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From continental rifting to collisional crustal shortening - Paleoproterozoic Kaleva metasediments of the Höytiäinen area in North Karelia, Finland

by Jarmo Kohonen

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FROM CONTINENTAL RIFTING TO COLLISIONAL CRUSTAL SHORTENING - PALEOPROTEROZOIC KALEVA METASEDIMENTS OF THE HÖYTIÄINEN AREA IN NORTH KARELIA, FINLAND

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JARMO KOHONEN

with 41 figures, 3 tables and one appendix

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The depositional history and provenance of Kaleva metasediments comprising the outer, deltaic to deep-marine sequence of the Paleoproterozoic Karelian continental margin have been studied using facies analysis, major element geochemistry and structural geology.

The sharp eastern contact of the turbiditic Kaleva association is interpreted as an ancient basin edge characterized by multiple normal faulting events during the extensional processes about 2.10 - 1.95 Ga. The unconformity at the contact seems to be younger in comparison to the lowermost graywackes in the central parts of the Höytiäinen basin. The conglomerates occurring along the contact are understood to represent several stratigraphic levels and are not regarded as basal (sensu stricto) to the Kaleva association.

Five different lithological assemblages (LA) have been defined in the study area. The easternmost LA 1 consists mainly of metapelites containing sharp-bounded interbeds of orthoquartzite. These are interpreted to be genetically linked to the phyllites of LA 3 in the central syncline of the study area. The assemblages seem to form the upper part of the system and represent shelf mud/tempestite and distal shelf mud deposits, respectively. The massive quartzites, metapelites and minor carbonate rocks of LA 2 may be lateral equivalents of LA1 and LA3, but the complex structural setting prevents confident correlation.

The alternating metagraywackes and metapelites of LA 4 form the major part of rocks studied. The assemblage is interpreted as representing rift basin turbidites derived from river deltas supplying detritus from the craton. The westernmost LA 5 consists of monotonous, thick-bedded sandy turbidites. A deep-marine turbidite fan system is tentatively suggested for this depositional environment.

The major element geochemistry confirms the similarity of the rocks studied and the metasediments of the Kiihtelysvaara-Tohmajärvi region further south, whereas differences with the Outokumpu metagraywackes, with exception to the LA 5, are obvious. Both petrography and geochemistry suggest that the Archean (Karelian Craton) rocks, spiced with rift related Paleoproterozoic mafics, form the probable source for most of the metasediments studied. A simple method for provenance estimation of sandstone-mudstone suites is introduced. The method, largely removing the effect of grain size variations, is based on computational mixing of co-existing sandstone and mudstone mean values by using constant Al_2O_3.

The structure of the studied part of the North Karelia - Kainuu foreland thrust belt is interpreted in terms of a tectonically shortened (inverted) extensional basin with significant reactivation of extensional faults. The observed structural sequence is interpreted by (partly) simultaneous layer parallel shortening of the cover and complex wrench movements of the Archean basement complex. The poorly understood behaviour of the basement complex in Svecofennian deformation restricts the validity of the structural models presented.

The presented rift basin model is connected to a regional context by using a 'Wilson cycle' type plate tectonic paradigm. Finally, the importance of the paleogeographic location of the craton margin for stratigraphic schemes is discussed, and caution is urged in making regional correlations between areas having very different locations with respect to major paleorifts.

Key words (GeorefThesaurus): schist belts, metasedimentary rocks, turbidite, lithofacies, chemical composition, structural geology, provenance, continental margin, rifting, plate tectonics, crustal shortering, Proterozoic, Paleoproterozoic, Karelian, Höytiäinen, North Karelia, Finland

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CONTENTS

"Time as a stratigraphic dimension has meaning only to the extent that any given moment in the Earth's history may be conceived as precisely coinciding with a corresponding world-wide lithosphere surface and all simultaneous events either occurring thereon or directly related thereto."

Harry E.Wheeler, 1964

INTRODUCTION

Background, methods and previous studies

Kaleva (or Kalevian) is a general term given to the metagraywacke-metapelite dominated association in the upper part of the Paleoproterozoic (Karelian) craton cover sequence in eastern and northern Finland. These rocks, described in the past as flysch-type or flyschoid (e.g. Simonen, 1986; Laajoki, 1986), form the major part of the North Karelia Schist Belt. The Kaleva rocks have often been collectively referred to as mica schists (e.g. Laiti, 1983; 1985; Lavikainen, 1985) or mica gneisses (Huhma, 1975), whose protoliths were evidently different kinds of sandstones.

In the Finnish province of North Karelia region the Kaleva rocks are bordered to the east and north by the platformal ('Jatuli') quartzites, while to the southwest they grade, without any clear break, into the migmatites of the Svecofennian (1.95 - 1.85 Ga) Savo foldbelt (Fig. 1). The depositional and tectonic development of this zone is important in understanding the evolution of Karelian craton margin as a whole.

The major aim of this study is to examine the primary features and lithological variations of Kaleva rocks in the Höytiäinen area (for location, see Figs. 1 and 2). Special emphasis is given to the relationship between the Kaleva and Jatuli associations. Structural data are also presented and interpreted, followed finally by a discussion of depositional and structural models.

Mapping was mainly carried out along profiles across strike. Selected key areas and the contact zone between the Kaleva mica schist and the Jatuli quartzites were investigated in detail. The subdivision of Kaleva rocks is based on facies successions rather than differences in rocktypes themselves. Consequently, small outcrops were considred to be of limited value in sedimentological work and continuous sections were sought where possible. The sections were mapped in detail, divided into lithofacies and drawn at a scale of 1:100. Comparisons with existing facies models were used in interpretation.

Petrographic work involved the examination of 177 thin sections. Etching and staining of samples and thin sections was done for K-feldspar identification. Representative samples for chemical analysis were selected covering the range of lithologies present throughout the study area. Major element analyses were done by XRF at the Geological Survey of Finland and the Rautaruukki Ltd laboratories. Structural observations were aimed at cover the whole study area evenly, but some concentration of data around the areas mapped in detail was unavoidable.

The first comprehensive review of the North Karelia Schist Belt is found in the explanation to the 1:400 000 geological map by Frosterus and Wilkman (1920) and much of the description is still of great value. The classic works of Wegmann (1928) and Väyrynen (1933, 1939) laid the foundation for both the stratigraphic and structural framework of North Karelia. More recent investigations include stratigraphic studies by

Pekkarinen (1979) and Kohonen & Marmo (1992), and structural reviews by Gaal et al. (1975), Koistinen (1981), Ward (1987) and Ward & Kohonen (1989). Mafic volcanism has also recently been studied by Pekkarinen and Lukkarinen (1991), Vuollo et al. (1992) and Vuollo & Piirainen (1992).

Apart from explanations to geological maps, published predominantly in Finnish (Huhma, 1975; Nykänen, 1968,1971), publications concerning the Kaleva mica schists are few. Ward (1987; 1988) presented litho-

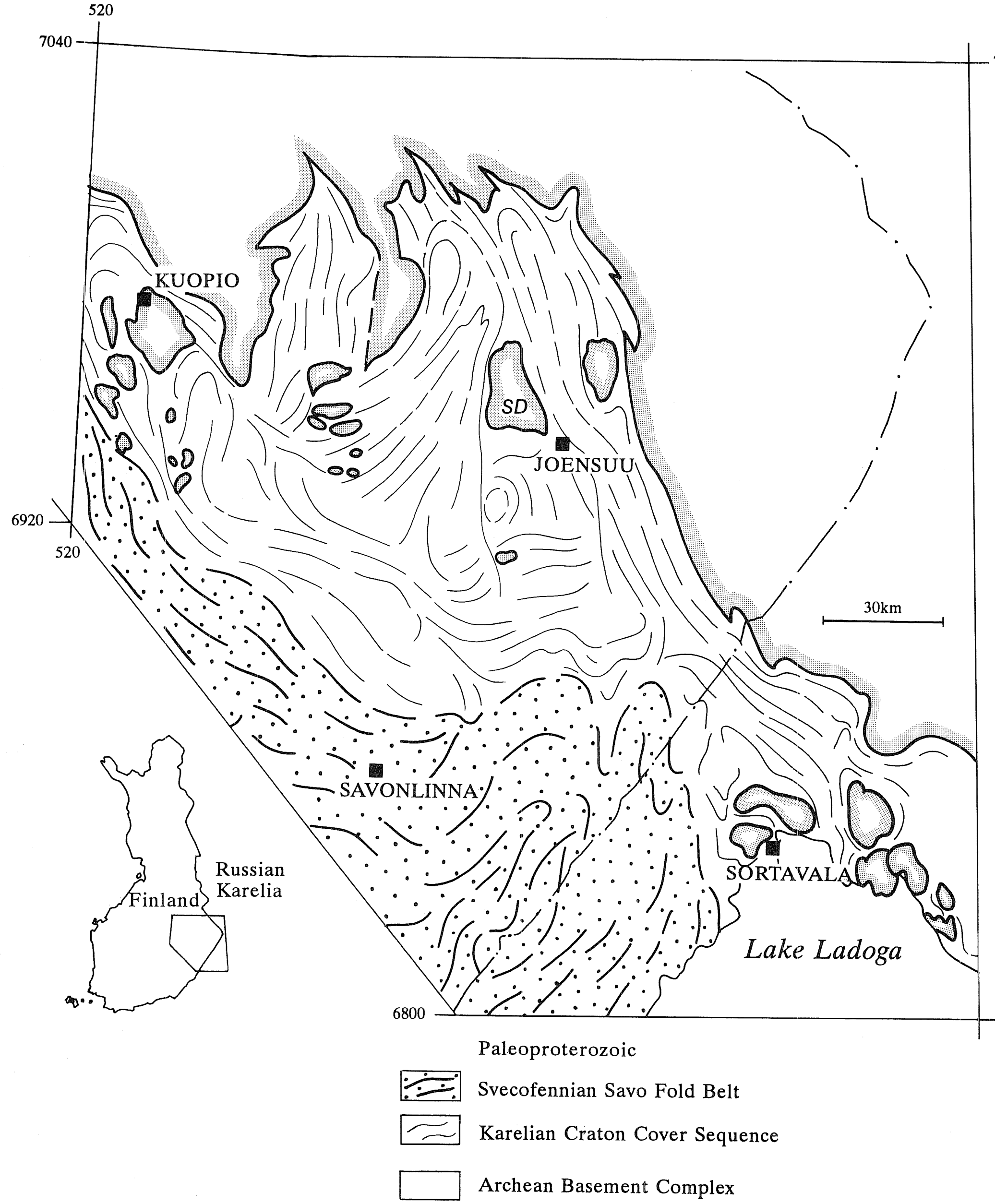


Fig. 1. Generalized map of the North Karelia Schist Belt. SD refers to Sotkuma basement inlier in Figure 2.

type division and a depositional model for the Hammaslahti-Rääkkylä district (for location, see Fig. 2). The geochemistry of the Kaleva metasediments has been investigated by Huhma & Huhma (1970) and Glumoff (1987), while Huhma (1986; 1987) and Karhu (1993) have studied Sm-Nd and carbon isotopes, respectively. Geochemical and structural characteristics of the Hammaslahti base metal (Cu-Zn-Au) deposit, hosted by Kaleva mica schists, have recently been summarized by Loukola-Ruskeeniemi et al. (1992).

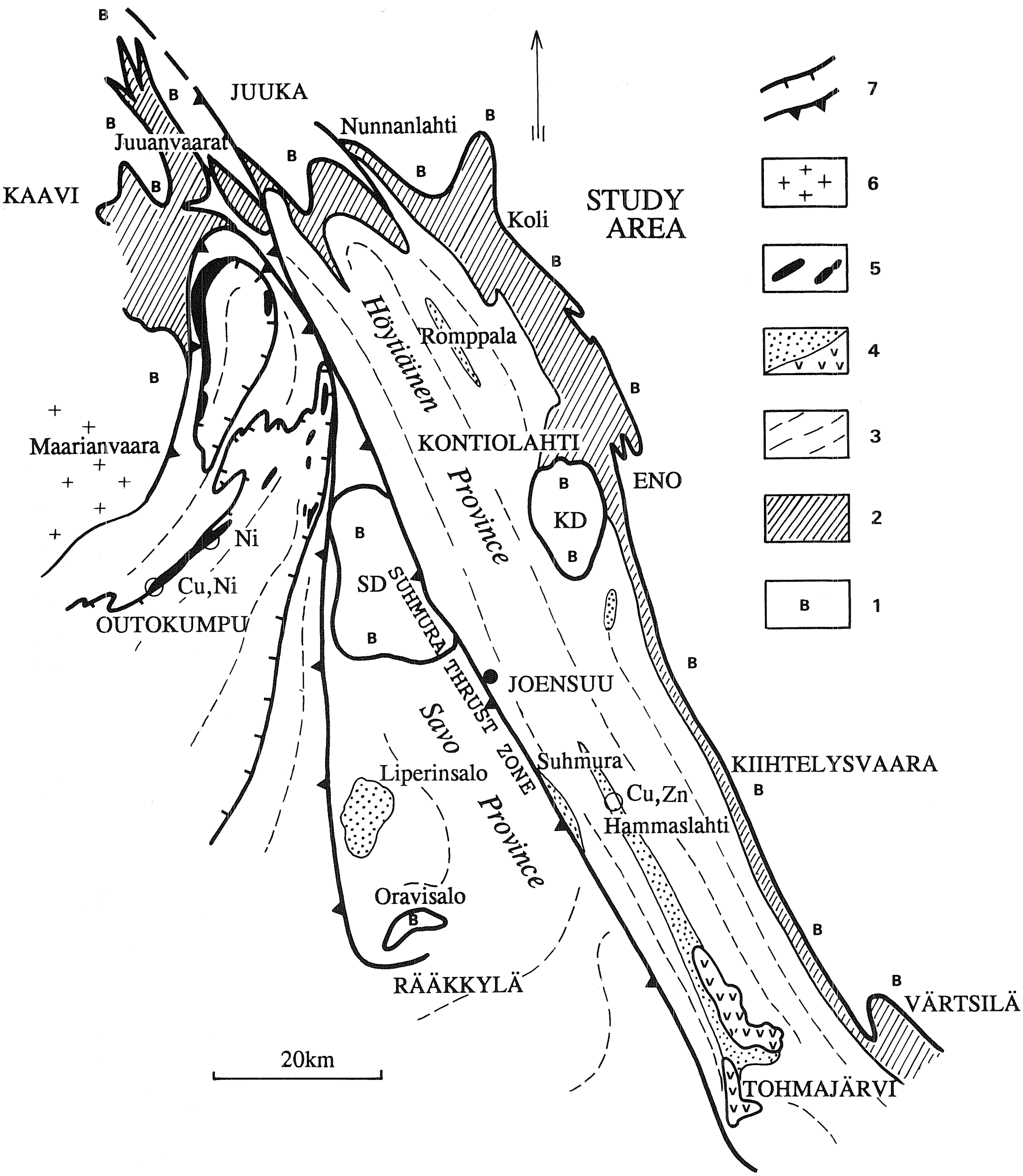


Fig. 2. Major geological units of the North Karelia Schist Belt (Karelian) and locality names referred to in text; SD and KD for Sotkuma and Kontiolahti inliers. 1= Archean Basement Complex, 2= Platform assemblage (quartzite, minor carbonate rock), 3= Basinal assemblage A (mica schist), 4= Basinal assemblage B (conglomerate/mafic volcanics), 5= Ophiolite assemblage (mainly serpentinite), 6= Maarianvaara granite, 7= Major thrusts.

Stratigraphic framework of the North Karelia Schist Belt

The Archean in eastern Finland consists mainly of granitoids, migmatite gneisses and minor greenstone-belt remnants. The Karelian formations, about 2.4 - 1.9 Ga in age, are separated from the Archean Basement Complex by a nonconformity. The conventional three-fold division of the Karelian formations into Sariola, Jatuli and Kaleva associations (e.g. Meriläinen, 1980) has generally been applied in North Karelia. The former two cratonic associations are intruded by numerous mafic dykes, indicating a minimum age of about 2.0 Ga for these units. Karhu (1993) redefined the traditional terms as parastratigraphic units and also introduced the term Ludian group (between the Jatuli and Kaleva groups) to the Finnish terminology of the Karelian formations.

The formal lithostratigraphic procedure is readily applicable to the autochthonous Sariola and Jatuli associations, and the reader is referred to reviews by Pekkarinen (1979), Pekkarinen & Lukkarinen (1991) and Kohonen & Marmo (1992) for a detailed consideration of the stratigraphic subdivision of these formations. However, an outline of the depositional sequence, together with some geochronological information, will be given here as a basis for the reasoning and discussion presented below.

The sporadically preserved basal Karelian formations (Pölkkylampi Fm.; Ilvesvaara and Urkkavaara Fms.) consist mainly of talus breccias, diamictites and polymict conglomerates that reflect a local basement provenance. An extensive regolith (Hokkalampi Paleosol) was developed within these sediments and the underlying Archean basement, prior to the deposition of quartz-pebble conglomerates and quartzites (Vesivaara and Koli Fms.). A depositional hiatus between these and the overlying alluvial Jero Formation is indicated by the irregular distribution and incomplete nature of the underlying strata, with the Jero conglomerates and arkosites lying in places directly upon the Archean basement. The Jero Formation is overlain in turn by a thick orthoquartzite (the Puso Fm. and correlatives; cf. Kohonen and Marmo, 1992, Fig. 37). All these units have been intruded by one or more suites of mafic dykes. In the Nunnanlahti-Koli-Kaltimo district, adjacent to the present study area, the platformal sequence terminates here but in the Kiihtelysvaara-Värtsilä region, some 50 to 80 km further south, the corresponding quartzite (the Haukilampi Fm.) is capped by a mafic volcanic horizon (the Koljola Fm.), which has a U-Pb zircon age of 2115 $\pm$6 Ma (Pekkarinen & Lukkarinen, 1991).

The Koljola mafic volcanics are in turn overlain unconformably by a hematite and carbonate bearing quartzite (the Kalkunmäki Fm.), which represents the lowermost unit within the Hyypiä Group (Pekkarinen & Lukkarinen, 1991). The upper part of the Hyypiä Group consists of dolomites, mafic volcanics and carbonaceous phyllites, together with minor oxide-facies iron formations and hematite bearing quartzite. Numerous mafic dykes and sills are associated with the volcanics, but no information concerning their age is available (Pekkarinen & Lukkarinen 1991).

The lithological assemblage dolomite-carbonaceous schist-mafic volcanics, represented in North Karelia by the Hyypiä Group in the Kiihtelysvaara-Värtsilä district, is traditionally referred as 'Marine Jatulian', whereas the rocks of the overlying Tohmajärvi Group are labeled to represent 'Kalevian' or 'Kaleva tectofacies' (Pekkarinen, 1979; Pekkarinen and Lukkarinen, 1991). According to Karhu (1993) the Jatulian-Ludian boundary can be located between the Annala and Petäikkö Formations within the Hyypiä

Group.

The Tohmajärvi Group is separated from the Hyypiä Group by an unconformity marked locally by a relatively thin (typically 0.2 - 4 m) conglomerate containing cobbles and boulders of quartzite, mafic igneous rocks and carbonaceous slate. The conglomerate and overlying massive, graded quartzites comprise the Kortevaara Formation. Upwards both the number and thickness of quartzite beds rapidly decreases and the dominant rock type gradually becomes mica schist of Heinävaara Formation (Pekkarinen, 1979; Pekkarinen & Lukkarinen, 1991).

Although the boundary between the Hyypiä and Tohmajärvi Groups is well defined in the type area at Kiihtelysvaara, the overall nature of the lower boundary of the Kaleva mica schists and even the term 'Kaleva' itself is unfortunately far from clear. In most stratigraphic overviews and general discussions (e.g. Simonen, 1980; Papunen and Vorma, 1985; Laajoki, 1986) Kaleva sediments are described as forming a 'flyschoid' unit younger than the platformal (Jatuli) quartzites. There are, however, many features which challenge this as being the only possible interpretation. For example, the Tohmajärvi volcanic complex (see Fig. 2), exposed as an anticlinal ridge (Nykänen, 1968), yields a U-Pb zircon age 2105 $\pm$15 Ma (Huhma, 1986) which is very similar to that obtained for the Koljola Formation (see above). Nevertheless, the coarse clastics associated with the Tohmajärvi volcanics seem to show a gradual transition into mica schists and phyllites regarded as 'Kalevian' (Nykänen, 1971b).

Ward (1987, 1988) classified the Kaleva mica schists within the North Karelia Schist Belt into two major divisions and assumed they to represent deposition in separate basins. The eastern Höytiäinen province is separated from the western, mainly allochthonous Savo province by the Suhmura thrust zone (see Fig. 2). Ward (1987) attempted to correlate volcanism at Kiihtelysvaara and Tohmajärvi by invoking intercalation and coeval deposition of platformal and deep marine clastic facies in different parts of the Höytiäinen province. Ward also argued that mafic graywackes representing distal parts of the Tohmajärvi volcanics were intercalated with the coarse clastic deposits of Tikkala. Therefore, unless these mafic sediments represent erosion of the Tohmajärvi volcanics long after volcanism had ceased, the age date for the volcanics (2105 $\pm$15 Ma) makes correlation of the Tikkala-Hammaslahti coarse clastics with the Kortevaara Formation at Kiihtelysvaara untenable.

In summary, very few solid information concerning the stratigraphy of the Höytiäinen province are available. In the Kiihtelysvaara-Hammaslahti-Tohmajärvi district Nykänen (1968, 1971a,b) tentatively placed the intraformational conglomerates and impure arkoses of the Hammaslahti - Tohmajärvi area below the mica schists and phyllites. In contrast, Ward (1988) suggested the Tikkala and Hammaslahti coarse clastics to represent prograding deposition over the pelites and correlated this phase to the deposition of the basal conglomerates of the Tohmajärvi Group in the east (see above).

Principal structural and metamorphic features

In contrast to the variable structural trends of the western Savo province the Höytiäinen province is characterized by a rather constant NW-SE strike of lithological units, stratification and pronounced foliations. The mica schists are tigthly folded and the younging directions are variable, even at outcrop scale. Within the present study area the main foliation and bedding planes dip generally 40 to 80 degrees to the WSW giving the overall appearance of eastern vergence for the major folding.

The quartzites in the east show a tectonic style distinctly different from that of the mica schists. The quartzites are imbricated rather than folded and invariably show younging away from the basement (Kohonen, 1988). In the Koli area the quartzites dip gently or moderately (20 to 60°) to the southwest, but in the Kiihtelysvaara-Värtsilä district the platformal sequence presently has a nearly upright attitude.

Taking the North Karelia Schist Belt as a whole, the metamorphic grade seems to increase towards the veined mica gneisses of the Outokumpu area in the west (Huhma, 1975), but detailed metamorphic studies have not been carried out. In the Kiihtelysvaara-Tohmajärvi district mica schist commonly contains porphyroblasts of staurolite or andalusite (parageneses Staurolite-Garnet-Biotite(-Muscovite)-Plagioclase-Quartz and Andalusite-(Garnet)-Biotite-Plagioclase-Quartz; Nykänen, 1968) and the growth of porphyroblasts seem to postdate the major foliation (Campbell et al., 1979). In the Joensuu-Hammaslahti district and in the Höytiäinen area porphyroblasts, excluding biotite and sporadic garnet, are rare, and the assemblage Biotite-Muscovite-(Chlorite)-Plagioclase-Quartz is typical for the metapelites (Nykänen, 1971a; Huhma, 1975). In spite of intensely developed mica foliations, metamorphic recrystallization has not obliterated primary grain textures, particularly in coarse clastic deposits.

Terminology

In the present study the term 'Kaleva' is used informally, without lithostratigraphic or geochronological connotations. Höytiäinen province is used to name the geographic zone corresponding to the definition of Ward (1987). The Höytiäinen area refers to the present study area in the northern part of the Höytiäinen province. It should still be noted that the location of the North Karelia Schist Belt (NKSB, for short) must not be confused with the area called as North Karelia, further to the north on the Russian side of the border.

In the absence of a well-established terminology for metamorphosed sedimentary rocks, the usage of rock names will be briefly explained. Mica schist has been used as a general term for metasedimentary rocks containing abundant mica, as well as referring to sequences consisting dominantly of this lithology. In the latter context the term quartzite refers to the platformal, originally arenitic sequence. The terms psammitic schist and pelitic schist has been used to describe the assumed original grain size of a mica schist and most typically reflects the distinction between graywackes and mudstones. In some cases additional descriptive attributes, such as 'quartzitic', have also been used. Arkosite, subarkosite and quartzite correspond to arkose, subarkose and quartz-arenite in non-metamorphic nomenclature, respectively.

RELATIONSHIPS BETWEEN KALEVA AND JATULI ASSOCIATIONS; DESCRIPTION

The main stratigraphic problem in the study area is the obscure and multifarious lower boundary of the Kaleva association. The eastern contact against the quartzites has been variously interpreted as thrusted (Frosterus and Wilkman, 1920; Gaál, 1964), gradational (Piirainen, 1968) and as an unconformity (Väyrynen, 1933; 1939; Piirainen et al., 1974). The unconformity concept - based largely on observations from the present study area - has been favoured and is, without doubt, at least locally correct (Pekkarinen 1979, Kohonen et al., 1990). Numerous quartzite-pebble conglomerates have been reported along the contact between Jatuli-type quartzites and Kaleva mica schists both within and outside the present study area (Väyrynen 1933, 1939; Piirainen et al., 1974). Abundant mafic dykes form an essential component within the platformal Jatuli association, and their apparent absence from the mica schists led Piirainen et al. (1974) to conclude that the Kaleva association is separated in the Höytiäinen area from the quartzites by a major unconformity and is thus distinctly younger in age.

Although the contact forms a sharp lithological boundary in the field, it cannot be traced using regional aerogeophysical or gravimetric data since neither the contact itself nor the lithologies on either side display diagnostic anomalies or patterns. The 2D-modelling of the deep seismic data (Tuomi, 1988) indicates that the contact zone could be steeply dipping. The VLF-R- and VAD-measurements across the zone are in accord with this interpretation: in places, at least, the lithological contact is clearly steeper than the bedding within quartzites immediately to the east of the contact (Kohonen et al. 1990).

The eastern contact of the Kaleva mica schists

To follow the procedure used by Väyrynen (1933) in his classic stratigraphic work, the conglomerates and other field observations are described commencing with the northern part of the study area (see Fig. 3 for locality names).

At Juntulankylä the cover sequence pinches out towards the north between two blocks of basement, and only a small remnant of quartzite has been preserved (Fig. 4). Immedeately to the west of the quartzite (contact not visible) a sequence of interlayered massive feldspathic, mica-poor psammitic schists is exposed. Pelitic alternations (bio-mus-quartz schist) gradually become more dominant upwards, but further west arkosic and quartzitic interbeds appear again (for detailed description see Kohonen et al. 1989). The sequence is finally truncated by a WSW-dipping mylonite zone which juxtaposes it against a diabase and basement gneisses .

Within the same area, a solitary outcrop of quarzite-pebble conglomerate (Fig. 5) is also exposed. The measured thickness is about 7 meters and some interbeds of coarse feldspar-bearing quartzite are also present. Quartzite clasts (in part cemented by carbonate) are dominant (Table 1), but in the upper part felsic plutonic fragments and feldspar pebbles are also present. Neither the upper or lower contacts are exposed.

Between Nunnanlahti and Ahmovaara the eastern contact of the mica schists is poorly exposed and strongly sheared. At Ahmovaara two different types of contact, very near to each other, have been examined. The section Ahmolampi 1 (Figs. 6 and 7) consists of fine-grained gray quartzite and semipelitic-pelitic mica schist. The overall grainsize clearly decreases upwards, but sharp bounded, 10 to 20 cm thick interbeds of carbonate cemented medium grained quartzite are sporadically present troughout the whole section. To the east of the measured section a solitary outcrop of medium-grained, light gray orthoquartzite, quite similar to that within the lowermost interbeds, was observed.

The section Ahmolampi 2 begins in the east with medium to coarse grained arkosites, which are intercalated with sheared amphibolites (originally mafic sills ?). The uppermost amphibolite layer is in contact with a strongly sheared (partly mylonitic), greenish colored biotite-albite-chlorite schist, which grades westwards into interbedded psammitic and pelitic schists. Both the amphibolite and the schist are very strongly sheared and disrupted near the contact, and a wholly tectonic relationship cannot be excluded. Nevertheless, a lensoid apophyse-like mafic inclusion was observed within the sheared schist, thus favouring an intrusive interpretation.

The mica schist sequence at Piili has an exceptional setting in that it forms a lens completely enclosed by quartzite and metadiabase and is isolated from the main area of mica schists. (Fig. 3). The overall structure has been interpreted as a tight, partly overturned syncline (Kohonen et al. 1990). The western, overturned limb of the structure is characterized by sharp-bounded alternations of quartzite and mica schist. The thickness of the quartzite layers increases towards the west, and near the sheared, but unexposed contact against the quartzite they attain thicknesses of more than 1 m and form the major part of the rock. Thus, although the contact is not seen, a depositional, interbedded relationship between the quartzite and mica schist units seems plausible.

West of the lake Pusonjärvi brecciated orthoquartzite was observed on the eastern side of the contact (Fig. 8) and in places rounded pebbles of vein quartz occur between the angular breccia-blocks. The outcrop most likely represents conglomerate deposited on top of a scree breccia but a possible origin as a tectonically brecciated quartz-pebble conglomerate cannot be excluded either.

The well preserved conglomerate near the small lake Tottolampi (Fig. 8) has been described by Väyrynen (1933; "Lääsölä conglomerate"), Piirainen et al. (1974; "Puso conglomerate"), Kohonen (1987) and Kohonen et al. (1990). The rounded pebbles consist of quartzite and chlorite schist, the latter probably representing originally mafic rocks (Fig. 9 and Table 1). The matrix is quartzitic and the conglomerate contains interbeds of purplish colored quartzite. The eastern concact of the conglomerate is, again, somehow obscure: the westernmost Jatuli-type orthoquartzites show brecciation with hematite and quartz filling the space between the blocks. In places the conglomerate seems to have been deposited directly on top of the breccia. West of the conglomerate exposures of the bedrock is deeply buried, and the first mica schist outcrops are more than 500 meters away. Southwards along strike a quartzite, similar to the interbeds in the conglomerate, was observed. The purplish quartzite is here associated with carbonate cemented quartzite and discontinuous interbeds of semipelitic schist.

The Hautajärvi region is modified by a late, dextral strike-slip fault. The contact was observed in a cliff NNW of the lake Hautajärvi (Fig. 10). The westernmost Jatuli-type rocks are epigenetically dark colored (see Kohonen et al. 1990) conglomeratic subarkosites of the Jero Formation. These are followed (after an unexposed interval of about 2 m) by gray, massive quartzites alternate upwards with, and then pass rapidly into mica schist. Some 300 m

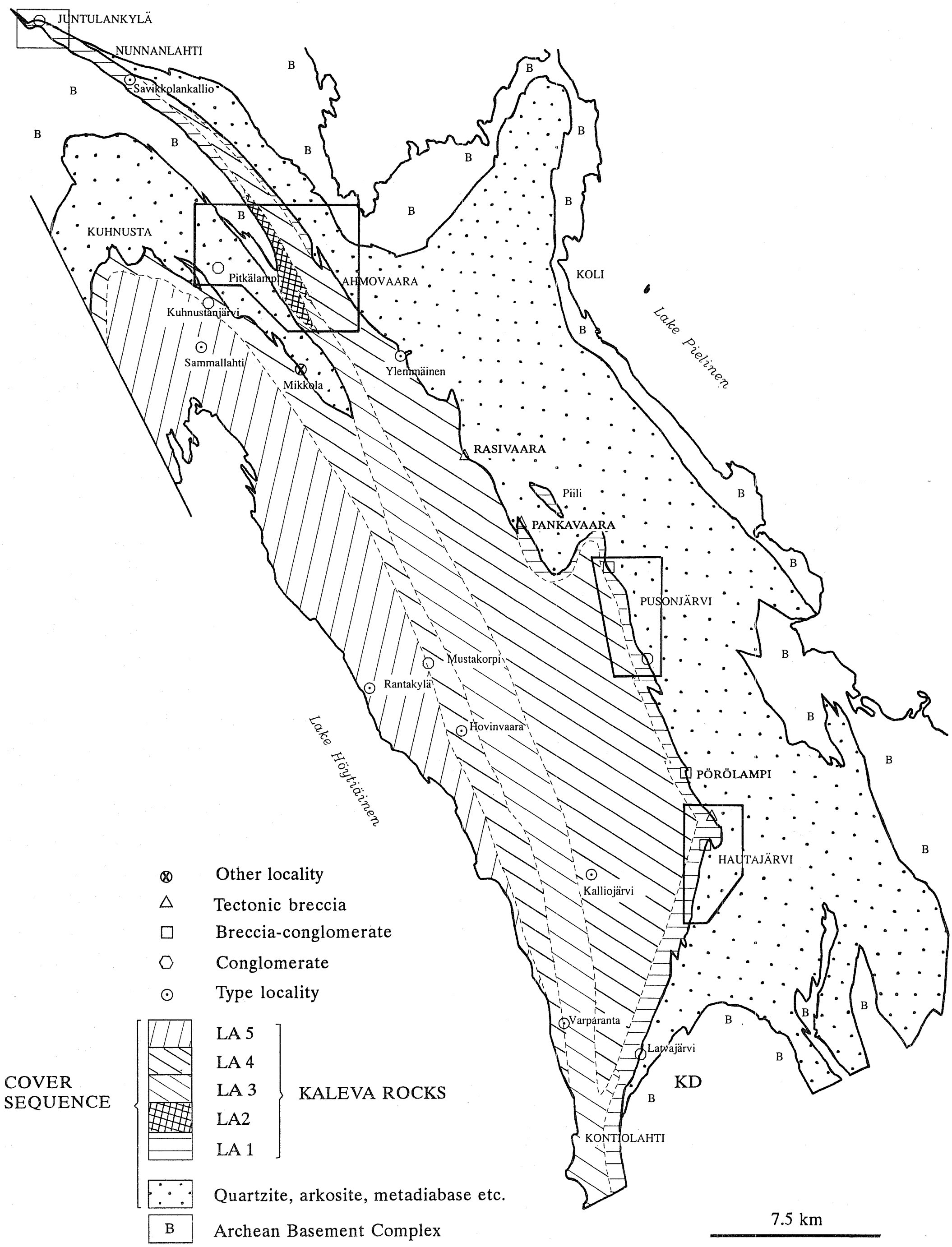


Fig. 3. Sketch map indicating location of the areas investigated in detail (boxes). Areal extent and type section localities for the lithological assemblages (LA 1-5) are also shown. For location see KD (Kontiolahti inlier), Koli and Nunnanlahti in Figure 2.

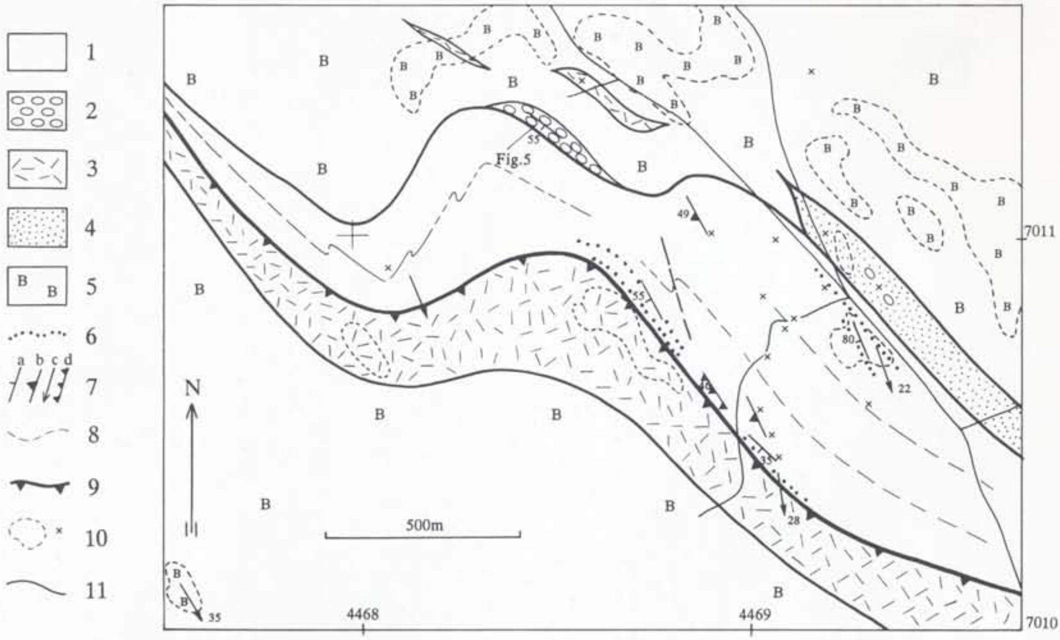


Fig. 4. Geological map of the Juntulankylä area (for location see Fig. 3). 1= Mica schist, 2= Conglomerate, 3= Metadiabase, 4= Quartzite, 5= Basement granitoids, 6= Quartzite and arkosite interbeds, 7= Structural symbols: (a) bedding, (b) main foliation, (c) lineation, (d) mylonitic foliation, 8= Trace of S_1 foliation, 9= Thrust, 10= Outcrops, 11= Roads.

Fig. 5. Quartzite pebble conglomerate at Juntulankylä (Tätilä conglomerate, for location see Fig. 4). Notebook is 17 cm long.

Table 1. Characteristics of conglomerate occurrences.

Locality	References		Clast composition (common/subordinate)	Clast size (cm) (max/average)	Matrix composition	Interpretation
TÄTILÄ	Kohonen et al., 1989	lower part	qtz, crbqtz / vq, chls	25 / 6	quartzite	fluvial (?)
		upper part	qtz, crbqtz / vq, chls, gr, fs	20 / 4	quartzite	fluvial (?)
TOTTOLAMPI	Väyrynen, 1933; Piirainen et al., 1974; Kohonen, 1987; Kohonen et al., 1990		qtz / chls, sers, vq, db	30 / 5	quartzite	fluvial
LATVAJÄRVI	Frosterus & Wilkman, 1920; Väyrynen, 1933; Gaál, 1964; Kohonen et al., 1990	lower part	qtz	60 / 20	quartzite	talus (?)
		upper part	qtz, qfss / bios, chls, db?	30 / 4	arkosite	fluvial (?)
MUSTAKORPI	Piirainen et al., 1974		gr, qtz / crbr, bs	20 / 3	calcareous arkosic wacke	turbidite channel
KUHNUSTANJÄRVI	Karhu, 1993		crbr, qtz, crbqtz	30 / 5	impure carbonate rock	turbidite channel
PITKÄLAMPI	Väyrynen, 1933; 1939		qtz, gr	25 / 5	chlorite-biotite schist	debris flow (?)

qtz = quartzite, crbqtz = carbonate-cemented quartzite, gr = felsic plutonic rock, fs = feldspar, db = diabase, crbr = carbonate rock, vq = vein-quartz, chls = chlorite schist, sers = sericite schist, bios = biotite schist, qfss = quartz-feldspar schist, bs = carbonaceous schist

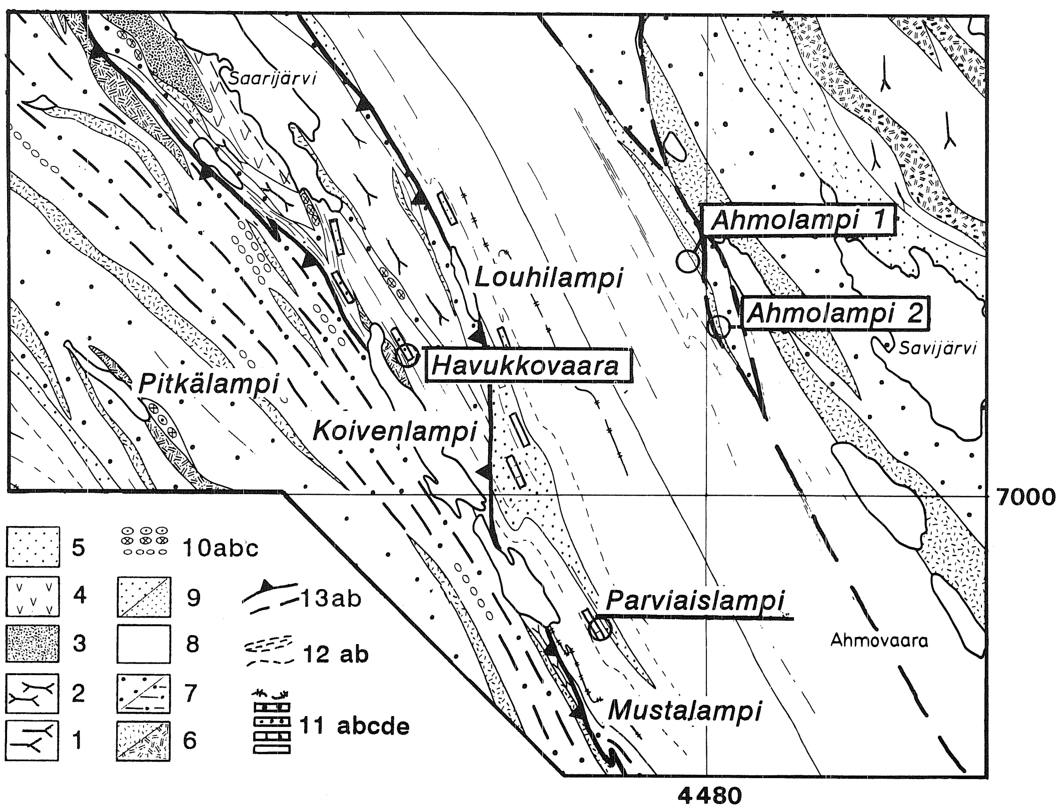


Fig. 6. Geological map of Ahmovaara area (modified from Kohonen et al., 1989; for location see Fig. 3). 1= Basement granitoids, 2= Basement (?) pegmatite, 3= Basement serpentinite, 4= Basement amphibolite, 5= Quartzite, 6= Metadiabase (tholeiite/karjalite), 7= Arkosite/Protomylonitic quartz-sericite schist, 8= Mica schist, 9= Quartzite/Quartzitic schist, 10= Conglomerate; (a) quartz clasts, (b) granitic clasts, (c) quartzite clasts, 11= Interbeds: (a) pelitic schist, (b) carbonate rock, (c) quartzite, (d) arkosite, (e) graphitic schist, 12 (a) trace of bedding, (b) phyllonitic (sheared) schist, 13 (a) shear zone, (b) thrust.

SW of here (see Fig. 10) a quite different kind of contact is exposed. The subarkosic quartzite in the east is separated from mica schist containing thin quartzite interbeds and laminae by a 1 to 2 meters thick quartz vein. At the southern end of this group of outcrops a peculiar breccia-conglomerate (Fig. 9) is present: angular blocks of quartzite are 'floating' in a dark, banded phyllite-like matrix (see also Väyrynen, 1933). The contacts of the breccia are not exposed, and both sedimentary (collapse breccia?) and tectonic modes of origin could be invoked. Nevertheless, a fairly similar rock-type is known from near the lake Pörölampi (3 kms north of Hautajärvi; see Figs. 3 and 7), where the breccia-conglomerate shows a gradual transition into a tectonically disrupted quartzite (see Frosterus and Wilkman, 1920; Kohonen, 1987). Even more convincing tectonic quartzite breccias and disrupted quartzites have been observed at several places along the contact zone; the locations of the Rasivaara, Pankavaara and Hautajärvi N breccias are indicated in Figure 3.

The impressive conglomerate along the western shore of the lake Latvajärvi in southern part of the study area (Figs. 3, 7 and 9) has been discussed by Frosterus and Wilkman (1920), Väyrynen (1933), Gaál (1964) and Kohonen et al. (1990). The lowermost part of the section consists of interbedded mica schist and massive, gray quartzite. These rock types are overlain by a conglomerate which in its lower parts contains large blocks of quartzite (some of them cross-bedded) in a psammitic matrix. Upwards the conglomerate becomes clearly polymict and contains, in addition to quartzite, fragments of quartzofeldspathic schist, grani-

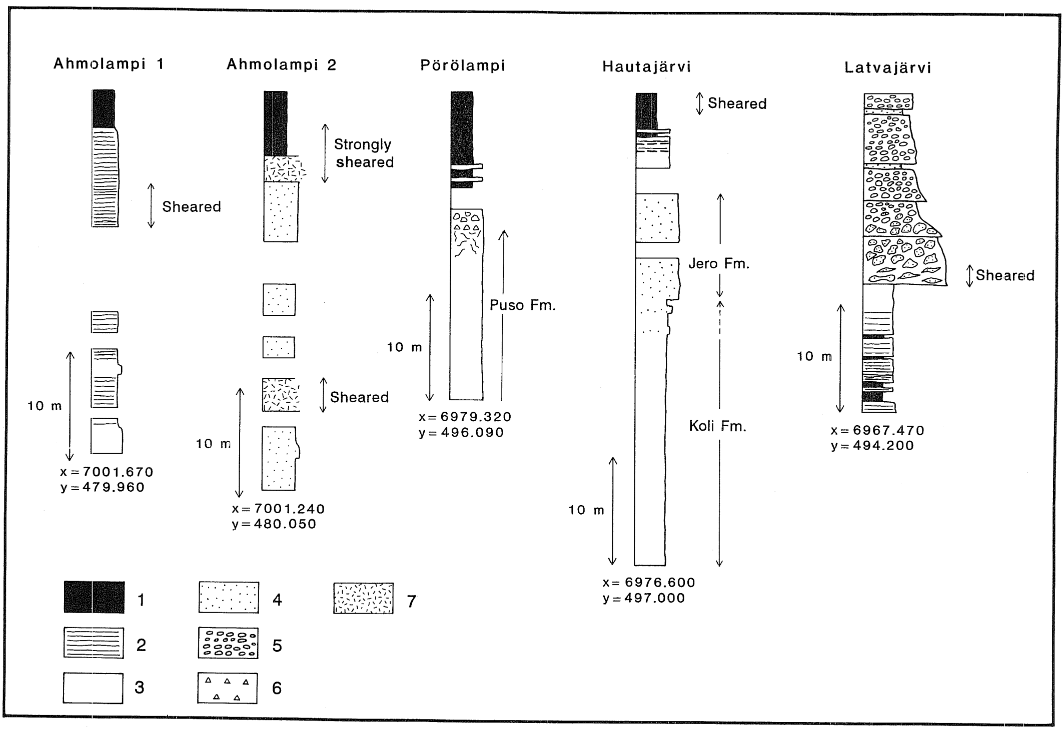


Fig. 7. Ahmolampi, Pörölampi, Hautajärvi and Latvajärvi contact sections (for location see Figs. 6, 3, 10 and 3, respectively). 1= Pelitic schist, 2= Gray quartzite with pelitic laminas, 3= Quartzite, 4= Arkosite, 5= Conglomerate, 6= Breccia, 7= Amphibolite.

toids, mica schist and mafic rocks (Table 1). The conglomerate is succeeded to the west by cross-stratified arkosic rocks and mica schists.

To conclude, several true (sedimentary) conglomerates, representing quite different types, occur in the contact zone between Jatuli-type quartzites and Kaleva mica schists. However, in every case, either or both of the upper and lower contacts are obscured, so that none of the conglomerate occurences can be shown to unequivocally represent a primary **stratigraphic** relationship between the Jatuli and Kaleva associations. On the other hand, the easternmost mica schists contain interbeds of quartzite and a conformable, interbedded relationship with quartzite seems plausible in some places (e.g. within the Piili district). The breccia-conglomerates may be either sedimentary or tectonic in origin, but quartzite breccias of demonstrably tectonic origin are also present. It must still be noted that only the localities of special interest have been included to this description; in most places the westernmost quartzite, or metadiabase outcrops are separated from exposures of folded mica schist by intervening marshy ground several hundreds of meters wide.

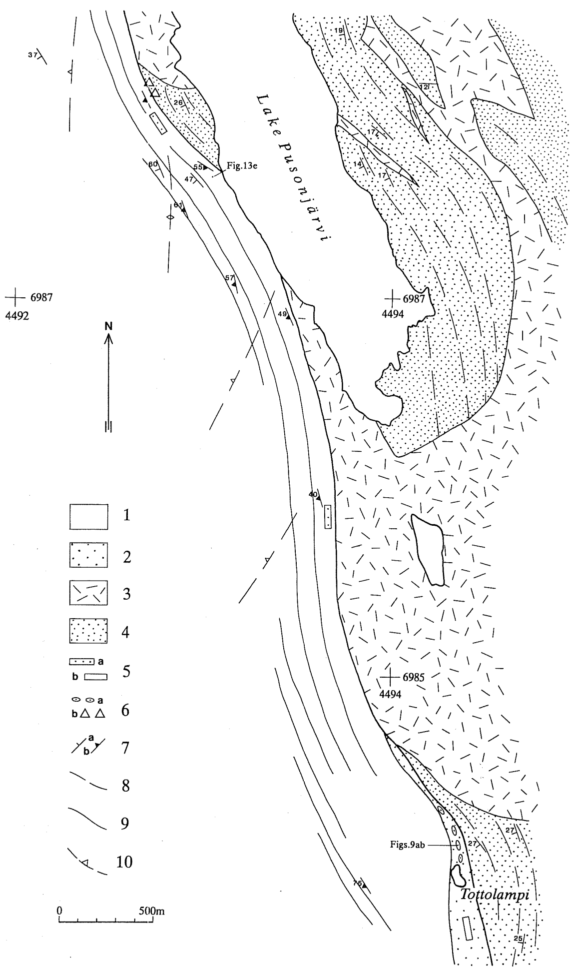


Fig. 8. Geological map of the Pusonjärvi area (modified from Kohonen et al., 1990; for location see Fig. 3). 1= Mica schist, 2= Younger quartzite, 3= Metadiadase, 4= Quartzite (Puso Fm.), 5= Interbeds (a) quartzite, (b) mica schist, 6= (a) conglomerate, (b) breccia-conglomerate, 7= Structural symbols: (a) bedding, (b) main foliation, 8= Trace of bedding in quartzites, 9= Trace of S_1 foliation, 10= Trace of S_2 foliation.

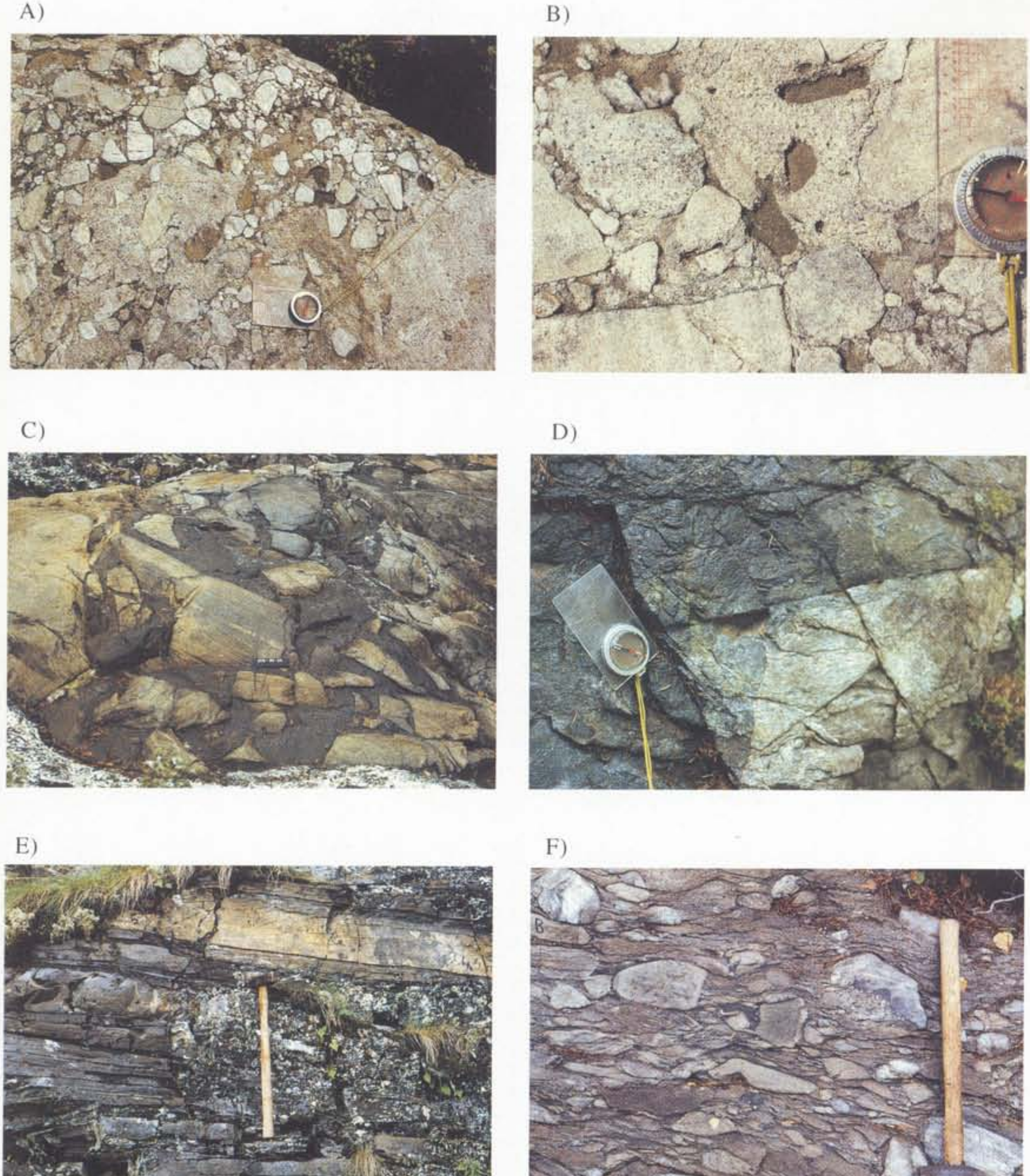


Fig. 9. (A) Tottolampi conglomerate; outcrop surface parallel to bedding and a quartzite interbed is visible at lower right. Compass is 12 cm long. (B) Detail from A, note the deeply weathered chlorite schist pebbles, (C) Hautajärvensalo breccia-conglomerate. Tag is 16 cm long. (D) Epigenetically dark colored tectonic breccia in orthoquartzite, Pankavaara. (E) Quartzite interbeds in mica schist below the Latvajärvi conglomerate. Hammer is 55 cm long. (F) Polymictic upper part of the Latvajärvi conglomerate. Photos (C), (E) and (F) by J. Marmo. For locations see Fig. 8 (A and B), Fig. 10 (C), Fig. 3 (D, E and F).

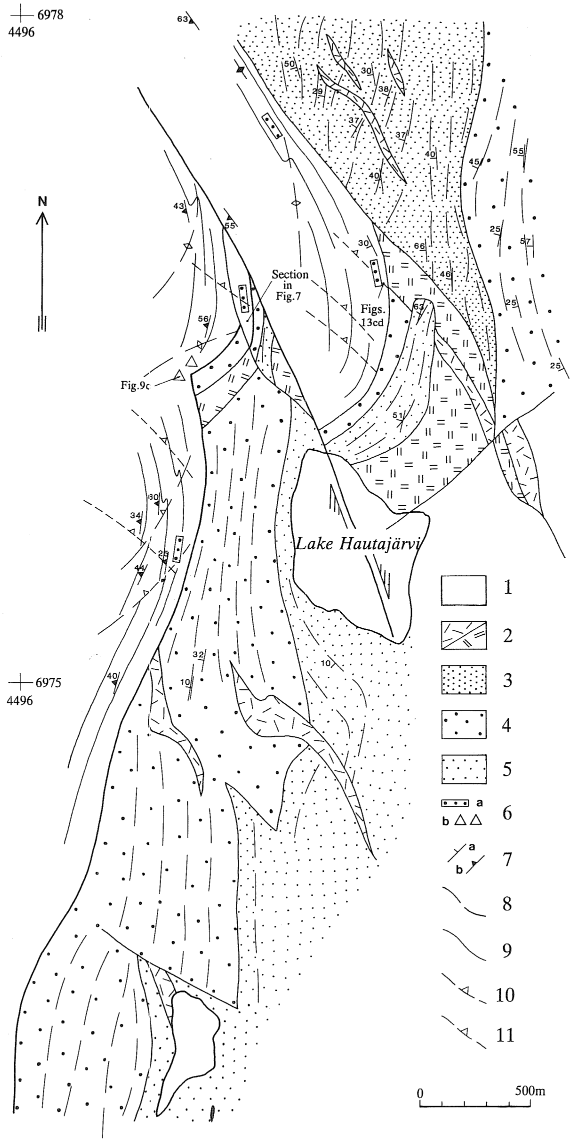


Fig. 10. Geological map of the Hautajärvi area (modified from Kohonen et al., 1990; for location see Fig. 3). 1= Mica schist, 2= Metadiadase (tholeiite/karjalite), 3= Quartzite (Puso Fm.), 4= Arkosite (Jero Fm.), 5= Quartzite (Koli Fm.), 6 (a) quartzite interbeds (b) breccia-conglomerate, 7= Structural symbols: (a) bedding, (b) main foliation, 8= Trace of bedding in quartzites, 9= Trace of S_1 foliation, 10= Trace of S_2 foliation, 11= Trace of Post-D_2 foliation.

Nunnanlahti - Kuhnusta area

In the Kuhnusta district (see Fig. 3 for location) the contact zone is poorly exposed. The Pitkälampi conglomerate (Fig. 6), assumed as basal Kalevian by Väyrynen (1933, 1939), is hard to interpret unequivocally. The conglomerate contains rounded fragments of quartzite and basement-derived granite-gneiss in a dark, chlorite-biotite rich matrix. Interbeds of very coarse-grained graywacke-like psammitic schist are also present. The western margin of the outcrop consists of a strongly sheared mafic rock (metadiabase ?) somewhat resembling the matrix of the conglomerate. The eastern contact is poorly exposed, but gives the impression of being discordant against the arkosite. The conglomerate is, however, not situated at the contact between the arkosite and mica schist but is surrounded by arkosite on both sides.

The western boundary of the Kaleva mica schists in the Nunnanlahti-Ahmovaara district is defined by a fault zone along which Archean gneisses and schists have been thrusted. Thrusting resulted in tectonic imbrication and the stratigraphy of the Koivenlampi-Saarijärvi-Louhilampi area (Fig. 6) is complicated. The basement-cover relationships require, however, some clarification. The basement inlier is bounded on both sides by basement derived conglomerates and associated arkosic rocks. The mica schists between Saarijärvi and Koivenlampi contain abundant arkosic intercalations (Fig. 11), but a direct stratigraphic relationship with the basal arkosites has not been proven. However, at Mustalampi, some 3 km SE from Koivenlampi (Fig. 6), mica schists are interbedded with arkosite to the west.

The mafic intrusions of the adjacent Koli area have recently (Vuollo et al., 1992) been divided into three groups on the basis of age and geochemistry: karjalites (gabbro-wehrlites of Hanski, 1984), Fe-tholeiites, and tholeiites with approximate ages around 2.2 Ga, 2.1 Ga and 1.97 Ga, respectively. As already concluded by Väyrynen (1933; 1939) and Piirainen et al. (1974) these appear not to intrude the Kaleva mica schists. However, at Havukkovaara a mafic intrusion truncates a sequence of interlayered pelitic and psammitic schists (see Fig. 6). The mica schists of the Koivenlampi - Saarijärvi district are lithologically quite typical Kaleva rocks, resembling for example those of the Juntulankylä district (see Kohonen et al., 1989 for detailed descriptions), but their close association with the basement derived conglomerates and the thrusted structure complicates the stratigraphic interpretation. This observation requires, however, that at least some of the Kaleva mica schists are indisputably older than the youngest set of mafic dykes.

About 5 km SSE from the Havukkovaara,

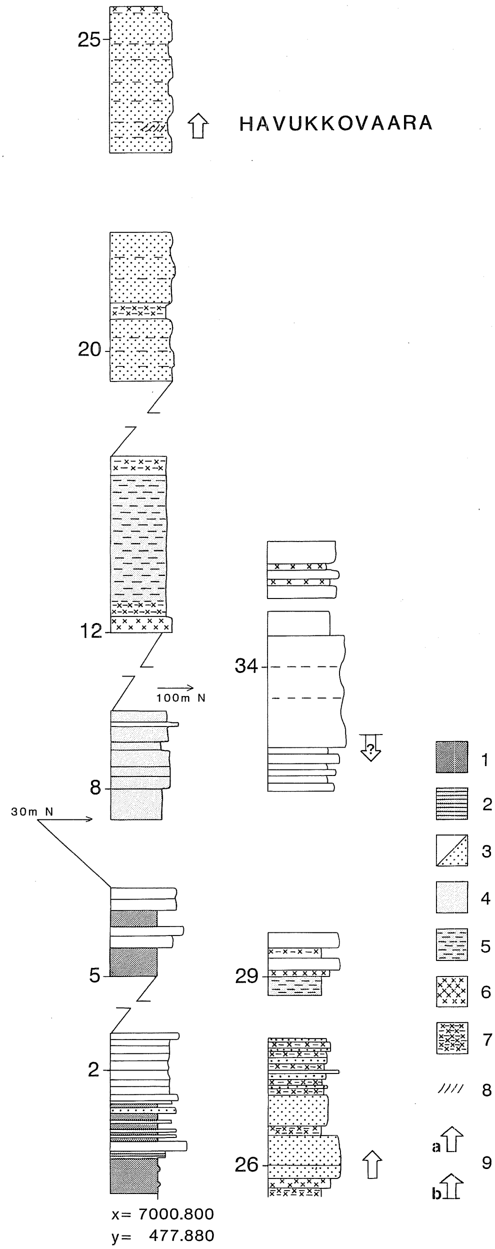


Fig. 11. Lithological section across the eastern flank of Havukkovaara hill (for location see Fig. 6). 1= Pelitic schist, 2= Semipelitic schist, 3= Massive arkosite/metagraywacke, 4= Gray, fine-grained quartzite, 5= Gray, fine-grained quartzite with abundant semipelitic laminas, 6= Calcareous massive arkosite, 7= Calcareous thinly bedded arkosite, 8= Cross-bedding, 9= Depositional top determined from (a) sedimentary structures, (b) S_0/S_1 relationships.

near the Mikkola farmhouse (see Fig. 3 for location), the contact between a karjalite intrusion and mica schist is exposed. The former occurs between a strongly foliated, protomylonitic arkosite in the east and the mica schist in the west. The western margin of the intrusion, against the mica schist, is sheared and contains large amphibole crystals and abundant garnet and the contact is conformable with bedding. The mica schist unit begins with a thin (about 10 cm) layer of dark, mica-poor schist containing scattered quartz pebbles up to 25 mm in size. This rock type is followed by a dark gray pelitic schist. Further away from the contact the mica schist contains interbeds of calcareous arkosite. The rock types in the contact could indicate hybridization caused by the intrusion, but a metamorphic modification seems possible as well (e.g. Tuisku, 1992). Because the contact is sheared and parallel to both the bedding and to the strong foliation, a tectonic, rather than primary relationship is tentatively suggested.

KALEVA LITHOTYPES AND LITHOLOGICAL ASSEMBLAGES; DESCRIPTION AND INTERPRETATION

The subdivision of lithological units is dependant on the scale and objectives of the study, and the usage of term 'facies' varies between different authors (cf. Walker, 1992). 'Lithofacies' (or 'facies') means here simply a specific lithological unit distinguished on the basis of composition and sedimentary structures. Informal and non-descriptive designations have been used for each section individually. The term 'lithotype' is used to refer to rock types occurring within a type area or section and forming a primary, depositional succession of different facies. Further, 'lithological assemblage' (later on LA, for short) is used as a broad, informal term consisting of units that are lithologically and spatially associated. Some of the assemblages are easy to define, but typically the boundaries between the LA's are gradational and indicated by differencies in the abundance or mode of occurence of a particular rock type. The characteristic features of lithological assemblages are presented in Table 2.

Depositional younging indications are not common and, consequently, the mutual lithostratigraphic relations between the assemblages are hard to prove. The study area could not be systematically remapped during the course of this study, so that descriptions and maps by Frosterus and Wilkman (1920, 1924), Pelkonen (1966), Huhma (1971, 1975) have been used in compiling Figure 3.

The interpretation of the lithotypes (facies associations) was made by comparing the observed sequences with current facies models. It is emphasized, however, that quite similar facies successions occur in different depositional environments, and that interpretation of relatively short sections of folded and metamorphosed strata cannot be final and conclusive in nature. In spite of the uncertainties, the description of grain size variations and primary structures provide valuable information concerning depositional conditions, and the models presented can be updated later by re-evaluation of the sections.

Table 2. Characteristics features of lithological assemblages.

Lithological assemblage	Type localities	Rock types (unmetamorphic names)	Psammite/pelite boundaries	Proportion of quartz in psammites	Thickness of psammite units	Sand/shale
LA 1	Savikkolankallio, Hautajärvi, Piili, Pusonjärvi W	Mudstone, quartz arenite, quartzose siltstone	gradational or sharp	high	variable	~ 0.5 - 1
LA 2	Louhilampi, Parviaislampi	Mudstone (± graphite), quartz arenite; minor dolomite	interbedded, gradational	high	variable	~ 1
LA3	Varparanta	Mudstone, quartzose siltstone	----	----	----	< 0.1
LA 4	Ylemmäinen, Kalliojärvi, Hovinvaara	Greywacke, mudstone	sharp	moderate to high	moderate	~ 1 - 5
LA 5	Rantakylä, Sammallahti	Greywacke, graphite-bearing mudstone	sharp	moderate	high	> 5

Lithological assemblage 1

LA 1 is exposed in a narrow zone parallel to the contact with the quartzites in the eastern part of the study area. Gray pelitic schists, locally phyllitic in appearance, and quartzite interbeds are characteristic. Thin, light and darker gray colored bands representing primary depositional laminations (or composite bedding and differentiated foliations in the more deformed types) are typical of these phyllites and schists.

The type section of the **Savikkolankallio lithotype** in the north is located along the shore of the bay Nunnanlahti (Fig. 3) and five lithofacies have been distinquished (Fig. 12). Lithofacies A is a thinly banded pelitic schist (Fig. 13). Quartz, biotite and sericite are major minerals together with some albitic plagioclase and chlorite. The quartz-rich and mica-rich bands may partly represent primary bedding, but the metamorphic segregation has evidently taken place as well (see Kohonen et al., 1989 for detailed description).

The facies B1, B2 and D are quartzitic schists and fine grained quartzites (Fig. 13). The rocks are light gray when weathered, but fresh surfaces are dark in color. They contain, in addition to quartz, some albite, chlorite and biotite. Thin micaceous laminae are typical of the B2 facies. Incontinuous layers and patches rich in carbonate, amphibole and garnet are common, especially within facies B1.

Lithofacies C consists of a massive, gray quartzite with grain size corresponding to fine to medium sand. In some places indistinct grading was observed at the sharp base of a bed; facies C typically passes gradually upwards into facies D. Lithofacies E occurs as sharp bounded layers of medium to coarse-grained quartzite cemented by carbonate and quartz, intercalated within facies A.

Because lithofacies A, B1, B2 and D have indistinct, gradational contacts with each other, they are interpreted to reflect continuous 'normal' fall-out deposition of silt and mud from suspension in low-energy conditions, and a distal shelf environment is tentatively suggested. The 'event beds' of the facies C and E could reflect episodal deposition of occasional storm generated currents bringing sand from adjacent siliciclastic and carbonate environments located above the storm wave base (e.g. Snedden and Nummedal, 1991).

Much of the northern LA 1 area indicated in Figure 3 consists of pelitic schist similar to lithofacies A at Savikkolankallio. The Savikkolankallio-type rocks seem to pinch out towards the south, but the northern Ahmolampi section (Fig. 7) is again very similar to this lithotype.

The **Hautajärvi lithotype** (location of the type area is indicated in Fig. 3) has been divided into three lithofacies. Lithofacies A is a pelitic, phyllite-like schist consisting of quartz (35-45%), biotite and sericite. Primary lamination is still distinct in the well preserved parts (Fig. 13). Other common sedimentary structures are varve-like grading of silty beds and small scale loading of thin beds of facies B.

Lithofacies B occurs as beds of massive white quartzite which typically show very sharp bases and tops (Fig. 13). The thickness of these interbeds ranges from millimeters to some tens of centimeters and the clastic material is almost exclusively quartz; minor amounts of feldspar and mica form the additional components. The grain size corresponds to that of medium or coarse sand. Typically the thickest quartzite beds occur near the contact zone against the quartzites, and bed thicknesses decrease very rapidly towards the west. Facies C is also interbedded with facies A, but this brownish colored quartzite contains up to 60% carbonate.

Lithofacies A is interpreted as representing mainly deep shelf muds. The graded silts deposited either by waning turbidity currents or as fallout from storm-generated suspension. The discrete beds of massive sand (facies B and C) are interpreted in essentially the same way as facies C and E of the Savikkolankallio lithotype. The absence of grading and sharp boundaries could indicate an origin as grain flow (cohesionless debris flow) deposits. However, the good sorting of the facies B sands is probably inherited from the source area of the (storm-generated ?) shallow marine debris flows and is thus not a result of the final depositional process itself.

Rocktypes quite similar to the Hautajärvi lithotype are also present to the west of the lakes Pörölampi and Pusonjärvi (Fig. 13). The schists of the Piili area, discussed in the previous chapter, represent this lithotype as well.

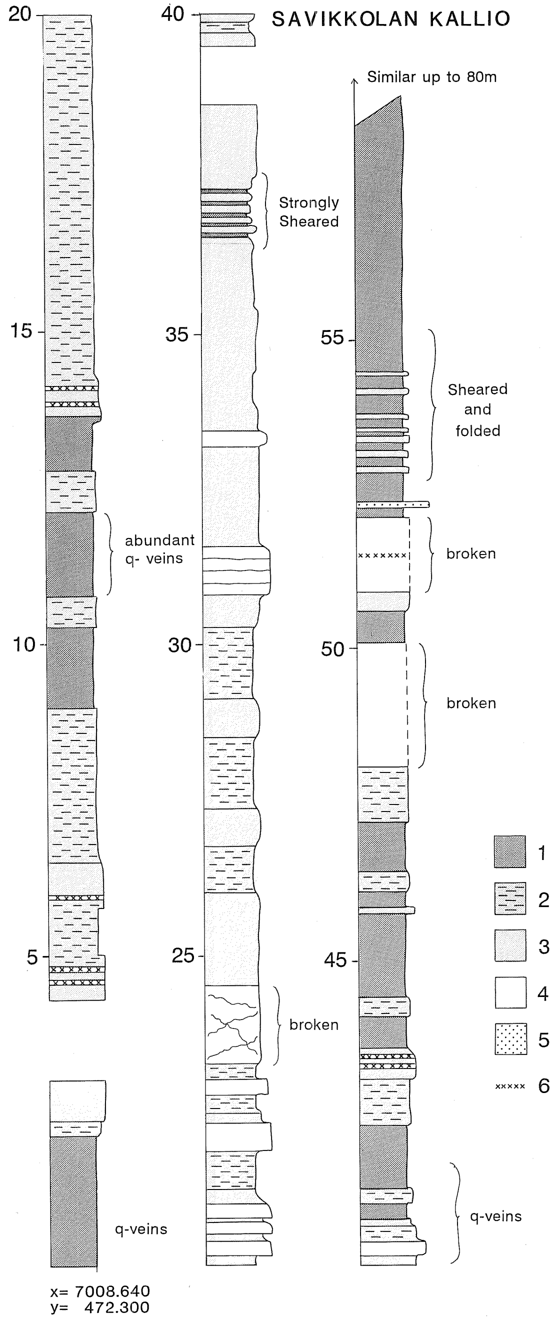


Fig. 12. Lithological section across the eastern flank of Savikkolankallio hill (for location see Fig. 3). 1= Pelitic schist (facies A), 2= Quartzitic schist (facies B2), 3= Gray, fine-grained quartzite (facies B1 and D), 4= Massive medium-grained quartzite (facies C), 5= Calcareous massive quartzite (facies E), 6= Calc-silicate bearing interbed.

A)

B)

C)

D)

E)

Fig. 13. (A) Pelitic mica schist belonging to Savikkolankallio lithofacies A. Note also abundance of tightly folded quartz-veins. The pen-marked corners form an area 30 x 40 cm in size. (B) Fine grained quartzite of Savikkolankallio lithofacies B1. Compass is 12 cm long. (C) Alternation of Hautajärvi lithofacies A (pelitic schist) and B (massive quartzite). (D) Detail from C; note the well preserved lamination in the fine-grained portions. Pencil for scale. (E) Dewatering (dish ?) structures in a quartz-rich psammitic schist, Pusonjärvi. Tag is 16 cm long. For locations see Fig. 3 (A and B), Fig. 10 (C and D), Fig. 8 (E).

Lithological assemblage 2

The quartzites, dark gray pelitic schists and minor carbonate rocks comprising the LA 2 are restricted to a small, lensoid (fault bounded ?) area within the Ahmovaara district (see Fig. 3). The **Louhilampi lithotype** (type area SSE from the lake Louhilampi; see Fig. 6) consists of two rocktypes. The quartzite units (lithofacies A) are at least several meters (probably tens of meters) in thickness. Internally the quartzite is seemingly massive and the original grain size is around 0.5 mm. Clasts consist predominantly of quartz with some plagioclase. The presence of layers containing some carbonate, thin micaceous laminations and (intraformational ?) fragments of pelitic schist (Fig. 14) are characteristic of this facies.

Lithofacies B consists of pelitic (quartz-biotite-sericite) schist containing some graphite and pyrite. Porphyroblastic biotite is locally characteristic. The contact between the two facies seems to be gradational in that the mica schists adjacent to facies A contain abundant interbeds of quartzite.

Further south, to the west of the small lake Parviaislampi (see Fig. 6), the contact between quartzite and mica schist is exposed in the bottom of a ditch. Unfortunately no evidence of depositional younging was found, but the **Parviaislampi lithotype** section begins in the east with quartzites similar to those of the Louhilampi facies A. These are followed by a dark, pelitic schist with thin interbeds and laminae of carbonate rock. Towards the west these interbeds disappear and strained crystals of pyrite characterize the schist. In addition, several loose boulders of carbonate cemented quartzite were found along the ditch. Between Louhilampi and Parviaislampi rocktypes are similar, but carbonate rock units almost a meter in thickness are locally present.

In the absence of any diagnostic primary structures the depositional nature of the Louhilampi and Parviaislampi lithotypes remains unclear. A lagoonal or shelf environment could, however, be tentatively suggested for the assemblage originally consisting of quartz arenite, mudstone and interbedded carbonate rock. On the other hand the massive quartzite seems to indicate gravity flow deposition.

Fig. 14. Fragments of dark pelitic schist in massive quartzite of the Louhilampi lithotype. Pen for scale. For location of Louhilampi see Figure 6.

Lithological assemblage 3

LA 3 forms a zone about 2 kilometers wide from the Varparanta district in the south to the Tuopanjärvi near Ahmovaara (see Fig. 3). The rocks are dominantly phyllite-like pelitic schists. Sericite, quartz and biotite in variable proportions are the main minerals. The color of the schist varies with the amount of graphite from medium gray to almost black. The area is poorly exposed, but the Varparanta district near the shore of the lake Höytiäinen may be considered as the type area. The **Varparanta lithotype** is monotonous and consists of dark phyllite containing thin laminae rich in fine grained quartz.

The Varparanta lithotype is interpreted as representing originally muds and quartzose silts deposited on the outermost part of a siliciclastic continental shelf.

Lithological assemblage 4

The major part of the mica schists have been included within LA 4. The assemblage is heterogenous, and a characteristic feature is the alternation of pelitic and graywacke-like, psammitic units with variable bed thicknesses. Although the differences within LA 4 are indistinct, three lithotypes are described.

The **Ylemmäinen lithotype** consists of frequently alternating psammitic and semipelitic units. The type area is located west of the lake Ylemmäinen (see Fig. 3) but similar lithologies are widespread troughout the eastern parts of the study area. The thickness of graywacke beds (lithofacies A) is 10 to 40 cm on average. In places indistinct grading has been observed. The fine grained units (lithofacies B) are typically about equal in thickness, but pelitic schists several meters in thickness are also present. The pelitic schist still displays a banded structure that conceivably represents primary alternation of silty and muddy laminae. The sand/shale ratio was estimated to be around 1 within this lithotype (Table 2).

The quartz in the psammitic schist still displays blastoclastic features, and the original grain size mostly corresponds to that of a fine or medium sand. Small intraclasts of dark phyllite are locally present. The mineral composition is quartz-plagioclase-biotite-sericite, with the main difference between the two lithofacies being the greater proportion of micas in facies B. Calcareous concretions are locally present in the psammitic schist; in addition to elliptical forms (Fig. 15) bedding parallel thin bands

Fig. 15. (A) Dark-colored calcareous concretions in psammitic schist of the Ylemmäinen lithotype at Rasivaara. (B) Erosional scour at base of a psammitic bed; LA 4, Ylemmäinen lithotype, Pankavaara. Tag is 16 cm long. For locations see Fig. 3.

rich in carbonate and calc-silicates are also present.

The Ylemmäinen lithotype is difficult to interpret conclusively. In spite of the absence (obliteration?) of Bouma sequences, the depositional style could be interpreted by using the turbidite concept. The grain size variation, composition of sandstones and the overall depositional style correspond quite well to the middle/lower fan deposits of the model proposed by Mutti and Ricci Lucchi (1972) or medial turbidites of Stow (1986). Nevertheless, a sequence of alternating mudstones and massive - or even graded - sandstones is not necessarily dianostic of a deep sea environment; they are reported to be common also in deltaic (e.g. Nemec, 1990) and shallow marine depositional settings (e.g. Ricci Lucchi, 1985).

Three lithofacies are distinguished within the **Kalliojärvi lithotype**. Facies A can best be described as siliceous semipelitic schist; no clasts are visible to the naked eye but the dark gray rocktype is rich in quartz and very hard. The facies probably originally represented quartzose silt. Lithofacies B is a dark pelitic mica schist. Facies C is medium to very coarse grained psammitic schist often displaying distinct grading in the basal parts (Fig. 16). The quartz grains still retain original clast boundaries, but the rest of the rock seems to have been extensively recrystallized (assemblage Biotite-Muscovite-Plagioclase) during metamorphism. In comparison with the Ylemmäinen lithotype the sand/shale ratio is considered to be greater, as is the general thickness and overall grain size of the individual psammitic beds.

The pronounced foliation obscures primary features in the fine grained portions and caution in interpretation is, again, needed. Nevertheless, the well developed grading indicates deposition from turbulent suspension flow and the dominance of sandstones, together with the moderate maximum grain size and bed thickness, suggest a location relatively proximal to the source.

The **Hovinvaara lithotype** is even more dominated by psammite, the sand/shale ratio of the type section (Fig. 17) being around 5. The psammitic units, originally consisting presumably of several amalgamated sandstone beds, may reach several meters in thickness whereas the fine grained beds seldom exceed 40 cm. Though the grading is poorly developed, the Bouma **Tae** and **Tabe** sequences may be discerned. Load casts, flame structures and erosional scours are locally present at the base of the psammitic units. The primary grain sizes of the psammitic schists of facies A (Ta) and B (Tb) correspond to coarse to very coarse and fine to medium sand, respectively. The coarsest parts are poor in matrix and display a nearly arenitic texture. These arkosic psammites

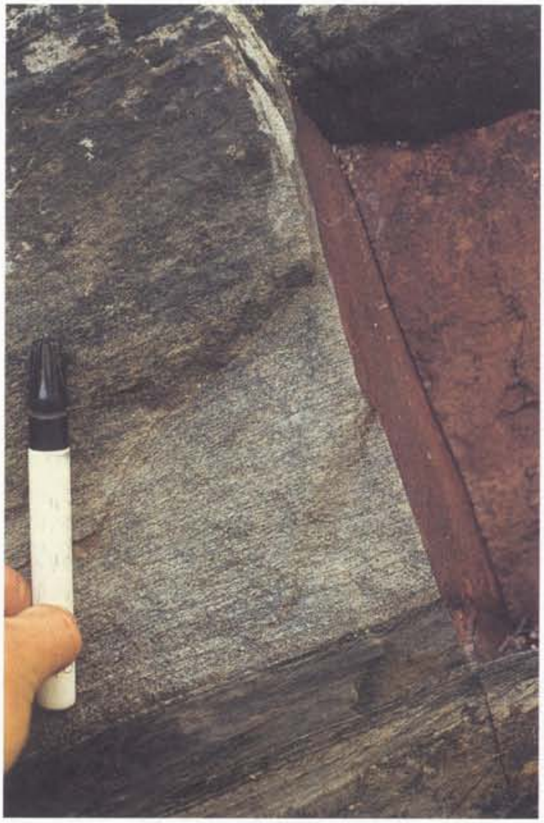


Fig. 16. Graded basal portion of a psammitic bed; LA 4, Kalliojärvi lithotype. Note also that S_1 foliation is slightly steeper than bedding in the semipelitic layer (bottom), consistent with depositional top indicated by grading. Pen for scale. Kalliojärvi type locality is indicated in Fig. 3.

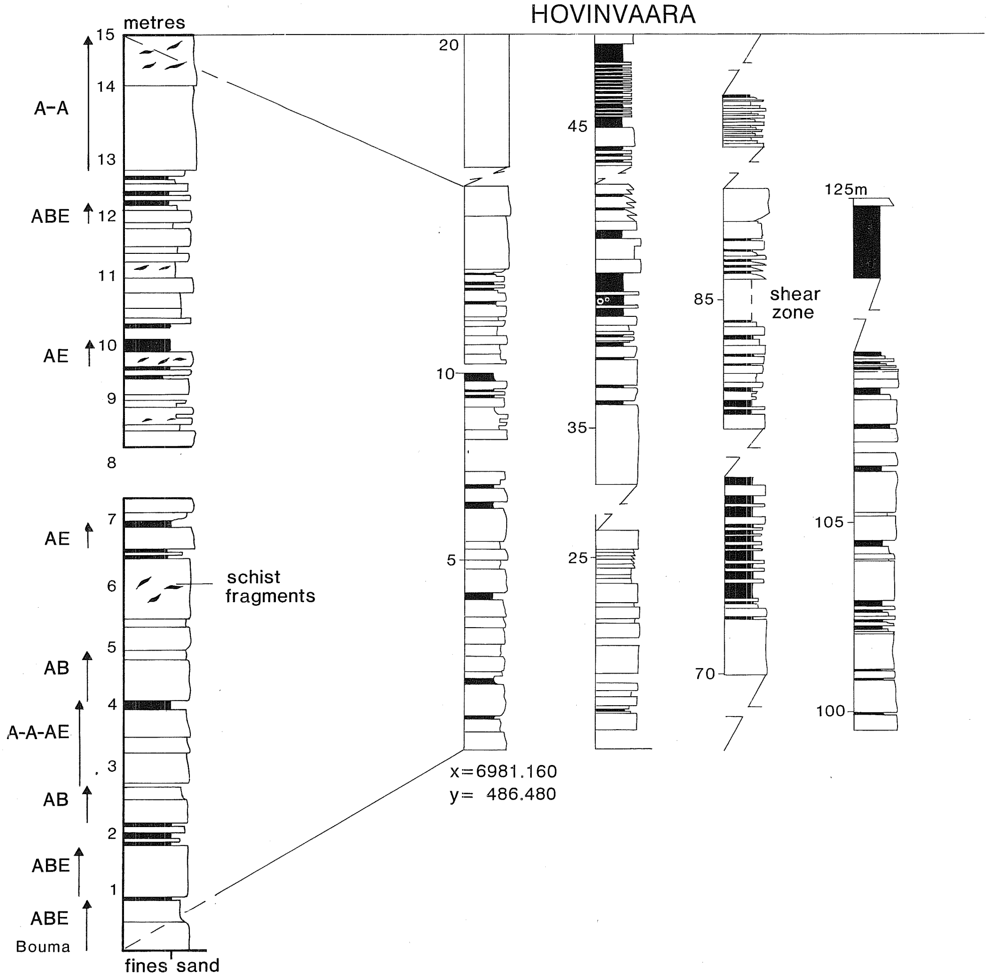


Fig. 17. Grain-size variation within the Hovinvaara section; mapped from west to east. Hovinvaara type locality is indicated in Fig. 3.

contain locally abundant clastic K-feldspar in addition to quartz and plagioclase, a feature which is exceptional when considering the study area as a whole (Kohonen, 1994). Intraclasts of dark phyllite are abundant in parts of the section. Facies C (Te) consists of dark gray pelitic schist.

The Mustakorpi conglomerate (Fig. 18) at Romppala, 3 km NW from the type section is also included within the Hovinvaara lithotype. The conglomerate beds are 40 to 100 cm in thickness and associated with coarse or very coarse arkosic psammites. The clasts within the conglomerate consist of granitic rocks, quartzite, carbonate rock and graphite bearing phyllite.

The presence of relatively well preserved primary structures, makes interpretation relatively straigthforward. All the features observed, including amalgamated, ungraded psammites, AE-sequences and the high sand/shale ratio, for example, correspond to those typical of proximal turbidites (cf. Walker, 1967; Stow, 1986). The Mustakorpi conglomerate fits well into this overall scheme as a channel fill deposit.

A) B)

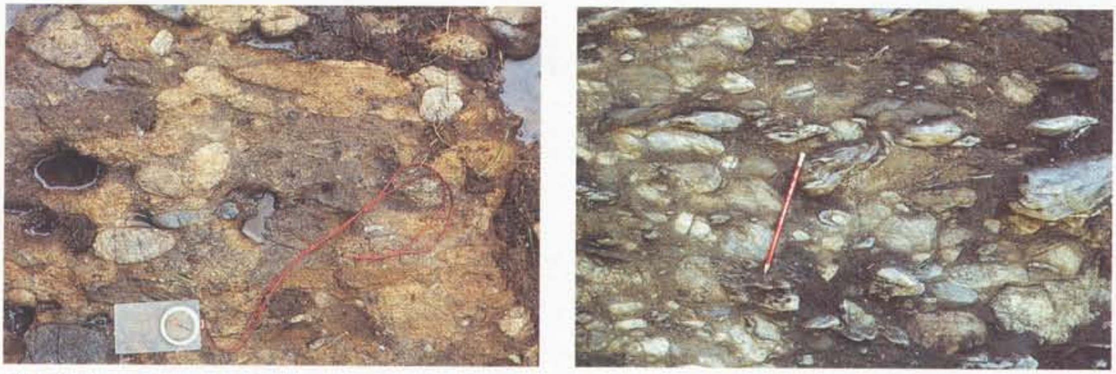


Fig. 18. (A) Mustakorpi conglomerate. Compass is 12 cm long. (B) Kuhnustanjärvi conglomerate. Pencil for scale. Both localities are indicated in Fig. 3.

Lithological assemblage 5

LA 5 is present along the eastern shore of the lake Höytiäinen and also to the north of it (see Fig. 3). The **Rantakylä lithotype** is dominated by monotonous light gray psammitic schist (lithofacies A). The graywacke schist is typically medium to coarse grained with quartz, plagioclase and micas as the main minerals. Most outcrops consist solely of this rocktype, the abundant calcareous concretions bringing the only variability. Individual beds are difficult to discern within the facies and distinct evidence of grading is absent. The psammitic schist is in places interrupted by a dark graphite-bearing pelitic schist (facies B); some of these units may be referred to as black schists. The **Sammallahti lithotype** further to the north is quite similar, but here too the psammitic rocks are often dark colored due to the presence of graphite.

LA5 is the only lithological assemblage that has a distinctive signature on aerogeophysical maps: the distribution of weak bedding-parallel electromagnetic anomalies (reflecting probably graphitic interbeds) coincides well with the mapped areal extent of LA 5.

The seemingly massive, monotonous psammites of these lithotypes are not very informative regarding depositional conditions. However, a high density flow origin seems most plausible (cf. Middleton and Hampton, 1976). Submarine slope apron (e.g. Hiscott and Middleton, 1979) and deep-sea fan channel (e.g. Walker, 1978; Wuellner and James, 1989) settings, at least, have been suggested for similar deposits consisting dominantly of massive sands. The carbonaceous, pyritic schist probably represents hemipelagic sapropelites, which is in accord with the assumed deep marine depositional environment.

A conglomerate occurrence at the eastern shore of the lake Kuhnustanjärvi is still to be described. The contacts of the conglomerate are not exposed, but on the eastern side of the occurrence, glacial boulders consist dominantly of pelitic schist whereas outcrops in the west are composed of dark gray monotonous graywacke, and associated black schist corresponding to the Sammallahti lithotype. The section, several metres in thickness, consists solely of conglomerate with clasts up to 30 cm in size (typically 1 to 10 cm; Fig. 18). The fragments are, in order of decreasing abundance: gray dolomite (weathered surface light brown), dark gray, impure carbonate rock, light gray quartzite (often containing carbonate bearing laminae) and carbonate cemented quartzite with a crumbling appearance on weathered surface. The pebbles, though deformed, still clearly retain well rounded outlines. The matrix is rich in carbonate. The conglomerate is tentatively interpreted as being submarine canyon incision into lithified platformal strata near a shelf edge.

MAJOR ELEMENT GEOCHEMISTRY

The main purpose of this chapter is to examine how the lithological assemblages described differ from each other in terms of their chemical compositions. Results from the study area are also compared with those from adjacent areas to the south and west.

Chemical data

Three separate data sets have been used: (1) analyses of the samples collected from the study area by the author (Table 3); (2) analyses from a larger area of North Karelia presented by Glumoff (1987; unpublished data, used by permission of Tapio Glumoff) and (3) unpublished data from the Kiihtelysvaara area provided by A. Kontinen. Only the mean values from this third set was used in the comparisons.

All major element analyses were determined by the XRF-method. The first set was analyzed at the laboratories of Rautaruukki Ltd. and Geological Survey of Finland, the second set by Outokumpu Ltd. and the third set by GSF. All analyses plotted on diagrams were first recalculated as 100% volatile free. Total iron is expressed as Fe_2O_3. Because of the lack of carbon dioxide determinations, recalculation on a carbonate-free basis was not possible.

General aspects of sediment geochemistry

Geochemical studies of clastic sediments are generally intended to reveal information concerning provenance area characteristics. However, the chemical composition of any particular sedimentary rock is controlled by the interaction of several factors, the most important being: (1) source rock composition; (2) sorting during depositional processes; (3) degree of chemical weathering; (4) amount of carbonate and other non-siliciclastic material and (5) compositional changes during post-depositional processes. In order to demonstrate the ambiguity inherent the source area interpretations the last four factors are briefly considered below.

The chemistry of sedimentary rocks is strongly influenced by the primary detrital grain size (e.g. Argast and Donnelly, 1987; Floyd and Leveridge, 1987; Korsch et al., 1993). A mineralogically biassed '**sorting effect**' is apparent in the high SiO_2 contents of coarse grained sediments and, accordingly, in the steady enrichment of most major and trace elements with decreasing grain size, due to progressive dilution of quartz and a concomitant increase in clay mineral (micas in metamorphic form) abundance (e.g. Roser and Korsch, 1986; Sawyer, 1986; Korsch et al., 1993).

Table 3. XRF-analyses of the metasedimentary rocks of the study area (Data set 1). Sampling sites are presented in Figure 23 and Appendix 1.

	Lithological assemblage 1									
	8306000	8311900	8502301	8702201	8702202	8702205	8702211	8707800	8800002	8803300
SiO_2	59.20	59.80	59.70	59.20	67.70	84.10	55.40	67.20	63.07	57.96
TiO_2	.91	.91	.82	.72	.78	.33	1.00	.75	.85	.87
Al_2O_3	16.50	16.10	16.80	19.60	14.40	6.12	17.70	14.20	15.64	16.54
Fe_2O_3tot	10.14	10.58	9.01	6.35	6.50	3.92	11.64	8.12	7.23	9.32
MnO	.07	.06	.06	.12	.11	.04	.11	.13	.08	.07
MgO	3.88	3.99	3.82	3.21	2.36	1.09	4.69	3.15	3.00	4.24
CaO	1.22	.90	.56	.62	1.58	.31	.42	.71	2.58	.67
Na_2O	1.66	1.39	.97	1.74	3.06	.98	.59	1.74	3.51	1.00
K_2O	3.46	3.12	4.10	4.94	1.89	1.42	4.16	1.94	1.72	3.98
P_2O_5	.13	.08	.08	.21	.18	.08	.15	.05	.10	.10
Total	97.17	96.93	95.92	96.71	98.56	98.39	95.86	97.99	97.78	94.75
CIA	65.50	68.82	70.45	67.75	59.14	62.25	73.96	69.44	55.93	69.77
$K_2O:Na_2O$	2.08	2.24	4.23	2.84	.62	1.45	7.05	1.11	.49	3.98
Na_2O+K_2O	5.27	4.65	5.29	6.91	5.02	2.44	4.96	3.76	5.35	5.26

	Lithological assemblage 2						Lithological assemblage 3				
	8616802	8617301	8617308	8617601	8801002	8806700	8615516	8616000	8708800	8709203	8709400
SiO_2	55.80	93.55	61.10	54.80	59.01	57.68	59.00	63.20	52.90	56.60	62.50
TiO_2	.81	.08	.85	.86	.83	.91	.89	.77	.88	.67	.70
Al_2O_3	18.70	3.56	16.50	19.10	18.64	16.95	18.30	14.40	20.30	18.40	17.20
Fe_2O_3tot	9.59	.47	10.12	9.92	6.31	9.84	7.55	9.42	9.92	10.32	6.87
MnO	.06	.12	.06	.18	.10	.15	.04	.06	.08	.05	.06
MgO	3.41		3.64	4.07	2.86	4.21	3.92	3.72	4.23	4.14	3.44
CaO	.39	.06	.35	1.63	2.78	1.36	.62	.23	.51	.19	.19
Na_2O	.64	1.12	.58	2.50	3.38	2.24	1.35	.76	1.05	2.29	2.11
K_2O	5.29	.55	4.60	2.34	2.79	2.71	4.62	3.08	4.55	3.68	3.17
P_2O_5	.10	.02	.09	.19	.07	.13	.10	.09	.16	.11	.08
Total	94.79	99.53	97.89	95.59	96.77	96.18	96.39	95.73	94.58	96.45	96.32
CIA	71.43	58.31	71.54	66.54	57.76	65.10	68.69	74.24	72.83	69.47	70.38
$K_2O:Na_2O$	8.27	.49	7.93	.94	.83	1.21	3.42	4.05	4.33	1.61	1.50
Na_2O+K_2O	6.26	1.68	5.29	5.06	6.38	5.15	6.19	4.01	5.92	6.19	5.48

Table 3. (cont.)

	Lithological assemblage 4									
	8610401	8610404	8615501	8615503	8615504	8615505	8617801	8506600	8508500	8608601
SiO_2	76.74	45.60	86.70	86.64	81.58	79.15	68.30	75.04	74.58	56.00
TiO_2	.57	1.03	.30	.33	.53	.47	.54	.55	.75	.78
Al_2O_3	9.82	22.10	6.46	6.19	8.85	9.28	13.10	11.20	10.53	18.90
Fe_2O_3tot	5.39	11.81	1.68	1.67	2.85	3.20	7.15	5.15	6.15	10.37
MnO	.03	.07	.02	.02	.02	.02	.07	.04	.04	.09
MgO	2.17	5.53	.57	.60	1.00	1.22	2.77	1.94	2.36	4.04
CaO	1.03	1.03	.40	.54	.41	.41	.47	.72	.69	.27
Na_2O	2.19	2.39	1.26	1.55	1.64	1.36	1.00	2.55	1.45	3.13
K_2O	1.66	5.33	2.01	1.49	2.57	2.79	3.73	2.35	2.81	2.68
P_2O_5	.07	.12	.04	.04	.07	.06	.11	.14	.25	.11
Total	99.67	95.01	99.44	99.07	99.52	97.96	97.24	99.68	99.61	96.37
CIA	57.46	65.64	56.50	54.62	58.72	60.74	66.73	58.20	61.19	68.90
K_2O:Na_2O	.76	2.23	1.60	.96	1.57	2.05	3.73	.92	1.94	.86
Na_2O+K_2O	3.86	8.13	3.29	3.07	4.23	4.24	4.86	4.92	4.28	6.03

	Lithological assemblage 4 (continued)									
	8608602	8608800	8609902	8610101	8617802	8703501	8707701	8707706	8707708	8707710
SiO_2	70.30	80.76	69.55	78.98	77.90	58.80	61.30	61.60	58.40	80.70
TiO_2	.56	.25	.70	.25	.44	.95	.78	.90	.23	.15
Al_2O_3	12.30	10.12	8.94	10.52	10.40	17.50	17.00	17.50	11.90	9.60
Fe_2O_3tot	5.03	2.16	5.47	3.26	3.92	8.92	8.74	7.69	1.56	1.22
MnO	.08	.06	.13	.02	.07	.23	.08	.07	.09	.03
MgO	1.87	.78	1.73	.93	1.41	3.66	2.72	2.54	.39	.34
CaO	1.77	.50	7.35	.26	.62	.37	1.92	1.61	11.30	1.57
Na_2O	2.88	4.22	1.68	3.31	2.65	.54	3.22	4.31	4.45	2.77
K_2O	2.37	.75	1.33	2.02	1.45	4.43	3.22	3.25	1.60	2.42
P_2O_5	.19	.09	.11	.09	.11	.11	.13	.16	.03	.04
Total	97.35	99.69	96.99	99.64	98.97	95.51	99.11	99.63	89.95	98.84
CIA	53.90	53.89	33.71	56.50	59.59	73.38	58.08	56.40	28.66	48.91
K_2O:Na_2O	.82	.18	.79	.61	.55	8.20	1.00	.75	.36	.87
Na_2O+K_2O	5.39	4.99	3.10	5.35	4.14	5.20	6.50	7.59	6.73	5.25

Table 3. (cont.)

	Lithological assemblage 4 (continued)									
	8708201	8708302	8804502	8805301	8807401	8808102	8902701	8902702	9000006	9000012
SiO_2	56.00	70.20	61.09	55.98	68.28	66.55	64.13	80.90	77.02	72.90
TiO_2	.83	.64	.72	.94	.64	.74	.81	.56	.33	.30
Al_2O_3	21.40	12.70	17.43	19.40	13.66	13.34	14.56	7.05	8.44	8.03
Fe_2O_3tot	6.79	6.67	7.30	7.39	6.70	7.10	7.93	4.71	2.47	3.03
MnO	.09	.05	.05	.09	.06	.05	.05	.08	.04	.05
MgO	3.40	2.78	2.72	2.63	2.69	3.04	3.41	1.89	1.64	2.52
CaO	1.51	.79	1.19	.50	.74	.51	.32	1.11	3.27	4.77
Na_2O	2.85	2.22	1.84	2.93	2.43	2.88	1.19	1.33	1.67	1.91
K_2O	4.64	2.52	4.01	5.78	2.46	2.29	3.94	1.22	2.09	1.74
P_2O_5	.12	.09	.10	.17	.14	.12	.13	.08	.07	.06
Total	97.63	98.66	96.45	95.81	97.80	96.62	96.47	98.93	97.04	95.31
CIA	63.22	61.92	64.66	61.83	63.06	62.11	68.17	56.06	43.51	36.94
K_2O:Na_2O	1.63	1.14	2.18	1.97	1.01	.80	3.31	.92	1.25	.91
Na_2O+K_2O	7.67	4.80	6.07	9.09	5.00	5.35	5.32	2.58	3.87	3.83

	Lithological assemblage 5									
	8615800	8616100	8616201	8616202	8616702	8707400	8800001	8807001	8901101	8901500
SiO_2	71.90	71.80	60.60	71.40	70.30	70.50	66.98	68.45	67.60	71.82
TiO_2	.48	.51	.70	.51	.55	.53	.68	.52	.58	.55
Al_2O_3	13.00	14.00	17.80	14.30	13.70	14.80	13.74	13.48	14.09	12.84
Fe_2O_3tot	3.92	4.33	7.67	4.02	4.99	3.40	5.45	4.36	5.10	4.42
MnO	.05	.05	.09	.04	.06	.04	.07	.08	.07	.07
MgO	1.69	1.92	3.26	2.05	2.50	1.89	2.51	2.00	2.40	1.89
CaO	1.61	1.47	.96	1.43	1.79	1.41	2.18	2.75	1.89	1.80
Na_2O	2.84	2.48	1.80	2.24	3.11	2.40	3.46	3.85	3.94	3.41
K_2O	2.17	2.82	4.24	3.14	2.31	2.99	2.04	2.33	2.32	2.48
P_2O_5	.13	.14	.14	.13	.13	.13	.15	.14	.14	.15
Total	97.79	99.52	97.26	99.26	99.44	98.09	97.26	97.96	98.13	99.43
CIA	56.66	58.82	65.71	59.63	55.76	60.30	53.67	49.31	53.14	52.61
K_2O:Na_2O	.76	1.14	2.36	1.40	.74	1.25	.59	.61	.59	.73
Na_2O+K_2O	5.12	5.33	6.21	5.42	5.45	5.49	5.66	6.31	6.38	5.92

The sorting effect makes the mean values sensitive to the grain size fraction sampled and in most cases hinder direct comparison of averages from published studies. These difficulties can be largely avoided by using sandstone-mudstone pairs, trend of individual data point arrays or by sampling of a particular grain size fraction (e.g. mudstones).

Chemical weathering results in hydration of silicate minerals and in removal of mobile elements, especially those incompatible in clay minerals. Calcium, for example, can be substantially depleted even in the early stage weathering processes. The amount of matter leached out during weathering is impossible to estimate quantitatively, and this is one of the major problems in provenance studies based on major element data.

The most typical weathering products of ferromagnesian minerals are chlorite, smectite and vermiculite, whereas feldspars yield the suite hydromuscovite-illite-kaolinite, respectively, reflecting increasing intensity of alteration. The extent of chemical weathering may be evaluated by the chemical index of alteration (CIA) introduced by Nesbitt and Young (1982), and based on molecular proportions of the major oxides: $Al_2O_3/(Al_2O_3+CaO^*+Na_2O+K_2O)/100$ where CaO^* represents CaO in the silicate phases only. The CIA value for fresh feldspars is 50, for illite 75 and for kaolinite and chlorite 100.

Samples rich in oxides or sulphides are generally avoided in sampling, but **carbonates** may seriously distort the results if no correction is made. An increasing amount of carbonate not only causes a spurious rise in CaO (MgO, MnO, FeO) levels, but tends to generate secondary 'trends' by dilution of other components. For example, on Harker variation diagrams individual data points become displaced towards the origin as the carbonate component increases, such that in extreme cases, a trend perpendicular to that defined by grain size variations, may result.

The origin of carbonate is also an important question in provenance studies. If the carbonate is an authigenic alteration product after Ca-bearing clastic minerals (e.g. plagioclase), it clearly should be included in calculations. In contrast a correction is needed for the carbonate of chemical origin, irrespective of whether it precipitated in situ, or represents reworked carbonate detritus from elsewhere. Furthermore, diagenetic concretions are difficult to take into account, because their abundance is generally impossible to estimate.

Finally, the most difficult and controversial question of **post-deposional changes in composition** of sediments is briefly addressed. This issue is obviously related to some fundamental aspects of metamorphism, such as closed versus open systems, and sedimentary petrology, notably the matrix problem and the problem of Na_2O in graywackes (see Pettijohn et al., 1972). Because common minerals occurring in different proportions may produce fortuitous similarities in chemical composition, the importance of supporting petrographic studies should not be underestimated. Significant changes in mineral compositions may occur during diagenetic and metamorphic processes, but the crucial question is whether the scale of diffusion related to these changes is large enough to affect the bulk chemical composition of the samples.

Diagenesis is a multifarious - and insufficiently understood - process characterized by the breakdown of unstable rock fragments, the appearance of authigenic mineral phases, various replacement reactions and decrease of porosity. Chemical reactions during diagenesis seem to be variable, or difficult to quantify, and mobility of many elements, including Si, Fe and Mn under certain conditions, is evident (e.g. Taylor and McLennan, 1985). From the present point of view the major difference between diagenesis and metamorphism is that most prograde metamorphic changes are water releasing dehydration reactions. The resultant aqueous fluids are plausibly capable of transporting mobile elements, such as alkalies and alkali-earths, and therefore the system can hardly be assumed to be 'closed' in the strict sense of the term.

The extent to which changes in bulk chemical compositions occur during diagenesis and metamorphism have been studied by comparing the compositions of sedimentary or metasedimentary rocks with those of corresponding unlithified sediments. Most workers (e.g. Ronov et al., 1977; Maynard et al., 1982; Argast and Donnelly, 1987) concluded that such effects are relatively minor. However, the K_2O/Na_2O ratios of lithified shales appear to be generally higher in comparison to recent muds (e.g. Maynard et al., 1982 and references therein), and the data presented by Taylor and McLennan (1985) indicate a similar trend. This feature seems even more striking when comparing mud-sand and shale-sandstone pairs in that the difference in K_2O/Na_2O ratios between lithified pairs is generally greater than in recent ones. Depletion of calcium and alkalies (especially Na) relative to aluminium during the lithification process has also been reported (e.g. Sayles, 1981; Boles, 1982; van de Kamp and Leake, 1985).

Major features and trends

Some of the main features of the present data set are readily apparent from the Harker variation diagrams (Fig. 19). Most elements correlate negatively against SiO_2 and, conversely, show positive correlation with Al_2O_3. The Al_2O_3 vs SiO_2 trend mainly reflects grain size variation (Fig. 20), and is simply a result of the higher proportion of quartz in the coarser grained sediments. TiO_2, Fe_2O_3, MgO and K_2O show trends similar to that of Al_2O_3 and are all relatively enriched in the pelitic rocks. In contrast the CaO and Na_2O contents seem to be largely independent of the main trend and show large scatter in all grain size classes.

In general the lithological associations described earlier do not define any distinctive groups in the Figure 19. However, the LA 5 psammites all plot within a small area, which cleatly indicates a restricted compositional variation within the group. The assemblages 1, 2 and 3 lack samples with SiO_2 between 70 and 85%, reflecting a distinct bimodal division into aluminous pelitic schists and quartzitic psammites. In contrast the LA 4 samples show a continuum from pelitic to nearly quartzitic compositions. All three samples showing SiO_2 less than 53 % in Figure 19 represent outcrops characterized by abundant quartz veins, so that the possibility of post-depositional SiO_2 depletion accompanying vein segregation seems tenable in these instances.

In Figure 21 the samples representing assemblages 1, 2 and 3 show slightly higher CIA values than those for groups 4 and 5. This is consistent with the observation that the psammites of the former groups are mature quartzites. The two group 1 samples with low CIA values are rich in albite, which is presumably secondary in origin. The psammitic rocks of group 5 again form a coherent group showing CIA values typical of slightly altered feldspars.

Carbonate does not appear to cause serious problems in the present study since according to both chemical analyses and petrographic studies, abundances are generally low. Because the composition of detrital plagioclase ranges from albite to oligiclase, samples containing carbonate are easily distinguished by having CaO/Na_2O values greater than 1 (Fig. 21).

The very high K_2O/Na_2O ratios of some pelitic schists (Fig. 20) might indicate local secondary enrichment of potassium in those layers presently rich in mica. The decreasing variability of alkalies when summed as Na_2O+K_2O (Figs. 19 and 20) could perhaps also be explained by mobility of these ions during diagenetic and metamorphic processes.

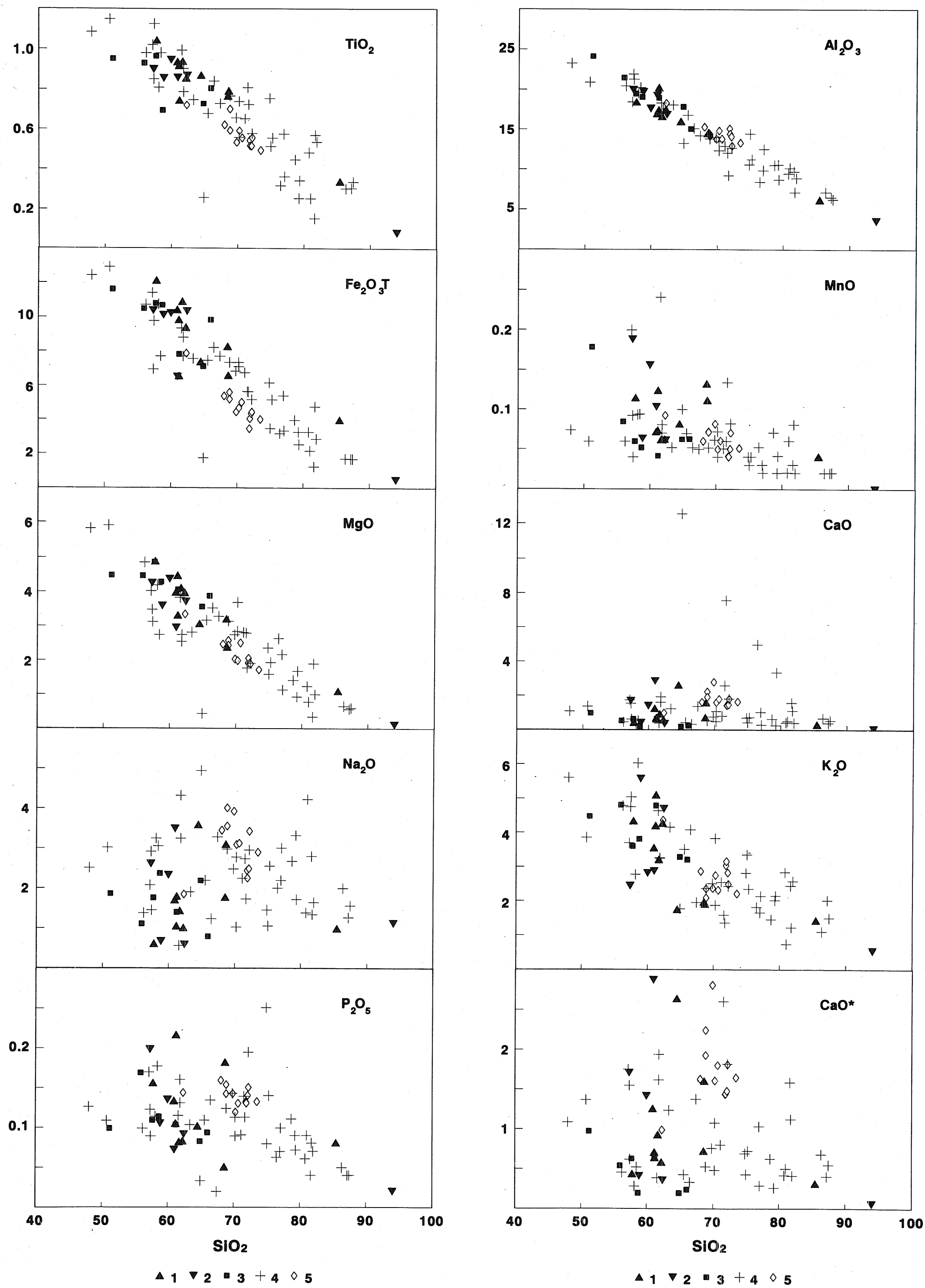


Fig. 19. Harker variation diagrams for the Kaleva rocks of the study area (data sets 1 and 2). Numbers for symbols correspond to respective lithological assemblages (1= LA1, etc.). Value of CaO* differs from CaO only by different vertical s

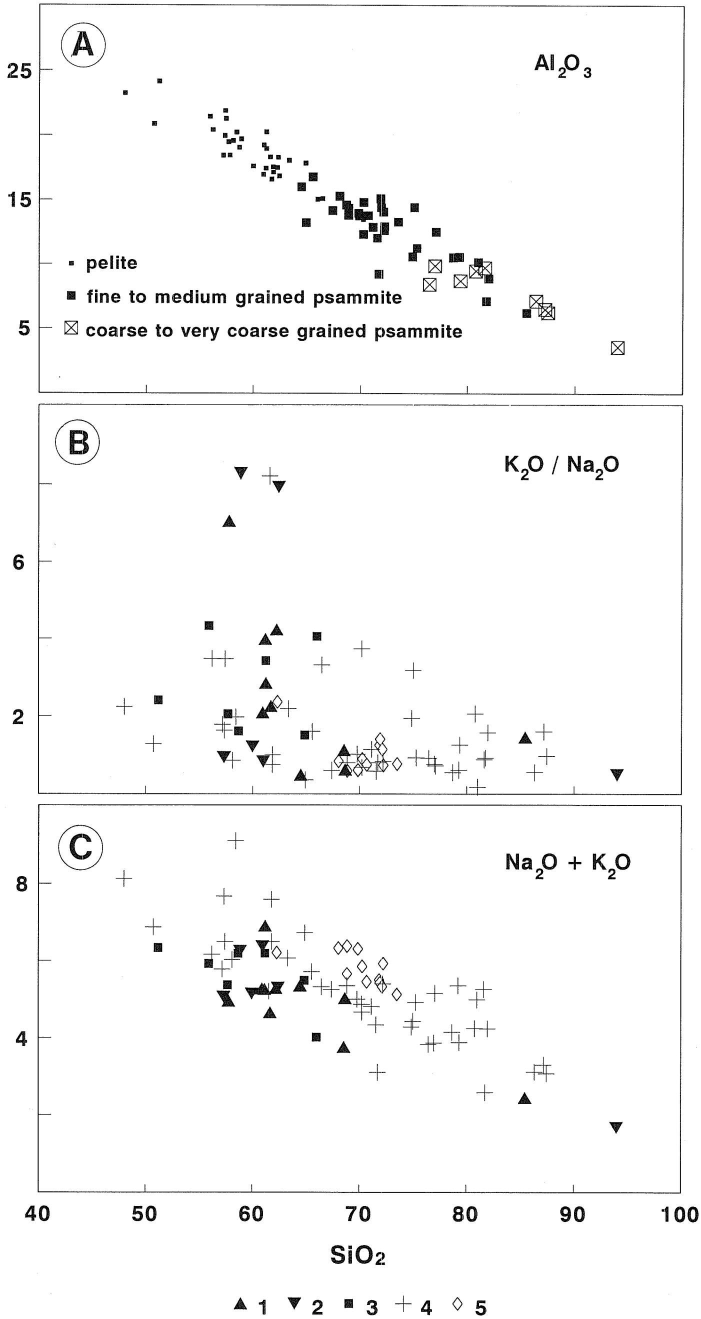


Fig. 20. Plots indicating (A) the relationship between grain size and SiO_2 content, (B) K_2O/Na_2O vs. SiO_2 and (C) Na_2O+K_2O vs. SiO_2.

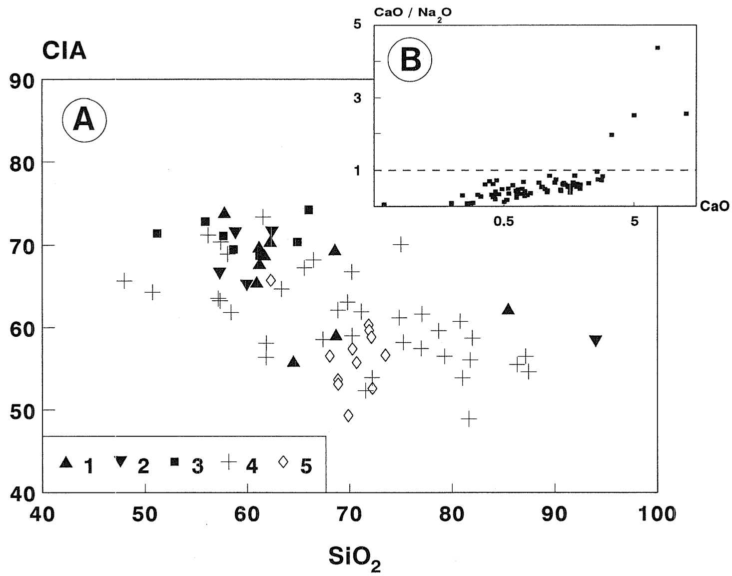


Fig. 21. (A) Chemical Index of Alteration plotted against SiO_2. For symbols see Fig. 19. (B) CaO/Na_2O vs. CaO plot indicating the samples containing carbonate.

Major elements and primary clastic components: the problem with K-feldspar

Even the applicability of unmetamorphosed sandstone mineralogy in provenance studies has been questioned, because diagenetic processes are capable of causing major modifications (Blatt, 1985 and references therein). Metamorphism may cause even more radical changes in mineral composition. Unfortunately, major element geochemistry does not necessarily reveal the nature of primary clastic components through such modifications. To clarify this assertion, consider a hypothetical, simple bulk sediment consisting of quartz, K-feldspar and kaolinite. Given that the proportions are appropriate, after breakdown of Ksp and illitization of the clay component, a composition indistinguishable from a primary mixture of quartz and illite may result. Numerous similar examples exist; for example a K-feldspar-chlorite (or smectite) mixture is, in its volatile free form, chemically equivivalent to the assemblage biotite-illite-quartz. Hence, it seems that the discrimination of clastic components by 'end member technique' (cf. Argast and Donnelly, 1987) is possible only in simple cases or by making major assumptions. However, if the premise that post-depositional compositional changes are only relatively minor is accepted, then it may be possible to identify a number of original clastic composition alternatives.

The K-feldspar was only rarely observed in the studied rocks, and the scarcity of Ksp within the Kaleva rocks in North Karelia was also noted by Pelkonen (1966), Nykänen (1968, 1971a) and Huhma (1975). In the following discussion possible reasons for the dearth of Ksp are outlined. In Figure 22A arkosic rocks, consisting of quartz and feldspars, tend to plot near the line connecting the alkali feldspars. Samples containing minerals rich in Al, Fe and Mg, such as chlorite or kaolinite, plot above the line connecting albite and biotite. The plot therefore clearly reflects the observed metamorphic mineral compositions (quartz, albitic plagioclase, micas and chlorite). The three samples containing Ksp plot near the 'arkosic line' indicating that the sediment was originally poor in clay minerals, but none of these samples have conspicuously high K_2O/Na_2O ratios. In general the K_2O/Na_2O ratios for the psammites plot near to value 1 (K_2O/Na_2O+K_2O = 0.5 in Figure 22A) on both sides. If the possibility of cation exchange between compositionally different layers during metamorphism - the formation of micas and secondary enrichment of K_2O in pelitic schists now rich in micas - is taken into account, the psammite values may be considered to record minumum estimates for original K_2O/Na_2O ratios, and the possibility of a source region exceptionally poor in potassium can be excluded.

K-feldspar is normally more resistant to weathering than plagioclase and should presumably be enriched in psammites showing

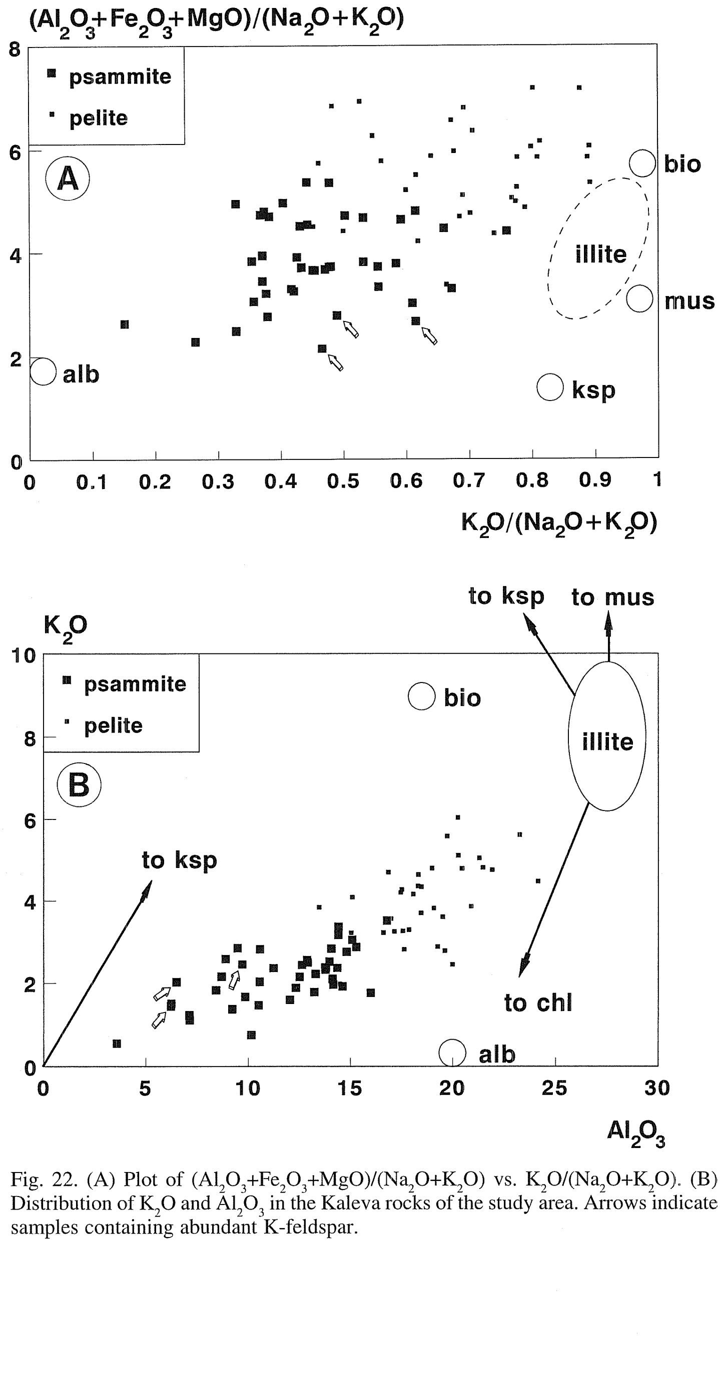


Fig. 22. (A) Plot of $(Al_2O_3+Fe_2O_3+MgO)/(Na_2O+K_2O)$ vs. $K_2O/(Na_2O+K_2O)$. (B) Distribution of K_2O and Al_2O_3 in the Kaleva rocks of the study area. Arrows indicate samples containing abundant K-feldspar.

moderate or high K_2O/Na_2O values. If both the present chemical composition and the absence of Ksp are assumed to be primary features, then a rather unusual apparent clastic composition is implied. In addition to quartz, the original sand would seem to have contained albitic plagioclase (3 % Na_2O requires nearly 30% pure albite component) with a nearly equal amount of detrital mica (or illite), together with some mafic material. However, an explanation implying a major source area consisting of rock types that were rich in potassium but effectively devoid of Ksp is considered unlikely.

When a three component system composed of illite, plagioclase and K-feldspar (in addition to quartz) is presumed, variations in K_2O/Al_2O_3 should, according to Argast and Donnelly (1987), represent different proportions of Ksp and illite. However, with the present data the diagram (Fig. 22B) does not discriminate between samples in which K-feldspar was observed and those that were devoid of Ksp. Consequently, it is clearly not valid to ignor chlorite (mafic component) when deducing primary clastic components.

In summary, although detailed chemical discrimination of clastic components is not possible, the relatively high average K_2O/Na_2O ratios imply that K-feldspar was presumably an essential component in the precursor sediments. As suggested by Kohonen (1994), some of the abundant metamorphic biotite may have been formed from reactions between chlorite and K-feldspar. In support of this argument, the local preservation of blastoclastic Ksp in some of the most coarse grained psammites seems to correlate with a relative scarcity of chlorite containing matrix.

Regional comparisons

Although a more comprehensive geochemical study of the Kaleva rocks of North Karelia and Kainuu is in progress, some preliminary regional comparisons are presented here. Four subregions within North Karelia have been distinguished (Fig. 23).

It seems that the rocks of Kiihtelysvaara area do not differ substantially from the rocks of the eastern (main) part of the present study area (Fig. 24). The striking difference in compositions between psammites and pelites in the east may reflect fractionation of elements due to weathering and sorting processes. According to their CIA values the eastern schists (LA 1,2,3,4 and Kiihtelysvaara) ap-

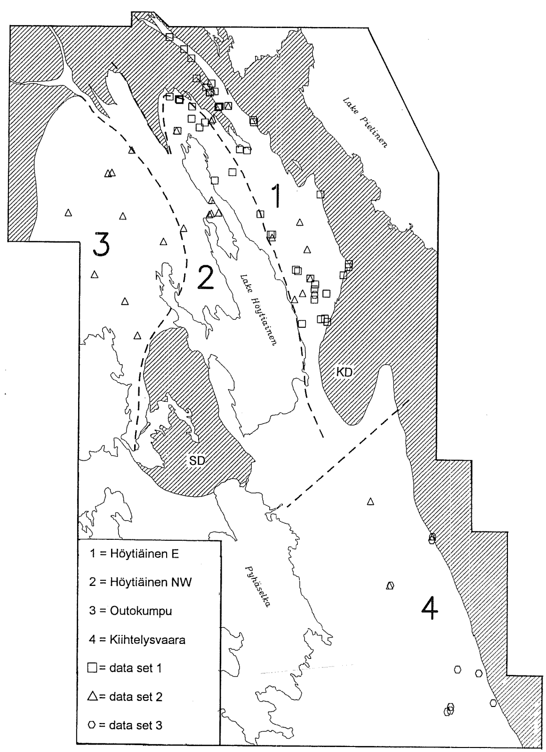


Fig. 23. Map indicating the subregions used in comparisons. 1= Höytiäinen E (LA's 1-4 of the study area), 2= Höytiäinen NW (Höytiäinen W + LA 5 of the study area), 3= Outokumpu, 4= Kiihtelysvaara. Sampling sites are also shown. Symbols indicate different data sets.

pear to be compositionally more mature when compared with the Höytiäinen W and Outokumpu subregions. The LA 5 psammitic schists in the present study area resemble the western rocks and especially those of the westernmost Outokumpu area. In most diagrams the psammites of these associations form quite a coherent group corresponding to the quartz-intermediate graywackes of Crook (1974). The Höytiäinen W area apparently represents some kind of intermediate between the areas to the east and west. Because the LA 5 schists of the study area clearly show geochemical similarities with the rocks from the western areas, the LA 5 area is amalgamated with the Höytiäinen W subregion in the following provenance comparisons.

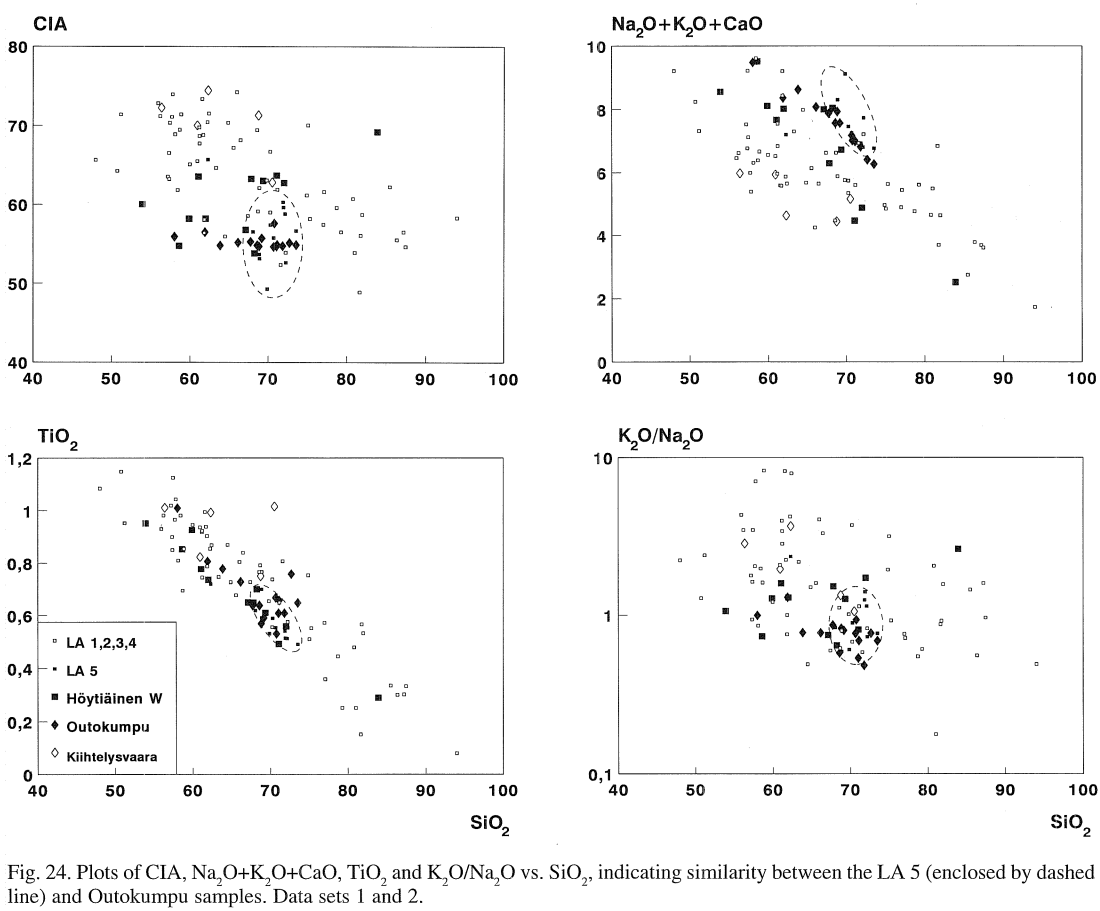


Fig. 24. Plots of CIA, Na_2O+K_2O+CaO, TiO_2 and K_2O/Na_2O vs. SiO_2, indicating similarity between the LA 5 (enclosed by dashed line) and Outokumpu samples. Data sets 1 and 2.

PROVENANCE

Evidence from detrital components

Both the pelitic and psammitic schists have an essentially metamorphic mineralogy, and therefore sophisticated techniques based on detrital modes, such as the Gazzi-Dickinson method, are not applicable to provenance studies. However, the proportion of quartz in psammites (Table 2) still has probably some significance. For example, the quartzite interbeds within the LA 1 indisputably indicate an exceedingly mature intimate source, whereas the graywackes rich in feldspar display immature or mixed natures.

The only direct indications of provenance are provided by the sparse conglomerate occurrences. The conglomerates and breccias at the eastern margin of the study area are characterized by a variety of quartzite fragments. The breccias and breccia-conglomerates are oligomict and, if sedimentary in origin, indicate a local source dominated by quartzite. The mafic fragments in the Tottolampi conglomerate may be perhaps best understood to represent detritus derived from the volcanics and dykes that are commonly associated with the eastern platformal quartzites. A noteworthy feature of the conglomerates at the eastern margin is the presence of granitic fragments; in both the Juntulankylä and Latvajärvi occurrences the abundance of these increases upwards over a short distance. This is attributed to rapid, probably fault related, exhumation exposing the basement in the source area and/or highly condensed, hiatal nature of the conglomerates.

Both of the conglomerate occurrences (Kuhnustanjärvi and Mustakorpi) within the mica schist area indicate the presence of quartzites and carbonate rocks in their respective source areas. Nevertheless, also felsic plutonic fragments are common in the Mustakorpi conglomerate. According to a zircon ($^{207}Pb/^{206}Pb$) age determination from a granodiorite fragment (Huhma, H. 1991, written communication), these very probably represent detritus derived from the Archean basement.

Evaluation of major element geochemistry

Various parameters based on major elements have been used in estimating source rock compositions and the tectonic setting of sandstone-mudstone suites (e.g. Crook, 1974; Bhatia, 1983; Roser and Korsch 1986, 1988). Although the abundance of most major elements, and many widely used parameters calculated from these (e.g. K_2O/Na_2O) are grain size dependent, some consistent general features doubtless exist. For example, both the SiO_2 and K_2O/Na_2O-ratio tend to be high in sands from passive margin settings with a cratonic source, whereas FeO+MgO and TiO_2 tend to be higher in sands derived from active volcanic arcs.

Geochemical discriminant diagrams should be used only after a close inspection of the reasons behind the variability of the data. In the present study it seems that differences in the grain sizes sampled is the major factor explaining the observed trends. However, the variability between samples having equivalent detrital grain sizes is, even within a single lithological unit, rather large. In particular, the large scatter observed for alkali elements and CaO quite plausibly indicate post-depositional mobility and question the applicability of many widely used parameters.

In provenance studies it is, however, perhaps not always crucial whether the variability between individual samples is due to primary, local differencies in compositions or caused by post-depositional mobility of certain elements. If a sample set is truly representative of a larger depositional unit, that can be reasonably assumed to be isochemical as a whole, then mean values can indeed be used to determine the composition of the source area. However, determination of such mean values depends substantially on the proportion of pelites and psammites in the data set, and some estimation of relative volumes is accordingly necessary. In the following an attempt to develop a more appropriate method of provenance evaluation is presented.

Fundamental to this approach is the observation that an Al_2O_3 abundance of 16% discriminates effectively between the psammitic and pelitic schists (Fig. 20A). According to Taylor and McLennan (1981) this value is very close to the estimated amount of aluminium in average upper continental crust. If we assume (1) that all the aluminium is transported in clastic form and (2) a source area producing detritus with Al_2O_3 16% on average, the proportions of psammite and pelite fractions can be calculated. Furthermore the composition of the average bulk sediment may be estimated by computational mixing using constant Al_2O_3 = 16%.

Firstly, mean values for both psammites (samples with Al_2O_3 < 16%) and pelites (Al_2O_3 > 16%) for a particular unit are calculated separately. During the second step a hypothetical mixture of the resulted average shale and sandstone is computed: (1) all the element percentages for the avarage pelite are multiplied by a factor A ((16-Mpsa)/(Mpel-Mpsa)) and correspondingly the multiplier for the average psammite percentages is B (Mpel-16)/(Mpel-Mpsa), where Mpsa and Mpel are the mean Al_2O_3 for psammites and pelites, respectively; (2) the resultant percentages for all the major elements are then summed. In this theoretical mixture Al_2O_3 is always 16%, and A and B represent the proportions of pelite and psammite in the compound.

Because this method largely removes the effects of variable grain sizes and post-depositional element mobility, it provides an estimate of the original clastic bulk composition of a given unit. Consequently, the major advantage of the method is the possibility of direct provenance estimation, instead of comparison with other sediments. The main drawback is that the second assumption (source material with Al_2O_3 = 16%) is unlikely to be exactly fulfilled. Consequently, when the Al_2O_3 content of the natural bulk sediment exceeds the value 16%, the elements associated with aluminium are underestimated and vice versa.

In Figure 25 the calculated compound analyses from different subareas are normalized with respect to average values of the present upper crust (Taylor and McLennan, 1981) and Archean upper crust (Taylor and McLennan,1985), respectively. In the first case the Kaleva sediments seem to be enriched in elements of 'mafic' affinity (TiO_2, Fe_2O_3 and MgO). In contrast, comparison with Archean upper crust indicates considerably higher amount of potassium in the Kaleva sediments, together with some depletion of MgO, Fe_2O_3 and TiO_2. The depletion of CaO and Na2O is obvious in both cases and the differences between the various subareas probably indicates the presence of different lithologies within the respective source areas, combined with variable degrees of weathering .

If the amount of element mobility due to diagenesis and metamorphism is assumed to be significant, it should be possible to compare and contrast corresponding grain size fractions of sediments of different age and deposited in different environments. In Figure 26 the sandstone (psammitic schist) means from different subareas in North Karelia are compared with the mean values for

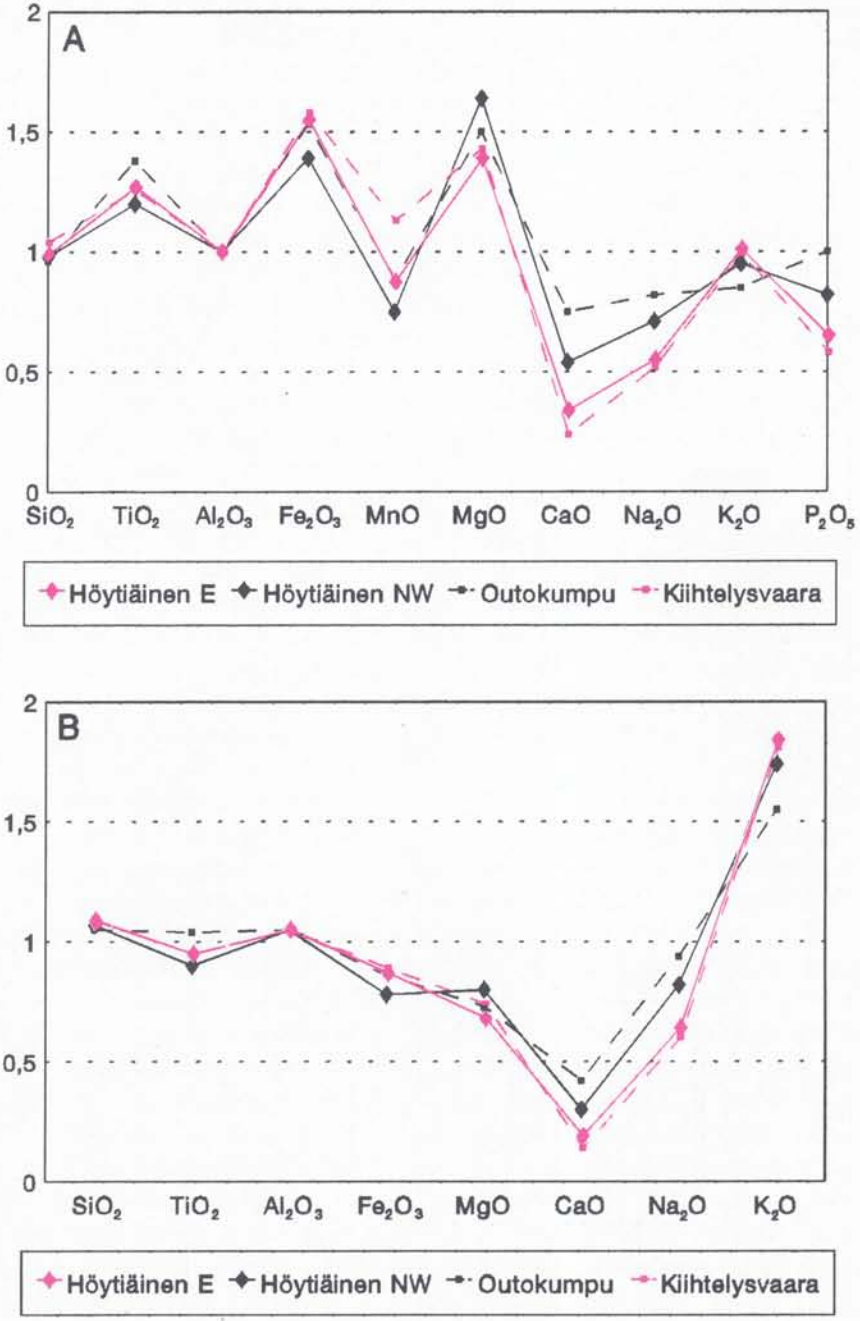


Fig. 25. Calculated compound (computational mixture of psammite and pelite) values from different subareas of the NKSB normalized (A) to the composition of the present upper crust and (B) to the estimated Archean upper crust. Data sets 1-3.

typical Paleozoic psammites from various tectonic settings (Bhatia, 1983). The distribution of the major elements seems to show passive margin/active continental margin characteristics in the eastern part of the area, as opposed to an active continental margin/ continental island arc affinity for the western Outokumpu area.

However, it should be noted that the main difference between the 'passive margin' and 'active continental margin' type sands simply reflects maturity (SiO_2 content), and that the 'active continental margin' 'continental island arc' and 'oceanic island arc' sands form a sequence with increasing amount of the mafic component (for definitions and typical source areas for PM, ACM, CIA and OIA, see Bhatia, 1983). Thus, the Figure 26 must be used in relation to provenance evaluation, not as a conclusive discrimination of tectonic settings.

The major element geochemistry suggests that the provenance area for the studied Kaleva rocks was relatively rich in K_2O, roughly corresponding to the mean values for the present upper crust. Consequently, an immature arc source can virtually be eliminated, particularly with respect to the present study area. However, the Sm-Nd results with T_{DM} model ages 2.7-2.3 Ga presented by Huhma (1987) preclude an exclusively Archean source for the Kaleva sediments. The only sample from the present study area (Kalliojärvi) shows ε_{Nd} of -3.4 (T_{DM} 2.47 Ga), and quite a small contribution from the mafic dykes and/or volcanics could here explain the observed Proterozoic component.

In summary, it seems that the Karelian Craton in the east fulfills the major conditions for the provenance and is thus the most natural

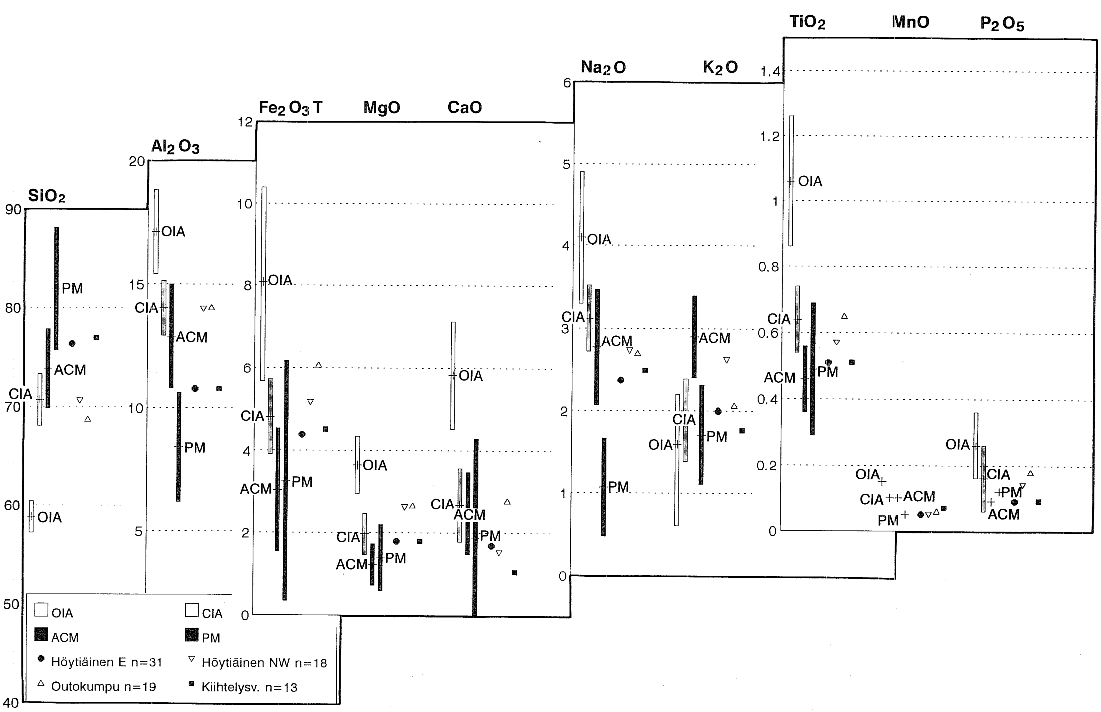


Fig. 26. Mean values of psammites from NKSB subareas compared to the average chemical composition of sandstones from various tectonic settings (data from Bhatia, 1983). OIA= oceanic island arc, CIA= continental island arc, ACM= active continental margin, PM= passive margin. Data sets 1-3.

source area for the Höytiäinen E and Kiihtelysvaara rocks. The relatively high values for K_2O are quite easy to explain with the abundant late Archean potassic granites (such as Kutsu-type granites, cf. Nykänen 1968, 1971a) intruding the basement gneisses. The major element geochemistry is also consistent with the presence of a minor component from the Proterozoic (syn-rift) mafic volcanics dykes. When compared to the Höytiäinen E area the provenance of the western Outokumpu schists seems to be less weathered, more rich in plagioclase and, conceivably, slightly more mafic.

STRUCTURAL GEOLOGY

Overall structural style

To generalize, the North Karelia Schist Belt may be described as a foreland thrust belt showing increasing basement involvement towards the Svecofennian orogenic front in the west. In the western parts of the NKSB the basement thrusts are interpreted as having followed a phase of cover nappe emplacement and related deformation (Park and Bowes, 1983), whereas these early deformation stages are weakly expressed or absent in the eastern part of the region, designated the Höytiäinen province (Ward, 1987).

Most studies in different parts of the Schist Belt (e.g. Koistinen, 1981; Park and Bowes 1983; Ward, 1987) agree in that the crustal thickening was followed by deformation related to wrench movements (for review of the regional structural sequence, see Park et al., 1984; Ward and Kohonen, 1989). According to geophysical studies (e.g. Kohonen and Elo, 1991; Korja et al., 1993) the thickness of the cover sequence seldom exceeds 5 kms, and it seems that the basement structures largely control the structural style in the cover rocks.

The **basement** gneisses display a complicated Archean history, but the style and areal extent of Svecofennian 'reworking' is poorly defined at the regional scale. However, studies from the Kaavi - Nunnanlahti area (WNW from the present study area) reveal some information about basement structures. Within the Kaavi-Niinivaara district an easterly vergent, moderately dipping thrust system of sliced basement is evident (Park and Bowes, 1983; Park and Doody, 1991). The area is characterized by SW plunging linear elements and N-S striking mylonitic foliation along the thrust zones. Apart from within the major zones of movement the basement gneisses do not appear to record significant strains relatéd to the cover deformation (Park and Bowes, 1983).

Further east, near Nunnanlahti, the streching lineations plunge towards the SE and the Archean granitoids show major reworking with subvertical zones of mylonitic (NW-SE) foliation. Kohonen et al. (1991) interpreted these features by a major sinistral shear zone ('Nunnanlahti-Holinmäki Shear Zone') affecting both the cover and the basement. Still further east the basement granitoids seem to have remained largely unaffected by the cover deformation. However, the moderately to steeply dipping basement-cover contact in Koli and Kiihtelysvaara districts, respectively, demands, on the ground of geometry alone, some ductile deformation of the basement.

A typical feature of the basement reactivation is the alternation of intensely deformed high strain domains and intervening areas that remain nearly unaffected. The rheology (and orientation?) of basement lithologies appears to be an important factor controlling the nature of structural reworking. The relatively imcompetent Archean greenstones localized strain (Kohonen et al., 1991), and these inhomogenities in the regional basement strain field are reflected also in the structural pattern of the cover.

The structures of the **autochthonous quartzite** are directly controlled by the behaviour of the underlying basement. Within the Koli area the quartzite overlies nearly unstrained granitic gneiss and indicates shear subparallel to the gently dipping bedding planes; imbrication, duplex-like geometries and relative lack of folding are characteristic. On the other hand, the same quartzites locally display ductile 'cross-fold' like deformation where underlain by Archean supracrustal rocks (see Kohonen, 1988).

In contrast to the quartzites, the **Kaleva mica schists** are tightly or isoclinally folded, and a pronounced, penetrative foliation is characteristic. The different style of deformation compared to the quartzites does not necessarily require the existence of a detachment between the units. Quite a good example is in the Juuanvaarat area (NW from the study area, see Fig.3) where the ductility gradually increases upwards in the sequence, and shortening represented by imbrication of the basement is accommodated instead by ductile folding in the overlying mica schist.

Structural sequence in the Kaleva mica schists

The use of overprinting criteria in the determination of the structural sequence in polyphase deformed rock (e.g. Hobbs et al., 1976) is in principle a straightforward and rigorous method. In practice, however, sequential relationships are often difficult to define and correlate unambiguously. Strong transposition and/or reactivation may produce composite fabrics that defy resolution. On the other hand, progressive refolding of mylonitic foliation has been reported (Bell and Hammond, 1984), and the rotation of the finite strain ellipsoid may result in anastomosing, or even apparently cross-cutting (C/S fabric), foliation patterns within a shear zone (see Ramsay and Graham, 1970; Hanmer and Passchier, 1991). These kinds of structures can easily be mistaken in the field as the unrelated products of separate deformational events.

Furthermore, the difficulty of correlating structures is a major problem confronting interpretations based on regional sequences. In the absence of metamorphic contrasts, correlation criteria such as style and orientation of structures are ambiguous (e.g. Park, 1969; Williams, 1970, 1985). Consequently, it seems that the attempts to extrapolate crustal scale structures from the outcrop scale data may not be successful without simultaneous consideration of the regional framework - both in terms of rheology and overall tectonic setting (cf. Williams, 1991). To emphasize the local and descriptive nature of the data - and to avoid any restrictions on any possible subsequent revisions or reinterpretations - the numbering of structures used here is unrelated to and independent from the previously published regional schemes, and the possible correlations are discussed separately.

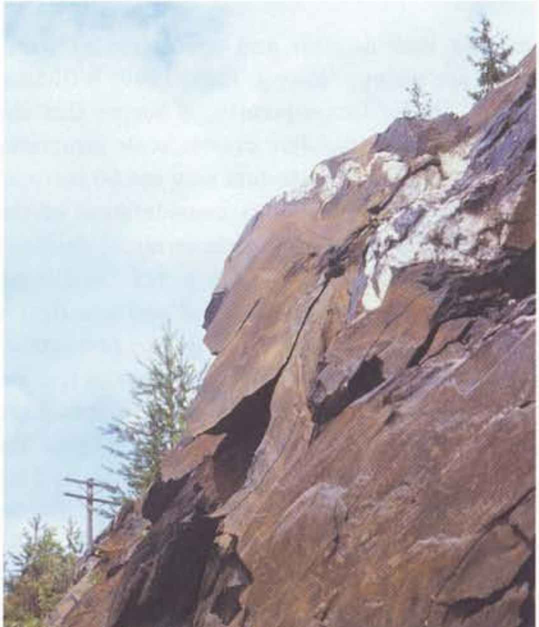
Depositional layering is designated as S_0. Because alternating psammitic and pelitic/semipelitic layers are characteristic of most outcrops, depositional bedding planes are generally obvious; thick psammitic units and some outcrops consisting solely of metapelite of phyllitic appearance form the only notable exeptions. Obvious, reliable depositional younging indicators were unexpectedly not common, precluding a satisfactory regional picture from emerging.

S_1, L_1, F_1 correspond to the first regionally expressed group of deformation structures. However, in some places a foliation classified as S_1 truncates an earlier segregation banding or a folded foliation was observed in an F_1 hinge. These rare features are collectively referred to as **pre-S_1 foliations**. Some of the locally abundant quartz veins also preceded the major (F_1) folding.

F_1 folds were only rarely found. This might be partly due to relative scarcity of vertical exposures suitable for making observations, since gently plunging folds are not easy to identify on horizontal outcrop surfaces. Strong transposition and shearing are also evident and may have obscured the fold hinges. Nevertheless, all the observations indicate tight, asymmetric, eastwards inclined and slightly overturned folding around a subhorizontal axis (Fig. 27).

The **S_1 foliation** is the most pronounced fabric in the mica schists. A distinct, continuous cleavage is characteristic of the pelitic schists. The semipelitic rocks also show a slaty cleavage but thin banding defined by alternating micaceous and quartz-rich folia is more typical. This banding is not always easy to distinguish from primary lamination (Fig. 28). In the psammitic schists the intensity of S_1 varies from a weak preferred orientation of micas to penetrative mylonitic foliation. The most characteristic features are closely spaced layer-silicate seams defining the foliation between thicker domains rich in quartz and feldspar. The S_1 foliation is typically par-

A)



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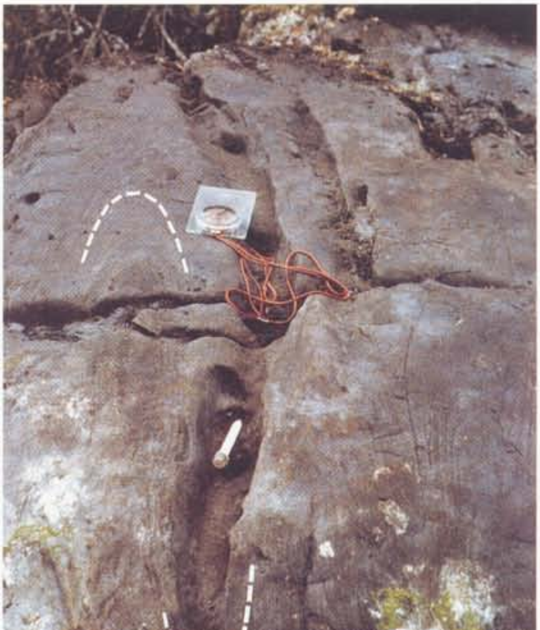


Fig. 27. (A) Parasitic upper limb F_1 structures deforming the bedding (top left). Kalliojärvi roadcut (for location see Fig. 3) viewed towards the north. (B) Tight F_1 fold in a psammitic schist; antiformal closure left of the compass and synformal closure bottom center. Pencil indicates the plunge of the fold axis. Viewed towards the south. Footwall of the Koivenlampi thrust (for location see Fig. 6).

A) B)

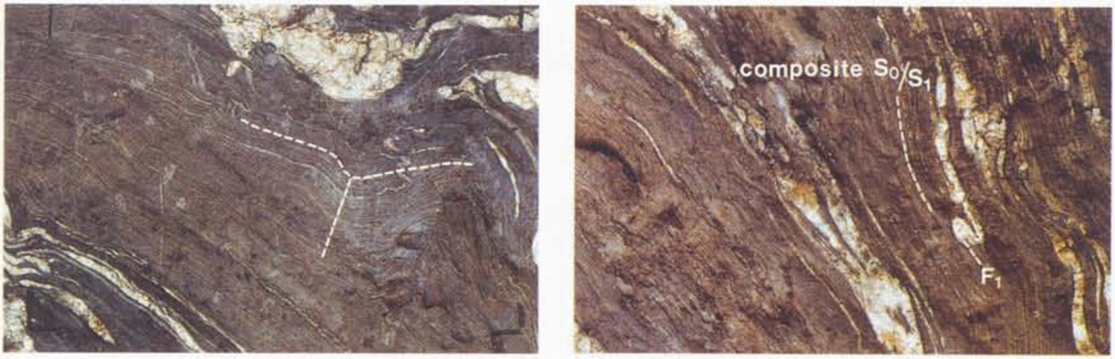


Fig. 28. (A) Isoclinal F_1 closure in a quartz-rich layer (center right). Note also the slight F_2 refolding of the structure. Nunnanlahti (NW of Savikkolankallio), outcrop on the shore of Lake Pielinen. (B) Composite S_0/S_1 banding in axial plane of tightly folded blind quartz vein (center right). Savikkolankallio; see Fig. 3 for Savikkolankallio locality. Both viewed towards the north.

A) B)

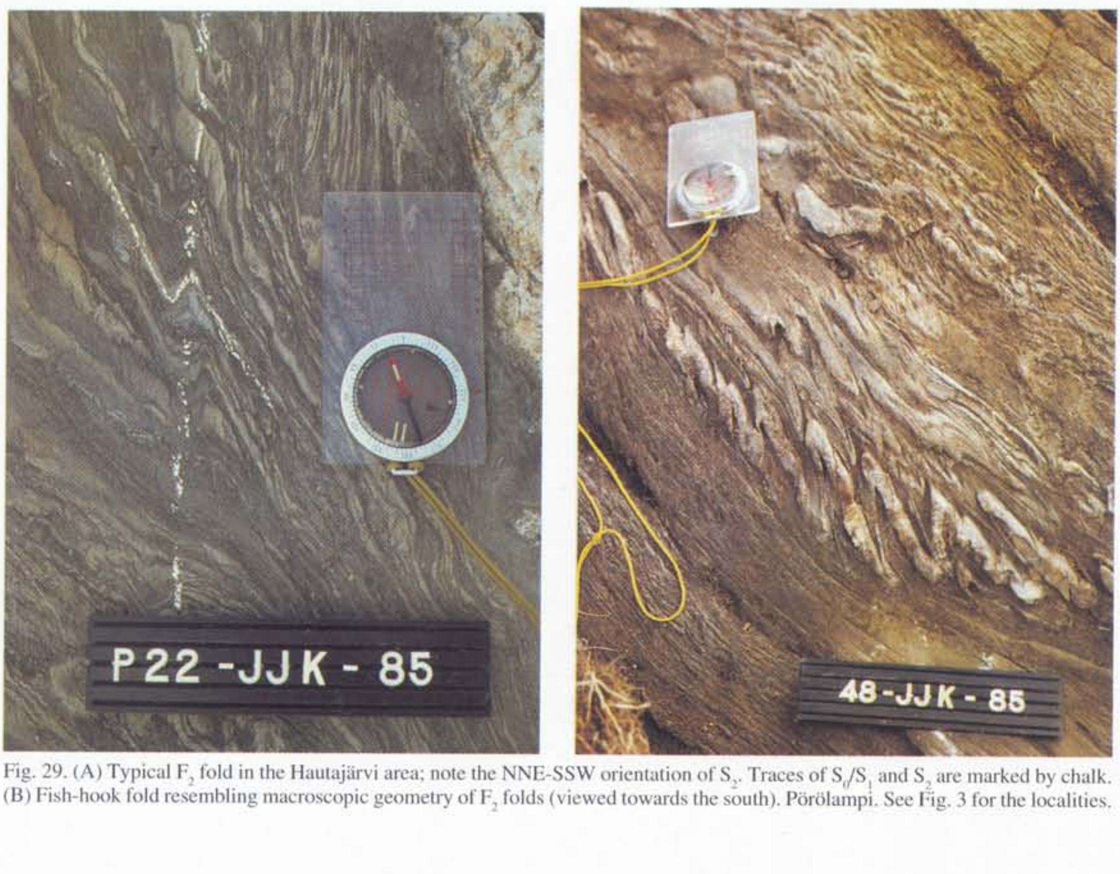


Fig. 29. (A) Typical F_2 fold in the Hautajärvi area; note the NNE-SSW orientation of S_2. Traces of S_0/S_1 and S_2 are marked by chalk. (B) Fish-hook fold resembling macroscopic geometry of F_2 folds (viewed towards the south). Pörölampi. See Fig. 3 for the localities.

allel to or slightly steeper than bedding.

The open to tight folds affecting the S_1 foliation and associated fabrics are defined as **D_2 structures**. The folds almost invariably show a sinistral down-plunge sense, and both the fold axes and the S_1xS_2 intersection lineations typically plunge 30-60° to the S or SW. Displacement along the axial plane resulted locally in a geometry resembling a 'fish hook fold' (Fig. 29). Except for the regional scale Kuhnusta fold (see Figs. 3 and 34 for location), large-scale F_2 structures (such as Valkealampi, as shown in Fig. 30) are not common, although minor folds and/or crenulation microfolds are present in most outcrops. The steep to upright position of the axial plane to F_2 folds reduces the risk of misinterpreting them as geometrically quite similar F_1 lower limb structures.

The **S_2 foliation** is expressed most typically as a nonpenetrative crenulation clevage (Figs. 30 and 31). However, several distinct morphologies were observed; in the psammitic schists spaced, discrete phyllosilicate seams are common, while in the phyllitic schists, S_2 varies in intensity from weak crack-like features to a slaty cleavage. The cleavage is most distinct in fold hinges, but also outcrops lacking macroscopic folds often show a faint pla-

Fig. 30. (A) Sinistral F_2 structures. Varparanta. (B) S_2 foliation (strike indicated by the pencil) perpendicular to the composite S_0/S_1. Hinge of the Kuhnusta fold. (C) Nearly symmetrical F_2 structure in hinge of the Valkealampi fold at Rasivaara. (D) S_2 crenulation cleavage forming axial plane of sinistral F_2 fold. Valkealampi. Tag is 16 cm long. All localities are indicated in Fig. 3.

nar feature cutting S_1 with a sinistral sense. In the major shear zones S_2 becomes nearly parallel to S_1 and distinction between the two in the field therefore becomes obscure. Both the S_1 and the S_2 foliations are defined by alignment of biotite, sericite and quartz-feldspar aggregates and no evidence for contrasting metamorphic conditions was found.

The **post-D_2 structures** are present only locally and expressed as kink bands and weak crenulations. The succession of these structures is, due to a lack of convincing overprinting relations, not possible to determine unequivocally. Three types of structures can be described. Most striking are the kink zones (Fig. 32), which are often spatially related to rel-

A) B)

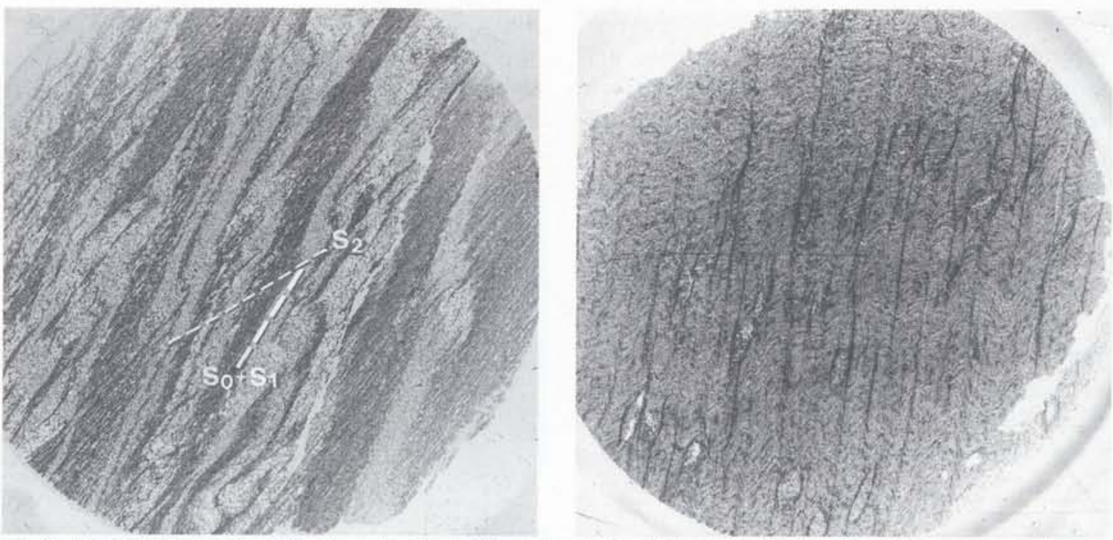


Fig. 31. (A) Typical composite S_0/S_1 - S_2 relationship in thin section. Orinlouhi, (SE of Savikkolankallio. (B) S_2 crenulation cleavage almost perpendicular to S_1 from hinge of the Kuhnusta fold; note also the lensoid microlithons between the discrete crenulation seams. Diameter of the drill core is 25 mm in both cases. Locations of Savikkolankallio and Kuhnusta are indicated in Figure 3.

A) B)

Fig. 32. (A) Open, kink-like post-D_2 folds in pelitic schist containing thin interbeds of quartzite (LA1), Orinlouhi (SE of Savikkolankallio). (B) Conjugate system of post-D_2 kink-bands, Nunnanlahti. Compass is 12 cm long. For localities see Figure 3.

atively young, brittle faults that transect all previous structures. Another group consists of microfolds (faint crenulations) on the S_1 foliation planes. In some places a weak crenulation cleavage, dipping steeply towards the east was observed, but more often these local features form conjugates with variable orientations. The last group also includes very open apparently E-W oriented warping of the dominant structures.

Thrusts and shear zones

In the Nunnanlahti area thrusts along WSW dipping planes and involving basement are evident. The folded basement-cover interface also indicates partly ductile behaviour of the basement gneisses. At outcrop scale the granitic gneisses show zonal deformation with a pronounced foliation (Fig. 33) and SSW plunging lineations along the major planes of movement. An elongation lineation forms the dominant fabric within the basement granitoids, whereas conglomerate clasts, even in the basalmost parts of the cover sequence, record oblate finite strains (Kohonen et al., 1989). The Nunnanlahti thrust becomes indistinct towards the south and is difficult to locate precisely within the folded mica schist area. Consequently, the overthrust part of the mica schist is not well defined.

A discrete magnetic and electromagnetic anomaly with a NNW trend (indicated as shear zone in Fig. 34) transects the whole study area. The anomaly zone itself is poorly exposed, but according to the few observations available strongly sheared mica schist and abundant pyrrhotite are characteristic. The shearing is plausibly linked genetically to the D_2 structures and development of the Nunnanlahti-Holinmäki shear zone (see Kohonen et al. 1991).

Fig. 33. Protomylonitic S_1 foliation in a small inlier of granitic basement gneiss. Compass is 12 cm long. Koivenlampi thrust, for location see Fig. 6.

Structural domains and regional correlation

The orientations of the major structural elements in different parts of the study area are presented in Figure 34. The dominant S_1 foliation shows rather uniform NNW-SSE strike over the whole area, but closer inspection reveals slight differences between each of the subareas. Both bedding planes and S_1 have nearly N-S orientations in the south (subarea KAL3) and show a gradual, anticlockwise rotation and steepening towards the north (subareas KAL4 and KAL5).

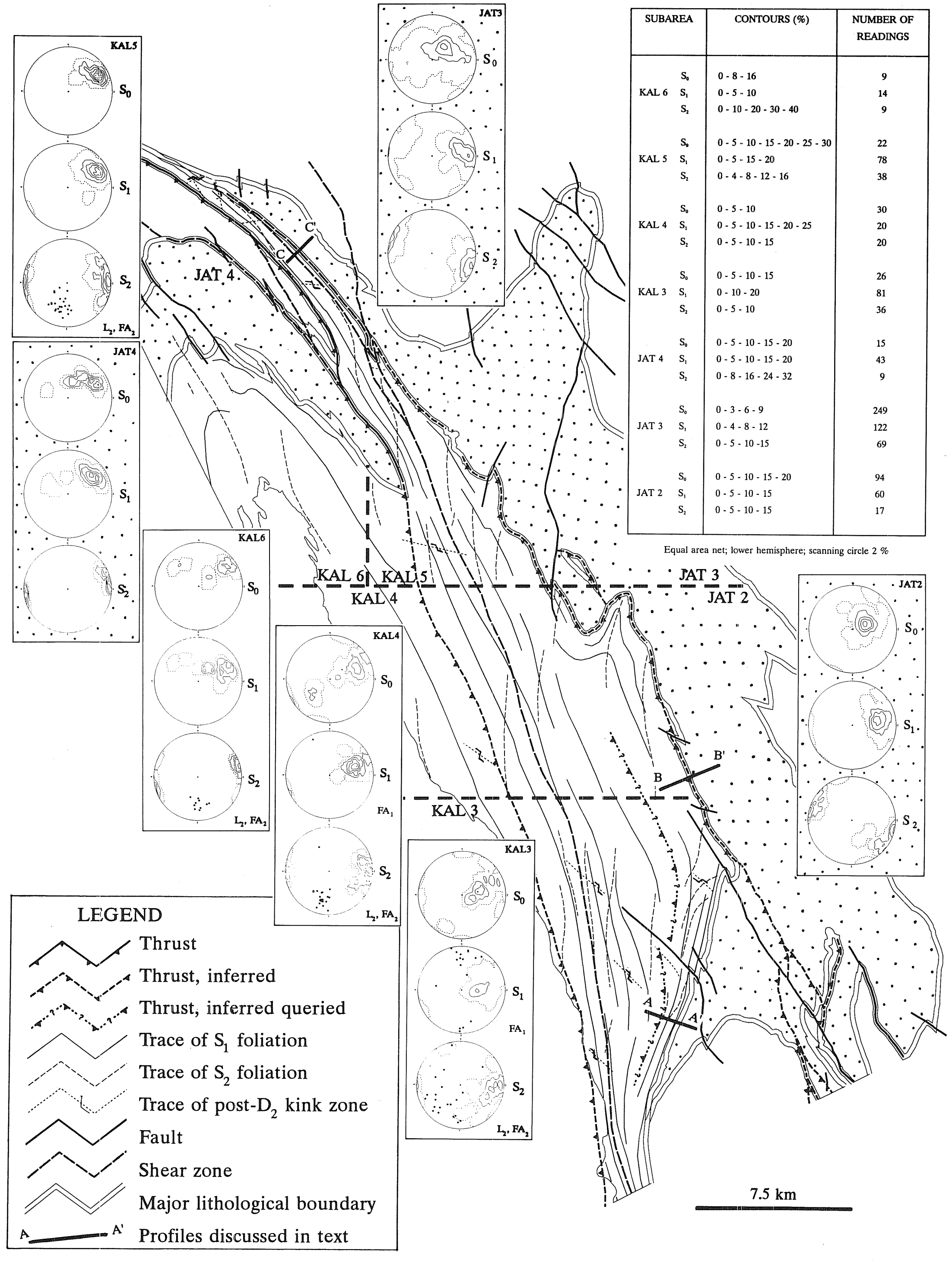
The orientation of S_2 shows also a gradual anticlockwise turning from the NNE trend in the south to N-S (KAL4 and KAL5) and NW-SE (KAL6) strikes in the north. The anastomosing pattern of the S_2 foliation is clearly evident in the equal area diagrams (Fig. 34). The Kuhnusta area (KAL6) represents a regional scale F_2 structure, and a weak maximum indicating NE-SW strikes is present in both the S_0 and S_1 data. In the hinge zone of the regional Kuhnusta fold dextral parasitic folds are also occasionally present.

The postD_2 kink zones show dextral sense and NW-SE orientations in the southern part of the area. Further north the dextral zones have a nearly E-W orientation, and in the northernmost part of the area a conjugate system of dextral and sinistral kinks with ENE-WSW and NNW-SSE orientations, respectively, is present.

In comparison to the adjacent areas, the Höytiäinen province constitutes a quite distinct structural domain, characterized by steep bedding and foliation planes that have a constant NW-SE strike. It therefore seems that the overall style of deformation is comparable within the province, whereas the surrounding domains (e.g. the Outokumpu area) may record a subtantially different tectonic history or, at least and more probably, a different response to predominantly similar crustal stresses. Accordingly, the correlation between the Höytiäinen and Hammaslahti areas appears to be obvious: D_1 and D_2 structures correspond, respectively, to the Group 3 and Group 5 structures of Ward (1987). In their attempt to develop a regional scheme Ward and Kohonen (1989) suggested correlation of these structures with the D_2 and D3 structures described by Koistinen (1981) and Park et al. (1984). The correlation of the local post-D_2 features is nevertheless difficult because their regional expresssion is not significant.

Interpretation of structures

The significance of the pre-S_1 foliations is difficult to evaluate, due to the strong D_1 transposition and scarcity of preserved F_1 hinges. In principle, however, three different explanations may be considered. Firstly, the pre-S_1 structures could record a major deformation preceding the D_1 structures. In the light of the presented regional schemes and evolution models (e.g. Koistinen, 1981; Ward, 1987) this possibility is realistic. On the other hand, the model seems to demand a detachment between the mica schists and the autochthonous cover, because no such early deformation was observed in the quartzites. Alternatively, the pre-S_1 foliations could represent local features developed during the early stages of the D_1 shear folding, and record the early stages of a progressive deformation (Fig. 35). The third - and most simple - explanation is that the S_2 feature has locally been misinterpreted as the S_1 foliation. This is also quite possible, because the intensely



SUBAREA		CONTOURS (%)	NUMBER OF READINGS
KAL 6	S_0	0 - 8 - 16	9
	S_1	0 - 5 - 10	14
	S_2	0 - 10 - 20 - 30 - 40	9
KAL 5	S_0	0 - 5 - 10 - 15 - 20 - 25 - 30	22
	S_1	0 - 5 - 15 - 20	78
	S_2	0 - 4 - 8 - 12 - 16	38
KAL 4	S_0	0 - 5 - 10	30
	S_1	0 - 5 - 10 - 15 - 20 - 25	20
	S_2	0 - 5 - 10 - 15	20
KAL 3	S_0	0 - 5 - 10 - 15	26
	S_1	0 - 10 - 20	81
	S_2	0 - 5 - 10	36
JAT 4	S_0	0 - 5 - 10 - 15 - 20	15
	S_1	0 - 5 - 10 - 15 - 20	43
	S_2	0 - 8 - 16 - 24 - 32	9
JAT 3	S_0	0 - 3 - 6 - 9	249
	S_1	0 - 4 - 8 - 12	122
	S_2	0 - 5 - 10 -15	69
JAT 2	S_0	0 - 5 - 10 - 15 - 20	94
	S_1	0 - 5 - 10 - 15	60
	S_2	0 - 5 - 10 - 15	17

Fig. 34. Structural trends within the study area. Bedding and foliation data of subareas are given as contoured equal area plots and linear elements as heavy dots. Data from adjacent quartzite domains are also included (diagrams with dotted background).

heterogenous strain related to the D_2 deformation has resulted in structures of quite dissimilar appearance.

The asymmetric, eastwards inclined geometry of the F_1 structures suggests that the folding was related to crustal thickening by thrusting. The nearly parallel cleavage and bedding planes indicate tight folds and significant shortening during the deformation. The genetic model for the D_1 structures must explain, at least, the following observations: (1) basement thrusts in the Nunnanlahti district, (2) rapidly waning intensity of basement reactivation towards the east, (3) considerable layer parallel shortening of the cover by folding, (4) contrasting structural style of the mica schists and the eastern quartzites.

In the Nunnanlahti district the deformation displays thick-skinned (the terms 'thick-skinned' and 'thin-skinned' are used in the following for structural interpretations with and without basement involvement, respectively) style: the basement and cover are folded together resulting in a basement-cored thrusted anticline. The seemingly intact Archean gneisses further east might even indicate that the Nunnanlahti zone represents a leading ramp anticline of an imbricate fan involving the Kaavi-Timonvaara-Nunnanlahti basement. However, the style and distribution of Proterozoic reactivation structures of the eastern foreland basement are poorly known, and it is premature to label this particular zone as the ultimate frontal thrust beneath allochthonous basement. Although the intensity of reworking rapidly wanes towards the east of the present study area (cf. Sorjonen-Ward, 1993), evidence from the craton interior in Russian Karelia (e.g. Sokolov et al., 1970; Systra, 1991) indicates local basement ductility over large areas within the foreland.

The imbricated gneisses of the Kaavi-Nunnanlahti region along the strike plausibly provide the best analogy for the structures of the basement underlying the study area. The Nunnanlahti area is clearly an exception - the basement consists largely of ductile greenstones that show structural sequence and geometry identical with that of the cover (Kohonen et al., 1989). Because the basement lithology and the relief of the basement-cover interface beneath the Höytiäinen area is not known, different structural models may be considered.

Thick-skinned models may be condensed

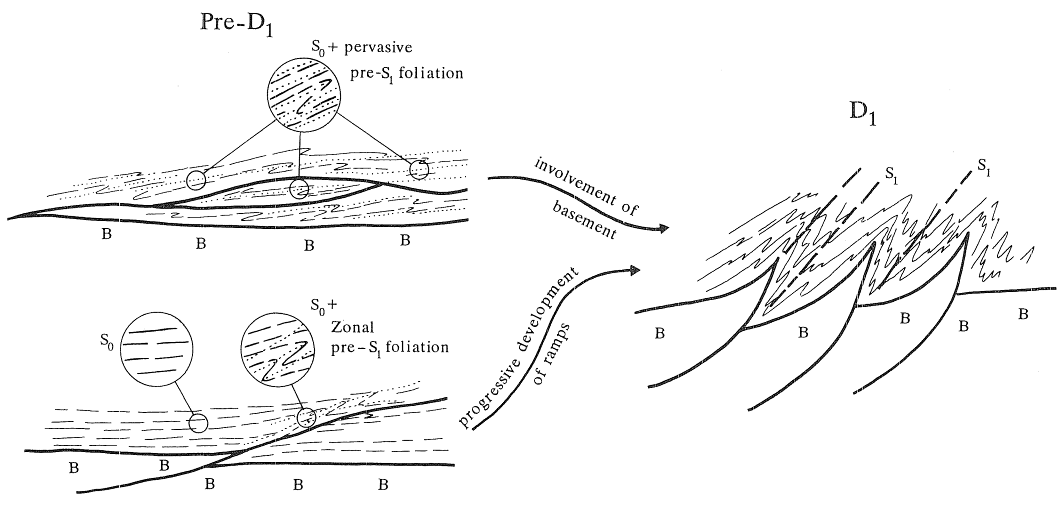


Fig. 35. Two alternative explanations for the pre-S_1 foliations.

into two end members. In Figure 36A a ductile basement is assumed, and major thrusts are not required to produce the observed D_1 geometries. In contrast, if basement is more competent (such as granitoid gneisses) then the steep bedding planes of the cover sequence favour substantial shortening by ramp structures (Fig. 36B).

The Kaavi-Nunnanlahti basement thrusts also limit the usefulness of pure thin-skinned applications for the main part of the study area. However, if buttressing against pre-existing steep normal faults and reactivation of horst structures is assumed, quite a realistic model can be produced without substantial shortening of the basement (Fig. 36C). This kind of model is in accord with the structure suggested for the Sotkuma basement inlier by Kohonen and Elo (1991) and also explains well the contrasting structural styles between the Kaleva mica schists and the quartzites in the east. To conclude, it seems that all the potential modes of shortening presented above were active during D_1 deformation, the importance of each varying from place to place.

The most noticeable D_2 characteristics include: (1) the NNW-SSE striking zones that record heterogenous and variable amounts of strain, (2) the nearly ubiquitous sinistral sense of the minor folds and (3) the regional Kuhnusta fold structure. All these structures can be attributed to sinistral simple shear with a subvertical shear plane. The composite S_1/S_2 foliation in the most ductile micaceous rocks is interpreted as resulting from intense shear along these planes, whereas the more northerly orientations of the S_2 crenulation cleavage reflect strain refraction between the major shear zones (Fig. 34).

The geometry of the Kuhnusta fold is reminiscent of a 'fish hook fold' and gives some clue to the formation of the structure. According to Ramsay and Huber (1987) this kind of fold is best explained by early buckling followed by rotational shearing. Within the overall structural framework of the study area this could mean that the cover fold initiated due to a locally increased strain perturbation caused by ductile (F_1) shortening of the adjacent greenstone basement. The final sinistral geometry then resulted in a shear zone corresponding to the overall nature of the F_2 folding. An important consequence of this model is that the effects of F_2 folding are expected to wane rapidly towards the south from the Kuhnusta hinge zone, so that the tight fold closures in the northern end of the lake Höytiäinen are actually in-

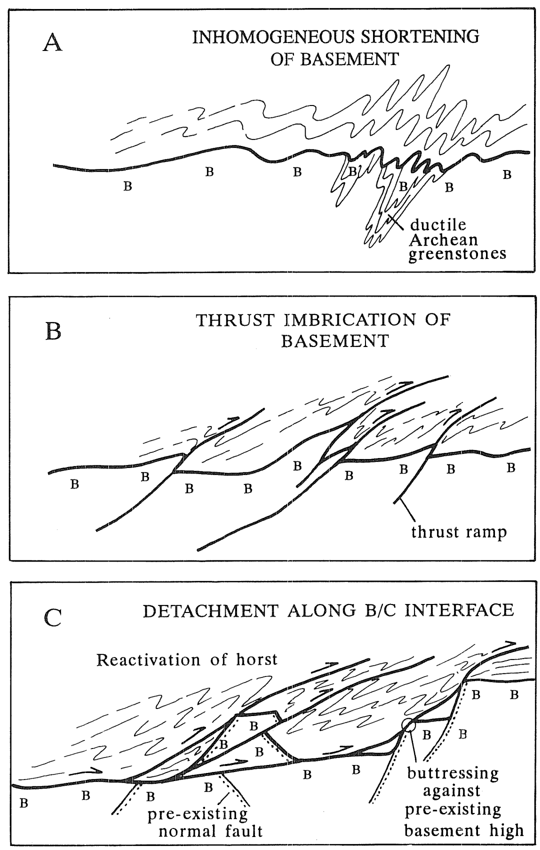


Fig. 36. Three schematic models for basement-cover relationships during shortening of the Höytiäinen province. (A) Inhomogenous buckle fold shortening due to the rheological contrasts within the basement. (B) Thrust ramp structures within relatively competent basement and ductile deformation of the cover. (C) Decollement along the B/C interface. Thrust activation of previous horst structures and localized high strain due to impedance by pre-existing basement uplifts.

terpreted as F_1, rather than F_2 structures.

In summary, the D_1 deformation of the cover mica schist may have formed in response to combined thrust imbrication and buckling of the underlying basement. The steep attitude of the basement-cover interface in the Nunnanlahti area appears to make a thin-skinned interpretation less tenable. The interpretation of the pre-D_1 structures is clearly dependent of the overall model; thrusts involving basement either followed nappe movements in time and space or developed progressively deforming local initial irregular features. The alternative deformation paths from easterly vergent thrusts to the sinistral structures have been discussed in previous papers (Ward and Kohonen, 1989; Kohonen et al., 1991). Within the present study area all features observed fit well with a model invoking the progressive development of D_1 and D_2 structures. However, it seems that confident correlations between the D_2 shear zones of the present study and major crustal scale shear systems proposed by Kärki et al. (1993) must wait for more detailed mapping of kinematic indicators in eastern and northern Finland.

Structural interpretation of the eastern boundary of the Kaleva mica schists

Current models of foreland deformation emphasize the role of pre-existing structures, and this opens new perspectives in interpretations. The styles of thrust reactivation of normal faults have been described by several authors (e.g. Kulik and Schmidt, 1988; McClay and Buchanan; 1992), and similar principles have been applied in construction of Figure 37 (for location of the profiles see Fig. 34).

The sharp boundary between the eastern platformal quartzites and the Kaleva mica schist is interpreted to represent the previous locus of a major synsedimentary normal fault (growth fault) zone near the edge of the Kaleva basin. The contact separates two distinctly different lithological provinces and, consequently, two domains having different deformation style in response to tectonic stresses. This kind of pre-existing fault, which may also form a rheological interface, clearly has potential to become intensely modified in course of the compressional deformation.

Within the present study area the major part of the mica schist/quartzite contact is interpreted as having subsequently been modified by major thrusts (Fig. 34). The direct field evidence for this is sparse, but the northern branch of the thrust along the eastern contact of the Kontiolahti basement inlier apparently converges into the contact in the Hautajärvi region (Fig. 34). Further north the brecciated quartzites and block breccias could be attributed to intense compressional buttressing and shear along the previous normal fault system presently represented here by the contact itself (Profile B-B' in Fig. 37). In contrast, the gently dipping and weakly deformed mica schists near the contact within the area south of Hautajärvi are interpreted to represent the footwall of a thrust ramp (Profile A-A' in Fig. 37) plausibly reactivating one of more western growth faults. In the northernmost part of the study area the distinct, linear aerogeophysical anomaly interpreted as a D_2 shear zone (see Fig. 34) coincides with the contact, and the straightness of the contact within the area north of Ahmovaara probably reflects final modification by these strike-slip movements (block model/profile C-C' in Fig. 37).

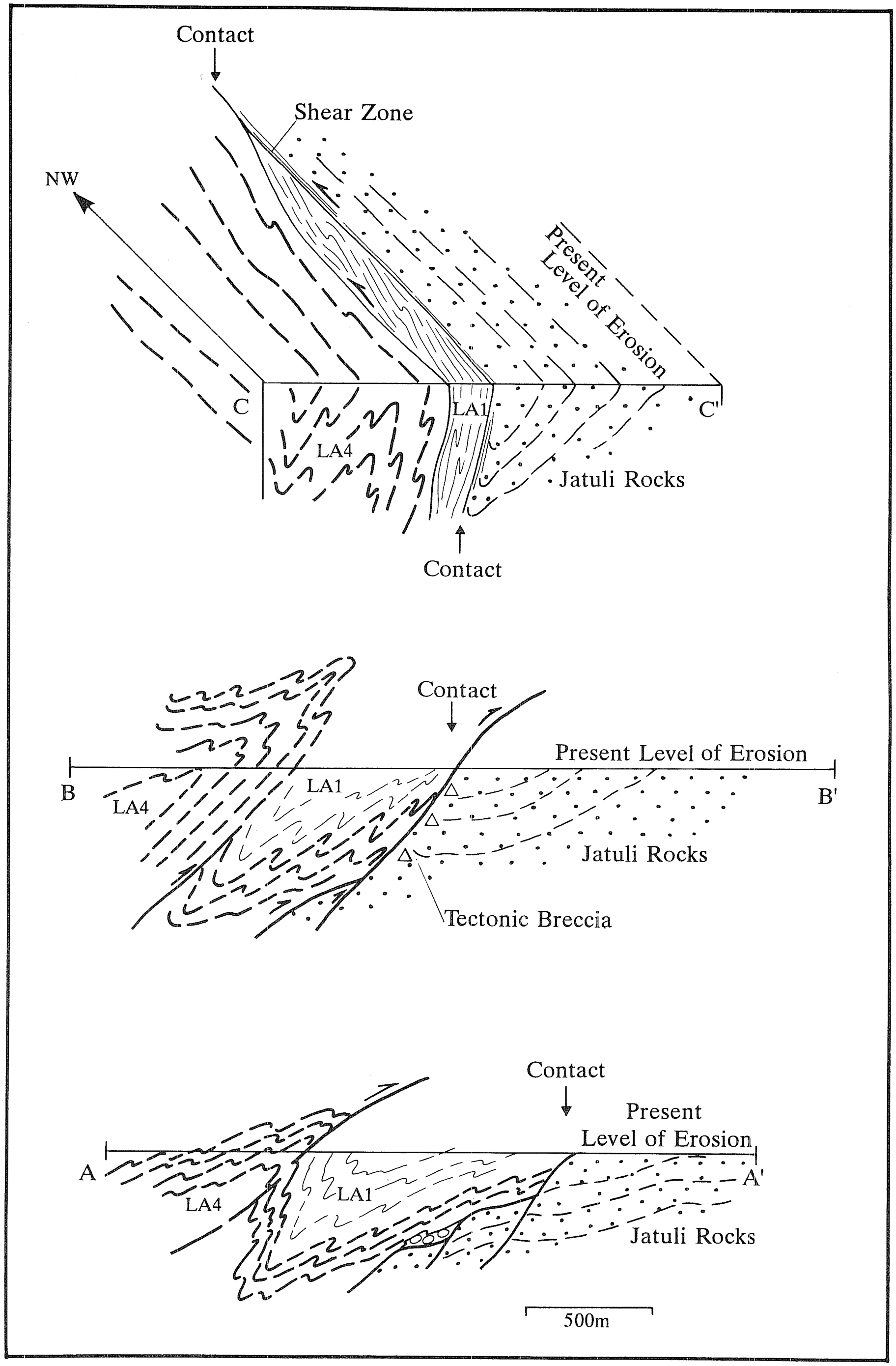


Fig. 37. Structural interpretation of the contact zone between the Jatuli and Kaleva rocks (modified after Kohonen et al. 1990). For regional structure and location of the profiles see Figure 34.

EVOLUTIONARY MODELS

At first, some clarification of the approach adopted here is perhaps needed. The paucity of depositional younging determinations, together with the poorly defined location and displacement magnitude of thrusts, means that the original relative stratigraphical relationships between the observed lithological assemblages are still partly equivocal. Consequently, it is not strictly justified to apply standard lithostratigraphic procedures resulting in columns showing the superposition of formally named units. Modern stratigraphy, however, emphasizes the importance of a three dimensional depository model in the interpretation of local vertical sequences (e.g. Miall, 1984; Galloway, 1989; Laajoki, 1991) instead of a formal, straight line correlation of similar lithologies. In the following synthesis the observed sequences, the few available geochronological data, and stratigraphic deductions derived from structural interpretations are integrated to define a coherent system within a presumed rifted craton margin setting.

The basis of the interpretation is the rift basin model and obviously, if this assumption can be proven incorrect, the ensuing interpretation is not necessarily valid anymore. Nevertheless, the presence of Archean crust beneath the Höytiäinen basin seems very probable and, on the other hand, the abundant mafic dykes with ages from 2.1 to 1.97 Ga cutting the cratonic sequence are difficult to interpret without invoking major extensional movements. In order to emphasize the equivocal nature of the approach selected, however, a complementary model assuming a more allochthonous nature for some of the units is also outlined.

Geochronological constraints

Within the study area the mafic dykes apparently stop abruptly at the eastern contact of the mica schists, which probably indicates the existence of a local unconformity younger than 1,97 Ga. However, the observations from Ahmolampi (the mafic apophyse in the mica schist) and Koivenlampi (the metadiabase cutting the mica schist-arkosite sequence) raise questions concerning the age of the Kaleva mica schists in general. A similar paradox related to the Tohmajärvi gabbro and associated volcanics at Hammaslahti has already been discussed in the introductory section.

In the western Outokumpu area the Horsmanaho gabbro, which intrudes an ultramafic (serpentinite) body, yields an age of 1972 Ma (U-Pb, zircon; Koistinen, 1981; Huhma, 1986). The serpentinites occur as tectonic lenses in the 'Upper Kalevian' (cf. Kontinen, 1987; Kontinen and Sorjonen-Ward, 1991) monotonous graywackes. These ophiolitic fragments of oceanic lithosphere (Koistinen, 1981; Vuollo and Piirainen, 1989) seemingly provide a constraint on timing of the advanced stages of extensional evolution. The recent detrital zircon age of 1920 Ma from the Outokumpu area ('Upper Kalevian' graywacke from Vuonos mine; Claesson et al., 1993) here constrains the maximum age for deposition. Thus, it is possible, in principle at least, that the oceanic-type crust locally formed the depositional base for the western 'Upper Kalevian' mica schists. The 1920 Ma age also gives the maximum age for the deformation, whereas the weakly deformed Maarianvaara granite (1857 Ma; Huhma, 1986) has been referred to as 'late orogenic' (Huhma, 1975) or 'late tectonic' (Park and Bowes, 1983).

A depositional model for the study area

Several proximal type conglomerates occur along the contact separating the mica schist and the quartzite, but their **stratigraphic** location and significance cannot be determined unambiguously in the field. Because the contact fault seems to have an attitude steeper than the bedding in the eastern quartzites (Kohonen et al., 1990), the conglomerates are not necessarily 'basal' with respect to the Kaleva mica schist but may represent different stratigraphic levels and stages of basin development. The proposed relationships between the lithological assemblages are diagrammatically presented in Figure 38. The model clearly implies that the easternmost Kaleva rocks - presently in contact with the platformal quartzites - do not represent the lowermost part of the mica schist unit.

Although the lower contact of the Kaleva mica schist is not possible to define in the study area, it is suggested that the lowermost graywackes are mainly underlain by coarse clastics and volcanics with an age of about 2.1 Ga and partly deposited unconformably

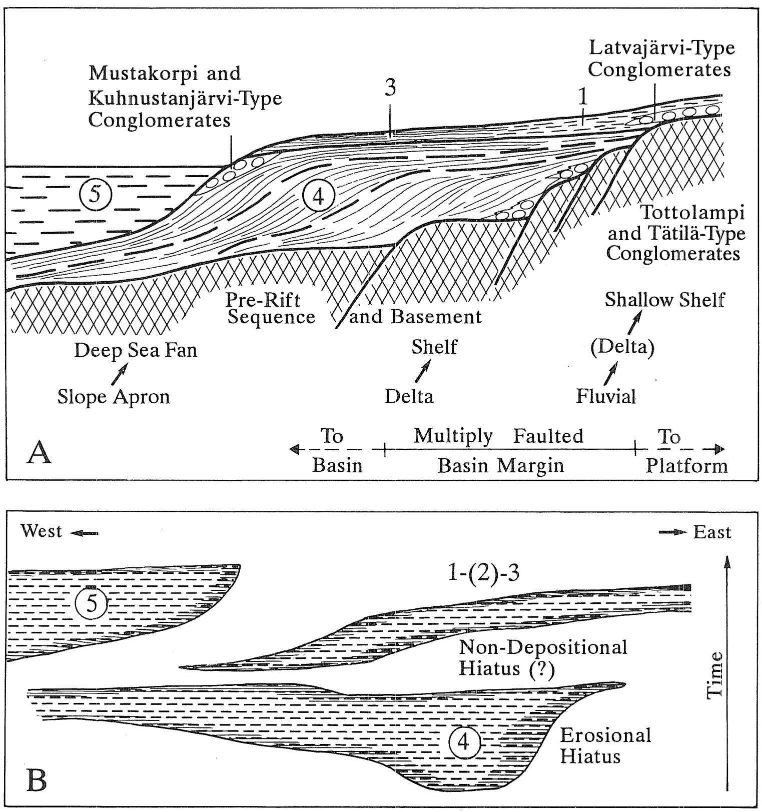


Fig. 38. Diagrammatic depositional model for the study area. (A) Depositional pattern in different parts of the basin (vertical exaggeration; not to scale). (B) Time-space diagram for the study area. Time-scale strictly schematic. Numbers in both figures correspond to specific lithological assemblages.

on still older platformal cover or even directly on the Archean basement. This surmise is based on a number of lines of evidence. Within the Koivenlampi-Saarijärvi area (part of the Nunnanlahti-Kuhnusta area; see Chapter 2), the mica schists are separated from the basement gneisses by a thin unit of arkosic conglomerates. The mica schist contains arkosic interbeds and therefore conceivably has a gradational transition into the coarse clastics. Rather similar basement cover relationships have been reported from around the Sotkuma dome (Frosterus and Wilkman, 1920; Väyrynen, 1939; Mattila, 1971; Huhma, 1975), and on the southern flank of the Kontiolahti dome the arkosic mica schist rests directly on the basement granitoids (Väyrynen, 1939). As suggested by Ward (1987, 1988) the coarse grained sediments, mafic volcanics and gabbros of the Tohmajärvi region (for descriptions see Pokki, 1965; Nykänen, 1968; Nykänen, 1992) may reasonably represent lower basin fill within the axial zone of the Höytiäinen Province.

The graywackes and mudstones forming LA 4 are considered here to form the lowermost division within the formations studied. These types of deposits occupy areas interpreted as antiformal structures and are in places (Ahmovaara and Koivenlampi localities) truncated by mafic intrusions of unknown age. The variable nature of lithotypes within the assemblage most probably reflects different depositional regimes (such as water depth, proximity to supplying fluvial deltas),but in the absence of additional knowledge of basin geometry, a more detailed interpretation is premature.

The depositional setting of the phyllite-quartzite assemblage (LA 1) and the phyllites of LA 3 in Fig. 38 is based on structural arguments and the observations indicating depositional age younger than 1.97 Ga; LA 1 is in several places in mutual contact with diabases of the 1.97 Ga age group (T2 diabases of Vuollo et al., 1992), but intrusive or contact metamorphic features are never seen. The sharp-bounded orthoquartzite interbeds (see Hautajärvi lithotype) strongly support a contemporary mature shelf as the immediate up-slope source. The observations from Ahmolampi 1 and Piili (LA 1 schist interbedded with quartzite) indicate that some quartzites in proximity to the contact zone could be considerably younger in age than the eastern quartzites intruded by the mafic dykes. The area occupied by LA 2 forms a fault-bounded lens, but the deposition could be coeval with the former assemblages or, alternatively, represent a remnant of cover older than the main stage of the Kaleva deposition.

The rocks of the LA 5 more closely resemble the metasediments of the Outokumpu area in depositional style and composition than the other rocks within the study area, and two alternative interpretations, at least, may be considered. In Fig. 38 these rocks are assumed to represent deep marine facies of a stepwise westwards deepening continental margin. In this view the Kuhnustanjärvi and Mustakorpi conglomerates could represent turbidite channel fills near the hiatal boundary between the shelf and the slope.

Another explanation for LA 5 is that the whole unit forms a klippe isolated from the thrust sheet represented by the allochthonous Outokumpu assemblage in the west (Fig. 39); in this case the conglomerates at the eastern margin of LA 5 plausibly represent relatively local detritus from the western margin of the intracratonic Höytiäinen basin (cf. Ward, 1988). It is to be noted, however, that even the former model does not exclude a temporal connection between the Outokumpu metagraywackes and the rocks of LA 5.

A model for the Kaleva sediments of the North Karelia Schist Belt

An important consequence of the supposed multiply rifted craton margin setting of the NKSB is that the subareas within the belt possibly record uplift-subsidence histories that differed substantially from each other throughout the course of basin evolution. This is clearly indicated by the presence of very different kinds of deposits - both in age and depositional setting - directly on top of the Archean. Block movements probably resulted in a complex basin configuration with escarpments separating areas of erosion and deposition. Hence, depositional style and source material may vary, even within a relatively small area, especially in the proximity of a marginal fault. A basin model based on lithostratigraphic correlation i.e. on the assumption that lithologically similar units represent even roughly same period of geological time may lead to fundamentally flawed conclusions.

An alternative 'space and time' approach is applied in construction of the Kaleva model in the North Karelia region. The local sequences form the starting point, but the correlation has been attempted using the available geochronological information and spatial distribution of formations. Unfortunately, reliable age determinations from the NKSB are quite few, and future work will hopefully refine probably reveal furher details concerning the extensional evolution. The mafic magmatic pulses at about 2.1 and 1.97 Ga ago (Vuollo et al., 1992) has been used as major rift event markers in the present evolution model (Fig. 39). It is emphasized, however, that the two stage rifting model is a simplification since ages of around 2060 Ma have been reported from eastern Finland (Koivusaari Fm., Pekkarinen and Lukkarinen, 1991). Following assumptions have been used in compilation of Fig. 39: (1) The Fe-tholeiitic dykes cutting the quartzitic Puso Formation in the Koli area (cf. Vuollo et al., 1992), the mafic dykes at the eastern margin of the Sotkuma inlier, and the amphibolites in the Juuanvaarat area are taken to represent the 2.1 Ga age group. (2) The suggested contemporaneity of the Hammaslahti-Tohmajärvi mafic intercalations and the Tohmajärvi volcanic complex with age about 2.1 Ga (see Ward, 1988) is accepted. (3) The schists of the Outokumpu area and associated ophiolite bodies (Fig. 2) are considered as an allochthonous unit (see Koistinen, 1981; Gaál, 1990). The question as to whether the root zone of the nappe is located along the Jormua-Kaavi zone (e.g. Park et al., 1984; Koistinen, 1986; Kontinen, 1987) or at the craton margin further west (Ward, 1987; Sorjonen-Ward, 1991) is not crucial from the present point of view.

The main corollary of the present model is that the contact zone between the Jatuli assemblage and the Kaleva mica schists was originally and primarily **a first-order boundary between different depositional regimes**. Local unconformities and hiatal gaps are characteristic of sequences deposited along a repeatedly active basin margin. A further point to note is that the precursors of the Sotkuma and Maariánvaara inliers are interpreted to have been initiated as basement horsts during extensional basin development and also supplied detritus to the basin. This is in accord with the ideas presented by Ward (1987, 1988). In the schemes of Ward (1987) and Park & Doody (1991) the Suhmura thrust zone is of fundamental importance in separating the western allochthonous Savo province from the autochthonous Höytiäinen province. In the present model part of the Savo province graywackes (e.g. flat lying sediments between the Sotkuma and Oravisalo domes) may have been deposited on top of exhumed Sotkuma horst block. Previously the 'Upper Kalevian' type graywackes have been reported as directly overlying basement in at least two localities (Niinivaara in Kaavi; Park, 1988 and Oravisalo in Rääkkylä; Ward, 1988). If some of the western schists are parautochthonous, the LA 5 may be connected to those as well as to the allochthonous Outokumpu graywackes. The triangle shaped horst block origin for the Sotkuma inlier (Fig. 39) is supported by geophysical studies indicating an abrupt termination of the basement high on the northern side of Sotkuma along the strike of the inferred Suhmura thrust zone (Kohonen and Elo, 1991).

In the Kiihtelysvaara region the conglomerates of the Kortevaara Formation (Pekkarinen and Lukkarinen, 1991) are here correlated, not with the Suhmura coarse clastics (Ward 1988), but rather with deposition of the LA 1 in the present study area, and are hence considered as younger than the major part of the Höytiäinen area graywackes. This interpretation, although tentative, nevertheless obviates the problems related to the intrusive relationships of mafic dykes of different age groups. On the other hand, the Tottolampi conglomerate and the associated purplish quartzite might represent

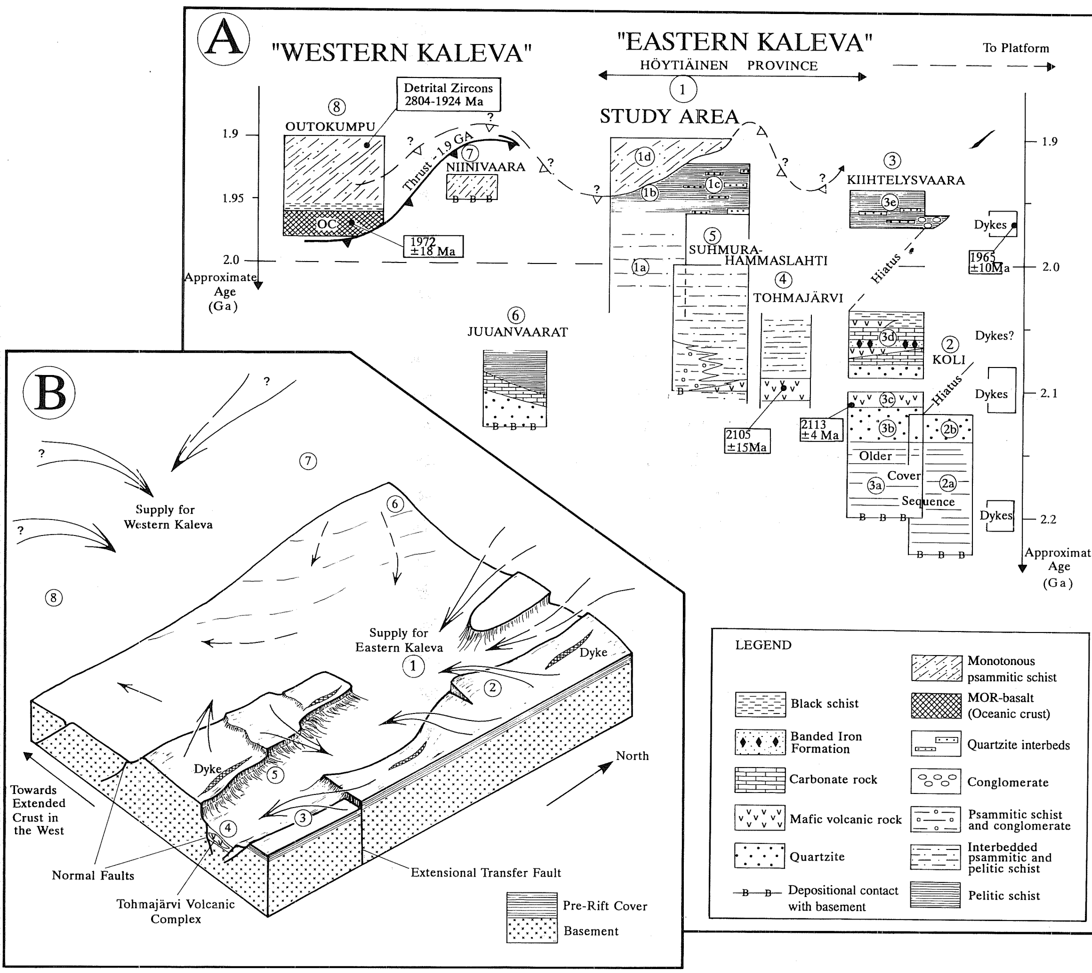


Fig. 39. Depositional model for the North Karelia Schist Belt. (A) Space and time diagram for supracrustal sequences. 1a= LA4, 1b= LA3, 1c= LA1, 1d= LA5, 2a= Kyykkä Group and lower part of the Herajärvi Group, 2b= Puso Fm., 3a= Viesimo Group, 3b= Haukilampi Fm., 3c= Koljola Fm., 3d= Hyypiä Group, 3e= Kortevaara and Heinävaara Fms.. Data from Kohonen & Marmo, 1992 (2); Pekkarinen & Lukkarinen, 1991 (3); Ward, 1987,1988, Nykänen, 1968,1971b (4); Ward, 1988 (5); Park, 1988 (7); Loukola-Ruskeeniemi et al., 1991 (8). All sections simplified and reinterpreted by the present author. Temporal relationships given to emphasize the concept of diachronous deposition; time scale only schematic. (B) Basin model for the eastern part of the NKSB. Approximate locations of sections presented in A are also indicated. Not to scale, for locality names see Figure 2.

deposition corresponding to the Kalkunmäki Formation (the lowermost part of the Hyypiä Group, Fig. 39) in the Kiihtelysvaara area.

The Juuanvaarat area between Kaavi and Nunnanlahti is the only place within the NKSB where the mica schist conformably, without an intervening conglomerate, overlies the platformal quartzite-carbonate sequence. The deposition is attributed to the development of half-grabens at about 2.1 Ga and plausibly represents platform-type, low-energy conditions on the shallow flanks of the asymmetric basin. No age data are available, but correlations based on the results of Karhu (1993) are not in conflict with the assumed depositional age.

Because the depositional ages of the studied lithological assemblages and other Kaleva subunits are not firmly established, the author feels that nomenclature implying a sequential relationship (such as upper or lower) should at present be avoided. It is suggested instead that the Höytiäinen area mica schists (and possible correlatives), deposited in marine basins of rift origin at the evolving continental margin could be informally referred to as the Eastern Kaleva (or Eastern Kaleva tectofacies). In North Karelia this unit largely corresponds to the 'Höytiäinen province' of Ward (1987). Within the present study area the Eastern Kaleva can be tentatively divided further into the syn-rift (LA 4) and post-rift (passive margin and/or foredeep sequences of LA 1, LA 2 and LA 3) assemblages. Typical features of the former group are hiatal proximal facies conglomerates and overall variable and immature lithology with arkoses, arkosic wackes, graywackes and mudstones. The latter group is in contrast characterized by mature detritus probably derived from the platform in further east. An overall Karelian Craton (Archean) provenance with contribution from rift related mafics is proposed.

The western mica schist (roughly corresponding to the 'Savo Province' of Ward, 1987; the 'Lower Kalevian' of Gaál, 1990 and the 'Upper Kalevian' of Kontinen and Sorjonen-Ward, 1991) presently enclosing the ophiolites is, accordingly, named as the Western Kaleva tectofacies. Within the NKSB the typical features for the Western Kaleva include monotonous depositional style and minor compositional variability of psammites, black schist intercalations and detritus as young as 1.92 Ga. Deposition from gravity flows in a deep marine fan environment has been suggested (Ward, 1988; Kontinen and Sorjonen-Ward, 1991). It seems not impossible that the basin was, at least partly, underlain by newly formed oceanic crust.

Crustal shortening stage; tectonic basin inversion

According to modern tectonic models, basin configuration and thrusting are not independent but intimately related to each other. Foreland thrust belts develop in most cases by progressive restacking and thickening of a rifted margin. Rejuvenation of old normal faults is an important process during at least the early stages of the shortening and inversion.

In North Karelia the arc-collage/continent collision about 1.9 Ga ago resulted in development of an easterly vergent foreland fold and thrust belt characterized by cover nappes and basement thrust ramps (Fig. 40). On a crustal scale, the cover everywhere in North Karelia forms only a relatively thin veneer (about 5-7 km in maximum; see Kohonen and Elo, 1991) on a thick (50-60 km; Luosto et al. 1990) Archean crust. The upper greenschist - amphibolite facies metamorphism of the parautochthonous cover (Campbell et al., 1979 and references therein) indicates, however, that the present erosion level of the NKSB represents the relatively deep levels of the collisional thrust belt. This view is supported by the widely reset, Proterozoic K-Ar ages of biotite and hornblende from the Archean domain in the eastern North

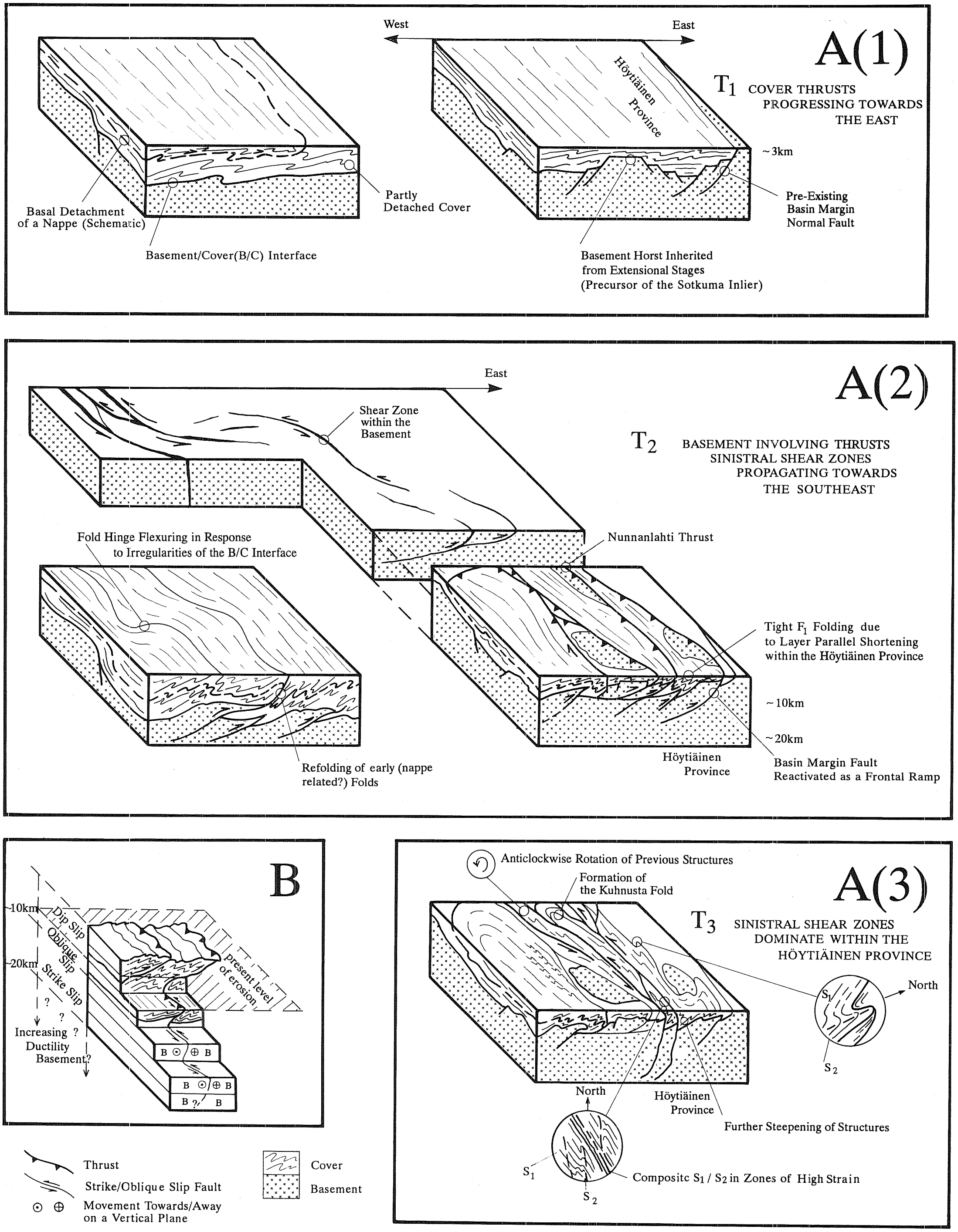


Fig. 40. (A 1-3) Three stage model for the structural evolution of the study area during collisional crustal shortening. Note the assumed diachronous development of structures. (B) Speculative depiction of contrasting styles of deformation at different structural levels.

Karelia (Kontinen et al., 1992).

Nevertheless, in terms of the basement structures the eastern margin of the present study area represents an important boundary. In the west imbricated slices of basement are characteristic (Park and Bowes, 1983; Park and Doody, 1991; Koistinen, 1993), whereas the eastern basement seems to have remained nearly intact. Concerning the present problem, this observation has fundamental consequences. The reverse reactivation of the westwards dipping basin margin normal fault led to development of a **younger-over-older** thrust along the mica schist/quartzite contact. It should still be noted that the presented explanation for the contact includes all the previous models (unconformity, gradational contact, thrust) as special cases.

The absence of the pre-rift sequences around the basement inliers at Nunnanlahti, Sotkuma and Kontiolahti indicate that those areas were 'positive' (located above the baselevel) during the early stages of the rift basin evolution. On the other hand, all the inliers show major reactivation and protomylonitic foliations along their eastern margins. Thus, it seems that easterly dipping horst margin faults having a 'non-ideal' orientations were also rotated and reutilised as thrusts (Fig. 40). A geometrically similar rotational reactivation of a basement-involving normal fault has been described from the Rocky Mountains by McClay et al. (1989).

An important consequence of the suggested inversional tectonic style is that the thrust ramp structures involving basement (basement relief) can **not** be used in estimation of the amount of crustal shortening or in construction of balanced sections. The Kontiolahti and Sotkuma basement inliers (see Fig. 2), for example, may have remained as basement highs since the early stages of the basin evolution. Also the apparently large amount of cover shortening may be resolved neatly by assuming inversion of a horst and graben system within the basement (e.g. Etheridge, 1986).

The kinematic evolution of the thrust belt was probably controlled by pre-existing basin margin irregularities (e.g. extensional transfer faults), basement topography and by differences in basement rheology (cf. Kohonen et al., 1991). The nappe movements at higher structural levels and progressively increasing vertical stresses at deeper levels might explain the gradual transition towards the development of upright structures. If the sinistral shear zone delineating the basement blocks gradually propagated southwards and was more or less continuously detached from the overlying cover, the present orientations of the cover structures would reflect the sum effect of cover thrusting and rotation caused by the block movements within the basement (Fig. 40). Accordingly, the present orientation of the cover structures may be of restricted value in the reconstruction of the regional paleostresses. The final complex architecture of the NKSB seems to display both features generated or modified during the shortening stage and those inherited from early extensional basin evolution stages.

SUMMARY AND TECTONIC SETTING

The Kaleva mica schist within the NKSB includes deposits with apparent ages ranging from 2.1 Ga (Tohmajärvi district) to younger than 1.92 Ga (Outokumpu district). Consequently, it is quite obvious that the Kaleva as a whole cannot be understood as the result of a single geological event or process but is a diverse sequence representing a long and complex geological history. Regarding the Eastern Kaleva, the results of the present study are in accord with the overall Höytiäinen rift basin scheme of Ward (1987). The studied lithological assemblages 1-4 include sequences interpreted as syn-rift turbidites and post-rift marine (shelf?) deposits. The sharp mica schist/quartzite contact in the east is understood to represent the locus of one of the original marginal faults bounding the Kaleva basin. The boulder conglomerates along the contact zone indicate proximal deposition adjacent to the paleoescarpment. It seems that the various conglomerates do not necessarily occupy the same lithostratigraphic positions nor need they be coeval with each other, but primarily indicate the general **position** of a multiply activated basin margin. Accordingly, similar deposits are expected to occur only in a narrow zone along the strike of the corresponding structures. Therefore, the potential for regionally successful lithostratigraphic correlation based on these conglomerates is low.

The geochemical studies suggest a cratonic source for the major part of the rocks studied. The granitoid fragments in the conglomerates also imply that the basement was exposed (as horsts blocks) during the rifting and supplied detritus to the Kaleva basin. Local basement exhumation within the Höytiäinen Province is supported by the observation that in some places the syn-rift sequence directly overlies Archean rocks. Unit LA5 seems to have a different provenance from the other assemblages and shows geochemical features similar to the Western Kaleva rocks.

Ward (1987) located the boundary between the Savo Province and the Höytiäinen Province along the 'Suhmura Thrust' (see Fig. 2). However, the boundary between the Eastern and Western Kaleva appears to be interfingering rather than sharp. The area north of the Sotkuma inlier is complicated by tectonic slices, and the rocks of the LA5 in the east still resemble the Western Kaleva, though ophiolitic serpentinite bodies are missing. The tectonic and depositional setting of the LA5 remain equivocal, but the assemblage is tentatively assigned to a stage in basin evolution later than the filling of the marginal rift basins.

Detailed plate tectonic speculations are beoynd the scope of the present paper, but some regional aspects are shortly considered. The incipient Kaleva deposition has been attributed to craton break-up and development of a continental margin by several authors (e.g. Park et al., 1984; Gaál and Gorbatchev, 1987). The NW-SE strike of the rift-related mafic dyke swarms within the North Karelia - Kainuu region plausibly represents the orientation of the cratonic edge and major axes of the marginal depositories. This overall geometry directly implies that different depositional environments would have prevailed in the craton interior and in the areas adjacent to the rifted margin at any given time during and after the continental margin formation. Interestingly, this view seems to be directly supported by some recent investigations. The Karelian depositional sequences show distinctly different carbon isotope characteristics in the northeastern and southwestern parts of the Archean Domain (Karhu, 1993), and the author interpreted this to represent primary division into cratonic and marginal subdomains, respectively.

After taking the potentially different geotectonic settings into account some remarks concerning the sequences referred to as 'Kalevian' in the literature are considered. Without going into great detail, it seems that formations apparently correlative with the Eastern Kaleva show rather different characteristics in different parts of the shield. Metagraywackes of the Jänisjärvi area (N of the Lake Ladoga; Hausen, 1930; Zagorodny et al, 1986) can be seen as the southeastern continuation of the NKSB, and quite similar lithologies, associated with mafic volcanics and arkosites, have been reported to directly overlay the basement in the Northern Ostrobothnia region (Honkamo, 1985). Although graywackes are also present in Kainuu, the abundance of quartz wackes, quartzites and phyllites is characteristic (e.g. Kontinen, 1986; Gehör and Havola, 1988; Kärki, 1988; Laajoki, 1988). The presence of iron-formations (see Laajoki and Gehör, 1990 and references therein) also is a feature that differs from the other areas.

A simple and idealized plate tectonic model is finally presented in Figure 41. Within the

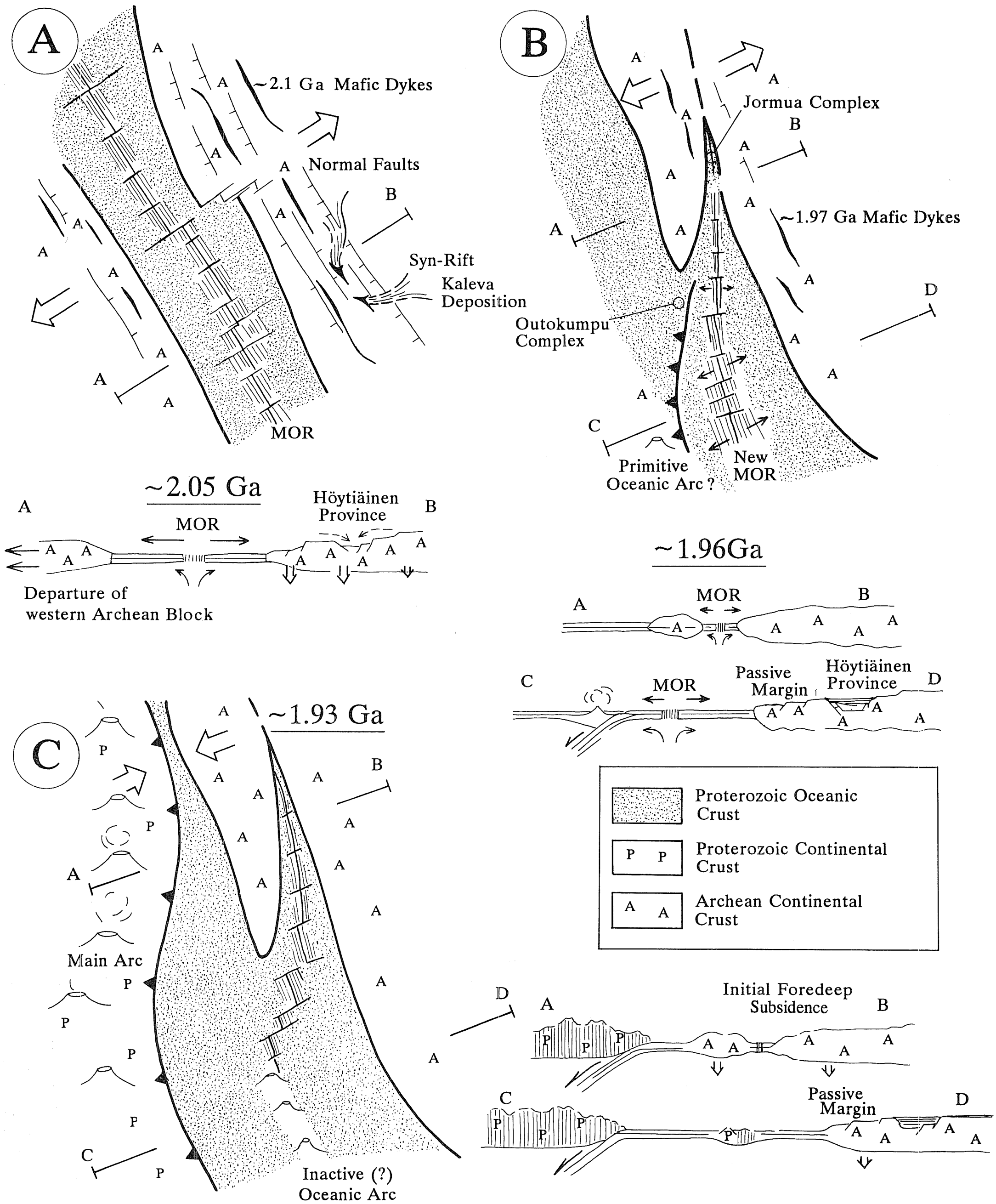

A
~2.1 Ga Mafic Dykes
Normal Faults
B
Syn-Rift
Kaleva
Deposition
MOR
~2.05 Ga
Höytiäinen
Province
MOR
Departure of
western Archean Block
B
Jormua Complex
~1.97 Ga Mafic Dykes
Outokumpu
Complex
Primitive
Oceanic Arc ?
New
MOR
~1.96Ga
MOR
Passive
Margin
Höytiäinen
Province
Proterozoic Oceanic
Crust
Proterozoic Continental
Crust
Archean Continental
Crust
C
~1.93 Ga
Main Arc
Inactive (?)
Oceanic Arc
Initial Foredeep
Subsidence
Passive
Margin

Fig. 41.

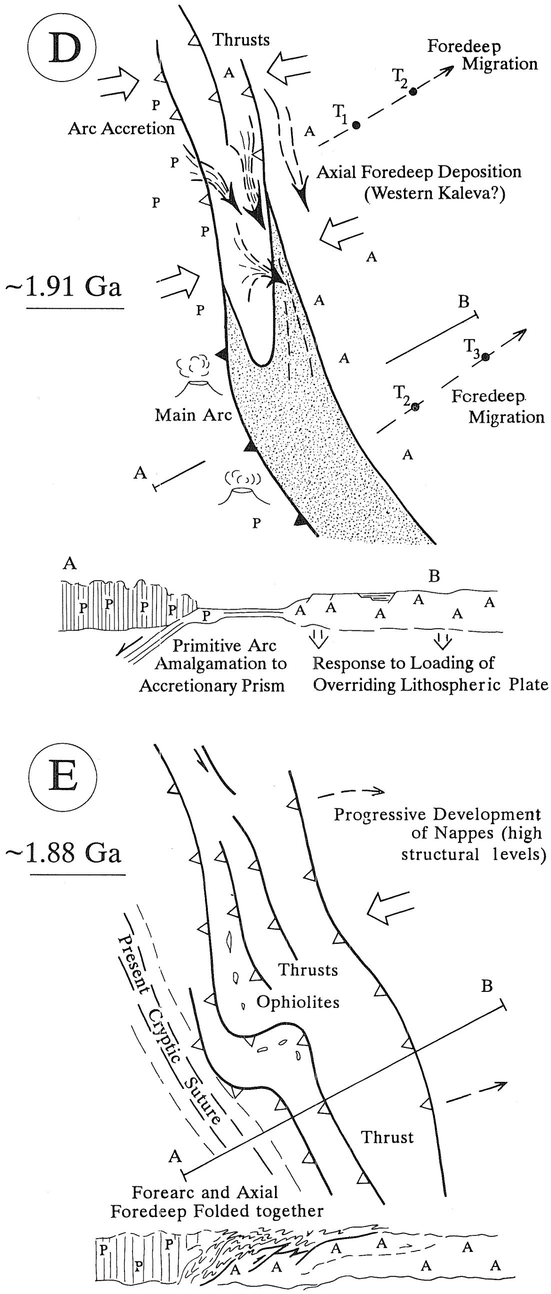


Fig. 41. Diagrammatic plate tectonic model for the NKSB. (A) Break-up of the Archean continent. A to B: development of a passive margin. (B) Formation of a narrow oceanic basin between two continental blocks (cf. Park et al. 1984, Fig. 2). For tectonic setting of Jormua and Outokumpu complexes see also Vuollo (1994) and references therein. B to C: initiation of a foredeep. (C) Position of the juvenile arc (or arc collage) prior to collision. C to D: foredeep migration towards the foreland. (D) Arc-continent collision in the north; subduction continues in the south. Note that during stages C and D forearc, trench and passive margin/foredeep deposition takes place in different parts of the same basin. D to E: development of a foreland fold and thrust belt. (E) Thrusting towards the foreland; basement involvement in deep levels and nappes on high structural levels. Scale strictly schematic.

model, the marginal-type graywackes (syn-rift Eastern Kaleva of this study) represent deposition in a relatively narrow zone of incipient rift basins. These are followed by subsequent passive margin/foredeep deposits (post-rift Eastern Kaleva). The development of the post-rift basin may be attributed to passive margin thermal subsidence (cf. Burke and Dewey, 1973) or alternatively to foreland bending due to load of the overriding plate (cf. the foredeep model of Hoffman, 1987). Deciding between these alternative models is difficult due to the poor timing constraints on deposition, and the time span of about 200 Ma between rifting and collision also sufficient to allow the development of both the passive margin and the foredeep. If a foredeep existed, the deposition is likely to have been diachronous, transgressing progressively from west to east and these sediments may have covered large areas within the foreland.

According to the present model, the rift basins are expected to recede towards the northeast, and the differences between the Eastern Kaleva deposits in North Karelia and Kainuu could, perhaps, be best understood by assuming the former principally as a rift basin and the latter as a passive margin/foredeep basin with minor syn-rift deposition. A foredeep model was first applied for the Kaleva rocks in Kainuu by Korkiakoski and Laajoki (1988), and later Laajoki and Gehör (1990) considered the presence of the iron-formations as supporting evidence for the proposed foredeep setting. The Western Kaleva rocks can be related to the scheme by further development of the foredeep towards a foreland flysh basin. It is pointed out, however, that the presented primitive Wilson-cycle application is unavoidably an oversimplification and should be taken as such; the potentially important role of the Kola-Lappidic Orogen (see discussion by Laajoki, 1990; 1991) has not been considered at all, for example.

The development of a collision zone in the west at about 1900 Ma ago resulted in thrust movements towards the foreland. The distended craton margin responded to the compressional stresses by shortening largely controlled by the architecture of the earlier extensional fault system. The rift related faults are interpreted as the main factor localizing the strong transposition observed within the study area. A single, progressively developing regional shortening event is proposed. The variable deformational styles over the North Karelia region are best explained by heterogenous shortening controlled by the complex pre-existing fault block geometries, variable and changing rheological properties and relative location within the thrust pile. As pointed out by Ward (1987) and Gaál (1990), the scarcity of subduction related magmatism within the Karelian (Archean) domain together with the observed easterly directed tectonic transport (thrusts) seem to imply westerly directed subduction prior the collision. However, if a minor amount of subduction (narrow ocean) and a doubly vergent ('inside-out') collision are assumed, an eastwards directed subduction (cf. Ekdahl, 1993) seems possible as well.

To conclude, the NKSB is interpreted here as a tectonically inverted extensional continental margin. The development of Kaleva basins begun about 2.1 Ga ago in relation to the extensive rifting of the Archean continent. Evidence of further (episodic) extension up to about 1.97 Ga is present in the form of mafic dykes. Multiple extensional events and repeated block movements resulted in a complex depository pattern. The local sequences record different depositional styles (settings within the basin) and highly variable completeness (degrees of preservation). For these reasons it seems that lithostratigraphic correlation of the Eastern Kaleva sequences will remain inherently difficult. The nature of the Western Kaleva depositories remains unresolved, but it seems possible that both oceanic and continental crust underlies those basins about 1.9 Ga in age.

CONCLUSIONS AND SUGGESTIONS FOR FURTHER WORK

Conclusions

(1) The variable depositional style, age and composition of the Kaleva rocks within the North Karelia Schist Belt (NKSB) cannot be explained by a uniform, single-stage genetic model.

(2) The Eastern Kaleva rocks of the Höytiäinen area have been interpreted as syn- and post-rift sequences within a marginal basin. The major element geochemistry points to a cratonic source area consisting predominantly of Archean rocks. The reported Proterozoic component (Huhma, 1987) is explained by a limited contribution from rift-related, continental mafic volcanics.

(3) Most of the studied lithological assemblages show major element characteristics similar to the Kaleva rocks of the Kiihtelysvaara region, but are different from the Western Kaleva rocks of the Outokumpu area. This is in accord with the models assuming a separate evolution for the Eastern and Western Kaleva in North Karelia (Ward, 1987). Lithological assemblage 5 correlates better with Western Kaleva and may conceivably represent a klippe of rocks lying to the west of the Suhmura Thrust Zone.

(4) The eastern contact of the Kaleva mica schist within the NKSB does not generally correspond to the basal contact of the unit. The conglomerates along the contact zone are interpreted as principally marking the paleolocation of a multiply-lived growth fault at the margin of the Eastern Kaleva basin.

(5) The rift related extensional features had a major influence to the style of deformation during the collisional crustal shortening. The eastern, originally westwards dipping basin margin normal faults developed into younger-over-older thrusts (e.g. the present eastern contact of the Kaleva mica schist), whereas the western marginal faults with a non-ideal, eastwards dipping orientation show rotational re-utilization as basement thrusts (e.g. the eastern contact of the Sotkuma inlier; see Kohonen and Elo, 1991).

Suggestions for further work

The swift progress of the Scandinavian joint project for a SIM-spectrometer (NORDSIM) opens new opportunities for provenance studies. Dating of single detrital zircons from the Eastern Kaleva is clearly a method for verifying or invalidating many of the suggestions presented here. In particular a test of the assumed cratonic provenance for LA 4, together with comparisons between the other assemblages could reveal critical new data. The apparent affinity of LA 5 with the Western Kaleva is also worth further study.

The current research into mafic dykes will probably provide new information about the age and spatial distribution of the extensional events. The analysis of these data together with the observed supracrustal sequences could resolve many questions pertaining to the evolution of the basin configuration.

The reasons for the remarkable strain partitioning into narrow zones within the basement should be studied in more detail. A suitable area for such studies is, for example, the region between the North Karelia and Kainuu. A regional and detailed investigation of the boundaries between the reworked and intact basement blocks could also produce important new results regarding the deformation style within the deep levels of a foreland fold and thrust belt.

Finally, the author believes that the regional and detailed data accumulated during last ten years already form an excellent basis for an exploration project. A rift basin setting has been proposed for most of the important sediment-hosted base metal deposits. It might therefore be beneficial to check whether the Hammaslahti ore body really is the only economically viable base metal deposit within the riftogenic Höytiäinen province.

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Appendix 1. XRF-analyses; location and short description of samples.

DATA SET 1 (samples collected by the author)

Sample	Area[1]	LA[2]	Map sheet	Grid coordinates		Description
8306000	SA	1	4224 12	6976.650	4497.405	pelitic schist
8311900	SA	1	4313 10	6986.770	4493.400	pelitic schist
8502301	SA	1	4224 12	6975.530	4496.700	pelitic schist
8702201	SA	1	4313 06	7008.630	4472.210	pelitic schist
8702202	SA	1	4313 06	7008.630	4472.210	quartzite;fine grained;secondary albite
8702205	SA	1	4313 06	7008.630	4472.210	quartzite;medium grained
8702211	SA	1	4313 06	7008.610	4472.190	pelitic schist
8707800	SA	1	4224 11	6969.040	4494.500	pelitic schist
8800002	SA	1	4224 12	6977.090	4497.460	graywacke schist;medium grained
8803300	SA	1	4313 06	7006.950	4474.220	pelitic schist
8616802	SA	2	4313 05	6998.860	4479.020	pelitic schist;dark
8617301	SA	2	4313 05	6999.030	4479.240	quartzite;coarse grained
8617308	SA	2	4313 05	6999.030	4479.200	pelitic schist;dark
8617601	SA	2	4313 05	6999.000	4479.120	pelitic schist;dark
8801002	SA	2	4313 06	7001.090	4478.500	pelitic schist;some carbonate
8806700	SA	2	4313 06	7002.220	4478.080	pelitic schist;dark
8615516	SA	3	4313 02	6981.160	4486.700	pelitic schist;dark
8616000	SA	3	4313 08	6992.840	4483.080	pelitic schist
8708800	SA	3	4224 11	6968.750	4490.960	pelitic schist
8709203	SA	3	4224 12	6976.260	4490.040	pelitic schist;dark
8709400	SA	3	4224 12	6976.060	4490.330	pelitic schist;dark
8506600	SA	4	4313 10	6988.300	4492.900	graywacke schist;fine grained
8508500	SA	4	4313 07	6989.790	4481.070	graywacke schist;medium grained
8608601	SA	4	4313 08	6996.810	4484.020	pelitic schist
8608602	SA	4	4313 08	6996.810	4484.020	graywacke schist;fine grained
8608800	SA	4	4313 08	6997.120	4483.900	graywacke schist;medium grained;quartz rich
8609902	SA	4	4224 12	6972.130	4492.660	graywacke schist;medium grained
8610101	SA	4	4224 12	6973.560	4492.670	graywacke schist;medium grained;quartz rich
8610401	SA	4	4224 12	6975.050	4492.080	graywacke schist;coarse grained
8610404	SA	4	4224 12	6975.050	4492.080	pelitic schist
8615501	SA	4	4313 07	6981.160	4486.480	graywacke schist;coarse grained;abundant Ksp
8615503	SA	4	4313 07	6981.160	4486.480	graywacke schist;coarse grained;some Ksp
8615504	SA	4	4313 07	6981.160	4486.480	graywacke schist;medium grained
8615505	SA	4	4313 07	6981.160	4486.480	graywacke schist;coarse grained
8617801	SA	4	4313 08	6999.100	4480.420	graywacke schist
8617802	SA	4	4313 08	6999.100	4480.420	graywacke schist;medium grained
8703501	SA	4	4313 06	7002.900	4475.990	pelitic schist
8707701	SA	4	4313 06	7000.900	4477.870	pelitic schist
8707706	SA	4	4313 06	7000.900	4477.870	pelitic schist
8707708	SA	4	4313 06	7000.970	4477.820	graywacke schist;medium grained;some carbonate
8707710	SA	4	4313 06	7000.970	4477.820	arkosic graywacke schist;coarse grained
8708201	SA	4	4224 11	6969.480	4494.170	pelitic schist;dark
8708302	SA	4	4224 11	6969.400	4493.600	graywacke schist;fine grained
8804502	SA	4	4313 06	7005.710	4475.220	pelitic schist
8805301	SA	4	4313 06	7001.630	4477.340	pelitic schist
8807401	SA	4	4224 12	6972.970	4494.340	graywacke schist;fine grained

8808102	SA	4	4224 12	6974.240	4492.750	graywacke schist;fine grained
8902701	SA	4	4313 08	6992.970	4482.070	pelitic schist
8902702	SA	4	4313 08	6992.970	4482.070	graywacke schist;fine grained
9000006	SA	4	4313 07	6984.000	4485.020	graywacke schist;coarse grained;some carbonate
9000012	SA	4	4313 07	6984.000	4485.020	graywacke schist;coarse grained;some carbonate
8615800	SA	5	4313 04	6988.620	4478.620	graywacke schist;medium grained
8616100	SA	5	4313 05	6996.040	4476.440	graywacke schist;medium grained
8616201	SA	5	4313 05	6995.540	4473.400	pelitic schist;dark
8616202	SA	5	4313 05	6995.540	4473.400	graywacke schist;medium grained
8616702	SA	5	4313 06	7000.000	4473.550	graywacke schist;medium grained;dark
8707400	SA	5	4313 05	6999.860	4473.670	graywacke schist;fine grained;dark
8800001	SA	5	4313 06	7000.320	4472.290	graywacke schist;medium grained
8807001	SA	5	4313 05	6996.750	4477.190	graywacke schist;medium grained;dark
8901101	SA	5	4313 05	6998.970	4475.400	graywacke schist;fine grained
8901500	SA	5	4313 05	6997.250	4475.320	graywacke schist;medium grained;dark

DATA SET 2 (data presented in Glumoff, 1987)

Sample	Area[1]	LA[2]	Map sheet	Grid coordinates		Description
8526776	Ki		4241 02	6939.300	4509.500	pelitic schist
8526777	Ki		4241 02	6939.300	4509.500	pelitic schist
8526778	Ki		4241 02	6939.300	4509.500	pelitic schist;staurolite
8526779	Ki		4241 02	6932.500	4503.500	pelitic schist
8526785	SA	5	4313 05	6995.570	4473.260	graywacke schist
8526786	SA	5	4313 05	6997.040	4478.070	graywacke schist;dark
8526787	SA	5	4313 05	6997.260	4478.270	pelitic schist
8526788	SA	4	4313 08	6999.080	4480.410	pelitic schist
8526789	SA	4	4313 08	6999.080	4480.410	graywacke schist
8526790	Oku		4313 01	6989.530	4463.830	graywacke schist
8526791	Oku		4313 01	6989.530	4463.830	graywacke schist
8526792	Oku		4313 01	6989.530	4463.830	graywacke schist
8526793	Oku		4313 01	6989.530	4463.830	graywacke schist
8526794	Oku		4313 01	6989.530	4463.830	concretion
8526795	Oku		4313 01	6989.660	4464.360	graywacke schist
8526796	Oku		4313 01	6989.660	4464.360	pelitic schist
8526797	Oku		4313 01	6989.660	4464.360	graywacke schist
8526798	Oku		4313 01	6983.640	4465.910	graywacke schist
8526799	Oku		4313 01	6983.640	4465.910	graywacke schist
8526800	Oku		4313 01	6983.640	4465.910	graywacke schist
8526801	Oku		4313 01	6983.640	4465.910	graywacke schist
8526802	Oku		4313 04	6980.090	4471.560	graywacke schist
8526803	Oku		4313 04	6980.090	4471.560	graywacke schist
8526804	HöW		4313 04	6981.960	4474.350	pelitic schist
8526805	HöW		4313 04	6981.960	4474.350	pelitic schist
8526806	HöW		4313 04	6981.960	4474.350	pelitic schist
8526807	HöW		4313 04	6983.910	4477.920	pelitic schist
8526808	HöW		4313 04	6983.910	4477.920	graywacke schist
8526809	HöW		4313 04	6984.070	4478.160	pelitic schist;dark
8526810	HöW		4313 04	6984.070	4478.160	pelitic schist;dark
8526811	HöW		4313 04	6984.180	4479.180	pelitic schist
8526812	HöW		4313 04	6985.920	4478.220	graywacke schist
8526813	HöW		4313 04	6985.920	4478.220	pelitic schist
8526814	HöW		4313 04	6985.920	4478.220	concretion

8526815	HöW		4313 04	6985.920	4478.220	graywacke schist
8526816	SA	4	4313 07	6980.750	4486.680	graywacke schist
8526817	SA	4	4313 07	6980.750	4486.680	graywacke schist
8526818	SA	4	4313 10	6982.910	4490.540	graywacke schist
8526819	SA	3	4224 09	6972.140	4489.890	pelitic schist
8526820	SA	3	4224 12	6973.000	4491.000	pelitic schist
8526821	SA	4	4224 12	6979.100	4491.600	graywacke schist
8526822	SA	4	4224 12	6979.100	4491.600	pelitic schist
8526823	SA	4	4224 12	6975.070	4492.050	pelitic schist
8526824	SA	4	4224 12	6975.070	4492.050	graywacke schist
8526825	SA	4	4224 12	6975.070	4492.050	graywacke schist
8527992	HöW		4313 02	6992.820	4467.110	graywacke schist
8527993	HöW		4313 02	6992.820	4467.110	graywacke schist
8527994	HöW		4313 02	6992.820	4467.110	graywacke schist
8527995	HöW		4313 02	6992.820	4467.110	graywacke schist
8627466	Oku		4224 03	6971.750	4466.300	graywacke schist
8627467	Oku		4311 10	6984.060	4458.360	graywacke schist
8627468	Oku		4224 02	6967.020	4468.070	graywacke schist
8627469	Oku		4224 02	6967.020	4468.070	graywacke schist
8627479	Oku		4224 03	6975.440	4462.070	graywacke schist
8627480	Oku		4224 03	6975.440	4462.070	graywacke schist
8627487	Ki		4241 03	6944.250	4500.770	pelitic schist
8627488	Ki		4241 03	6944.250	4500.770	pelitic schist

DATA SET 3 (unpublished data from Kiihtelysvaara area provided by A. Kontinen)

Sample	Area[1]	Map sheet	Grid coordinates		Description
29PW88	Ki	4241 02	6932.500	4503.650	pelitic schist
30PW88	Ki	4241 02	6932.500	4503.650	graywacke schist
31PW88	Ki	4241 02	6939.320	4509.500	pelitic schist
32PW88	Ki	4241 02	6939.320	4509.500	quartzite
2HÖ89	Ki	4232 06	6915.600	4512.200	graywacke schist
4HÖ89	Ki	4232 06	6915.100	4512.200	graywacke schist
5HÖ89	Ki	4232 08	6913.200	4522.060	pelitic schist
6HÖ89	Ki	4232 06	6916.170	4518.210	graywacke schist
7HÖ89	Ki	4241 04	6920.300	4516.150	pelitic schist
9AHÖ89	Ki	4224 12	6972.800	4492.700	graywacke schist
9BHÖ89	Ki	4224 12	6972.800	4492.700	pelitic schist
151HÖ89	Ki	4241 02	6938.800	4509.400	graywacke schist
152HÖ89	Ki	4241 02	6938.820	4509.400	pelitic schist
171HÖ89	Ki	4241 03	6920.300	4516.120	graywacke schist
172HÖ89	Ki	4241 03	6920.300	4516.120	pelitic schist
181HÖ89	Ki	4241 03	6920.800	4513.150	graywacke schist
182HÖ89	Ki	4241 03	6920.800	4513.150	pelitic schist
19HÖ89	Ki	4232 06	6915.600	4512.250	graywacke schist
201HÖ89	Ki	4232 06	6914.920	4511.760	graywacke schist
202HÖ89	Ki	4232 06	6914.920	4511.760	pelitic schist

1) SA= Study area, HöW= Area west from the lake Höytiäinen, Oku= Outokumpu area, Ki=Kiihtelysvaara area; see Fig. 23 for definition of the areas.

2) LA=Lithological assemblage (only within the study area).

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