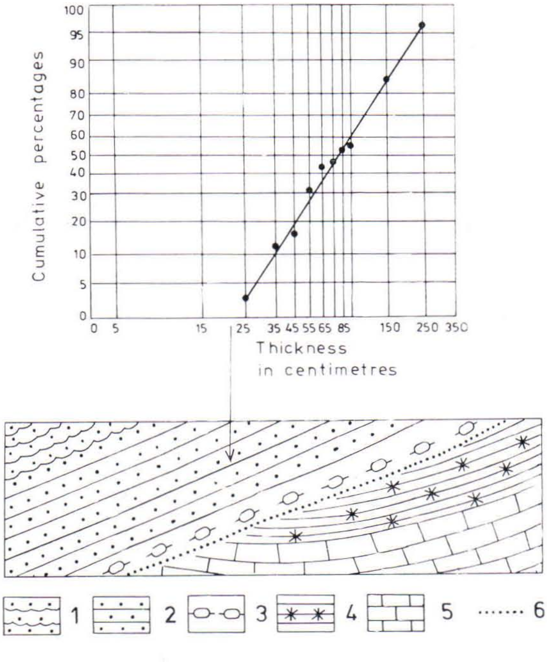

Fig. 62. Schematic representation of bedding thickness in quartzites of the Turbidite conglomerate—quartzite Formation plotted on logarithmic probility paper. Symbols: 1 = graded-bedded phyllite; 2 = graded-bedded quartzite; 3 = conglomerate; 4 = carbonaceous slate; 5 = dolomite; 6 = unconformity.

solidated when it was submitted to erosion and that the rounding took place in the course of fairly short transport.

The volcanite and diabase fragments, e.g. those in the Kortevaara conglomerate (Fig. 7 and p. 21), are from a few mm to three metres in size. They vary from subrounded to rounded, obviously depending on the distance they were transported. The carbonaceous slate and carbon phyllite fragments, which are seldom more than 20 cm long, are the most angular of the fragments and occur at the base of the conglomerate layer.

In the graded-bedded quatzite that deposited on the conglomerate (Figs. 60 and 61) the layers are from 15 cm to three metres thick. This variation in the thickness of the layers (80 layers) does not exactly obey the log-normal distribution (Fig. 62), and thus indicates somewhat irregular sedimentation. Each graded bed (Fig. 60) consists of a quartzitic lower part (sand) and a phyllite or mica schist (shale) upper part. They are AE turbidites (proximal facies) in the sense of the term as applied by Bouma (1962). Upwards the quartzitic lower parts become rapidly thinner while the phyllitic upper parts become thicker and grade into graded-bedded mica schist (distal facies; Fig. 20). In the lower part of the quartzite deposit the sand to shale ratio is 10 : 1 or higher; upwards, over a distance of 15 to 50 m, it grades into 1 : 1. Still farther upwards it is even lower. Hence, the boundary between the graded-bedded quartzite and mica schist marked on the maps is only statistic (Fig. 15).

Intense recrystallization has obliterated the primary boundaries of the mineral constituents in the graded-bedded quartzite and mica schist; thus, it is impossible to determine the grain-size distribution or the roundness of the grains.

Compositional characteristics

The Kiihtelysvaara Turbidite conglomerate is polymictic, containing fragments from various Karelian rocks but, except for redeposited vein quartz pebbles, none from the Prekarelidic basement (Figs. 7 and 59; Table 15). The matrix in the lower part of the conglomerate contains abundant mica-rich matter deriving from carbonaceous slates, as well as dark minerogenic matter from volcanites and diabases. In some places the basal matrix contains appreciable carbonate. As a result of metamorphism, garnet porphyroblasts that reach 3 cm in diameter have formed in the mica-rich portions. Upwards the matrix grades into a quartz-sandy variety.

The turbidite quartzite is frequently pale grey or grey and shows a glassy fracture surface (Figs. 60 and 61A). The lower part of each graded bed is composed almost entirely of medium- to coarse-grained granoblastic quartz with some minor sericite, biotite and accessories (Table 16, No. 53). The dark hue of the rock may be largely due to the very fine-grained graphite that occurs as inclusions in the recrystallized quartz. The more fine-grained upper part of each graded bed is composed predominantly of micas and quartz, with minor chlorite, feldspar and accessories.

The graded-bedded phyllite, or rather mica schist, is made up mainly of micas and quartz, with minor feldspar, chlorite and accessories (Table 16, No. 54). Staurolite porphyroblasts are often present, particularly in the mica-rich upper parts of the graded beds (Fig. 61B; Table 16, No. 55). Garnet and andalusite porphyroblasts are also rather common.

Studies on the chemistry of the rocks of the Kiihtelysvaara Turbidite Suite are still at a preliminary stage; hence, their chemical characteristics have not been fully established.

Nevertheless, enough data are available to show how clearly their chemical composition differs from that of the sediments lower down in the sequence (Fig. 57 and pp. 102—107), as is to be expected from the difference in mineral composition.

Table 15

Compositional characteristics of the conglomerate member of the Turbidite conglomerate—quartzite Formation in the Kiihtelysvaara—Värtsilä area.

Characteristics	Turbidite conglomerate
Gravel-sized fragments	
of Prekarelidic	
— plutonic rocks	not observed
— supracrustal rocks	not observed
— vein quartz and fine-grained quartz[1])	common
of Karelian	
— carbonaceous slate	abundant
— dolomite	rare
— volcanite	common
— diabase	common
— quartzite	common
— arkosite	not observed
Matrix	upper part quartzite-like, commonly rich in micas; lower part consisting of rock and mineral debris of all sizes, rich in micas and locally rich in carbonate
Typical heavy minerals	tourmaline, magnetite, zircon, titanite

[1]) redeposited pebbles and cobbles from the eroded quartz conglomerate interbeds in the Jatulian quartzites.

Table 16

Mineral compositions of various rocks in the Kiihtelysvaara Turbidite Suite. Determined by point-counting method.

Minerals	53	54	55
Quartz	98.0	35.2	31.8
Feldspars		0.7	0.6
Biotite	+	33.0	43.2
Sericite and chlorite	2.0	30.9	21.4
Staurolite			2.8
Carbonate	+	+	+
Titanite	+	+	+
Zircon	+	+	+
Tourmaline	+	0.2	0.2
Opaque	+	+	+
Total	100.0	100.0	100.0

+ detected minor constituent

53. Graded-bedded quartzite from the lower part of the Turbidite conglomerate—quartzite Formation (drill core No. R 311/4241/73.65), Viistola, Kiihtelysvaara.
54. Mica schist from the lower part of the Mica schist Formation (drill core No. R 311/4241/63.40), Viistola, Kiihtelysvaara.
55. Staurolite-bearing mica schist from the lower part of the Mica schist Formation (drill core No. R 311/4241/4.55), Viistola, Kiihtelysvaara.

Conclusions concerning the depositional environment and origin of the material

The erosional period was followed by a new stage in Karelian sedimentation, and the configuration of the eastern margin of the sedimentation basin was altered as a result of the evolution of the geosyncline that had developed farther west. The new Karelian sedimentation stage obviously had a violent start, because the weathering material produced by erosion seems to have been whipped by turbulent currents to the margins of the basin. The conglomerate that deposited on the bottom forms a 20 cm to 15 m thick wedge-shaped blanket that covers the sedimentary and volcanic rocks but probably disappears entirely towards the middle of the basin (westwards). The graded-bedded quartzite that succeeds the conglomerate behaves in a similar manner.

There is no doubt that the sediment and volcanite piles that rose from the sea during the erosional stage preceding the sedimentation stage were affected by erosion that, in some places, cut quite deep into the rocks. How else could one explain that the conglomerate at the base of the Kiihtelysvaara Turbidite Suite contains carbonaceous slate, quartz-bearing dolomite, cross-bedded quartzite, mafic effusives and hypabyssic dyke rocks as fragments, i.e. the same rocks as in the stratigraphic horizons

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below the conglomerate. Further, the depositional basement of the conglomerate varies: sometimes it is dolomite, sometimes carbonaceous slate and sometimes volcanite; in other words, the conglomerate is separated from the underlying beds by an angular unconformity. The material of the graded-bedded quartzite derived predominantly from eroded quartzite horizons, although in the course of sedimentation the quartz sands were intermixed with more fine-grained clay-bearing constituents from an unknown source. This material as well as the more fine-grained variety in the mica schists possibly derived from the internal zone of the geosyncline in the west. This topic has not been considered further, however, and it poses a separate problem.

EFFUSIVE AND HYPABYSSAL ROCKS OF KIIHTELYSVAARA TYPE

Volcanic activity was marked in the Kiihtelysvaara—Värtsilä area during the Karelian period although the effusive beds thus formed were not as voluminous as, say, in Soviet Karelia (see Sokolov *et al.* 1970). Nevertheless, the volcanite beds have facilitated the correlation of the Karelian formations with those in the other areas, particularly since some of the ancient discharge channels of the lava flowe have been identified and also dated with the aid of their diabase-like rocks (p. 135). Because the effusive/diabase-like rocks were folded and metamorphosed to a variable extent during the Svecokarelian orogeny (about 1800 to 1950 Ma ago) the effusive rocks are called metavolcanites and the diabase-like rocks metadiabases.

The associations and their mode of occurrence

In the study area the lowest stratigraphic horizon in which effusive rocks occurs is a metavolcanite bed 30 to 80 m thick that rests on the Lower quartzite Formation and extends fairly coherently from the Sääperi area (Fig. 21) to Karsikkojärvi (Fig. 5). This bed, called the Volcanite Formation, is composed of two or three lava flows. Fine-grained rocks dominate in the lower and middle parts of each lava flow; the upper part contains amygdaloidal variants, lava breccia or fragmentary lava (Fig. 63). The upper lava flow is overlain by layered tuffs and tuffites. The volcanite layers dip 50 ° to 60 ° westwards except in the Sääperi syncline (Fig. 21) where they form a subhorizontal plate that dips towards the middle of the syncline.

The quartzites and dolomites succeeding the Volcanite Formation contain local tuff and tuffite intercalations a few tens of centimetres or a few metres thick (Fig. 16). In the upper part of the Dolomite—Volcanite Formation in the Viistola section (Fig. 15), however, there is a lava bed roughly 10 m thick (the Volcanite Member) that is a still better indication of volcanic activity. In the Haluksenlammit area (Fig. 19A), the corresponding volcanite has probably been eroded to such an extent that only the feeding channels, i.e. the metadiabase dykes cutting the dolomites, have survived.

Volcanic intercalations are markedly more abundant in the upper part of the Dolomite—Carbonaceous slate—Volcanite Formation,

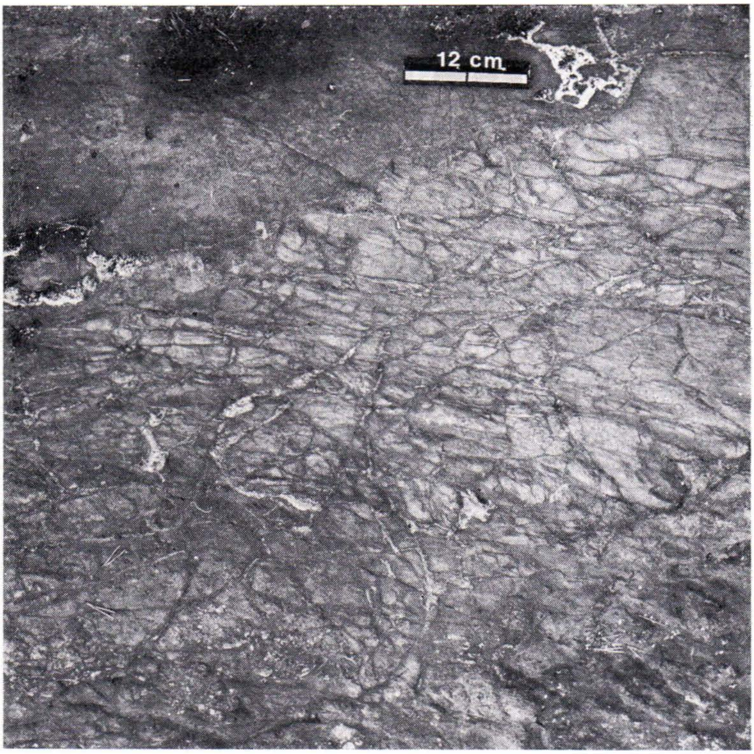


Fig. 63. Contact between first (below) and second (above) lava flow in the Volcanite Formation. Notice fragmentary lava in upper part of first lava flow. Length of tag 12 cm. Hyypiä, Kiihtelysvaara. Photo by E. Halme.

where the volcanic layers (the Volcanite Member) alternate with layers of dolomite and carbonaceous slate. They are predominantly lavas but include some tuffs as well. These layers are often a few tens of centimetres or a few metres thick.

Abundant metadiabase dykes (Fig. 21) crop out from under the Karelian sedimentary-volcanite piles, which, in the study area, are cut transversally by the present erosion surface. Most of the dykes occur in the Prekarelidic basement although numbers of them also cut the Karelian formations (Figs. 5, 6A, 8, 19A, and 21). The majority of the dykes trend roughly NW in such a way that the major maximum is at N 50 ° W and minor maxima at N 75 ° W and N 30 ° W. The dykes trending N 85 °, 60 ° and 40 ° E are less frequent. The trends of the metadiabase dykes are also marked fracture trends even though their maxima show a different preference, viz. N 20 ° W, N 85 ° E, N 50 ° E and N 60 ° E, listed in the order of falling frequency. Two other directions have also been noted, viz. N 85 ° W and N 10 ° W, which are not among the trends preferred by dykes. The dykes that occur in the Prekarelidic basement are more or less rectilinear and subvertical whereas those that extend to the Karelian formations are bent and tilted (Figs. 6A and 6B), presumably as a result of folding.

The individual metadiabase dykes are rather small in dimension: excluding the apophyses, they are from 10 to 30 m wide (Fig. 23), although some of them are as much as 100 m wide. The thin dykes may be a few hundred metres long, but the thickest ones may extend for 3 to 5 km. The thinnest dykes are fairly homogeneous darkish metadiabase, whereas the widest dykes exhibit differentiation (cf. Meriläinen 1961; Nykänen 1968). The contacts of the dykes often show fine-grained sheared margins or breccia. In some places, xenoliths of wall rock,

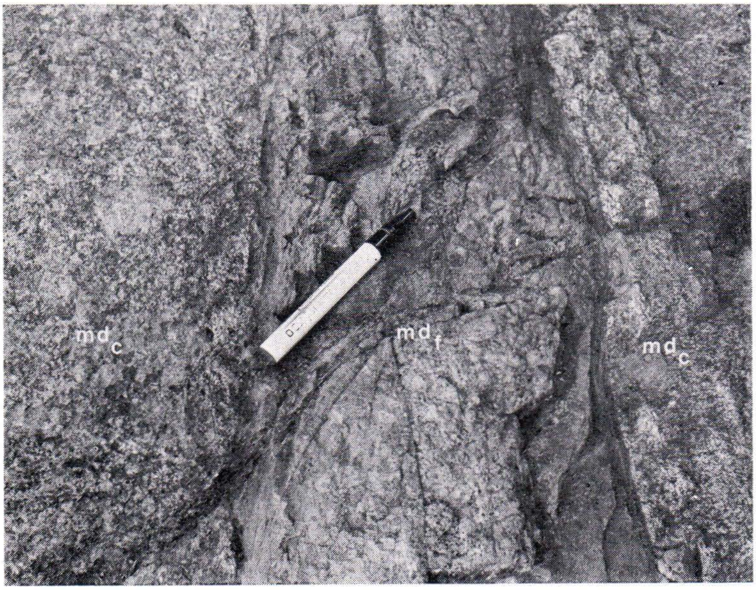


Fig. 64. A narrow internal dyke of fine-grained metadiabase (md_f) cutting medium-grained metadiabase (md_c) in the Kortevaara dyke. Length of pen 14 cm. Kortevaara, Kiihtelysvaara. Photo by E. Räisänen.

such as quartzite, have been observed in the marginal parts of the dyke. The contact effect of the intruded rock on the wall rock has been mild; in some places, however, e.g. in the contacts with quartzite, a seam of glassy quartzite some 20 cm thick has been found.

The intersections of the dykes that trend in various directions and cut the Prekarelidic basement have not been found exposed; hence, we do not know whether the dykes cut each other. On the basis of the relation of the metadiabase dykes to the Karelian formations, however, we do know that not all the dykes are contemporaneous. It seems likely that the majority of the dykes exposed never extruded to the surface. Nevertheless, there are some fairly common dykes that cut the arkosite-quartzite piles and extend up to the Volcanite Formation, obviously representing ancient discharge channels of the lava flows that gave rise to that Formation (Figs. 5, 6A, 8, 19A, and 21). Younger than them are the dykes that intruded through the arkosite-quartzite-volcanite piles as far up as into dolomites and carbonaceous slates and acted as feeding channels for the effusives that occur in the carbonaceous slates.

Many of the wider dykes are often cut by younger metadiabase dykes 0.2 to 5 m wide that run subparallel to the main dyke (Fig. 64), suggesting that at least some of the older dykes were reactivated during later periods of volcanism.

Mafic magma also intruded as sills between the upper dolomite and carbonaceous slate layers of the Dolomite—Carbonaceous slate—Volcanite Formation. The biggest of them are the Hyypiä sill, which is about 80 m wide (Fig. 8), and the Sääperi sill, which is about 200 m wide (Fig. 21). The middle parts of these sills contain coarse variants, but towards the footwall and hanging wall the grain size decreases and the contact is occupied by a fine-grained seam. Owing to the scarcity of exposures, the internal structure of the intrusions is not well known. It is noteworthy that the sills are more gabbroic than diabasic (Fig. 67F) and thus resemble the hypabyssal intrusive rocks encountered in Soviet Karelia, e.g. in the Upper Jatulian formations, which Soviet research workers call gabbro-diabase (cf. Kharitonov 1966; Sokolov *et al.* 1970).

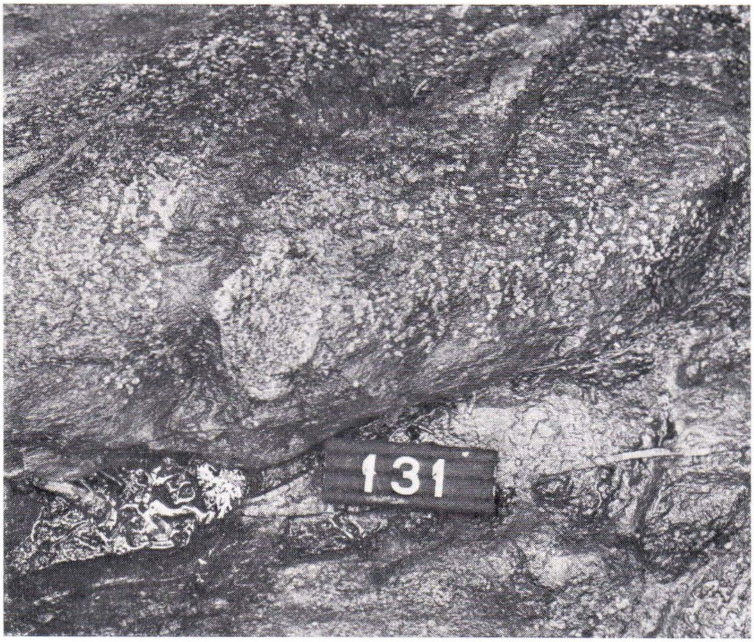


Fig. 65. Variolitic portion in first or second lava flow of the Volcanite Formation. Length of tag 12 cm. Sääperi, Värtsilä. Photo by E. Halme.

The mineralogical character of the effusive and hypabyssal rocks

Effusive rocks

The Volcanite Formation, which is the most marked occurrence of effusive rocks in the study area, is composed predominantly of lavas but also includes some pyroclastites.

The Kiihtelysvaara Volcanite Formation is composed of two or three lava flows whose lower and middle parts consist predominantly of slightly porphyric lavas, although small amounts of micro-ophitic variants occur as well. In the lower lava flow there are albite phenocrysts 0.2 to 0.8 mm long in a matrix composed of chlorite, albite, biotite and magnetite (Fig. 67A). The second and third lava flows are similar to the first except that, as well as albite, they often contain light-coloured amphibole as phenocrysts and in the matrix (Fig. 67B). The micro-ophitic variants are composed of the same constituents but without phenocrysts. In the Sääperi area, the second or third lava flow exhibits variolitic variants with abundant varioles 2 to 15 mm in diameter (Fig. 65). The varioles often have a radial chlorite shell and an albite core that also contains small amounts of quartz. The spaces between the varioles are filled with a fine-grained mass of biotite and quartz, which locally also contains hornblende phenocrysts. Magnetite, titanite, pyrite and chalcopyrite are the characteristic accessories in the volcanites. Albitic feldspar seems to be typical of the volcanites in the study area, a feature which they share with the spilitic rocks in the Karelidic schist zone in other regions (Meriläinen 1961; Piirainen 1968, 1969; Piispanen 1972).

Amygdules from 2 to 30 mm long (Fig. 66) are common in the upper parts of the first and second lava flows, where they are oriented more or less parallel to the contacts of the lava flows. The amygdules are usually filled with quartz or calcite, less often with epidote and hematite. In the Hyypiä area, the hanging-

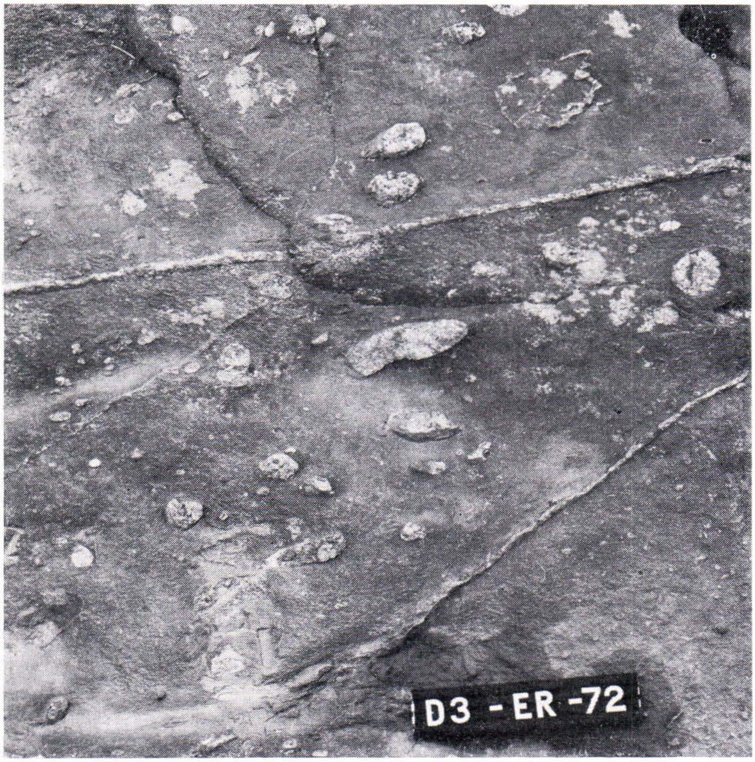


Fig. 66. Quartz amygdule-bearing portion in upper part of second lava flow of the Volcanite Formation. Length of tag 12 cm. Hyypiä, Kiihtelysvaara. Photo by E. Halme.

wall of the lava flow has fractured for a thickness of a couple of metres or so (Fig. 63). This heterogeneous rock is mainly composed of chlorite, albite, epidote, biotite and quartz. The amygdules in the rock are filled with milky quartz and epidote.

The topmost part of the third lava flow contains heterogeneous rocks that are presumably tuffs in origin. They are often fine-grained and contain chlorite, albite, epidote, quartz, titanite and magnetite. Minor agglomerate beds rich in chlorite and epidote have been encountered in the Sääperi area (Nykänen 1968).

The volcanic interlayers in the upper quartzites and in the dolomites that rest on them are mainly tuffs deposited in water. In many cases, however, the tuffaceous matter deposited together with quartz sands, and hence mixed rocks, or tuffites, were formed (Fig. 16). As a rule the tuffs are composed of chlorite, albite, biotite and quartz. The accessories are epidote, titanite and magnetite. No distinct lava rocks have been recognized apart from the lava bed in the upper part of the Dolomite—Volcanite Formation (the Volcanite Member) in the Viistola section. This lava bed is composed of chlorite, biotite, albite, quartz, titanite and magnetite. It also contains nodules of carbonate, quartz and hematite, presumably amygdules in origin.

The volcanic layers (Volcanite Member) that alternate with the dolomite and carbonaceous slate layers in the upper part of the Dolomite—Carbonaceous slate—Volcanite Formation consist mainly of somewhat porphyric lavas but also contain small amounts of tuffs. The lavas exhibit hornblende or plagioclase (albite-oligoclase) phenocrysts in a groundmass of hornblende, plagioclase, biotite, epidote, quartz, titanite and magnetite. Hornblende has partly altered into actinolitic amphibole or locally also into cummingtonite. The tuff layers associated with the lavas usually contain the same constituents as the lavas except that the abundance of amphiboles is lower.

Hypabyssal rocks

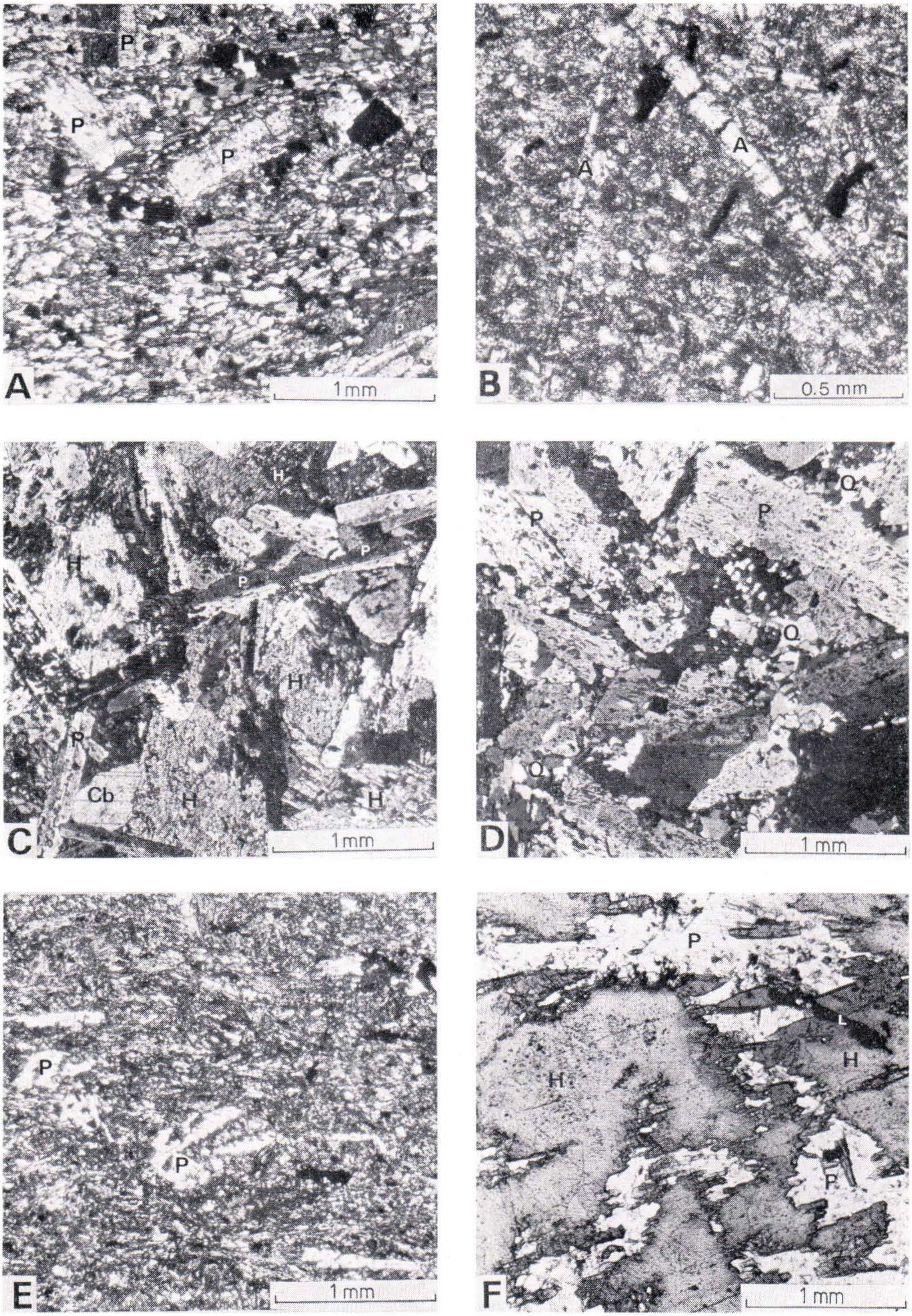
This group consists of those Karelian (Jatulian) mafic intrusive rocks that resemble diabases in structure and mode of occurrence. Since their mineral composition was altered during metamorphism they are often called metadiabases. Although hornblende is the predominant mafic mineral in these rocks today, it seems often to have been preceded by pyroxene. In spite of recrystallization, the primary structure is still often visible as blasto-ophitic texture. In some cases, however, the hornblende grains are so well developed that they may be primary. It is noteworthy that, as in the other rocks in the area, the degree of metamorphism varies from one site to the other, which is also reflected in the variation in mineral composition. Furthermore, some variation in composition is due to differentiation.

In the metadiabase dykes farther on within the Prekarelidic basement, the main constituents are green hornblende and plagioclase (Nykänen 1968, 1971b; Kallio 1976). The dykes also contain small amounts of biotite, quartz, epidote, carbonate, titanite and magnetite. Here and there hornblende exhibits slight chloritization. Some wider dykes have a differentiated and more silicic core, which, being paler in colour than the margins, is composed of plagioclase (albite-oligoclase) with some minor quartz, chlorite, biotite, titanite and magnetite.

Table 17
Mineral compositions of Karelian hypabyssal rocks in the Kiihtelysvaara—Värtsilä area. Determined by point-counting method.

	56	57	58	59	60	61	62	63	64
Amphibole	48.4	23.9	—	—	—	66.4	58.7	53.2	41.4
Biotite	14.5	4.2	+	0.9	+	0.3	+	1.9	8.5
Chlorite	0.3	11.2	56.2	68.8	20.4	+	1.8	+	2.7
Plagioclase	19.2	46.4	34.0	0.3	44.0	31.9	28.4	36.9	39.7
(An)	(30—33)	(27)	(18)		(8—10)		(28)	(27—32)	(30)
Epidote	13.7	6.0	0.4	5.1	+	+	7.3	+	3.9
Sericite	+	+	+	+	+	+	+	+	+
Carbonate	+	4.8	6.1	0.4	6.0	+	+	0.9	+
Quartz	+	+	+	24.1	20.4	+	+	+	+
Apatite	+	+	+	+	+	+	+	+	+
Titanite and leucoxen	3.7	3.3	2.3	0.2	7.8	+	3.6	+	3.7
Opaque	0.2	0.2	1.0	0.2	1.4	1.4	0.2	7.1	0.1
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

56. Medium-grained metadiabase from the margin of the differentiated Piippolanmäki dyke, from its eastern end, where the dyke is in a Prekarelidic quartz diorite (drill core No. R307/4241/66.85), Hyypiä, Kiihtelysvaara.
57. Medium-grained metadiabase from the southern margin of the Piippolanmäki dyke, from its western end, where the dyke is in the Lower quartzite (drill core No. R304/4241/44.60), Hyypiä, Kiihtelysvaara.
58. Medium-grained metadiabase from the northern margin of the Piippolanmäki dyke, from its western end (drill core No. R301/4241/56.65), Hyypiä, Kiihtelysvaara.
59. Chlorite-rich lens from the middle of the Piippolanmäki dyke, from its western end (drill core No. R304/4241/19.00), Hyypiä, Kiihtelysvaara.
60. Medium grained albitic metadiabase from the middle of the Piippolanmäki dyke, from its western end (drill core No. R301/4241/18.25), Hyypiä, Kiihtelysvaara.
61. Medium-grained metadiabase from a dyke that cuts the rocks of the Dolomite—Carbonaceous slate—Volcanite Formation (drill core No. R306/4232/38.65), Valkeavaara, Kiihtelysvaara.
62. Coarse gabbroic metadiabase from the middle of the Hyypiä sill (Specimen No. 9a/PV—73), Hyypiä, Kiihtelysvaara.
63. Medium-grained gabbroic metadiabase from the Sääperi sill (Specimen No. 11/LJP—75), Röykynvaara, Värtsilä.
64. Coarse gabbroic metadiabase from the middle of the Sääperi sill (Specimen No. 10/LJP—75), Jänisjoki, Värtsilä.


A
P
P
P
P
1 mm
B
A
A
0.5 mm
C
H
H
P
P
H
P
Cb
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1 mm
D
P
P
Q
Q
Q
1 mm
E
P
P
1 mm
F
P
L
H
H
P
1 mm

Note that in the metadiabase dykes that extend from the Prekarelidic basement to the Karelian formations the mineral composition changes rather sharply to the west of the contact between the rock complexes. This is well exemplified in the dyke at Piippolanmäki (Fig. 8), where hornblende predominates until close to the contact of the Prekarelidic basement with the Karelian formations (Table 17, No. 56). To the west of the contact, hornblende is intensely chloritized (No. 58) although there are some portions in the dyke in which hornblende has partly survived (No. 57, Fig. 67C). At the same time, plagioclase has graded into a more albitic variant (oligoclase-albite) containing epidote and carbonate as alteration products (No. 58). The differentiate (Fig. 67D) in the core, where the dyke is paler than at the margins, is composed predominantly of albitic plagioclase, quartz and chlorite (No. 60). The rock occasionally exhibits pyrite and chalcopyrite dissemination and some richer nests. A chlorite-rich lens with small carbonate amygdules has been encountered in the central part of the dyke (No. 59). A similar variation in mineral composition has also been noted in other dykes of the same type.

Some differentiated dykes, such as those at Kortevaara and Piipolanmäki, show mottled rock on both sides of the pale central differentiate near the orifice of the discharge channel. The mottles are biotite and magnetite aggregates. The rock is more magnetic than the wall rocks and leads to the bipartition observed in the magnetic anomalies of the dykes (Fig. 6B).

The younger metadiabase dykes that cut the wider metadiabase dykes (Fig. 64) are composed chiefly of hornblende and plagioclase (andesine). They also contain small amounts of biotite, epidote, titanite and magnetite. Whenever the dykes are wider than a couple of metres the core is medium-grained and ophitic, and the dykes have slightly porphyric, fine-grained margins. The narrow dykes are composed entirely of this kind of fine-grained rock. As a rule the phenocrysts are of plagioclase, sometimes also of hornblende (Fig. 67E). Small calcite amygdules have been encountered in the cores of some dykes. Westwards the younger dykes exhibit grading in mineral composition similar to that in the host dykes; in other words, the younger dykes were metamorphosed at the same time as the host dykes.

The metadiabases that cut the dolomites and carbonaceous slates are dark and medium-grained rocks that in spite of recrystallization, have locally preserved their ophitic structure. The main constituents are green hornblende and plagioclase (oligoclase-andesine). Minor biotite, quartz, carbonate, epidote, titanite and magnetite also occur (No. 61). In many dykes, hornblende is intensely altered into pale actinolitic amphibole. Chlorite is also present, although mainly in the shears.

Fig. 67. Photomicrographs of Karelian metavolcanics of the Kiihtelysvaara type.

A) Slightly porphyric lava with idiomorphic plagioclase crystals in a fine-grained groundmass consisting mainly of plagioclase, biotite, chlorite, magnetite and quartz from the lower part of the Volcanite Formation (Thin section No. 19530). Crossed nicols. B) Slightly porphyric lava with idiomorphic amphibole and plagioclase crystals in a fine-grained groundmass consisting mainly of amphibole, plagioclase, biotite and magnetite from the central part of the Volcanite Formation (Thin section No. 19528). Ordinary light. C) Medium-grained metadiabase consisting mainly of plagioclase, hornblende, chlorite, biotite and carbonate from the southern margin of the Piippolanmäki dyke (Thin section No. 17935). Crossed nicols. D) Medium-grained more albitic metadiabase consisting mainly of albitic plagioclase, quartz, chlorite, carbonate and leucoxene from the central part of the Piippolanmäki dyke (Thin section No. 17735). Crossed nicols. E) Slightly porphyric fine-grained metadiabase from a narrow dyke cutting the Kortevaara metadiabase dyke (Thin section No. Ku 00296). Ordinary light. F) Gabbroic metadiabase from the central part of the Hyypiä metadiabase sill (Thin section No. Ku 00297). Ordinary light. Markings: Q = quartz; P = plagioclase; H = hornblende; A = amphibole; Cb = carbonate; L = leucoxene. Photo by E. Halme.

In the wider metadiabase sills, e.g. at Hyypiä and Sääperi, the coarsest variant occur in the cores of the sills, and the grain size diminishes towards the contacts. In these two intrusions the upper and lower parts are composed mainly of medium-grained gabbroic rocks with green hornblende and plagioclase (oligoclase-andesine) as major minerals (Table 17, No. 63). Other mineral constituents are biotite, epidote, carbonate, chlorite, titanite and magnetite. The cores of the sills are composed of rather coarse-grained variants (Fig. 67F) in which hornblende grains from 0.5 to 2 cm exhibit poikilitic intergrowth, e.g. with plagioclase (Nos. 62 and 64). In the western part of the Sääperi sill the rocks are intensely altered; this has led to chloritization of the mafic minerals and the formation of epidote and carbonate due to the decomposition of plagioclase. It is likely that during the Svecokarelidic folding the western part of the Sääperi sill tilted steeply westwards and was metamorphosed under the conditions of greenschist facies.

The chemical character of the effusive and hypabyssal rocks

The metalavas and metadiabases in the study area bear a great resemblance to each other in chemical composition. There are, however, some marked differences (Table 18), the most significant being due to the CaO content, which is lower in metalavas (Nos. 34 and 35) than in metadiabases (Nos. 29—33). The somewhat higher silica content in the effusives may be partly attributed to secondary gain (cavities and amygdules filled with quartz).

The analytical data have been visualized by two conventional triangular diagrams (Fig. 68A—B), which plot the data from Table 18. For the sake of comparison, the compositions of the six metadiabases reported by Piirainen (1969) from the Koli area have also been plotted in the diagrams. Furthermore, the boundaries of the field that describes the composition of the differentiated tholeiitic diabase sheets in Tasmania (after McDougall 1964, p. 124) have been drawn on one of the diagrams. The diagrams show that the compositions of the metadiabases in the study area (solid circles) fall rather close to each other in the same part of the field as do the compositions (open circles) of the rocks of the tholeiitic association at Koli (Piirainen 1969); even so, the composition of the sample taken from the orifice of the discharge channel (No. 30) is already rather near that of the rocks of spilitic association (open squares). Moreover, the variation in composition largely takes place within the boundaries of the composition of the Tasmanian differentiated diabases (Fig. 68A), most of the points plotting at the tholeiitic end of the field, while the composition of the variant richest in silica (No. 31) is already close to the granophyre end.

In the diagram the compositions of the metalavas in the study area (solid triangles 34—35) as well as those of the rocks of the spilitic association at Koli plot outside the area occupied by the above points. This is because of the lower CaO and higher Na_2O contents in the metalavas than in the rocks of the spilitic association and also because of the increase in the Fe_2O_3/FeO ratio. The lowest flows in the metalavas (No. 35) are slightly more siliceous than the upper flows (No. 34), which is also reflected in the mineral composition (p. 117). The chemical composition of the metalavas is somewhat similar to that of basalts, although their low CaO content is anomalous.

Even though the rocks that crystallized under the conditions of greenschist facies, as did the rocks of the Volcanite Formation, contain albite as an essential constituent, the fundamental differences in chemical composition

Table 18
Chemical composition of various Karelian effusive and hypabyssal rocks in the Kiihtelysvaara—Värtsilä area (weight per cent).

	29	30	31	32	33	34	35
SiO_2	49.55	50.45	64.40	50.75	50.66	52.50	56.37
TiO_2	1.80	1.64	0.82	1.72	1.09	1.57	1.84
Al_2O_3	12.57	15.58	13.30	13.25	16.47	16.46	13.63
Fe_2O_3	2.75	3.90	2.01	2.10	1.44	4.98	4.75
FeO	12.20	9.37	7.47	11.76	8.87	9.00	7.21
MnO	0.27	0.22	0.17	0.25	0.16	0.14	0.09
MgO	6.47	4.82	1.32	6.19	7.21	4.81	7.14
CaO	8.78	6.94	3.40	8.44	8.36	1.68	0.43
Na_2O	2.40	3.39	5.67	2.78	3.12	3.96	2.75
K_2O	1.30	1.27	0.16	0.51	0.70	2.05	0.52
P_2O_5	0.08	0.22	0.14	0.16	0.12	0.15	0.35
CO_2	0.00	0.00	0.05	0.47	0.00	0.00	0.00
H_2O^+	1.48	1.98	0.77	1.67	1.46	2.55	4.39
H_2O^-	0.03	0.09	0.05	0.07	0.04	0.07	0.07
S	0.17	—	0.04	—	—	—	0.01
	99.85	99.87	99.77	100.12	99.70	99.92	99.55
$—O = S_1$	0.04		0.01				
	99.81		99.76				

35 a

Emiss. Spec.	» Ba	30 ppm
	» Zr	0.03 %
	» V	0.02 %
	» Co	0.004 %

35 b

XRF	» Rb_2O	0.00 %
	» SrO	0.00 %
	» BaO	0.06 %

29. Medium-grained metadiabase from almost the undifferentiated dyke that cuts the rocks of the Prekarelidic basement complex, Tervavaara, Värtsilä. Analyst: A. Heikkinen (Nykänen 1968 p. 28).
30. Medium-grained metadiabase from the end of a dyke that cuts the Lower quartzite Formation, from the ancient orifice in the eruption channel (Specimen No. 91b/EJL—61), Silmunvaara, Kiihtelysvaara. Analyst: P. Ojanperä.
31. Medium-grained, more albitic portion from the core of the differentiated dyke that cuts the rocks of the Prekarelidic basement complex, Kutsu, Tohmajärvi. Analyst: A. Heikkinen (Nykänen 1968, p. 28).
32. Medium-grained metadiabase from the roughly 10 m wide dyke that cuts the rocks of the Dolomite—Carbonaceous slate-Volcanite Formation (drill core No. R1/Kvh—61/30.50), Haluksenlammit, Kiihtelysvaara. Analyst: P. Ojanperä.
33. Coarse-grained gabbroic metadiabase from the core of the Hyypiä sill, Hyypiä, Kiihtelysvaara. Analyst: A. Heikkinen and R. Saikkonen (Nykänen 1971b, p. 40).
34. Slightly porphyric metalava from the Volcanite Formation (Specimen No. 73/PO—61), Valkeavaara, Kiihtelysvaara. Analyst: P. Ojanperä.
35. Amygdaloidal metalava from the Volcanite Formation, Raatevaara, Kiihtelysvaara. Analyst: P. Ojanperä. Emission spectr. analyses by A. Löfgren. XRF analyses by V. Hoffrèn (Nykänen 1971b, p. 40).
35 a, 35 b. Trace-element abundances, same rock as analysis 35.

between the effusive lavas and those crystallized in the discharge channels cannot be attributed solely to metamorphism. Hence, it is likely that the chemical composition of the tholeiitic magma that flowed in the discharge channels changed radically either during or soon after eruption. This may have been caused indirectly by the concentration of volatiles in magma (see Meriläinen 1961; Piirainen 1969), viz. the abundance of amygdules. The spilitic variants were possibly mainly produced when the mafic magma was contaminated locally while penetrating to the surface through psephitic and psammitic layers saturated with sea-water (cf. Piispanen 1972). It is, however, possible that the alterations took place later metasomatically when the effusives were covered by water-bearing sedimentary piles (e.g. Levi 1969).

The distributions of the trace elements Cu, Ni, Zn, Pb and Co and of sulphur and phosphorus in the metavolcanites and metadiabases in the study area were studied statistically. The analytical data were processed by ADP into graphic presentation with the aid of expanded

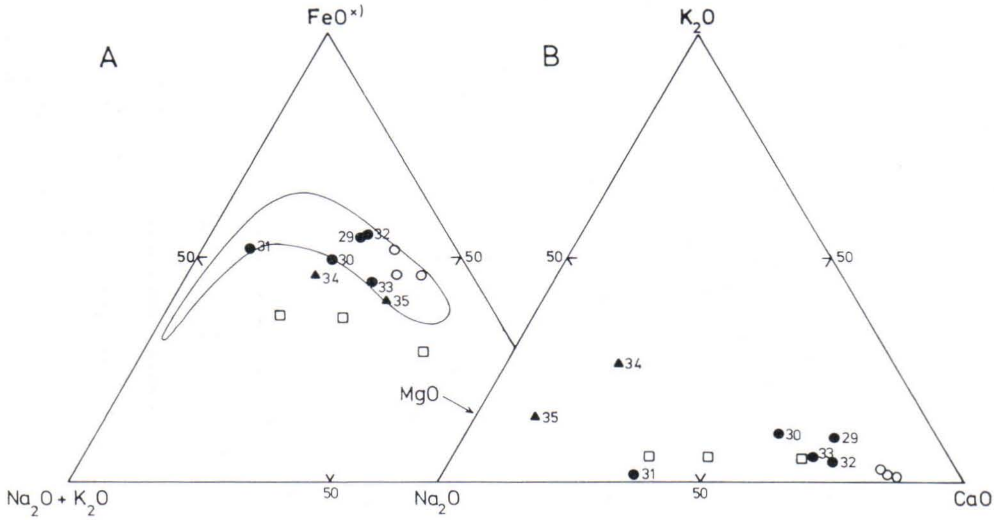

Fig. 68. Plot of analyses (weight proportions of oxides) of two Karelian metavolcanites and 11 Karelian metadiabases, and a shaded field enclosing diabases and granophyres of three Tasmanian sheets; after Mc Dougall (1964, p. 124). Solid circles for individual analyses of metadiabases, and solid triangles for individual analyses of metavolcanites in the Kiihtelysvaara—Värtsilä area. Open circles for individual analyses of metadiabases of tholeiitic association, and open squares for individual analyses of metadiabases of spilitic association in the Koli area; after Piirainen (1969, Table No. 2 and Table No. 6). FeO includes only Fe^{++}.

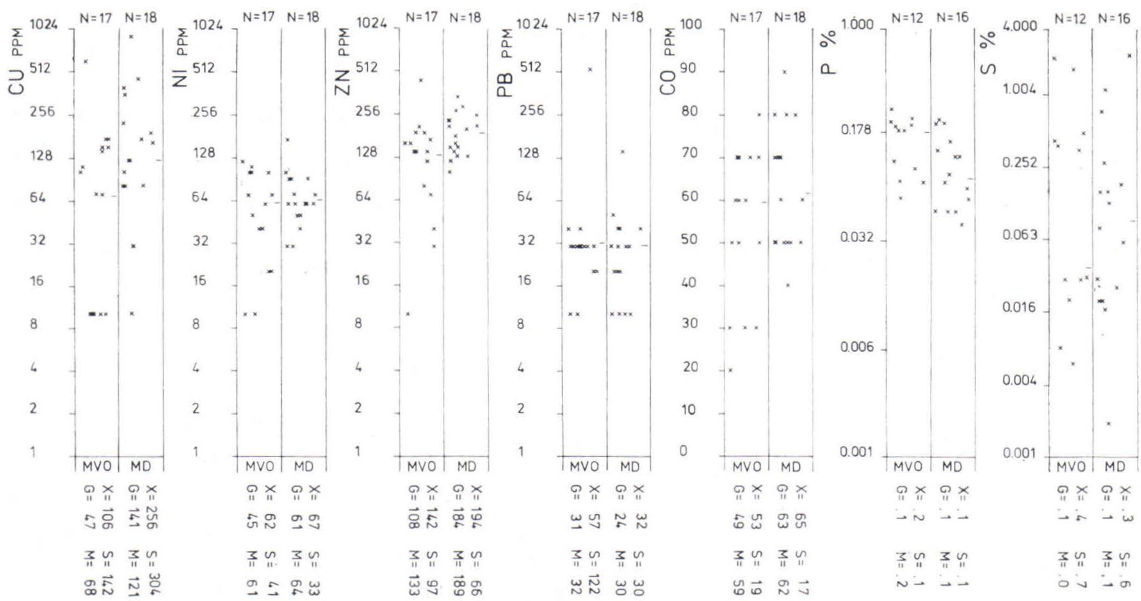

Fig. 69. Scatter diagram showing the distribution of copper, nickel, zinc, lead, cobalt, phosphorus and sulphur in Karelian metavolcanites and metadiabases in the Kiihtelysvaara—Värtsilä area. Cu, Ni, Zn, Pb and Co were assayed by AAS and sulphur by Leco M 621 sulphur analyser (A. Nurmi). Phosphorus was analysed at the ore laboratory of the Chemistry department. Scales of columns logarithmic. Symbols: MVO = metavolcanites; MD = metadiabases; X = arithmetic mean; M = median; S = standard deviation; N = number of observations. The medians marked on the column by short horizontal lines.

scatter diagrams, which also give the arithmetic mean, geometric mean, median and standard deviation (Fig. 69). The analytical data show that in the study area the abundances of the above elements are in many cases higher in the metavolcanites and metadiabases than in the other rocks (Figs. 43, 51 and 58). Locally some metadiabase dykes exhibit higher Cu abundances in their cores (in nests 0.3—0.5 % Cu) e.g. at Kortevaara (Fig. 6A) and Piippolanmäki (Fig. 8). They are, however, without economic significance.

The ore laboratory of the Geological Survey analysed uranium from twenty-five metavolcanite and metadiabase samples from the study area. Most of the samples show U values less than 1 ppm; only two contained 3 ppm U. Hence, the slight enrichment of U in the contacts of some dykes with the quartz conglomerate beds is merely a secondary phenomenon (cf. Piirainen 1968). Some samples were also assayed for Th by gamma spectrometer; the Th contents were found to be slightly higher than those of U.

In general the chemical composition of the volcanogenic rocks in the study area corresponds to the type composition of continental Fe-rich tholeiitic rocks described by Carmichael, Turner and Verhoogen (1974).

Conclusions concerning the cause of the eruptions and the intrusion mechanism

The first volcanic episode that can be recognized in the Kiihtelysvaara—Värtsilä area took place during the marine regression stage that interrupted the deposition of quartz sands of the Orthoquartzite Suite at Kiihtelysvaara. Movements in the Earth's crust associated with regression produced ruptures that later filled with mafic magma. In some places these ruptures reached the surface through the partly consolidated quartz sand piles and thus the magma discharged onto them. On account of the regression, most of the eruptions were onto dry land, although the presence of fragmented lavas indicates that some were into a shallow littoral sea (Fig. 63). No distinct pillow-lava structures that would refer to submarine eruptions have been found. The discharging magma was obviously very fluid and spread over a large area; hence, only few pyroclastites were formed. The variation in mineral and chemical compositions in the effusives (page 117 and Table 17, Nos. 34—37) suggests that there were several successive eruptions. The effusives that erupted at that time constitute the Volcanite Formation, which today is from 30 to 80 m thick. Originally it was probably even thicker because the erosion following the episode of volcanism seems to have substantially denuded the volcanites.

The thin tuff and tuffite interbeds in the upper quartzites and dolomites suggest that some kind of weak volcanic activity continued even after the period of more intense volcanism.

During the marine regression that temporarily interrupted the formation of carbonates, the Earth's crust became unstable and the movements in it resulted in volcanic activity that manifested itself in the study area as minor eruptions (Volcanite Member of the Dolomite—Volcanite Formation).

Later, during the marine regression preceding the Flysch stage, the movements in the crust resulted in more intense volcanic activity. This is revealed as the intrusion of mafic dykes and sills in the sediments of the upper part of the Kiihtelysvaara Carbonate—Black slate Suite and as minor eruptions of effusive rocks (Volcanite Member of the Dolomite—Carbonaceous slate—Volcanite Formation). Particularly in the carbonaceous slate the intrusion took place as conformable sills, possibly on account of the well-developed parting parallel to the bedding (see Leaman 1975).

Closer examination of the bedrock shows that the directions of the metadiabase dykes are distributed so that the expansion was greatest perpendicular to the dykes that trend roughly NW—SE. Kallio (1976) has come to a rather similar conclusion. This suggests that the tension field was oriented in such a way that the maximal tension was perpendicular to the NW—SE-trending dykes. Hence, the maximal compression was effective roughly parallel to the dykes. This would imply that most of the dykes filled the tension joints of the field with the above orientation, although it is true that some of the dykes trend more or less parallel to the shear directions.

There is, however, one point that contradicts the intrusion mechanism suggested above. As noted previously, intensification of volcanic activity seems always to have been intimately associated with marine regression, which presumably resulted from tangential compression perpendicular to the sedimentation basin trending roughly NW—SE. The compression probably broke the basement into NW—SE-trending platy blocks, which were slightly translated in relation to each other and led to the change in the configuration of the basin. On account of the slightly sinuous bending in the basement, the compressional field produced tension joints in the basement principally perpendicular to the compression as well as diagonal shear points (cf. De Sitter 1956, Fig. 56). Some deep-seated fractures were obviously formed, along which the magma later invaded the upper levels of the crust. The elastic release of the NE—SW compression as well as the intrusion of the mafic magma into the tension joints and some of the shear joints may have resulted in the formation of the dykes. The adjacent maxima parallel to the trends of the dykes may refer to different injection periods. Since, however, the intersections of dykes with different orientations have not been found exposed and since their datings are merely tentative, it has not been possible to group them according to age. The issue is further complicated by the fact that some dykes have demonstrably been active more than once (p. 116).

SUMMARY AND DISCUSSION

The geologic evolution of the Karelian formations

In the development history of the Karelian formations the initial stage succeeding the Presvecokarelidic times (2600—2800 Ma) is characterized by the denudation of the mountain ranges, the cratonization of the continent with subsequent levelling almost into a peneplane, and the formation of a weathering crust covering the surface. This older, i.e. Early Karelian weathering crust, seldom escaped erosion and only survived when covered by younger sediments. The present author believes that the breccia-conglomerates overlain by Karelian sediments are almost synchronous with the Early Karelian weathering crust.

It is the opinion of this author that the rocks of the Kiihtelysvaara Arkosic Suite (Appendix 2), which rest on the Prekarelidic basement with an angular unconformity in between, represent the feldspathic residual sediments of the first cycle composed of the disintegration and weathering products of rocks of the basement. This fluvial association begins with alluvial-fan types of deposits that developed under desert-like conditions. They are followed by braided-river deposits that, presumably owing to levelling of the relief and changes in climate, were formed on low-lying land. The deposits probably reached their thickest in tectonically active areas, such as graben bordered by faults, where several adjacent and overlapping alluvial fans were associated with each other.

The poorly consolidated arkose sandstone layers of the Kiihtelysvaara Arkosic Suite were later weathered in situ down to a depth of several tens of metres. Although today the weathering products occur as authigenic sericite, the weathering was in fact of kaolin type. The kaolinization of that time hardly has any direct relation to other kaolin occurrences in sericite schists or arkosites and overlain by Jatulian quartzite beds, as e.g. at Pihlajavaara, Puolanka (Väyrynen 1928, 1954; Laajoki 1975a). These occurrences are probably associated with later fractures, and their kaolin is demonstratably of later origin, perhaps Preglacial (Laajoki 1975a). This younger or Early Jatulian weathering crust did not, however, survive as such; erosion had time to denude it, locally even the underlying fresh layers as well, before the sedimentation of quartz sands commenced.

The rocks of the Kiihtelysvaara Orthoquartzite Suite rest on the rocks of the Kiihtelysvaara Arkosic Suite with an unconformity marked by an erosion surface in between. As a matter of fact, the deposits are separated from each other by a gentle angular unconformity. The rocks of the lower part of the Orthoquartzite Suite, that is, the Lower quartzite Formation, are, in the opinion of the present author, paralic sand sediments of the second cycle at least (Appendix 2) and were formed from the rocks of the Arkosic Suite in the course of a long history of transport and reworking. The grading of the rocks in the section of the Lower quartzite Formation and the change in structural and textural features suggest transgressive-regressive sedimentation. Cross-bedded quartzites with quartz conglomerate and clay slate interbeds occur in the footwall. Higher up these are followed by cross-bedded quartzites poorer in micas. In the hanging wall there are cross-bedded quartzites with quartz conglomerate interbeds and fresh feldspar increment indicating uplift of the erosion area, i.e. the onset of regression.

The release of the tension field built up in the Earth's crust during regression presumably caused ruptures in the crust; some of these reached the surface and acted as channels for fissure eruptions. The structures of the volcanites suggest that the mafic lava erupted mainly on the partly emerged quartz sandstone piles but partly also in the shallow littoral sea. The eruptions probably took place in succession, two or three lava flows having been discerned in the Volcanite Formation.

The change in topography during the volcanic stage gave a good start to erosion. In more intensely uplifted areas the erosion wore away the volcanic rocks and even cut the quartzites of the Lower quartzite Formation, as e.g. northwest of Karsikkojärvi, Kiihtelysvaara (Fig. 5).

The quartzites in the upper sequence of the Kiihtelysvaara Orthoquartzite Suite, or the Upper quartzite Formation, which were deposited after the volcanism-erosion stage, contain decomposition and weathering material from mafic effusives and are thinner than the quartzites in the Lower quartzite Formation. The grading of the rocks in the section of the Upper quartzite Formation (Appendix 2) and the change in their structural and textural features show that sedimentation was transgressive. In the footwall of the formation there are often coarse and cross-bedded quartzites with quartz conglomerate and tuff and tuffite interbeds. Higher up in the sequence there are more fine-grained and cross-bedded or layered quartzites with tuff, tuffite and clay slate interlayers. The hanging wall exhibits fine-grained horizontally layered or cross-bedded quartzites with abundant clay slate interlayers and mica—carbonate matrix. The presence of carbonate in the matrix of the hanging-wall quartzite indicates that the terrigenic rocks (quartz sandstone, shales) graded into carbonate sediments, i.e. conditions became marine.

The quartzites of the Upper quartzite Formation graded into the carbonate sediments of the Kiihtelysvaara Carbonate-Black slate Suite. The irregular distribution of the lower section of the above Suite, or the carbonate rocks of the Dolomite—Volcanite Formation, and the variable thickness of the occurrences suggest that the conditions were not everywhere equally favourable for the formation of carbonates. The rate of carbonate formation was probably highest in small offshore basins and in sea gulfs. The grading of the rocks as revealed in the section of the Dolomite—Volcanite Formation (Appendix 2) and the change in their structural and textural features indicate that the carbonate rocks were largely mechanically deposited, i.e. they are allochthonous and that sedimentation was transgressive-regressive. Thin quartzite and clay slate interlayers occur in the footwall of the formation between the dolomitic carbonate rock layers. Higher up the dolomite carbonate rocks with calcite nodules contain less quartz and clayey matter but, instead, some tuffaceous interlayers and abundant cherty constituents. In the hanging wall, clay slate interlayers and minor volcanic beds occur once more, demonstrating that transgression switched to regression.

Upwards in the sequence the carbonate sediments grade into fresh water terrigeneous clay slates of the Hematite rock—Quartzite Formation; some minor hematite ores were formed in them locally, perhaps in the same way as in the lake ores of karst type (cf. Strakhov 1969, pp. 149—153). In the final stage of regression these iron-rich layers were covered by a thin blanket of cross-bedded quartz sands.

The rocks of the upper section of the Kiihtelysvaara Carbonate—Black slate Suite or the Dolomite—Carbonaceous slate—Volcanite Formation, rest unconformably on the quartzite of the Hematite rock—Quartzite Formation, perhaps with a hiatus surface in between. The grading of the rocks noted in the section of the Dolomite—Carbonaceous slate—Volcanite Formation and the change in their structure and texture (Appendix 2) show that the carbonate sediments were deposited in situ, i.e. they are autochthonous and that the sedimentation was transgressive-regressive. At first the configuration of the shore line probably changed to allow the formation of carbonate once more. Later short-lived fluctuations in the sea level resulted in changes in the configuration of the small basins and sea gulfs. It is possible that semi-isolated minor basins and bays were formed offshore, and that under reducing conditions the prevailing gyttja and carbonaceous matter were slowly deposited; carbonate suspension on the other hand was deposited only sporadically. The occurrence of sandy interlayers among the alternating dolomite and carbonaceous slate layers in the hangingwall of the Dolomite—Carbonaceous slate—Volcanite Formation suggests that eventually the deposition took place in rather shallow water.

The changes in the littoral configuration obviously reflect movements deeper in the Earth's crust. During the final stage of sedimentation these movements triggered volcanic activity, which is manifested by the mafic dykes and sills in the sediments of the Dolomite—Carbonaceous slate—Volcanite Formation and the minor effusive beds on top of them.

Probably on account of bending in the sedimentary piles at the margin of the sedimentation basin, the existing sedimentary piles and associated volcanites later emerged in many places and underwent substantial erosion before the sediments of the Kalevian flysch started to deposit.

The rocks of the Kalevian Kiihtelysvaara Turbidite Suite usually rest on the rocks of the Kiihtelysvaara Carbonate—Black slate Suite with an unconformity marked by an erosion surface in between. A more or less continuous layer of polymictic conglomerate delineates the lower boundary of the Kalevian group rather accurately. The fragments of this conglomerate (Table 15) suggest that erosion cut the underlying layers locally to such a depth that in places the conglomerate rests on quartzites or effusives, to mention but two rocks. The grading of the rocks shown in the section of the Turbidite conglomerate—quartzite Formation and the Mica schist Formation (Appendix 2) and the change in their structure and texture indicate that the Turbidite conglomerate—quartzite Formation represents the coarse turbidite sediments of the early geosynclinal stage that derived from a craton and on top of which the normal greywacke-turbidite type of sediments of the Mica Formation later deposited. The conglomerate and quartzite probably pinch out farther west, i.e. the boundary of the Kalevian flysch is not well established there. No such case, however, was encountered within the study area.

Sedimentation, volcanism and tectonics

In terms of structural position, two facies types occur in the Karelian Sequence in the Kiihtelysvaara—Värtsilä area (Appendix 2): that of the stable platform, or cratonic facies, and that of the unstable mobile zone, or geosynclinal facies.

The Prejatulian Kiihtelysvaara Arkosic Suite represents sediments of steeply depressed craton blocks, or Graben Assemblage. The fundamental cause of graben sedimentation may have been the switch of the compressional field into a tensional field during the denudation of the Prekarelidic mountain ranges. This led to uplift and fault subsidence in the Earth's crust, on the margins of which the sedimentation mainly took place. No signs of simatic volcanism, which is frequently associated with the formation of horst and grabens (Aubouin 1965, p. 101; Pettijohn 1975, p. 574), have been noted in the study area, although it has been reported elsewhere in the Karelidic zone (cf. Isohanni 1971, p. 31; Silvennoinen 1972, p. 41; Sokolov *et al.* 1972, p. 78; Perttunen 1975; Laajoki and Saikkonen 1977, p. 119 and Fig. 72).

The cratonic Jatulian Orthoquartzite—Carbonate Assemblage (Appendix 2) seems to represent sedimentation at the margin of a flat-lying and stable land mass. Even so, the deposition surface seems periodically to have emerged. The Black slate Assemblage, which succeeds the Orthoquartzite—Carbonate Assemblage i.e. the euxinic shallow-water assemblage, indicates significant changes in the configuration of the sedimentation basin, possibly the bending of the sedimentary piles at the margin of the basin. Volcanic activity took place from time to time during the deposition of the quartzite, carbonate rocks and carbonaceous slates. The mode of occurrence, structure and texture of the effusives and hypabyssal rocks of Kiihtelysvaara type are suggestive of cratonic fissure eruptions of flood-basalt type (cf. Svetov 1976, pp. 46—48). The volcanic activity seems to have been distinctly correlated with the regression stages, and the eruptions took place principally during the three injection periods.

The onset of the geosynclinal Kalevian flysch sedimentation with its clastic sediments (conglomerates, quartzites) is not typical of flysch; rather, on account of its transgressive characteristics, it shows molassic features. According to Auboin (1965, p. 133), however, flysch may be locally transgressive when in a miogeosynclinal basin close to the foreland, as seems to be the case here (cf. Huhma 1976, p. 22). Hence, it may be presumed that in the Kiihtelysvaara—Värtsilä area the lowest flysch conglomerates and quartzite represent a sediment wedge that was chaotically washed out to the margin of the miogeosynclinal basin (the Höytiäinen basin in the west) from the foreland (the Prekarelidic basement complex in the east). Later the sedimentation continued, however, with normal greywacke-turbidite types of sediments (graded-bedded mica schists). The absence of mafic volcanites, expecially ophiolites, from the Kalevian Kiihtelysvaara Turbidite Suite suggests the existence of a miogeosynclinal domain in the Höytiäinen basin. The occurrence of volcanites of the lower horizon as fragments in the conglomerates at the base of the Turbidite Suite shows indisputably that volcanic activity came to an end in the study area before the onset of Kalevian miogeosynclinal flysch sedimentation.

Lithostratigraphic correlation

In the Kiihtelysvaara—Värtsilä area the rocks of the Arkosite Formation rest on the Prekarelidic basement with an angular unconformity in between. Lithologically they are very similar to the conglomerates and associated arkoses described by Eskola (1919) from Soviet Karelia, e.g. from east of Lake Seletskoje, and which he called the Sariolan formation. These and the associated quartzites were included by Ramsay (1906) in the Jatulian formations. Väyrynen (1933, 1954), like Eskola (1948, 1963), held that the Sariolan and Jatulian formations were closely related to each other; he maintained that the Jatulian formations apparently derived from the Sariolan when the latter were undergoing chemical weathering while still unconsolidated. The protagonists of this concept did not consider the Sariolan formations as separate from the Jatulian even though there seemed to be a hiatus in places; instead, they included them as a Sariolan facies in the Jatulian.

The presence of kaolin type of weathering crust separating Prejatulian from Jatulian formations (Early Jatulian weathering crust in the hanging wall of the Arkosite Formation) is indisputable in the Kiihtelysvaara—Värtsilä area (cf. Negrutsa 1965, 1971). The weathering crust itself, however, does not demark a well-defined correlation horizon, several weathering crust horizons having been encountered at different levels in various areas. Two of these are known in the study area but up to three in the Jangozero area in Soviet Karelia (Negrutsa 1965). The rocks of the Arkosite Formation, as a partially consolidated deposit, underwent not only regional chemical weathering but also marked erosion before the deposition of the Jatulian quartzites. On account of erosion, there is an angular unconformity between the Prejatulian and Jatulian formations (Arkosite Formation/Lower quartzite Formation) but at such a low angle that in many places the deposits are subparallel (cf. Sokolov *et al.* 1972).

In general, the Graben or half-graben Assemblage represented by the Kiihtelysvaara Arkosic Suite constitutes such an isolated occurrence in the Karelian Sequence that it cannot be understood as a Jatulian subfacies; it is more logical to consider it as a discrete group. Owing to overuse of the term »Sariolan», however, this group should preferably be designated Prejatulian, even though beyond the Kiihtelysvaara—Värtsilä area the group may differ conspicuously from this »basic» type.

Somewhat north of the study area, in the Koli—Kaltimo area, the Kiihtelysvaara Arkosic Suite does not occur as such, although the basal formations in the Koli—Kaltimo area (Väyrynen 1954; Gaál 1964; Piirainen 1968, 1976) may be compared with it (Table 19).

Still farther north, in the South Puolanka area (Väyrynen 1954; Laajoki 1973, 1976; Laajoki and Saikkonen 1977), there is an arkosite and conglomerate deposit, or the Arkosite Formation (Table 19), that may be comparable; comparison is, however, made difficult by the lack of exposures and more intense deformation of these rocks.

The Lower quartzite Formation in the Kiihtelysvaara—Värtsilä area (Table 19) and the quartz conglomerate interbeds and pelitic intercalations associated with its lower part correlate with Orthoquartzite I in the Koli—Kaltimo area (Piirainen 1968; Piirainen *et al.* 1974). In contrast, the correlation between the Upper quartzite Formation and the Arkose quartzite Formation in the Koli—Kaltimo area (Piirainen 1968; Piirainen *et al.* 1974) is not so straightforward because the volcanite horizon corresponding to the Volcanite Formation in the study area is absent or has disappeared (Table 19). It is possible that in the Koli—Kaltimo area events took place that were analogous to those in the northern part of the study area, where the effusives of the Volcanite Formations were eroded away during the erosion stage following the volcanic stage. This is suggested by such features as the hiatus between Orthoquartzite I and the Arkose quartzite Formation (Piirainen *et al.* 1974), and the volcanite fragments in the conglomerate interbeds in the footwall of the Arkose quartzite Formation in the Höytiäinen arc. The relation between the Upper quartzite Formation of the study area and Orthoquartzite II in the Koli—Kaltimo area (Piirainen 1968; Piirainen *et al.* 1974) is considerably more obscure, partly because, in the Koli—Kaltimo area, the dolomite deposit that succeeds the quartzite is either absent or was not developed at all (Table 19).

The Lower quartzite Formation in the study area correlates with Quartzite I in the South Puolanka area (Laajoki 1973, 1976; Laajoki and Saikkonen 1977), although the correlation between the upper boundaries is rather vague (Table 19). The correlation between the Upper quartzite Formation of the study area and Quartzites II and III in the South Puolanka area may correspond to each other, since they have certain features in common, e.g. pelitic intercalations and carbonate in the uppermost layers (Table 19).

The Marine Jatulian rocks defined by Väyrynen (1933, 1954) constitute one of the most easily recognisable stratigraphic horizons in the Karelidic schist zone in eastern Finland. With the increase in our information, however, concepts of the regional distribution of the association have changed and the term has come to be applied in a somewhat wider sense: it now refers to the dolomite—phyllite—carbonaceous slate formation with associated volcanic rocks that rests on the Jatulian quartzites (cf. Nykänen 1971c; Silvennoinen 1972; Laajoki 1973). The most striking feature is that the rocks corresponding to the Marine Jatulian are lacking from the Koli—Kaltimo area (Piirainen 1968; Piirainen *et al.* 1974). It should be remembered, however, that these rocks are also absent from the northernmost part of the study area or, if present, they are of minimal thickness (p. 48). By and large, the Dolomite—Volcanite Formation together with the Hematite rock—Quartzite Formation and the Dolomite—Carbonaceous slate—Volcanite Formation correlate with the Dolomite—Phyllite Formation in the South Puolanka area (Laajoki 1973; Laajoki and Saikkonen 1977) although between individual members the correlation is not complete (Table 19).

As for the term Marine Jatulian proposed by Väyrynen (1933, 1954), it must be pointed out that the sedimentation environment was not always as marine as the term implies; for example, in the study area, the Marine Jatulian comprises two transgression—regression cycles and during the regression clastic sediments were also deposited.

The position of the »Kalevian» phyllites and mica schists in the Karelidic schist zone of North Karelia (Frosterus 1902, Ramsay 1902) in relation to the Jatulian formation had been

Table 19

Comparison of the stratigraphy of the Karelian rocks in the Kiihtelysvaara—Värtsilä area with the Karelian rock sequences in the Koli—Kaltimo area and the South Puolanka area and with the Karelian rocks of the Suojärvi—Pialozerian type in Soviet Karelia.

SOUTH PUOLANKA AREA (Laajoki 1973, 1976; Laajoki and Saikkonen 1977)		KOLI—KALTIMO AREA (Piirainen 1968, and Piirainen et al. 1974)	
		(Höytiäinen arc)	*(Koli)*
	Pyssykulju Quartzite Fm — quartzite — conglomerate	Greywacke schist Fm	
	— — — — — unconformity — — — — —	— — — — — — — — — hiatus — — — — — — — —	
(western margin) Dolomite—Phyllite Fm — black schist and phyllite			
— iron formation			
— quartzite and phyllite		?	
— — unconformity? — — — —			
		?	
— dolomite — local tuffite			
		?	
	Quartzite III	Orthoquartzite II	
	Quartzite II	Arkose quartzite Fm	
		— — — — — — — — — — — — — — hiatus — ?	
	Quartzite I		Orthoquartzite I
— — — — — — — — — — — major unconformity — — — — — — — — — — —			— — — hiatus — ?
Arkosite Fm — feldspar-bearing quartzite, arkosite — local metavolcanics — basal arkosite and conglomerate			Basal Fm
— — — — — — — — — — — great discordance — — — — — — — — — — — —			great discordance — —
Prekarelidic Basement Gneiss Complex		Basement Complex	

a subject of controversy from the turn of the century right up to when Väyrynen (1933, 1954) convincingly demonstrated that the Kalevian phyllites and mica schists rest unconformably on rocks of the Jatulian group, as is shown by some Kalevian basal conglomerates.

On account of polarity in the geosyncline and the non-contemporaneous development of the basins, the Kalevian flysch does not make a good correlation horizon. In the study area, however, the miogeosynclinal flysch is in a special position, its eastern margin being auto-

KIIHTELYSVAARA—VÄRTSILÄ AREA (present author)		SUOJÄRVI—PIALOZERIAN TYPE SOVIET KARELIA (mainly according to Sokolov et al. 1970 and Sokolov et al. 1972)		
(northernmost part)	*(main area)*	Strata	Cycle	Sequence
Mica schist Formation Turbidite	conglomerate — quartzite Fm			
— — — erosion and	— — — unconformity — — — — — —			
(missing)	Dolomite—Carbonaceous slate—Volcanite Fm — alternating dolomite and carbonaceous slate member with some volcanics (volcanite member) — carbonaceous slate member — laminated dolomite member — — — — — — hiatus — — — — — —	Shungite—carbonate slate with gabbro-diabase intrusions in places Red-coloured dolomite—dolomite Dolomite—quartzitic sandstone or limestone—dolomite—clay slate ? — — — — — — — — — — — — — — —	sedimentary III	Shungite—carbonate slate
(missing)	Hematite rock—Quartzite Fm	Basic effusives Quartz sandstone–hematite, clay slate	sedimentary II	Terrigenic carbonate
(missing)	Dolomite—Volcanite Fm — local volcanite member — dolomite member	Cherty dolomite, dolomite—quartzitic sandstone—clay slate or clay slate—dolomite—limestone		
(Upper quartzite Fm, locally?)	Upper quartzite Fm — fine-grained quartzite with some clay slate and tuffaceous and tuffitic intercalations — coarse-grained quartzite with some tuff and tuffite and quartz conglomerate interbeds	Quartzitic sandstone— (clay slate)—dolomite Limestone—breccia—clay slate or limestone—dolomite—quartzitic sandstone	sedimentary I	
— erosion and —	— — — — — unconformity — — — —	— — — — — — — — — — — — — — —		
	Volcanite Fm	Basic effusives		
Lower quartzite Fm	Lower quartzite Fm — local feldspathic quartzite member — orthoquartzite member — sericite quartzite member	Quartzite—quartzitic sandstone		Quartzite—sandstone
— erosion and —	— — — — unconformity — — — —	? — — — — — — — — — — — — — — —		
(Arkosite Fm locally?)	Arkosite Fm — upper arkosite member — lower arkosite member — local arkosic conglomerate member			
— erosion and — — — — — — —	unconformity — — — —			
	Breccia-congl. Fm — graded unconf. —			
Prekarelidic	Basement Complex	Granites and gneiss granites of basement		

chtonous, i.e. its basement is a more or less fixed horizon. The correlation with the other subareas of the Karelian zone is not straightforward, and it is not easy even to correlate formations in the study area with those in the nearby Hammaslahti—Tohmajärvi antichlinorium ridge (see Väyrynen 1954; Nykänen 1968, 1971b). The conglomerate at the base of the Kalevian group in the study area is much like the Kirkkoniemi conglomerate at Tohmajärvi, which is associated with the above-mentioned ridge. Although its position is somewhat controversial, there are a number of authors who suspect that it is a Kalevian basal conglomerate (Wegmann 1929b; Väyrynen 1954; Nykänen 1968).

The Kalevian Mica schist Formation in the study area and the underlying quartzite with its Kalevian basal conglomerates correlate rather well with the conglomerates and associated graded-bedded greywacke schists (Table 19) of the Greywacke schist Formation in the Koli—Kaltimo area (Piirainen 1968; Piirainen *et al.* 1974). It is much more difficult to decide which rocks in the South Puolanka area correspond to the Kalevian (Laajoki 1973; 1976; Laajoki and Saikkonen 1977) as it is understood in the study area (Table 19). There may be some slight similarities, for example in conglomerates, between the Turbidite Conglomerate—quartzite Formation in the study area and the Pyssykulju Quartzite Formation in the South Puolanka area. A certain resemblance also exists between the Mica schist Formation in the study area and Phyllite II (but not Phyllite I) and the Staurolite—Mica schist Formation in the South Puolanka area (Laajoki 1973, p. 51; Laajoki and Saikkonen 1977, p. 127). This correlation has, however, not been well established (Table 19).

The sedimentation environment and sedimentation tectonics some areas of the Karelidic zone, e.g. the Kuopio area (Wilkman 1938; Preston 1954), the Rukatunturi area in Kuusamo (Silvennoinen 1972) and North Finland in general (Härme 1949; Nuutilainen 1968; Paakkola 1971; Perttunen 1971; Piispanen 1972), differ appreciably from the Kiihtelysvaara—Värtsilä area. Thus only rough correlations can be made with the other Karelidic areas.

The relation between the quartzite formations in the study area and the Jatulian formations in the Rukatunturi area (Silvennoinen 1972) is obscure. The Volcanite Formation in the study area may share some features with the Greenstone Formation III at Rukatunturi; likewise, the Upper quartzite Formation is similar to the Rukatunturi quartzite Formation. The underlying formations, however, excluding the Sericite quartzite Formation, bear little resemblance to each other (Table 19). The Dolomite—Volcanite Formation in the study area together with the Dolomite—Carbonaceous slate—Volcanite Formation correlate to some extent with the Dolomite Formation and the overlying Amphibole schist Formation in the Rukatunturi area (Silvennoinen 1972). The correlation is also vague with the limestone—amphibolite horizon in the Kuopio area (Wilkman 1938, Preston 1954) and the dolomites and associated mafic schists in the Kemi area (Härme 1949). As a whole, early Karelian magmatism was more intense in the Rukatunturi area (Silvennoinen 1972) and throughout North Finland (Härme 1949; Paakkola 1971; Piispanen 1972) than in the Kiihtelysvaara—Värtsilä area.

The Jatulian formations in the Kiihtelysvaara—Värtsilä area correlate rather well with some of the Jatulian formations in Soviet Karelia, perhaps best with the Suojärvi—Pialozero type (Metzger 1924; Sokolov *et al.* 1970). On account of the ambiguity of the terms in use, the correlation is presented schematically as a table (Table 19) in which the corresponding horizons are arranged side by side. In general, the quartzite, dolomite and carbonaceous slate or shungite schist deposits correspond to each other rather well. The most distinct correlation horizons are the Volcanite Formation in the study are and the Lower Jatulian mafic effusives of Suojärvi—Pialozero type, which rest on the quartzite—sandstone sequence; likewise the hematite layer in the Hematite rock—Quartzite Formation - - the hematite layers in the upper section of the second sedimentary cycle; the carbonaceous slates with associated dolomite interlayers in the Dolomite—Carbonaceous slate—Volcanite Formation - - the shungite—carbonaceous slates of the third sedimentary cycle, etc. It is noteworthy that the formations corresponding to the Kalevian formations in the study area seem to be entirely absent from the Suojärvi—Pialozero type.

The Kalevian rocks of the study area seem to continue unchanged southeastwards into Soviet Karelia (Väyrynen 1954; Nykänen 1968). Detailed comparisons with the Soanlahti (Hausen 1930; Hackman 1933) or Pieni-Jänisjärvi areas in Soviet Karelia (Kharitonov 1966) have not been undertaken. It should be mentioned, however, that the Kalevian basal conglomerate in the study area correlates well with the Partanen conglomerate in the Soanlahti area (Eskola 1921; Wegmann 1929b; Hausen 1930; Hackman 1933; Väyrynen 1933, 1938; Kharitonov 1966), although the position of the Partanen conglomerate has been questioned by some investigators (Inina 1975).

Time-stratigraphic correlation

The sedimentation in the Kiihtelysvaara—Värtsilä area fits roughly between two time limits: 2600—2800 Ma (p. 53) and 1800—1950 Ma (the main stage of the Svecokarelidic orogeny). Current information allows the events that took place during that period to be expressed somewhat more accurately.

The zircon from the mafic intrusions overlain by Jatulian formations (Isohanni 1971, p. 31; Silvennoinen 1972, p. 41; Piirainen *et al.* 1974; Perttunen 1975; Laajoki and Saikkonen 1977, p. 119 and Fig. 72) gives a U-Pb age of about 2450 Ma (Kouvo, personal communication). Kratz and his coworkers (1976) obtained practically the same U-Pb age, 2455 Ma, for the zircon from a quartz porphyry overlain by Jatulian formations in Soviet Karelia. Hence, the Prejatulian (or Sariolan) formations are younger than that.

In the study area a preliminary investigation was conducted on the occurrence of the Mustalampi type of quartz (2340 Ma, p. 54) in quartzites and arkosites. The preliminary results of fluid inclusion studies (K. Kinnunen, personal communication) show that the Mustalampi type of quartz commonly occurs not only in Jatulian quartzites but also in Prejatulian arkosites, except in some basal layers; hence almost all Prejatulian rocks of the Kiihtelysvaara Arkosite Suite are younger than 2340 Ma.

Most of the U-Pb ages obtained from the zircons of the albite diabases cutting the Jatulian quartzites are between 2160 and 2050 Ma [1]) (Sakko and Laajoki 1975); the youngest ages are around 2000 Ma (Kouvo 1976). For the zircon of a leucocratic diabase that cuts the Lower Jatulian quartzite in Soviet Karelia, Kratz and his coworkers (1976) obtained 2180 Ma, an age that is very close to those given above. In the Kiihtelysvaara—Hyypiä area the zircon from the metadiabase cutting the Lower quartzite Formation gives a U-Pb age of 2112 Ma (Sakko, personal communication), which roughly indicates the time when the lavas of the Volcanite Formation (p. 125) erupted (Table 19). Inspite of numerous efforts, attempts to date younger eruption episodes in the study area have failed.

In the Jatulian, quartzites were locally overlain by Marine Jatulian dolomite. Wampler and Kulp (1962) have analysed the dolomite of Kalkkimaa in the Kemi area (Härme 1949) for lead isotopes and uranium. After reduction for radiogenic lead, the dolomite showed a Pb-Pb of 2050 Ma. It is often overlain by a kind of Pre-flysch-phase euxinic sediment — a black slate that locally contains carbonates, clay-sand interlayers and iron precipitates of iron formation. Sakko and Laajoki (1975) have published an isochron based on the total lead in the rocks of iron formation in the Pääkkö area, South Puolanka, which shows an age of 2080 Ma.

[1]) If the new decay constants are used (Jaffey *et al.* 1971) the ages would be about 30 Ma younger.

Roughly speaking, the Marine Jatulian rocks are overlain by the miogeosynclinal greywacke flysch known as Kalevian, although there is often a slight but locally distinct unconformity between them. The dating of a formation of that kind is not easy. It is certainly younger than the above Marine Jatulian rocks (about 2050 Ma), but older than the late tectonic or syntectonic granitoids (1800—1900 Ma) that cut it; the age is probably somewhere around 2000 Ma (see Kouvo 1976). This is only true, however, for the miogeosynclinal flysch; the eugeosynclinal flysch may be considerably older owing to polarity in the geosynclinal system.

In the light of the above, it is evident that the Karelian Sequence developed during a fairly long period and that information on specific events is still rather inadequate. Thus, the dating of volcanic periods in different subareas might contribute to our knowledge of the events and make it easier to correlate the subareas.

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APPENDIX 1.

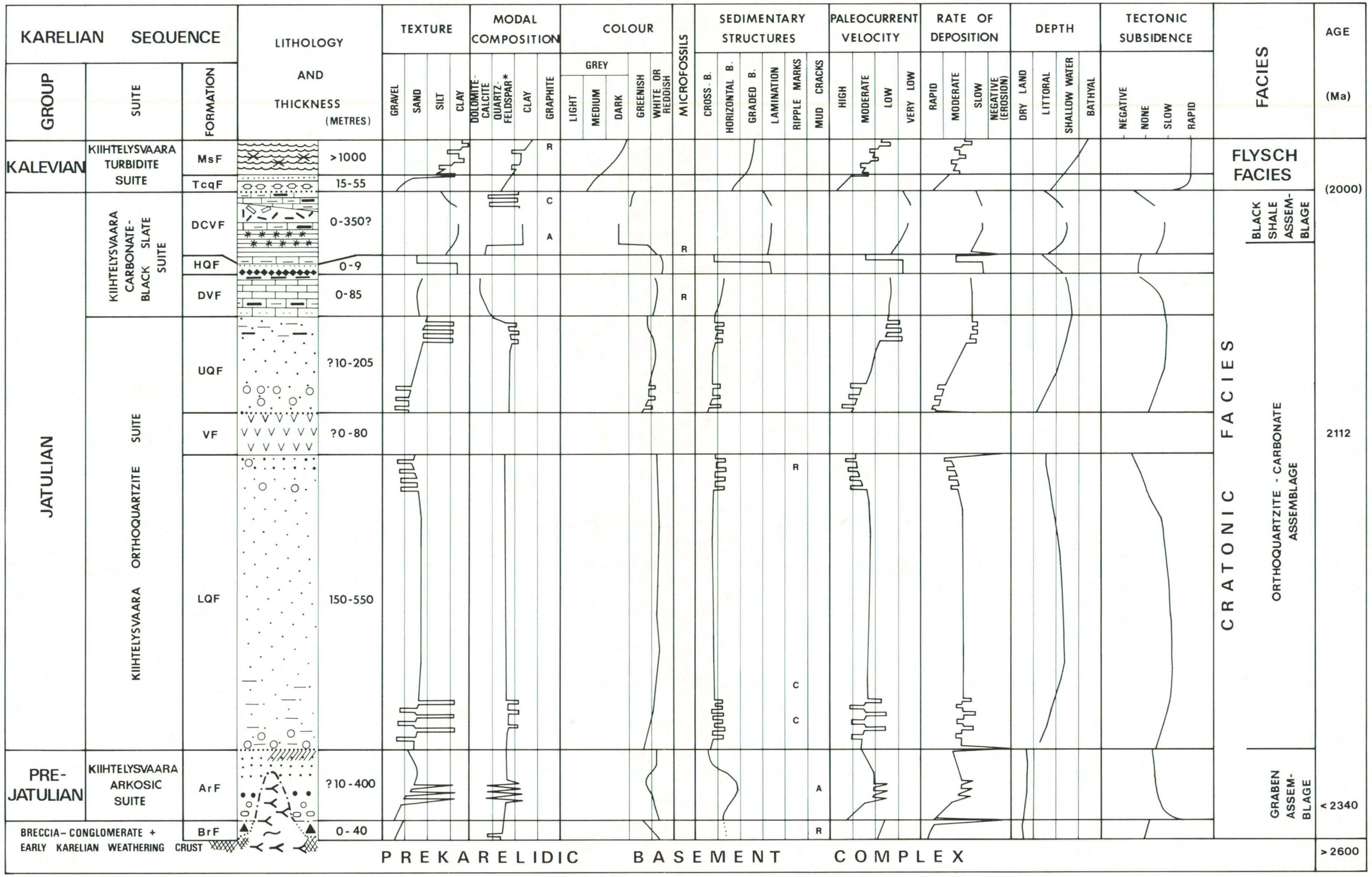
List of thin sections referred to in the study

Nmber of thin section	Rock name	Occurrence	Location: Drill hole/depth (or × =, y =)	References
17735	Albitic metadiabase	Piippolanmäki	301/ 18.25	Fig. 66 D
17838	Laminated dolomite	Viistola	310/ 87.65	Fig. 55 E
17851	Hematite rock	Viistola	310/118.70	Fig. 55 C
17854	Dolomite	Viistola	310/164.75	Fig. 55 B
17860	Quartzite	Viistola	310/116.20	Fig. 55 D
17861	Carbonaceous slate	Viistola	310/ 75.50	Fig. 55 F
17863	Quartz-bearing dolomite	Viistola	310/190.60	Fig. 55 A
17869	Carbonate-bearing upper quartzite	Viistola	309/ 53.20	Fig. 48 F
17874	Tuffaceous upper quartzite	Viistola	309/188.60	Fig. 48 E
17877	Upper quartzite	Viistola	309/186.20	Fig. 48 D
17935	Metadiabase	Piippolanmäki	304/ 44.60	Fig. 66 C
19257	Orthoquartzite	Särkilampi	× = 6923.91 y = 516.73	Fig. 48 B
19528	Metavolcanite	Kalkunmäki	× = 6922.74 y = 516.56	Fig. 66 B
19530	Metavolcanite	Kalkunmäki	× = 6922.75 y = 516.61	Fig. 66 A
19618	Sericite quartzite	Kortevaara	321/ 23.45	Fig. 48 A
Ku 00062	Basal arkosite	Karsikkojärvi	× = 6933.46 y = 513.54	Fig. 39 A
Ku 00283	Upper arkosite	Särkilampi	329/ 25.75	Fig. 39 D
Ku 00296	Fine-grained metadiabase	Kortevaara	× = 6929.40 y = 514.83	Fig. 66 E
Ku 00297	Gabbro-like metadiabase	Hyypiä	× = 6926.06 y = 515.18	Fig. 66 F
Ku 00302	Feldspathic quartzite	Kalkunmäki	× = 6922.73 y = 516.64	Fig. 48 C
Ku 00325	Sericitized upper arkosite	Särkilampi	329/ 1.55	Fig. 39 F
Ku 00635	Uraninite-bearing quartz rock	Mustalampi	× = 6923.72 y = 519.27	Fig. 26

Vertical profile and facies analysis of the Karelian formations in the Kiihtelysvaara—Värtsilä area.

Symbols of formation:

MsF = Mica schist Formation; TcqF = Turbidite conglomerate-quartzite Formation; DCVF = Dolomite — Carbonaceous slate — Volcanite Formation; HQF = Hematite rock — Quartzite Formation; DVF = Dolomite — Volcanite Formation; UQF = Upper quartzite Formation; VF = Volcanite Formation; LQF = Lower quartzite Formation; ArF = Arkosite Formation; BrF = Breccia-conglomerate Formation. Other symbols as in Fig. 24.



C= common R= rare A= abundant *) or rock fragments

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