DEPOSITION AND DEFORMATION AT A LOWER PROTEROZOIC CRATONIC

MARGIN, SOUTH-EAST FINLAND

Peter Ward

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CORRIGENDUM   to   p. 77

It is acknowledged that the term 'lithofacies' is used here in a somewhat different sense than in typical sedimentological studies, but it nevertheless seems more appropriate than "formation", "suite" or mere lithological descriptors such as "metapsammite" or "mica schist".

Likewise, the hazards of introducing a plethora of new geographical terms, and the possibility of frustrating, rather than simplifying stratigraphical correlation is freely admitted. On the other hand, it seems preferable at first to be overcautious, even extravagant in classification rather than to enthusiastically group diverse units : superfluous terms can be subsequently abandoned but amalgamated data are more difficult to subdivide. Terms and classifications introduced here are therefore regarded as informal, not inviolate.

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ABSTRACT

In south-eastern Finland, lower Proterozoic sediments were strongly deformed and metamorphosed to amphibolite facies during the Svecokarelian orogeny approximately 1.9 Ga. Three early schistosities and associated folds share an enveloping surface parallel to primary lithological layering whereas the six younger groups of structures recognized have steep attitudes and zonal distribution; only the earliest of these younger groups of structures are regionally significant in the study area. Deformation commenced with northwards translation of thrust nappes onto the adjacent Archaean craton and continued during prograde metamorphism. Internal deformation of the metasedimentary sequence became progressively more intense, as recorded by recumbent F_2 and F_3 folds and S_1, S_2 and S_3 schistosities, in particular S_2 differentiation layering. S_3 development included mylonite formation with local transposition of Archaean structures near the basement - cover interface.

Younger deformation is recorded by NW-trending ductile shear zones and minor folds (with rarer antithetic conjugate structures) followed by substantially vertical displacements along N-trending normal faults and shear zones. These are intrepeted dynamically as recording a progressive relative increase in vertical crustal stresses in response to gravitational and thermal disequilibrium induced by burial of Archaean basement beneath the earlier nappe structures and/or ultrametamorphism and accumulation of buoyant magmas in the lower crust.

In spite of this complex structural history, lithofacies investigations were practicable and resulted in the recognition of two distinct stratotectonic provinces that represent the deformed remnants of two separate, though possibly penecontemporaneous basins. The more westerly Savo province contains chemogenic and mafic volcanogenic lithologies, serpentinites and graphitic metapelites but is dominated by monotonous quartz - plagioclase - biotite - garnet metapsammites. Rarely preserved sedimentary structures suggest that these metapsammites originated as thick-bedded mass flow deposits in a prograding submarine fan setting. Near the present eastern margin of the province, intercalated metapsephites contain exclusively intraformational detritus consistent with deposition from debris flows and high-density turbidity currents in a submarine canyon or channelized inner fan environment. Abundant Na-plagioclase and biotite in Savo province metapsammites is considered to reflect sediment supply from a relatively K-poor terrane rich in chlorite-montmorillonite but petrographical criteria alone cannot discriminate between a possible Archaean provenance (with derivation from both greenstones and granitoids) or penecontemporaneous intermediate plutonic or

bimodal volcanogenic sources. Preliminary Nd-Sm whole rock and U-Pb detrital zircon isotopic data indicate however, the likelihood of a sedimentary mixing of detritus of both Archaean and lower Proterozoic igneous origin.

In contrast, the Höytiäinen province to the east contains better preserved and more diverse primary depositional features, with unambiguous evidence for a more local provenance, including Archaean granitoids and gneisses, Jatulian ortho-quartzites and penecontemporaneous metabasites. This diversity of source material is also reflected in a wider scatter in εNd ratios, including a component from penecontemporaneous tholeiitic volcanism derived from a mantle that was REE depleted with respect to CHUR.

Near the present western margin of the province, coarse-clastic lithofacies are interpreted as proximal, partly channelized prograding fan sequences with several fining and thinning upward sequences recognized, passing upward and laterally into quartzose metapsammites less diagnostic of a suprafan - middle fan environment. Stratigraphical relationships with associated pelitic and chemogenic lithogies are obscure, due to complex deformation but the coarser lithologies may record a terrigenous clastic influx prograding across finer-grained and more distal deposits, correlating with the upper Jatulian - Kalevian disconformity and facies transition widely recognized in eastern Finland.

On the basis of this dual intrabasinal volcanogenic and Archaean basement provenance, the Höytiäinen province precursor is thus perceived as having been a linear intra-cratonic trough within a more extensive system of rifting or fragmentation of the Karelian cratonic margin during the period 2.1 - 1.9 Ga. The Savo province pre-cursor is envisaged as a large submarine fan complex that was evidently flanked to the east by the Archaean craton, which locally acted as both depositional base-ment and source terrain but the ultimate nature and location of provenance for the bulk of the sediment remains problematical.

Such en echelon basin development at the craton margin is independent from, though compatible with extant and conceivable plate-tectonic paradigms: a geometrical analogue for the Finnish Proterozoic is distilled from the Neogene of the Australia - Banda arc region, incorporating continent-arc collision, a curvilinear subduction zone trace and transcurrent deformation but it is advocated that alternative mechanisms to crustal growth involving subduction and fore-arc accretion be con-sidered tenable until a more comprehensive understanding of Finnish geology is achieved.

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PART I :   INTRODUCTION

1   OUTLINE OF FINNISH GEOLOGY

Application of actualistic principles to the Precambrian of the
Baltic Shield was pioneered in the lithostratigraphic studies of
Sederholm (1899), and the Alpine tectonic concepts introduced by
Wegmann (1928). In addition, the major contributions of Sederholm
(1907) and Eskola (1914) to the understanding of metamorphism and
magma genesis derive from their study of the Proterozoic rocks of
south-west Finland. Broadly contemporaneous regional studies in
eastern Finland (Frosterus and Wilkman 1920, Metzger 1924, Väyrynen
1928, 1933) established major tectono-stratigraphic subdivisions that
are still accepted in spite of modifications in detail (compare reviews
by Sederholm 1930, Eskola 1963, Simonen 1980 and Laajoki 1983).

From these works an essentially tripartite classification of
the Finnish Precambrian has evolved. The oldest unit is confined to
eastern Finland and consists dominantly of granitoids and granitoid
gneisses with subordinate areas of greenstone and metasediment.
Isotopic age determinations have indicated a late Archaean age but
the widespread preservation of unconformably overlying metasediment
alone led early workers to recognize this unit as the most ancient
(Figure 1).

The other two subdivisions were originally made according to
contrasts in lithology, metamorphic grade and structural style,
rather than lithostratigraphic relationships. South-west Finland (and
the Precambrian of Sweden), with abundant volcanogenic and plutonic
lithologies was thus defined as Svecofennian while the metabasites
and quartz-dominated metasediments occurring as platform cover on
the Archaean basement were classified as Karelian. When radiometric
age determinations showed Karelian and Svecofennian plutonic rocks
to be broadly coeval and of early Proterozoic age (Kouvo 1958),

the two units were amalgamated and renamed Svecokarelian (see Eskola 1963) although Wegmann (1961) argued cogently for retention of the original scheme on the basis of contrasts in tectonic style. Laajoki (1983) also adopted this approach in his subdivision of the Sveco-karelian Orogen into the Svecofennidic and Karelidic realms.

The only substantial re-appraisal of this regional picture derives from recent work in north-east Finland and Soviet Karelia: Barbey and others (1984) interpret an extensive sequence of high grade metamorphic rocks as a distinct lower Proterozoic orogen effect-ively bisecting the Archaean craton, and marginally older than Sveco-karelian deformation. Figure 1 indicates the principal tectonic units of the central and eastern Baltic shield in a generalized manner, namely, the Inari-Kola and Karelian cratons, of late Archaean age, separated from each other by the 2.1 - 1.9 Ga Lappland granulite belt, and flanked to the south-west by the somewhat younger Sveco-karelian fold belt; this latter unit is distinctive in lacking detrital and isotopic evidence for an Archaean basement provenance (Huhma 1985).

The transition from the Svecokarelian fold belt to the Karelian craton records complex deformation, particularly overthrusting directed towards the craton, in places incorporating allochthonous Archaean basement (Wegmann 1928, Väyrynen 1939, Park and Bowes 1983). Subsequent deformation involved transcurrent shear strain along the craton margin with abundant syntectonic intrusions, including Ni-bearing mafic to ultramafic plutons and plagiogranites (Gaál and Rauhamäki 1971, Gaál 1972, Halden 1982). This Svecokarelian oro-genesis culminated at 1.9 - 1.8 Ga, since when the entire Finnish crust has behaved as a stabilized craton.

2   SCOPE OF PRESENT PROJECT

The area chosen for study has afforded the opportunity to examine the transition zone between the Karelian craton and the Svecokarelian fold belt, with Archaean basement both unconformably overlain by and tectonically intercalated within diverse Proterozoic lithologies. Some of these basement inliers were discussed by Eskola (1948) as type examples of mantled gneiss domes. Thus, a major objective of the project has been to examine the nature of Svecokarelian deformation, particularly in the Proterozoic cover but also in associated allochthonous basement. The results of this structural analysis are presented and discussed in the first part of the thesis, including correlation of structural elements, domain designation, geometry of major structures, microstructural expression and development, relationship to metamorphism and finally, their kinematic interpretation.

The second part of the thesis may be treated as a separate entity, dealing with lithofacies designation and their environmental interpretation. Such interpretations are as valid as the sedimentological models from which they derive, irrespective of structural complexity, since they are deduced directly from outcrop observations. On the other hand the succeeding section discussing stratigraphical correlation of lithofacies and reconstruction of basin morphology very much depends upon a thorough understanding of structural geometry. To this extent, results of both parts of the thesis are necessary to produce a coherent understanding of the study area as an example of an early Proterozoic cratonic margin basin.

The final part of the thesis is briefer, and more speculative, considering the relationship of the study area to the regional development of Finland during the Early Proterozoic.

3   FINNISH TERRAIN AND MAPPING

Topographical relief in the study area is slight, with a base level of 77 m defined by the lakes Orivesi and Pyhäselkä and the highest points being around 200 m, along a ridge formed by Jatulian orthoquartzite at Kiihtelysvaara. Exposure mostly comprises roches moutonnées protruding through glacial deposits and is easily located due to accurate positioning on the Finnish Cartographical Survey maps.

The precision of the 1:20 000 topographic series makes them practical base maps for field work, particularly in areas of low outcrop density. Each sheet covers an area of 100 km^2 and shows 1 km^2 grids in red (based on longitude 027° E) and black (based on longitude 030° E). Exposures referred to in the text are positioned according to the latter grid, using the last two digits of northings and eastings respectively, with a third figure more precisely defining outcrop location. For example, the reference 'Pyhäselkä 97.4 28.3' indicates an exposure on the Pyhäselkä 1:20 000 sheet, within the 1 km^2 grid having co-ordinates $_4$97 E and $_{69}$28 N. Only the 1:20 000 sheet boundaries appear printed on the accompanying 1:50 000 map of the study area but a transparent overlay with a 1 km^2 grid is provided for ease of reference to discussed outcrops. A similar grid is provided for Map 1 at 1:200 000 scale. All place names referred to in the text can be located on Maps 1 and 2 and appropriate figures by cross-reference to Appendix 1.

Orientations are given with respect to true north on the maps and in the text, and this is virtually coincident with grid north. During the period of field studies, and across the study area magnetic declination changed from $^+$6.5^o to $^+$8.0^o - this variation is considered negligible in comparison with that of structural trend at outcrop scale, when combined with measurement errors in using a hand-held Brunton compass.

4. MAJOR GEOLOGICAL ELEMENTS OF THE STUDY AREA

Map 1 shows the geological units present in North Karelia and adjacent parts of south-east Finland at a scale of 1:200 000. As mentioned in the preceding section, this area straddles the transition zone between the Svecofennian fold belt and the Archaean basement of the Karelian craton. Important features evident on Map 1 are the lower Proterozoic Jatulian platform deposits unconformably overlying the western margin of the Archaean basement between Koli, Kiihtelysvaara and Värtsilä and the allochthonous basement inliers further west, near Joensuu, Kontiolahti, Oravisalo and Outokumpu. This basement overthrusting accompanied deformation of the Proterozoic cover sequences, including the northwards translation of the Outokumpu nappe, whose trace is clearly defined by the distribution of the serpentinites, metabasites, chemical metasediments and sulphide ores of the Outokumpu assemblage (Wegmann 1928, Väyrynen 1939, Koistinen 1981). Synorogenic intrusive rocks postdated horizontal tectonism in the southwestern part of the map area, being coeval with dextral NW-trending ductile faults, in particular the Suvasvesi and Haukivesi shear zones (Gaál and Rauhamäki 1971, Parkkinen 1975, Halden 1982).

The study area itself is readily classified into two major provinces based on the distribution of lithofacies and major structures. The lithofacies of the Savo province in the west are grouped into stratotectonic assemblages, of which the Pyhäselkä assemblage is volumetrically the most significant (Map 2, Map 5). This consists dominantly of monotonous quartz-plagioclase-biotite metapsammites and local metapsephites derived from intraformational conglomerates. The minor occurrences of metabasite and quartzite in the Oravisalo assemblage are associated with allochthonous thrust schuppen of Archaean basement (Map 2, Figure 2). A small inlier of Archaean basement is also found above the Kettämö thrust zone near Hammaslahti (Map 2, Map 3).

The Kettämö thrust zone delineates at least in part, the boundary between the Savo province and the Höytiäinen province to the east, and is likely to continue northwards from the study area along the eastern margin of the Sotkuma basement inlier, constituting one of the major structural features of North Karelia (Map 1). Within the study area, allochthonous Archaean basement is not present east of the Kettämö thrust; eastwards from Kiihtelysvaara, Archaean basement is extensively exposed, devoid of Proterozoic cover but the basement-cover contact is a well-preserved unconformity (Map 1, Map 2, Map 5).

Metasediments of the Höytiäinen province tend to have better preserved depositional structures than in the Savo province, a greater abundance of metapelite and a lower proportion of plagioclase to quartz in metapsammites. In addition, coarse resedimented clastic deposits clearly indicate a local basement provenance for much of the province.

5.   GLOSSARY OF SELECTED TERMS

    Chirality is employed to describe the sense of displacement
in a ductile shear zone: left- and right-hand shears have sinistral
or dextral chirality respectively. This is to distinguish such dis-
placements from the sense of deformation indicated by the vergence
of minor structures on recumbent folds.

    Continuous cleavage is, according to Borradaille and others
(1982), a foliation defined by parallel alignment of mineral grains,
lacking any zonal distribution or differentiation (at the scale of
observation).

    Crenulation cleavage is defined by microfold axial planes, or
zones of solution transfer truncating any extant mineral alignment;
such cleavage tends to be zonal and spaced with either discrete or
gradational boundaries between cleavage domains and relict micro-
lithons.

    Differentiation layering is a non-depositional or non-magmatic
compositional anisotropy ascribed to metamorphic processes, particularly
solution transfer.

    Facing relates to tectonic inversion prior to the folding in
question and is indicated by variations in depositional or strati-
graphical younging: a downward facing F_2 fold occurs, for example,
on the inverted limb of an F_1 structure.

    A leucosome is a light-coloured unit within a gneiss or
migmatite, formed by metamorphic differentiation, in situ anatexis
or either concordant or discordant intrusion of felsic magma.

    A 'melanosome' is a dark-coloured unit within a gneiss or
migmatite, formed by metamorphic differentiation by either concordant
or discordant intrusion of magma, or as restite, remaining after the
extraction of felsic magma.

'Mesosome' refers to the portion of rock between any specified neosome component, irrespective of composition or origin, in the sense of Johannes (1983). It thus includes both unmodified palaeosomes and modified or restitic components remaining after metamorphic differentiation or partial melting.

Metapelite is employed with respect to modal grain size, as presently observed. Composition is indicated separately, as in calcareous, mafic, siliceous or micaceous metapelites.

Metapsammite refers to a rock with a sand-sized modal mineralogy at present, rather than a meta-arenite of particular composition. Where relict depositional features are evident, terms such as metagraywacke, metaturbidite or orthoquartzite are used.

Metapsephite refers to metamorphosed rudites, classified according to clast type.

A neosome is a component, commonly vein-like and discordant, that has intruded into, or differentiated within a body or rock.

NO : nicol prisms not mutually orthogonal.

NX : nicol prisms crossed orthogonally.

PPL: plane-polarized light.

Assignment of tectonic translation direction depends upon kinematic interpretation of relationships between fold hinges and extension lineation and furthermore can only be established in the presence of depositional younging determinations or an obvious, asymmetrical décollement.

Vergence refers to sense of overturning of antiforms in reclined to recumbent folds, using minor fold and cleavage - layering relationships.

FIGURE 1 : MAJOR TECTONIC ELEMENTS OF THE BALTIC SHIELD

Terminology adapted from Simonen (1980), Gaál (1982a) and Barbey and others (1984). The term Karelian is used here with reference to that part of the Svecokarelian orogen containing detritus of Archaean derivation and allochthonous Archaean basement, in contrast to the Svecofennian which has a wholly Proterozoic provenance and evolution (Huhma 1985).

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INARI - KOLA CRATON

LAPLAND GRANULITES

FOLD BELT

KARELIAN

CRATON

area of Map 1

SVECO-

FOLD

BOTHNIA-LADOGA ZONE

Onega

CALEDONIAN

FENNIAN

BELT

Ladoga

Leningrad

Helsinki

N

■ study area

Stockholm

CALEDONIAN

GOTHIAN < 1.6 Ga

SVECOKARELIAN 1.8-1.9

1.7 - 1.9 Ga

SVECOFENNIAN KARELIAN

1.9 - 2.1 Ga

GRANULITE BELT

2.6 - 2.9 Ga

ARCHAEAN CRATON

GOTHIAN FOLD BELT

Oslo

0   100   200   300 km
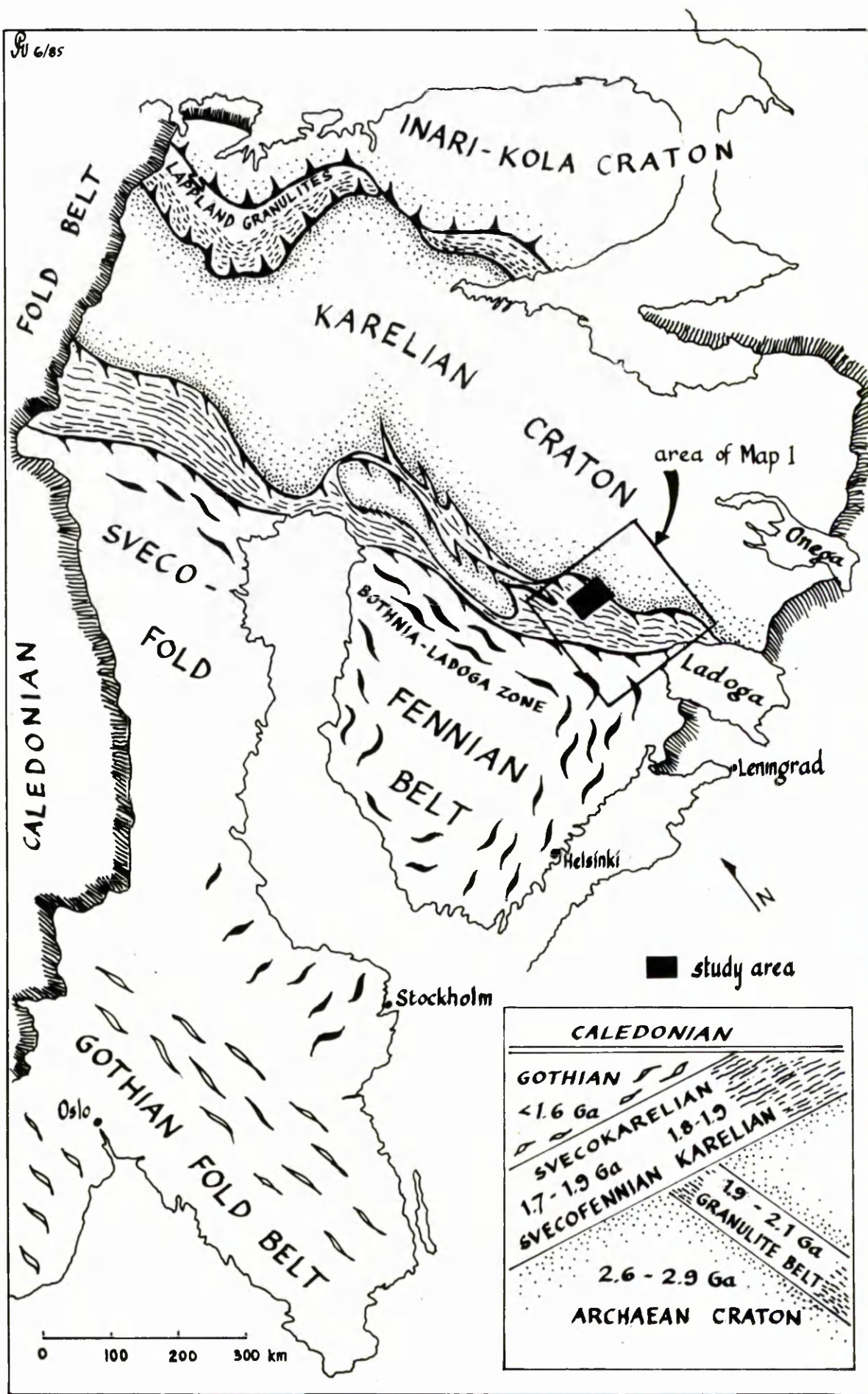
PART II : STRUCTURAL INVESTIGATIONS

1.   PROCEDURES IN STRUCTURAL ANALYSIS

1.1   GENERAL METHODS AND LIMITAIONS

Analysis of structurally complex terrane requires careful, though straightforward application of simple, well established principles such as outlined by Hobbs and others (1976) and Hopgood (1980). These principles include recognition of successive truncating or over-printing relationships - as indicated by intrusive neosomes or mineral segregation, growth and recrystallization, - or from de-formation and displacement of earlier elements, as recorded by folds and fractures.

While these procedures may be applied objectively to individual exposures, correlation of structural sequences between outcrops is a statistical rather than empirical process; confidence of such correla-tions depends upon several factors.

i) Abundance of exposures; in an area of good exposure there are better opportunities for detecting rare structural elements, or documenting geometrical variations of a particular element.

ii) Abundance and diversity of structural features; as the range and variety of elements increases, the probability of random development of consistent local structural sequences diminishes.

iii) Restricted rather than ubiquitous development of struct-ures; if some element is of local or zonal expression, then its absence from some exposures may lead to confusion.

iv) Variations in style or geometry of structures; if a particular element shows variation in style, it may be difficult to recognize at different exposures (cf. Williams 1970). Similarly, difficulties may arise if several structures of different age possesses a locally congruent geometry.

v) Availability of isotopic age determinations; although these results are also subject to statistical uncertainties, discrete, late stage neosomes are potentially valuable as absolute time-markers in correlation.

1.2 FORMAT FOR PRESENTATION OF RESULTS

The foregoing principles and limitations have been applied to outcrop mapping in the present study, in investigating the deformational history of the region. In general, a combination of poor exposure, zonal development and congruent geometry of superposed structures has hindered correlation such that no single exposure was found to record all observed features and relationships; in particular, overprinting relationships amongst the younger, zonal and weaker structures are rarely seen.

Therefore, the structural section of the thesis begins with descriptions of the more confidently established and most complete local structural sequences, from which a regionally correlated scheme is then derived, firstly for the Archaean basement (Section 2) and then for the Proterozoic cover rocks (Section 3). Following this demonstration of deformational sequence, variation in the nature and expression of each structural element is described in turn (Sections 4 and 5), along with definition of regional geological domains based on vergence and the disposition of major fold hinges, shear zones or thrusts. Origin and development of microstructures is also considered in Sections 4 and 5. Moreover, since metamorphic studies are not a major aspect of this study, is seems most apposite to discuss metamorphism in this context of regional deformation and microstructural development rather than in an independent chapter.

Section 6 concludes the structural analysis with an appraisal of the likely kinematic development of structures, and an attempt to integrate this into a regional deformational pattern.

Structural data for the whole study area are recorded at 1:50 000 scale on Map 2, whereas Map 3 presents detailed observations near the Kettämö thrust zone at 1:5 000; Map 4 attempts to convey the disposition and geometry of major structures based on a dimetric projection.

2.   STRUCTURES IN ARCHAEAN BASEMENT ROCKS

2.1   CRITERIA FOR RECOGNITION OF BASEMENT

Within the study area, lithological distinctions alone are
sufficiently diagnostic to distinguish the schistose metasediments
of the cover sequence from the banded migmatitic gneisses and
granitoids of the basement. No evidence for in situ ultrameta-
morphism of cover rocks exists and most exposures of metasediment
reveal some indication of relict depositional layering. In addition,
the presence of metamorphosed conglomerates and regoliths derived
from and unconformable upon gneissose lithologies at Kiihtelysvaara
and Oravisalo attests to a significant interval of time between
gneiss development and final burial beneath cover rocks. This
inference has been confirmed by isotopic age determinations: zircon
fractions from granitoid gneiss near Kiihtelysvaara yield a U-Pb
concordia upper intercept at 2.765 Ga while metabasite dykes in-
truding both gneissose basement and basal cover units range in age
from 2.16 Ga to 2.05 Ga (Pekkarinen 1979a).

2.2   STRUCTURAL SEQUENCE IN SELECTED REFERENCE AREAS

Sequential relationships of structures have been established
in five places within the basement of the study area. Although some
structures are only sporadically developed, others have been
correlated between the five reference areas with a high degree of
confidence. For convenience of reference in the following text
descriptions, numerical subscripts are used that correspond to this
finally correlated sequence, rather than according to the local
schemes, but both may be compared in Table 1. Roman numerals are
given with respect to structures in the basement, to avoid confusion
with the structural sequence in cover lithologies (for which Arabic
numerals are used).

2.2.i Oravisalo quarry area (Oravisalo 75.7 15.7)

Recent enlargement of a quarry by the Liperi road, 2 km north or Oravisalo post office, has exposed various features, including an unconformity between Proterozoic metasediments and the Archaean basement (Figure 2). The earliest and most conspicuous structural element recognized in the basement is a stromatic migmatite banding (S_I) defined by thin quartz-feldspar leucosomes alternating with thicker homogeneous, darker grey mesosomes (Figure 3e, 23b). This stromatic banding is disrupted by minor ductile thrusts, (S_{II}), filled with a similar leucosome to that in S_I, or deformed into intrafolial folds with attenuated limbs (F_{II}), (Figure 3e). Intrusion of mafic magma probably accompanied or preceded this deformation and is recorded by thin, discontinuous amphibolite bands broadly conformable with the stromatic banding of S_I. Because of this near concordance of neosomes, S_I and S_{II}, a composite S_I-S_{II} fabric commonly occurs.

Younger structures are only sporadically developed so that mutual relationships are not clearly seen. Within the quarry itself occur moderately tight folds (F_{III}) whose attitude differs from those of F_{II} and with which are associated quartzose neosomes, but seldom any axial planar foliation. In contrast S_{IV} is more widespread, and particularly prominent near the basement-cover unconformity (Figure 24 c). Although S_{IV} and S_I are divergent by $20°$- $40°$ at the unconformity itself, elsewhere S_I has been effectively transposed into S_{IV}. At exposures some 50 m structurally below the quarry, S_{IV} records intense ductile strain, with extreme flattening of quartz and feldspar grains and the development of quartz-ribbon mylonites. However, diffuse and anastomosing metasomatic zones, as shown in Figure 3 f may be a different manifestation of S_{IV} in more massive and competent lithologies, rather than belong to a later deformation.

Distinctly younger structures include a weak N-trending upright cleavage (S_{VI}) and a NE-trending fault (S_{VII}) that disrupts basement and cover alike.

2.2.ii Kakkari - Vannila area (Oravisalo 73.16 ; 75.17)

The most informative exposure was found at a small bluff
(Oravisalo 74.3 17.2), showing a range of lithologies but which all
show a distinct stromatic banding, referred to S_I, and intrafolial
folds (F_{II}), which also deform early, broadly concordant metabasites
(Figure 3 c). In addition there occurs a distinct mica-free pegmatite
phase, usually discordant to S_I but containing indications of syn-
tectonic intrusion (Figure 3 a,b). S_{IV} is also nearly concordant with
earlier structures, but clearly superimposed, exhibiting a variation
in intensity from quartz-ribbon and blastomylonites (Figure 3 g,h)
to foliations defined by syntectonic biotite crystallization
(Figure 3 c). The only neosomes associated with this fabric are
sporadic quartz veins (Figure 23 a,e).

2.2.iii Archaean basement at Hammaslahti (Pyhäselkä 95.6 27.0)

A small inlier above the Kettämö thrust consists of stromatic
gneiss containing some concordant mafic intrusions. Banding defined
by a pegmatitic leucosome and gneissosity is identified as S_I. Intra-
folial folds and concordant deformation at the margins of metadolerite
intrusion may be defined as F_{II} and S_{II} structures. A differentiated
metamorphic layering is developed only at the margins of the intrusion;
the central part preserves original textures, or has recrystallized
isotropically.

S_{IV} and L_{IV} are manifest as a discordant schistosity and
mineral elongation lineation, usually defined by biotite or amphibole
growth. It is consequently only weakly developed in the massive
granitoid gneisses or the more competent metadolerite.

Two younger sets of structures have been noted: S_{VI} may be
described as a weak crenulation cleavage in the schistose parts of
the metadolerite, or as a crenulation lineation in the felsic

stromatic gneisses. S_{VII} is represented by a NE-trending fault zone
in the metadolerite.

3.2.iv Archaean basement near Kiihtelysvaara (<u>Viesimo</u> 16.2 28.7)

Several kilometers east of Kiihtelysvaara, granitoid gneisses
and stromatic migmatites appear to have a relatively simple structure
dominated by leucosomes and concordant gneissose banding (S_I). These
are intruded discordantly by numerous metadolerite dykes known to be
Proterozoic since they also intrude lower Proterozoic orthoquartzites
and arkoses (Pekkarinen 1979a). The dykes contain a NW-trending
schistosity S_V and accompanying metamorphic recrystallization which
is also present as a zonal deformation in the enclosing gneisses,
particularly those richer in biotite.

3.2.v Archaean supracrustal rocks at Särkilampi
(<u>Viesimo</u> 17.3 24.0)

Several kilometers south of Kiihtelysvaara, an enclave of
supracrustal Archaean rocks occurs, with metasediments showing well-
preserved depositional layering (S_0). These are folded isoclinally
and refolded about more open folds with NW-trending axial planes.
Foliations to these folds are not well developed but parallel
intrusions of tourmaline-bearing pegmatite are boudinaged (Figure 3 d).
Rb-Sr isotopic ages of 2.63 Ga have been obtained for these, indicating
a maximum age for fold initiation (Pekkarinen 1979 a). These are in
turn intruded by unfoliated oligoclase pegmatites.

2.3 CORRELATION OF STRUCTURES IN ARCHAEAN BASEMENT

In view of the paucity of data and tectonically disjunct nature of basement areas, correlations presented in Table 1 should be treated cautiously. It is nevertheless apparent that the dominant fabric is the ubiquitous migmatitic or gneissose banding, likely to be itself a composite fabric. It is not known whether even earlier structures have been transposed, or obliterated by recrystallization and anatexis or how this banding correlates with the early structures seen in the supracrustal association. Similarly it is not always possible to discriminate between S_I and the concordant neosomes defined as S_{II}.

S_{IV} is the only other fabric consistently observed and it is plausible that the presently concordant geometry of S_I and S_{II} was at least locally accentuated, if not induced by transposition into zones of intense S_{IV} development. These structures, particularly as expressed at Oravisalo, closely resemble structures described from the Kaavi district by Park and Bowes (1983), both in fabric development and tectonic setting near the basement-cover interface.

2.4 RECOGNITION OF SVECOKARELIAN STRUCTURES IN ARCHAEAN BASEMENT

The existence of such younger deformation is apparent from two sources, namely, thrusting of basement gneiss into Proterozoic rocks and deformation of Proterozoic dykes intruding basement.

Within the study area compelling evidence for tectonic intercalation of basement and cover units exists at Oravisalo. The basement of the Oravisalo quarry area is dissected by several thrusts related to S_{IV} and which cause repetition of the basement-cover interface (Figure 2 b). A distinct arcuate topographic depression occurs between the quarry area and the Vannila-Kakkari basement to the N and may well be the surficial expression of a structural or

lithological boundary between separate basement allochthons (Figure 2 a). This evidence alone is sufficient to suggest that S_{IV} formed during a later, distinctly independent event but even more convincing is that S_{IV} below the unconformity is contiguous with the major S_2- S_3 Svecokarelian schistosity in the immediately overlying metasediment. More pronounced S_{IV} development subjacent to the unconformity probably reflects a visible increase in muscovite content, ascribed to illite formation during pedogenesis (Figure 24 c).

Additionally, equal area projections 2 and 3 in Figure 5 illustrate the virtual congruence of the major shape fabric lineation in juxtaposed basement gneiss and Proterozoic metapsammites, supporting a unified, concurrent fabric development.

The second source of evidence for Svecokarelian deformation of basement rocks comes from the structural relationships of early Proterozoic metabasite dykes. Many of these discordantly cut basement structures but themselves contain foliations, commonly accompanied by metamorphic alteration. Zircon populations have yielded isotopic ages between 2.16 and 2.05 Ga and their post-Archaean tectonic setting is further demonstrated in that they intrude Proterozoic, basement-derived orthoquartzites (Pekkarinen 1979 a). Their presence is therefore convenient in attempting to discriminate Svecokarelian from earlier deformations.

NW-trending zones in particular show diverse subparallel structures of both Archaean and Svecokarelian origin in addition to the Proterozoic dyke swarms. Further evidence of the longevity of such features and their persistent reactivation is provided by Pekkarinen (1979 a), referring to a 30 km shear zone with quartz veins containing epigenetic U-Th mineralization. Uraninite from this zone has been analysed, and yields a U-Pb concordia upper intercept at 2.34 Ga.

I	Mafic intrusions	S_i Stromatic migmatite banding and gneissosity	S_i Stromatic migmatite banding and gneissosity	Mafic intrusions	S_i Stromatic migmatite banding and gneissosity	S_i Migmatite banding and gneissosity in granitoids; 2.765 Ga	F_i Isoclinal folds with only weak schistosity; F_{ji} open to tight folds with NW-trending axial planes
II	S_{ii} Minor thrusts with weak schistosity	F_{ii} Intrafolial and isoclinal folds; Syn-tectonic leucosome	Metadolerite	Pegmatite	S_{ii} Transpostition of S_i and syntectonic neosomes; F_{ii} Intrafolial folds; Syn-tectonic pegmatites	S_{ii} Differentiation banding in metadolerite; F_{ii} Intrafolial folds in metadolerite; ? coeval and syntectonic tourmaline pegmatites (2.63 Ga)	F_{ii} Intrafolial folds in granitoid gneiss
III	F_{iii} local open to tight folds, weak axial plane cleavage	Mafic dykes; Felsic neosome			S_{iii} local mylonite foliation		
IV	S_{iv} Quartz-ribbon mylonites and sulphide-bearing fracture zones	S_{iv} Quartz-ribbon mylonites, blasto-mylonites and biotite schistosity in meta-basites; L_{iv} Amphibole lineation	L_{iii} rodding lineation		L_{iii} mineral elongation lineation	Jatulian mafic dykes	L_{iv} lineation in felsic gneiss
V					S_{ii} NW-trending zonal foliation with biotite recrystallization		
VI	S_v weak N-S trending cleavage	S_{vi} crenulation cleavage in differentiated dolerite					
VII	S_{vii} NE-trending dextral fault	S_v NE-trending fault					

TABLE 1

COMPARISON OF STRUCTURAL SEQUENCES IN ARCHAEAN BASEMENT

FIGURE 2 : ARCHAEAN BASEMENT AT ORAVISALO

a 1:20 000 scale map showing distribution of lithologies, major
   structural features and colinearity of the principal tectonic
   lineation in both basement and cover, designated as Svecokarelian
   L_3. Stippled pattern denotes biotite-plagioclase-garnet meta-
   psammites (Rääkkylä lithofacies); dashed pattern indicates
   quartz-feldspar-muscovite schists (Kankaala lithofacies);
   unoriented v-pattern represents amphibolite (Oravisalon posti
   lithofacies) while Archaean gneisses are represented devoid of
   patterning.

b Sketch section in the vicinity of Oravisalo quarry, showing
   thrusted repetition of lower Proterozoic unconformity on
   Archaean basement. Depiction of folds in Archaean basement is
   diagrammatic and has no relation to the northwards vergence of
   Proterozoic Group 3 thrusts and folds.

a — Legend:

- S₃ thrust zone
- L₃ fold hinge
- L₃ mineral lineation
- S₃ foliation
- S₀–S₂ composite layering
- S₁–S₁₁ banding in Archaean gneiss

Kakkari

Vannila

Basement-cover unconformity

Section N-S

Oravisalo quarry

Kankaala

Oravisalo post office

400 m / 200 m / 0 m

TN

b — Cross-section:

Group 3 quartz-ribbon mylonites in Archaean gneiss

Proterozoic unconformity

Oravisalo quarry lithofacies B

N — S

meta-arkose of Oravisalo quarry lithofacies A

Group 3 thrust

late Archaean or Jatulian mafic dykes

F₃

100 m

FIGURE 3 : STRUCTURES IN ARCHAEAN ROCKS

a S_I gneissosity truncated by a slightly discordant neosome
   consisting of elongate quartz and feldspar grains (depicted in
   black and white respectively), suggesting intrusion late during
   S_I if not recording younger deformation. (Oravisalo 74.3 17.2).

b Similar neosome to that in 'a' but with similar folds occurring
   where pegmatite apparently intruded at angles more discordant to
   S_I (Oravisalo 74.3 17.2)

c S_I gneissosity with concordant mafic and felsic layers, isoclinal-
   ly folded into F_{II} folds, with development of S_{II}. Further
   irregular felsic pegmatites are locally discordant to both S_I and
   S_{II}. Mafic units show subsequent recrystallization and a biotite-
   defined S_{IV} schistosity correlates with Svecokarelian S_3 in cover
   lithologies (Oravisalo 74.3 17.2).

d Banded tourmaline-pegmatite disrupted by boudinage concordant
   with differentiated S_{II}, Särkilampi (Viesimo 17.5 23.8).

e S_I leucosome-mesosome banding displaced along a neosome-filled
   shear (S_{II}). Intrafolial F_{II} fold at bottom shows a similar sense
   of deformation (Oravisalo 75.7 15.8).

f Anastomosing, diffuse metasomatized zones, rich in biotite and
   sulphide porphyroblasts, regarded as manifestations of S_{IV}
   (Oravisalo 75.7 15.8).

g S_{IV} quartz-ribbon mylonite developed in granitoid gneiss, with
   anastomosing mica-rich seams and relict feldspar porphyroclasts,
   possibly in part recrystallized rather than merely deformed.
   Natural size (Oravisalo 74.3 17.2).

h S_{IV} mylonite fabric superimposed upon earlier, fine- to medium-
   grained gneissosity. Natural size. (Oravisalo 74.3 17.2).

a

quartz-feldspar neosome

1 dm

b

1 dm

S_{I-II}

quartz-feldspar neosome

c

pre-D₂ amphibolite

post-S_I pegmatite

S_IV = Svecokarelian S₃

S_I

F_II folding S_I

S_I

1 m

S_I transposed into S_II

d

S_II

e

S_II

S_I

F_II

N

f

S_IV

g

h
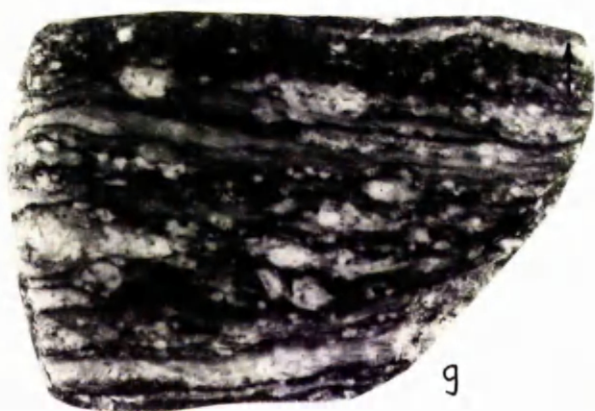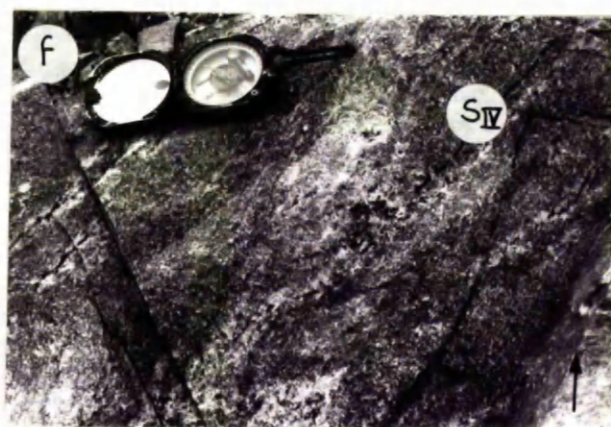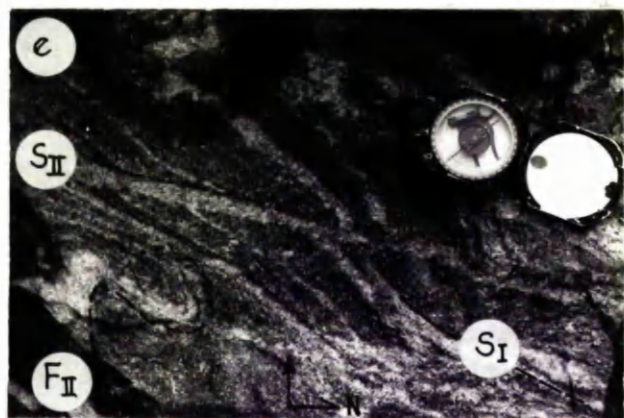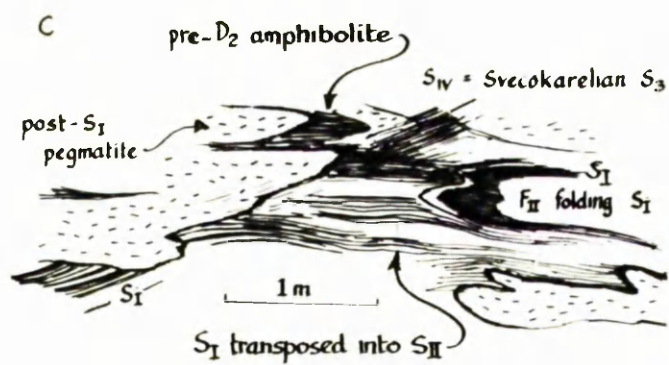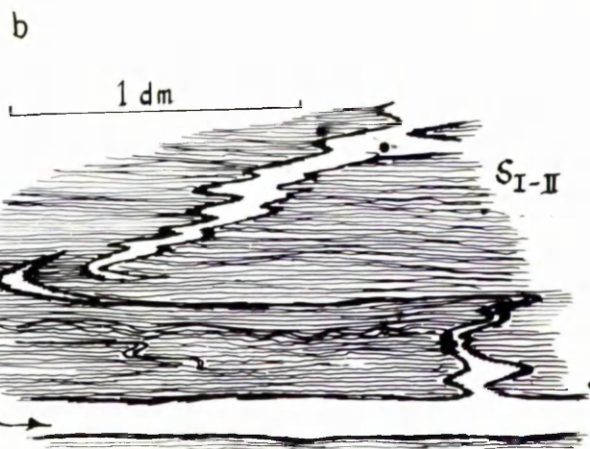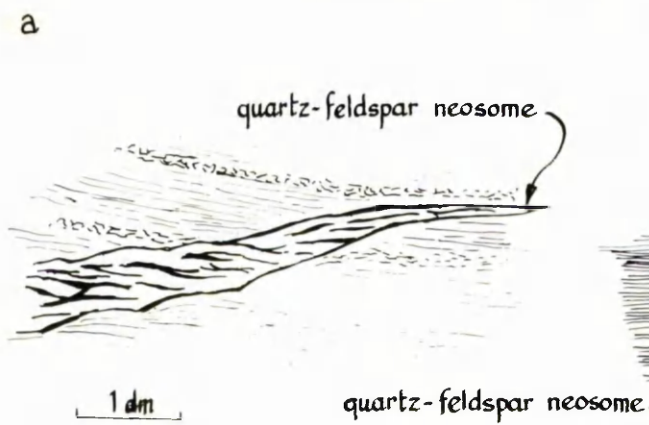
3. DEFORMATIONAL SEQUENCE IN PROTEROZOIC COVER ROCKS

3.1 INTRODUCTION

Because no individual outcrops were found containing the complete range of structures observed, it has been necessary to derive a regional deformational sequence from a combination of the most informative and representative exposures available (typical examples of which are illustrated in Figure 4). Consequently, uncertainties in correlation may exist, as discussed in Section 1, due to superimposed congruence or local absence of particular structures. As a result of this, the term 'Group' is used throughout the text, rather than the more usual and precise 'D_1...D_n' classification, such that 'Group 3 structures' for example, include elements having the same fabric expression, fold style and geometry as S_3 and F_3 in exposures where the structural sequence is empirically established, but whose exact relationships to other structures are not everywhere demonstrable. Use of upper case is purely for ease of reading and is not intended to indicate formalized terminology.

Coincidence of the Kettämö thrust zone and major lithostratigraphic boundaries justifies a geographical subdivision of the study area into two major zones - the Savo province in the south-west and Höytiäinen province in the north-east. This distinction is also valid in a structural sense, with the two areas exhibiting slight differences in fabric expression and structural history. Therefore, descriptions of reference exposures from each province are presented separately; deformational sequence in the Savo province is considered first, as it appears both more complete and straightforward. Structural sequences for each reference locality are tabulated and correlated with the proposed regional scheme in Tables 2 and 3 (Savo and Höytiäinen provinces respectively). However, throughout the text descriptions, subscript notation refers to the finalized, regional, rather than local schemes. Table 4 summarizes the structural sequences for both provinces.

3.2   SEQUENTIAL RELATIONSHIPS IN THE SAVO PROVINCE

Relict depositional layering is evident in most exposures and is designated S_o. Successive deformation of this surface at both small and large scales is best displayed near Ukonlahti, on the island Oravisalo but several other localities are also described from which overprinting relationships are clearly established.

3.2.i   Ukonlahti area   (Oravisalo 73. 18.)

The earliest generation of folds identified are only rarely seen (Figures 4b,c, 6h) and are classified as F_2 since they at least locally deform a differentiation layering (S_1) parallel to S_o (Figures 4c, 6g). Except where axial planar to such F_2 folds (Figures 4c, 6g) S_2 is a prominent quartz-biotite differentiation layering effectively parallel to relict depositional banding (Figures 4b, 8c, 9d). This composite S_{o-1-2} surface is deformed into open to isoclinal recumbent F_3 folds (Figures 4b, 8c, 9a,b). Axial planar S_3 is observed in more pelitic lithologies (Figures 8c, 9a) but L_3 is commonly a more prominent feature, defined by elongate knots of biotite (Figure 4c). All these early fabrics are deformed by zonal NW-trending Group 4 structures, typically minor dextral shear zones (Figure 20a). These are in turn overprinted by the N-trending S_5 cleavage (Figure 20a) which is related to a regional scale reorientation of all earlier structures, passing through Ukonlahti (Maps 2 and 4). Sporadically observed younger and weaker fabrics are difficult to constrain sequentially; the most frequently seen is a E-trending cleavage, in places axial planar to open warps.

Apart from quartz veins, the only neosomes seen are pegmatite dykes broadly concordant with, but locally truncating S_2 and deformed by S_3 (Figure 13b,d).

3.2.ii Rääkkylä area (Rääkkylä 81. 14.)

In this area relationships amongst early structures are well eatablished, due to situation in a major F_2 hinge zone, Hence, S_2 differentiation layering is commonly at a high angle to lithological layering and the early differentiated fabric S_1 is readily distinguishable (Figure 6a-f). L_3 also is demonstrably independent and cuts F_2 folds (Figure 9f), but mesoscopic F_3 folds and S_3 foliations are rarely seen. More open folds nearly colinear with L_3 could be late stage F_3 or else F_5 folds (Map 4) but otherwise, the only record of younger deformation is in the form of N-trending fractures and zones of alteration (Figure 20 g).

3.2.iii Hypönniemi area (Varpasalo 68. 19.)

The earliest recognized folds are F_2, with angular hinges and a prominent axial planar S_2 differentiation layering (Figure 6 h). S_3 is a less conspicuous fabric, usually parallel to S_2 but its independent origin can be demonstrated in F_3 fold hinges (Figure 9c,e). L_3 is a prominent SW-plunging biotite lineation and in Figure 17c is clearly seen to have been disrupted by NW-trending S_4 cleavage and shear zone development. Although mutual overprinting relationships have not been found between these and S-SW plunging open folds, the latter are regarded as younger, Group 5 structures on the basis of analogy with structures at Ukonlahti (Map 4). All these structures are commonly overprinted by a weak E-trending cleavage axial planar to open folds (Figure 6h); a similar fabric trends ESE but mutual relationships are not apparent. All earlier fabrics are disrupted by small faults and veinlets striking N and even younger steep arrays of fractures or thin veinlets with dihedral angles of 30°- 50° in horizontal exposures (Figure 6h).

3.2.iv Niva - Sammallahti (Pyhäselkä 93.24; 91.20.)

Metapelitic lithologies at Niva sometimes show transposition
of S_0 into a slightly discordant S_2 foliation, indicating the presence
of inverted F_2 fold limbs (Figure 28cd). Metapsammitic schists show
a more distinct S_2 differentiation layering characterized by biotite
-rich foliae normally parallel to depositional layering but sometimes
distinctly divergent. This fabric is classified as S_2 as in places it
cuts across an earlier but similar S_1 foliation (Figure 26e).

S_3 is recorded in metapelites as a discrete crenulation cleavage
axial planar to moderately tight F_3 folds with round hinges. S_3 in
metapsammites is more typically a continuous fabric defined by growth
of individual biotite crystals. The S_5 crenulation cleavage is similar
but generally weaker and distinguishable by its steeper axial plane,
striking between S and SW. F_5 folds tend to be subparallel to the L_3
lineation defined by elongate aggregates of biotite.

Weak, later structures are sporadically expressed but only
rarely can mutual relationships be ascertained: Figure 20h shows
a weak, disjunctive E-trending S_7 cut by thin N-trending veinlets
(S_8), in turn overprinted by E and SE arrays of fractures or veinlets,
classified as S_9.

3.2.v Niittylahdenrantatie area (Niittylahti 91.31.)

Depositional layering is well-preserved and is locally deformed
into intrafolial F_2 folds, occasionally with transposed hinges. Con-
sistent directions of depositional younging indicate that these are
minor folds on a larger F_2 inverted limb (Figure 5, Map 4) S_3 is a
continuous cleavage or locally a differentiation layering slightly
divergent to S_0 and S_2. F_3 folds are not common but tight to iso-
clinal and are consistently downward-facing. S_0, S_2 and S_3 are all
steeply disposed but a younger generally sub-parallel crenulation
cleavage (S_5) can sometimes be distinguished (Figure 8f). Even younger
is a locally developed NE-trending zonal crenulation cleavage (S_6).

Group	Ukonlahti	Rääkkylä	Hypönniemi	Niva - Sammallahti	Niittylahdenrantatie
S_0	Depositional layering	Depositional layering	Depositional layering	Depositional layering	Depositional layering
1	S_a differentiation layering parallel to S_0	S_a differentiation layering and intrafolial F_a folds			
2	F_b isoclinal fold with spaced axial planar S_b differentiation layering	F_b isoclinal folds with axial planar S_b differentiation layering	S_a differentiation layering axial planar to F_2 isoclinal folds	F_a recumbent folds, S_a transposing depositional layering and biotite differentiation layering	F_a recumbent folds,
3	F_c tight to isoclinal folds with axial planar S_c and L_c shape fabric	F_c asymmetric recumbent folds with axial planar S_c biotite schistosity and L_c shape fabric	F_b tight to isoclinal folds with axial planar S_b, locally as differentiation layering	F_b tight to isoclinal folds with axial planar S_b differentiation layering	F_b recumbent folds, axial planar S_b with biotite growth
	Quartz veins and pegmatites	Quartz veins	Quartz veins and pegmatites	Quartz veins	Quartz veins
4	Minor F_d folds of dextral chirality, with NW-trend		Minor F_c folds of dextral chirality with NW-trending S_c crenulation		
5	Steepening of earlier structures with minor F_e folds and N-trending S_e crenulation cleavage		S-SW-plunging F_d folds with weak axial planar S_d crenulation cleavage	S_c crenulation cleavage	Steepening of earlier structures with S_c crenulation cleavage
6					S_d NE-trending crenulation cleavage
7	S_f, weak E-trending cleavage		E- and ESE-trending weak S_f; open F_7 folds	S_d, weak E-W cleavage	
8		S_d N-trending fractures	S_g, N-trending fractures	S_e, N-trending fractures	
9			Conjugate fracture arrays	Conjugate fractures with E-ESE trend	

TABLE 2: CORRELATION OF SVECOKARELIAN STRUCTURES IN THE SAVO PROVINCE.

3.3 SEQUENTIAL RELATIONSHIPS IN THE HÖYTIÄINEN PROVINCE

In addition to depositional layering being readily recognizable, except in the more schistose metapelites, depositional younging directions are commonly found, enabling better constraints to be made concerning fold vergence and tectonic translation direction than in the Savo province. The area near Kettämönniemi shows most structures and their relationships clearly, but observations from several other localities are presented, to indicate the regional consistency of structural parameters.

3.3.i Kettämönniemi - Suhmura area (Pyhäselkä 95.25, Niittylahti 94.30)

Due to a comparative abundance of exposures and depositional younging criteria, structural geometry has been resolved better here than elsewhere and results are presented at a scale of 1:5 000 on Map 3. The earliest fabric, S_2 is seen independently as a differentiation layering only in the hinge zones of recumbent F_3 folds (Figures 7c, 32a,b) but anomalous zones of inverted depositional grading indicate local high-strain. Group 2 structures also include isoclinal folds and transposition or other disruption of sedimentary features (Figure 29e, Map 4). S_3 shows limited development as a differentiation layering in metapsammites, particularly in minor fold hinges (Figure 11d) but is more typically defined by deformation of pre-existing elements such as flattened clasts and transposed lithological boundaries (Figure 14e); in places this flattening has culminated in mylonite formation (Figure 14f).

NW-trending F_4 folds can be mistaken for F_3 normal-limb minor folds of similar trend, except the presence of a steep crenulation cleavage axialsurface shows them to be younger, as does the deformation of Group 2 and Group 3 microstructures (Figures 11g, 17a, 18a,c). Group 5 structures deform these, with fold interference relationships sometimes seen (Figure 4e, 17e). S_5 is a prominent crenulation cleavage

axial planar to open folds plunging S-SW at gentle to moderate angles
(Figures 4e, 8e, 32f). Apart from sporadically developed fractures of
late, but otherwise uncertain affinity, the only younger feature
consistently seen is a weak E-trending cleavage (S_7), in places axial
planar to open folds (Figure 4h).

3.3.ii Mulonsalo area (Niittylahti 95.37.)

The principal lithology is a micaceous metapelite, conducive
to the development of the zonal but predominant S_3 crenulation cleavage
(Figures 7a, 11e, 34d,f). Where this is axial planar to F_3 folds, S_2
schistosity can be discerned parallel to any relict lithological
layering and isoclinal F_2 folds are also seen in places (Figure 7a,b).

S_5 is a discrete crenulation cleavage similar to S_3 (Figure
19a-f) and where the latter is reoriented into a steep attitude, it
is commonly difficult to discriminate between the two cleavages;
unequivocal examples of overprinting relationships have nevertheless
been found (Figure 11f). The only younger structure consistently
observed in this area is an E-trending crenulation cleavage (Figure
20f).

3.3.iii Kalliojärvi area (Heinävaara 03.32.)

In general a well-developed differentiation layering parallel to
S_0 is present in metapsammites and is regarded as S_{0-2} (Figures 4d, 8d)
but since neither F_2 nor F_3 minor folds have been found it is
possible that the fabric generally represents composite S_{0-2-3}.
However sporadic depositional younging determinations and cleavage
relationships where S_3 is demonstrably divergent to S_0-S_2 indicate
the presence of overturned F_3 folds. Mutual overprinting relationships
between a sporadic NW-trending and abundant N-trending crenulation
cleavages have not been established at Kalliojärvi, but they are
regarded as S_4 and S_5 respectively, by analogy with Group 4 and
Group 5 structures at Kettämönniemi. S_5 has associated quartz veins

and is axial planar to moderately tight F_5 folds (Figure 17h). Younger than these structures are NE-trending disjunctive kink zones and S_6 crenulation cleavage (Figure 20c).

3.3.iv   Hammalahti mine area   (Rauansalo 01.28.)

Depositional features are generally well preserved in the vicinity of Hammaslahti mine and the earliest tectonic deformation is recorded by transposition into S_2, with accompanying boundinage of coarse-grained depositional units (Figure 37c). S_3 is a lithology--parallel cleavage axial planar to asymmetrical F_3 folds with highly attenuated limbs (Figure 11b). Conjugate quartz vein arrays are deformed by S_3 and in turn truncated by Group 5 veins (Figure 12d). A NW-trending crenulation cleavage cutting F_3 folds is regarded as S_4 (Figure 11b) and is distinct from N-trending Group 5 structures (Figures 12d, 37c). Still younger are small NE-trending faults and kinks, commonly with dextral displacement, classified as Group 6 structures.

3.3.v   Tikkala area   (Onkamo 04.17.)

Group 2 structures are manifest as a micaceous differentiation layering in metapsammites and boudins developed in thicker units, deformed by F_3 and F_5 folds (Figures 7e, 10a, 12g,h, 38c). Except where the S_2 schistosity is present, it is difficult to discriminate between Group 2 and Group 3 structures, with only rare examples of F_2 folds overprinted by S_3 (Figure 7d). F_3 folds are tight to iso-clinal, frequently with transposed inverted limbs (Figures 10a, 11a). Shear or slide zones, although parallel to S_3 axial planes, could be earlier if related to Group 2 boudins. However, unlike boudins, no examples were found of such shears deformed around F_3 folds. It is therefore possible that they are even younger, Group 5 structures.

The present steep N-trending attitude of these early features is attributed to reorientation by Group 5 structures; consequently

Group	Kettämönniemi – Suhmura	Mulonsalo	Kalliojärvi	Hammaslahti mine	Tikkala
S_0	Depositional layering	Depositional layering	Depositional layering	Depositional layering	Depositional layering
1	Not observed	Not observed	Not observed	Not observed	Not observed
2	Localized slide zones ans S_a differentiation layering	Local F_a isoclinal folds and weak axial planar S_a differentiation layering	S_a differentiation layering in metapsammite	F_a recumbent isoclinal folds S_a transposition layering	Sporadic S_a differentiation layering in metapsammites
3	F_b asymmetric recumbent folds with axial planar S_b continuous cleavage and L_b shape fabric lineation Quartz veins	F_b tight to isoclinal folds and S_b transposition and crenulation cleavage Quartz veins	F_b isoclinal folds and S_b transposition and crenulation cleavage Quartz veins	F_b isoclinal folds with S_b transpostion and con-(slaty) cleavage; L_b boundins and shape fabric lineation Quartz veins	F_b isoclinal folds with S_b transposition and continuous (slaty) cleavage; L_b boundins and shape fabric lineation Quartz veins
S_4	NW-trending S_c crenulation cleavage axial planar to minor folds of dextral chirality; rarer NE-conjugate crenulations		NW-trending S_c crenulation cleavage	NW-trending S_c crenulation cleavage	
5	S-SW-plunging F_d folds with axial planar S_d crenulation cleavage	S-SW-plunging F_c folds with axial planar S_c crenulation cleavage	Asymmetric F_d folds with axial planar S_d crenulation cleavage	Asymmetric F_d folds with steepening of earlier structures, quartz veins	Asymmetric F_c folds with steepening of earlier structures and S_d crenulation cleavage, quartz veins
6			NE-trending kinks and S_e crenulation cleavage	NE-trending dextral faults	
7	E-trending S_e crenulation	E-trending S_d crenulation			

TABLE 3: CORRELATION OF SVECOKARELIAN STRUCTURES IN THE HÖYTIÄINEN PROVINCE.

it is difficult to distinguish between S_3 and S_5 foliations away from F_5 fold hinges (Figures 10, 19g, 20d).

3.4 OVERALL STRUCTURAL SEQUENCE DETERMINED FOR THE STUDY AREA

Based on the foregoing local structural sequences, a total structural scheme has been derived for both the Savo and Höytiäinen provinces, and is presented in Table 4. Figure 4a gives a diagrammatic summary of geometrical relationships. Correlation between the two provinces has been made on the basis of the close correspondence amongst Group 3, Group 4 and Group 5 structures in particular. Validity of the proposed sequence is supported, though not proven, by similarities between these structures and those of other published schemes from elsewhere in Karelia, considered in the succeeding section. Therefore, since mutual overprinting relationships amongst the later and weaker (Group 6 - Group 8) structures in the study area are not confidently established, their sequence has been adopted from the literature, on the basis of corresponding trends. This procedure seems justifiable in that the structures are zonal and nearly vertical, so that their original trends are unlikely to have been substantially reoriented. On the other hand it may be unacceptable if trends were originally curvilinear.

It is evident from each of the local, as well as the overall schemes that a broad two-fold subdivision of the structural sequence is tenable. That is, Group 1, Group 2 and Group 3 structures are all contained within an enveloping surface roughly conformable with depositional layering and characterized by recumbent folds, thrusts and penetrative schistosities. Younger structures in contrast, are characterized by more zonal development and upright axial surfaces. These empirical distinction are considered significant in interpreting the deformational history of the region and consequently, the following discussion of the nature and expression of each group of structures is subdivided accordingly.

GROUP	SAVO PROVINCE		HÖYTIÄINEN PROVINCE	
9		E-ESE-trending conjugate fractures and veinlet arrays		Not observed
8	S_8	N-trending faults, sporadically with pseudotachylite		Not observed
7	S_7	Weak disjunctive E-trending cleavage	S_7	Weak E-trending crenulation cleavage
	F_7	Open minor folds		
6	S_6	Rare NE-dextral faults	S_6	NE-ENE dextral faults and sporadic crenulation cleavage and kink zones (sometimes sinistral)
5	S_5	Weak N-trending cleavage	S_5	N-trending faults and shear zones, crenulation cleavage, locally intensely expressed
	F_5	S-SW-plunging folds and shear zones with vertical displacement	F_5	Dextral minor folds (viewed down plunge) and shear zones
4	S_4	Weak but widespread NW-trending cleavage	S_4	NW-trending discrete crenulation cleavage; more rarely NE trend.
	F_4	Dextral folds; NW-trending shear zones, sporadically conjugate NE-trending zones	F_4	Dextral folds and occasional conjugate shear zones
3	S_3	Continuous cleavage, mylonitic thrusts	S_3	Continuous and crenulation cleavages, thrusts
	F_3	NNW-NNE verging reclined to recumbent folds. Later more open folds	F_3	NNW-NNE verging reclined to recumbent folds
	L_3	Shape-fabric lineation defined by elongate knots of biotite	L_3	Defined by elongate clasts and biotite-porphyroblasts
2	S_2	Pervasive quartz-biotite differentiation layering, quartz veins	S_2	Differentiation layering, locally developed in metapsammites; narrow slide zones, transposition and boudinage of S_0
	F_2	Tight to isoclinal recumbent folds	F_2	Rare isoclinal folds, with transposed hinges
1	S_1	Bedding-parallel differentiation layering; quartz veins		Not recognized
	F_1	Intrafolial folds		
0	S_0	Load casts, slumps	S_0	Load casts, slumps, sedimentary dykes

TABLE 4: CORRELATION OF SVECOKARELIAN STRUCTURES BETWEEN HÖYTIÄINEN AND SAVO PROVINCES.

3.5   DEFORMATIONAL SEQUENCE IN RELATION TO OTHER PARTS OF KARELIA

The complex nature of Proterozoic deformation in Karelia has long been recognized. Metzger (1925) and Wegmann (1928) described a later upright folding, with a NNE-trending axial plane, super-imposed upon earlier NE-vergent reclined folds and thrusts. In addition to this dominant early deformation, Väyrynen (1939), Gaál (1964), and Gaál (1972) also identified and recognized the regional importance of NW trending ductile faults and shear zones, particularly the Suvasvesi and Haukivesi zones (Map 1), aspects of which were studied in greater detail by Parkkinen (1975) and Halden (1982).

Regional studies stimulated by mineral exploration produced integrated pictures of structural development to the south-west of the present study area, near Savonlinna (Gaál and Rauhamäki 1971) as well as to the north, including the Outokumpu mining district (Map 1) (Gaál and others 1975, Bowes 1976, Gaál 1977, Koistinen 1981, Park 1983, Park and Bowes 1983). The recent review papers by Bowes and others (1984) and Park and others (1984) present a regional classification based on the above studies and this scheme may be compared with that of the present study area in Table 5.

Correlation of the structural schemes is based on similarities using numerous parameters, particularly in the case of earlier structures. The distinct, differentiated nature of Group 1 and Group 2 foliations in the study area and recumbent F_2 folds compares well with pre-D_1 and D_1 structures elsewhere in Karelia, and the Group 3 lineation defined by aggregates of biotite in the study area can also be confidently correlated with L_2 of Bowes and others (1984). Likewise, the form of F_3 folds and the presence of Group 3 mylonites near the basement-cover interface at Oravisalo corresponds well with D_2 features north of Outokumpu (Park and Bowes 1983).

Group 4 structures closely resemble those described as D_{2c} by Koistinen (1981) and Halden (1981), in having consistent NW trends,

with dextral chirality. In the study area however, the antithetic conjugate NE-trend is difficult to identify if relationships to younger structures are obscure, since the D_4 structures described by Bowes and other (1984) share a similar trend. Group 5 and D_3 structures also correspond closely in geometry, orientation, fabric expression and overprinting relationships. Microstructural evidence also supports this correlation of schemes from the two areas, with a thermal metamorphic peak indicated between Group 4 - Group 5 and D_{2c}- D_3 deformations, respectively.

Comparisons cannot be readily made for younger structures since no convincing overprinting relationships have been established amongst Group 6 and Group 7 structures in the study area. It seems likely that Group 6 structures, while weakly and rarely expressed in the study area are better developed elsewhere in Karelia. As discussed in Section 3.4, correlation with the scheme of Bowes and others (1984) has been attempted primarily based on similarities in geometry. The rare examples of overprinting amongst the weak Group 7, 8 and 9 structures of the study area are however consistent with the sequence proposed from similarities in geometry with D_5, D_6 and D_7, respectively.

STUDY AREA	REGIONAL SCHEME OF KOISTINEN (1981), BOWES AND OTHERS (1984)	
GROUP 1	pre-D_1	Nappe emplacement; segregation quartz veins.
GROUP 2	D_1	Isoclinal folds (F_1); axial planar metamorphic segregation banding (S_1).
GROUP 3	D_2	Asymmetrical open to tight folds (F_2); axial planar schistosity (S_2); thrusts; mylonite-phyllonite in S_2 transposition; strong mineral lineation (L_2).
GROUP 4	D_{2c}	Asymmetrical folds, conjugate in places (F_{2c}); crenulation cleavage (S_{2c}); expression largely confined to major NW-SE-trending wrench-fault zones.
GROUP 5	D_3	Open, upright N-to NNW-trending folds, commonly asymmetrical (F_3); crenulation or fracture cleavage (S_3) and intersection lineation (L_3) develop locally.
GROUP 6	D_4	Open, upright NE-trending folds (F_4); axial planar cleavage (F_4); prominent crenulation or rodding (L_4) in parts.
GROUP 7	D_5	Open, upright, E-trending folds (F_5); axial planar cleavage or fractures (S_5).
GROUP 8	D_6	Open, upright, N-trending folds (F_6); axial planar healed cleavage or fractures (S_6); local pseudotachylite.
GROUP 9	D_7	Conjugate set of open, upright WNW- and WSW-trending folds (F_7) that pass into fractures (S_7).

TABLE 5.   COMPARISON OF KARELIAN STRUCTURAL SCHEMES.

FIGURE 4: SEQUENTIAL STRUCTURAL RELATIONSHIPS

a   Schematic depiction of overprinting relationships observed for
     Svecokarelian planar and linear structural elements throughout
     the study area.

b   Coaxial fold interference pattern due to tight F_3 refolding of
     intrafolial F_2 folds. Biotite selvedges enhance   lithological
     boundaries parallel to S_2 and may also represent rare S_1 differentiatic
     layering. S_3 axial planar growth is less distinct (Oravisalo
     72.8 11.7).

c   Hinge zone of mesoscopic F_2 fold in metapsammite, with axial
     planar S_2 differentiation layering across depositional boundary
     and S_1 biotite foliae. Elongate knots of biotite represent
     superimposed L_3 (Oravisalo 72.1 17.3).

d   Exceptionally distinct S_2 differentiation layering in metapsammites,
     deformed into F_5 folds, with considerably more attenuation of limbs
     in interbedded laminated metapelite (Heinävaara 03.7 32.5).

e   Lithological layering and parallel S_3 are deformed by dextral
     F_4 minor folds with NW-trending axial planes. These have been
     refolded by F_5 with N-trending upright axial planes (Pyhäselkä
     95.6 27.6).

f   Sinistral F_5 folds deform composite S_2-S_3 parallel to depositional
     la yering. Curvature of S_5 axial plane indicates the presence of
     open E-trending F_7 fold (Niittylahti 94.5 31.3).

a

F₂ L₃ F₃ S₃ S₄ S₆ N

S₂ S₀/₁ S₃ thrust S₇ S₈ S₅ F₅

b F₃ N F₂ S₀₂ 1 dm

c S₀-₁ L₃ S₂

d F₅ S₀-₂

e F₄ S₅ S₀-S₃ S₄ N

f F₅ S₅ F₇ S₀-₃ N
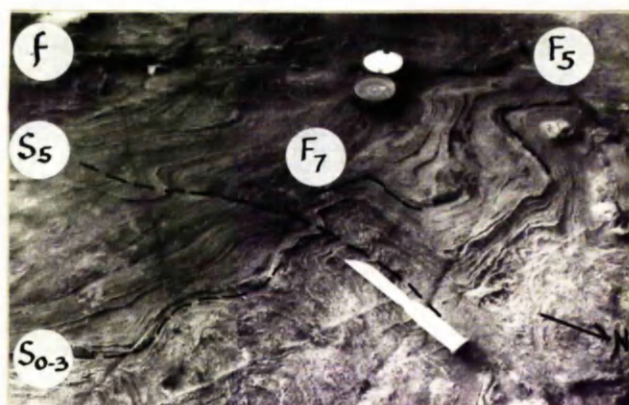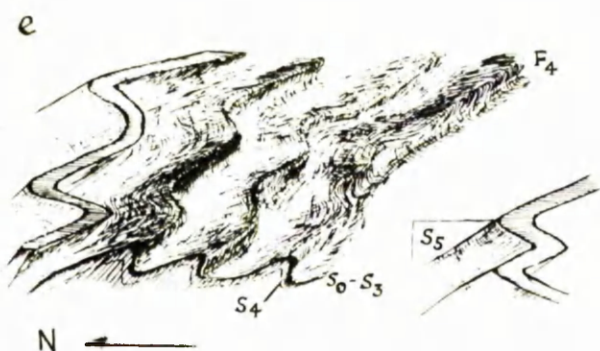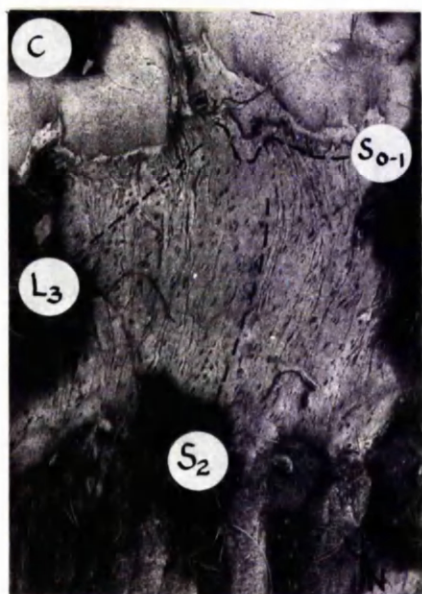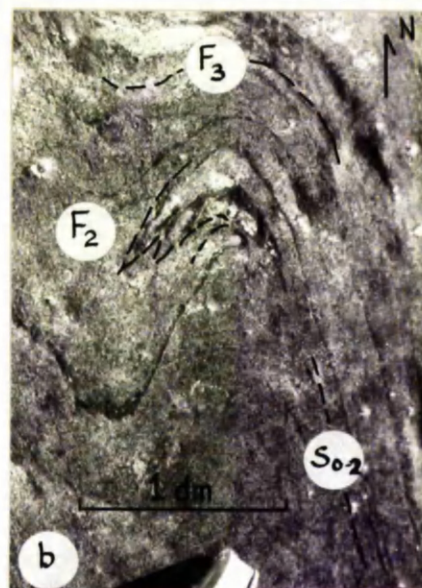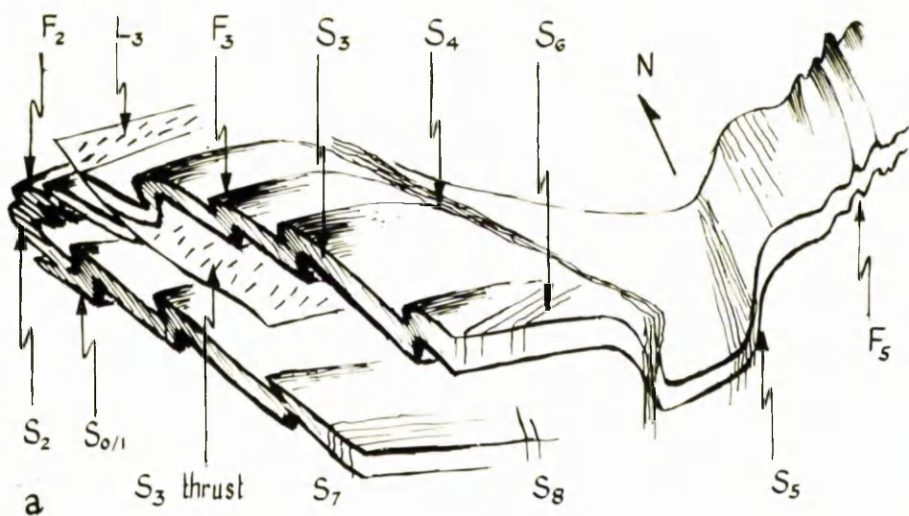
4. <u>NATURE AND EXPRESSION OF EARLY, DOMINANTLY LOW-ATTITUDE</u>
   <u>STRUCTURES IN PROTEROZOIC COVER ROCKS</u>

These structures are characterized by enveloping surfaces
roughly concordant with depositional layering. Younger deformation
has substantially modified the geometry of these older structures,
but an originally shallow attitude is inferred from relatively
undisturbed domains between Oravisalo and Nieminen (Map 2), and
the characteristic style of recumbent isoclinal folds.

In spite of their near concordance, it has been possible to
recognize up to three distinct foliations and linear fabrics,
locally distinguishable on the bases of overprinting criteria,
fabric differences, scale of structures and vergence relationships,
particularly where depositional younging features are preserved.
It is emphasized that these distinctions are descriptive and
remain valid regardless of whether all three sets of structures
developed during a single progressive deformation, or under
completely unrelated stress systems.

Figure 5 shows geographical distribution of domains for
Group 1, Group 2 and Group 3 structures recognized with respect to
normal- or inverted-limb vergence and major thrusts and folds.
Equal area projections 1-9 record differences in orientation
of these structures across the study area - phenomena ascribed to
younger, post-Group 3 deformation.

4.1   GROUP 1 STRUCTURES

4.1.i Distribution and structural domains

Since they all share nearly concordant enveloping surfaces,
Group 1 structures have commonly been recrystallized, transposed

or obliterated by superimposed Group 2 and Group 3 fabrics. Therefore, the most suitable place to examine the existence or nature of a bedding-parallel S_1 fabric is in the hinge zone of an F_2 fold, where S_2 is locally discordant to lithological layering (Figure 6c,d). A major structure of this kind is recognized near the village of Rääkkylä, trending ENE for more than 4 km and indeed it is only from this particular F_2 domain that S_1 has been consistently identified (Map 2).

4.1.ii Fabric elements, microstructures and metamorphism

In general, S_1 is a differentiation layering with persistent but anastomosing biotite-rich foliae up to 1 cm thick. Concordant quartz veins are also present and have been recorded elsewhere in North Karelia (Koistinen 1981, Park and others 1984). It is apparent that the development of this layering was very dependent upon lithological type, being considerably more distinct in thinner, fissile units than in massive ones (Figure 6a,b).

Linear fabric elements have not been recognized, except for the intrafolial fold evident in figure 6b; there is thus no indication of magnitude of F_1 structures or their vergence. It would indeed be logical to regard these structures as early, anchimetamorphic structures due to lithostatic load although evidence presented by Koistinen (1981) indicates substantial tectonic dislocations at this stage, namely emplacement of the Outokumpu nappe. The study area has yielded no evidence to confirm or refute this, nor any reliable indication of ambient metamorphic conditions; the present biotite-dominated assemblage may be a later, largely mimetic recrystallization from a former chlorite-illite-muscovite assemblage.

4.2   GROUP 2 STRUCTURES

The Kettämö thrust zone, defining the boundary between the Savo and Höytiäinen provinces may also be used to separate two major structural domains in which S_2 and F_2 have somewhat different development: in the Höytiäinen province, S_2 is less pervasive and more zonal in distribution than in the Savo province.

4.2.i   Structural domains and distribution in the Savo province

Three distinct domains are recognized, although for much of the province, magnitude and vergence of F_2 structures is largely unknown. Rare depositional younging determinations nevertheless suggest that the bulk of the metasedimentary sequence is right way up and not on the inverted limb of recumbent F_2 nappes. Unfortunately, this assertion is difficult to demonstrate unequivocally because S_2 is effectively parallel to lithological layering over a substantial part of the province. There are however, two areas, designated by cross-hatching in figure 5 that yield convincing evidence for the existence of major F_2 structures.

The more westerly of these domains trends in an ENE direction through the village of Rääkkylä. No depositional younging directions have been determined but many outcrops show very distinctive widely spaced and anastomosing, biotite-rich cleavage domains at a high angle to lithological layering and axial planar to F_2 minor folds, thus indicating an F_2 hinge zone of regional significance (Map 4).

Between the Kettämö thrust zone and the shores of the lake Pyhäselkä ((Figure 5, Map 2), at Niittylahdenrantatie (Pyhäselkä 91.31.)),   metaturbidites yield consistent downward younging

determinations with respect to S_2, thus delineating the inverted limb of a major F_2 fold nearly 8 km along strike and at least 300 m in thickness. South-west of this zone, at Niva, mafic metapelites may occupy the corresponding hinge zone to this inverted limb, with normal-limb vergence indicated by younging and cleavage criteria between Niva and Sammallahti (Figure 5, Map 2).

It is possible that this major structure is contiguous with the F_2 hinge domain at Rääkkylä but S_2 vergence relationships are so far lacking from the intervening exposures between Nieminen and Sammallahti. In this interpretation, originally gentle F_2 plunges would have been preserved at Rääkkylä but subsequent, dominantly F_5 folding would have reoriented Group 2 structures near Hammaslahti exposing an oblique cross-section through the major isoclinal fold, as depicted in Map 4.

Such a conclusion is of significance in connexion with observations of tectonic style at regional scale. Both of the major F_2 fold sections described have a common position structurally above a major décollement expressing tectonic intercalation of Archaean basement and Proterozoic cover sequences (Maps 1 and 2). Elsewhere in North Karelia, the deformed basement-cover interface reinforces this tectonic sequence, particularly between Outokumpu and the western margin of the Sotkuma basement inlier (see Gaál and others 1975, figures 13 and 26). Similar features are recognized further to the north by Väyrynen (1939), implicit in his definition of the 'Outokumpu nappe' and the 'Kaavi imbricate structure'.

Throughout the western part of the study area mesoscopic F_2 folds and associated S_2 are sporadically observed but are too rare to permit definition of any major structures. Geometry is further complicated by superimposed, nearly coaxial F_3 and

and more open F_5 folds. In general however, these F_2 folds are distinctive in their angular profile and intensity of axial planar S_2 differentiation layering (Figure 6g,h).

4.2.ii Structural domains and distribution in the Höytiäinen province

Through most of the province it is difficult to identify Group 2 structures: S_2 is only readily discernible where at a high angle to S_3 in F_3 fold hinges (Figure 7a-c) or as a differentiation layering in some metapsammites (Figure 7f).

Two zones of Group 2 structures are indicated by anomalous younging directions in metagraywackes above the Kettämö thrust (Map 3, Figure 5). There the upper limb of a younger, F_3 structure includes two narrow, downward-younging zones, indicating localized Group 2 'slide zones' of relatively high strain.

Depositional younging criteria delineate an apparently extensive inverted F_2 limb in the north of the area at Kumpu ((Map 2), Co-ordinates Niittylahti 98.37.) This is difficult to demonstrate conclusively, due to the complexity arising from superposed, coaxial F_3 and F_5 structures but younging reversals of a similar nature some 10 km southwards at Hammaslahti mine (Co-ordinates Rauansalo 01.28.) likewise led Gaál (1977) to deduce the existence of major early structures. Since Kumpu and Hammaslahti share lithological similarities in addition to being along strike from one another, they may well form parts of a single, complex zone of F_2 deformation.

4.2.iii   Group 2 microstructures and fabric development

The distinctive differentiation layering, parallel to
depositional banding and axial planar to F_2 folds is characteristic
of, and virtually unique to Group 2 structures, particularly within
the Savo province. This layering is evident microscopically as
alternations of domains richer and poorer in mica (Figure
8a,c,d,e). The latter type are interpreted as relict micro-
lithons approximating original composition, although recrystallized,
whereas the former are considered to have resulted from the
preferential dissolution and removal of silica in particular,
by solution-transfer processes (Williams 1972, Stephens and
others 1979). This mechanism implies that, given similar external
conditions, the development of S_2 will have been lithology dependent
and is therefore better expressed in quartz-rich metapsammites
than in micaceous metapelites. A consequence of this is that
because metapsammitic biotite schists are the dominant lithology
in the Savo province but subordinate to muscovite-rich metapelites
in the Höytiäinen province, intensity of S_2 development is itself
a useful parameter in discriminating between the two provinces.
In addition, relict depositional features and detrital fabrics
are more abundant and better preserved in the Höytiäinen
province. The precise reasons for these differences remain
unclear, but they may reflect regional variations in heat flow
or depth of burial of the provinces or more local influences such
as contrasts in anisotropy, mineralogy, water content and porosity
and resistance to fracturing, facilitating or impeding removal
of components from the site of dissolution.

4.2.iv   Group 2 mineral parageneses and metamorphism

In general, microstructural evidence supports interpretation
of Group 1, Group 2 and Group 3 fabrics as having developed during
a progressive metamorphic and deformational continuum. Thus the

assemblage extant during S_2 formation may be regarded as part
of an overall transition from a detrital mineralogy - whose
composition is only inferred - to an extensively recrystallized
syn- or post-S_3 climactic assemblage.

The Höytiäinen province metagraywackes have sufficient
well-preserved detrital fabrics to permit inferring of the
following primary mineral assemblage:

$$\text{quartz + feldspar + illite} \pm \text{chlorite} \pm \text{calcite} \pm$$
$$\text{Fe-sulphide} \pm \text{hydrocarbons}$$

Apart from the clay minerals, which presumably dehydrated and
recrystallized to produce the present muscovite and biotite,
and hydrocarbons, now present as graphite, these phases would
evidently have been in equilibrium with syn-S_2 metamorphism.
Major processes in fabric development were apparently intra-
granular deformation and solution truncation of detrital grains
with remobilization of components in sites of relatively high
stress (Figure 8b). Solution transfer was probably effected
by trapped connate water combined with that released during
clay dehydration reactions.

With the exception of mica, the only non-detrital phase
evident in metapsammites during S_2 formation is garnet, although
even this is rare from pre-S_3 assemblages in the Höytiäinen
province. Calcareous concretions in Savo province metapsammites
show a more diverse mineralogy including the following pre-S_3
assemblage:

$$\text{calcite + biotite + quartz + plagioclase} \pm \text{scapolite}$$

Alternating laminae defined by microcline, graphite and calcite
abundances occur in the calcareous Mulo lithofacies and are
considered to represent S_2 differentiation layering; none of these
parageneses seems however, to be particularly diagnostic of
metamorphic conditions.

In the Savo province metapsammites, the predominance of
biotite over muscovite probably reflects more mafic detritus
such as chlorite or montmorillonite than in the Höytiäinen
province source material and while relict detrital grain boundaries
are virtually absent, a primary origin for the abundant plagio-
clase is indicated by its fairly sodic composition: no other
suitable Na-bearing detrital precursor is readily envisaged.

To summarize, Group 2 microstructures formed during a
prograde metamorphic event, concomitant with recumbent folds
and thrusts initiated during the development of Group 1 structures.
The absence of accompanying plutonic rocks at the present erosion
level, or volcanic rocks known to be contemporaneous with
deformation suggests that metamorphism was not, at least in the
early stages, primarily due to anomalous increased heat flow.
Instead, lithostatic load due to basin filling is postulated to
be the initial cause, embracing the transition from diagenesis
to burial metamorphism. The major S_2 differentiation layering
was produced at upper greenschist to lower amphibolite facies
conditions during ensuing tectonic overthrusting.

4.3 GROUP 3 STRUCTURES

As with Group 2 structures, there are distinctions in style that justify separate consideration of the Savo and Höytiäinen provinces, and the Kettämö thrust zone is itself correlated with S_3. Within both of these provinces, smaller structural domains are recognized and relatively abundant $S_3 - S_0$ relationships and younging determinations have resulted in a better definition of geometry and vergence than for other sets of structures.

4.3.i Structural domains and distribution in the Savo province

Figure 5 shows Group 3 domains on the basis of fold vergence from which it is evident that only local inverted limbs exist, at least in those areas amenable to study. The variation between equal area plots in Figure 5 is largely a result of younger deformation with the data from the Oravisalo-Rääkkylä, Oravisalo basement and Nieminen plots recording attitudes suffering least from reorientation. In this region the L_3 biotite lineation shows gentle ENE and WSW plunges but elsewhere it has a moderate plunge and trends consistently between SSW and WSW.

F_3 minor folds tend to be parallel to this L_3 lineation but at Oravisalo for example, they are commonly divergent, as shown in the relevant equal area plots in Figure 5 . In Figure 9 d, the two lineations are actually mutually perpendicular: L_3 has the same expression and trend as elsewhere, while the fold is indistinguishable from others ascribed to F_3 on the bases of style, attitude and axial planar fabric. Such folds are therefore interpreted as later aspects of Group 3 deformation rather than assigned to a distinct deformational phase. Similarly, large, low amplitude folds virtually coaxial with L_3 are recognized between Oravisalo and Rääkkylä and since they have low-attitude axial planes, they too are regarded as late Group 3 structures (Map 4).

Various profiles of more typical F_3 minor folds are illustrated in Figure 9; these and other such structures, are indicated on both Figure 5 and Map 4.

a) The inverted limb of an F_3 fold is apparent in the south-western part of the island Oravisalo, with a broad hinge zone several hundred meters in width. S_3 is present as a differentiation layering in this hinge zone, in contrast to its usual occurrence as a continuous cleavage with a penetrative alignment of mica and elongate quartz grains. This hinge zone may be traced along the western shore of Oravisalo and as such could account for the preservation of relatively abundant $S_2 - S_0$ relationships.

b) Much of the remaining area of Oravisalo shows normal-limb vergence with respect to Group 3 structures but two relatively narrow inverted-limb zones are depicted in Figure 5. Each may be traced for more than 2 km along strike and commonly show a more intensely developed S_3 foliation than elsewhere. Such narrow but laterally extensive inverted-limbs evidently belong to a continuum of Group 3 structures, ranging in style from relatively open recumbent folds to mylonitic thrusts showing considerable ductile strain and transposition of earlier features both in cover and Archaean basement lithologies (Figure 9 a,b).

c) In contrast to Oravisalo, exposures in the Rääkkylä and Nieminen areas consistently show normal-limb vergence, and considerably less evidence for local and zonal intense ductile strain indicating the association of such structures with proximity to the basement-cover interface.

d) In the north-eastern part of the province, Group 3 structures are not so well defined; relatively open and symmetrical folds coaxial with L_3 at Sammallahti (Map 2 reference: Pyhäselkä 91.20.) are likely to be younger, F_5 features but this has not been satisfactorily demonstrated. The

metasediments adjacent to the Kettämö thrust zone have however
been invaluable in understanding the nature and geometry of F_3
structures only little affected by younger structures and aided
by numerous determinations of depositional younging direction.
Detailed mapping, presented at 1:5 000 scale on Map 3 has revealed
the existence of a major F_3 fold verging to the NNW whose develop-
ment was connected with that of the Kettämö thrust itself. Such
an association between F_3 recumbent folds and thrusts near ·the
basement/cover interface shows a striking analogy with the
tectonic sequence at Oravisalo. It is evident from Map 3 and
the accompanying cross-sections that the amplitude to wavelength
ratio of this structure increases rapidly northwards and in all
probability, the fold terminates or develops into a narrow zone
of high shear strain, possibly another thrust. The Kettämö thrust
may also have thus developed from a transposed, F_3 fold limb but
such evidence of intense ductile strain is not invariably present
- Figures 7 c and 11 d show F_3 hinge zones with both axial planar
S_3 and preserved sedimentary structures. It is therefore
reasonable to envisage this area in general as displaying an
anastomosing network of ductile faults gradational into recumbent
folds or as narrow zones of high strain enclosing lenticular, less-
deformed units. Such a pattern of deformation is evident in the
western part of Map 3, between Pitkänurmi and the farmstead
'Erkkila', where a narrow inverted limb increases in thickness
northwards, and thence contracts again into a narrow zone of higher
strain.

The Kettämö thrust zone may be extrapolated northwards with
confidence at least as far as the eastern margin of the Sotkuma
inlier - a proposal adumbrated in the studies of Wegmann (1928)
and Väyrynen (1939). The zone is expected to extend to the SE as
well but poor exposure and probable displacement across the later
Haapajärvi fault preclude tracing it. The postulated course on
Map 1 conforms to constraints placed by lithological changes in

the Ladoga area (Hackman 1933) and changes in metamorphic assemblage (Korsman and others 1984).

4.3.ii Structural domains and distribution in the Höytiäinen province

Although Group 3 structures tend to be the most dominant in this province, uncertainties arise in the field as to whether a given structure is of Group 3 origin, subsequently reoriented, or belongs to an entirely independent, younger deformation. This ambiguity is especially evident among F_3 and F_5 folds which can have similar style, attitude, vergence and axial planar fabrics (Figure 10). Distinction is most difficult in the Tikkala and Kukkupää lithofacies between Kumpu, Hammaslahti and Tikkala (respective map 2 reference: Niittylahti 98.27; Rauansalo 01.28; Onkamo 04.17.) and the problem is further compounded by the complexity of F_2 structures in this region. Nevertheless, four informal Group 3 domains have so far been identified in the province.

a) Below the Kettämö thrust and trending parallel to it, a zone of metapsammite about 3 km wide is comparatively well exposed and consequently F_3 structures have been relatively well defined. No younging criteria were found but $S_3 - S_0$ and minor fold relationships indicated exclusively F_3 upper-limbs except for a single, narrow inverted limb zone possibly only 100 m wide and passing along strike into a thrust (Map 2).

b) Eastwards from the Haapajärvi fault, numerous vergence and some younging determinations have been made, mostly indicating a position on upward-facing F_3 upper-limbs. However, exposures tend to be too sparse and scattered to permit confident definition of the scale of F_3 structures. This difficulty is exacerbated by a tendency for F_4 and F_3 upper limb minor structures to show congruent trend and vergence in horizontal exposures and common reorientation of F_3 structures into locally intense zones of F_5 deformation.

c) Exposures at the Hammaslahti mine and at Tikkala railway siding affirm the general observations of zonal and variable strain, usually with transposed inverted limbs and boudinage of massive layers in sequences with significant lithological contrasts (Figures 9 a and 10). Examples where transposition has occurred on normal-limbs of minor folds also exist (Figure 11 b).

A zone of intense shearing is exposed in the open cut at Hammaslahti mine and hanging wall collapse has been a persistent problem with underground stoping. This zone is part of a Group 3 fault system that effectively bisects the Höytiäinen province from N to S and which is evident as a zone of cataclasis on the KONTIOLAHTI 1:100 000 sheet (Huhma 1971). The southerly extension of this zone may coincide with the eastern margin of the Tohmajärvi volcanic complex; the TOHMAJÄRVI 1:100 000 sheet (Nykänen 1967) and limited data from this study indicate that this boundary is of shallow dip, consistent with an interpretation of the metabasite complex as being thrust over younger metasediments (Maps 2 and 4).

d) The region between Hammaslahti and Kiihtelysvaara has not been studied in detail but younging and fold vergence criteria in limited areas indicate that here too, inverted F_3 limbs are present. This is best seen in road cuttings at Kalliojärvi (Heinävaara 03.7 32.5) and Heinävaara (Heinävaara 09.8 39.4). The latter outcrop is in close proximity to Archaean basement but along most of this zone Proterozoic sediments still lie with original depositional contacts intact (Pekkarinen 1979a).

4.3.iii Group 3 structures in relation to quartz veins

Although quartz veins have been noted as very early
structures throughout North Karelia (Koistinen 1981, Park and
others 1984), their major development apparently occurred
between S_2 and S_3 formation, with the youngest showing distinct
geometrical relationships to S_3 (Figure 12c). It is likely
that these correlate in space and time with differentiation
layering induced by solution transfer processes, as advocated
by Beach (1977) and Stephens and others (1979). In view of the
heterogeneous nature of S_3 strain, and the existence of super-
posed structures, it is difficult to relate vein arrays uniquely
to a single set of structures. Therefore, some examples are
described, but without any attempts to quantify or partition
strains.

Very typical examples of syn-S_3 quartz veins are shown
in Figure 12c; such a development appears to relate to shear
zones and minor folds at a low angle to lithological layering
and S_2. A similar geometry is evident in Figure 12a,b and
Figure 37c where composite S_2-S_3 cleavage bisects the acute
dihedral angle between conjugate quartz veins. Despite these
symmetrical relationships, coeval and cogenetic development of
veins and foliation need not be assumed, nor should oversimplified
deductions concerning stress systems be made.

Irrespective of conditions of origin, these conjugate
veins can nevertheless reliably record subsequent deformation.
Those in Figure 12a,b and Figure 37c share many geometrical
features, even though the latter has been reoriented into a
zone of relatively high Group 5 strain. Both exposures show
upward younging and normal-limb vergence with respect to an S_3
foliation which bisects evidently conjugate quartz-vein arrays.
In both however, deformation of vein arrays as recorded by
buckling and boudinage suggests an opposite sense of vergence to

that derived from S_3 cleavage-bedding relationships. Similar observations have been made elsewhere and, in the absence of distinctively superposed structures, indicate a more complex nature than their geometry would at first suggest.

These complex relationships are also well expressed in a transposed, F_3 inverted-limb at Tikkala railway siding, depicted in Figure 10, and Figure 12 e-h. Although thin quartz veins are discordant to Group 2 and Group 3 structures in detail, yet their overall geometry is consistent with syn- or late- Group 3 structural development. Interpretation is further complicated however, by Group 5 structures being congruent with earlier fabrics at this locality.

4.3.iv Group 3 structures in relation to the
        Varpasalo pegmatite swarm

As metamorphic facies grade increases towards the south-west
so does the abundance of intrusions. Within the study area however,
no lithologies have sufficient neosome to be classified as migma-
tites, whether as the products of in situ partial melting, meta-
morphic differentiation or injection of exotic magma. Instead, they
occur sporadically as felsic pegmatites on the island Varpasalo and
on smaller islands in the channel between Varpasalo and Oravisalo.

Dykes are tabular or lenticular at outcrop scale and vary
from 5 cm to over 40 in thickness. They are generally concordant
with respect to lithological layering and early cleavages but
apophyses are locally discordant to S_2 and F_2 minor folds (Figure
13b). Some dykes are coarse-grained and massive, commonly with
graphic intergrowths of quartz and feldspar, as in Figure 13a. More
usually however, some indication of foliation exists parallel to S_3;
this locally imparts a resemblance to augen gneiss (Figure 13d).

The examples illustrated in Figure 13h may well be represent-
ative of general dyke relations, in that several lenticular bodies
of pegmatite are confined to distinct lithological horizons. This
may suggest an en echelon arrangement analogous to that of quartz-
vein arrays or alternatively they are related to adjacent minor
folds of characteristic F_3 form and vergence (Figure 13g,h).

Massive vein quartz is associated with a number of pegmatites;
lack of foliation may reflect ease of recrystallization of quartz
rather than indicate their age with respect to foliated dykes. In
Figure 13h it is unclear which of the pegmatite or vein-quartz is
the earlier but the occurrence of quartz at vein tips suggests it
developed after the pegmatite, during late-stage propagation of the
veins. In Figure 13f, there is no such ambiguity and massive quartz
forms septa-like veins approximately perpendicular to the S_3 folia-
tion, and are thus likely to be late S_3 structures.

4.3.v   Fabric elements and microstructures

The expression of microstructures depends to a large extent
upon lithology. Thus, in the monotonous metapsammites of the Savo
province, Group 3 textures are typically defined by coplanar biotite
growth and recrystallization but in the Höytiäinen province they
are more usually zonal crenulation cleavages. Only in major hinges
or inverted limbs such as in the south-east of the island of Varpa-
salo does S_3 show the differentiation layering so characteristic
of S_2. The general congruent geometry of S_2 and S_3, combined with
them both having formed during the same progressive metamorphism,
can clearly result in a composite schistosity, making it difficult
to identify the amount of grain growth directly attributable to S_3.
Hinges of F_3 microfolds tend to show substantial deformation of the
pre-existing S_2 fabrics, with only minor amounts of entirely new
and discordant axial planar grain growth (Figure 8c). In addition
to the alignment of biotite in higher strain zones, polygonal
mosaics of elongate quartz and plagioclase grains show a preferred
dimensional orientation, enhancing anisotropy. Where S_3 cuts them
at a high angle, S_2 biotite-rich foliae may preserve their overall
morphology while individual biotite grains tend to have totally
recrystallized with their c-axes perpendicular to S_3.

Figure 14 illustrates fabric modification with increasing
strain, following a sequence noted at both Oravisalo and in the
Kettämö thrust zone, culminating in highly laminar quartz-ribbon
blastomylonites (see Figure 14f). The first noticeable changes
are a general elongation of detrital quartz grains in the matrix
with an increase in sutured grain boundaries. This is accompanied
by a highly concordant alignment of layer silicates and develop-
ment of mortar texture in response to recrystallization at the
margins of larger detrital grains (Figure 14a).

Feldspar displays degradation into tiny sub-grains (Figure 14c) or the progressive appearance of small irregular domains of quartz as in Figure 14b. This latter is a common feature in plagioclase porphyroclasts in the thrusted granitoid gneisses at Oravisalo. Extensive recrystallization of quartz into polygonal, strain-free grains results in a true blastomylonitic texture but anisotropy is maintained by extreme flattening of former feldspar grains, which become anastomosing films with no strain-recovery recrystallization (Figure 14d).

The L_3 lineation is best developed in the Savo province metapsammites where it is defined by elongate knots of biotite crystals (Figure 9d,f). It is rare for the individual grains to show any dimensional alignment along this lineation and while the aggregates could represent syn-tectonic breakdown of a larger porphyroblastic phase such as garnet, or staurolite, there are no discernible relicts of these minerals. In metabasites and Archaean gneisses at Oravisalo, this L_3 lineation is recorded as a c-axis alignment of small amphibole grains or as a "rodding" effect defined by variations in amphibole content.

The same uncertainties apply to identifying microstructures in the Höytiäinen province as to major structures - frequently congruent $S_3 - F_3$ and $S_5 - F_5$ cleavage styles, attitudes and fold vergence reduce confidence in correlation. A penetrative mica alignment is characteristic where S_3 transposes lithological layering but in laminated metapelites S_3 is most typically a well differentiated crenulation cleavage, zonal and discrete in the terminology of Gray (1977) and Borradaile and others (1982). This is superbly displayed in the metapelites illustrated in Figure 11e where mica-rich cleavage domains crenulate and truncate quartz veins, depositional layering and S_2 (being the earliest tectonic fabric identified in the province). Though thicker than typical crenulation cleavage domains, being up to 7-8 mm in width, they appear otherwise analogous to those that Gray (1979) ascribes to solution transfer of quartz and feldspar from tightened buckle

folds. Accurate and consistent distinction between such similar Group 3 and Group 5 structures is only possible where S_5 is superimposed at a divergent trend, as in Figure 11 f or where post-S_3 biotite porphyroblasts are deformed by S_5 (Figure 18 h).

4.3.vi   Mineral parageneses and metamorphism in the Savo province

No mineral textures provide compelling evidence for separate metamorphic events connected with S_2 and S_3 formation, but in the Savo Province metapsammites, garnet porphyroblasts record the relationship of metamorphism to S_3 development. Figure 15 a,b shows post-S_2 garnet growth with porphyroblasts free from all but quartz inclusions. No relict earlier fabrics are discernible and in rims (Figure 15a)inclusions of any kind are uncommon. Rims also appear to have been more resistant to breakdown, giving rise to a very characteristic 'atoll' form in altered grains (Figure 15 c). Garnet breakdown has in places progressed even further towards total replacement, as shown in Figure 15 d,e. These pseudomorphs do not have any well-oriented fabric, in contrast to the S_2-S_3 enveloping foliation, and it is therefore inferred that garnet breakdown was a post-kinematic phenomenon with respect to S_3. Usually, pseudomorphs contain the following mineralogy:

biotite + quartz + plagioclase + fibrolite ± prismatic sillimanite ± opaque grains.

Each of these phases appears also to have been in microstructural equilibrium with garnet prior to its breakdown except for sillimanite, though this has evidently crystallized in equilibrium with biotite in the pseudomorphs (Figure 15 e).

Garnet breakdown is here interpreted as a prograde reaction, with the local formation of sillimanite during the post-S_3 thermal metamorphic peak. Figure 16 indicates that this was more prevalent in the south-west of the study area ; elsewhere garnets evidently coexist with the climactic assemblage thus :

quartz + Na-plagioclase + biotite + muscovite ± garnet ± hornblende ± calcite.

At Niva, at the eastern margin of the Savo province, mafic and calcareous metapelites show S_2 differentiation banding (defined by alternations of quartz and biotite with aggregates of very fine-grained muscovite) overgrown by biotite porphyroblasts, some of which are oriented in S_3. Most diagnostic of these lithologies is however, a distinct post-S_3 growth of rosettes and brush-like aggregates of actinolite and small (0.5 mm) porphyro-blasts of chloritoid (Figure 18 g). This assemblage is considered here to represent a somewhat lower peak of metamorphism at Niva than in the region of garnet breakdown, though still within the amphibolite facies. Whilst this widespread post-S_3 porphyroblast growth is in general taken to indicate restoration of a thernal equilibrium with respect to crustal depth, the foregoing observations also afford evidence of lateral thermal gradients across the study area. This is not attributed to late-orogenic differential uplift but to abundant post-S_3 and in part syn-S_4 pluton emplacement south-west of the study area, as established by Gaál and Rauhamäki (1971) and Halden (1982). Metamorphic grade is known to continue to increase in this direction from biotite-sillimanite-quartz assemblages to K-feldspar-sillimanite-garnet-cordierite gneisses; in the latter case, garnet and cordierite occur in mutual equilibrium (Korsman and others 1984). These studies by Korsman (1977) and Korsman and others (1984) have also clearly delineated thermal metamorphic domes over-printing early deformational features, again consistent with a post-S_3 metamorphic peak.

4.3.vii Mineral parageneses and metamorphism in the Höytiäinen province

The widespread preservation of depositional features immediately suggests a lower degree of metamorphism than is seen in the Savo province, with less recrystallization or grain growth and evident stability of much of the detrital mineralogy

(Figures 42,43). In the Höytiäinen province, metapsammites, including metagraywackes are devoid of garnet, except adjacent to the Archaean basement (Nykänen 1971a, Campbell and others 1979, Pekkarinen 1979a). Typical syn-S_3 assemblages appear to be of

quartz + muscovite + biotite $\pm$ feldspar $\pm$ sulphide
$\pm$ graphite $\pm$ calcite.

At Reijola (Map 2 reference: Niittylahti 91.38), several kilometers east of the Kettämö thrust zone, syn-S_3 porphyroblasts resembling cordierite were found in quartz metapsammites, but nowhere else was it identified with confidence.

Metapelites in the eastern part of the province frequently show spectacular post-S_3 porphyroblasts, up to 25 cm in length, of staurolite, andalusite and less frequently garnet (Nykänen 1971a, Campbell and others 1979, Pekkarinen 1979a). This probably results from local compositional peculiarities, but Nykänen (1971a) also considers it likely that the porphyroblasts grew during contact metamorphism surrounding late tectonic granitoid intrusions, particularly near Tohmajärvi. The mineralogy is recognized by Korsman and others (1984) as recording a distinct metamorphic milieu, the boundary of which appears to correspond to the postulated south-easterly continuation of the Kettämö thrust zone.

Between Hammaslahti and Tikkala, metagraywackes sporadically contain chlorite-muscovite-biotite pseudomorphs after staurolite; this replacement was evidently post-S_3 but how much later is not clear (Figure 18e). The sample illustrated is also very rich in tremolite which has overgrown S_3 fabrics and forms rosettes and symplectic intergrowths with quartz. Dimensional alignment of deformed staurolite pseudomorphs within the axial planes of F_3 folds suggest that staurolite initially crystallized prior to or during S_3 formation.

Between the Kumpu-Hammaslahti-Tikkala zone and the Kettämö thrust zone, metapelites contain biotite both within S_3 and as a later unoriented porphyroblastic growth. This again provides evidence for the thermal metamorphic peak being younger than S_3 development, but preceding S_5 formation, which is seen in general, to deform porphyroblasts (Figure 18h).

The same conclusions are supported by the presence of post-S_3 tremolite porphyroblasts in calcareous metapelites at Mulo (Map 2 reference: Niittylahti 90.36). The spectacular, unoriented crystals overgrow and enclose the earlier differentiated fabric consisting of the assemblage

microcline + calcite + graphite + pyrrhotite $\pm$ chalcopyrite $\pm$ pyrite $\pm$ sphene $\pm$ sphalerite $\pm$ chlorite $\pm$ quartz $\pm$ plagioclase.

These particular graphitic schists are in fact some of those from which derived the postulate of grain growth inhibition in the presence of graphite (Eskola 1963). While conceivable that calcite or dolomite growth was impeded due to buffering between carbonate and reduced carbon, the examples shown in Figure 18b,c,d do not appear to indicate inhibition of either tremolite porphyroblastesis or $S_2 - S_3$ differentiation layering. On the other hand the low silica and mica content may have affected metamorphic grain growth. Furthermore it is likely that in addition to H_2O, other molecular fluid phases would have been important, such as CO, CO_2, O_2, H_2S, S_2 and CH_4, creating the potential for highly complex and buffered reactions. There are textural indications in support of this, such as the K-feldspar rims surrounding pyrite and pyrrhotite grains (Figure 18c), K-feldspar overgrowths resembling that in drusy cavities (Figure 18d) and K-feldspar 'shadows' fringing pyrrhotite grains that have evidently crystallized within syn-S_3 or younger calcite veins.

4.3.viii Concluding remarks on Group 3 fabric development

Differences between Group 1, 2 and 3 fabric development and
expression are ascribed to variations in stress distribution,
strain rates and changing rheologies of rocks as they deformed
under conditions of prograde metamorphism. Deformation itself
could have indirectly influenced mineral assemblages by increasing
lithostatic load and thus rate of pressure increase with respect
to temperature. It is further surmised that whatever thermal
anomalies may have existed during sedimentation or instigated
deformation, a period of apparently rapid basin filling followed
by essentially overthrust deformation would lead to depressed
geothermal gradients. Restoration of a 'normal' crustal P-T
regime was evidently attained after S_3 and hence after the bulk
of crustal 'thickening' or 'loading'. It may be therefore that
the presently observed microstructures derive from essentially
continual recrystallization and modification of progressively
evolving, dynamic fabrics. The products of certain aspects of
deformation survive, for example S_2 differentiation layering or
S_3 blastomylonites because they formed by distinctive processes
other than the small scale intergranular adjustments such as
grain boundary recrystallization or diffusion.

FIGURE 5 : GROUP 1, GROUP 2 AND GROUP 3 STRUCTURAL FEATURES

Diagram gives location of major F_2 hinge zones and inverted limbs and major F_3 folds for which inverted limbs are of more restricted development. Trajectory of L_3 is also shown and equal area plots 1 - 9 indicate attitudes of Group 1 - Group 3 structures from representative domains affected to varying degrees by reorientation during younger deformation. These domains do not exactly coincide with the reference localities upon which structural sequence is based (Section 3).

Grid coincides with 1:20 000 sheet boundaries (sheet area = 100 km^2); names of map sheets are underlined and in lower case.

DOMAINS RELATING TO EARLY GROUP OF STRUCTURES S_1, S_2, S_3 AND LINEAR ELEMENTS

Schmidt (equal area, lower hemisphere) projections:

· poles to lithological and differentiation layering (S_0, S_1, S_2) and S_0 continuous cleavage
· poles to gneissosity and crosscut bending in basement
✳ poles to syntectonic pegmatite dykes
✳ shape fabric lineation, dominantly L_3
• late F_3 fold hinges, generally not parallel to L_3
General symbols:
⋯ undifferentiated Archaean basement
⌃⌃ established or inferred thrust surface, dominantly S_5

⊞ inverted limb or large zone of major F_3 fold
↗ normal limb of recumbent F_3 fold
⌇ inverted limb of F_3 fold
⫽ trace of composite S_0, S_1, S_2, or S_3 layering
⫽ depositional younging indicates upward facing F_3 fold
⊙ F_3 superimposed on undisturbed or normal limb of earlier structures
◐ depositional younging indicates downward facing F_3 fold
✕ inverted sequence prior to F_3 development
↗ trend of L_3 shape fabric lineation

ORIVESI SHEAR ZONE

VAARASALO–
HYPÖNNIEMI

Varposalo
Hyponniemi

Lappviluoto
Ninakkosaari
Uukoniahti
Tulqumiemi

ORAVISALO–
RÄÄKKYLÄ

Oravisalo
Rääkkylä

ORAVISALO
BASEMENT

Rasivaara
Tikansaari

NIEMINEN

SAVO   PROVINCE

Pohäselkä
KETTÄMÖ

Kompakka
Sammakki

KETTÄMÖ THRUST ZONE
Niva
Niittylahti

HAAPAJÄRVI FAULT
Haapajärvi township
Hammaslahti township

SUHMURA

KUMPU
Kallojärvi

Onkamo

Rauansalo
HAMMASLAHTI
TIKKALA
Hammaslahti mine

HAMMASLAHTI MINE–

HÖYTIÄINEN   THRUST   ZONE

TOHMAJÄRVI THRUST ZONE

PROVINCE

Vaikeasuo

Vuosimo

Kiihteysvaara

Hendvaara

Tohmajärvi

FIGURE 6:   GROUP 1 AND GROUP 2 STRUCTURES IN THE SAVO PROVINCE

a   Broad, diffuse S_2 differentiation banding at high angle to relict
     depositional layering (S_0). S_2 is weaker in thinner depositional
     unit (below compass) but a slightly oblique S_1 schistosity is
     well expressed, defined by anastomosing biotite-rich foliae.
     Small dark blebs in metapsammite at top right are knots of biotite,
     elongate in L_3 (Oravisalo 79.8 12.8).

b S_1 biotite-rich foliae and quartz veins, subparallel to
     depositional layering and displaying intrafolial F_1 fold at lower
     left.   Faint S_2 differentiation banding is evident in upper,
     massive unit, almost perpendicular to S_0 (Oravisalo 79.1 12.9).

c   Well-defined S_2 axial planar to F_2 minor folds that have deformed
     depositional layering and local S_1 biotite-rich foliae
     (Oravisalo 79.8 12.7).

d F_2 minor folds with diffuse, fanning S_2 foliae in hinge zones
     Earlier S_1 fabric is evident, modifying depositional layering
     (S_0). L_3 is seen as small blebs of biotite, particularly at top
     left (Oravisalo 79.8 12.7).

e S_2 differentiation layering axial planar to F_2 fold.   Form of
     composite S_0-S_1 surface is indicated by lenticular calc-silicate
     lithology (Oravisalo 77.8 11.5).

f   Detail of fold in 'e' showing nature of S_1 biotite foliae slightly
     oblique to S_0 (Oravisalo 77.8 11.5).

g   Discrete biotite foliae defining S_2 differentiation in mesoscopic
     fold hinge.   Elliptical pods at bottom right are presumably
     transposed, disrupted relics of a calcareous layer or concretion
     (Oravisalo 72.1 17.3).

h S_2 axial planar to F_2 folds deforming lithological layering and
     gently warped about open F_7 folds.   Still younger fracture, or thin
     vein arrays occur but have not been classified in the numbered
     structural sequence (Varpasalo 67.7 20.2).
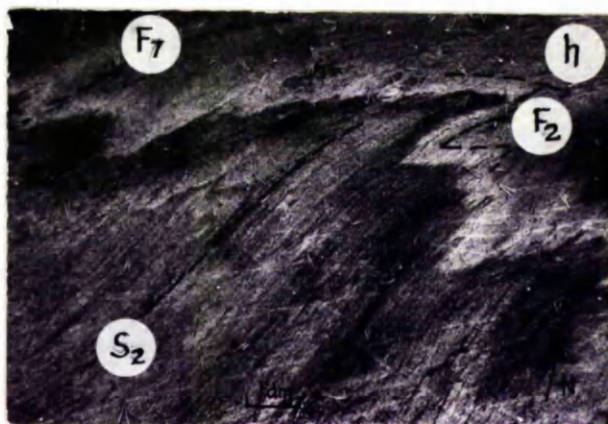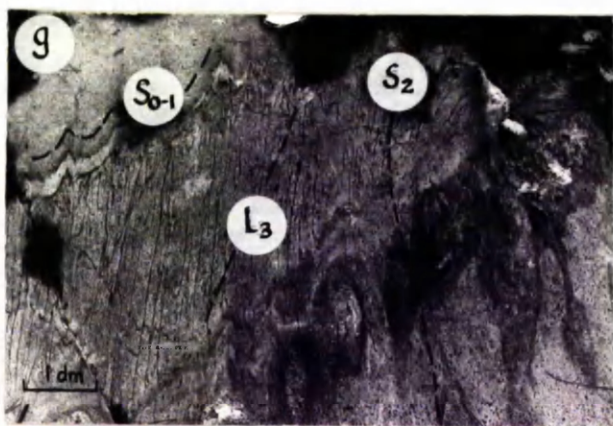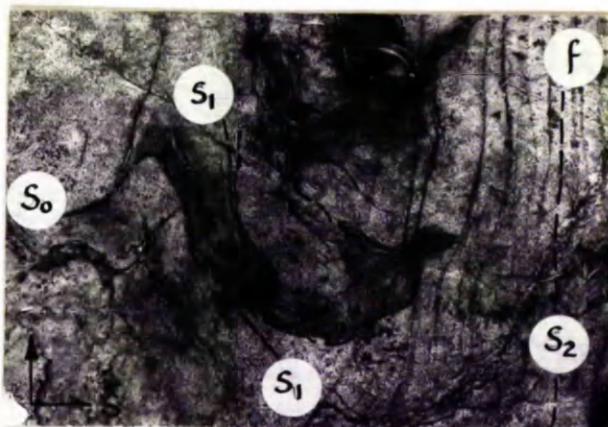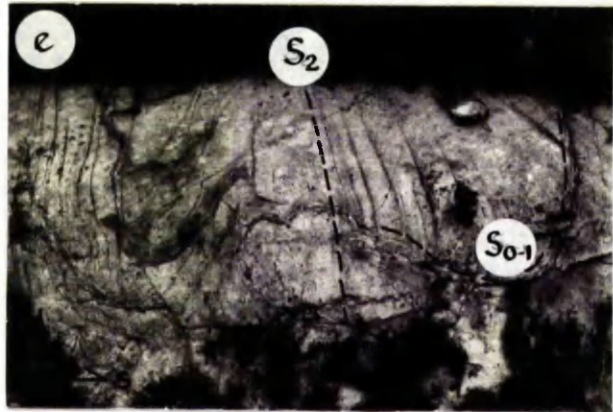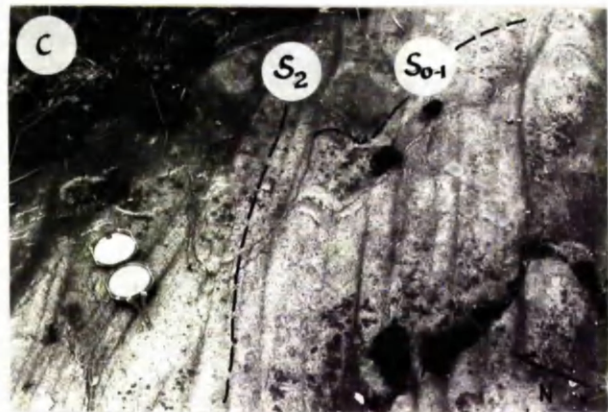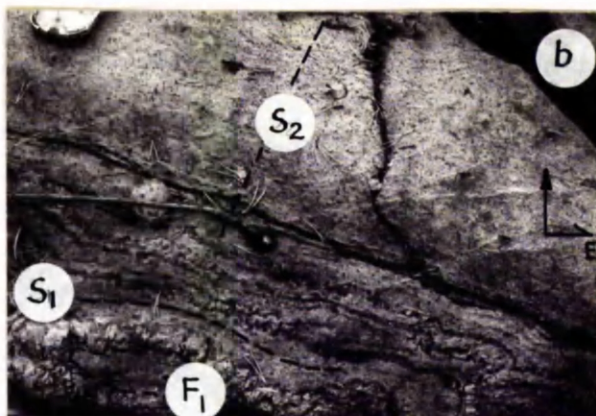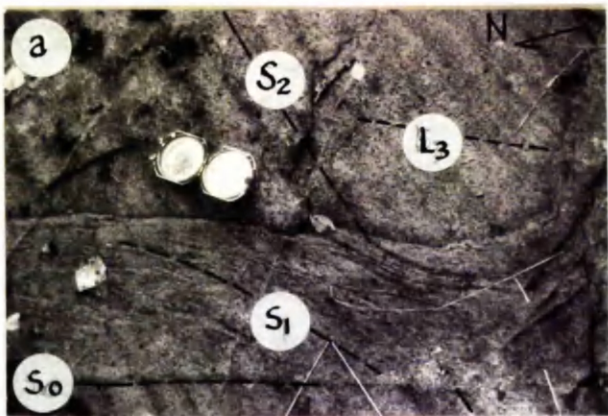
FIGURE 7: GROUP 2 STRUCTURES IN HÖYTIÄINEN PROVINCE

a   Anastomosing S_2 differentiation layering in metapelite, crenulated
     by intensely developed S_3 (Niittylahti 94.8 37.3).

b   Interference pattern produced by F_3 and steep crenulation cleavage
     S_3 superimposed upon F_2 fold (hinge almost within plane of
     exposure) (Niittylahti 94.8 37.3).

c   In hinge zone of recumbent F_3 fold, S_3 is only expressed as faint
     light-coloured streaks at right. Development of S_2 differentiation
     layering has been lithology-dependent and fails to cross from the
     metagraywacke at left into the overlying bed (younging at right
     indicated by basal load casts) (Pyhäselkä 95.2 27.4).

d   Metapsammite layer transposed in minor F_2 fold. S_3 zonal
     crenulation cleavage is expressed in intercalated metapelite and
     shows opposite vergence with respect to S_0. S_3 is in turn
     crenulated by weaker N-trending S_5 (Onkamo 01.2 17.5).

e   With S_3 cleavage almost perpendicular, earlier deformation can
     be identified by boudinage and transposition of sedimentary
     structures in addition to S_2 fabric development (Rauansalo
     00.5 24.1).

f S_2 differentiation layering and intrafolial F_2 folds more
     distinct in quartz-rich layers than in interbedded metapelite
     (Valkeasuo 17.4 18.9).
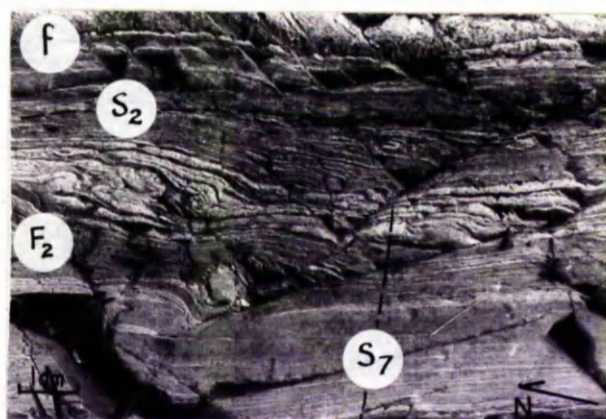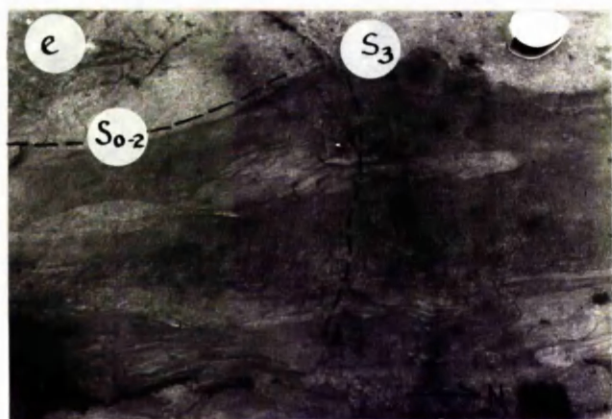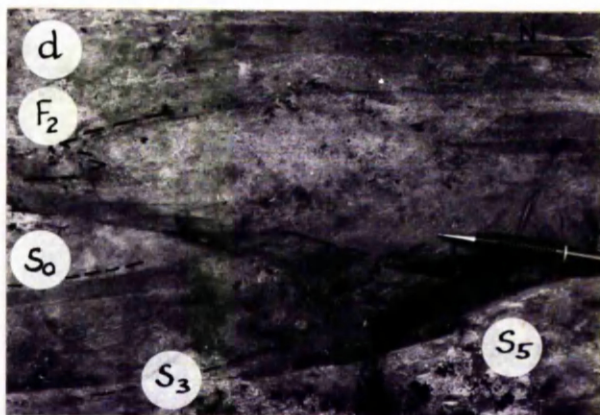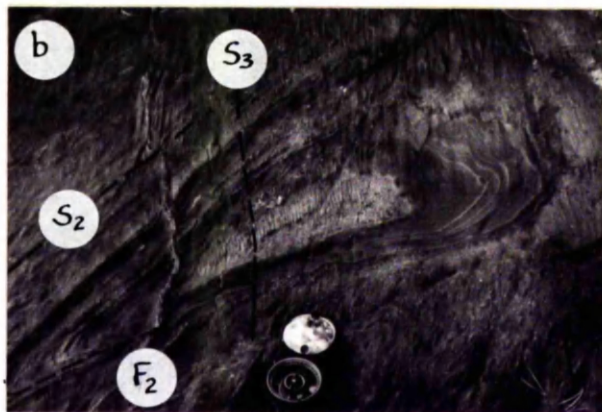
FIGURE 8: GROUP 2 MICROSTRUCTURES

Scale bar represents 1 mm

a S_2 differentiation layering, defined by alternating quartz-
    rich and micaceous domains. Small mica grains oblique to this
    layering do not necessarily belong to a younger, cross-cutting
    fabric but might have grown with S_2 in such an attitude, analogous
    to larger scale cleavage refraction in competent units
    (Välkeasuo 17.4 18.9). PPL.

b   Metagraywacke with deformed quartz grains and mica-rich,
    discontinuous S_2 cleavage domains (Niittylahti 98.2 35.9). PPL.

c S_2 differentiation layering deformed into F_3 microfold. S_3
    axial plane contains only a little new grain growth (Oravisalo
    75.7 15.8). PPL.

d S_2 or S_3 differentiationlayering with anastomosing biotite-
    muscovite domains subparallel to depositional layering. In
    darker, fine-grained band S_5 crenulation cleavage is distinct
    (Heinävaara 03.7 32.5). NO.

e   Composite S_2 - S_3 differentiation layering with sharply defined
    mica-rich cleavage domains. F_5 microfolds have an axial planar
    growth of fine-grained mica that differs from the typical S_5
    crenulation cleavage domain (Niittylahti 94.4 31.3). PPL.

f   Sulphide grains are oblique to but not necessarily younger than
    composite S_2-S_3 continuous cleavage in metapelites. Their
    distribution appears to have located the nucleation of F_4 or F_5
    microfolds and discrete, zonal crenulation cleavage domains
    (Niittylahti 91.3 31.8). NO.
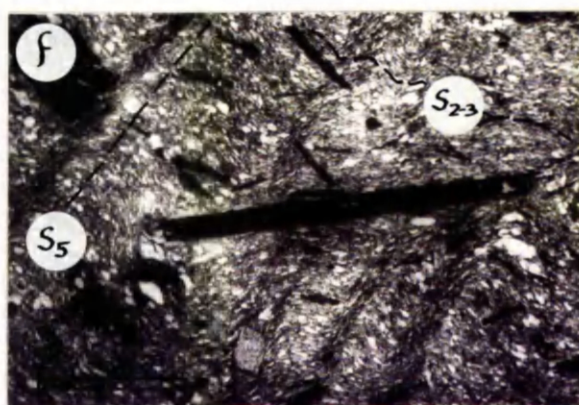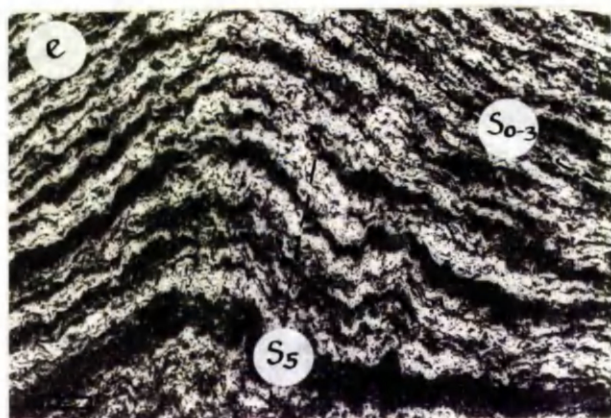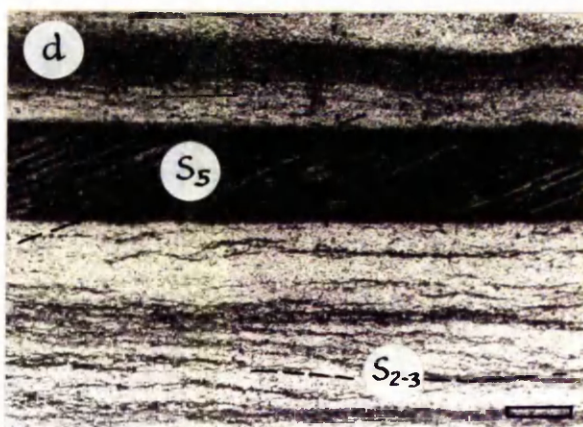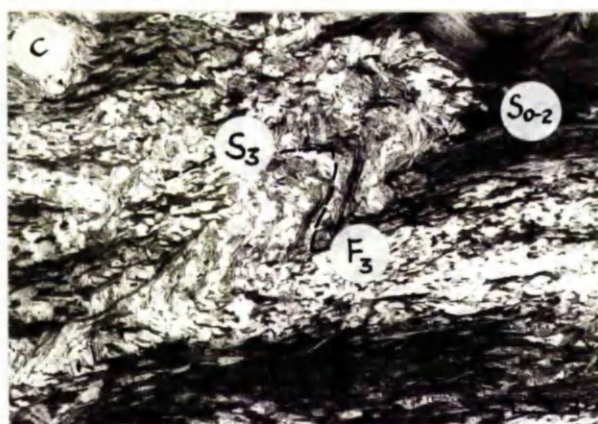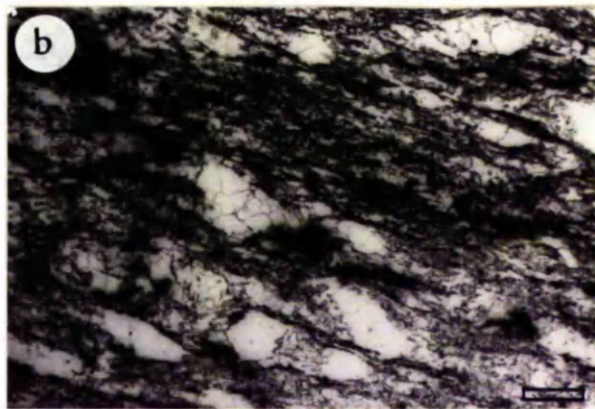
FIGURE 9:   GROUP 3 STRUCTURES IN SAVO PROVINCE

a   Disharmonic recumbent F_3 folds deforming metapsammite, with axial planar S_3 most conspicuous in more fissile, metapelitic intercalations (Oravisalo 74.2 17.5).

b   Isoclinal recumbent F_3 fold associated with underlying ductile zone deforming S_{0-2} (Oravisalo 73.9 17.5).

c   Relict depositional layering with subparallel S_2 biotite foliae and quartz veins deformed into a typical, disharmonic F_3 fold (Oravisalo 70.6 19.8).

d   Minor F_3 fold showing inverted-limb vergence and L_3 shape fabric lineation perpendicular to fold hinge (Oravisalo 74.7 18.8).

e   Minor F_3 folds oblique to lithological banding and subparallel S_3 foliation, with incipient transposition of calc-silicate lenses (Oravisalo 70.6 19.8).

f   Irregular, diffuse leucosome with thin seams rich in biotite porphyroblasts and sulphide minerals, transecting the main gneissosity (S_{I-II}) (Oravisalo 75.7 15.8).
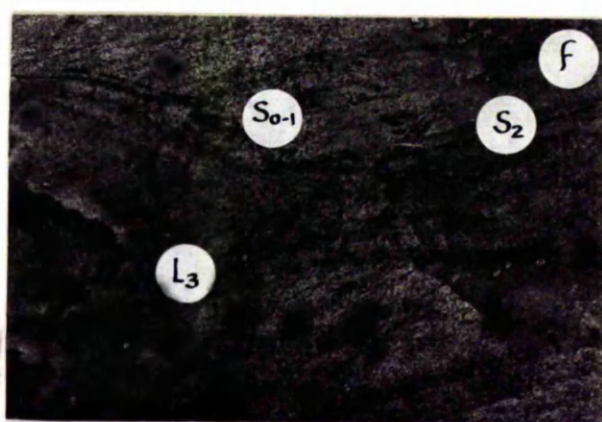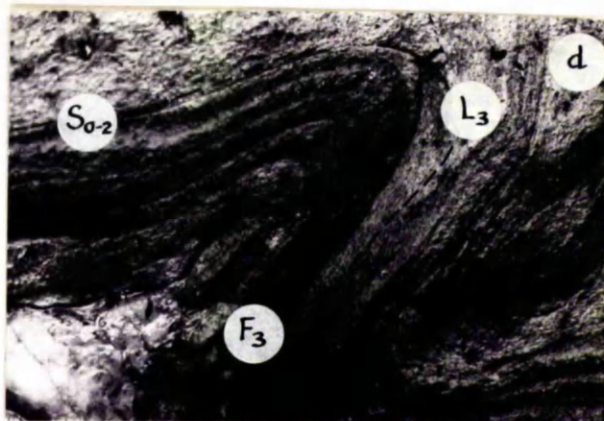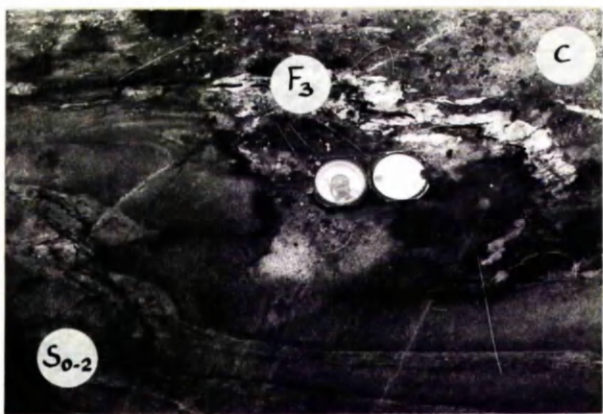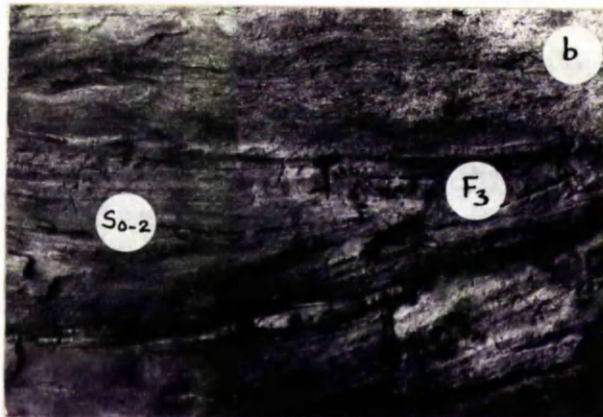
FIGURE 10 : STRUCTURES AT TIKKALA RAILWAY SIDING

a   Outcrop map showing complex Group 2 and 3 deformation including
     tight to isoclinal folds, boudinage of metapsammite units
     and transposition, subsequently disrupted by subparallel
     Group 5 structures. Some veins relate to this later event
     but the majority are probably late-stage Group 3 structures.

b   Diagrammatic example showing how it is possible to dis-
     criminate early folds if hinges remain at a high angle to
     superimposed structures.

c   Diagrammatic example of the situation at Tikkala illustrating
     the difficulty in discriminating between new structures and
     reorientation of earlier structures.

STRUCTURAL FEATURES

TIKKALA RAILWAY SIDING

a

metagraywacke

N

5

meters

5

10

F_5

boudins

shear zones

S_2

S_5

quartz veins

F_3

metapelite

F_3

b

N

N

c

schematic depiction
of reorientation
of Group 3 structures
by Group 5 structures
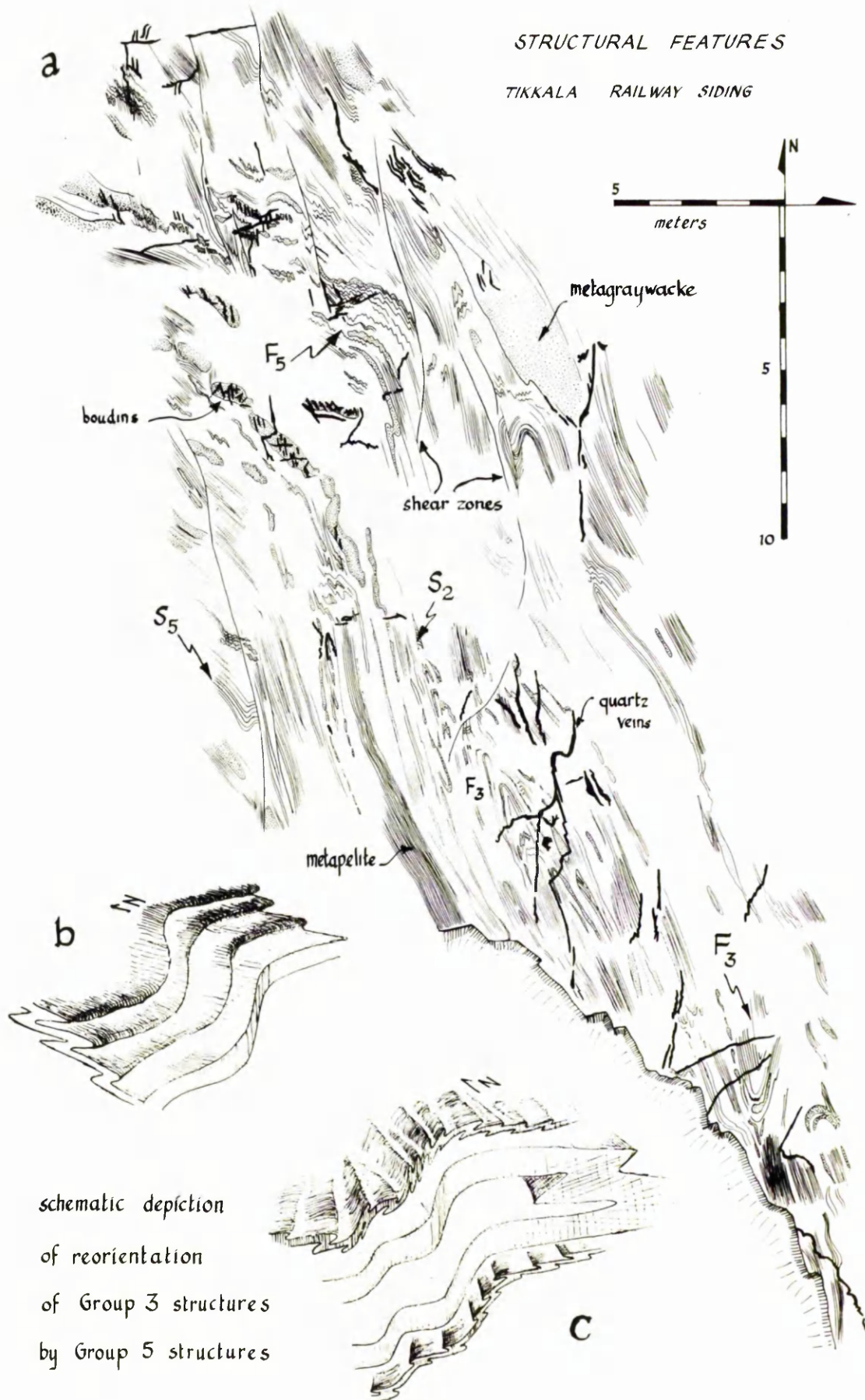
FIGURE 11:   GROUP 3 STRUCTURES IN THE HÖYTIÄINEN PROVINCE

a   Inverted limb of F_3 fold with transposition of load casts at base of graded metaturbidite (Onkamo 04.6 17.5).

b   Heterogeneous strain in metagraywacke and metapelite, recorded by localized S_3 truncation and transposition of F_3 or earlier minor fold, with later S_4 crenulation cleavage and quartz vein perpendicular to S_5 (Rauansalo 02.4 28.2).

c   Isoclinal F_3 folds with prominent axial planar S_3, warped and crenulated by NE-trending S_6 or conjugate S_4 crenulation cleavage (Niittylahti 92.5 35.2).

d   Axial planar S_3 differentiation layering in feldspathic metagraywacke, with transposition of basal load casts (Pyhäselkä 95.5 27.2).

e   Group 2 quartz veins subparallel to depositional layering buckled and truncated during S_3 zonal crenulation cleavage development (Niittylahti 94.8 37.3).

f   Thin, discrete S_3 crenulation cleavage domains separating lithons with relict earlier, presumably S_2 cleavage. At bottom left weaker, S_5 crenulation cleavage is visible. Approximately natural size (Niittylahti 94.8 37.1).
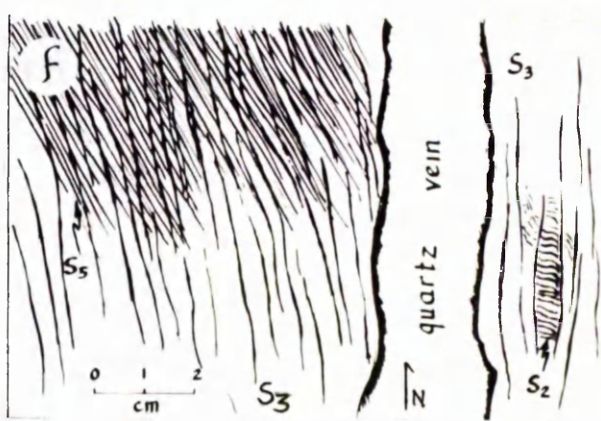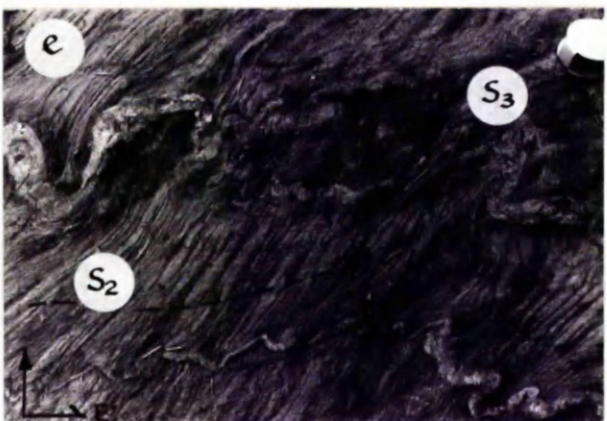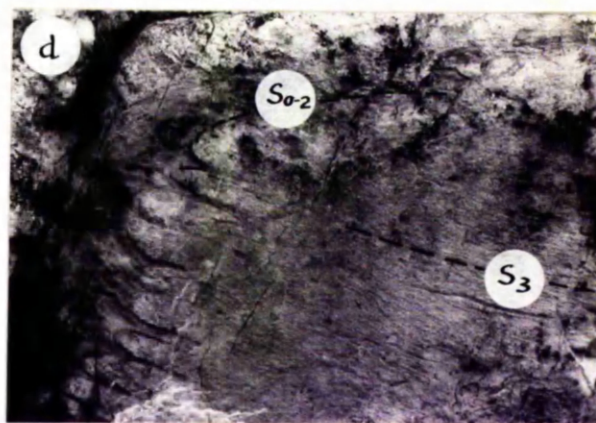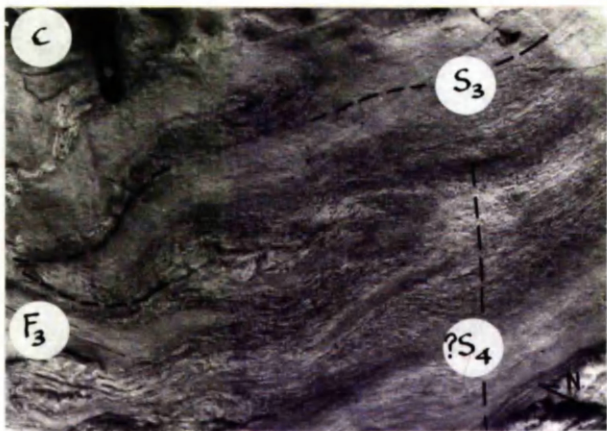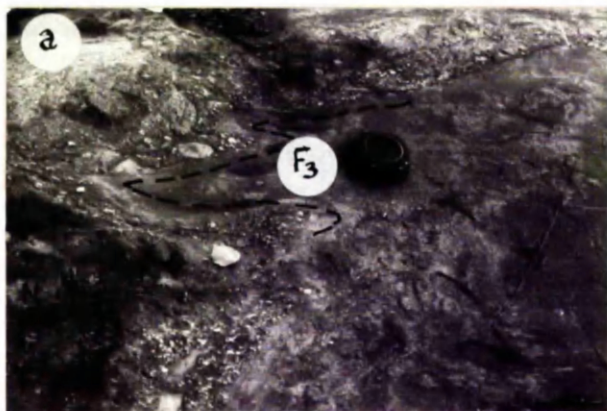
FIGURE 12:   QUARTZ VEINS IN RELATION TO GROUP 3 STRUCTURES

a,b   Probable syn-S_3 or earlier conjugate vein array, with acute
      bisectrix subparallel to S_3. Left-dipping veins are locally
      buckled and right-dipping veins tend to be discontinuous,
      suggesting continued deformation after vein development
      (Pyhäselkä 94.3 28.8).

c   Typical angular, discontinuous quartz-feldspar vein development
      within S_3 ductile thrust or shear zone (Oravisalo 74.6 19.3).

d   Two generations of quartz veining: on either side of the
      compass veins coeval with or antedating S_3 occur. The thicker is
      conformable with S_3 and boudinaged and to its left occurs a
      mesoscopic F_3 fold. All are truncated by an undeformed, almost
      perpendicular vein that apparently consists of a series of
      en echelon segments. These are perpendicular to both L_5 and
      S_5 (Rauansalo 01.8 28.6).

e,f   In metagraywackes, syn- or post-S_3 vienlets show both increased
      thickness and angle with respect to composite S_0-S_3, compared
      with metapelites (Onkamo 04.6 17.4).

g,h   Large pre- or syn-S_3 boudins traversed by conjugate quartz veins
      whose geometry is consistent with either late S_3 or syn-S_5
      development (Onkamo 04.6 17.4).

a

clasts

quartz vein

S_3

N

b

c

S_{0-2}

S_3

d

F_3

N

S_{0-2}

e

metapelite

metagraywacke

quartz vein

S_2

f

S_{0-2}

g

normal F_3 limb

S_3

inverted F_3

quartz

graywacke boudin

1 m

N

h
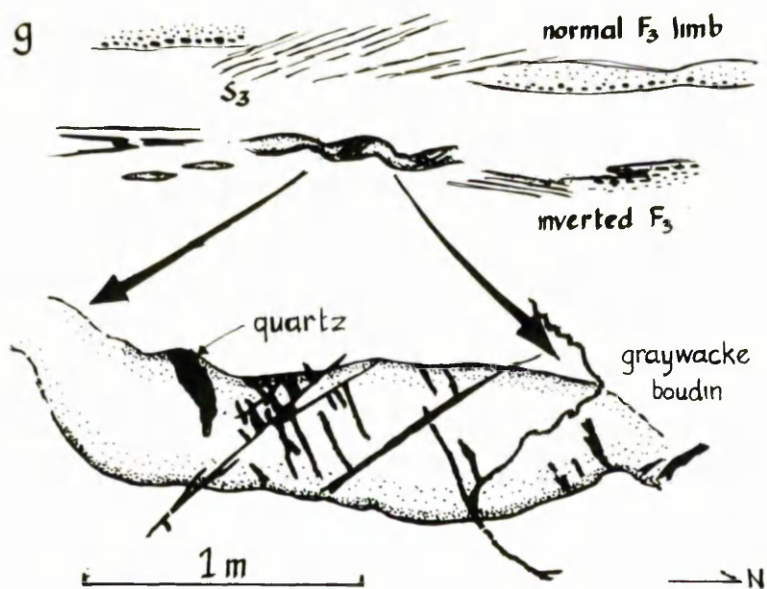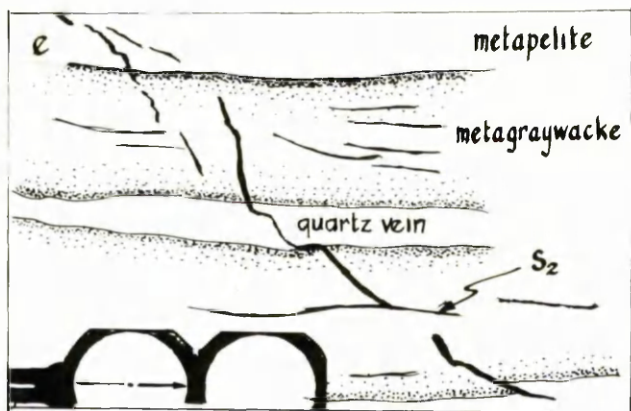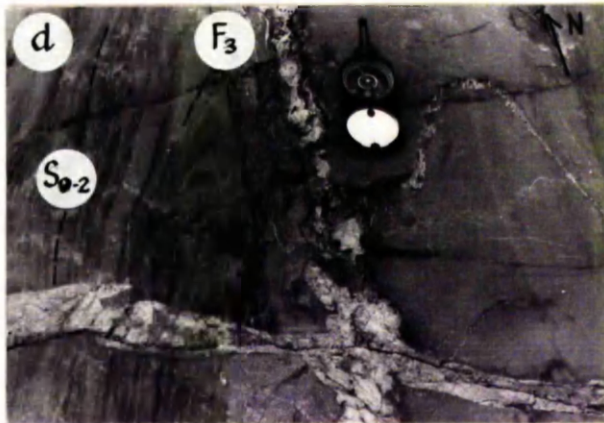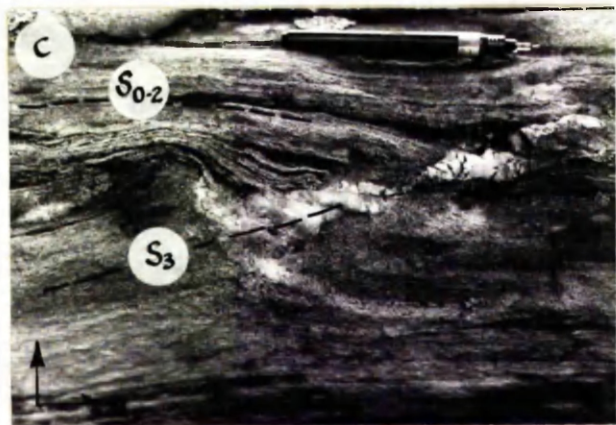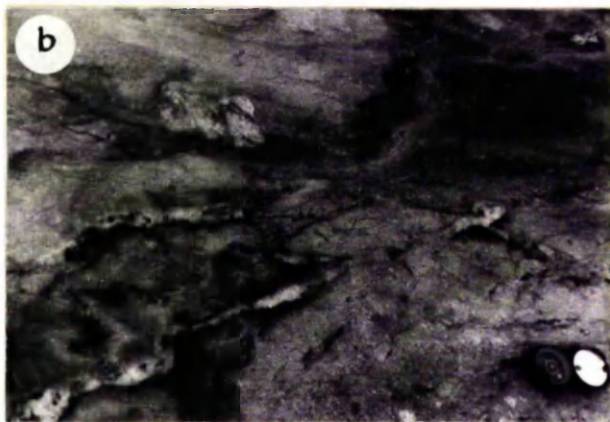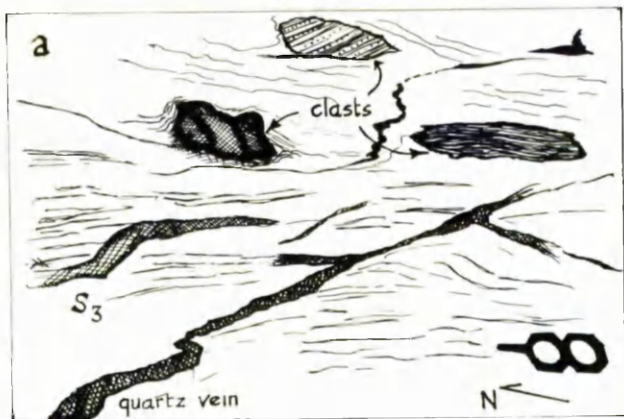
FIGURE 13:   VARPASALO PEGMATITE IN RELATION TO GROUP 3 STRUCTURES

a   Porphyritic, unfoliated apophyses transecting S_2 differentiation layering but possibly syn-tectonic with respect to F_3 folding (Oravisalo 72.7 12.7).

b   Pegmatite dyke   truncates S_2 differentiation layering axial planar to folds deforming S_0-S_1 (Oravisalo 72.8 12.8).

c   Pegmatite at right slightly discordant to lithological layering and S_2 in host rock but foliated itself by subparallel S_3, attesting to syn-tectonic intrusion (Oravisalo 72.8 13.2).

d   At top left, further example of syn-tectonic S_3 pegmatite with almost augen-like texture developed in some feldspar phenocrysts (Oravisalo 72.7 13.4).

e   Symplectic intergrowth of quartz, muscovite and plagioclase in syn-S_3 pegmatite (Oravisalo 72.8 12.8).

f S_3 foliation in pegmatite, with subsequently developed quartz veins approximately perpendicular (Oravisalo 72.7 13.4).

g   Pegmatites syn-tectonic with respect to S_3 and related to F_3 minor folds (Oravisalo 72.7 13.4).

h   Pegmatite lenses concordant with and probably coeval with F_3 minor folds. Associated quartz veins may have developed with pegmatite intrusion - the geometry of the upper lense suggests quartz is younger but the lower middle lense has the morphology of pegmatite forming in later-stage propagation (Varpasalo 67.2 11.2).

a — S_2

b — S_{0-1}, S_2

c — S_{0-2}, S_3

d — S_3, 1 dm

e — K-feldspar, 1mm, plagioclase, muscovite, quartz

f — S_{2-3}, pegmatite, vein quartz, 1m.

g — F_3, S_{0-2}, pegmatite, 1 dm.

h — metapsammite, F_3, S_{0-2}, vein quartz, 1m.
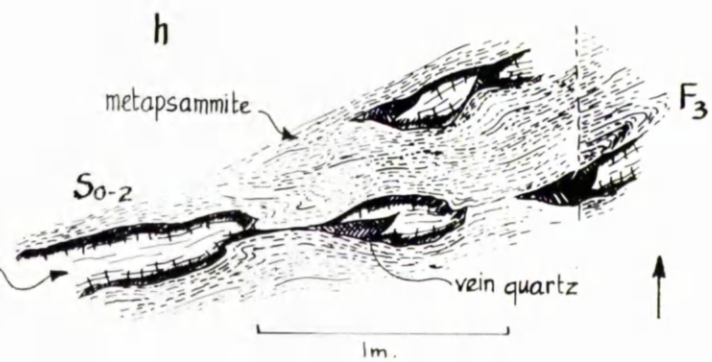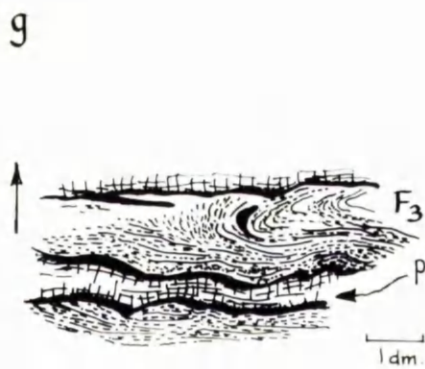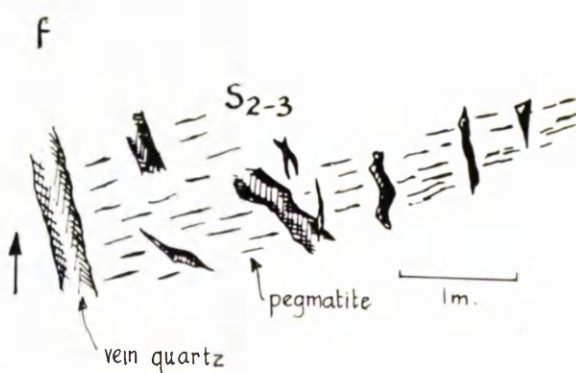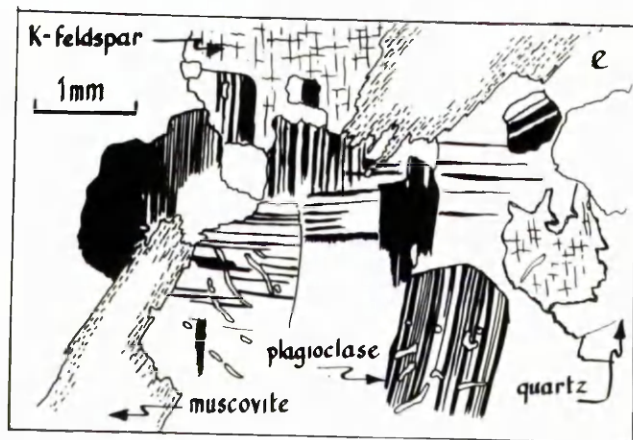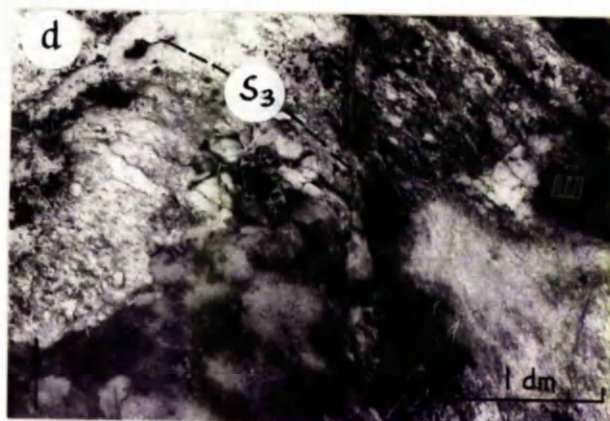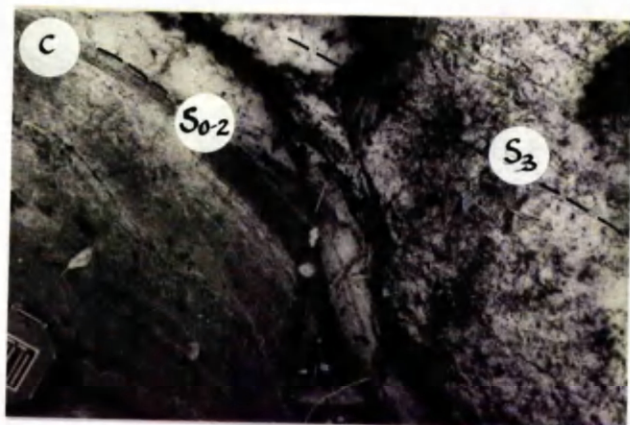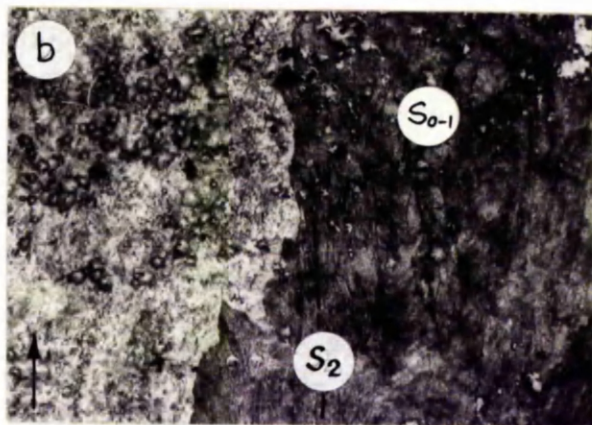
FIGURE 14: MYLONITIC GROUP 3 STRUCTURES

Scale bar in photomicrographs represents 1 mm

a   Relict detrital feldspar (mottled grain at left) and quartz
     grain showing sutured grain contacts with recrystallized
     quartzose matrix; finer quartz grains are elongate and subparallel
     to S_3 as defined by small muscovite grains (Pyhäselkä 94.6 27.4).
     NO.

b   Irregular domains of quartz in plagioclase, characteristic of
     incipient S_3 mylonite development in basement granitoids
     (Oravisalo 75.7 15.8). NX.

c   Muscovite blades aligned parallel to S_3 in mylonitic Archaean
     gneiss. Twinned plagioclase has deformed along zone of sub-
     grain development and conjugate microkinks (Oravisalo 73.4 15.4).
     NX.

d   Advanced stage of high-strain deformation in arkose. Mosaic
     of quartz has recrystallized from larger grains, deformed as
     quartz ribbons. Feldspar grains have been attenuated parallel
     to S_3 but sub-domains have not recrystallized into a new mosaic
     (Oravisalo 94.9 28.1). NO.

e   Anastomosing mylonitic seams have effectively obliterated relict
     clast boundaries in arkosic metapsephite (Pyhäselkä 94.9 28.1).

f   Laminar, quartz-ribbon mylonite having completely destroyed any
     evidence for depositional fabric (Pyhäselkä 94.9 28.1).
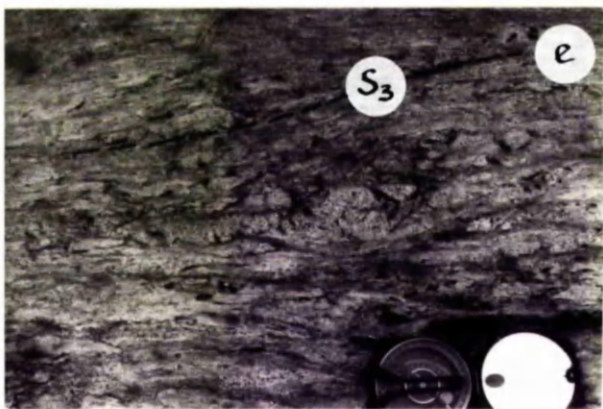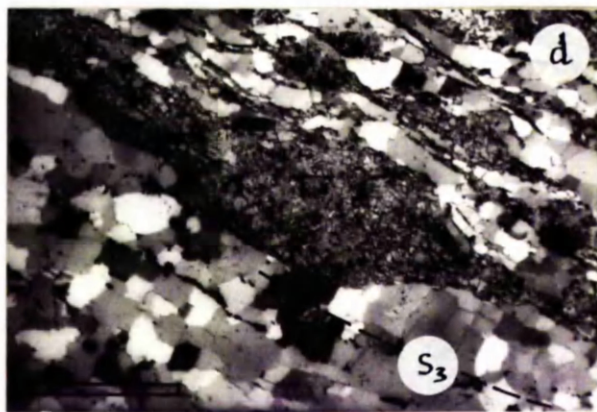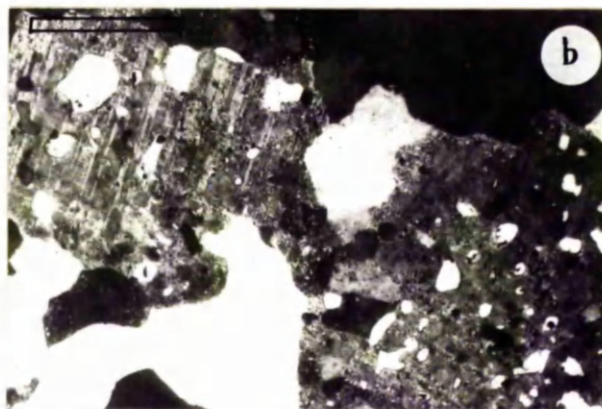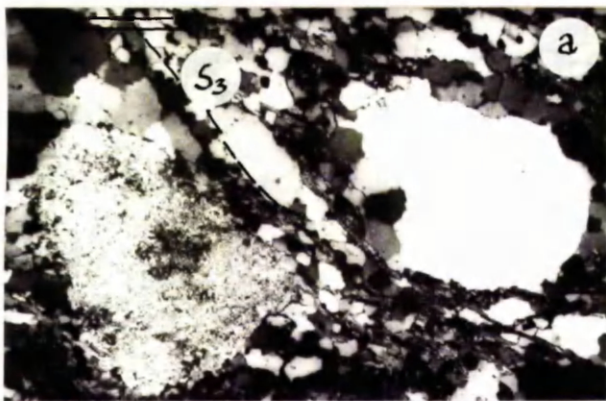
FIGURE 15:   GARNET PORPHYROBLASTS IN RELATION TO S_3

Scale bar represents 1 mm
a - e: Progressive stages in prograde garnet breakdown.

a   Garnet and biotite porphyroblasts evidently overgrowing S_2
        differentiation layering but deformed during S_3 formation
        (Oravisalo 75.7 15.8).   PPL.

b   Quartz-biotite-plagioclase mosaic presumably in equilibrium
        with garnet.   Quartz inclusions are not considered to reliably
        indicate the nature of any fabric existing prior to garnet
        formation (Varpasalo 67.7 15.7).   PPL.

c   'Atoll garnet' whose internal texture is interpreted as a post-S_3
        prograde alteration to quartz + biotite + plagioclase $\pm$
        fibrolite (Varpasalo 67.7 15.7).   PPL.

d   Complete pseudomorph of garnet porphyroblast indicated by
        deflexion of enclosing S_2 - S_3 fabric and unoriented biotite-
        plagioclase replacement growth (Oravisalo 75.7 15.8).   PPL.

e   Post-S_3 pseudomorph of garnet porphyroblast with abundant
        biotite, fibrolite and a little prismatic sillimanite.
        Replacement biotite has formed with c-axis nearly perpendicular
        to that of S_3 biotite (Varpasalo 67.7 15.7).   PPL.

f   Poikiloblastic garnet in syn-S_3 quartz - feldspar - muscovite
        pegmatite. S_3 defined by dimensional alignment of quartz and
        muscovite (Oravisalo 72.2 13.7).   NX.
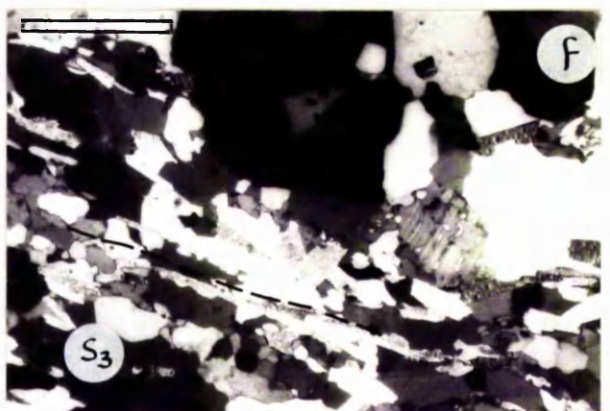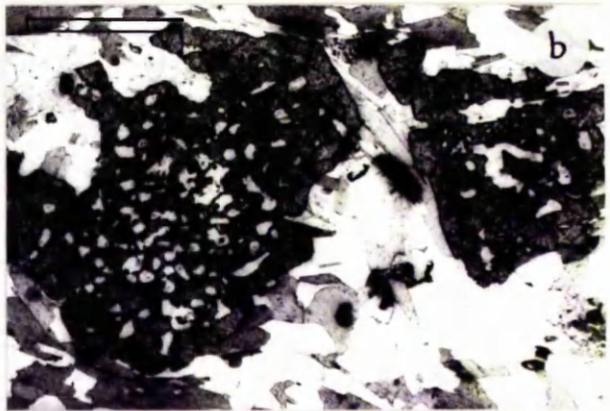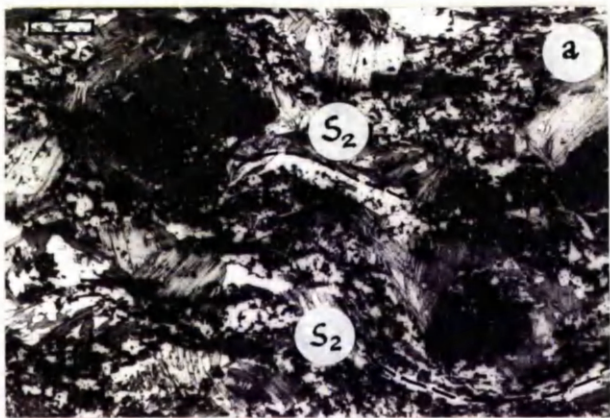
5. NATURE OF YOUNGER, DOMINANTLY STEEP AND ZONAL PROTEROZOIC STRUCTURES

5.1 GROUP 4 STRUCTURES

5.1.i Distribution and structural domains

These structures differ from earlier ones in style, attitude and intensity and no distinct domains are recognized in the study area. Rather, F_4 minor folds and S_4 show similar vergence and geometry wherever observed: these are asymmetrical dextral minor folds with associated axial planar foliation, frequently a crenulation cleavage (Figures 11b and 17a). No large scale structures have been identified within the study area, unless a lithological boundary between Niittylahti (Map 2 reference: Niittylahti 93.0 35.3 and Mulo (Niittylahti 90.1 36.0) is an S_4 fault zone. Since this is subparallel to S_3 trends, so that F_4 and F_3 dextral minor folds may be mistaken for each other, it must be admitted that other similar structures might have been overlooked, especially in poorly exposed areas.

A conjugate shear zone geometry has been described for correlates of Group 4 structures by numerous authors (Gaál and Rauhamäki 1971, Gaál 1972, Parkkinen 1975, Koistinen 1981, Halden 1982). Within the study area a few minor folds appear to correspond to the anticipated sinistral, NE-trending antithetic conjugate set (Figure 17b). This trend is however similar to that of the equally rare S_6 - F_6 so that unambiguous recognition requires consistently different vergence or presence of intervening Group 5 structures.

Despite the uniformity of Group 4 structural style, strain gradients are heterogeneous and irregular in that to the SW, the Suvasvesi and Haukivesi wrench fault - shear zones apparently record displacements of the order of tens of kilometers (Gaál

1972, Halden 1982). There they are associated with diverse and abundant syntectonic intrusions (Map 1). Though no such lithologies are recorded from within the study area, analogous structures occur in the form of quartz veins developed in minor dextral shear zones (Figures 17c and 20a).

No attempt to integrate Group 4 strain across the study area has been attempted as frequency of minor structures and mean displacement across them is not known.

5.1.ii   Group 4 fabric elements, microstructures and metamorphism

Synkinematic mineral recrystallization is associated with S_4 and manifest as compositionally modified "aureoles" to quartz veins in shear zones. This modification has tended to obliterate earlier fabrics where they have not merely been reoriented (Figures 17c and 20a). Syn-S_4 grain growth is not pronounced in axial planes to F_4 crenulations at outcrop scale (Figure 17a) or microscopically (Figure 18a,b).

F_4 microfolds in Figure 17a clearly crenulate biotite porphyroblasts grown within $S_2 - S_3$, but the F_4 crenulation cleavage is itself overprinted by biotite and staurolite porphyroblasts (Figure 18a) and tremolite porphyroblasts (Figure 18b). Thus there is some evidence, from widely distributed exposures, of idioblastic crystal growth, post-kinematic with respect to S_4. This may correlate with the biotite, staurolite and andalusite growth in Höytiäinen province metapelites, already constrained as post-dating S_3 (see section 4.3.vii above), but because S_4 is generally not a penetrative, ubiquitous fabric, it has not been possible to establish relationships to blastesis in every exposure. This mineral growth may also have been coeval with the post-S_3 garnet breakdown reaction in the south-west part of the study

area, and hence probably records the regional thermal metamorphic climax. Such timing is consistent with Koistinen's (1981) conclusions concerning the metamorphic peak at Outokumpu, and also those of Campbell and others (1979) from the eastern part of the Höytiäinen province, at Heinävaara. Based on the stoichiometry of coexisting mineral phases, and also solving simultaneous equations using thermochemical data for inferred reactions, the latter authors computed maximum metamorphic conditions, recorded by staurolite-muscovite breakdown, as T = 675 $\pm$ 25°C and P = 500 $\pm$ 25 MPa.

5.2   GROUP 5 STRUCTURES

As with Group 4 structures these display a characteristic
and homogeneous style across the study area while showing higher
displacement gradients in certain zones.   In general, S_5 is more
perceptible in metapelites of the Höytiäinen province than in
the Savo province, although a major Group 5 structure clearly
passes through Oravisalo and Varpasalo causing a regional deflexion
of earlier structures (Map 2 and Figure 5).   This is termed the
Orivesi shear zone and may be contiguous with a similar feature
found to the north between Outokumpu and the Sotkuma basement inlier
(Gaál and others, 1975).

5.2.i   Distribution and structural domains in the
Savo province

The Orivesi shear zone separates the province into two
Group 5 domains with vertical movement along the shear zone having
raised the eastern one with repect to that in the west; the shear
zone is thus of normal, rather than reverse sense, in contrast
to the earlier Group 3 overfolds and thrusts.   In the western
domain the composite S_0-S_3 foliation and associated early folds are
reoriented about large scale, relatively open, SSW- to SW-plunging
folds and it is these structures that give rise to the gross
attitude of lithological layering (Map 4).   At outcrop scale,
parasitic minor folds are frequently observed as coaxial with the
L_3 mineral elongation lineation but detailed examination shows
them in may cases to be slightly divergent.   This in itself does
not unequivocally indicate two separate phases of deformation
since such relationships could arise in a transected fold with
L_3 lying within the axial plane, oblique to the hinge-parallel
cleavage intersection lineation.   However, microfolds have been
seen with crenulation cleavage deforming L_3 biotite aggregates,
affording better evidence of superposed deformation.

In contrast to this domain, Group 5 structures are virtually absent from a large area to the east of the Orivesi shear zone, as might be inferred from the low-angle attitude of early foliations and gentle F_3-L_3 plunges in the Oravisalo-Rääkkylä-Nieminen area (Map 2). This change is not attributed to rheological contrasts between lithologies as metapsammites are effectively identical to either side of the shear zone.

However, immediately east of the zone at Oravisalo, and also to the north of the study area at Sotkuma, Archaean basement is exposed (Map 1). These may be the exhumed portions of a large contiguous unit of basement buried at relatively shallow depth; it is envisaged that such a competent block could to some extent inhibit the formation of large amplitude Group 5 structures in overlying metasediments.

Approaching the eastern margin of the Savo province, at Sammallahti and Niva, reorientation by Group 5 structures again becomes more noticeable, with SW-plunging F_5 folds nearly coaxial with L_3 but related to a steeply dipping axial planar crenulation cleavage, S_5. (Maps 2 and 4).

5.2.ii Distribution and structural domains in the Höytiäinen province

To the north-east of the Kettämö thrust zone, reorientation of early structures is not substantial at a regional scale but most exposures record some kind of Group 5 structure in the form of crenulations, asymmetrical to disharmonic minor folds (Figure 17d,e,f) or ductile shear zones (Figure 17g). In general, overprinting relationships are distinct with respect to S_3 and earlier structures and in some cases F_5 folds are clearly superimposed upon Group 4 structures (Figure 4e).

With few exceptions, F_5 folds in this region show a
consistent sense of asymmetry. When viewed down plunge they
typically show an open S-shaped profile; close inspection reveals
that the steeper, longer limbs are the higher strain portions.
Sense of asymmetry is the same as for the larger, but analogous
Orivesi shear zone: 'normal' displacement, with relative uplift
to the east. These relations are present at microstructural
scales as well but in this case alternative processes such as
buckling and pressure-solution truncation have also contributed
to final fabric morphology (Figure 19).

A notable feature of S_5 in many outcrops is its variation
in strike from about 175° to 220°. This is manifestly primary
and not a result of later superimposed deformation. In general,
Group 5 shears or faults are almost N-S as is for example, the
Orivesi shear zone, but in areas of lower strain, S_5 crenulations
swing into a more southwesterly trend, seeming to define lenticular
and anastomosing domains in which Group 5 structures tend to be·
more intensely expressed, as depicted in Map 3.

A line passing northwards along the eastern shore of the
lake Haapajärvi (Map 2 reference: Niittylahti 95.30) marks a major
change in lithology and also intensity of Group 5 structures. The
coarse clastic lithofacies and Group 2 - Group 3 structures of
Kettämönniemi are abruptly truncated by this fault zone, immediate-
ly east of which, depositional layering (where evident) and S_2 - S_3
foliations have been reoriented into an attitude almost invariably
steeper than 50 (Maps 2 and 3). In such zones, and particularly be-
tween Kumpu, Hammaslahti and Tikkala, a persistent difficulty.has
arisen in attempting to distinguish real Group 5 structures from
Group 3 or earlier structures merely passively reoriented into
steeper attitudes. Discrimination is only confidently made where
contrasts in cleavage style occur, such as the typical S_2 - S_3
differentiation in metapsammites overprinted by an S_5 crenulation
cleavage (Figure 17h). In the absence of these distinctive

criteria, horizontal exposures like that in Figure 10a and
Figure 12d display fold morphology appropriate for both F_3
upper-limbs and F_5 folds. Moreover, the frequent coaxial nature
of L_3 and the L_5 intersection lineations precludes ready discrim-
ination even in 3-dimensional exposures (Figure 37c). Examples
have nevertheless been found where presumable syn-S_3 boudins
are seen to pitch more steeply on S_3 planes than the intersection
lineation formed with S_5.

S_5 generally dips steeply westwards, except in a zone S of
Hammaslahti mine where dips are from 60° - 90° eastwards. This
does not seem to represent axial plane fanning about a major F_3
fold as no shallow hinge zone dips were recorded and minor folds
plunge northwards rather instead of southwards, preserving the same
sense of vergence throughout.

5.2.iii Group 5 structures in relation to quartz veins

Group 5 structures are expected to include quartz veins,
given the evidence presented below (section 5.2.iv) for volume loss
by solution transfer of quartz from S_5 crenulation cleavage domains.
Some veins are located with S_5 axial planes but a more typical
attitude appears to be perpendicular to this and the L_5 inter-
section lineation. This relationship, of E-W trending veins
and S-plunging L_5 are well expressed throughout Hammaslahti mine
(T. Karppanen spoken comm.) and is illustrated in Figure 10a,c.
Figure 12d clearly indicates an en echelon vein set perpendicular
to F_3 fold axes and boudinaged syn-F_3 or earlier veins. Much
smaller veins of similar attitude are shown in Figure 20b, record-
ing evidence of solution processes in a deformed gneissose clast.

Yet another vein configuration is represented by Figure
20c,d. This array is deduced as a Group 5 structure in that it
transects bedding-parallel S_3 and is of appropriate attitude in

having its acute bisectrix perpendicular to S_5. The orthogonal
vein sets within the metapsammite boudins shown in Figure 12 g,h
possess a similar geometry but do not unequivocally cut S_3, and
hence cannot be proven as Group 5 structures.

5.2.iv Fabric elements, microstructures and metamorphism

Expression of S_5 is not only dependent upon zonal distribution
of major Group 5 structures but also to a large extent upon
lithology. Thus S_5 is only rarely evident as a crenulation cleavage
in metapsammites of the Savo province and even in the Höytiäinen
province, metapsammites tend to lack syn-S_5 mineral growth, apart
from recrystallization within the more highly strained limbs of
asymmetric microfolds.

In metapelites of the Höytiäinen province, S_5 is frequently
a spectacular crenulation cleavage that records much evidence
concerning mechanisms and processes of origin. Figure 19a-h
illustrates the variation in morphology of these fabrics. Using
terminology proposed by Gray (1977, 1979) and Borradaile and
others (1982) they may usually be described as discrete zonal
crenulation cleavages as opposed to having gradational boundaries
between the relict microlithons and cleavage domains. Both kinds
may however be juxtaposed, as in Figure 19 e.

Individual cleavage domains may vary in length from several
mm to over 1 dm. In thin section, discontinuous cleavage domains
pass gradually lengthways into microfolds that have little effect
on pre-existing fabric, other than reorienting it, as in the right
hand part of Figure 19 d.

Diffuse cleavage domains become more distinct as the angle
between the reoriented foliation and S_5 axial surface decreases,
as in Figure 19 e. The left hand part of figure 19 d shows another
good example, with a strong dimensional alignment of quartz grains

further enhancing the reoriented S_2-S_3 fabric. This probably developed in a complex manner, with a combination of body rotation, incremental grain deformation and recovery in addition to solution transfer processes, as envisaged by Gray (1979) and White and Johnston (1981).

The predominating, discrete cleavage domains tend to be thinner than diffuse ones (about 1 mm compared to up to 7 mm) and also relatively depleted in quartz. They may consist of multiple anastomosing foliae defining a distinct zone, as in Figure 19 f, through which relict earlier foliation may still be traced, or comprise single discordant seams truncating early fabrics and showing a strong preferred orientation of mica, as in Figure 19 d, b and c. Figure 19 g,h shows mica aligned within axial planes of F_5 microfolds but such distinct mineral growth does not normally accompany S_5.

The various kinds of crenulation cleavage observed may each be members of a sequence that progressed through stages as follows.

1) Probable low-attitude relict depositional and S_2-S_3 differentiation layering deformed at first by bulk flattening, forming small, discontinuous S_5 cleavage domains with localized pressure solution, effectively analogous to stylolites (Figure 19 b,d).

2) The role of pressure solution may have temporarily diminished as the rock responded by buckling into microfolds, typically of wavelength 2-7 mm (Figure 8 e). Earlier fabrics could have been enhanced by body rotation of individual grains, intracrystalline deformation and recovery and crystallization of new grains as a result of progressive reorientation with respect to imposed forces (Figure 19 d,e,h).

3) Asymmetrical microfolds produced zonal crenulation cleavage

with gradational boundaries between relict microlithons and developing cleavage domains (Figures 8 f and 19 e).

4) Renewal of solution transfer in tighter microfolds results in zones of anastomosing foliae depleted in quartz, enclosing and separating microlithons which can nevertheless be traced coherently through the cleavage foliae (Figure 19 c,d,f). From such criteria, Gray (1979) argued against the occurrence of shear dislocations along the cleavage domains, at least as a primary cause of apparent offsets across them.

5) Final stage is amalgamation of foliae, or possible lateral growth due to continued dissolution of more mobile components, obscuring relict fabrics and abutting microlithons sharply, as though truncating them (Figure 19 a,d,g). A new anisotropy has thus been imparted to the rock by a combination of microfolding and grain-scale adjustments, including volume loss, rather than by faulting or shear transposition (Figure 19 c).

Metamorphic conditions during this stage of the deformational history are not well constrained. S_5 clearly kinks or truncates post-S_3 and post-S_4 biotite porphyroblasts, suggesting that the regional thermal metamorphic peak occurred after S_4 formation (Figure 18 a) but prior to the S_5 crenulation cleavage (Figure 18 h). At Heinävaara (Map 2 reference: Heinävaara 08.39), Campbell and others (1979) present evidence for limited garnet and staurolite growth being pre- or syn-kinematic with respect to S_5 and relate this to staurolite instability. Pseudomorphs, presumably after syn- or post-S_3 staurolite are also present at Hammaslahti and Tikkala (Figures 16 and 18e) though mica, quartz and tremolite are noted instead of garnet, staurolite and biotite. If indeed the results from Heinävaara indicate a somewhat anomalous, younger thermal maximum than elsewhere, this is explicable according to the geometry of Group 5 structures established for the study area - W-dipping shear zones of normal sense would be expected to progressively expose deeper crustal sections by relative uplift to the east.

5.3   GROUP 6 STRUCTURES

5.3.i   Distribution and structural domains

Group 6 structures trend NE or ENE but are only recognized
in several exposures, where they truncate or deform Group 5 and
older structures. In the northeastern part of the study area,
they are present as an S_6 crenulation cleavage axial planar to
kink bands (Figure 20e). Elsewhere, at Hammaslahti mine and the
Oravisalo quarry Group 6 structures are expressed as brittle faults,
sometimes quartz-filled,with at least some dextral horizontal
component of displacement. If this is consistent for S_6 structures,
it could diagnostically discriminate the antithetic Group 4
conjugate which also trends NE but has sinistral chirality.

5.4   GROUP 7 STRUCTURES

5.4.i   Distribution and structural domains

In the strictest sense, the subscript notation should not
be used here with a sequential connotation - although they
definitely deform Group 5 structures in Figure 4f, the relationship
between Group 6 and Group 7 structures is not yet established.
The gentleness of folds and lack of asymmetry precludes definition
of vergence domains and their general weakness and uniformity
prevents recognition of any zonal variations in expression. S_7
is usually discontinuous and disjunctive, most evident on
weathered surfaces or where it traverses feldspar grains. In
metapelites it can occur as a distinct crenulation cleavage, with
only incipient development of cleavage domains due to relatively
low amplitude of microfolds (Figure 20f). Sometimes two distinct

trends, to 080° and 095-110° are discernible but it is uncertain whether they could be conjugate or of wholly different origin. F_7 folds are usually open or irregular flexures whose plunge is dependent upon disposition of pre-existing foliations.

6.5   LATE STRUCTURES DIFFICULT TO CLASSIFY

Spaced, N-trending zones of alteration, in places several cm wide are sporadically seen, commonly containing epidote and chlorite or narrow seams of pseudotachylite (Figure 20 g). They are likely to correspond to the S_6 cleavage of Koistinen's (1981) classification and similar structures have been reported by Bowes (1976) and Park (1983).

Figure 20 h shows various late sets of veinlets, fractures and metasomatic or retrogressive zones that are commonly seen on weathered metapsammite surfaces. They appear to record a weak E-trending fabric (S_7?) cut by N-trending veinlets and still younger ESE-trending en echelon fractures. Such a conclusion should be substantiated from other exposures but is compatible with the sequence and orientation of the latest structures described by Bowes (1976) and Koistinen (1981).

FIGURE 16 : METAMORPHIC ASSEMBLAGES AND LATE STRUCTURAL DOMAINS

Most of the distinctions on the map are ascribed to compo-
sitional differences in protoliths, except for the presence
of fibrolite in the south-west of the area. Where established,
the climactic thermal metamorphism occurred between the formation
of Group 4 and Group 5 structures. Major Group 5 structures
shown on the diagram are the Haapajärvi fault and the Orivesi
shear zone, both of normal sense with uplift to the east. Grid
coincides with boundaries of 1:20 000 map sheets.

SAVO PROVINCE

* post-S3 chloritoid and actinolite in mafic metapelite
* S1-S3 garnet and L5 biotite aggregates in quartz-plagioclase-mica metapsammitic schists
* post-S5 garnet breakdown to fibrolite and prismatic sillimanite + biotite + plagioclase + quartz
* post-S5 garnet breakdown to fibrolite and prismatic sillimanite + biotite + plagioclase + quartz
* syn-S5 biotite and hornblend in Archean gneiss
* S5 Orivesi shear zone

HÖYTIÄINEN PROVINCE

* chlorite porphyroblasts post-kinematic with respect to S5
* post-S5, pre-S6 garnet, staurolite, andalusite in metapelites
* post-S5 staurolite pseudomorphed by phyllosilicates
* pre-S5 tremolite overprinting S4-S5 in calcareous metapelite
* syn-or-post-S5 staurolite and cordierite
* hornblende + chlorite in mafic volcanogenic rocks
* post-S5, pre-S6 biotite porphyroblasts in metaturbidites

Orivesi

Oravisalo

Rääkkylä

Pyhäselkä

Niva

Mulo

Rekola

Hammaslahti township

HAAPAJÄRVI FAULT ZONE

Hammaslahti mine

Heinävaara

Kiihtelysvaara

Tohmajärvi

HÖYTIÄINEN PROVINCE

SAVO PROVINCE
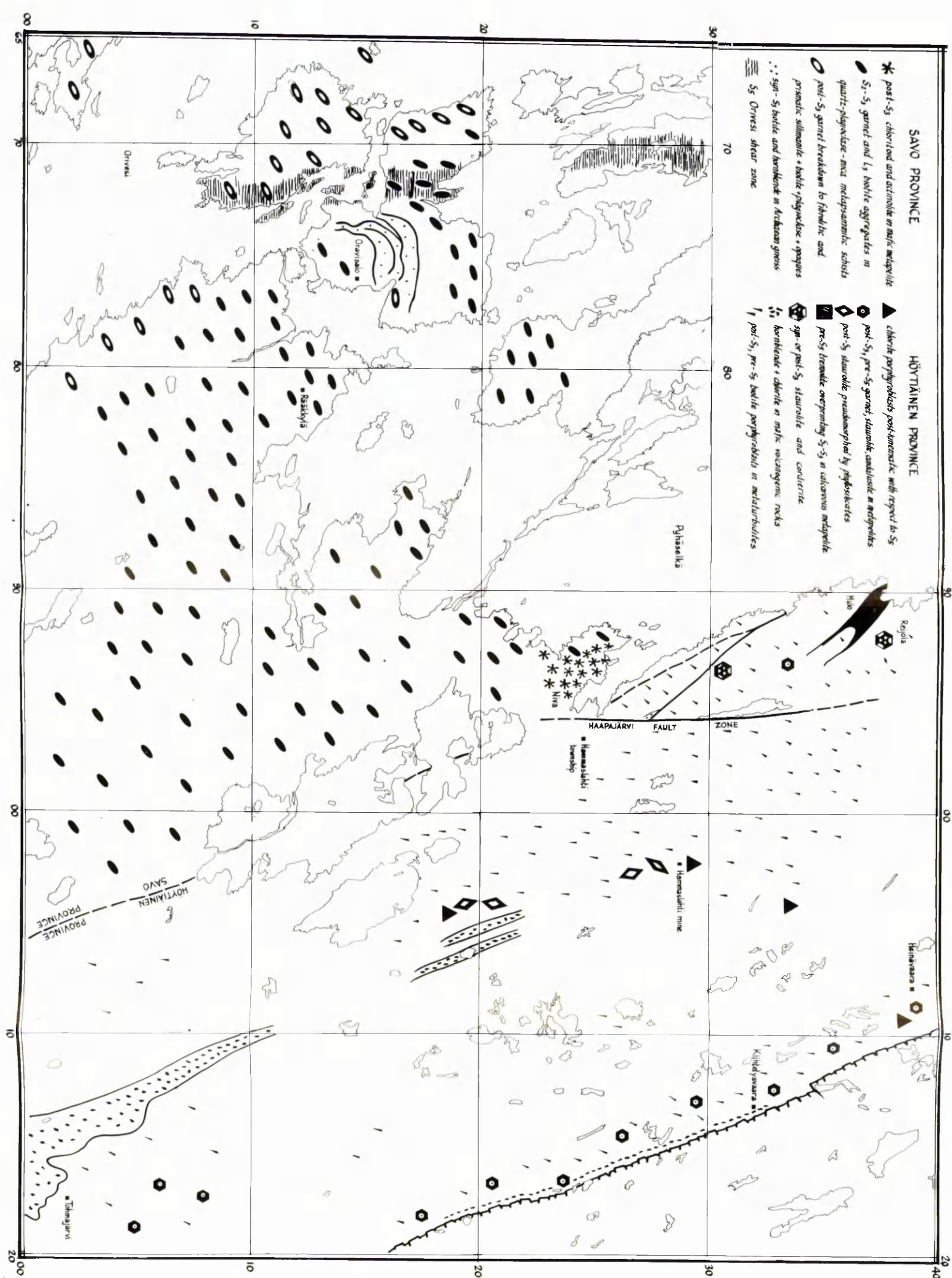
FIGURE 17:   GROUP 4 AND GROUP 5 STRUCTURES

a F_4 crenulations deforming earlier composite S_0-S_3 fabric. Unless
     they underwent subsequent mimetic growth, it seems clear that
     biotite porphyroblasts were syn-tectonic with respect to S_3
     and definitely pre-date S_4 (Niittylahti 91.7 35.6).

b   Isoclinal F_3 folds are overprinted by NE-trending crenulations.
     These could be assigned to S_6-F_6 but sinistral chirality in
     adjacent exposures suggests that they are antithetic conjugates to
     the usual NW-trending S_4.   (The few examples found where S_6 is
     established as cutting S_5 have indicated the opposite chirality)
     (Niittylahti 92.4 35.2).

c   Minor S_4 shear zone containing a quartz vein and causing both
     distortion and compositional modification of composite S_2- S_3
     fabric (Oravisalo 70.5 19.6).

d S_5 shear zone, probably with some vertical component of displacement.
     Deformed fabrics at right are difficult to assign, but are either
     S_3 or S_4 axial planar to F_3 or F_4 dextral folds, deforming
     lithological layering and parallel S_2 quartz veins (Pyhäselkä
     94.0 29.3).

e   Asymmetrical S_5 crenulations developed in laminated metapelite but
     absent from thicker massive metapsammite (Pyhäselkä 93.9 29.3).

f   Composite S_2-S_3 differentiation fabric with abundant quartz veins,
     folded by F_5, with no evidence of accompanying S_5 (Pyhäselkä
     93.5 29.2).

g   Discrete, zonal S_5 crenulation cleavage across well-preserved
     depositional laminae and subparallel but weak S_3 (Rauansalo 00.7 23.9).

h   Axial planar S_5 crenulation cleavage deforming S_2 differentiation
     layering in metapsammites. Vein quartz is also present in axial
     planes (Heinävaara 03.7 32.5).
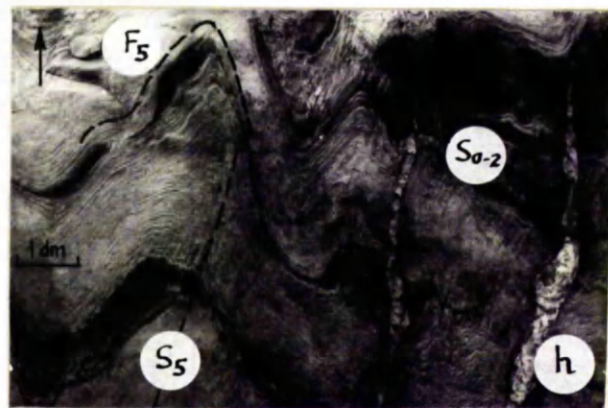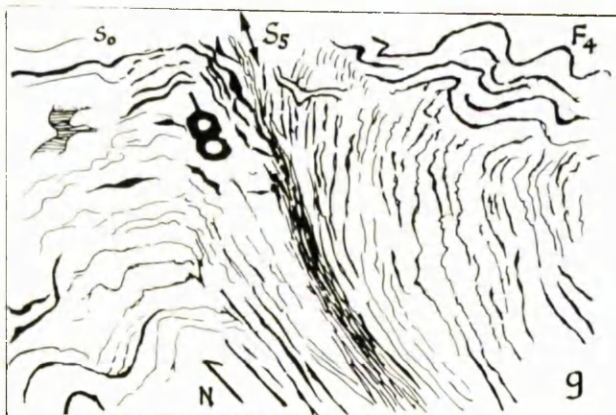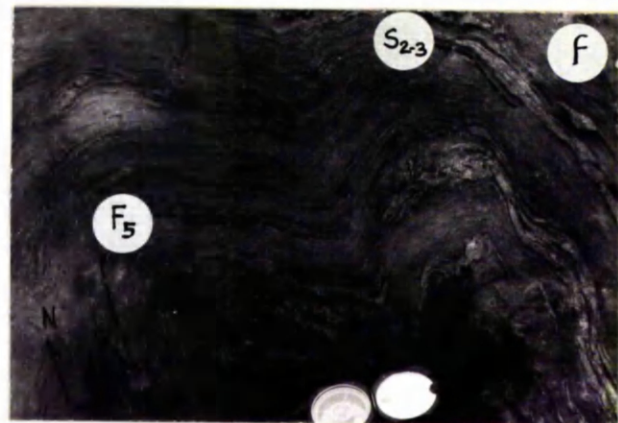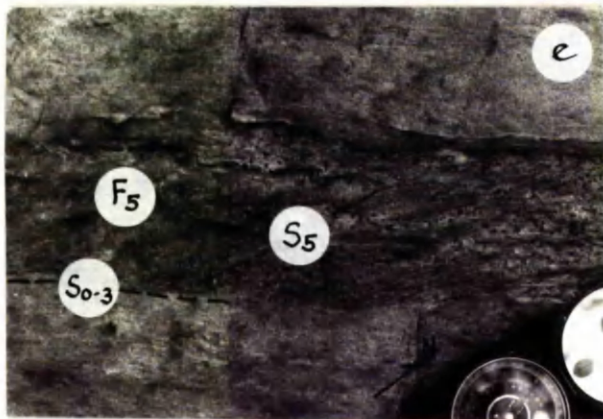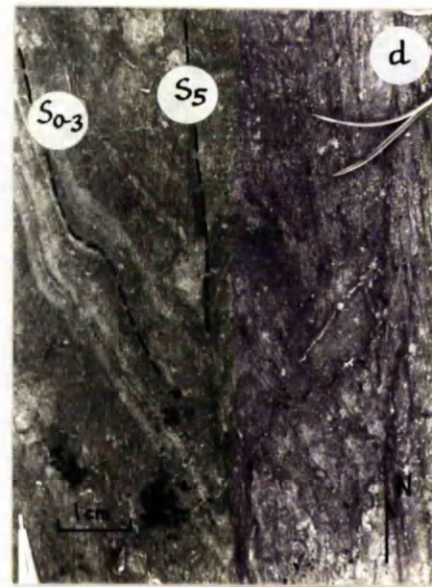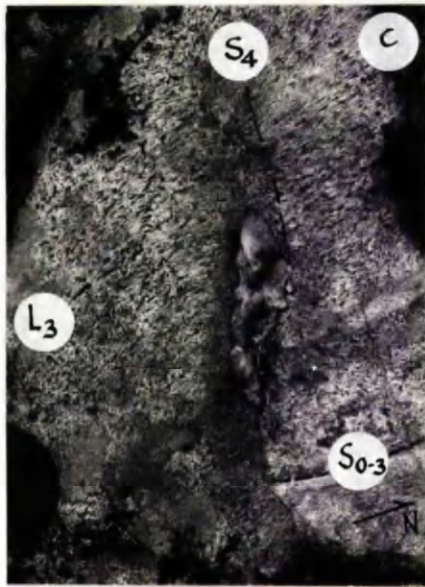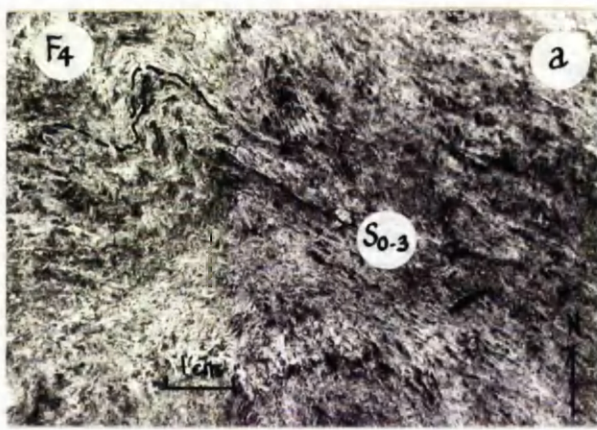
FIGURE 18:   PORPHYROBLAST RELATIONSHIPS TO GROUPS 4 AND 5
MICROSTRUCTURES
Scale bar represents 1 mm

a   Twinned biotite porphyroblasts overprinting F_4 microfolds that
crenulate composite S_2-S_3 in graphite-bearing metapelite. Hexagonal
grain of high relief may be staurolite, though is not pleochroic
and has sector twinning like that of cordierite (Pyhäselkä 94.7
28.6). PPL.

b   Pre- or syn-S_5 tremolite porphyroblasts superimposed upon S_2-S_3
differentiation layering in calcareous metapelite (dark bands are
graphite-rich). Open crenulations are late F_3 or F_4 microfolds
(Niittylahti 91.3 35.6). PPL.

c F_3 or F_4 microfolds deforming S_2-S_3 differentiation layering, with
darker bands being graphite-rich. Disrupted pre- or syn-S_3 veins
or sulphide porphyroblasts are intergrown with or replaced by
K-feldspar at margins (Niittylahti 91.3 35.6). PPL.

d   Apparent multiple growth of feldspar and sulphide resembling vugh
growths, except that here, the 'vugh' is filled with quartz, possibly
part of a deformed syn-S_2 veinlet (Niittylahti 91.3 35.6). NX.

e   Chloritic and micaceous pseudomorphs after staurolite porphyro-
blasts, syn- or post-kinematic with respect to S_3. Accompanying
this replacement, or younger still are acicular tremolite crystals
intergrown with quartz (Hammaslahti mine). PPL.

f   Porphyroblast, pre- or syn-kinematic with respect to S_5, encloses
a composite S_2-S_3 fabric defined by dimensionally aligned quartz
and mica, more distinct within the blast than outside it. Nature
of blast is unclear - pseudomorphing of former staurolite by
sulphide is a possibility (Onkamo 04.6 17.5). PPL.

g   Post-S_3 dendritic actinolite in quartz-mica-chloritoid metapelite
(chloritoid not evident in this photograph) (Pyhäselkä 93.9 24.6).

h   Post-S_3 biotite porphyroblasts deformed and truncated by discrete,
anastomosing S_5 crenulation cleavage (Niittylahti 97.7 30.5). PPL.
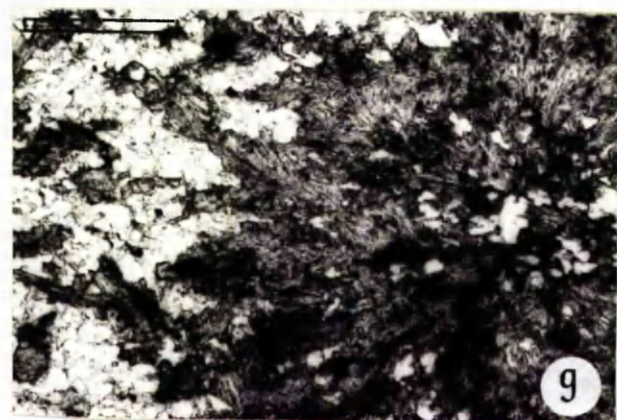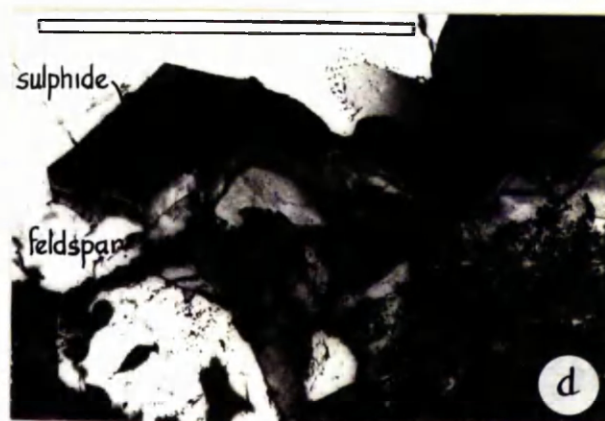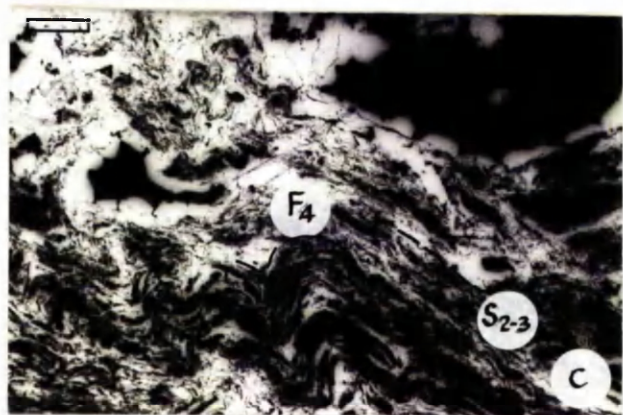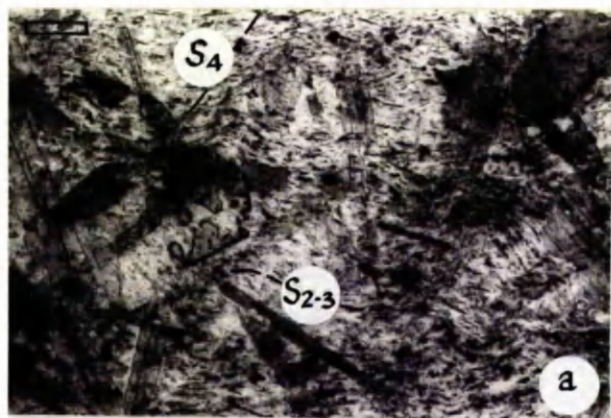
FIGURE 19: S_5 CRENULATION CLEAVAGE IN HÖYTIÄINEN PROVINCE METAPELITES
Scale bar represents 1 mm

a   Apparent offsets of lithology-parallel quartz veinlet during formation of S_5 crenulation cleavage domains. Spacing between domains is least where S_0 is reoriented subparallel to S_5, at bottom left (Niittylahti 94.6 38.4). PPL.

b   Detail of S_3 veinlet showing deflexion of S_5 crenulation cleavage around microfolds and also distinct thinning of veinlet where it crosses S_5 cleavage domains (Niittylahti 94.6 38.4). PPL.

c   Finely spaced S_5 cleavage domains transecting tectonically modified silty laminae (Rauansalo 01.8 28.3). PPL.

d   Asymmetric F_5 microfolds deforming composite S_2-S_3 foliation. S_5 cleavage domains may be described as incipient and gradually bounded (far right) or as discrete and truncating (Niittylahti 94.6 38.4). PPL.

e   Composite S_2-S_3 foliation continues across several thin, wispy and discontinuous S_5 cleavage domains at right but is completely truncated and reoriented or recrystallized within the thicker, dark seam-like domain. To the left of this, S_2-S_3 has an enhanced dimensional alignment of quartz and mica through a gradationally bounded rather than discrete S_5 cleavage domain (Niittylahti 93.3 37.4). PPL.

f   Composite S_2-S_3 mica-rich cleavage domains and microlithons are seen to pass continuously through discrete S_5 cleavage domain composed of anastomosing foliae, themselves crenulated by F_7 microfolds (Niittylahti 94.6 37.9). PPL.

g S_3 differentiation layering left of center, deformed into angular F_5 microfolds, with S_5 fortuitously parallel to lithological layering. In contrast, silty layer at right does not show any distinct S_3 fabric (Rauansalo 01.3 21.4). PPL.

h   Detail of S_3 differentiation layering in 'g' to show reorientation or recrystallization of phyllosilicates in mica-rich domains, so that they lie in an S_5 axial planar attitude (Rauansalo 01.3 21.4). PPL.
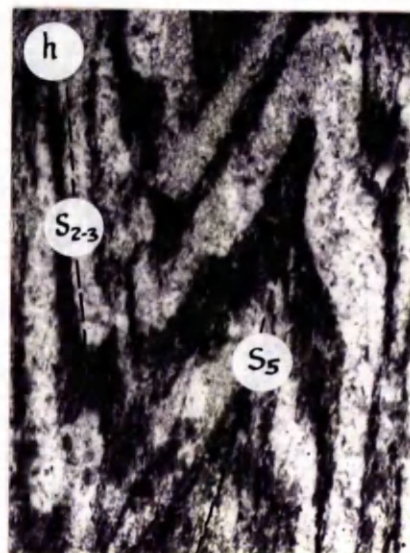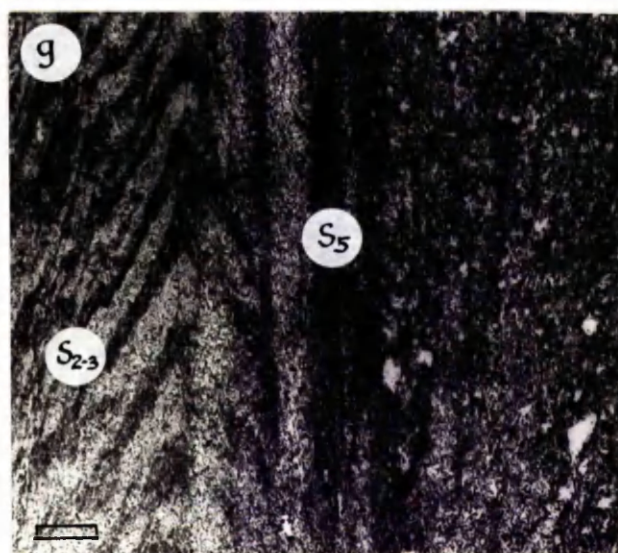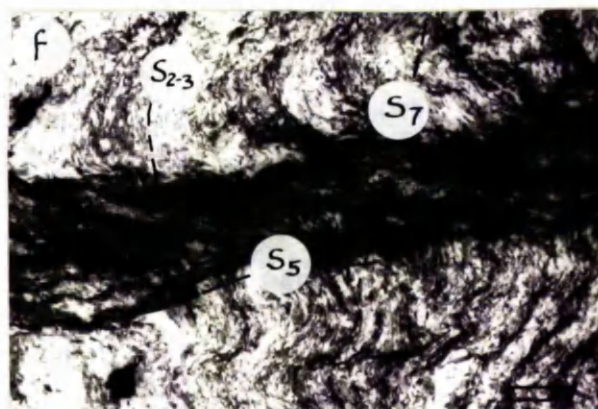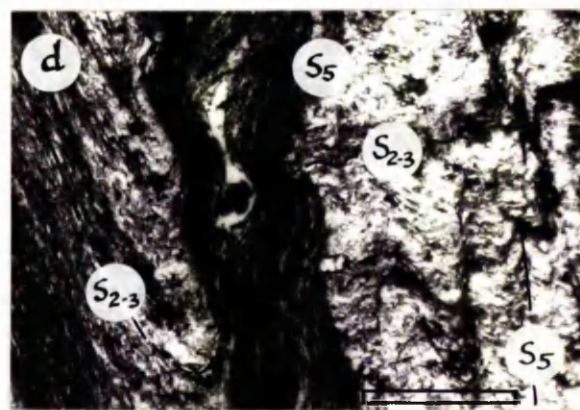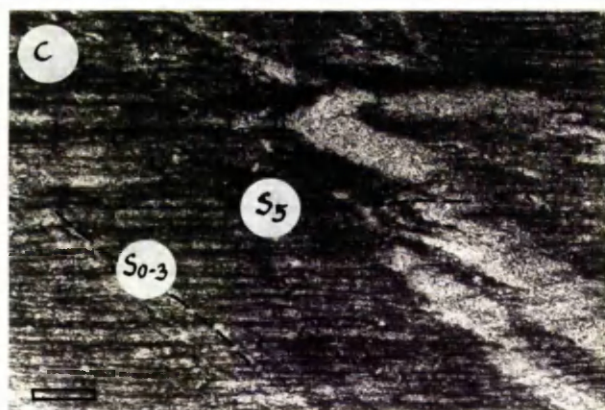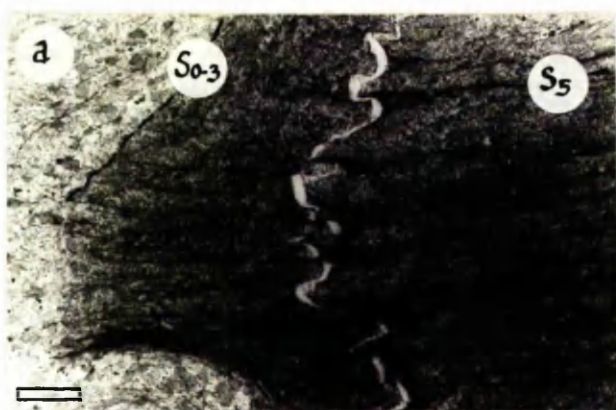
FIGURE 20:   QUARTZ VEINS AND LATE STRUCTURES

a   Massive quartz development associated with dextral NW-trending
        S_4 shear zone. Note deflexion and obliteration of S_2-S_3 composite
        fabric approaching the zone (Oravisalo 74.7 19.8).

b   Small veins perpendicular to L_5 and S_5, associated with deformation
        of gneissose clast. Outer margin of vein contains feldspar in
        addition to quartz. Approximately natural size (Rauansalo 02.4 28.2).

c,d   Because of the geometrical congruence of composite S_0-S_3 and S_5
        fabrics, it is difficult to establish affinities of the quartz
        veins. Their undeformed, cross-cutting nature does however
        suggest that they developed during or after S_5 formation
        (Onkamo 04.6 17.5).

e   Broad sinistral kink band and NE-trending S_6 crenulation cleavage
        distinctly cutting composite S_2-S_3 (Heinävaara 02.8 33.2).

f   Unusually distinct S_7 crenulation across earlier S_2-S_3
        differentiated fabric (Valkeasuo 17.4 18.9).

g   Epidotized fracture zones (S_8), containing sporadic seams
        of pseudotachylite (Rasivaara 86.5 07.4).

h   Unclassified sets of late fractures and veinlets, lacking in
        relationships to earlier structures. They may however be
        E-trending S_7 (parallel to compass) cut by N-trending S_8
        veinlets (parallel to knife) in turn cut by S_9 ribbed zones
        (Pyhäselkä 91.4 20.6).

a — S_4, S_5, S_{2-3}, N

b — S_5, S_{0-3}, N

c — graywacke, vein (post S_3), $S_0 \parallel S_3 \parallel S_5$, N

d — N

e — S_6, S_{2-3}, N

f — N, S_7, S_{2-3}, L_7

g — S_8, N

h — S_9, S_7, S_8, N
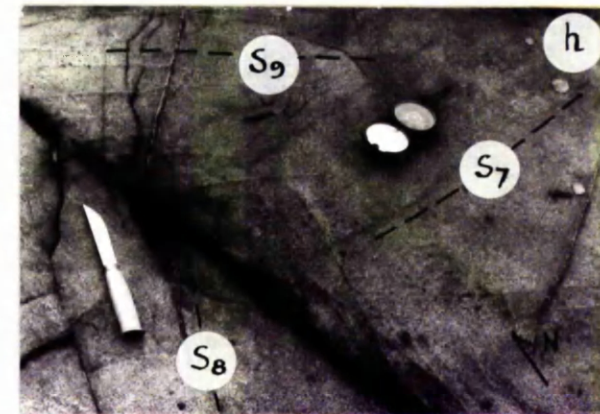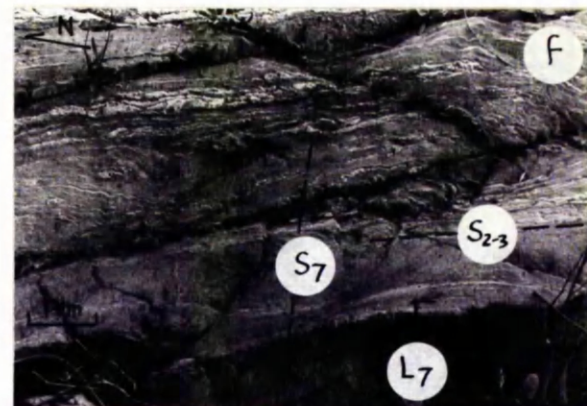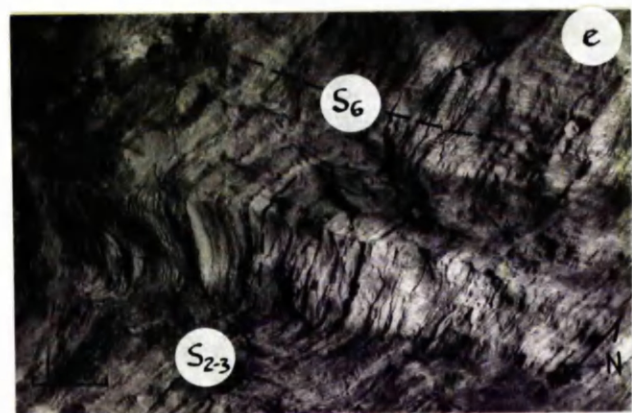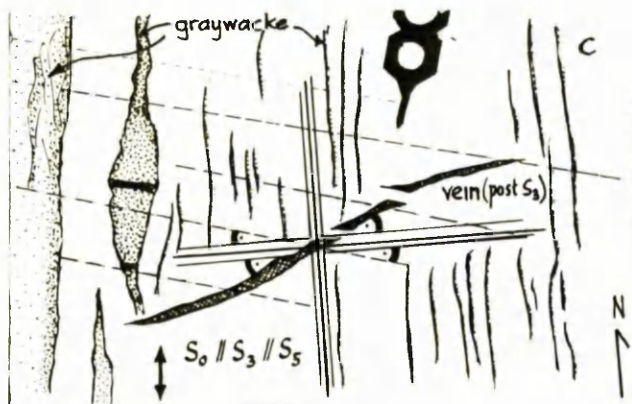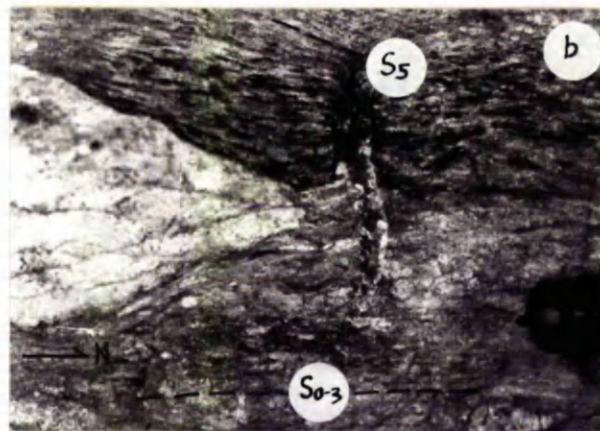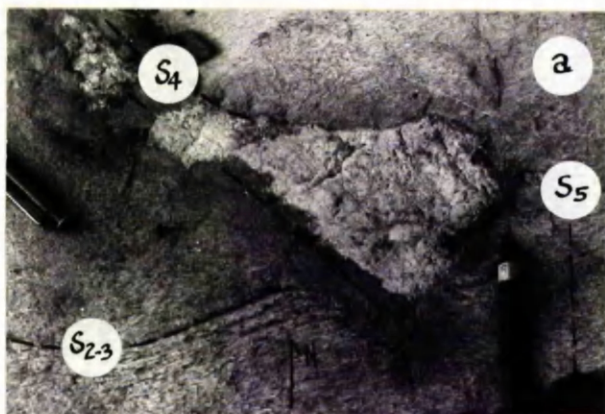
6.   KINEMATIC INTERPRETATION OF STRUCTURES

It is logical to regard the differences in style and
attitude between early and late groups of structures as reflecting
different mechanisms or stress systems during their formation.
On this basis they are therefore discussed separately with regard
to kinematic development.

6.1   NATURE OF EARLY SVECOKARELIAN DEFORMATION

At all scales, Group 1, Group 2 and Group 3 structures
resemble those typical in moderately metamorphosed, overthrust
sequences such as the Pennine Alps, or Otago schists of New
Zealand (Hobbs, Means and Williams 1976). Isoclinal, recumbent
F_2 folds (where not reoriented by younger structures) and overturned,
tight to isoclinal F_3 folds, locally associated with mylonite
zones are all features indicative of substantial horizontal
shear strains.

Subject to the interpretation of the L_3 lineation, a
tectonic transport direction to the NE or NW is deduced, based
on abundant Group 3 vergence criteria in association with younging
directions. Group 2 structures apparently share this geometry,
recording translation onto the Karelian craton or somewhat
obliquely along the margin. No translation vector can be
determined for Group 1 structures but if they correlate with the
major dislocation and emplacement of the Outokumpu nappe, as
advocated by Koistinen (1981), then a substantial northwards
component is again implied. The absence of Group 1 structures
from the entire Höytiäinen province is also consistent with
deformation having been initiated in the south and west and
advancing towards the craton.

Archaean basement was not significantly affected by this deformation but remained as a competent nucleus, as maintained by Wegmann (1928, 1929), Hausen (1930) and Väyrynen (1939). Even near the cover interface, within allochthonous thrust slices, basement shows little indication of substantial bulk shortening.

The first stage of Svecokarelian deformation may therefore be regarded as a loading of the craton margin beneath a series of fold and thrust nappes; exposure and stratigraphical constraints are too limited to permit recognition of any duplex morphology but the absence of Group 1 structures, and localized nature of Group 2 structures in the Höytiäinen province accords with a scenario involving progressive craton-ward propagation of splays from a basal thrust zone.

At outcrop scale, diagnostic evidence for non-coaxial deformation such as simple-shear strain, is not abundant. However, some Group 2 or early Group 3 quartz veins have morphologies consistent with intitial development as extensional arrays followed by deformation within a rotational stress field (Figures 12 a,b and 37 c). That is, sigmoidal vein shapes are not necessarily primary nor due to rotation of primary stress fields in shear zones (Beach 1975, Rickard and Rixon 1983) but can result from subsequent, superimposed shear parallel to one quartz vein direction. Microstructural expression of non-coaxial deformation, such as curved inclusion trails within or pressure shadows around porphyroblasts are, however not common (Figure 15).

6.2   SIGNIFICANCE AND INTERPRETATION OF GROUP 3 LINEATIONS

Many theoretical and field-based studies relate to the origin of shape fabric lineations (see review by Hobbs, Means and Williams 1976, and Sanderson 1973, Bell 1978). Frequently they

are regarded as indicating direction of maximum principal
elongation, parallel to the vector of thrusting. This inter-
pretation has been difficult to reconcile with theories and
experiments demanding that such lineations be perpendicular
to fold hinges since, in the field, these lineations are
commonly parallel. This is typically so in the study area as
elsewhere in Karelia (Koistinen 1981, Park and others 1984),
but at Oravislo, some folds indistinguishable from F_3 in other
respects (style, attitude and overprinting) are almost perp-
endicular to the SW-trending L_3 biotite lineation and may
be relevant to interpretation of the kinematic development
of Group 3 structures in general. Before this is discussed
as an apparent paradox, several other possibilities should
be examined.

6.2.i L_3 as an intersection lineation

L_3 may not be a shape-fabric lineation but instead
defined by the intersection of biotite porphyroblasts within
S_3 and the composite S_o - S_2 foliation, in which case L_3 should
be coaxial with F_3. Bell (1978) provides a succinct explanation
for this : at low strain states in a rock with a foliation
containing or defined by oblate grains, any intersection with
a superposed metamorphic fabric will produce an apparent mineral
elongation lineation, parallel to any associated fold axes. This
seems tenable for some Group 3 structures but less so where L_3
is defined by prolate, pre-tectonic elements such as conglomerate
pebbles or where F_3 and L_3 mineral lineations are divergent.

Apart from the examples mentioned above , in which this is
in fact the case, sufficient evidence exists to show that L_3 is
commonly defined by elongate minerals or mineral aggregates and
deformed metapsephite clasts, and hence cannot be an intersection
lineation in all cases.

6.2.ii L_3 and syn-kinematic mineral growth

It was proposed in Section 6.1 that Group 3 structures formed partly under simple shear, consistent with extension perpendicular to fold hinges. On the other hand, the elongate knots of biotite defining L_3 could be taken to indicate extension parallel to fold hinges, particularly if their present shape derives directly from both homogeneous and irrotational Group 3 deformation. This difficulty in interpretation might be obviated if the biotite knots had crystallized syn-kinematically, even during non-coaxial deformation. Under such conditions, growth of nucleating porphyroblasts might have been preferential parallel to F_3 hinges since shear strain would be relatively insignificant in this direction; perpendicular to this, accommodation of incremental shear strains could cause continued disruption to nucleating grains by slip along the [001] plane, inhibiting their growth. Similarly, in a petrofabric analysis of quartzite from Koli, north of the study area, Hietanen (1938) found acicular kyanite parallel to the major tectonic lineation (correlated here with L_3) and also to c-axes in quartz grown perpendicular to the walls of extensional veins. If these veins were coeval with Group 3 structures, then L_3 would seem to be the direction of maximum elongation but alternatively, they may be younger, related to congruent Group 5 structures, as is demonstrably the case at Hammaslahti mine (Figure 12 b).

Regardless of which interpretations and mechanisms are realistic and acceptable for synkinematic mineral aggregates, a satisfactory explanation is necessary to account for the parallel orientation of F_3 and deformed, elongate clasts; this matter is addressed in Section 6.2.iv.

6.2.iii L_3 and rotation of F_3 fold axes

Sanderson (1973) and Escher and Watterson (1974) have offered theoretical explanations for the reorientation of contemporary fold axes into the tectonic translation direction, as a result of high shear strain during thrusting. Williams (1978) has studied progressive reorientation of fold axes in proximity to major thrusts in two superposed Norwegian nappes, using garnet pressure shadows to define the X direction of the finite strain ellipsoid. Quinquis.and others (1978) likewise emphasize the role of intense simple shear in strongly deformed blueschists in Brittany and relate this to the progressive development of sheath folds. Although no structures of this nature were identified in the study area, this might reflect poor exposure as much as lower amounts of finite strain.

In his study of mylonites from the Woodroffe thrust of central Australia, Bell (1978) affirms the conclusions of the above authors but further demonstrates the inhomogeneity of strain and deduces the existence of a bulk flattening ('pure shear') component in addition to that due to simple shear. Moreover he shows that large elliptoidal blocks with relatively little internal strain can be passively rotated within anastomosing higher strain ductile mylonite zones. Such observations demonstrate the heterogeneous nature of this kind of deformation, a feature also evident from the variation in Group 3 fold morphology and fabric intensity in the study area (Section 4.3).

Other similarities with Group 3 structures are described by Bell, resembling the late F_3 folds at Oravisalo,   having fold axes at a high angle to the mineral elongation lineation and are interpreted as recording diminishing simple shear strain. However, most other examples of Group 3 structures, including mylonites examined at Oravisalo and Kettämö, lack convincing evidence for the preservation of F_3 fold axes in various degrees

of rotation towards the L_3 elongation direction. Since the mylonites in particular, are most prevalent in massive, though anisotropic Archaean migmatites this may merely indicate that their rheology was at no stage conducive to the formation of folds, either by bulk shortening or simple shear.

6.2.iv L_3 in relation to deformed metapsephite clasts

Unlike syn-kinematic minerals, sedimentary clasts share the complete deformational history of their enclosing lithologies but the extent to which they record the same finite strains will depend upon clast : matrix viscosity contrasts and initial clast shape. This situation was modelled by Gay (1968 a,b) who found that for rigid bodies undergoing homogeneous flattening in a viscous matrix, those of prolate shape will reorient with long axes parallel to the maximum elongation direction whereas oblate bodies will rotate such that their XY planes are perpendicular to the axis of short-ening . Metapsephites of the study area have oblate clasts lying within $S_2 - S_3$, consistent with this result while prolate clast long axes tend to parallel F_3 fold axes.

In contrast, Gaál and Rauhamäki (1971) recorded X : Y : Z ratios of nearly 100:1:1 for clasts in the Savonlinna conglomerate southwest of Haukivesi shear zone (see Map 1) and found the lineation thus defined to be commonly divergent to contemporary fold axes; the elongation lineation was consequently regarded as parallel to the tectonic translation direction. This interpretation is of particular relevance as the Savonlinna conglomerates are only 30 - 40 km south -west of the present study area and the elongation lineation trends virtually parallel to L_3.

Clasts in the study area are less prolate than at Savon-linna though this is difficult to establish quantitatively since clast : matrix viscosity contrasts and initial clast shapes are unknown. Furthermore, the lack of suitable exposures for measuring all three principal clast dimensions precludes plotting of data on conventional Flinn diagrams or performing strain analyses using Dunnet's (1969) method. Nevertheless, in Figure 21a , limited

results are presented from clasts measured parallel and perpen-
dicular to the composite S_2 - S_3 foliation and at a consistent
angle to L_3 , enabling a comparison of the relative dimensions
of different kinds of clasts. Intrabasinal clasts are the most
elongate, particularly with increasing size, in contrast to the
more equidimensional granitoid and vein quartz pebbles. This
probably reflects an originally more tabular clastic shape for
intrabasinal clasts although they may also have deformed more
readily than the more indurated and less anisotropic pebbles
of extrabasinal origin.

Thus, reliable strain estimates need to consider initial
shape variation, although relative differences in finite strain
from place to place may be compared with respect to a specific
clast type. For example, figure 21 c shows granitoid and quartz-
ose clast dimensions in a major F_3 fold hinge near Petäjikkö-
kallio to be more equidimensional (measured perpendicular to L_3)
than those adjacent to the Kettämö thrust zone, suggesting that
strain increases in proximity to the thrust.(compare Figures 31
and 14 e,f).

Hence, evidence exists for both a syn-sedimentary anisotropy
and a superimposed tectonic strain causing the short axes of clasts
to be perpendicular to the S_2 - S_3 foliation, which is the orient-
ation usually associated with the XY plane of the finite strain
ellipsoid (Hobbs and others 1976). Figure 21 b indicates that
flattening perpendicular to this plane (as recorded by the ratio

Z / Y) has been more substantial than elongation within it(as meas-
ured by the ratio Y / X) . That is , at this locality , on a normal
F_3 limb at Pitkänurmi (Map 3) , clasts are more oblate than prolate,
though most granitoid and all graywacke clasts are to some extent
elongate parallel to L_3 as defined by syn-kinematic mineral growth
and F_3 fold hinges .

6.2.v Concluding remarks concerning origin of L_3

In studies advocating progressive reorientation of
fold axes into the maximum elongation direction (Sanderson
1973, Escher and Watterson 1974, Bell 1978, Quinquis and
others 1978, Williams 1978), zones of very high strain, and
typically mylonitic are invariably described. Although such
structures are present within the study area, as are examples
in which contemporary fold axes are perpendicular to L_3 min-
eral lineations, abundant, well-preserved sedimentary
structures indicate that such intense strains are not
ubiquitous and that in some cases, L_3 might not correspond
to maximum elongation direction and tectonic translation
direction. Similarly , Ehlers (1976) has found deformed
markers to be very prolate in the hinge of the Kumlinge
synform (south- west Finland) but oblate elsewhere, indicating
homogeneous flattening on fold limbs. An explanation for such
observations was provided by Beutner (1978) who partitioned
finite strains into incremental components, of which the earlier
ones seem applicable to Group 3 (and also Groups 1 and 2)
deformation.

1) Synsedimentary and diagenetic compaction accompanies
volume loss, producing oblate ellipsoids whose major and inter-
mediate axes are contained within lithological layering.

2) Layer-parallel shortening produces prolate ellipsoids
due to extension perpendicular to compression direction.

3) Buckle folds develop with fold hinges parallel to
the existing lineation defined by prolate ellipsoids (cf. Bell 1978).

4) Under shear strain, earlier structures may be transposed
and reoriented, particularly in proximity to mylonite zones.

This sequence seems generally analogous to proposed Group
3 (and earlier) structural development but before definitive
conclusions can be drawn, more detailed studies are necessary,
particularly of mylonite fabrics and systematic strain analysis
of both deformed clasts and syn-kinematic porphyroblasts.

6.3 ORIGIN OF GROUP 4 STRUCTURES

Although they are of much smaller scale, F_4 folds in the study area have a geometry analogous to the larger structures described from adjacent regions by Gaál and Rauhamäki (1971), Parkkinen (1975) and Halden (1982). These are NW-trending ductile shear zones with consistently dextral horizontal components of displacement which have deformed and controlled the emplacement of diverse syntectonic intrusions (Map 1). It is therefore apparent that Group 4 structures result from a deformation that was homogeneous in a regional sense but showing substantial strain variations at more local scales. Relatively intense strain, such as within the Suvasvesi and Haukivesi shear zones, is attributed to dextral transcurrent faulting along the south-west margin of the Archaean craton (Gaál 1972). Between such zones, the crust at depth may have remained relatively rigid but such a postulate would be difficult to sustain against evidence for more ductile behaviour, such as indications that Archaean basement itself underwent syntectonic partial melting.

Weaker, NE-trending antithetic conjugate zones occur rarely in the study area, as elsewhere (Gaál and Rauhamäki 1971, Parkkinen 1975, Koistinen 1981) and provide a further indication of the nature of Group 4 deformation. According to stress-system analyses by both Gaál (1972) and Parkkinen (1975), these conjugate geometries are more likely to result from principal stress oriented N-S than from a simple shear couple resolved in a north-westerly direction across the whole region (Figure 22). A corollary of this deduction, combined with the evidence for transcurrent displacements, is that σ_2 would have been vertical, and the occurrence of such a stress system in association with syn-tectonic magma intrusion is pertinent to discussion of overall deformational history (see Section 6.6).

6.4   ORIGIN OF GROUP 5 STRUCTURES

Although Group 5 structures of different magnitude have
similar attitude and style, deformational processes at different
scales may have varied. A likely example of this is afforded by
Group 5 microstructures resembling the crenulation cleavages
described by Gray (1979) and White and Johnston (1981), for which
an origin by microfolding, solution transfer and recrystallization
is advocated. This obviates the need to postulate simple-shear
displacements along cleavage domains, although it does not preclude
subsequent shear strains from being accommodated preferentially
within cleavage planes. Hence, although they are geometrically
similar to macroscopic shear zones, microstructures do not
necessarily correspond to them mechanistically.

This conclusion leads to problems in interpretation of Group 5
structures, such as whether the lack of evidence for shear strain
at microstructural levels in fact precludes simple shear as a
viable mechanism in generating major Group 5 features. For instance
it could be argued that large-scale structures result from the
regional accumulation of apparent-offset increments across individual
cleavage domains that developed during homogeneous pure shear
flattening (and hence representing the XY plane of the finite-
strain ellipsoid). A corollary of this contention is that the
volumetric abundance of cleavage domains should relate directly to
the amount of finite strain. Although such a relationship is im-
practical to examine quantitatively at a regional scale, no clear
indication of more intense S_5 cleavage development exists within
the Orivesi shear zone or near the Haapajärvi fault zone. Rather,
both of these major structures seem better interpreted as ductile
faults or shear zones of normal sense, which is also consistent
with the vergence of most mesoscopic F_5 folds. This interpretation
could be reconciled with that based on microstructures if simple
and pure shear strains were consecutive rather than simultaneous or,

if crenulation cleavages derived by buckling or solution-transfer processes are not exactly parallel to major ductile shear zones or faults (so that the XY plane of the strain ellipsoid is not a shear surface). This may be difficult to discern in zones of relatively high strain, but near the Haapajärvi fault, at Kettämö, S_5 crenulations swing from parallelism with the main N-striking zone into a SSW trend, typically with more widely-spaced crenulations. Similarly, the Orivesi shear zone strikes almost due N but mesoscopic F_5 folds and crenulations can plunge between SSW and SW.

A complex dynamic analysis of such structures cannot be attempted here, but the simplest interpretation of the major 'normal-sense' Group 5 shear zones would have σ_1 nearly vertical as a result of isostatic reequilibration driven by the buoyancy of buried relatively low-density Archaean basement or large amounts of syn-kinematic felsic magma. At a smaller scale the occurrence of quartz veins perpendicular to F_5 and occasionally within S_5 may also signify that σ_1 was oriented almost vertically (Figure 22).

6.5 ORIGIN OF GROUP 6 AND YOUNGER STRUCTURES

The paucity of data inhibits interpretation of these later structures, although faults correlated with S_6 at Hammaslahti have a dextral sense of displacement and may correlate with NE-trending antithetic conjugate faults described from Kerimäki and Haukivesi by Gaál and Rauhamäki (1971) and Parkkinen (1975). The sense of displacement on these faults is effectively the opposite to that of Group 4 structures.

No consistent evidence for the origin of Group 7 structures was found and uncertainty exists over the nature of S_7 itself: if it were an extensional 'fracture' cleavage, a different stress field would be required than if they resulted from solution-transfer processes. In the former case, σ_1 could have been within S_7; in the latter case perpendicular to it.

6.6   INTEGRATION OF DEFORMATIONAL HISTORY

It is possible for successive deformations to be completely
distinct and independent from each other - Svecokarelian Group 3
structures superimposed upon Archaean basement structures are
pertinent examples of this. On the other hand, instead of envisaging
complex deformation as the result of various unrelated stress
systems, it is possible to consider the observed structural sequence in
terms of an area's evolving response to a more or less constant,
externally imposed system of forces. Accordingly, the following,
speculative outline of deformational history adopts the premise of
relative changes in the magnitudes of principal stress axes whose
variation in orientation is less significant. Although geological
factors are suggested that might have led to stress variations, the
nature of transitional stages and the reason why superimposed groups
of structures are so distinct from each other remains unknown.
Accepting the above premises and possible permutations the following
summary of Svecokarelian deformation in the study area is ventured.

(i)   Group 1 fabric development may have been initiated during
later stages of deposition in the Savo province, as a result of
increasing lithostatic load, taking σ_1, as vertical. There is no
direct evidence from the study area indicating the possibility of
significant horizontal stress, causing substantial horizontal dis-
placement but Koistinen (1981) recognizes a northwards translation
in the Outokumpu ore bodies at this stage of deformation. There is
no record of corresponding deformation in the Höytiäinen province.

ii)   The nature of major Group 2 structures indicates sub-
stantial tectonic translation and also strain accommodation by
recumbent folding; stratigraphical younging criteria and rare fold
hinge measurements (trending ENE-E) suggest translation in a NNW - N
direction onto the Archaean craton with possible stress configura-
tions shown in Figure 22.

In the Höytiäinen province, Group 2 structures are less significant and have been recognized as narrow zones of transposition or isoclinal recumbent folds. This appears consistent with deformation initiated in the south and west, and prograding northwards, affecting the Savo province.

iii) Group 3 structures are seen as products of continuing shortening along a N - NNW vector with L_3 in most cases representing synkinematic mineral growth or the intersection of S_3 with pre-existing elements (Figure 22). Evidence exists in high-strain zones for partial reoreintation of L_3 towards the proposed tectonic translation (or maximum principal elongation) direction. At higher structural levels within the Savo province recumbent moderately open to isoclinal folds are characteristic but at deeper levels, mylonites occur, suggesting more intense shear strain nearer the basement-cover interface. Alternatively, strain may have been accommodated by shear translations ("tectonic stacking") and bulk flattening (folding and layer-parallel shortening) at different stages of deformation (see Bell 1978). Indeed, limited evidence from pre- and syn-S_3 quartz veins supports the notion of more homogeneous deformation followed by rotational strain, associated with mylonite development and thrusting.

Group 3 structures form the dominant features of the Höytiäinen province, but are almost everywhere reoriented to some extent by younger structures.

iv) After Group 3 structural development, it is possible that the regional N-directed maximum principal stress diminished in an absolute sense. However as long as it remained greatest in a relative sense, the development of Group 4 structures can be explained by a concomitant increase in the vertical component of stress such that it changed from being the least to the intermediate principal direction (Figure 22). This situation is consistent with

appropriate trend and chirality for Group 4 structures. Geologically realistic factors for a steady increase in the vertical stress component are the increasing metamorphic grade of the overthrust craton margin (thermal climax between S_4 and S_5) and the abundance of mafic to intermediate syntectonic intrusions found in the Suvas- vesi and Haukivesi shear zones.

v) The geometry of Group 5 structures is consistent with a continued increase in vertical stresses, eventually exceeding the horizontal N-S principal component - the latter may in fact have steadily decreased in an absolute sense as well (Figure 22). Relative uplift in the east with respect to the west may be cor- related with a westwards increase in the abundance of synorogenic intrusions and climactic metamorphic grade (Korsman and others 1984). This suggests that in the study area, uplift associated with Group 5 structures was an isostatic response to the burial of relatively low-density Archaean basement. An essentially rigid-block response is implied, rather than gravitational re-equilibration by partial- melting and intrusion, possibly due to a combination of insufficient metamorphic grade and the relatively anhydrous nature of the mica- poor Archaean gneisses.

vi) It is more difficult to integrate the youngest structures into this deformational scenario. The stress analyses of Gaál and Rauhamäki (1971) and Parkkinen (1975) portray the N-S horizontal component as σ_3, with an E-W σ_1 and vertical σ_2, during the deforma- tion correlated with the development of Group 6 structures. This is probably best interpreted as representing a concomitant decrease in both vertical stresses (due to attainment of gravitational and thermal re-equilibration) and magnitude of the externally imposed forces initially responsible for deformation. Interpretation of Group 7 structures is subject to a convincing explanation of their origin.

FIGURE 21 :   DEFORMED METAPSEPHITE CLASTS

a   Variety of clast types measured in horizontal exposures,
    perpendicular to $S_2 - S_3$ foliation and oblique to L_3
    (Hammaslahti mine and Tikkala railway siding).

b   Dimensions of granitoid and graywacke clasts measured within
    $S_2 - S_3$ foliation (using ratio Y / X) and approximately perp-
    endicular to L_3 (ratio Z / Y).   Granitoid clasts are commonly
    prolate parallel to L_3 (indicated by $Y / X < 1$) whereas gray-
    wacke clasts are invariably so.

c   Granitoid and quartzite clasts measured perpendicular to L_3
    show more evidence of flattening in the vicinity of the
    Kettämö thrust zone than in F_3 normal limbs or fold hinges.

    In each diagram, diagonal reference line indicates ratio of
    unity (ie. equal values) for abscissa and ordinate.

a

+ dolomite   * intrabasinal graywacke

o granitoid   : quartzite and vein quartz

b

o granitoid Y / X

· granitoid Z / Y

* graywacke Y / X

+ graywacke Z / Y

c

· F_3 normal limb and hinge zone

o near Kettämö thrust zone
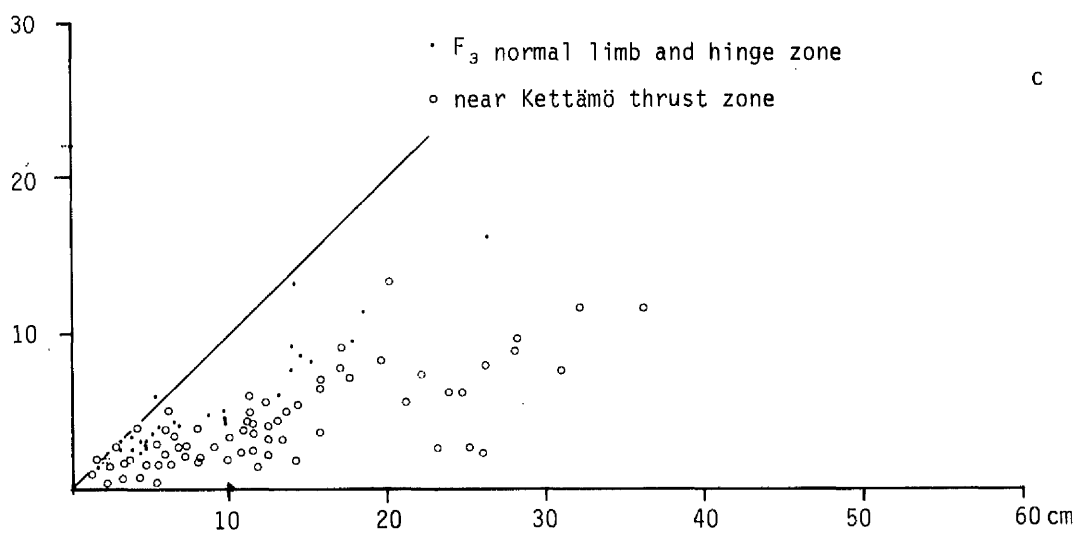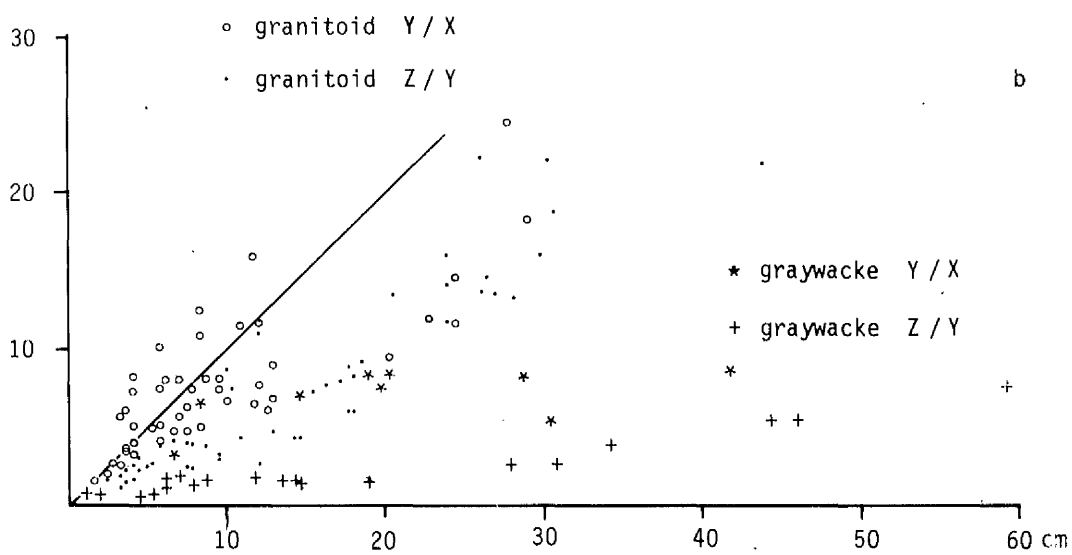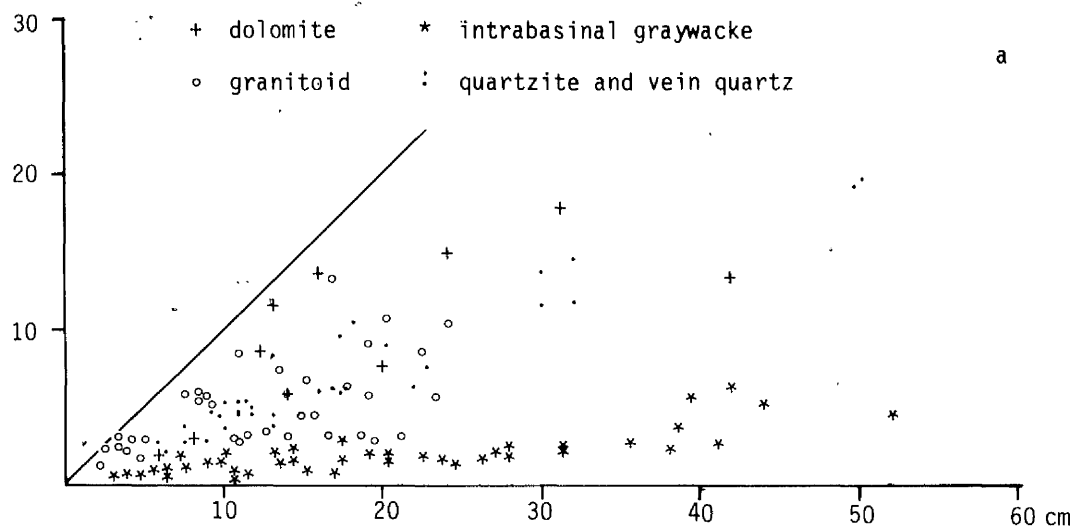
FIGURE 22 : INTEGRATED DEFORMATIONAL HISTORY

Envisaged principal stresses in relation to various stages of Svecokarelian deformation. Relative variation in magnitude of principal stresses is emphasized, rather than successive changes in orientation of externally applied forces.

(i)   Deposition occurred in extensional basin setting, with vertical stress as maximum, due to increasing lithostatic loading.

(ii)   Onset of compressive deformation along an approximately northwards vector caused major horizontal translation and F_2 recumbent folds.

(iii)   Continued compression resolved along thrusts and in F_3 recumbent folds. Mylonitic zones indicate high shear strain due to increasing applied external stresses (σ_1) or decreasing vertical stress (σ_3), since shear stress

$$\tau = \frac{\sigma_1 \quad \sigma_3}{2}$$

(iv)   Increase in vertical stresses resulted in Group 4 NW-trending shear zones with dextral chirality (such as Suvasvesi, Haukivesi). Such an increase in vertical stress correlates with the regional metamorphic peak and major plutonism, and is ascribed to thermal and gravitational buoyancy.

(v)   Continued increase in vertical stresses, in a relative, if not absolute sense is consistent with the geometry of F_5 folds and associated shear zones and is ascribed to isostatic uplift induced by the negative buoyancy of Archaean basement buried beneath the overthrust cover allochthon.

(i)

(ii)

(iii)

(iv)

(v)

PART III :   LITHOFACIES INVESTIGATIONS

1. INTRODUCTION TO LITHOFACIES STUDIES

1.1 DISCUSSION OF APPROACHES TO MAPPING

Absence of recognizable fossils, masking of large areas of bedrock beneath lakes and Quaternary glacial deposits, and a complex deformational history combine to preclude a straightforward determination of stratigraphical and lateral facies relations. Nevertheless, observations from individual exposures, and comparison with others in the vicinity can be useful in a sense analogous to incomplete drill hole data, being essentially one-dimensional analyses of a three-dimensional entity. Therefore, the most apposite manner in which to classify the lithologies of the study area was to define readily identifiable lithofacies devoid of interpretative, stratigraphical and temporal connotations. Though no constraints are placed on their distribution, most lithofacies are confined to discrete stratotectonic units; the gamut of lithofacies recognized within each stratotectonic unit constitutes an assemblage.

1.2 USE OF THE TERM 'LITHOFACIES'

The term 'facies' is in widespread and valid usage in a descriptive and also an interpretative sense, to indicate depositional processes and environments. This duality is evident in the following definition from Mutti and Ricci Lucchi (1975):

> "A facies is a layer or group of layers showing
> lithological, geometrical and sedimentological

characters which are different from those of
adjacent layers. A facies is considered to be
the product of a specific depositional
mechanism or several related mechanisms
operating at the same time."

While it is desirable that recognition of a facies leads
readily to a process or environmental interpretation, not
all observed features are diagnostic of a particular event or
setting, if indeed they are amenable to interpretation. This
latter applies especially to metamorphosed, recrystallized
sediments in which the only preserved parameter may be bed
thickness (subject itself to strain considerations) inferred
from gross changes in lithology.

Hence the term 'lithofacies' is employed here in a
descriptive sense similar to that in the first part of the
definition quoted above, and is intended primarily for field
observation at outcrop scale. That is, a single lithofacies
may describe a sequence of layers that display a diversity of
differing lithological, geometrical and sedimentological
features (where preserved), repeatedly interbedded, though
not necessarily in a consistent pattern or sequence.
Diagnostic metamorphic minerals could also be included in
the description, where they relate to a particular primary
lithology (as for example the distinctive tremolite
porphyroblasts of the Mulo lithofacies).

1.3   USE OF THE TERM 'ASSEMBLAGE'

This is preferred here to the term 'facies association'
which connotes a genetic or environmental relationship

amongst adjacent lithofacies (Reading 1978). Within the
study area there is the likelihood of some lithofacies being
in tectonic rather than depositional juxtaposition;
'assemblage' is intended here to be an informal, nonexclusive
term encompassing the lithofacies found in a stratotectonic
unit. Five such units are identified within the study area
and hence five assemblages whose geographic distributions are
outlined below, and displayed on Map 5.

1.4   DISTRIBUTION AND CHARACTERISTICS OF ASSEMBLAGES

1.4.i Pyhäselkä assemblage

This occurs in the southern and western parts of the study
area and extends to the north eastern shore of the lake
Pyhäselkä where its margin is defined by the coarse clastic
Niittylahdenrantatie and Sammallahti lithofacies. There is
also a region of mafic metapelitic schists nearby at Niva
included within the assemblage from inferred structural
criteria, but the bulk of the assemblage is composed of the
monotonous and homogeneous metapsammitic schists of the
Rääkkylä lithofacies.

1.4.ii Oravisalo assemblage

This is present as a tectonic window within the Pyhäselkä
assemblage at Oravisalo and possibly to the N at Liperinsalo
(based on the studies of Saksela, 1933). There are a number
of distinct lithofacies associated with allochthonous inliers
of Archaean granitoid gneiss, namely the Quarry and Kankaala
lithofacies consisting respectively of metapsephites and impure
metaquartzites and the Oravisalo post office lithofacies, com-
prising metabasites or mafic metapelites.

1.4.iii Kettämö assemblage

This is represented by a well-defined stratotectonic
unit between the Niittylahdenrantatie lithofacies of the
Pyhäselkä assemblage and the Hammaslahti assemblage to the
NE; it abruptly juxtaposes the latter as the hanging wall of
the Kettämö thrust zone. The assemblage is characterized by
well preserved depositional features within the coarse clastic
metasediments of the Kettämönniemi lithofacies and the more
thinly-bedded Salonkylä lithofacies. A small occurrence of
granitoid gneiss and metadolerite is associated with coarse
metasediments of obvious granitic provenance but their
mutual relations are not apparent.

1.4.iv Hammaslahti assemblage

The most heterogeneous assemblage within the study area,
this includes a large proportion of calcareous, micaceous
and siliceous metapelite (Mulo and Kukkupää lithofacies),
terrigenous metasediments, locally as thick-bedded coarse
clastic deposits (Tikkala, Kilpelänkangas and Suhmura
lithofacies) and rather more mafic metasediments and possible
metatuffs (Rauansalo and Kalliojärvi lithofacies). It also
contains the Hammaslahti Cu-Zn ore bodies.

1.4.v Kiihtelysvaara assemblage

This was comprehensively described by Pekkarinen (1979a)
and was only briefly examined during the course of this study,
primarily for comparative purposes as the region contains a
particularly well preserved and complete succession of
Lower Proterozoic strata. Relationships with Hammaslahti
assemblage lithofacies have not been determined.

2.   DEPOSITIONAL BASEMENT TO LOWER PROTEROZOIC SEQUENCES

2.1   RECOGNITION, DISTRIBUTION AND SIGNIFICANCE OF ARCHAEAN BASEMENT

Granitoids and gneisses are widespread throughout easternmost Finnish Karelia and many have been isotopically dated as late Archaean in age (Kouvo & Tilton 1966). These have intruded a yet older, diverse sequence of metavolcanic and metasedimentary lithologies whose base has not been identified (Saxén 1923, Nykänen 1971b). It is demonstrable in numerous places that all these Archaean lithologies are unconformably overlain by Proterozoic metasediments and hence that they were exposed as source terrains and as depositional basement during Early Proterozoic time.

The major unconformity is well documented and illustrated by Pekkarinen (1979a) in his study of the Proterozoic sequence between Kiihtelysvaara and Värtsilä on the Soviet border. Although steeply dipping, it has not been substantially disturbed tectonically and frequently preserves a transition from pristine basement through weathered horizons (now metamorphosed palaeosols) to metasediment derived wholly from the underlying basement (see Pekkarinen 1979a p. 29). For further description of lithologies and their distribution the reader is referred to both Pekkarinen's study and those of Frosterus and Wilkman (1920) and Nykänen (1971b).

Further west (including the area of the present study), gneissose granitoids are exposed as inliers in structural culminations, surrounded by Proterozoic metasediments; thrusted tectonic contacts are common. (Frosterus & Wilkman 1920, Gaál 1964, Park & Bowes 1983). Fortunately, unconformable

relations between the two lithologies have also been preserved,
affirming that these granitoid gneisses represent depositional
basement and not the more highly metamorphosed variants of
Proterozoic metasediments.

One such basement inlier is found within the field area
on the island Oravisalo, and a small, somewhat contentious
occurrence is present near Hammaslahti, above the Kettämö thrust.

2.2   ORAVISALO BASEMENT LITHOLOGIES

Gneissose granitoids crop out within a crescentic belt
4 km in length in the SE part of Oravisalo and appear to constitute
at least four thrusted tectonic units that protrude through
Proterozoic metasediments of the Rääkkylä lithofacies.

The largest group of outcrops is near the low ridge of
Kakkari; immediately to the N and W metasediments of the
Rääkkylä lithofacies are juxtaposed. Though the contact is not
exposed, the lack of a coarse clastic basal facies in the
metasediments nearest the basement outcrops suggests an
intervening thrust rather than an unconformity.

Lithologies dominated by quartz and feldspar are well
exposed in a bluff at locality 73.4 16.4 and at 74.2 17.2. These
vary from medium grained and equigranular to foliated with
extremely elongate quartz grains and sporadic feldspar augen,
invariably smaller than 1 cm. In addition to this intense, in
places mylonitic foliation, planar stromatic migmatites are
present with alternating leucosome and melanosome layers
several cm thick (Figure 23 a). Only rarely are there massive
developments of more mafic granitoid material (Figure 23 e).

Homogeneous granitoid gneiss and stromatic migmatites alike are associated with concordantly deformed mafic bands attaining 30 cm in thickness, and having mineralogy frequently dominated by retrogressive biotite (Figure 23 d). The only obviously transgressive neosome identified occurs as quartz-feldspar veins less than 5 cm thick and these too variously exhibit fabrics from granular to elongate (Figure 23 c).

An arcuate topographic depression without exposures may reflect a lithological variation, or structurally weaker horizon that separates the basement at Kakkari from the second major group of gneissic granitoid outcrops near Oravisalo quarry. At least three thrusted tectonic units are inferred from the presence of two interdigitated metapsephite horizons, one having been deposited unconformably on gneissic granitoid basement as seen in a quarry at locality 75.7 15.8,(Figure 2). Using the combined descriptive nomenclature of Mehnert (1968) and Johannes (1983) these basement lithologies are best described as stromatic migmatite with mesosome predominating and consisting of medium to coarse-grained homogeneous granite and granodiorite (Figure 23b). Also present are felsic leucosomes less than 10 cm thick which from their colour contrast with the mesosome serve to delineate fold morphology and also small dislocations (Figure 3 e).

In addition to these well defined lithological alternations are more widely spaced diffuse leucocratic zones often accompanied by abundant sulphide mineralization and clots or selvedges of biotite porphyroblasts (Figure 23 f). In the quarry area mylonitic or cataclastic textures are only expressed near the unconformity in lithologies which are notably richer in muscovite. A further lithology was exposed in 1981 but entirely removed by subsequent quarrying. This was of thin discordant

dykes, dark in colour due to an abundance of very fine grained biotite. Unfortunately their intrusive relations with respect to the unconformity could not be ascertained.

Structurally below the quarry, at locality 75.7 15.9 there occurs another metapsephite horizon and below this a variety of amphibole granitoid that tends to be more gneissose than the quarry lithologies and also less differentiated into leucosome and mesosome; neither was any discordant neosome apparent. A mylonitic foliation is pervasive at locality 75.2 15.9 with individual quartz grains forming lamellae up to 10 cm long parallel to the principal tectonic lineation.

Some of the more felsic horizons are particularly difficult to distinguish from deformed meta-arkose unless the latter contains obvious coarse detritus. Distinctly more mafic, impersistent amphibolite bands are distributed throughout the sequence, as in the basement at Kakkari. They differ from the metadykes in the quarry both in their mineralogy and in their concordance with the major gneissosity, rather than transection of it.

2.3   BASEMENT NEAR HAMMASLAHTI

Immediately above the Kettämö thrust are two adjacent exposures consisting of felsic granitoid gneiss and metadolerite, first located by Pekkarinen (1979b) after the recognition of granitic detritus in nearby coarse clastic metasediments (Kettämönniemi lithofacies of this study). A massive leucocratic, equigranular lithology constitutes the bulk of both exposures with the only neosome material being distinctly discordant dark smoky quartz veins and somewhat thinner and paler quartz veins that are concordant with and help delineate folds within the granitoid gneissosity.

The middle part of the exposure at 95.8 27.1 consists
however of a metadolerite unit over 20 m thick. The central
part shows evidence of fracturing and a weak crenulation
schistosity and preserves a fabric of randomly oriented
amphibole prisms up to 2 mm long in a feldspathic groundmass.
As both margins are approached a gneissose fabric becomes
increasingly evident: it is conformable with the contact
itself and is present there as thin leucocratic to melanocratic
alternations depending upon amphibole content.

The nature of the contact itself and temporal relations
of the metadolerite with respect to gneiss development in the
enclosing granitoid is unclear. It is worth noting however
that discordant metadolerite intrusions are common within the
basement at Kiihtelysvaara (Nykänen 1971b) and the basement
inlier at Sotkuma to the N (Huhma 1975).

Though correlation of these Kettämö exposures with
Archaean basement lithologies is not in dispute, their
relationship with associated metasediments is problematical;
the possibility of them being large glacigenic erratics needs
to be examined. Evidence against the exposures being in situ
could be adduced from the fact that no unconformity is seen and
attempts to excavate between the gneiss and the nearest abutting
metasediments (Pekkarinen 1979b) indicate that the latter
belong to an erratic block some 10 m across. Nevertheless,
when taken in concert there are several lines of reasoning
leading to the conclusion that the exposures do represent
outcrops of bedrock. Firstly, exposures are an order of
magnitude larger than any other erratic blocks encountered;
boulders of gneissic granite within glacial deposits and
metapsephites of the Kettämonniemi lithofacies seldom exceed

1 m in diameter. The nearest known exposures of basement
are 20 km to the NW at Sotkuma, along the most recent
direction of ice movement, and a similar distance to the
E near Kiihtelysvaara. If the Kettämö basement rocks were
indeed transported such a distance it remains still to

(1) explain the seeming absence of other large
erratic blocks, and

(2) dismiss as coincidence the fact that these two
surviving blocks were finally deposited immediately
above a major thrust in association with a restricted
development of basement-derived coarse clastic deposits.

From this evidence, the gneiss is interpreted as
an isolated tectonically incorporated enclave of Archaean
basement, and its existence is of significance to
discussions of Proterozoic sedimentation and subsequent
deformational events.

FIGURE 23 : ARCHAEAN BASEMENT LITHOLOGIES

a   Stromatic migmatite. The white quartz vein and open flexure
     above the hand lens are ascribed to Svecokarelian Group 3
     deformation. (Oravisalo 74.2 17.2).

b   Stromatic migmatite with homogeneous grey mesosome and thinner
     leucosomes enabling small folds and dislocations to be
     recognized. (Oravisalo 75.7 15.8).

c   Slightly discordant equigranular felsic neosome with biotite
     selvedges possibly constituting melanosome. Mesosome, or host
     lithology, is a typical banded gneiss. Natural size.
     (Oravisalo 74.2 16.9).

d   Fine-grained metadykes, now rich in biotite, discordant with
     respect to banding in stromatic migmatites. (Oravisalo 75.7 15.8).

e   Homogeneous medium to coarse-grained gneiss considerably
     more mafic than is usual for Oravisalo basement. Quartz
     and associated buckling are considered to be products of Group 3
     Svecokarelian deformation, unrelated to gneiss development
     (Oravisalo 73.6 16.9).

f   Irregular, diffuse leucosome with thin seams rich in biotite
     porphyroblasts and sulphide minerals, transecting the main
     gneissosity. (Oravisalo 75.7 15.8)
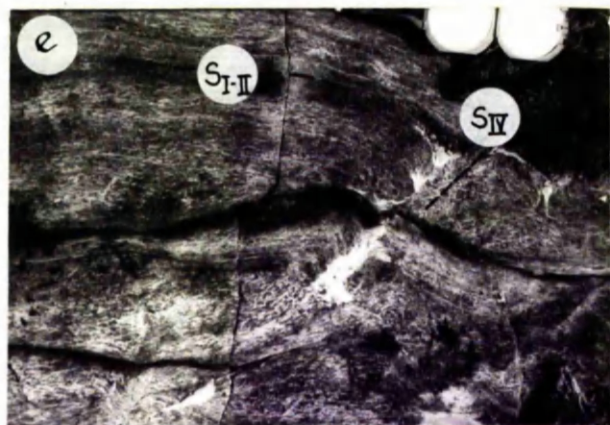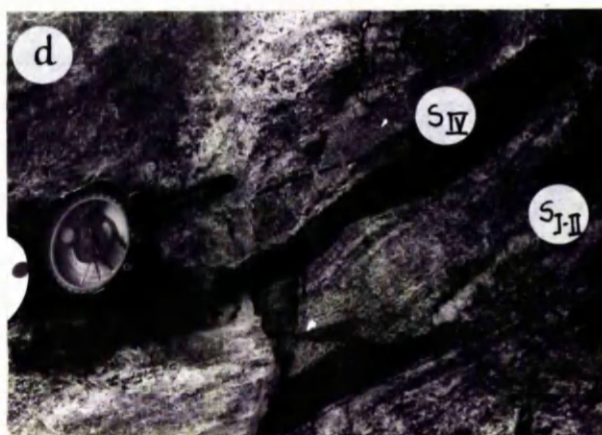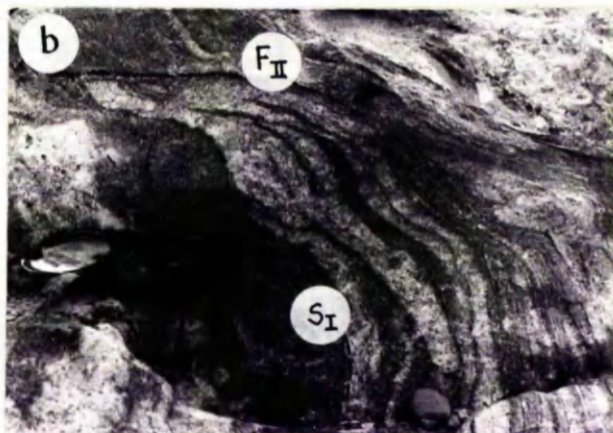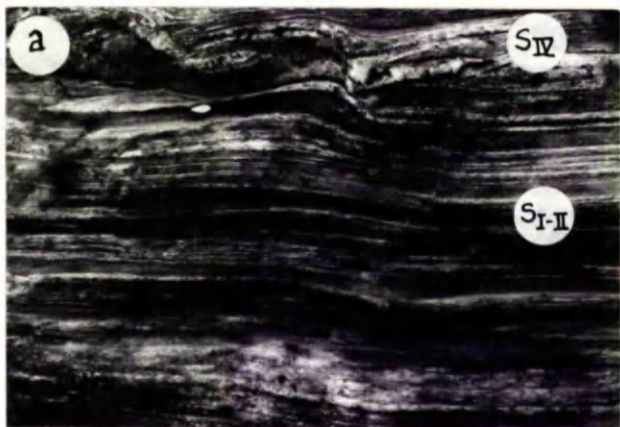
3.   LOWER PROTEROZOIC LITHOFACIES

3.1   ORAVISALO ASSEMBLAGE

3.1.i Oravisalo quarry lithofacies
Distribution and reference exposures:

At locality 75.8 15.9, 2 km N of Oravisalo post office, a conglomerate horizon is present between two thrusted inliers of Archaean granitic gneiss. The upper thrust sheet itself preserves an unconformity, in turn overlain by another metapsephite, well exposed nearby in a recently enlarged quarry. The two conglomerates are lithologically distinct from each other, and neither were encountered elsewhere.

Depositional units and structures:

Lithofacies A (between the two gneiss thrust sheets) yielded no younging criteria, unless the granitoid cobble in Figure 24 b is interpreted as a drop stone (see discussion below). The exposed sequence consists of several meters of micaceous schist overlain by 3-5 m of arkosic metapsammite. The meta-arkoses appear very similar to the more felsic, massive granitoid against which they are tectonically juxtaposed, and clear discrimination between them is some-times only possible where distincly coarser granitoid clasts are present, or more particularly mafic clasts, since no mafic xenoliths were observed in the granitoids. Bands of amphibolite are seen however and their weathered products may be present in the meta-arkose as clasts attaining 1 m in maximum dimension. Amphibole porphyroblasts are also common throughout the meta-arkose and may denote metamorphosed palaeosols, such as recognized in similar contexts, mantling basement gneiss at Sotkuma (Gaál and others 1975), and Kiihtelysvaara (Pekkarinen 1979a).

Granitoid clasts seldom exceed 10 cm in size, granules and finer grains being ubiquitous and only slightly finer than equivalent mineral types in the granitic gneiss. Deformation and recrystallization have obscured the original grain shapes but for pebbly clasts, rounding and sphericity is generally low; mafic clasts tend to be more elongate which may be due in part to relatively higher ductility during deformation. Obvious depositional features are also lacking, apart from the erratic distribution and poor sorting of coarser clasts (Figure 24 a) and interbedding of meta-arkose layers with somewhat better sorting, lacking pebbly detritus.

Lithofacies B is much easier to recognize as a metasediment because the matrix enclosing pebbles is considerably finer-grained and richer in biotite than either the clasts or the immediately underlying basement gneiss, (Figure 24). An unconformity appears to be represented and preserved intact although limited displacement is possible, given the platy, sheared nature of the uppermost basement gneiss.

Along some 20 m of section this basal metapsephite does not exceed 50 cm in thickness and locally clasts are very sporadic. Clasts measure less than 10 cm in maximum dimension and all are of gneissic or equigranular felsic granitoid, closely resembling subjacent basement lithologies. Sections parallel to the major tectonic lineation, as in Figure 24e and f, show correspondingly elongate clasts. This seems to have been tectonic in origin since sections perpendicular to this direction reveal clast cross-sections of lower eccentricity and less alignment of long axes.

The majority of clasts are subangular and it should not be assumed that the metapsephite represents a sedimentary deposit and not a differentiated, metamorphosed palaeosol. The presence of some rounded clasts (Figure 24 c, e) and the fact that the metapsephite matrix is indistinguishable from the immediately overlying metapelites suggests however that it is a clastic deposit. The strongly sheared, uppermost part of the basement is markedly richer in chlorite and particularly muscovite than that further below the unconformity. This difference could be tectonic in origin, as with phyllonite development but conversely, shearing may have been located preferentially in this horizon as it already contained a higher proportion of phyllosilicates as a result of weathering and pedogenesis.

These metapsephites resemble a more extensive sequence mapped by Saksela (1933) in an anticlinal culmination some 12-15 km to the N. At Torohvenniemi, towards the eastern margin of this domal structure a poorly sorted arkosic metabreccia occurs, containing angular granitic fragments and larger orthoquartzite clasts up to 30 cm across (Figure 28 e, f). Although Saksela described "satroliths" on Niinikkosaari in the west of the area, as being transitional into gneissic granite, all are interpreted here as more schistose metapsephites. At locality 69.6 29.5 schistose metapsephites and meta-arkoses are quite clearly interbedded with massive metapsammites indistinguishable from those of the Rääkkylä lithofacies, affirming their sedimentary origin.

3.1.ii Kankaala lithofacies

Distribution and reference exposures:

Schistose metaquartzite and meta-arkose are seen in an arcuate line of outcrops, best exposed at 74.8 15.4, 74.7 15.2

on the property "Kankaala" and on the Liperi-Rääkkylä road
at locality 75.3 15.1. Southwards and probably structurally
below occurs the Oravisalo post office lithofacies whereas to
the N are granitoid gneisses upon which the metaquartzites
are either thrust or unconformable.

The very high quartz content, though only sometimes
evident as detrital rather than deformed igneous grains,
and the greater abundance of muscovite and pyrite (with
accompanying sulphurous odour) best distinguish the lithofacies
from the felsic granitoid gneisses of the basement.

The extent of metamorphic recrystallization is not
clear but it appears that in places an almost mylonitic
fabric has developed from a protolith of feldspathic quartzite,
containing granules of both quartz and feldspar up to 6 mm
across. Matrix content would have been low, reflected in
the present overall dearth of mica (Figure 24g, h).

3.1.iii. Oravisalo post office lithofacies

Distribution and reference exposures:

This was represented by several outcrops near the
Oravisalo post office and is probably not of greater extent
than indicated on the lithofacies map. Support for this
notion is inferred from the absence of a notable positive
geomagnetic anomaly on the published reconnaisance maps,
such as would be expected from these lithologies, if
widespread.

Depositional units and structures:

At 74.8 14.8,dark, schistose, fine-grained amphibolite
prevails, mottled by small lenticular feldspar grains, and

occasionally alternating with lighter pink to grey, equally
schistose, felsic layers. No primary textures have been
identified, of intrusive, extrusive or depositional nature,
unless the small lensoid feldspar grains are deformed
phenocrysts. Several hundred meters along strike a coarser-
grained, less schistose metabasite preserves a gabbroic
texture, with stubby amphibole crystals up to 5 mm long and
apparently interstitial feldspar.

It still remains to demonstrate that this lithofacies
belongs to the Proterozoic, and not to the Archaean
basement sequence. Contacts with adjoining lithofacies are
not seen but thrusted intercalation of allochthonous slivers
of basement is demonstrable nearby. Though mostly gneissic
granitoid, these and other exposures of basement do
contain some mafic lithologies which can however be distinguished
from the Oravisalo post office lithofacies using the following
criteria:

1) basement amphibolites tend to be thinner than
30 cm, though this observation does not preclude them from
being thicker.

2) they share a structural unity with the granitic
gneisses, both having been deformed into isoclinal folds
which are in turn overprinted by a schistosity ascribed to
Svecokarelian deformation.

3) syntectonic Svecokarelian metamorphism of mafic
basement lithologies is invoked to explain them as typically
being fine-grained biotite-rich schists. These features
contrast with those in the dominant amphibole-feldspar
assemblage of the Oravisalo post office lithofacies but do not
unequivocally indicate cover or basement affinity.

Partly surrounding the metapsephites in the domal
culmination mapped by Saksela (1933) and mentioned above,
is an association of graphitic and sulphidic metapelites,
and metabasites, including a gabbroid. The Oravisalo post office
lithofacies might well be a correlative of this whence, in
deference to Saksela's study, they could be together
designated the Karhunsaari lithofacies. This correlation
does not solve any dispute over absolute age, but if valid,
indicates the lithofacies to be of regional stratigraphical
relevance.

3.2   PYHÄSELKÄ ASSEMBLAGE

3.2.i Rääkkylä lithofacies
Distribution and reference exposures:

In areal extent, this is the most abundant lithofacies
but at the present erosion level it is not known to occur
anywhere east of the Niittylahdenrantatie lithofacies. It
resembles a lithology widespread outside the study area,
occurring at Outokumpu to the NW where it has been classified
as homogeneous massive and bedded mica schist by Gaál and
others (1975) and southwards near Savonranta (Halden 1982)
and Kerimäki, where it has been more metamorphosed and
disrupted by intrusive bodies. Typical lithologies are well
exposed on the north western shore of the island Oravisalo,
particularly at locality 67.4 19.6, in bluffs at locality
74.2 17.5 and near the eastern limit of its distribution, at
94.1 20.2.

Depositional units and structures:-

The lithofacies is remarkably monotonous, in that
metamorphic recrystallization has resulted in the more massive
metapsammites superficially resembling a fine-grained

granodiorite or diorite (Figure 26 a, b). At many exposures
there exists no evidence of sedimentary layering and even
where original bedding may be discerned, depositional structures
are only rarely preserved (Figures 25a, b, 26 c). However,
impressive examples of these have been found at locality
94.1 202 near the eastern margin of the lithofacies' present
extent (Figure 25c, d, e, f, g), where three successive beds
preserve dune foresets in their lower portions, each passing
upward into planar stratified laminae. Bed forms are not
tectonically transposed as are for example the load casts in
Figure 25b, so that the maximum foreset angle of repose, of
15° is probably close to its original value, and bed forms
have an amplitude generally less than 10 cm and wavelength of
20-80 cm.

These beds are immediately overlain by a massive, much
thicker metapsammite which has caused spectacular basal load
casts and flame structures (Figure 25c, d, g). The thin
biotite selvedge in Figure 25d silhouettes the upper surface
of the load casts, but does not represent a pelitic film
between the two beds, or the topmost part of the underlying
bed.

In fact, laminated metapelite is virtually absent from
the entire lithofacies, one exception being a disrupted layer
or horizon of ragged, intra-basinal clasts at locality
79.1 12.8.

Several kilometers S of the study area at Puhos
(locality 92.7 97.1), typical Rääkkylä lithofacies metapsammites
are interbedded with fissile graphitic metapelites that may
attain 1 m in thickness. But within the study area, even
readily recognizable thinly bedded layers, such as those in

Figure 26 c and d, with respective modal thicknesses of
20-30 cm and 5 cm, cannot be properly described as metapelites.

Calcareous concretions are encountered throughout the
lithofacies but are nowhere seen to exceed 1 m in length,
usually being only 10-30 cm and about 5-10 cm thick (Figure 26a).
They are usually well-differentiated mineralogically with a
pale rim surrounding a darker, greenish core rich in randomly
oriented amphibole crystals up to 3 cm in length. Where
abundant, the concretions tend to be concentrated within or
define a particular lithological horizon but no evidence was
found conclusively demonstrating them to represent boudinage
of a continuous calcareous layer, rather than post-
depositional chemical phenomena.

3.2.ii   Sammallahti lithofacies
Distribution and reference exposures

Like the Niittylahdenrantatie lithofacies, this in many
ways resembles the Rääkkylä lithofacies, with which it is
apparently intercalated. It is restricted to an area of
several km^2 inland from Sammallahti (Map 2). The most
representative exposures are 91.7 20.0 and 91.5 20.3 .

Depositional units and structures

The lithofacies is characterized by an abundance of
quartzose granules 2-8 mm in diameter though they are never
sufficiently abundant to constitute an entire, supporting
framework (Figure 26 e, f). Because of their original massive
nature, and metamorphic recrystallization, no relict
depositional features are evident. Sharp boundaries between
granule-rich layers and Rääkkylä facies type are seen,
frequently marked by biotite selvedges or foliae, but more
usually lithological variations are poorly defined, with diffuse

aggregates richer in granules amongst finer massive metapsammite. Occasionally these granule-rich pods do appear as discrete intraformational clasts, up to 20 cm across, but no other large clasts of any kind were noted. As in the Rääkkylä lithofacies, calcareous lithologies are present in the form of zoned concretions or more extensive lenticular layers several cm thick. Metapelitic lithologies are absent.

3.2.iii Niittylahdenrantatie lithofacies
Distribution and reference exposures:

Outcrops of the lithofacies may be traced for 6 km along a low but persistent strike ridge, parallel to the NE shore of the lake Pyhäselkä. Neither tectonic nor depositional contacts are seen with the Rääkkylä facies to the W or Salonkylä lithofacies to the E. Lithological features are well displayed at localities 91.3 31.7, 91.2 32.4 and 91.3 32.2 .

Depositional units and structures:

Two principal lithological types were identified - the first, internally homogeneous metapsammites, both massive and stratified, are very similar to lithologies in the Rääkkylä lithofacies (Figure 27c, g). The other, which is diagnostic for the lithofacies is diamictite, comprising homogeneous clasts of Rääkkylä facies type, with very variable size and abundance, in poorly stratified metapsammites and metapelites.

Beds of homogeneous type are from 5-80 cm thick. The thinner beds are tabular at outcrop scale and differ from those of the Rääkkylä facies only in having a higher, yet variable proportion of interbedded metapelite (contrast Figure 26 c, d, with Figure 27g). Some metapsammites show

intercalated or transitional contacts with metapelite,
suggesting that they are codepositional, but distinct graded
bedding was only once noted, the majority of beds being
medium-grained and massive throughout. The thicker beds
however can show planar or low-angle cross-stratification,
with occasional preservation of scoured channels (Figure 27d).
Calcareous concretions are erratically distributed throughout
the metapsammites, but tend to be absent from metapelites
(Figure 27 c,e,g).

The diamictites have either a stratified metapelitic,
or more commonly, a metapsammitic matrix (Figure 27 a, b, c),
enclosing very variable proportions of clasts, ranging from
thin metapelitic flakes to rafts of metapsammite over 5 m in
length and up to 0.5 m thick (Figure 27a, b, e). Clasts are
dominantly internally uniform and generally indistinguishable
from interbedded metapsammites and those of the Rääkkylä
lithofacies. The remaining clasts are metapelitic flakes,
and some calcareous concretions which may have remained
coherent during reworking and comminution of their metapsammitic
host layers.

Most clasts are elongate within depositional planes,
reflecting to a large extent their initial tabular shape but
clearly tectonic strain is also to be expected (Figure 27 b).
Though many clasts retain sharp boundaries with matrix, as
seen in Figure 27 a, b, some such as the topmost bed represented
in Figure 27e, merge gradually and poorly defined aggregates of
metapsammite are seen.

3.2.iv Niva lithofacies
Distribution and reference exposures:

This lithofacies comprises a number of scattered outcrops
at Niva, several km SW of Hammaslahti township. There is some
evidence for a transition into lithologies of the Rääkkylä
lithofacies to the W but other boundary relations are
concealed beneath the bay Hammaslahti and surficial deposits.
The most informative exposures were 93.8 24.6 and 93.9 24.6 .

Depositional units and structures:

Three distinct lithologies are distinguished. Dominant
is a homogeneous dark grey metapelite, fissile only where a
discrete spaced cleavage dissects it and with abundant randomly
oriented biotite porphyroblasts. The metapelites constitute
uniform beds up to at least 5 m in thickness, intercalated
with the other two more complex lithologies: finely
stratified but massive grey to white alternations and
distinctly green, internally amorphous layers (Figure 28 a, b)
containing chloritoid and rosettes of actinolite.

Commonly, the latter form bands 0.5 - 5 cm thick, alternating
with the massive grey metapelites, from which actinolite is
absent. Boundaries between the two layers are generally
diffuse but at 93.8 24.6 one of these actinolite-rich layers
10-30 cm thick, displays a very sharp contrast with the
(structurally) underlying metapelite, with an almost purely
biotite selvedge between. Furthermore, the boundary is cuspate
upwards into the amphibolite-bearing layer, suggesting an
interface instability analogous to basal load casts in clastic
deposits. Depositional features are lacking from this bed
however, it being massive and knobbly due to intergrowth of
the porphyroblastic actinolite rosettes. The upper surface

of this layer is not so sharply defined but above it is seen
the third lithological type - the massive grey and white
laminites.

These have a tough, chert-like though not vitreous
appearance and show fine (1-5 mm) laminae which are not always
planar and may be laterally impersistent (Figure 28c, d).
Sporadic massive, but lithologically identical layers up to
7 cm thick are interbedded (Figure 28c, d).

Diffuse porphyroblasts of chloritoid or amphibole up to
5 mm may remain in relief on weathering, creating a false
impression of grading - no clastic grains are visible with the
unaided eye (Figure 28 c, d). Possible truncations of laminae
suggest that they are sedimentary in origin, and preserved
due to lithological competence of the bed, rather than the
products of metamorphic differentiation. They are not unlike
lithologies commonly encountered within the Kukkupää
lithofacies.

In the most westerly outcrops available for study, the
metapelitic lithology appears to be interbedded with biotite
metapsammites like those of the Rääkkylä lithofacies. The
relationship is a potentially valuable one for stratigraphical
interpretation but is unfortunately insufficiently exposed
for verification.

3.3   KETTÄMÖ ASSEMBLAGE

3.3.i Kettämönniemi lithofacies
Distribution and reference sections:

Metapsephites are diagnostic of this lithofacies which
occupies several km² NW of Hammaslahti township. Any

continuation to the SE along strike is obscured by surficial
lacustrine deposits and moraines whereas to the E the
lithofacies is truncated at the Haapajärvi fault zone. The
Kettamö thrust delineates its northeastern extent; in the
footwall of the thrust the Kilpelänkangas lithofacies
resembles finer lithologies within the Kettämönniemi
lithofacies. Northwards and westwards is a gradual transition
into the finer-grained Salonkylä lithofacies. Useful exposures
are referred to in descriptions below.

Depositional units and structures:

Four subfacies are defined within the lithofacies.
Although all are lithologically and texturally very similar,
the subdivision is according to dominant though not
necessarily exclusive clast type. This procedure is both
valid and convenient as the petrological distributions so
defined correspond closely to discrete stratigraphical or
tectonic horizons. Locally intense deformation has not
precluded recognition of numerous depositional younging
features, enabling the mutual stratigraphical relations of
three subfacies to be established. There also exists evidence
for lateral facies transitions between them across intervals
of 2-3 km.

The Pitkänurmi and Karvanurmi subfacies typically
contain recognizable granitoid detritus and an appreciable
number of sedimentary metaclasts, excepting orthoquartzite
fragments. Because they are structurally separated from each
other they are described individually. The Petäjikkökallio
subfacies contains more granitoid detritus than intra-basinal
sedimentary metaclasts but diagnostic is the presence of
orthoquartzite clasts. The Lentosärkkä subfacies lacks
pebbly clasts of granitoids or orthoquartzite but contains
sedimentary meta-clasts similar to enclosing host lithologies.

Karvanurmi subfacies

Three lithologies are expressed, with a tendency for detritus to become finer away from the basal Kettämö thrust, which also corresponds to the overall stratigraphical younging direction deduced from preserved depositional features.

Typical of the lowermost part is meta-arkose of modal grain size 2-6 mm (Figure 29 a,e) with additional granitoid clasts that seldom exceed 10 cm in maximum dimension. At locality 94.9 28.1 the Kettämö thrust is exposed, with mylonite developed within pebbly arkose. Magnitude of strain has not been resolved but a minimum estimate for thickness of these meta-arkoses is in excess of 10 m. No depositional features are in evidence here though this is not surprising given the massive nature of the beds and the amount of deformation. Where grain relations have not been tectonically obliterated it is however evident that sorting was poor and many clasts were not well rounded when deposited. Matrix content is evidently low which contrasts markedly with another lithological type that is commonly interbedded (Figure 29a, b).

These could be termed metadiamictites, as 'metapsephite' fails to convey the frequent chaotic nature of the deposit and extreme variation in size and proportion of clasts. Individual bed thicknesses may attain 3 m and clasts in excess of 1 m are sporadically present, though the modal size is 5-25 cm (Figure 29 c - g .). These are almost exclusively sedimentary metaclasts, granitoid detritus seldom attaining 4 cm in size. Many clasts preserve a distinct internal stratification, suggesting an intra-basinal origin. For example, the large clast in Figure 29 c and d (locality 94.5 28.5') is a well-bedded meta-arkose, itself containing granitoid clasts up to 5 cm across. It may be that isolate granitoid detritus of similar size derives from attrition of these large arkosic clasts, rather than from direct erosion of exposed granitoids.

Figure 29e shows a massive, graded metaclast with feldspar granules at the base, which is more characteristic of the higher beds of the subfacies. Alternating quartzose metapsammite and less abundant metapelite overlie these forming massive beds 10-50 cm thick and coarse clasts are altogether absent.

A thrust zone, or attenuated lower limb of a recumbent fold separates these beds from the Petäjikkökallio and Pitkänurmi subfacies to the W.

Pitkänurmi subfacies

This superficially resembles the Karvanurmi subfacies but displays more evidence of grading and bed thickness: granules of quartz and feldspar in the basal parts of meta-arkoses give way to massive quartzose metapsammite, with metapelitic tops thin or absent and bed thickness varying between 10 cm and over 10 m (Figure 30e, f). Associated with these is a polymictic metapsephite with a variety of sedimentary metaclasts and also abundant vein-quartz and granitoid detritus rarely attaining 20 cm in maximum dimension. This is usually within the major tectonic schistosity, but original grain shapes and roundness can be retained (Figure 30a).

As in the Karvanurmi subfacies, chaotic diamictite with a highly variable amount of matrix occurs; here (localities 94.4 28.4, 94.4 28.5) they are represented by two distinct but intercalated lithologies (Figure 29f, g). Those beds with a higher proportion of metapelite have fewer clasts and these tend to be lithologically similar to the enclosing matrix. These contrast with the interbedded layers containing almost exclusively granitoid metaclasts, in sufficient abundance for the rock to be locally clast supported (Figure 30c, d). Internal bed forms are absent, except for pockets or lenses of weakly stratified finer-grained lithologies and clast size varies considerably, from several mm to 1.4 m (Figure 30c, d).

One of these boulders appears to be fractured yet not
disaggregated (Figure 30 d).

Of significance is the fact that these coarser beds
are not the most basal but overlie the finer deposits depicted
in Figure 29 f; ascending the sequence, granitoid detritus
becomes coarser and predominates over intra-basinal metaclasts.
Notable too is the lateral impersistence and thickness
variations of the clast-rich and clast-poor lithologies,
giving rise to lenticular and intimately intercalated strata,
as illustrated in Figure 30 b.

Stratigraphically above this locality, and to the NW
along strike, coarse detritus is only infrequently encountered
and metapsammites and metapelites are not clearly distinguished
from those of the Salonkylä lithofacies and Lentosärkkä
subfacies. To the SE however, progressively lower horizons
appear to be exposed in the core of a major recumbent fold,
where the Petäjikkökallio subfacies is recognized.

Petäjikkökallio subfacies

More than in the preceding subfacies, graded meta-arkose
beds are prevalent with thickness varying from 20 cm to 1 m
and sporadic thin metapelite. In general the meta-arkoses
do not contain pebbly clasts: where any detritus coarser than
granules occurs it is usually significantly so (more than
20 cm), in a discrete chaotically sorted deposit (Figure 31).

As in the Pitkänurmi subfacies metapsephite beds may
contain boulders exceeding 1 m in diameter that are seen to
overlie finer-grained deposits, rather than rest unconformably
(as a basal facies) on markedly older basement. This is
clearly seen at locality 95.2 95.2 (Figure 31a). Large granitoid

clasts are relatively fewer than in the Pitkänurmi and
Karvanurmi subfacies, with orthoquartzite fragments being
diagnostic. These are generally well-rounded and composed
of homogeneous greenish-grey medium-grained orthoquartzite
(Figure 31a, b). Within the clasts sub-angular and coarser
quartz detritus, up to 1 cm in size, and some feldspar
detritus may also be present. Grains of an opalescent blue
quartz, though volumetrically trivial, are especially
conspicuous on freshly broken surfaces. Occasionally such
granules are seen within orthoquartzitic clasts and in one
light-coloured pebble they rather resemble quartz
phenocrysts in a felsic volcanic rock.

Lentosärkkä subfacies

In general this is indistinguishable from the finer
lithologies within the Karvanurmi and Pitkänurmi subfacies
and is likely to be transitional with the Salonkylä
lithofacies. It may be recognized as distinct from the latter
however, in that it contains rather thick beds of impure
meta-arkose or feldspathic metagraywacke while lacking in
extra-basinal detritus coarser than 1 cm. These coarser and
thicker beds may attain 5 m in thickness and contain
intrabasinal pelitic metaclasts, as at 94.8 27.2).

At 94.6 27.3 is a small normal fault accompanied by
ductile deformation near but not precisely along the boundary
between two of these thick metagraywackes. The fault predates
the earliest identified cleavage, so may be a consequence of
slumping penecontemporaneous with sedimentation.

3.3.ii   Salonkylä lithofacies

Distribution and reference exposures:

The lithofacies appears to be transitional southwards
with the finer lithologies of the Kettämönniemi lithofacies
but the nature of the western boundary with the Niittylahden-
rantatie lithofacies is unclear.   Northwards along strike it
may be contiguous with the lithologically similar Suhmura
lithofacies.   Best exposures are at locality 93.5 29.2.

Depositional units and structures:

Quartzose metapsammites dominate and are usually massive,
from 5-50 cm thick, without detritus coarser than 1.mm
(Figure 32 e).   Laminated metapsammites comprise another
variant and one exposure showed discontinuous rafts of
metapsammite 30-40 cm long in a stratified finer-grained
matrix, all now quite deformed (Figure 32 f).   If any internal
depositional features existed elsewhere, they have been
obliterated during the ubiquitous and pervasive S_{2-3} cleavage
differentiation.

3.4   HAMMASLAHTI ASSEMBLAGE

3.4.i Kilpelänkangas lithofacies

Distribution and reference exposures:

This is exposed in the footwall of the Kettämö thrust
zone and further to the NE it may be conformable with the
Suhmura lithofacies.   It terminates along strike to the SE
against the Haapajärvi   fault zone.   It is well displayed at
locality 94.6 28.7., where some Mulo lithofacies is interbedded.

<u>Depositional units and structures</u>:

Metapsammites from 10-100 cm thick alternate with
metapelite, the former tending to become coarser at higher
structural levels (which seem also to coincide with
depositional younging direction). The thicker beds may
contain 5-6 mm granules of quartz and feldspar detritus in
their lower parts, grading upwards to metapelite and recall
finer lithologies of the Kettämönniemi lithofacies. Since
the two lithofacies are separated by a thrust zone containing
mylonite, they are therefore considered separately here.

3.4.ii Suhmura lithofacies

<u>Distribution and reference exposures</u>:

This is present over a large part of the Niittylahti
1:20 000 sheet, from Suhmura in the S, to Reijola in the N.
It appears to be at least locally conformable with the Mulo
lithofacies, and may be equivalent stratigraphically to the
Salonkylä lithofacies, which it closely resembles.
Relationship to the metapelitic Kukkupää lithofacies has not
been ascertained, except that at Haapajärvi it appears to be
a faulted one. Representative exposures are at Niittylahti
92.4 32.6 and also localities 94.9 31.4 and 94.3 32.9 .

<u>Depositional units and structures</u>:

The lithofacies consists of uniform fine-grained
metapsammites and lesser metapelite, with no detritus observed
to be coarser than 1 mm. Consequently depositional features
such as grading are difficult to discern, except as transitions
from metapsammite to metapelite. Nevertheless, rhythmic
alternations of these, with metapsammite dominant and often
showing one sharp contact indicate depositional units with a
modal thickness from 10-30 cm and occasionally exceeding 1 m.

Where the alternating couplets are thinner, or within the
rare laminated metapelites, folds of small amplitude are very
prevalent, though such folding only rarely affects the disposition
of thicker beds. In proximity to the sulphide-rich Mulo
schist, as at locality 92.3 35.0, beds otherwise similar to the
Suhmura lithofacies are themselves quite ferruginous and weather
to a very rusty yellow-brown colour instead of the usual light
grey and ⸱ny lithological details are obscured.

3.4.iii   Mulo lithofacies
Distribution and reference exposures:

The observed outcrop distribution of this lithofacies
corresponds closely to that inferred from a positive anomaly
shown by reconnaisance geomagnetic data. It appears to occupy
or define a large fold hinge with the lithofacies disappearing
southwards along attenuated or thrust-out limbs. The magnetic
anomaly maps suggest however that it may continue further to
the NW than indicated on the facies diagram, under the lake
Pyhäselkä. It appears to be enveloped by rocks of the Suhmura
lithofacies and there is some evidence for the two having a
stratigraphically conformable transition. This is also possible
at Kilpelänkangas 10 km to the S where there is a restricted
occurrence of the Mulo lithofacies enclosed by metapsammites
and metapelites. The lithofacies is exceptionally prone to
weathering but informative reference exposures are found at
92.7 35.6 on the Joensuu-Lappeenranta highway and in a small
abandoned quarry at 91,3 36.5 .

Depositional units and structures:

When weathered, the dominant lithology is rusty and
disintegrates along a penetrative foliation and reticulate
jointing. Even when fresh it is difficult to identify any
relict depositional fabric because of the prevalence of

intergrown rosettes and brush-like aggregates of tremolite.
These are black, owing to graphite pigmentation and smooth
broken surfaces are vitreous, resembling anthracite or PVC
plastic. Occasionally primary lithological bands are observed,
defined by variations in sulphide content, these being the
only minerals other than tremolite that are obvious to the
unaided eye (Figure 33a,b).

At the quarry mentioned already some other lithological
variants are encountered: a calcareous metapelite (also
black) with scattered, randomly oriented prismatic tremolite
porphyroblasts, up to 3 cm long, appears to constitute a
discrete massive bed nearly 1 m thick. Underlying this are
laminated white and grey lithologies that are recognizable
as metadolomite (Figure 33 c) although it is unclear whether
the finest laminae could be depositional in origin. Non-
oriented tremolite rosettes clearly overgrow any major layer
parallel schistosity and may themselves be pseudomorphed by
invading sparry calcite veins which in this lithology are
quite abundant.

3.4.iv Kukkupää lithofacies

Distribution and reference exposures:

This is most extensive in the N of the study area whereas
S of Hammaslahti it may be faulted out or otherwise restricted
in development. A faulted boundary along Haapajärvi is invoked
to explain its juxtaposition with coarse clastic deposits of
the Kettämönniemi lithofacies but to the E a conformable
transition to the Tikkala lithofacies is indicated. Represent-
ative exposures are E of Haapajärvi on the Joensuu-Lappeenranta
highway (95,7 33.2)and 94,8 37.3 at Mulonsalo.

<u>Depositional units and structures</u>:

Three distinct lithologies are recognized, all metapelitic, lacking in detritus coarser than quartz silt:-

1) Dark grey micaceous metapelite, massive to fissile depending upon degree of cleavage development, displays no depositional features. Beds are tabular when boundaries are seen but thicker units may comprise an entire exposure. Fine-grained disseminated pyrite is frequently present, elongate in any schistosity developed.

2) Massive blue-grey siliceous metapelites seldom display any schistosity to the unaided eye and fracture conchoidally (compare lithologies in Figure 34f). They form sharply bounded tabular beds 5-25 cm thick and weathered surfaces often reveal fine planar laminae or elongate lenses, though neither of these appear to be concretionary or due to grain size variations. Overall, this lithology probably constitutes as little as 5% of the lithofacies.

3) Though typified by fine laminae several mm thick, this tends to be a massive rather than fissile lithology with a schistosity more weakly expressed than in interbedded micaceous metapelites. It consists of rhythmic alternations of calcareous metapelitic laminae 2-5 mm thick with thinner and darker graphite-rich laminae (Figure 34a, c). The calcareous laminae weather to a fawn-brown colour and also contain some detrital quartz silt, and abundant biotite porphyroblasts which mimic the grading direction that the detrital grains actually suggest. Low angle cross-stratification is surprisingly well preserved in these laminae (Figure 34 a, c) but no unambiguous directions of palaeocurrent flow were obtained.

3.4.v Tikkala lithofacies
Distribution and reference exposures:

This is exposed in a narrow meridional zone about 25 km
in length; its continuation to the N and S is conjectural
due to burial beneath unconsolidated glacial and fluvio-deltaic
deposits. It may have a coformable relationship with the
Kukkupää lithofacies in the N but the eastern boundary could
be controlled by high-angle faults. Three separate areas
were found to be particularly informative - Kumpu in the N,
the vicinity of Hammaslahti copper mine in the central region
and at Tikkala, near the southern end of the facies' exposed
distribution.

Depositional units and structures:
At Tikkala-

Exposures on either side of the Joensuu-Helsinki railway
at Tikkala siding (localities 0 4.6 17.4, 04.7 17.4 , 04.6 17.5)
have been very useful for observing both lithology and
deformation of the rocks.

Strata at 04.7 17.4 on the NE side of the railway siding
are deformed into a recumbent F_3 fold with an attenuated,
transposed lower-limb, subsequently reoriented into a vertical
attitude (Figure 10). The upper-limb, seen at locality 04.6 17.5
and illustrated in Figure 37c retains original sedimentary
features, apart from possible boudinage having affected the
thickness of metapsammites, and the further possibility of
layer-parallel dislocations with metapelites. The lower part
of the sequence is mostly laminated metapelite, constituting
layers usually thinner than 5 cm. Two lenticular or
boudinaged metapsammites are seen in the lower part and a
little higher are two very distinct planar beds identified as
partial Bouma sequence. They are of interest in being calcareous,

and also quite ferruginous, which is to be expected from the generally high sulphide content of adjacent metapelites.

The lithology then changes abruptly from laminites to massive, poorly sorted quartzose metapsammites. The lowest bed appears to be the thickest (up to 4 m) but may represent several amalgamated deposits, with discontinuous metapelitic layers denoting original depositional units. The beds cannot be regarded with confidence as being channeled because of the possibility of tectonic boudinage having modified depositional thickness. The thinner overlying metapsammites may indeed record depositional irregularities in that some beds thicken in mutually opposite directions but this too depends upon the homogeneity of strain.

On the southwestern side of the siding (04.6 17.4) cleavage and fold development is minimal and depositional features consistently show these beds to overlie those at 04.6 17.5 discussed above. The total section is illustrated in Figure 35. It begins with a 10 m sequence of metapsammites tending overall to become thinner and finer-grained upwards (Figure 36 a). Individual beds usually show grading from granule sized quartz and feldspar to finer quartz-dominated detritus, with thin metapelite partings (Figure 36 c,e). Other depositional features are developed in the uppermost beds: stratified metapsammite and occasionally cross-lamination in thin metapelites. These are features commonly associated with the Bouma sequence (Bouma and Nilsen, 1978) and though the complete sequence (T_{abcde}) is not recognized here, the massive graded beds, and stratified metapsammite may correspond to parts T_a and T_b respectively. The thin planar and cross-laminated metapelites can be equated with T_{cd} of the Bouma sequence.

Detritus discernible to the unaided eye include quartz and feldspar and occasionally small granitoid fragments.

Particularly distinctive are opalescent blue quartz grains
1-4 mm across, found also in the Petäjikkökallio subfacies.
Textural maturity inversions are typical, exemplified by rounded
quartz-grains in poorly sorted association with highly angular
quartz and feldspar detritus, and a high pelitic matrix
content.

In the laminated metapelites and thin metapsammites,
1-2 mm blobs of pyrite are common but are not clastic grains
in that they transgress sedimentary lamination. Larger subhedral
grains up to 2 cm are sporadically distributed through meta-
psammite (Figure 36 b). A high sulphide content is also evident
in the overlying sequence of laminated metapelites whose
weathered surfaces are particularly rusty, obscuring sedimentary
features and sometimes forming gossanous masses up to 1 m
across. A fault of probable small displacement interrupts the
succession but on both sides metapelite is overlain by
markedly coarser material (Figure 35, Figure 36 e, f). This
is at least 15 m thick and its basal layer, though generally
lacking in pebbly detritus contains one anomalously large
boulder, composed of stratified quartzose metapsephite
(Figure 35, Figure 36 f). Somewhat surprisingly there is very
little distortion of underlying strata and although boundaries
are partly obscured by rusty weathering, three successive beds
may be traced around it, thicker and more steeply inclined to
the left of the boulder. To its right, the second of these
beds shows an internal stratification resembling foresets,
although they are quite steep.

The overlying metapsammites are poorly sorted, yet well
stratified as discontinuous layers (Figure 36 e). Occasionally
pockets of larger clasts are present. These are mostly
intrabasinal but also include cobbles of vein quartz and one
rounded 5 cm gabbroid clast was noted. Clast diversity increases

upwards through the sequence, although granitoid detritus
is notably scarce. Clasts of stratified quartzite or other
metapsammites are most abundant,typically well-rounded and
10-40 cm in maximum dimension. Most interesting is the
occurrence in one horizon of abundant pink dolomite metaclasts
and rafts, up to 1 m in length. Some of these are quite
weathered but a few have primary lamination preserved
(Figure 36g). No discrete metadolomite layers have been
recognized within the Tikkala lithofacies though Nykänen
(1971a) records their presence southwards towards Tohmajärvi.

The Tikkala lithofacies at Kumpu

Interbedded with metapelites ascribed to the Kukkupää
lithofacies are quartzose metapsammites that exhibit grading
from granule- to pelitic-sized detritus. Bed thickness may
attain 1 m and granules are almost exclusively quartz. This
is well displayed at 98.2 35.8. The metapsammites become more
prevalent eastwards, where a further lithology is apparent -
a poorly sorted but stratified metapsephite, with elongate
and irregular intrabasinal metaclasts, lithologically identical
to associated coherent metapsammitic and metapelitic layers
(Figure 38 a, b). No granitoid or extrabasinal detritus was
positively identified in the field. Where clasts appear very
elongate, as above the hand lense in Figure 38a, it is difficult
to discriminate between tectonic deformation and syndepositional
processes.

The Tikkala lithofacies near Hammaslahti copper mine

In addition to the ore bodies themselves, two other
lithologies are developed in the vicinity of the mine, inter-
bedded with or in faulted juxtaposition with the more modal
variant of the lithofacies. To this category belong at least
two horizons of graphite metapelite, each less than 3 m thick
where exposed in the abandoned main open pit. These appear to

have been preferred horizons for fault movement but magnitude
and sense of displacement has not been established. Metapsammites
representative of the Tikkala lithofacies are now found on
either side of these graphitic schists; neither repetition of
strata nor mutual depositional contacts have been demonstrated.
Another lithology was well exposed during 1983 at a new open
pit (Z-Avolouhos) and was clearly seen to be interbedded with
the quartzose metapsammites of the Tikkala lithofacies.

This consisted of sharply bounded layers from 5-25 cm in
thickness but individual beds are difficult to trace laterally
more than 20 m, partly due to deformation. They are distinct
from the metapsammites in containing no trace of clastic detritus
at outcrop scale but instead are coloured dull green by the
abundance of rosettes and prismatic clusters of tremolite
porphyroblasts. An origin syndepositional with the metapsammites
is indicated however by their occasional presence as clasts in
coarser lithologies.

The associated metapsammites comprise the bulk of the
exposed sequence and constitute beds attaining 1 m in thickness,
the thicker ones clearly showing grading with quartz and
feldspar granules about 1 cm in size at the base (Figure 37 c).
Intrabasinal clasts may be larger, up to 20 cm and do not differ
lithologically from the enclosing metasediment. Thick metapelitic
units are not common, but thin alternations with graded meta-
psammite frequently occur, (Figure 37 a). Tectonic slip between
beds appears to be commonplace but it is possible that some
deformation was synsedimentary. Some peculiar small-scale
features are illustrated in Figure 37 b which may be examples
of the latter. They differ from typical basal lload casts,
showing thin (less than 5 mm) dykes of metapelite amongst
exceptionally well and finely stratified metapsammite.

About 1 km E of the above exposure, at locality 02.4 28.2, there occurs a 4 m thick polymictic metapsephite, overlying finer metapsammites and metapelites. Intrabasinal metaclasts are common, and usually very elongate, which can be ascribed to both initial tabular form and subsequent tectonic strain. More equi-dimensional granitoid,vein quartz and orthoquartzite meta- clasts are also abundant, none exceeding 20 cm in size. One apparently intrabasinal metaclast nearly 1 m long contains granules of the distinctive blue quartz, as at Tikkala and Petäjikkökallio.

A wide region to the W and N of Tikkala contains very few bedrock expsoures. Those present near Kostamo frequently display graded metapsammites that lack detritus coarser than 1 mm, except for some intrabasinal clasts. These resemble the finer lithologies of the Tikkala lithofacies but also recall the Suhmura lithofacies metapsammites and also in places the metapelitic Kukkupää lithofacies.

3.4.vi Rauansalo lithofacies
<u>Distribution and reference exposures</u>:

Sporadic outcrops of this define a low strike ridge E of the Joensuu-Lappeenranta highway; immediately to the W occurs the Tikkala lithofacies, while the eastern limit of the Rauansalo lithofacies has yet to be defined. Southwards along strike it appears to correspond to similar lithologies described by Nykänen (1968, 1971a) and to the N it may gradually merge with the Kalliojärvi lithofacies. A useful reference exposure was made at 04.8 22.5 during roadworks in 1982, and is depicted in Figure 39 c.

<u>Depositional units and structures</u>:

A sequence of graded, mostly massive, mafic metapsammites yield consistent depositional younging directions. Other

depositional features are generally lacking except in the
finer upper part of one bed in which ripple-cross lamination
was noted. Some thin laminated metapelite is present, but
the dark green metapsammites dominate, with beds between 10
and 80 cm thick. An intraformational, penecontemporaneous
dislocation, rather than later tectonic fault appears to be
recorded by an angular discordance in the upper part of the
sequence. In several beds below this discordance, there is
evidence of brecciation with angular to sub-rounded clasts of
metapsammite in a veinous, light-coloured matrix, also
containing brush-like aggregates of dark green amphibole and
a little malachite. The breccia does not constitute a discrete
depositional unit; neither does it resemble the usual results
of tectonic deformation (contrast Figures 39a and b)..

At locality 05.121.5, 2 km to the S, there is a higher
proportion of laminated metapelites and metapsammite layers
tend to be thinner than 10 cm. There are frequent alternations
between the two lithologies, a buff or light grey metapsammite
having recognizable clastic quartz grains and more mafic
layers, usually graded and much recrystallized; evidence for
detrital origin is better preserved in their finer, upper
portions. A few beds are particularly rich in feldspar that
seems to be detrital, and some metapelites are disrupted by
possible synsedimentary microfaulting.

3.4.vii Kalliojärvi lithofacies
Distribution and reference exposures:

This is developed to the E of the Tikkala lithofacies
but their mutual stratigraphical relations have not been
established. Southwards the lithofacies may be transitional
into the Rauansalo lithofacies, and to the E, towards the
area studied by Pekkarinen (1979a) similar lithologies were
noted but not mapped. Representative exposures are found in

road cuttings near the lake Kalliojärvi, in particular at the localities 02.7 33.2, 03.6 32.5 and 03.5 32.6 .

Depositional units and structures:

Chlorite and muscovite impart a grey-green sheen to the abundant thin-bedded mica schists. Buff to pale green meta-psammites are also typical, and the lithofacies is character-ized by alternations of these two lithologies, usually 2-20 cm thick but with psammite: pelite ratio varying considerably between couplets (compare Figure 39d with e). A differentiated cleavage parallel to bedding often precludes discernment of sedimentary features, as in Figure 39e but size grading from metapsammite to metapelite can be preserved, along with sedimentary lamination in the latter. In general the lithofacies contains more clastic detritus than the Kukkupää lithofacies yet lacks the coarser and thicker deposits of the Tikkala lithofacies. In addition, the apparent petrological difference, reflected in the abundance of chlorite instead of biotite further justifies it being classified separately.

3.5   VARPASALO PEGMATITE SWARM

Felsic pegmatite dykes are the only synorogenic intrusive rocks present in the study area and even these are of restricted occurrence, on the island Varpasalo and the south-western part of Oravisalo (map 2). Although they have no relationship to any depositional lithofacies a brief description of their nature and origin is apposite.

Individual dykes are up to 40 m thick and are commonly foliated, although very coarse, massive,graphic intergrowths of quartz and feldspar are also typical (Figure 13a). Muscovite is less common but usually present; biotite, beryl and garnet are accessory minerals with the latter probably a result of dykes having intruded under conditions of regional amphibolite facies metamorphism (Figure 15f). Evidence of synkinematic emplacement with respect to S_3 is dimensional alignment of small quartz grains and foliate aggregates of muscovite, and intracrystalline deformation of plagioclase phenocrysts, or as in some cases porphyroclasts (Figure 13c,d).

The mineralogy of the pegmatites provide few clues as to their origin. The composition overall appears to be near that of granite eutectic; this and the lack of xenoliths preclude placing any constraint on source rock chemistry. The magma was probably quite anhydrous, with muscovite only crystallizing at a late stage in myrmekitic intergrowths with quartz, interstitial to quartz and feldspar phenocrysts (Figure 13e).

FIGURE 24:   ORAVISALO QUARRY AND KANKAALA LITHOFACIES

ORAVISALO QUARRY LITHOFACIES A

a   Irregularly distributed felsic granitoid clasts in a matrix of
     meta-arkose. Also visible are some dark clasts which could be
     of retrogressed amphibolite or metamorphosed mud clasts. Pencil
     is lying above part of an amphibolite clast exceeding 1 m in
     length. This, and the granitic clasts are very similar to lith-
     ologies in tectonically juxtaposed basement. (Oravisalo 75.8 15.9)

b   Pebbly meta-arkose lying on a better sorted layer lacking in
     coarse detritus. The large granitoid cobble (approximately
     12 cm) shows a marked deflection around its base though the
     underlying meta-arkose layer remains undisturbed. Whilst
     resembling distortion due to dropstones, this may be rather the
     result of tectonic deformation; this would seem very probable if
     the elongate dark clasts below the cobble were of a pre-existing
     amphibolite and not ductile mud chips at deposition.
     (Oravisalo 75.8 15.9)

ORAVISALO QUARRY LITHOFACIES B

c   Granitoid clasts elongate parallel to principal tectonic lineation (
     in metapsephite unconformably overlying sheared, muscovite-rich
     basement gneisses and migmatites. (Oravisalo 75.8 15.7)

d-f Sections of metapsephite from above the unconformity, cut perpend-
     icular to the major schistosity. Section 'a' is also perpendicular
     to major tectonic lineation while sections 'b' and 'c' are parallel
     to this direction. Note the presence of a granular quartz-feldspar
     mosaic within all clasts and the variation in roundness and shape
     of clasts. Approximately natural size. (Oravisalo 75.8 15.7)

KANKAALA LITHOFACIES

g,h Intensely foliated impure metaquartzites, or quartzose metapsammites
     Note particularly in 'h' the darker finer grained part richer in
     muscovite and well preserved clastic grain outlines.
     Approximately natural size. (Oravisalo 75.3 15.2)
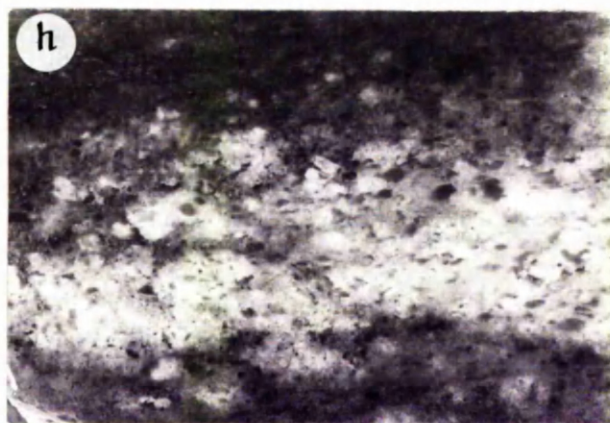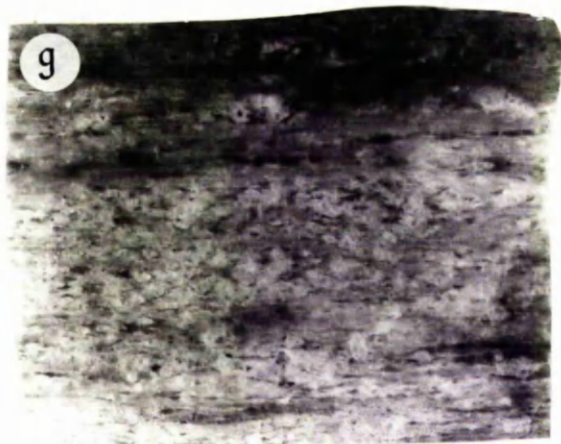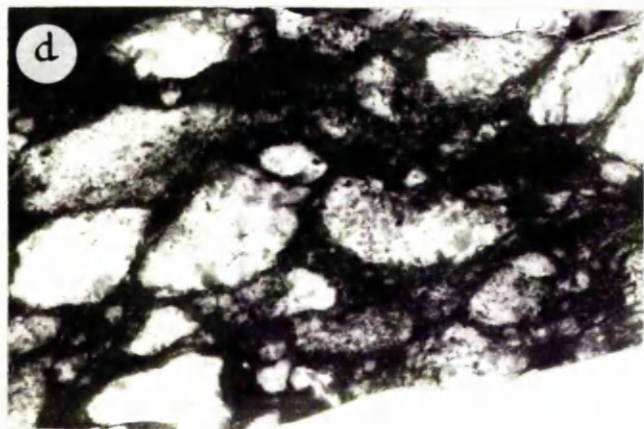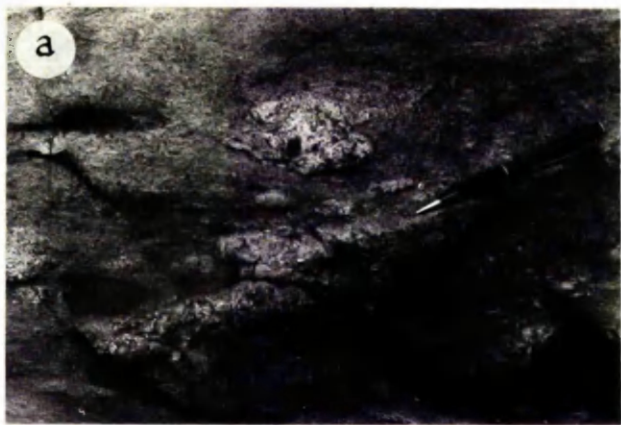
FIGURE 25 : RÄÄKKYLÄ LITHOFACIES

a S_3 is obvious in upper part of two metapsammite layers, showing incipient transposition of basal bed irregularities. (Varpasalo 67.8 20.4).

b   Composite S_2 - S_3 fabric has transposed load casts, preserved at base of a massive metapsammite layer. The lenticular biotite aggregates in the underlying lighter coloured horizon are characteristic of these metasediments. (Oravisalo 70.6 19.4).

c-g   Exceptionally well preserved load casts and flame structures at the base of a massive metapsammite and examples of planar and ripple laminations in the three subjacent layers. Figure f in particular shows ripples indicating that bed forms prograded from right to left and a superimposed differentiation layering expressed within the darker bands. (Pyhäselkä 94.1 20.2).
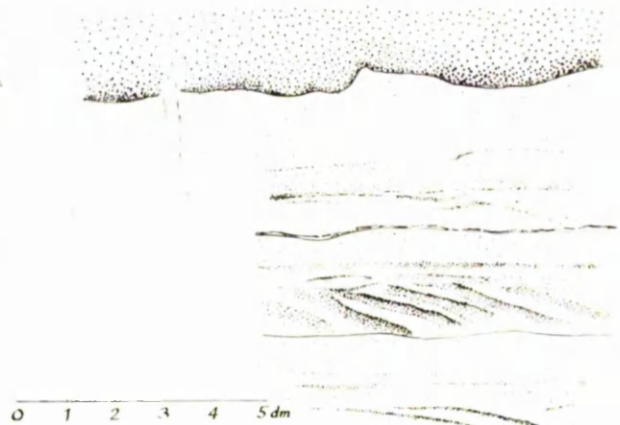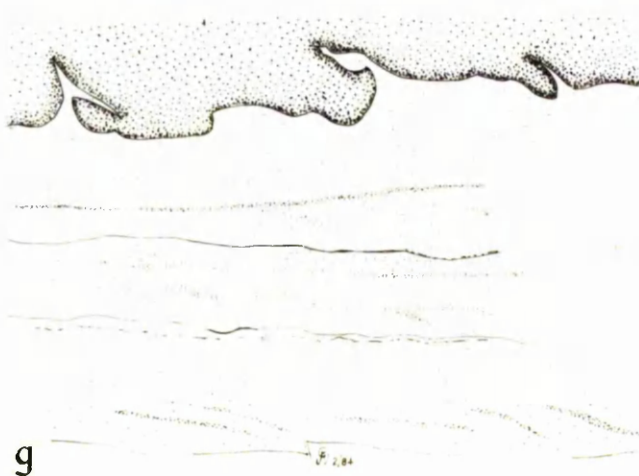
FIGURE 26 : RÄÄKKYLÄ AND SAMMALLAHTI LITHOFACIES

RÄÄKKYLÄ LITHOFACIES

a   Calc-silicate pods up to 12 cm in diameter exposed within
massive Rääkkylä lithofacies metasediment, unusually dark
due to it being a freshly blasted exposure (Varkaus road
near Viinijärvi). More commonly such pods appear lenticular,
confined within or defining lithological horizons.

b   Metapsammites completely lacking in evidence of depositional
features. The outcrop surface here intersects irregular,
anastomozing S_2 biotite foliae, hence the dark streaks giving
the appearance of a deformed xenolithic plutonic rock.
(Rääkkylä 80.8 12.0).

c   Planar beds 20-30 cm thick within a more thickly-bedded
sequence with grading and internal bed forms being either
absent or obliterated during metamorphic recrystallization.
(Varpasalo 67.9 19.4).

d   Example of the rarer, thinly bedded variant of the lithofacies.
(Oravisalo 74.3 17.6).

SAMMALLAHTI LITHOFACIES

e   Biotite selvedge developed at boundary of bed resembling
typical Rääkkylä lithofacies (at left) and metapsammite
bearing quartzose granules diagnostic of this lithofacies.
Compass face is 6.4 cm across. (Pyhäselkä 91.7 20.0).

f   Further example of Sammallahti lithofacies metapsammite rich
in granules of quartzose detritus having a poorly defined transition
to finer-grained metapsammite. (Pyhäselkä 91.7 20.0).
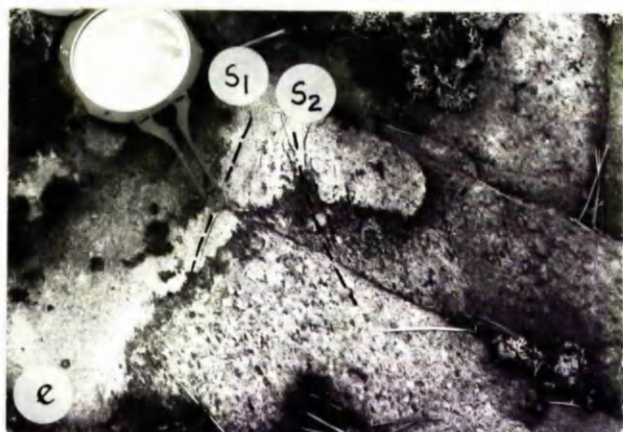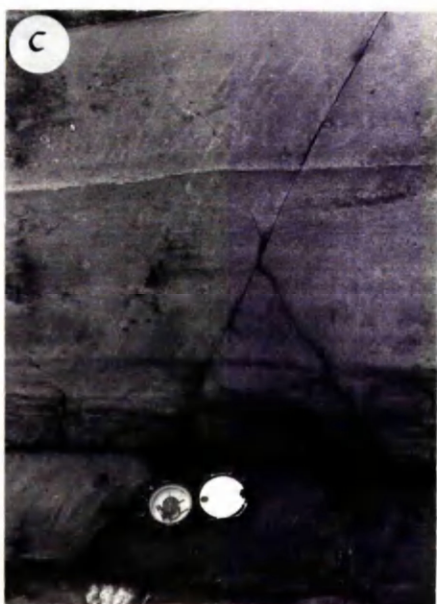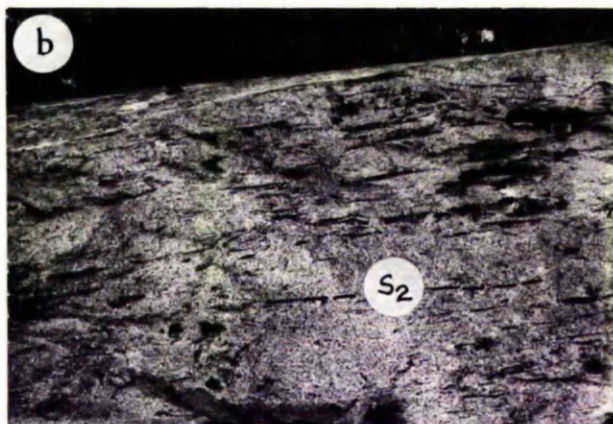
FIGURE 27 : NIITTYLAHDENRANTATIE LITHOFACIES

a   Large rafts of metapsammite apparently show ductile
deformation within diamictite. (Niittylahti 91.2 32.4).

b   Tectonic distortion S_2 of tabular intrabasinal clasts; white
quartz at bottom right is a discontinuous vein. (Niittylahti 92.2
32.4).

c   Between two massive metapsammite beds occurs this well
stratified metapsephite with an abundance of highly elongate
intrabasinal metaclasts. (Niittylahti 91.2 32.4).

d   Rare example of channelled scour in a 40 cm thick metapsammite
that also preserves internally planar stratification.
Calcareous pods are rich in green amphibole.
(Niittylahti 91.2 32.4).

e   Sketch of outcrop section showing massive metapsammites of
variable thickness and interbedded dark metapelite. Beds II
through VII could be interpreted as a thickening and
coarsening upward sequence culminating in the intra-basinal
diamictite of bed VII. Note that this does not display
chaotic slumping but is weakly stratified, and clast-matrix
boundaries are commonly poorly defined suggesting they were
poorly consolidated and partially disaggregated during
deposition. Lithologically they are indistinguishable from
the underlying metapsammites. (Niittylahti 91.3 32.2).

f   Photograph of outcrop sketched in 'e'; beds young from top
left towards bottom right and Brunton compass (total length
20 cm) rests on large metapsammite block in bed VII.
(Niittylahti 91.3 32.2).

g   Photograph of the more thinly bedded metapelite-metapsammite
alternations in the outcrop sketched in figure e. Note the
calcareous pods confined to the metapsammitic layers.
(Niittylahti 91.3 32.2).

a

b

c

d

e

f

g
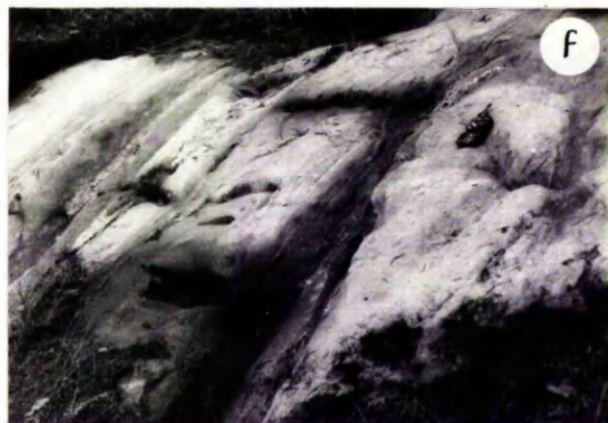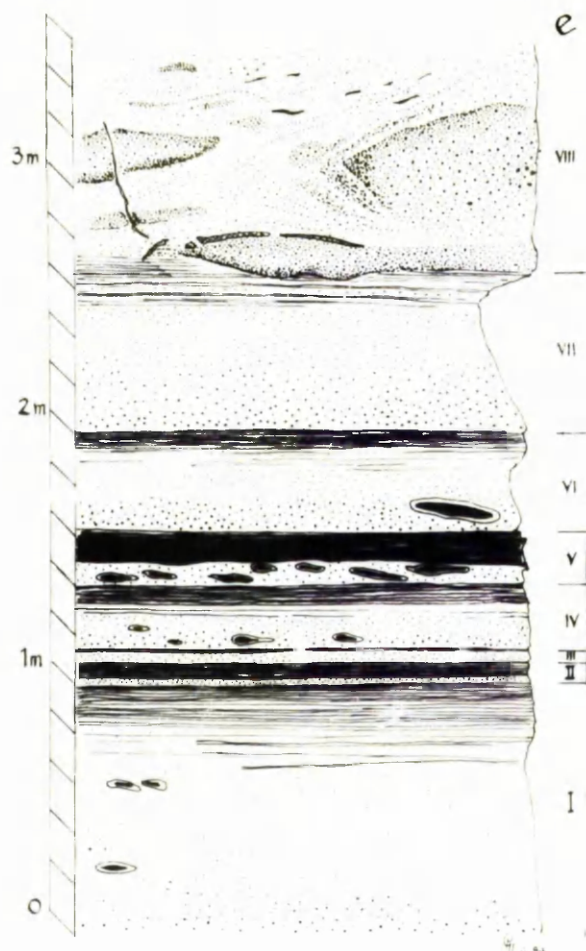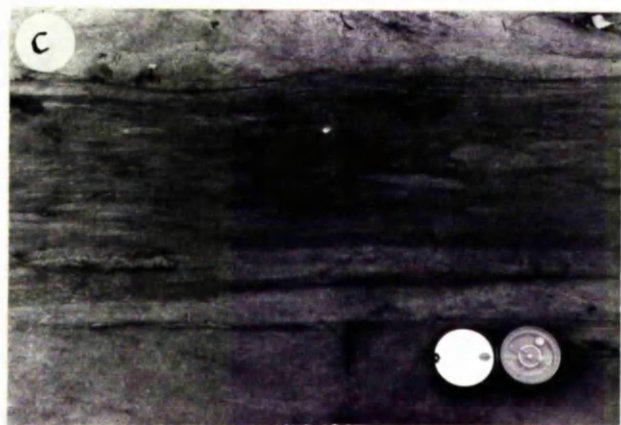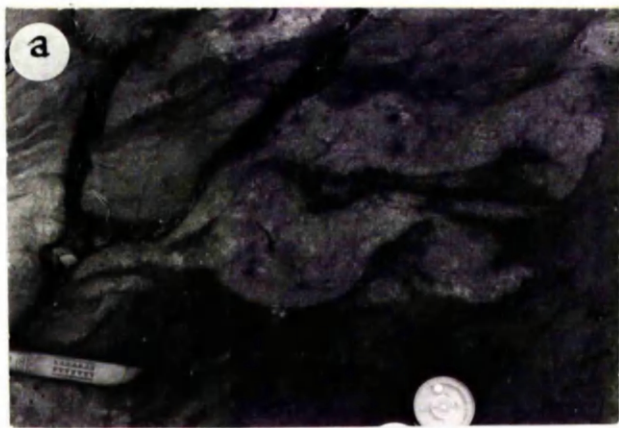
FIGURE 28 : NIVA AND NIINIKKOSAARI LITHOFACIES

NIVA LITHOFACIES

a,b Several different, spatially associated lithologies. Compass
     rests on massive grey metapelite with several ovoid calc-
     silicate pods at the contact with the adjacent laminated,
     but not fissile siliceous metapelites. This contact is also
     present at the base of the photograph where it appears more
     gradual. The third lithology is present as two thin massive
     bands (possibly folded repetition of a single layer) green
     in colour due to an abundance of actinolite. These may
     represent thin altered mafic tuffs or even sills rather than
     metamorphosed clastic deposits. (Pyhäselkä 93.9 24.6).

c,d The same strata several meters along strike. Apparent non-
     planar lamination may be tectonic in origin but note
     especially the massive metapelitic layer near the base of
     the photograph. The small nodules weathered in relief do
     not represent coarser detritus at the base of a graded bed
     but are diffuse actinolite or chloritoid porphyroblasts
     presumably reflecting a compositional difference across the
     bed. (Pyhäselkä 93.9 24.6).

NIINIKKOSAARI LITHOFACIES

e   Niinikkosaari lithofacies metapsephites substantially modified
     by composite S_2- S_3 foliation. Clasts include vein quartz and
     granitoid fragments. (Leppälahti 69.6 29.5).

f   Poorly sorted diamictite with angular clasts of a granitoid and
     darker orthoquartzite. Compare lack of deformational fabric with
     intense foliation in 'e'. (Roukalahti 72.8 30.2).

Figure a (diagram):
biotite - actinolite metapelite
cherty   laminite   concretion
concretion
* * * * * sill? * * * * *
cherty   laminite   or   metatuff
* * * * * sill? * * * * *
cherty   laminite   or   metatuff
massive biotite - actinolite metapelite

Figure c (diagram):
S_2
S_0
* * * sill? * *

S_{2-3}
S_2
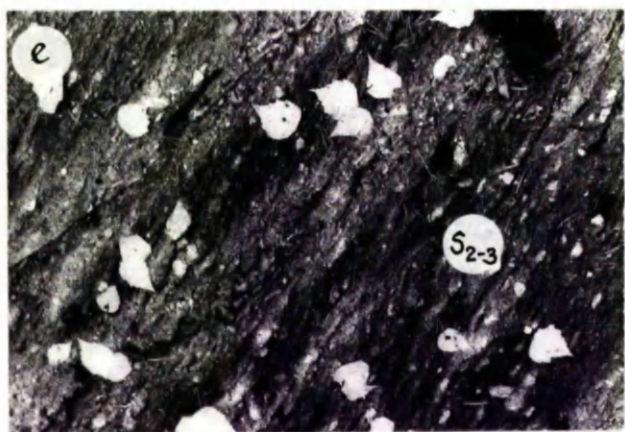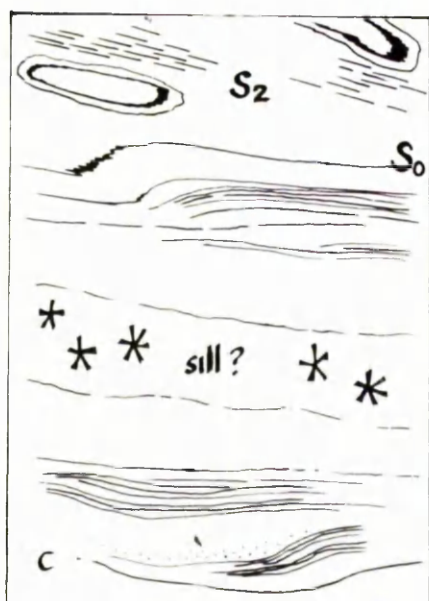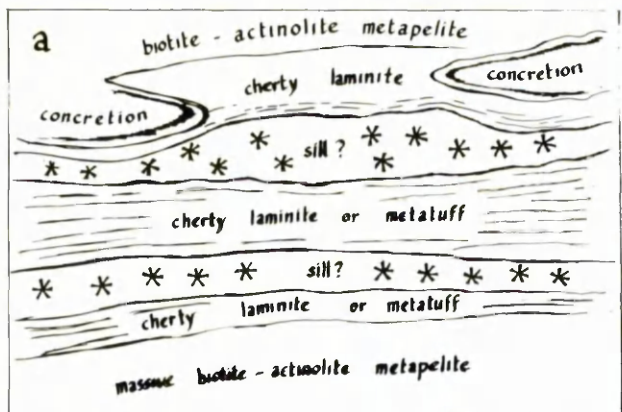
FIGURE 29 :   KARVANURMI AND PITKÄNURMI SUBFACIES

KARNAVURMI SUBFACIES

a   Typical lithological association in the upper part (presumed
     more distal aspect) of the subfacies: moderately well sorted
     meta-arkose lacking in both pebbly detritus and metapelitic matrix
     comprises the bed on the left while to the right is a diamictite
     with scattered intra-basinal metaclasts and somewhat finer
     granitic detritus with an abundant, unstratified metapelitic
     matrix. Compass gives scale. (Pyhäselkä 94.4 28.7).

b   As in 'a', a more homogeneous metapsammite interbedded with a
     poorly sorted lithology containing intra-basinal metaclasts up
     to 30 cm in diameter (Pyhäselkä 94.4 28.7).

c   Large boulder in metapsephite sequence containing predominantly
     intra-basinal detritus. The boulder itself displays internal
     stratification, with granitoid pebbles up to 5 cm in size. Much
     of the isolate granitic detritus in the metasephite in general
     may therefore be reworked from comminution of these pre-
     existing pebbly sediments rather than derive directly from
     an exposed granitic terrane. (Pyhäselkä 94.5 28.5).

d   Sketch of outcrop showing boulder in figure c. Unfortunately
     no depositional younging criteria were recognized but the
     coarsest detritus has clearly been transported with a depositional
     unit rather than forms a basal lag deposit. (Pyhäselkä 94.5 28.5).

e   Poorly defined, tectonically transposed intra-basinal clast;
     note coarser granules near base of clast, at bottom right of
     photograph. (Pyhäselkä 95,4 27.6).

PITKÄNURMI SUBFACIES

f   Stratified metapsephite layer with high matrix to clast ratio
     overlying fissile metapsammite or metapelite. Intra-basinal
     metaclasts prevail over granitoid detritus. (Pyhäselkä 94,4 28.3).

g   Intercalation of irregular, lenticular poorly sorted metapsephites
     with better stratified metapsammites and metapelites lacking
     in pebbly granitoid detritus. (Pyhäselkä 94.4 28.4).

a

b

c

d

e

S₃

So-2

dm
8
7
6
5
4
3
2
1
0

f

dm   granitic and arkosic clasts   intrabasinal clasts
10
9
8
7
6
5
4
3
2
1
0

g
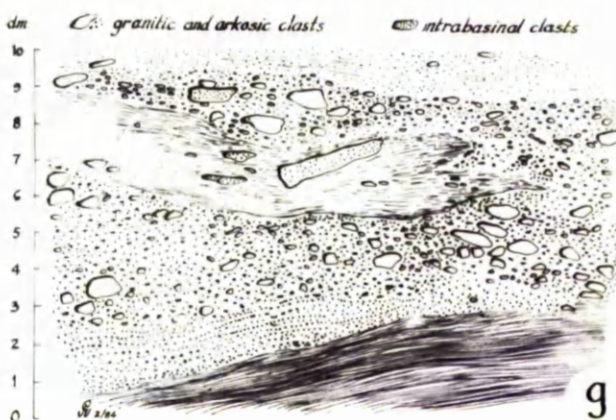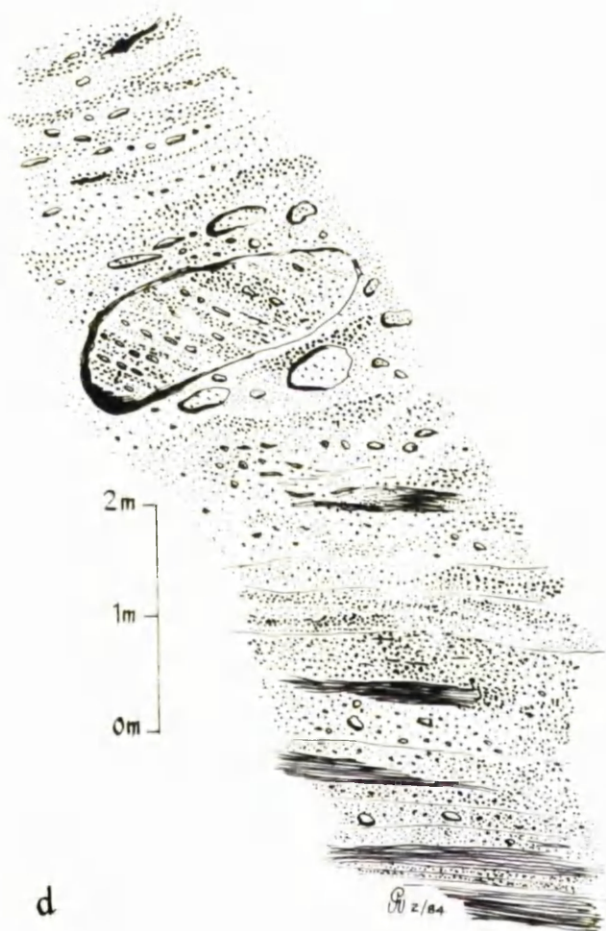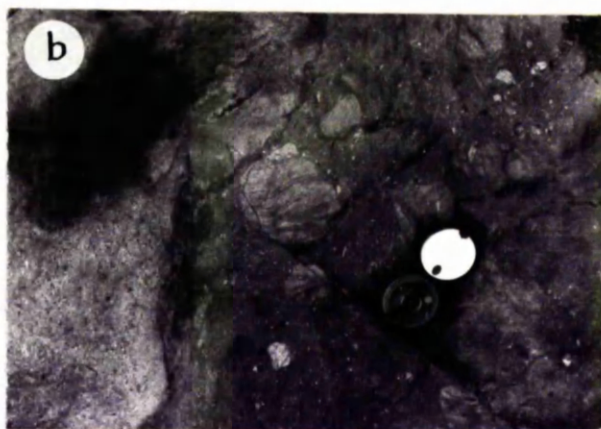
FIGURE 30 : PITKÄNURMI SUBFACIES

a   Angular clast of gneissic granitoid in coarse-grained meta-
arkose. Compass face is 6.4 cm across. (Pyhäselkä 95,5 27.2).

b   Very poorly sorted metapsephite with abundant granitoid
detritus. Note however that discontinuous stratified lenses
rich in metapelite are present. (Pyhäselkä 94.4 28.4).

c   Some of the largest granitic boulders encountered, exceeding
1 m in maximum dimension. They occur within a thick
diamictite horizon overlying the beds seen in figure 'b' and
consist of poorly stratified and sorted blocks of granitic
detritus. The large boulder to the left of the compass has
a fracture traversing it (now moss-filled) which may predate
lithification of the sediment. Attitude of these large
boulders is weakly imbricate, with a tentative suggestion
for a component of the palaeocurrent vector from bottom
left to top right of the photograph. (Pyhäselkä 94,4 28.5).

d   Less oblique view of same exposure, from out of pine tree.

e,f   Better-sorted and modally finer-grained than in the foregoing,
a sequence of graded meta-arkose and metapsammite exposed
in the lower limb of an F_3 recumbent fold. Note especially
the transposition of basal load casts within the axial plane
schistosity. Bed thicknesses are from 5-50 cm and display
only some features in common with the Bouma sequence.
(Pyhäselkä 95 .5 27.3).

5dm

S₃
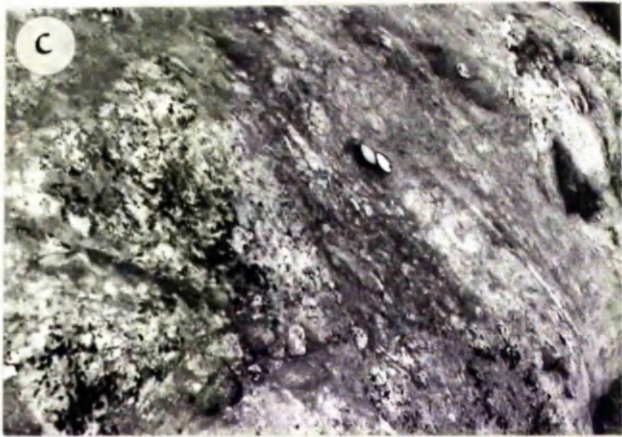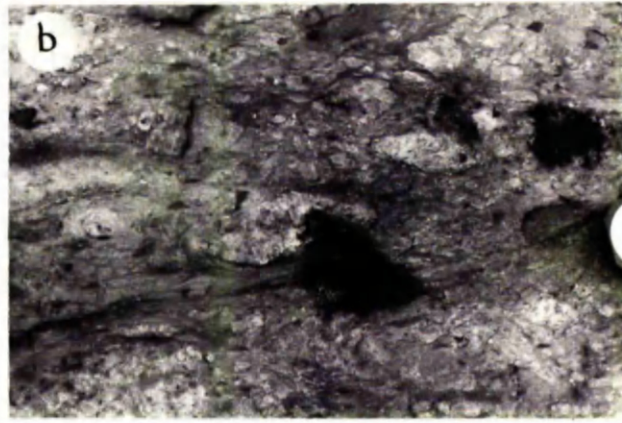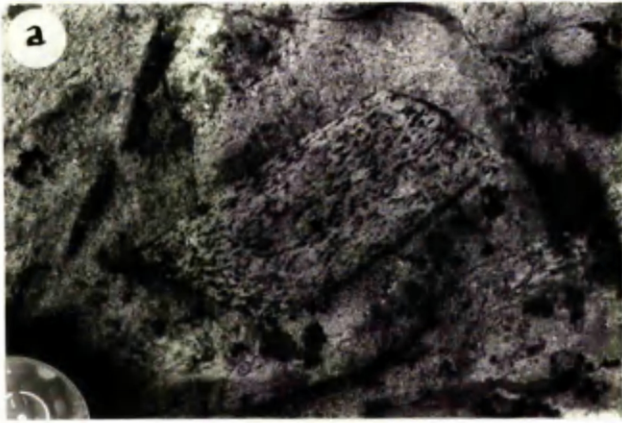
FIGURE 31 : PETÄJIKKÖKALLIO SUBFACIES

a   Sequence of graded impure meta-arkoses, bed thicknesses
      attaining as much as 5 m. Note that the bed containing
      the largest clasts (of greenish grey orthoquartzite) has
      been deposited on finer-grained beds. (Pyhäselkä
      95.2 27.4).

b   Impure meta-arkose with rounded orthoquartzite clast
      (diagnostic of the subfacies) and smaller pebble of vein
      quartz at bottom right of photograph. (Pyhäselkä 95.3 26.7).

c,d   Polymictic metapsephite with rather high matrix to clast
      ratio. In addition to quartzose metaclasts there are intra-
      basinal metapelitic fragments up to 20 cm in length and small
      granitoid pebbles. (Pyhäselkä 95.2 27.2).

b

c

d

area of
figure 'c'

orthoquartzite

metapelite

granitoid
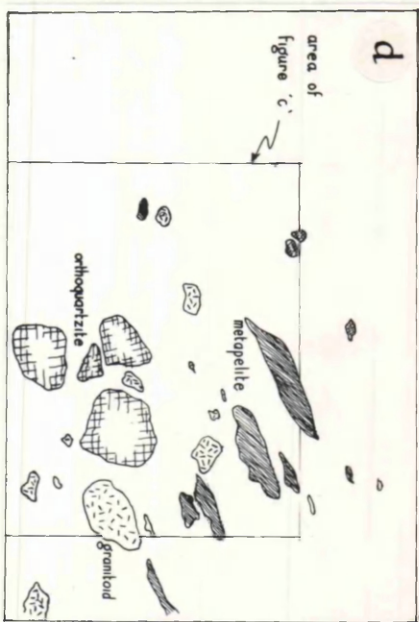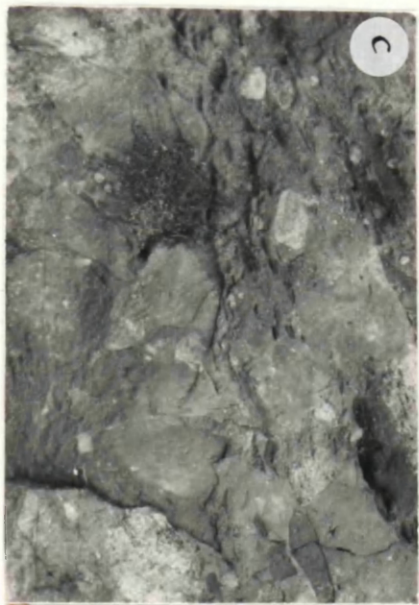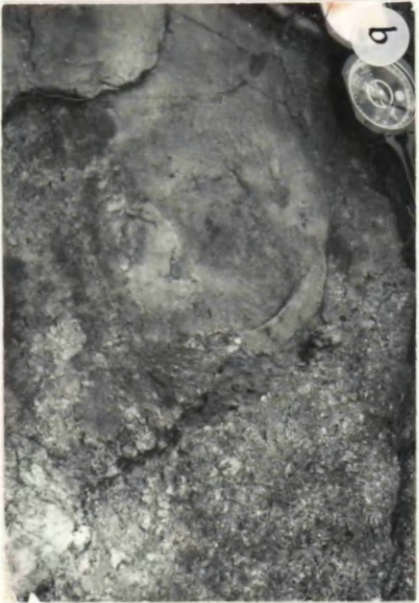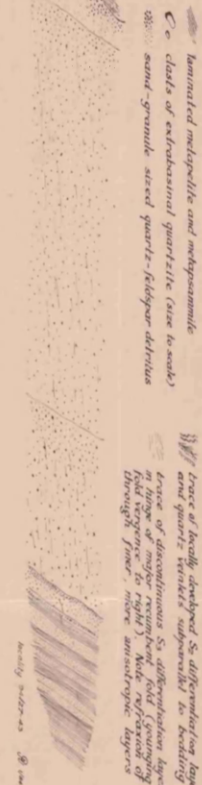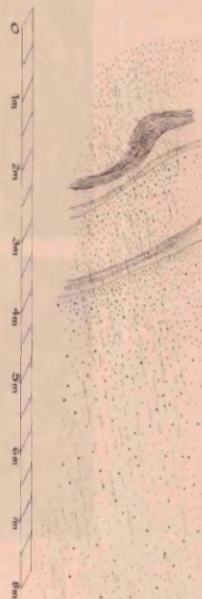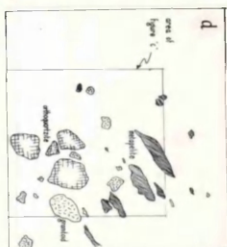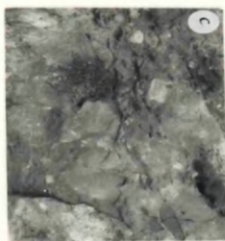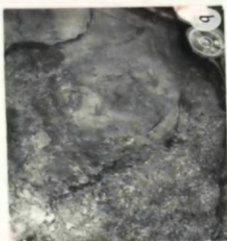
laminated metapelite and metapsammite

clasts of extrabasinal quartzite (size to scale)

sand-granule sized quartz-feldspar detritus

trace of locally developed S_2 differentiation layering and quartz veinlets subparallel to bedding

trace of discontinuous S_2 differentiation layering in hinge of major recumbent fold (younging and fold vergence to right). Note refraction of S_2 through finer, more anisotropic layers

FIGURE 32: LENTOSÄRKKÄ SUBFACIES AND SALONKYLÄ SUBFACIES

LENTOSÄRKKÄ SUBFACIES

a,b   The boundary between two graded feldspathic metapsammites
      can be traced from top left to bottom right of the photograph,
      with the bed to the right being the younger. This boundary
      corresponds closely to a small pre-tectonic ductile fault,
      recorded by drag in adjacent strata. (Pyhäselkä 94,6 27.3).

c   More detailed view of fault zone with rough spaced cleavages
      S_2 and S_3 cutting it. (Pyhäselkä 94.6 27.3).

d   Massive graded bed with unoriented and angular metapelitic
      clasts in basal portion. (Pyhäselkä 94.8 27.2).

SALONKYLÄ LITHOFACIES ·

e   Minor folds with associated crenulation cleavage (S_5) have
      deformed discontinuous thin metapsammite layers that may
      represent intrabasinal clasts. Host lithology is better
      stratified though also metapsammite. (Pyhäselkä 93.5 29.2).

f   Segregation or invasion of thin quartz veins imparts a
      laminated appearance to this massive quartzose metapsammite.
      (Pyhäselkä 93.5 29.2).
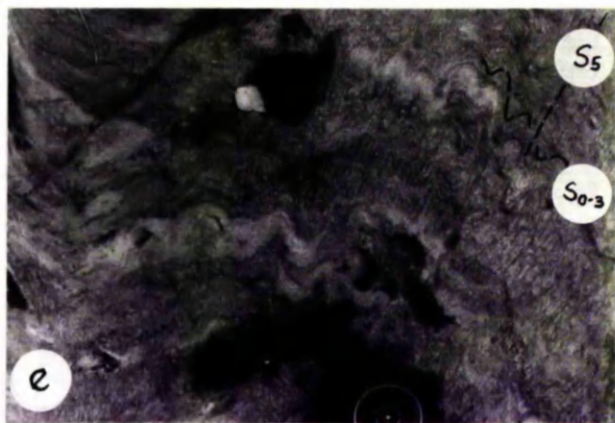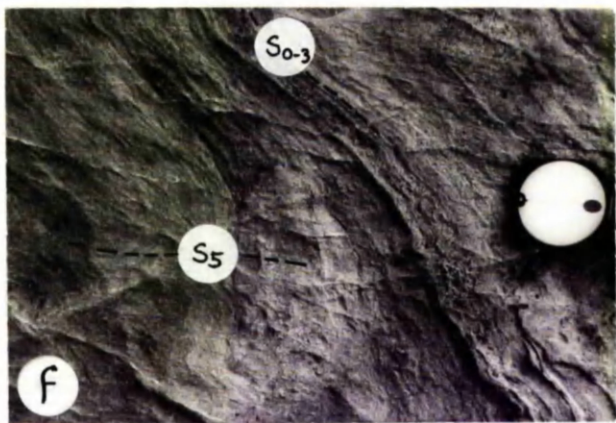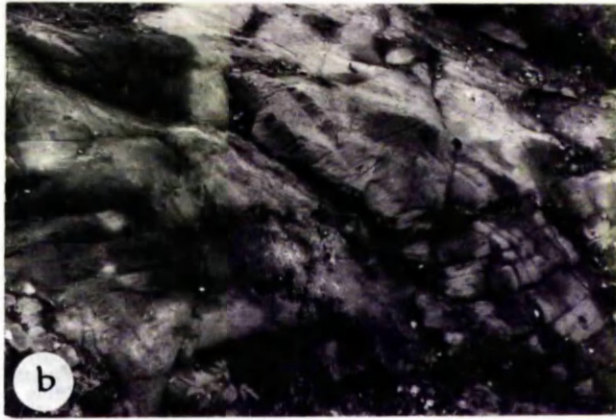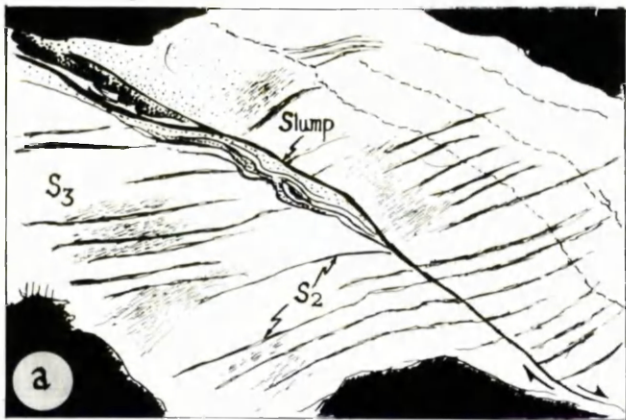
FIGURE 33: MULO LITHOFACIES AND HAMMASLAHTI ORE

MULO LITHOFACIES

a   Major tectonic lineation (L_3) is evident on this bedding-parallel
    cleavage plane, typically rich in pyrite. Approximately 2/3
    natural size. (Niittylahti 92.7 35.6).

b   Same specimen sectioned perpendicular to bedding-parallel
    schistosity. Lithological layering of a primary nature is
    evident only from the variation in sulphide content; the
    black colour is due to abundant graphite in a calcareous
    metapelite matrix, with abundant tremolite porphyroblasts.
    Small lenses of sulphide perpendicular to the lithological
    layering define the weaker S_4 cleavage. (Niittylahti
    92,7 35..6).

c   Metadolomitic lithology; dark bands are coloured by graphite
    but it is not certain whether the lighter laminae are of
    primary depositional origin. They do differ from the obvious
    white calcite vein that traverses the middle part of the
    specimen. Approximately natural size. (Niittylahti
    91.3 36.5).
    HAMMASLAHTI ORE

d   Clasts of brecciated metagraywacke in a matrix of mobilized
    chalcopyrite. Natural size. (Hammaslahti mine)

e   Tectonic brecciation of graphitic metapelite by pyrrhotite.
    (Hammaslahti mine)

f   Primary, pre-tectonic banding of pyrrhotite in metagraywacke;
    clustered aggregates are presumably metamorphic.
    (Hammaslahti mine)

g   Chalcopyrite-bearing layer (lighter) isoclinally folded
    (probably F_2) with quartzose metagraywacke. (Hammaslahti mine)
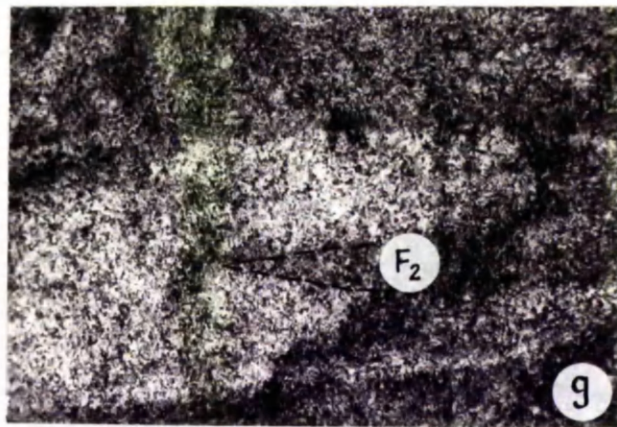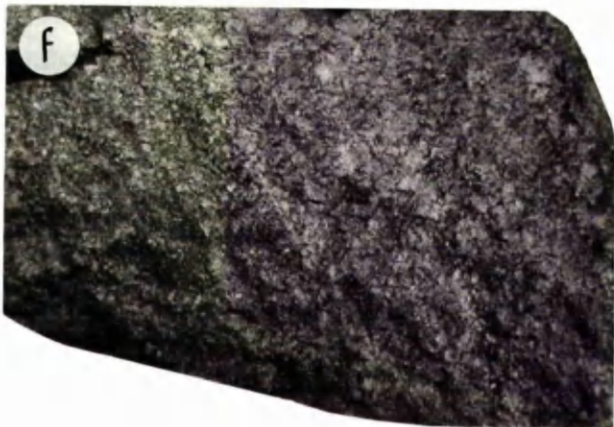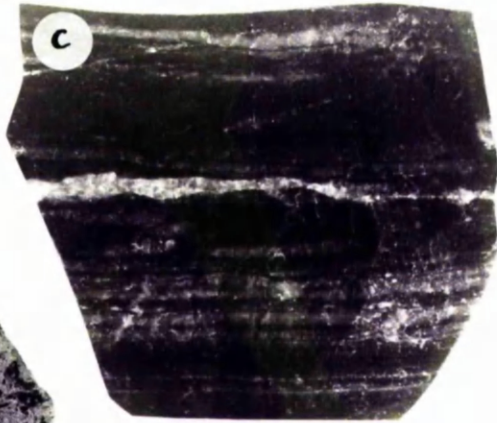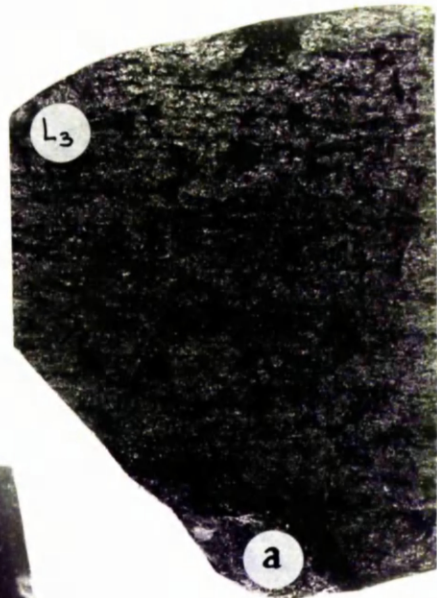
FIGURE 34 : KUKKUPÄÄ LITHOFACIES

a   Calcareous variant of the lithofacies with thin darker
     laminae richer in graphite. Small black specks are not
     detrital grains but biotite porphyroblasts. Approximately
     natural size. (Niittylahti 97.7.30.1).

b   Siliceous and calcareous laminated metapelites with an
     indication of cross lamination in the lighter middle horizon.
     Note also the truncating nature of the calcareous lenticular
     pod, affording evidence of its post-depositional, diagenetic
     origin. (Outokumpu Oy monument, Kivisalmi)

c   Thin (2-5 mm) rhythmic alternations of pale brown quartz-
     carbonate lithology with thin dark laminae richer in graphite
     and mica which could be interpreted as thin graded beds.
     Note especially the well preserved bed truncation and cross
     lamination delineated by the darker films. Approximately
     natural size. (Niittylahti 97.7 30.1).

d   Outcrop sketch showing distribution of massive grey siliceous
     metapelite amongst more common fissile micaceous metapelites,
     whose primary depositional features have been obliterated
     or transposed by pervasive S_3 crenulation cleavage.
     (Niittylahti 94.8 37.3).

e   Discontinuous siliceous-calcareous bands within metapelites
     traversed by S_5 crenulation cleavage. (Rauansalo 02.5 22.2).

f   Interbedded cleaved metapelites and massive siliceous layers,
     the latter showing less evidence of deformation fabric, and
     retaining primary depositional laminae. (Niittylahti 94.8 37.3).

1 dm

d

m
6
5
4
3
2
1
0

intrafolial F₂ fold
with axial planar S₂

S₃ well developed as crenulation cleavage
and differentiation layering in micaceous
metapelites

massive, sometimes concretionary metapelites

S₂ cleavage and quartz veins
parallel to lithological layering

S₅ crenulation cleavage

S₅

dm

f

S₃

S₀
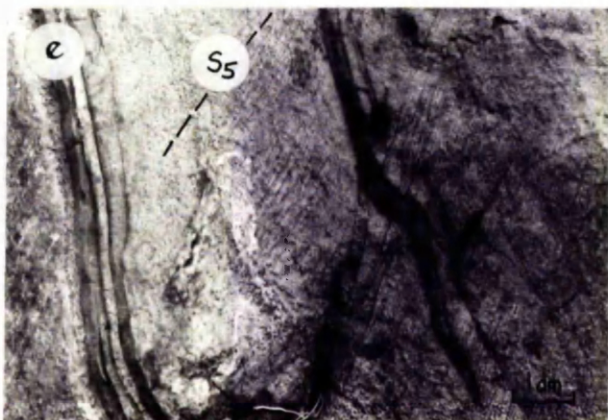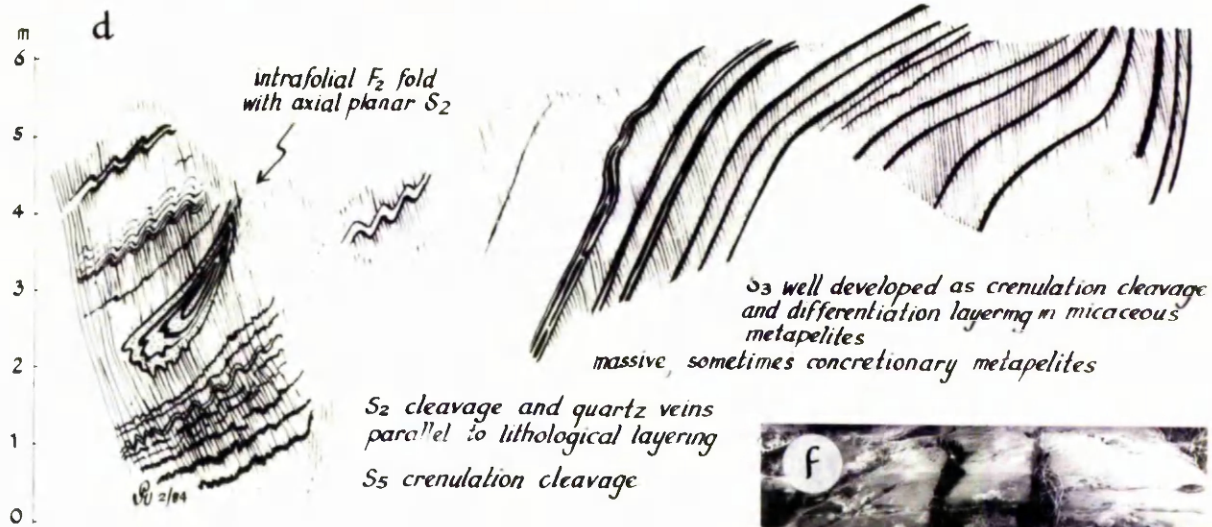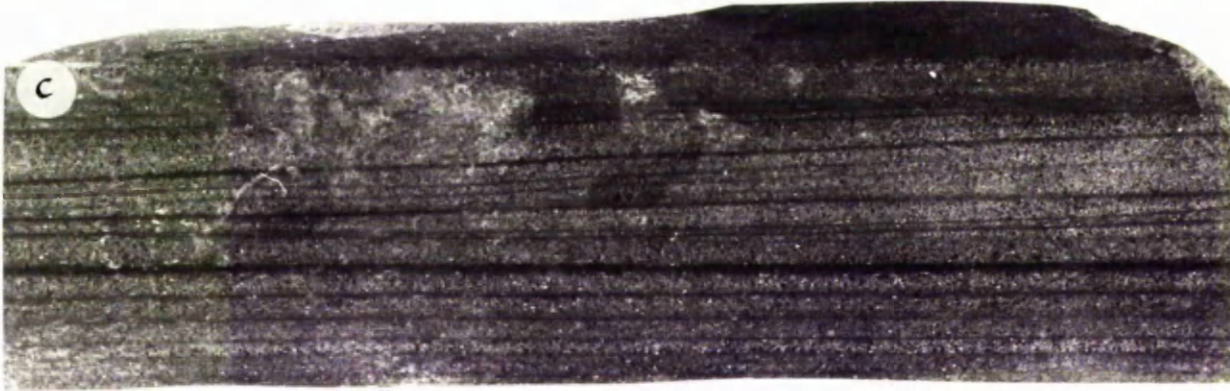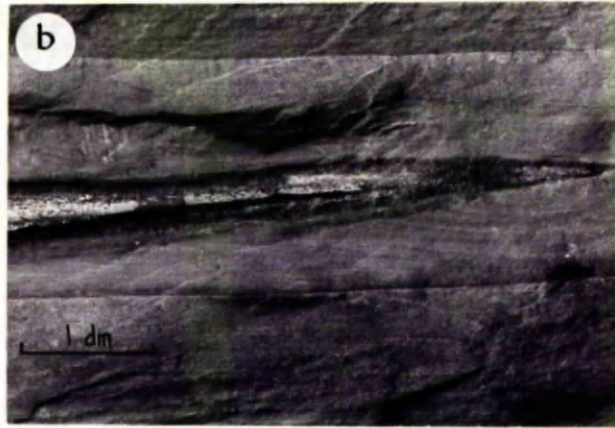
FIGURE 35 : TIKKALA LITHOFACIES AT TIKKALA-1

Sketch of sequence exposed on western side of Tikkala
railway siding. Features to note in particular are the
metapsammites at the base of the exposure. This 10 m sequence
may be described as a fining and thinning upwards cycle;
in the context of the submarine fan model (Mutti and Ricci
Lucchi, 1975; Walker, 1978) it can be interpreted as a mid-
fan depositional lobe progressively starved of coarse
sediment and consequently blanketed by the overlying sulphide-
rich metapelites. The upper part of the exposure shows a
renewed development of coarse clastic deposits, with
intrabasinal conglomerates containing a diversity of clast
types including pink dolomite of uncertain provenance. The
large boulder depicted in the inset diagram appears to
record a direction of palaeocurrent flow from left to
right. (Onkamo 04.6 17.4).

N

laminated, sulfide-bearing metapelite

gossanous features

massive metagraywacke with quartz-feldspar detritus up to 8mm

stratified, poorly sorted metagraywacke

~ pelitic clasts

graywacke clasts

quartzitic clasts

dolomitic clasts

granitoid-gabbroid clasts

Bouma sequence correlates

trace of S_3 crenulation cleavage

trace of S_3 cleavage
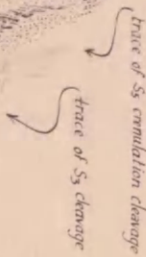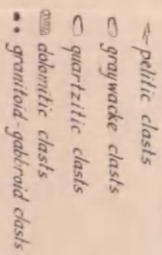
FIGURE 36 :   TIKKALA LITHOFACIES AT TIKKALA-2

a   Graded metapsammite beds interpreted as the upper part of a
     fining and thinning upwards sequence (left) overlain by
     sulphide-rich metapelites (right).   (Onkamo 04.6 17.4).

b   Sawn section showing grading in metapsammite interpreted as the
     T_a division of the Bouma sequence.   Note the preservation of a
     detrital fabric with angular white feldspar granules and the
     high proportion of matrix.   The protolith could be considered as
     a feldspathic graywacke.   Approximately $2/3$ natural size.
     (Onkamo 04.6 17.4).

c   Alternating lamination of micaceous metapelite and quartzose meta-
     psammite with very little recrystallization of grains.   Blobs of
     sulphide 1-2 mm across tend to be aligned within a later cleavage
     and transect depositional layering indicating that they are not
     detrital.   Note the deflection of metapelite laminae around the
     large pyrite grain. Approximately natural size. (Onkamo 04.6 17.4).

d   A rare, almost complete Bouma sequence with a thick massive graded
     basal part (T_a), planar-stratified upper part (T_b) and thin cross-
     and planar-laminated metapelite (T_{cd}) below succeeding T_a
     deposit.   (Onkamo 04.6 17.4).

e   Banded but poorly sorted metapsammites with sporadic pebbly
     detritus passing upwards (to top left) into layers with more abund-
     ant intrabasinal clasts.   (Onkamo 04.6 17.4).

f   Large isolated boulder of internally stratified quartzose meta-
     psammite enclosed within poorly stratified metapsammite.   Draping
     of several layers around it and possible foresets within one layer
     at right suggest palaeocurrent flow from left to right.   There is
     an absence of slumping or deflection of underlying strata.
     (Onkamo 04.6 17.4).

g   Sub-angular clast of pink metadolomite; this lithology is confined
     to one horizon within a thick intrabasinal metapsephite.   Primary
     lamination and irregular calcite veins or impurity selvedges are
     surprisingly well preserved.   (Onkamo 04.6 17.4).
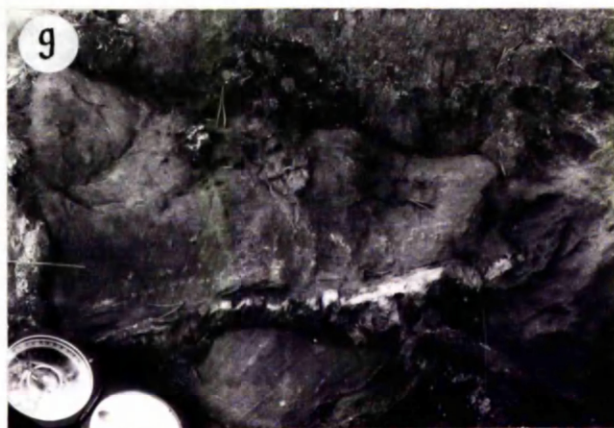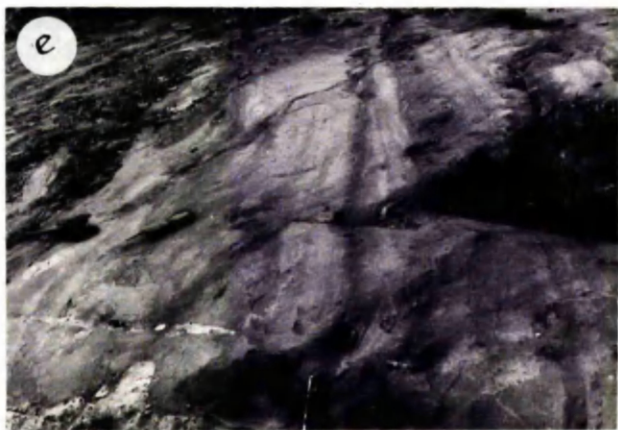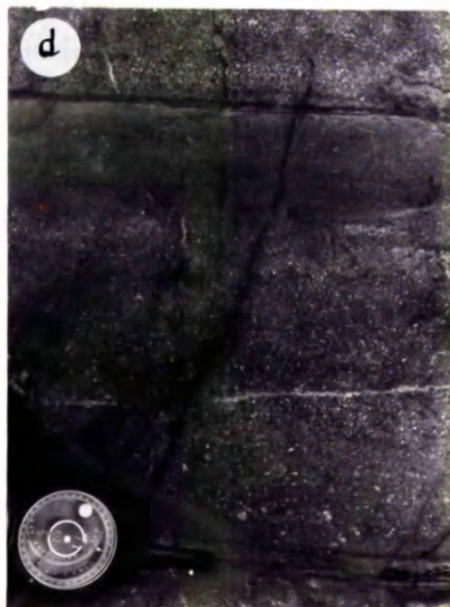
FIGURE 37 : TIKKALA LITHOFACIES AT HAMMASLAHTI MINE

a   At left, thinly bedded alternations of metapelite and
     metapsammite, disrupted to the right by a minor S_3 shear
     zone with associated vein quartz. Field of view approximately
     40 cm across. (Rauansalo 01.7 28.8).

b   Exceptionally finely stratified metapsammite overlying
     metapelite which appears to intrude as a small dyke, possibly
     a dewatering phenomenon. A weak cleavage (?S_5) appears to be
     parallel to this structure but it is likely to be fortuitous
     and later. Approximately natural size. (Rauansalo 01.7 28.8).

c   Interbedded graded metapsammites and metapelites exposed
     during 1983 at the Z-Avolouhos open pit, Hammaslahti copper
     mine. Note consistent direction of depositional younging
     to right. Variations in thickness and lenticular nature of
     some metapsammites is a consequence of tectonic deformation,
     though intrabasinal metapsephites with elongate clasts are
     associated. (Rauansalo 01.7 28.8).

d,e Deformed polymictic metapsephite; the most elongate clasts are
     intra-basinally derived metapelites which were doubtless
     already of low sphericity at time of deposition. Rounder
     cobbles are of othoquartzite or more commonly various kinds
     of massive to gneissose granitoid. Compass length is
     20 cm. (Rauansalo 02.4 28.2).

a S_3

b $?S_5$

c

1m

180°   270°

1m   1m

d

gneiss

vein
quartz

granitoid

graywacke

orthoquartzite

pelite

e S_{0-3}
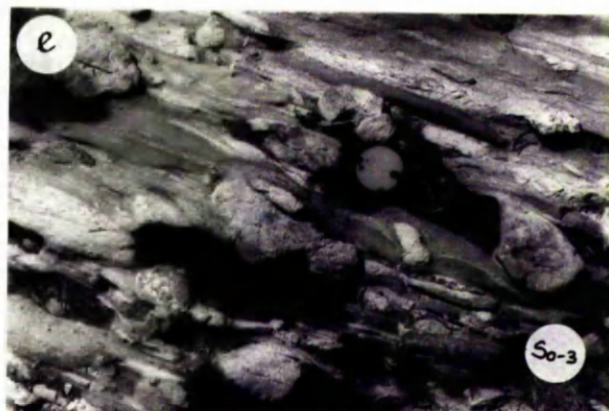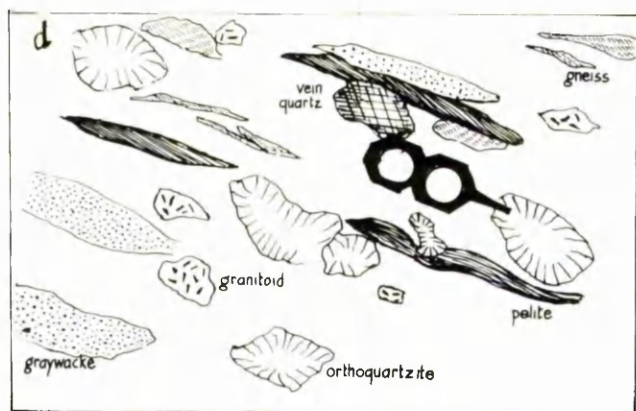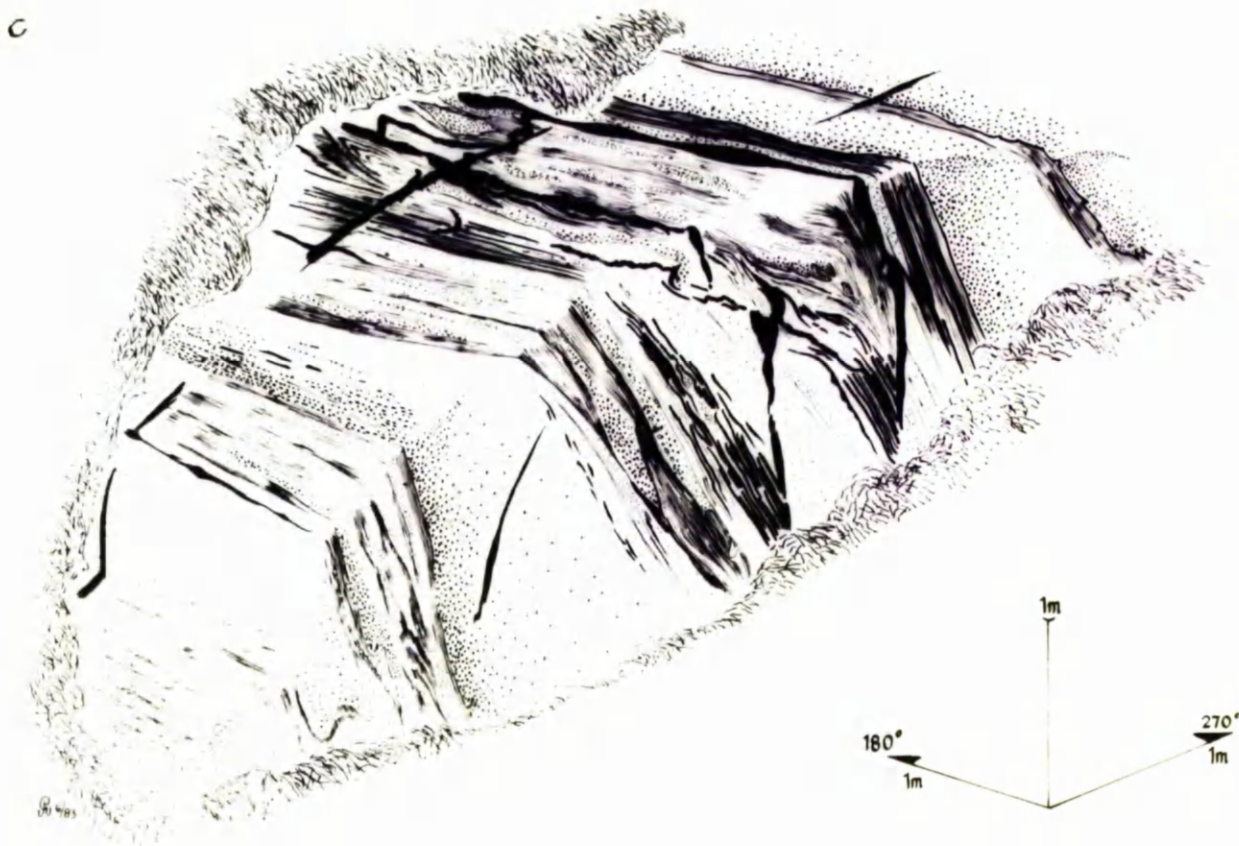
FIGURE 38 : TIKKALA LITHOFACIES

a,b   Poorly sorted metapsephite containing intrabasinal clasts
      identified as deriving from interbedded metapsammites and
      metapelites. Clast elongation probably results from a
      combination of initial shape, and both synsedimentary and
      tectonic deformation. (Niittylahti 98.9 36.5).

c   Metapelite sequence overlain by a sequence of massive quartzose
      metapsammites that could be interpreted as a fining and
      thinning upward sequence, stratigraphically underlying that in
      Figure 35. Apparent lenticular nature of the thickest bed
      in particular could be a result of tectonic boudinage,
      instead of depositional. Within the sulphide-rich metapelites
      occur two planar graded beds, as almost complete Bouma
      sequence, which are also unusually calcareous. (Onkamo
      04.6 17.5).

d   Matchbox lies on rounded boulder of Jatulian orthoquartzite;
      below this are mafic volcanic clasts, all in a matrix of
      degraded (subsequently metamorphosed) volcaniclastic material.
      Turbidite Conglomerate Formation of Pekkarinen (1979).
      (Viesimo 14.1 29.5).

e   Clasts of Jatulian metadolerite and orthoquartzite in
      Turbidite Conglomerate Formation of Pekkarinen (1979).
      (Viesimo 14.1 29.5).

a

S_{0-3}

b

c

1 2 3 4 5 m

000°

090°

5 m

b

b

b

a

e a

e

d

g

d

c
c

g   g

f

h

f

a   depositional base of massive quartz metagraywacke sequence

b   pelitic laminae, disrupted or coherent

c   graded calcareous metagraywacke

d   laminated sulfide-bearing metapelite

e   early tectonic displacement associated with S_3 or S_2

f   boudinaged metagraywacke

g S_5 crenulation cleavage and accompanying minor fold

h   discordant quartz vein perpendicular to S_5 and F_5

d

e

FIGURE 39 : RAUANSALO AND KALLIOJÄRVI LITHOFACIES

RAUANSALO LITHOFACIES

a   Non-depositional breccia that does not appear to relate to a
    specific fault zone, and which contains a matrix lithologically
    different from surrounding metasediments. (Rauansalo 04,8 22.5).

b   Tectonic discordance and quartz veining in graded mafic
    metapsammites at base of the section sketched in figure c.
    (Rauansalo 04,8 22,5).

c   Sketch section depicting consistent depositional younging
    direction indicated by grading from coarse metapsammite. Note
    also the intraformational discordance in the upper part,
    which may be causally related to the non-depositional breccia
    in the underlying beds. The disposition of beds at the base
    of the section however, is interpreted as resulting from Group 2 or 3
    deformation. (Rauansalo 04.8 22.5).

KALLIOJÄRVI LITHOFACIES

d   Thinly bedded metapsammite (light-coloured) and darker
    metapelite alternations showing disharmonic F_5 folds. Due to
    an abundance of chlorite and muscovite the rocks are greenish
    grey. Couplets are from 4-8 cm thick. (Heinävaara 02,7 33.2).

e   Same lithologies as in figure a but more thickly bedded with
    metapsammite dominant and displaying a well developed S_2
    differentiation layering, crenulated by S_5 and accompanying F_5
    folds. Compass is about 7 cm in size. (Kalliojärvi 03,6 32,5).

4.   INTERPRETATION OF LITHOFACIES AND PALAEOENVIRONMENTAL ANALYSIS

4.1   SOURCES OF INFORMATION

Valuable information may be obtained from a number of sources.

(i) Depositional processes may be deduced from any distinctive lithological and sedimentary structures present; some processes can be diagnostic of particular depositional environments.

(ii) Even if depositional processes are unknown, certain groups of structures or sets of strata can be identified as belonging to a distinct depositional milieu; this is frequently a uniformitarian deduction, based on observations from contemporary sedimentary environments.

(iii) Composition of recognized detrital components (including chemical and isotopic characteristics) can yield information concerning source region lithologies, weathering processes and direction of provenance, although there may be less information preserved in metamorphosed, schistose rocks.

(iv) Directional sedimentary structures may indicate palaeoslope and palaeocurrent trends and so help to delineate basin morphology. These are difficult to interpret meaningfully in complexly deformed rocks.

(v) Lateral and vertical facies transitions can be related to major events in basin evolution. Again, in

intensely deformed, poorly exposed terrain such as that
under consideration, these relationships are all too seldom
preserved.

In this analysis, depositional processes and environments
are considered together, in relation to each of the lithofacies
designated and described above. Succeeding this, provenance
and indications of weathering are discussed, for relevant
lithofacies, followed by inferences from palaeoslope
indicators. Finally, an attempt is made to integrate all of
these into a model of basin evolution and deposition, within
a regional stratigraphical context.

4.2   ORAVISALO ASSEMBLAGE INTERPRETATION

4.2.i   Oravisalo quarry lithofacies interpretation

It is difficult to interpret depositional processes and
environment in this instance because of the scarcity of
outcrops and lack of continuity with other lithofacies.
Distinctive features of lithofacies A are massive units
containing clasts of granite and amphibolite up to 1 m across,
with abundant feldspar,suggesting an originally sandy arkosic
matrix free of finer detritus. The diamictite texture seems
appropriate for a debris flow origin but even so, it cannot
be used to discriminate between the most probable environments:
submarine canyon, alluvial fan or glacigenic deposition
(e.g. Middleton and Hampton 1976, Rust 1979). The compositional
and textural immaturity favour proximality to the source with
little reworking and rapid burial.

In contrast, lithofacies B shows sub-rounded clasts in
a distinctly finer psammo-pelitic matrix indistinguishable
from Rääkkylä lithofacies metapsammites and is more clearly a

product of subaqueous deposition. This facies is clearly a
thin, local development and directly overlies basement; the
latter is relatively enriched in muscovite for several meters
below the tectonized unconformity which may be indicative of
limited pedogenesis. Overall, this appears to record a
rapid transition from subaerial conditions to a depositional
environment lacking in coarse detritus, possibly consistent
with transgression over a surface of low relief.

4.2.ii Kankaala lithofacies interpretation

No sedimentary structures have been noted at all and it
is not totally certain that the granule-sized quartz and
feldspar are truly detrital rather than recrystallized during
metamorphism. Although the nature of depositional processes
cannot be established, the presence of abundant, apparently
non-detrital sulphide suggests deposition under water rather
than in an aeolian environment. Neither sulphides nor
feldspars are characteristic minerals in the widespread
Jatulian orthoquartzites of Karelia, within which a large
terrestial (including aeolian) component is evident
(Ojakangas 1965, Pekkarinen 1979a).

The alternative to a psammitic detrital quartz-feldspar
protolith would be a chemical silica-feldspar precipitate.
This has analogues at Outokumpu, where a now metamorphosed,
cherty horizon was apparently associated with ore genesis and
hydrothermal leaching and exhalation (Huhma and Huhma, 1970).
In the present instance a highly alkaline brine would be
required, to facilitate subsequent feldspar precipitation;
deep weathering of a nearby granitoid terrain would be a
logical source for such brines.

4.2.iii Oravisalo post office lithofacies interpretation

As in the previously described facies, limited, poor
exposure, tectonic modification and lack of stratigraphic
context make interpretation difficult. An additional difficulty
is uncertainty as to whether the protolith was of effusive or
pyroclastic nature.

4.3   PYHÄSELKÄ ASSEMBLAGE INTERPRETATION

4.3.i Rääkkylä lithofacies interpretation

Although abundantly exposed, at least by Finnish standards,
the lithofacies is monotonously homogeneous and apparently
bed thickness can exceed outcrop dimensions. Sporadic
thinner beds with distinct planar boundaries are observed
while still less commonly, basal load deformation features
attest to successive rapid accumulations of unconsolidated
sediment (Figure 25).

Only at one exceptional locality (94.1 20.2) have sedimentary
structures within a bed been noted. These comprise megaripples
(wavelength to amplitude ratio about 4) and foreset accum-
ulations, the latter giving rise to cross-lamination within
several successive tabular beds. Rare transitions from psammite
to pelite are the only indications of grading that have not
been obliterated by metamorphic recrystallization; it is not
clear to what extent present grain sizes reflect original
characters, but 0.1 - 1 mm seems a likely range. Because of
the general lack of grain size and mineralogical variation
irrespective of bed thickness, the facies is in a sense well
sorted; this could be a function of detrital sizes inherited
from the source area, rather than simply a result of final
depositional processes.

The abundance of feldspar, particularly plagioclase, and the vast volume of this lithofacies suggest abundant supply and rapid dispersal and burial of compositionally immature sediment. All these features would be consistent with detritus from an intermediate plutonic or bimodal volcanic source, deposited in a deltaic or shallow marine environment and subsequently reworked and redeposited by high density turbidity currents. The general thickness of beds showing large load casts seems more consistent with the latter than stormed-induced reworking of shelf deposits. The cross-bedded sets referred to above are not diagnostic of a particular environment. Though they could form fluvially, such bed forms are encountered in coarse clastic pebbly sandstone, associated with more typical turbidites (Walker 1979). The difference to note here is the absence of pebbly detritus, presumably due to weathering or winnowing processes at an earlier stage of sedimentation.

Further indications that the Rääkkylä lithofacies was formed by sediment gravity flows comes from lithologies in the apparently interbedded Sammallahti and Niittylahdenrantatie lithofacies discussed below (Sections 4.3.ii and 4.3.iii).

Since it has not been possible to identify abundant sedimentary structures and lateral bed geometry it is not easy to assign the lithofacies to a particular setting, for instance within a submarine fan complex. The lithologies most closely resemble the arenaceous Facies C_1 (low pelite: psammite ratio) of Mutti and Ricci Lucchi (1975), or massive sandstone facies of Walker (1978); the Bouma turbidite sequence is not strictly applicable. Within the context of the submarine fan depositional model, these occur in the commonly channelized proximal, suprafan area or beyond this region, comprising major aggradational depositional lobes (Figure 40b).

Tectonic strain and lack of continuous outcrop preclude
recognition of broad lenticular geometry within the Rääkkylä
lithofacies. Furthermore, the dominance of this lithofacies
in North Karelia may indicate a departure from the concept
of a single source with radial dispersal onto a fan area -
possibly multiple sources gave rise to an accumulation of
coalesced fans. Alternatively the scale of fan development
may have been considerably larger than the area presently
preserved with more distal metapelitic deposits eroded away
or tectonically dislocated.

Estimations after allowance for structural modification
suggest the facies is at least 3 km thick at Oravisalo with
a preserved areal distribution exceeding 2500 km^2; it may
thus approach a volume of 10 000 km^3.

4.3.ii Sammallahti lithofacies interpretation

Absence of grading and depositional current structures
limits interpretation of this lithofacies. Pertinent however
are sharp transitions between matrix-supported coarse clastic
layers and alternating Rääkkylä facies lithologies. This grain
size contrast may reflect variations in intensity of the same
type of depositional process, or else requires a somewhat
different provenance for the coarser grains whose composition
indicates an extrabasinal origin. Both lithologies correspond
well to the massive and pebbly sandstone facies in Walker's
(1978) review of deep-sea coarse clastic deposits. Walker does
not consider it appropriate to classify such deposits according
to the conventional Bouma turbidite sequence, but rather as
derivatives of higher-density turbidity currents, with some
degree of fluidized flow in the final stages.

Stanley (1975) assigns grain flow and fluidized flow
deposits to high gradient environments (4.5° - 10°) such as

canyon axes, slopes and fan valleys. Lowe (1976) contends
however that natural grain flows of sandy material (dispersion
maintained by granular collisions) cannot exceed 3 cm in thickness
and that fluidized flows (dispersion maintained by upward flow
of escaping pore fluid) can only transport material finer
than medium-grained sand. Thus these mechanisms·alone could
not have been responsible for transport of the coarser detritus
in the Sammallahti lithofacies, which may be up to 1 cm in size,
occurring in beds over 10 cm thick.

Middleton and Hampton (1976) speculate on progressive
devolution of sediment-gravity flows from slumps (which possess
internal cohesion and finite yield strength) through debris
flows to high then lower density turbidity currents with fluid-
ized and liquefied (grain) flows during the waning stages of
deposition (Figure 40a). Lowe (1976) recognizes another category,
that of density modified grain flow in which interstitial grains
aid in maintaining and supporting a dispersion of coarser, even
sandy conglomeratic material. This could well be represented
by the Sammallahti lithofacies, with intervening finer meta-
psammites being the products of more conventional turbidity
current flows. If the interstitial matrix of a density-modified
grain flow attains a degree of cohesion it may evolve into a
debris flow, or the converse may occur on loss of cohesion by
fluidization. This may also be recorded in the Sammallahti
lithofacies as coarse-grained pockets,(as opposed to discrete
layers) are sometimes seen, indicating massive transport and
progressive disaggregation of loosely consolidated intra-
formational deposits.

This implied reworking, and the inferred processes indeed
suggest a proximal, relatively high-gradient environment within
a submarine fan complex - a broad channel or suprafan feeder.

It must be stressed however that tractional deposition in a marine shelf-slope environment, with poor sorting, cannot be categorically refuted.

4.3.iii Niittylahdenrantatie lithofacies interpretation

The abundant, coarse and stratified intraformational conglomerates, lacking slump structures, strongly suggest deposition from debris flows. According to Middleton and Hampton (1976) these may develop on slopes as low as 1°, with an interstitial matrix composed of water and clay or sand often considerably less than 20% of the flow volume. Because of matrix cohesion, and moreover, higher density than that of ambient fluids, debris flows are able to passively transport very coarse material; in the Niittylahdenrantatie lithofacies lenticular or tabular clasts may exceed 5 m in length.

Other depositional mechanisms are indicated for the intervening massive sandstone metapsammites. Figure 27d shows a scoured channel with the overlying bed displaying basal cross-bedding and planar stratification in the upper part. Apparent lenticular nature could have been induced by tectonic boudinage but the overall aspect resembles deposits of facies A, B_1 and C_1 in Mutti and Ricci Lucchi's (1975) classification which are assigned to canyon or channel fill in the proximal or inner part of a submarine fan complex. This applies especially to the debris flows deposits (Mutti and Ricci Lucchi facies A_2).

Beds II through VIII in Figure 27e represent a section that corresponds closely to a small scale coarsening and thickening upward sequence as defined by Mutti and Ricci Lucchi (1975). These are considered to develop as an actively prograding submarine feeder channel builds a suprafan depositional lobe (Figure 40c). Mutti and Ricci Lucchi's model is derived from

ancient sequences but accords with the descriptions from
modern fans by Normark (1977);in his classification, suprafan
deposits belong to the middle fan environment. A sequence
thinner than the normal 30-40 m is considered by Walker (1978)
to be found in shallow suprafan channels resulting from rapid
channel switching.

The Niittylahdenrantatie lithofacies as presently exposed
may be several hundred meters thick and can be traced along
strike for nearly 10 km; it cannot be ascertained whether it
is a linear or areally extensive lithofacies. Walker (1978)
tabulates some characteristic features of ancient feeder channels;
a wide ratio of length : width : depth is evident which encompasses
the apparent dimensions of the Niittylahdenrantatie lithofacies,
but detritus is frequently finer. Stanley (1975) also reports
fan valley widths of 1-10 km with coarse clastic fill.

The debris flows of this lithofacies almost certainly
developed over finer-grained sequences and do not represent the
earliest deposits associated with basin subsidence. The inference
is supported by the ubiquitous presence of detritus resembling
the Rääkkylä lithofacies, indistinguishable from intercalated
metapsammites. It is therefore concluded that the deposits
represent a period of active fan progradation due to an increase
in supply of detritus, or a relative lowering of sea level. A
corollary of this is that a more proximal channelized environment
is superimposed or incised on more "distal" deposits as a major
coarsening and thickening upward sequence (Walker 1978). A
further corollary is that the ultimate source area would become
more distant, hence the presence of only reworked, intraformational
detritus. Because of the tectonic position of the exposures on
the truncated, inverted limb of a major F_2 fold, it cannot be
seen in what manner the originally overlying strata recorded
cessation of progradation and fan avulsion.

4.3.iv Niva lithofacies interpretation

This differs substantially from the rest of the Pyhäselkä assemblage even though its upper part may be stratigraphically intercalated with metapsammites of the Rääkkylä lithofacies. It has not been possible to identify depositional features or bed thicknesses within the massive grey metapelites. Consequently it is not known whether they represent successive dilute turbidite mud flows, pelagic deposition or subaqueous deposition of tuffaceous clays.

The fine lamination in the greyish cherty layers are probably depositional in that they are deformed by early structures which predate the regional S_2 metamorphic segregation (Figure 28). Some of these laminae may show depositional truncation or cross-bedding (Figure 28) but these cannot be used to deduce depositional environment. Equally these layers could have been thin siliceous (or leached mafic) tuffaceous deposits, mud or silt turbidites or partly chemical precipitates and pelagic clay deposits, all subject to mild current reworking and winnowing.

The thin, green, amphibole-rich layers show a distinct competence contrast with respect to the above two lithologies (Figure 28b) and tend to show sharp contacts. It may well be that they represent thin mafic sills intruded within a sequence of poorly consolidated, water-rich sediment. The latter qualifications are made to account for

1) lack of obvious contact aureoles, probably due to rapid convective heat loss,

2) "stratabound" nature of sills: overburden and water load may have prevented brecciation or extrusion of magma as a submarine flow, and

3) one contact which shows features analogous to load casts , except that relative cleavage / bedding vergence suggest that the layers are on an F_2 inverted limb (that is, lobes bulge upwards, not downwards). A tentative interpretation is that this is analogous to both pillow lavas and sedimentary load casts , in that both essentially represent the interface between two fluids with a considerable density contrast (complicated by the temperature difference between magma and water).

Failure of the sill to rise to a higher level during injection may have led to small scale doming of the coherent but ductile upper surface before freezing of the magma. Such interpretations are nevertheless constrained by uncertainty concerning the effects of tectonic boudinage, as is possible in the example shown in Figure 28b.

If they are not sills, these layers are possibly altered tuffaceous deposits or even chemogenic. The absence of obvious terrigenous detritus indicates quiescent conditions with either low detrital input to the basin, or else deposition on the distal basin floor.

4.4 KETTÄMÖ ASSEMBLAGE

4.4.i Kettämönniemi lithofacies interpretation

The four lithological subfacies recognized do not appear to differ significantly in observed depositional structures except in the following.

1) There is a tendency towards increased muddy matrix (pelitic) content from the Petäjikkökallio and Pitkänurmi

subfacies through the Karvanurmi subfacies to the Lentosärkkä
subfacies. Relatively higher cohesion might have influenced
depositional mechanisms though Middleton and Hampton (1976)
suggest, for instance, that debris flows are feasible with as
little as 2% fluid-clay component.

2) A general decrease in coarse detritus occurs towards
the Lentosärkkä subfacies and Salonkylä lithofacies. This is
apparently an original stratigraphic younging feature (see
Map 3) relating directly to sedimentary features and the
depositional mechanisms may have varied accordingly.

The thick-bedded metapsephites of the lower part of the
lithofacies may be readily classified as diamictites resulting
from debris flows. Since clast composition and size suggest
proximality to source it should not be assumed that such debris
flows occurred only within a submarine canyon-fan context.
Terrestrial alluvial fan deposition would also be considered
possible, since the scale and overall trend of the sequence is
consistent with examples given by Rust (1979). It is also
possible to envisage an alluvial fan developed at a basin
margin leading to redeposition in a submarine fan environment,
further blurring distinction between the two systems.

Walker (1979) suggests that the problem of such ambiguous
coarse clastic deposits can be resolved by examining their
context with respect to enclosing deposits. For instance in this
case, debris flows associated with classical Bouma turbidite
sequences would indicate a submarine rather than terrestrial
setting. Complete Bouma sequences are very rare in the study
area but well preserved graded bedding with basal load casts
indicate turbidity current deposition between coarser clastic
deposits in the lowermost exposed Petäjikkökallio facies.

Carter and Lindqvist (1977) describe the development
of a submarine fan and canyon complex in New Zealand with an
abrupt transition from a cliffed shoreline wedge to proximal
canyon deposition. In the former, deposits include cross-
bedded sands with traction features,but lack turbiditic graded
beds; this contrasts with deposits of the Kettämönniemi litho-
facies. In a review of resedimented rudites, Kelling and
Holroyd (1978) describe a canyon fan sequence with many
characteristics in common with the Kettämönniemi lithofacies.
These authors have measured certain parameters such as maximum
clast size, rudite bed thickness and depth of basal scour with
which the following qualitative observations from the
Kettämönniemi lithofacies correlate:

1) maximum size of extrabasinal clasts (e.g. quartzites,
granitoids) gradually decreases upwards through a sequence
100-200 m thick,

2) intrabasinal clasts are more abundant in the upper
part, but maximum size is not so variable; they are presumably
more buoyant and readily transported, and

3) thickness of rudite units remains fairly constant
(considering presence of abundant detritus coarser than 1 cm as
defining rudite), seldom exceeding 3 m. Such a sequence is
interpreted by Kelling and Holroyd (1978) as gradual starvation
of extrabasinal detritus, but maintainence of similar depositional
environment and energy levels throughout. In general, the
Kettämönniemi rudites (or metapsephites) can be correlated
with Kelling and Holroyd's Type 2 deposits, being poorly channelled
with a variable degree of organization, assigned to a distal
rather than proximal canyon environment. The more proximal facies
could have been exposed and reworked or tectonically displaced by
movement on the Kettämö thrust.

In addition to this evident overall progression from
debris flow dominated deposition in an inner fan or canyon to tur-
bidity current deposition of arenites on the suprafan, there
are some smaller scale features of interest. While detailed
stratigraphical correlation may be hampered by poor exposure
and the possibility of undetected dislocations, it appears
that the Pitkänurmi subfacies belongs to a younger cycle of
rudites, overlying the Petäjikkökallio and Karvanurmi subfacies
(see Map 3). Thus through a sequence possibly 50 m thick a
transition is observed from quartz metapsammites and metapelites
to deposits at first resembling those of thin, matrix-supported
debris flows (Figure 29f) then an increasing proportion of
granitic detritus (Figure 29g) and finally more massive clast-
supported diamictite with granite boulders exceeding 1 m in
maximum dimension (Figure 30e). This passes relatively abruptly
upwards (about 30 m) into feldspathic metagraywacke which
gradually diminishes at the expense of thinner quartz meta-
psammites of the Lentosärkkä subfacies and Salonkylä lithofacies.
This total sequence has the appropriate magnitude and properties
of a coarsening and thickening upward sequence reflecting
progradation of a canyon or suprafan system, followed by
relatively rapid avulsion due to lateral channel migration or
filling of the channel system. Such progradation could result
from increased detrital supply, relative   lowering of sea level
or increased gradient of basin margin due to tectonic activity.

The interpretation of the underlying Petäjikkökallio and
Karvanurmi subfacies is less clear, but both appear to have
 arkosic metagraywackes lowermost with coarser material increasing
upwards followed by a fairly abrupt transition to progressively
finer lithologies, in turn succeeded by the Pitkänurmi sequence.
Figure 31a shows the uppermost coarse deposits seen in the
Petäjikkökallio subfacies and records, within 20 m, a transition
from deposition of conglomerate containing meter-sized extra-

basinal detritus to thin-bedded quartz psammites and pelites.
This again suggests a major progradation and fan abandonment or
channel migration but is also of suitable scale and morphology
for a subaerial alluvial fan (Rust 1979). As discussed earlier,
the former interpretation is favoured because of the resemblance
of the massive graded metagraywackes to turbidites.

The second general observation is of two distinct but
frequently juxtaposed lithologies, especially within the
Karvanurmi and Pitkänurmi subfacies. The first is a relatively
homogeneous, almost arkosic sediment lacking in internal structures
and containing occasional coarser granitoid or quartzite clasts
while the second is a disorganised diamictite with a muddy matrix
and irregularly distributed, dominantly intrabasinal clasts. The
two types are shown in figures 29 and 30.

The difference may derive from depositional mechanisms, or
source material variations. In the former case, the muddy
diamictites could have been deposited as cohesive debris flows,
whereas the more homogeneous arkosic beds could have been
separated from finer material by turbulence and gravity settling
and settling during less dense, non-cohesive sediment-gravity
flows. In envisaging differences due to different source, the
absence of the muddy diamictites from the Petäjikkökallio
lithofacies is pertinent. Thus these apparently develop higher
in the sequence than the somewhat better sorted arkosic conglom-
erates. This may indicate that there were no accessable muddy
horizons available for reworking and downslope transport in the
early stages of deposition, but rather only coarse-grained
winnowed detritus. An additional possibility is that the arkosic
conglomerates were deposited by sediment-gravity flows along the
canyon or channel axis and the muddy diamictites from slumps
devolving into debris flows derived from the steeper channel flanks

and walls; a higher proportion of finer detritus would potentially be available from erosion of adjacent levees and interchannel areas.

The Lentosärkkä subfacies contains minor amounts of both these lithologies but also shows the only definitive example of slumping seen. It appears however that the slumped material was quite consolidated and coherent as the discordance transects two depositional units rather than following the pelitic top of the lower bed (Figure 32a).

In summary, the Kettämönniemi lithofacies can be interpreted in terms of two major coarsening and thickening upward sequences associated with basin margin subsidence and canyon - submarine fan progradation, culminating in thin- to medium-bedded planar deposits more characteristic of the middle fan environment.

4.4.ii Salonkylä lithofacies interpretation

There is no precisely defined boundary with the presumably underlying Kettämönniemi lithofacies and the lithofacies seems likely to record progressively more distal deposition in the same submarine fan system. There is no indication however of thicker metapelite units indicative of the outer fan or basin plain environment, though metamorphic recrystallization may have obscured some grain size distinctions. Rather planar, thin- to thick-bedded quartz-rich metapsammites (with no discernible grading) suggest correlation with Mutti and Ricci Lucchi's (1975) turbidite facies C_1 and D_1, assigned to suprafan depositional lobes and interlobe deposits respectively. Rare channelized environments may have been present, containing intrabasinal clasts as evident in Figure 32f.

4.5   HAMMASLAHTI ASSEMBLAGE

4.5.i Suhmura lithofacies interpretation

It is possible that this overlies the Mulo lithofacies, at
first being represented by rusty massive metapsammites, but it
is difficult to demonstrate this conclusively. The relationship
is potentially an important one as it records a major change
in depositional environment and mechanisms, by way of influx of
quartzose terrigenous detritus. For the most part the Suhmura
lithofacies is indistinguishable from the Salonkylä lithofacies
and only their occurrence on opposite sides of the trace of the
Kettämö thrust zone precludes their correlation in this study.
These monotonous quartz-rich thin- to thick-bedded units there-
fore probably represent deposition from turbidity durrents, but
this should not be tacitly accepted,in view of the absence of
very thick beds and preserved sedimentary features of any kind.
Other mechanisms are plausible, if not demonstrable - for
example, storm-induced or other forms of tractional reworking
in a shelf environment. Between Suhmura and Niittylahti the
lithofacies is deformed in two major, reclined F_3 antiforms
separated by a narrow inverted limb zone. Simple trigonometrical
calculation from limb dip and area suggest a present thickness
exceeding 1 km.

4.5.ii Kilpelänkangas lithofacies interpretation

Like the Salonkylä lithofacies, this has a lithological
counterpart above the Kettämö thrust zone, being clearly
comparable to the Lentosärkka subfacies. It is possible that
in truncating the basin sequence, thrusting has juxtaposed this
proximal sequence with more distal deposits. The Kilpelänkangas
lithofacies does contain some thick-bedded graded feldspathic
metagraywackes which can likewise be interpreted as deposits from
high-density turbidity currents. There is however a relatively
high proportion of thin-bedded deposits, including metapelite

that could have formed as suprafan interlobe deposits, or more
distally. There are additional minor components resembling
the dark calcareous metapelites of the Mulo lithofacies (but
lacking in feldspars) which, if stratigraphically autochthonous,
may indicate typical "background" hemipelagic or chemical
sedimentation at low rates of terrigenous input, punctuated
and maybe reworked by deposition from turbidity currents.

4.5.iii Mulo lithofacies interpretation

The Mulo lithofacies is distinctive in that it lacks
identifiable clastic material, including mica that could have
had a detrital clay precursor. Its relationship to other
lithofacies is not well established except that rusty meta-
psammites interbedded in the (inferred) upper part could be
contiguous with the more normal quartz metapsammites of the
Suhmura lithofacies.

Since clastic provenance is not relevant, it seems most
appropriate to discuss origin in direct relation to depositional
environment. The two most likely environments suited to
accumulation and preservation of non-clastic deposits are a
distal or restricted basin floor, or a shallow-marine shelf with
a low-relief hinterland. Unfortunately, in this instance no
criteria appear to convincingly discriminate between these.

With regard to processes of origin, there would seem to
be three major possibilities occurring individually, or in
any combination.

a) Pelagic chemical precipitation took place whenever
saturation in any ambient sea-water component was attained.

b) Deposition was by brine refluxing, by upwelling onto
a shelf, or a flow of denser brines into a ponded basin. This

could exist as a closed chemical system as long as some form
of chemical gradient was present.

c) Hydrothermal fluid exhalation and precipitation
(with or without volcanic activity) occurred with leaching of
underlying strata including any intrabasinal intrusive rocks
at depth.

d) This possibility is not strictly chemogenic, namely
distal air-fall tuffs extensively leached and recrystallized.

Apart from the latter possibility, for which no direct
evidence is available, and in spite of metamorphic recrystall-
ization, it seems that the original mineralogy can be expressed
as the following four major components.

1) Alkali feldspar, now present as microcline and lesser
Na-plagioclase.

2) Carbonate component. Though now exclusively calcite,
abundant Mg-tremolite porphyroblasts, yet virtual absence of any
magnesian phyllosilicate phase may indicate that the original
carbonate was dolomitic.

3) Carbonaceous component. The present graphite may
possibly result from denaturing of hydrocarbons.

4) Sulphide and possible original sulphate component.
This is now expressed as abundant pyrite and pyrrhotite with
minor chalcopyrite and sphalerite.

Each of these components will be discussed with respect to
the three possible mechanisms already outlined although it should
be recognized that treating each component independently may
not be entirely realistic. For example, sulphide and organic

carbon contents can vary sympathetically with each other
(Trudinger and Williams, 1981) and authigenic feldspar could
conceivably form diagenetically from an illite-dolomite reaction
thus being linked to the carbonate component. The depositional
age of the Mulo lithofacies is also relevant to discussions
of its origin. Unfortunately only indirect evidence is available
but an age between 2.0-2.1 Ga will be deduced below to be
probable. A lower, conservative estimate comes from U-Pb dating
of detrital uraninite in a clast from the earliest Karelian
metasediments at Kiihtelysvaara (Pekkarinen 1979a). This is
2.34 Ga so that the deposits in question are younger still and,
in turn appear to pre-date development of the basin in which
the Mulo lithofacies was deposited. It is generally accepted
that by this stage of the Early Proterozoic, coeval with or
younger than the main development of Superior-Hamersley type
banded iron formations, the atmosphere and upper ocean levels
were weakly oxidizing but that deeper ocean conditions remained
anoxic and reducing (Button and others 1981). It also seems
that dissimilatory sulphate reduction by anaerobic bacteria
and photosynthetic O_2 production were already established as
metabolic pathways (Trüper 1981). Both these inferences are
pertinent to discussions of the origin of Mulo facies components.

(1)   Alkali feldspar component - authigenic versus
      metamorphic origin:

In the Mulo lithofacies, tremolite porphyroblasts overgrow
primary and/or differentiation layering defined by variations
in sulphide, feldspar, calcite or graphite proportions
(Figure 33a). In adjacent lithofacies porphyroblastic mineral
growth has also taken place post-tectonically with respect to
the major $S_2 - S_3$ fabric development. This is mostly biotite,
with very rare garnet or cordierite; overall the metapelitic-
metapsammitic assemblage is dominated by quartz, biotite and

muscovite, with no evidence of K-feldspar or sillimanite.
These observations indicate that Mulo lithofacies microcline
existed before and remained in equilibrium during the maximum
metamorphic conditions attained, which was below the K-feldspar
sillimanite isograd for micaceous metapelites. It is not so
straightforward to establish whether the feldspar is a product
of S_2 - S_3 differentiation itself. This might have been feasible
from an illite precursor during progressive lowering of ambient
pH_2O, but it then needs to be explained why feldspar has not
replaced mica elsewhere, for instance in the graphitic and
calcareous biotite metapelites of the Kukkupää lithofacies, and
in the Outokumpu area generally (Peltola, 1960). Furthermore,
Mather (1970) inferred that during biotite growth, detrital
microcline is an unstable reactant, and is consumed with chlorite
and muscovite, releasing water in the process. On this basis,
the bulk of the Mulo lithofacies feldspar is regarded as
authigenic.

Origin of authigenic feldspar: diagenetic replacement or chemical precipitate?

Authigenic feldspar is frequently associated with dolomites.
Regular alternations have been interpreted as possibly cyclic
in origin (Braun and Friedman 1969) with deposition of carbonate
occurring at the sediment-air interface under supratidal
hypersaline conditions, and feldspar precipitating from per-
alkaline brines derived from more abundant groundwater during
wet seasons, having a pH of 9.5-10 and high $a[K^+]/a[H^+]$. Baskin
(1956) reviewed occurrences of authigenic feldspar and noted
invariably relict detrital cores, with authigenic overgrowths.
K-feldspar overgrowths have been noted in the Mulo lithofacies
but are late metamorphic in origin (Figure 17c), while it seems
unlikely that original grains have avoided extensive
recrystallization.

In a study of feldspathic dolomites, Buyce and Friedman (1975) suggest either a diagenetic origin or overgrowths on clastic components,(whose size is less than 100 μm)possibly of aeolian if not detrital origin. They conclude that the diagenetic processes could be alteration of illite during dolomitization or the alteration of pyroclastic tephra first to zeolites and then to feldspar under conditions of increasing alkalinity. Nevertheless they do admit the possibility of direct precipitation of K-feldspar in restricted hyperalkaline basins, referring to contemporary examples in western USA.

Russell and others (1984) have assumed a non-detrital, non-pyroclastic source for massive celsian deposits associated with exhalative barite-Pb-Zn-Ag mineralization in upper Precambrian rocks of the British Isles. Mixing of sources from different environments is implicit: dense alkaline brines from a cratonic shelf being neutralized by metal-enriched, acidic solutions exhaled onto a ponded basin floor. Anomalous celsian and barite deposits correlate well with barium content and concentrations in stratigraphically underlying formations and isotopic characteristics of the ores are also consistent with. derivation from these; there is thus no need to invoke a cryptic volcanogenic source, and extensive hydrothermal convection is rather attributed to increased heat flow and fracture access in a kratonic margin under tension.

Using the above examples as a guide, the following model may be proposed for feldspar formation within the Mulo lithofacies.

A) Peralkaline, maybe also peraluminous groundwaters were derived from extreme weathering of a low-relief terrain composed dominantly of Archaean granitoid gneisses and providing negligible terrigenous detritus to adjacent basin margins.

This postulate is consistent with the high chemical index of
alteration for Lower Proterozoic terrigenous sediments, a feature
indicative of severe weathering (Eriksson and Soegaard 1983).
Under such conditions, K-feldspar would remain in soil profiles
after leaching of Na-feldspar but typical mature and residual
deposits appear to have been quartz and kaolinite, the latter
usually altered to illite by potassium fixation or sericite and
kyanite during metamorphism (Eriksson and Soegaard 1983,
Metzger 1924). A high abundance of K^+ in groundwater, and
ultimately in oceans is implied.

Garrels and Howard (1959) have experimented with hydrolysis
of K-feldspar at atmospheric conditions with variable pH,
governed largely by $a[K^+]:a[H^+]$. At higher values of $a[H^+]$, K^+
is replaced in feldspars, and to a lesser extent muscovites,
and relevant reactions are:

$$KAlSi_3O_8 + H_2O \;\rightarrow\; HAlSi_3O_8 + K^+ + OH^- \tag{1}$$

$$2KAlSi_3O_8 + 3H_2O \;\rightarrow\; H_4Al_2Si_2O_9 + 4SiO_2 + 2K^+ + 2OH^- \tag{2}$$

$$HAlSi_3O_8 + 4OH^- \;\rightarrow\; Al(OH)_4^- + 3H_3SiO_4^- \tag{3}$$
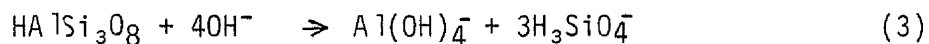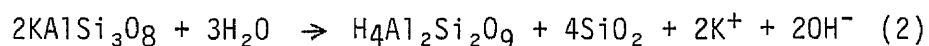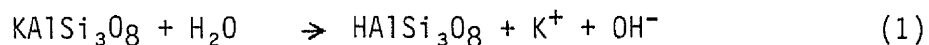
It is clear that for Al to be transported as $Al(OH)_4^-$, the
solution must already be alkaline, a state favoured by reactions
(1) and (2). Production of kaolinite in reaction (2) is also
of interest in that aluminosilicate may thus be transported
in suspension and subsequently flocculate.

B) Increased concentration and density due to evaporation
could have led to flow of alkaline brines from the basin shelf
to a deeper basin environment, in which the Mulo lithofacies
was deposited. These flows could have contained any evaporite
minerals formed on the shelf but there is no evidence for these,
unless carbonate derives from such a source. Alternatively,
carbonate and authigenic feldspar precipitated by brine refluxing
on a shelf according to the suggestion of Braun and Friedman (1969)

and were subsequently transported by turbidity currents into a deeper basin environment.

C) In the case of transport in solution, a mechanism must be found to facilitate precipitation of feldspar or its precursor, which should be favoured by a general lowering of pH while activity of K^+ remained high. In the example of Russell and others (1984) acid solutions are thought to have derived from hydrothermal circulation, partly by fixation of Mg and OH^- in chlorite. The absence of obvious signs of unusual metal enrichment in the Mulo lithofacies make it more difficult to invoke such a second exotic brine source in the present case. Instead, the high graphite and sulphide content point to the possibility of microbial biota having been active in an anoxic basin floor environment. Several mechanisms could feasibly participate in decreasing ambient pH:

(i) There could have been microbial oxidation of sulphur compounds by non-photosynthetic anaerobes, using H_2S as an electron donor.

(ii) Dissimilatory sulphate reduction may have been operative, using organic carbon and H_2O as electron donors. Such a mechanism would also generate CO_2 and in the presence of Ca^{++} or Mg^{++} could lead to dolomite or calcite precipitation, again yielding H^+. High activity of H^+ might also explain formation of kaolin, feldspar or a hydrated precursor to feldspar, rather than flocculation of illite or Mg-montmorillonite. Production of kaolinite in reaction (2) is also of interest in that aluminosilicate may thus be transported in suspension and subsequently flocculate.

To summarize, feldspar is regarded as authigenic, due to severe weathering of a felsic terrain yielding little if any clastic detritus. Precipitation of feldspar or a precursor or

kaolin flocculation took place on neutralization of alkaline brines, either on a shallow marine platform or more probably in a restricted basin in which reducing, relatively acidic conditions existed, to some extent maintained by microbial activity. An external brine system is not deemed necessary but on the other hand such activity cannot be disproven.

(2)   Calcareous component of Mulo lithofacies

Carbonate deposition was widespread in the Proterozoic and should require less explanation in this context. Though it could be described as micritic, the abundance of graphite and sulphide inclusions make recognition of grain boundaries difficult; it is however doubtful whether this records original grain size and only rare indications of carbonate laminae occur where content of other minerals is low. Depositional environment has therefore not been established. There are however some aspects to carbonate sedimentation that deserve mention.

a) Dolomite is invariably more abundant in Precambrian than Phanerozoic carbonates and this is considered to correlate with secular trends in hydrospheric evolution, dissolved Mg being higher in the Precambrian (Veizer 1976). This alone does not prove that dolomite precipitated from seawater directly, or whether it was diagenetic as in most (but not all) environments today. Due to the concomitant high activity of dissolved silica, Mg-sepiolite or chlorite precursors would have readily flocculated, at the expense of dolomite (Eriksson and Truswell 1978).

Microprobe analyses of the Mulo lithofacies show carbonate to be exclusively calcite; magnesium is however abundant in tremolitic porphyroblasts. Since Mg-bearing phyllosilicates are only present as fine-grained accessories, it seems likely that Mg was originally present as carbonate; the low detrital

input may have precluded flocculation of clay minerals.

b) For precipitation on the basin floor, or redeposition from precipitation elsewhere to have been preserved, slightly acidic conditions would need to be maintained, in the presence of high Ca (and Mg) activity. This was also suggested above as a requirement for feldspar precipitation. Carbonate precipitation could also have been related to CO_2 production during dissimilatory reduction of sulphate.

(3)   Carbonaceous component of the Mulo lithofacies

Given the moderate degree of metamorphism, it is possible that carbon isotope ratios may still discriminate between a volcanic and organic origin for the graphite. It may be more difficult to identify a component due to abiogenic reduction of carbonate. Sympathetic trends in sulphide and organic carbon content have been noted by Trudinger and Williams (1981) who also reiterate general observations of abundant biota associated with present day volcanic exhalative centres. A biogenic origin seems plausible for the graphite of the Mulo lithofacies, if its deduced age and observations inferred from Precambrian microfossils are valid. According to a review by Awramik (1981) a diverse bacterial assemblage existed in the Warrawoona Group of Australia at 3.5 Ga and dissimilatory sulphate reduction by 2.2 Ga is implied by the morphological similarity of microfossils in the Gunflint cherts (Canada) to present day cyanobacteria. It is not precisely known when oxygen-releasing photosynthesis developed even if it was responsible for significant O_2 production; accumulation of O_2 was presumably buffered by abiogenic oxidation of $S^=$ to $SO_4^=$. Biogenic sulphate reduction can occur anaerobically (dissimilatory) generating CO_2 in the absence of O_2, or aerobically (assimilatory), which yields further O_2 and $S^=$ (Trüper 1981).

Table 6 shows in a simplified manner the potential interaction between sulphur, oxygen and bacteria that were likely to have been extant during the early Proterozoic. Of the various kinds of metabolism, dissimilatory sulphate reduction was probably then, as now, the most important form of anaerobic respiration, and was probably most prolific in the uppermost layers of sediment, or at the sediment-water interface where nutrient supply is presumably greatest (Trudinger and Williams 1981). A source of organic carbon is also required and two may be envisaged - from decay of lithotrophic anaerobes (forming a closed system sulphuretum) or a pelagic contribution from decay of pelagic biomass.

The foregoing indicates the potential role of organic matter during deposition of the Mulo lithofacies, principally in oxidizing hydrocarbons to CO_2, reducing sulphate, yielding $S^=$ for metal fixation and tending towards decreased pH in an overall reducing environment. If a redox stratification existed in the basin, with an upper oxidized layer (not necessarily corresponding to the photic zone), O_2 generation may have been of considerable importance in that abiogenic oxidation of sulphide to sulphate may have exerted a major buffering influence on the supply of sulphate available for bacterial dissimilation, or $S^=$ available for cation fixation. This leads naturally to consideration of sulphide formation of the Mulo lithofacies.

(4)   Sulphide component of the Mulo lithofacies

This is obviously stratified, except where present as euhedral crystals in late mobilized calcite veins, and syngenetic, depositional or diagenetic origin is indicated. Firstly, sources for sulphur and cations (dominantly Fe) will be examined, and then potential mechanisms for their fixation.

microbial reduction
of sulphur compounds

microbial oxidation
of sulphur compounds

OXIC

progressive generation
of O_2, accumulation at
first buffered by
abiogenic oxidation of
$S^=$ to $SO_4^=$

ANOXIC

assimilatory sulphate reducers
- require $SO_4^=$ reservoir
- generate O_2 and $S^=$

if high ambient O_2 must
compete with abiogenic
oxidation

dissimilatory sulphate reducers
- utilize $C_{organic}$ + H_2O as e^- donors
- utilize $SO_4^=$ as recipient of $8e^-$
- generate CO_2

anaerobic photolithotrophs
- utilize light as energy source

direct assimilation of $S^=$
- requires ambient $S^=$ in environment

anaerobic chemolithotrophs
- utilize H_2S as e^- donor and
  assimilation of sulphur

Data abstracted and greatly simplified from Trüper (1981); cladistic relations not strictly phylogenetic.

There is as yet no direct evidence for any sulphate
deposits that could have formed as evaporites on a tidal shelf
and been redeposited by density gravity flows, or for precipit-
ation at the site of deposition. It seems most logical to
postulate derivation of sulphur species directly from ambient
seawater, and the information in Table 6 , indicates that
sulphur and dissolved H_2S availability could have been controlled,
or at least buffered by bacterial activity. Dissimilatory
reduction of seawater is at present the most discernible
bacterial process (Trudinger and Williams 1981); assimilatory
sulphate reduction requires oxygenated conditions and seems less
applicable in the present instance.

Sulphides of the Mulo lithofacies are dominated by pyrite,
though pyrrhotite, chalcopyrite and sphalerite are also present.
It is not clear whether pyrrhotite represents a metamorphic
replacement of pyrite, though textural evidence from the
Tikkala lithofacies metaturbidites suggest increasing sulphide
instability as metamorphism progressed, indicating a general
increase in oxygen fugacity, which could preclude pyrrhotite
replacement of pyrite. Such an inference may be vindicated
by the observation of pyrite within late quartz veins in the
Tikkala lithofacies, and in calcite veins in the Mulo lithofacies
(Figure 33c).

Goulevitch (1980) and Nicholson (1980) have both examined
sulphide deposition in the Middle Proterozoic Koolpin and
Kapalga formations of northern Australia and have related
pyrrhotite-bearing base-metal deposits to both hydrothermal
deposition derived from diagenetic devitrification of tuffaceous
strata and hot, metal- and boron-enriched brines, while pyrite
occurs in interbedded pelites lacking in other mineralization
and apparently representing normal periods of basin deposition.

From thermodynamic and experimental considerations, the
pyrrhotite deposits would have required relatively higher
temperature, and lower activity of oxygen and dissolved sulphur
species; these conditions also favour transport of Pb, Zn, and
Cu. These deductions may be used to postulate that Mulo litho-
facies sulphide deposition took place without metal enrichment
from an external or diagenetic hydrothermally circulating
brine source.

More sphalerite and chalcopyrite may merely reflect
relatively low Cu and Zn concentrations in ambient seawater
in general, or more local influences from contemporaneous
volcanism and weathering in the basin drainage system.
According to Button and others (1981) the early Proterozoic
oceans would have displayed a marked redox stratification, with
an upper oxidized layer rich in dissolved Fe, particularly as
$Fe(OH)_2$. This would be precipitated as Fe-oxides or carbonate
on shallow marine shelves but as sulphides in acidic basin-
floor environments.

(5)   Summary of depositional interpretation of Mulo lithofacies:

The distinctive mineralogy of the Mulo lithofacies is
interpreted in terms of chemical deposition in an anoxic,
slightly acidic basin-floor environment. Carbonate may have
precipitated directly from seawater as dolomite while sulphide
formed syndepositionally or diagenetically from components
present in the ambient seawater. There is a likelihood of
bacterial sulphate reduction having played a roled in sulphide
generation, but this was probably indirect rather than critical
or limiting.

Feldspar probably developed syndepositionally or dia-
genetically from a hydrated precursor, transported in K-rich,
peraluminous brines introduced from a shelf environment whose

hinterland was subject to severe weathering and restricted
supply of detritus. Apart from this, no external source of
components appears to be necessary, though neither diagenetic
leaching nor volcanogenic exhalation can be discounted as
having been active in initiating or sustaining hydrothermal
circulation.

4.5.iv Kukkupää lithofacies interpretation

Since they are intimately interbedded with one another,
the three different lithological types defined and described
earlier would appear to belong to the same depositional
environment, whatever difference there may have been in their
respective depositional mechanisms and sources. Nowhere was
any indication of coarser than pelitic detritus found, except
in the apparently conformable transition with overlying
Tikkala lithofacies metagraywackes. From the features, distal
or starved basin deposition is inferred.

(1)   Micaceous metapelites in the Kukkupää lithofacies

Sedimentary structures have not been noted in these
lithologies due to extensive recrystallization or transposition
into S_3 and S_5 foliations. A detrital origin is however
inferred from microscopically observed silt-sized quartz
grains, preserved in spite of mica grain growth and reorient-
ation. If such an origin is accepted, then a combination of
hemipelagic, and mud or silt turbiditic processes is indicated.
Fine-grained turbidites have been reviewed by Piper (1978) who
found them difficult to assign to specific settings within the
turbidite-submarine fan model. Nelson and others (1978) suggest
however that grain size and rate of sediment supply influence
fan morphology such that when fine-grained,a fan does not

develop; rather, sediment is delivered to the basin plain
via long leveed valleys. According to Piper's review, at
least half of the sediment in basin plains beyond the cont-
inental or basin slope is delivered by turbidity currents; this
applies in particular to thicker pelitic units, with thinner
and laminated silts or muds being more characteristic of
proximal environments and other mechanisms of deposition.
Because tectonic obliteration of fine laminations may have
occurred in the Kukkupää lithofacies, it is not easy to make
use of these conclusions and difficult to distinguish between
laminated and massive deposits. Nevertheless, a primary
origin is apparent for the thicker units shown in Figure 33f
and deposition from individual turbiditic flows is probable.

(2)   Massive siliceous metapelites in the Kukkupää
        lithofacies

There are two reasonable explanations for the origin
of these more competent layers, especially conspicuous on
weathered surfaces (Figure 34f), namely, chemical precipitation
of impure cherty deposits, or resulting from pyroclastic
deposition. In the former case, variations in abundance of
detrital clay deposition or flocculation, and in pH of ambient
seawater, would permit development of cherty layers, since it
is widely held that the Proterozoic ocean was almost
saturated with respect to silica (Veizer 1976). Since the Mg
content was also apparently greater, with a high Mg/Ca ratio,
slightly alkaline waters would lead to flocculation of the
clay sepiolite as well   (Eriksson and Truswell 1978). Thus
chemical deposits need not always be pure siliceous precipitates,
an observation consistent with the composition of the present
examples in the Kukkupää lithofacies. The presence of post-
depositional concretions also affords strong evidence for syn-
sedimentary chemical precipitation (Figure 34b). Deposition

from sporadic pyroclastic falls or flows may also have been a mechanism of formation of these lithologies. In the studies referred to earlier, Goulevitch (1980) and Nicholson (1980) described a sedimentary transition from sulphide-bearing pelites to turbidite graywackes analogous to the Kukkupää-Tikkala lithofacies transition considered here. The lower part of the sequence also contains abundant tuffaceous pelites, and tuffaceous cherts, sometimes albite-rich. These have a highly variable chemical composition resulting from considerable syndepositional to diagenetic interaction with seawater and contamination or dilution by syndepositional silica precipitation and clay flocculation. A similar, complex origin seems highly plausible for this component of the Kukkupää lithofacies, but cannot be demonstrated in the absence of relict pyroclastic features such as grading and devitrified glass or crystal shards.

3. Calcareous lithologies in the Kukkupää lithofacies

These constitute a widespread, though minor component of the lithofacies. In this regard the laminated lithologies shown in Figure 34 a, c are of particular significance. These laminations are generally less than 1 cm thick, consisting of gradational alternation of silt-sized quartz with abundant calcite and biotite in the lower part, and thinner graphite-mica-rich laminae. A sedimentary rather than metamorphic origin is favoured, based on the presence of low-angle cross-lamination. Accepting this, there are a number of conceivable mechanisms of deposition or formation.

a) Each couplet represents a distinct depositional event, indicating a discrete source for carbonate and reduced carbon with lamination due to subsequent differential settling or current winnowing. Two principal modes of deposition can further be envisaged, namely tractional currents in a shallow marine environment, reworking authigenic carbonate and authigenic or

detrital silica and clay, and secondly thin, relatively low energy turbidity currents transporting terrigenous and/or authigenic shelf detritus into a deeper basin environment. The latter seems more tenable in view of the nature of associated sediments and in that it provides a mechanism for both gravitational settling and preservation of reduced carbon which would seem less probable in a shallower oxidized environment conducive to carbonate precipitation. Depositional mechanisms of thin fine-grained turbidites are not so well understood as for coarser and thicker deposits, and are also more difficult to distinguish from alternative processes such as the winnowing effect of contour currents. Nevertheless Piper's (1978) definitions and descriptions of graded and laminated turbidite silts and muds show a close resemblance to the laminites being discussed here, in thickness being less than 5 cm and the presence of irregular low-angle lenticular bedding. Gravitational, graded settling in silt and mud turbidites is considered to be due to variations in extent of clay flocculation or degradation of flocs leading to changes in flow density and hence load-carrying capacity. These deposits appear to be relatively distal or else belong to levee environments in more proximal locations (Nelson and others 1978).

b) A variation on the above mechanism is that only the lower quartz-carbonate part represents turbidity current deposition whereas the upper layer results from intervening hemipelagic deposition, comprised of clays and the decayed remains of pelagic algae or bacteria. Such deposition might have been continuous, or periodic depending on nutrient supply or climatic seasonality such as controls algal "blooms". However such speculation should be restrained until an organic origin for the carbon can be demonstrated, either by a chemical analysis of any remaining kerogen, or by stable isotope analysis.

c) A detrital, resedimentary origin is accepted, but
carbonate reduction took place in the upper parts of layers
because of redeposition in a reducing environment, or biological
activity at the sediment -water interface. The latter seems
untenable in that the major metabolic pathway, namely dissim-
ilatory sulphate reduction also tends to oxidise hydrocarbons
to CO_2. Inorganic reduction seems less easy to refute, but if
this was the case it seems improbable that laminations could be so
well preserved,without evidence of "reduction fronts" transgress-
ing them. It is therefore concluded that some combination of
mechanisms a) and b) is more logical and appropriate.
Interpretation of deposits as thin laminite -turbidites is
favoured rather than as tractional shelf sediments chiefly
because of their context within sulphide-bearing and massive
metapelites.

Thus two differing origins and sources for carbon may be
envisaged, with possible isotopic fractionation, the carbonate
reflecting ambient seawater values and organic carbon being
relatively enriched in ^{12}C. (It is doubtful however whether
different metabolisms or types of environment such as stromatolitic,
pelagic open sea or deep-sea benthonic could be accurately
discriminated). Carbon isotope analysis would be of considerable
interest in evaluating whether homogenization occurred between
these two components during the regional metamorphism as is
widely held to have taken place (Schidlowski 1981).

4.5.v Tikkala lithofacies interpretation

The three areas described earlier (Section 3.4) Tikkala,
Hammaslahti and Kumpu have each yielded valuable information about
depositional environments and processes.

(1)   Interpretation of deposits at Tikkala

It has already been noted that the sequence depicted in Figure 35 is an excellent example of a fining- and thinning-upward sequence which could be expected to form as a submarine channel terminates at a suprafan depositional lobe within a submarine fan system; the vertical fining is ascribed to lateral migration of the feeder channel. The upper part of the sequence in Figure 38c shows similar features although there is a possibility that it records more channelized deposits with erosion of underlying strata. To reiterate, these exposures form two successive, typical fining- and thinning-upward sequences representative of an inner fan distributary system in the sense of Mutti and Ricci Lucchi (1975), or the synonymous middle fan of Bouma and Nilsen (1978). As at Tikkala, such deposits are frequently enveloped by fine-grained laminites representing hemipelagic deposition or overbank interchannel deposits. In this case, the regular, fine banding suggests the latter, rather than the former mode of deposition (see Figure 40b).

An interesting feature of these metagraywackes is the uniformity of maximum grain size (about 8 mm) suggesting an upper limit due to early sedimentary processes in a fluvial, deltaic, or shallow marine environment or physical constraints on the slope and density of the turbidity current. Where coarser detritus is present it is markedly so suggesting a natural paucity of detritus in the size range 1-3 cm. This could be attributed to inherent mechanical properties of the parent material (for example a granitoid with grain size less than 1 cm) or could reflect the transition between turbidity current deposition and higher density debris flows.

Deposits ascribed to the latter are abundant in the upper part of the section in Figure 35 with the weak, discontinuous stratification and coarse clasts, yet absence of slumping being

quite diagnostic. Their presence indicates either increasing proximity of a major channel due to lateral migration or incision, or a tectonically induced change in depositional environment. This would evidently have been a sufficiently major event to permit erosional access to a variety of lithologies, with transport of clasts exceeding 1 m in diameter. To have remained coherent yet somewhat rounded, these clasts would need to have been partly lithified prior to reworking but it is not certain whether they represent penecontemporaneous basin-margin deposits or derive from more ancient, extrabasinal sequences exposed by tectonic uplift. In this respect the dolomitic clasts are of interest as they could be the only surviving evidence for penecontemporaneous carbonate deposition at the basin margin. Other clasts now weathered to limonite may once have been sideritic, a common component of Proterozoic shelf and platform sequences (Button and others 1981).

The presence of these carbonate clasts leads to speculation concerning water depth and composition. Clearly turbidity currents and debris flows could be initiated in shallow, oxygenated water, carrying carbonate clasts into a deeper euxinic environment. Given rapid burial and sedimentation, calcareous detritus can be preserved below the carbonate compensation level in Phanerozoic basins (Hesse 1975), and there are indeed two calcareous metaturbidites enclosed within laminated sulphide-bearing metapelites at Tikkala (Figure 38c). Observations from the Mulo and Kukkupää lithofacies, and the lack of evidence for detrital pyrite seems to indicate that oxygenated water existed in upper levels, while it was apparently not incongruous to have stable coexistence of sulphides, graphite and carbonate.

(2)   Interpretation of Tikkala lithofacies at Hammaslahti

With the exception of one polymictic horizon containing
cobbles of granitoids and orthoquartzite (Figure 37f),coarse
detritus seems to be intrabasinal and lithologically similar
to host sediments, suggesting more distal deposition or alter-
natively a later stage of depositional history when reworking of
earlier deposits dominated. Nevertheless, thick-bedded turbidites
closely resemble those at Tikkala in lithology, structure and
grain size and a similar mode of deposition is inferred. Some
interesting post-depositional structures are illustrated in
Figure 37b. These appear to be small clastic dykes probably
intruded due to overburden loading rather than fluidization
though this too may have occurred in the overlying graywacke.

The actinolite bands remain problematical, though a dia-
genetic origin is ruled out as they fail to cross lithological
boundaries. A chemically altered mafic tuff, or even intrusive
sill seems more likely as a protolith than sediments of detrital
or chemical origin.

(3)   Mineralization in the Tikkala lithofacies at Hammaslahti

The Tikkala lithofacies is here host to several massive,
brecciated Cu-Zn ore bodies whose genesis has not been examined
in this project. Nevertheless, some apposite observations can
be presented here.

In a chemical and isotopic study of the ore deposits,
Hyvärinen and others (1977) concluded that lighter δ^{34}S values
and a constricted range of isotope ratios in the ore with respect
to country rock, as well as a negative correlation between δ^{34}S and
S content, indicate remobilization of synsedimentary sulphides.
Substantial remobilization is ascribed to tectonism and both the
local structural analysis of Gaál (1977) and more regional aspects
of this study confirm that the ore deposits have been completely
deformed by two isoclinal fold generations, (F_2 and F_3) succeeded

by reorientation into steep, N-S trending shear zones (S5). The irregular fracturing of fragments in much of the brecciated pyrrhotite-chalcopyrite ore also suggests that graywacke clasts were lithified prior to disruption (Figure 33d, 43g). It is not so clear whether this tectonic mobilization was itself responsible for the principal concentration of sulphide minerals; the northern sphalerite-rich body in particular frequently shows a disseminated stratiform banding in altered metagraywackes, and also appears to be deformed by the earliest tectonic structures observed (Figure 33 f.). If this indicates a syngenetic to dia- genetic origin for ore formation, then porous Tikkala lithofacies graywackes may have been advantageous in allowing fluid circulation and concentration but would seem less likely as a source for metal cation leaching. The Pb-Pb isotope systematics modelled by Vaasjoki (1981) do suggest a mixing of ambient Pb with that from an older source, for which graywackes derived from the Archaean granitoid basement would be appropriate. Nevertheless, litho- geochemical studies have delineated a well-defined aureole immediately enclosing the deposit, particularly with respect to Mg enrichment and K depletion (T. Karppanen spoken comm.). This would seem to be of too small a volume to correlate with merely a closed system, local source of enrichment sufficient to account for the ore bodies present.

In the event of formation through a larger scale hydro- thermal system it is necessary to consider potential source rocks, mechanisms of initiating and sustaining hydrothermal brine circulation and environments or conditions of deposition. The extent of deformation precludes recognition of evidence for deposition from dense brines ponded in restricted depressions, or for instance identification of depositional breccias with a sulphide matrix. The stratified banding referred to earlier is therefore tentatively assigned a post-depositional origin within consolidated, but porous clastic sediments.

Contemporaneous igneous rocks frequently accompany sulphide ore deposits and heat from a magmatic source is widely considered to be crucial in establishing and maintaining hydro-thermal brine circulation,causing leaching of adjacent strata, including consanguineous intrusive or volcanic horizons. These processes are considered relevant for example in the Pyhäsalmi-Pielavesi and Outokumpu massive sulphide deposits NW of the study area (e.g. Mäkelä 1974, Huhtala 1979, Koistinen 1981). At Hammaslahti however there are no convincing volcanogenic lithologies identified with certainty, unless the actinolitic schists prove to be altered, subsequently metamorphosed mafic sills or tuffs. Though they do not appear to be volumetrically substantial, they may be significant as general indicators of contemporaneous igneous activity elsewhere in the depositional basin. In particular, correlation with the Tohmajärvi meta-basite complex is appealing, though other stratigraphical and petrographical inferences suggest that the Tikkala lithofacies is in general younger.

Neither is there as yet compelling evidence for a (preserved) proximal feeder zone to the ore deposit such as the stockwork proposed at Outokumpu (Koistinen 1981) or stringer ore at Pyhäsalmi (Helovuori 1979). Given these uncertainties, an alternative mechanism to direct magmatic involvement may be sought.

In the study referred to earlier, Goulevitch (1980) presents an attractive theory of ore development in a sedimentary sequence stratigraphically and lithologically resembling those of the Hammaslahti assemblage. Essential similarities include

1) an early period of deposition characterized by disseminated pyrrhotite and pyrite (the former reflecting

relatively higher temperature and lower activity of dissolved
sulphur species), with intercalated cherty, calcareous and
carbonaceous horizons; this is analogous to the Kukkupää
lithofacies of the present study,

2) temporally and spatially associated volcanogenic
lithologies, mostly tuffaceous that can be likened to the
Rauansalo lithofacies in part, and

3) an overlying sequence comprising graywacke-mudstone·
turbidite deposits recalling those of the Tikkala lithofacies.

Though sulphides are abundant as disseminations in the
lower sequence, Goulevitch maintains that the absence of major
accumulations indicates a causal connection between distribution
of ore bodies and the younger tuffaceous deposits but not in
the usual sense of volcanic exhalation. Rather he suggests that
burial of the tuffaceous sediments beneath the younger turbiditic
deposits led to brine generation as a consequence of diagenetic
alteration. In particular, devitrification of volcanic glass -
an exothermic process - is believed to yield a solution that is
mildly acidic (due to incorporation of OH⁻ into diagenetic clay
minerals), with low dissolved sulphur content and high salinity
increasing its efficiency as a metal-carrying brine. This is
subsequently expelled during compaction and dewatering,
invading the overlying sediment, or gaining access to the basin
floor where the brine, being denser than ambient seawater, could
be expected to concentrate in topographic depressions. Basin
floor irregularites could be of tectonic origin but depositional
units such as the coarser metagraywackes of the Tikkala lith-
facies might well themselves produce a low-relief topography,
thus ponding brines, or else due to their relatively greater
porosity, may be potential sites of ore precipitation within
sediment. Such a scenario appears consistent with at least
some pre-tectonic aspects of the Hammaslahti ore bodies.

(4)   Interpretation of deposits in the Kumpu area

A debris flow origin seems again to be the most logical interpretation for the deposits in Figure 38a and b, showing contorted (also tectonically deformed) irregular intrabasinal clasts in a poorly sorted sandy matrix. In particular the example in 'b' may imply internal cohesiveness sufficient to develop slumping, which has not been noted anywhere else in the lithofacies; the debris flows at Tikkala are moderately well stratified. The lithology approaches Mutti and Ricci Lucchi's (1975) Facies A₂ and F which belong to inner fan channel fill or channel flanks and foot of basin slope respectively. The latter alternative seems more likely on the basis of their context, adjacent lithologies tending to be sulphide-bearing and thin-bedded. The massive meta-turbidites that are present contain notably less feldspar than at Hammaslahti or Tikkala implying a different source or distal fractionation effect, but are relatively carbonate-rich.

The Tikkala lithofacies is found in a zone of complex deformation. At Kumpu, for example, depositional younging and S_2/S_3 relations indicate that the sequence youngs eastwards on the inverted limb of an F_2 fold, tightly folded again by F_3 and reoriented into a steep N-S attitude by S₅ shear zones. In spite of this, it is possible to infer that nowhere is the Tikkala lithofacies adjacent to a depositional basement, although granitic and orthoquartzitic detritus are present. Rather the evidence is that these coarser clastic sediments were deposited away from the basin margin on top of an earlier sequence of pelitic sediments, partly detrital, partly chemical and possibly in part volcaniclastic, comprising the Kukkupää lithofacies. This offers no constraint on water depth, apart from the basin slope necessary to initiate and sustain the sediment-gravity

flows responsible for this relatively abrupt influx of coarse terrigenous detritus. A further implication of this observation is that these deposits aresome of the youngest currently represented in the area.

4.5.vi Rauansalo lithofacies interpretation

At this stage, analysis is based on only two exposures (Figure 39) in which a turbidity current origin is indicated for somewhat channelized graded units 0.1-1.0 m thick. Nevertheless there seems to be no diagnostic discriminant between turbidity currents carrying epiclastic volcanogenic detritus and a true submarine pyroclastic flow. In either case, the mafic composition and tectonostratigraphic continuity with the Tohmajärvi area to the south suggest deposition on the submarine flank of a contemporaneous volcanic complex. Perhaps one criterion for distinguishing pyroclastic flows from more conventional epiclastic turbidites would be the presence of quartz or granitic detritus in the coarser parts of the beds. If quartz phenocrysts were absent from the dominantly mafic Tohmajärvi complex then the finding of detrital grains would indicate the sedimentary mixing of different source components. These are indeed rarely observed but the majority of coarse detritus is feldspathic and angular which is opposite to the generally observed trend of quartz occurring in increasing abundance with progressive sediment maturity. This, and the mafic (now hornblende-rich) matrix argue against considerable detrital supply from Archaean basement or early Proterozoic clastic deposits.

The low-angle discordance shown in Figure 39c, and the irregular brecciation in the massive underlying bed may be causally connected; increasing fluid pressure from below could have initiated instability and slumping of uppermost beds with

brecciation and fluidization in more competent layers. An external fluid source is suggested in that the breccia matrix is different from the host rock but the proximity of volcanic or hydrothermal activity cannot be asserted on this basis alone.

In summary, the few outcrops of this lithofacies examined indicate a volcaniclastic origin but the relative contribution of submarine pyroclastic flows is unclear. Submarine deposition on the flanks of a volcanic edifice, presumably to the south, is indicated, with absence of basement detritus due to burial beneath lava flows, or development of the volcanic complex wholly within the basin itself.

4.5.vii Kalliojärvi lithofacies interpretation

There are occasionally clear relicts of original structures, indicating a turbidity current origin but this explanation need not be universally applicable - for instance storm or tidally induced reworking of shallow marine shelf sediments would be difficult to either prove or refute. The most notable distinction of this lithofacies is the higher proportion of metapelite and consistently thinner depositional units than in the other clastic lithofacies. If a turbidite origin is accepted, then a more distal environment is indicated, correlation being made with Mutti and Ricci Lucchi's Facies C, and especially D, representing non-channelized outer fan or basin-plain deposits.

Extent and stratigraphical position of the lithofacies is at this stage unclear.

4.6 SUMMARY OF DEPOSITIONAL ENVIRONMENTS (see Figure 41)

(i) Although distinction between the Savo and Höytiäinen provinces remains appropriate, similar depositional processes would have been operative in both. The latter province has, however, better preserved and more diverse primary depositional features; only the Niittylahdenrantatie lithofacies, along the NE margin of the Savo province shows such features abundantly.

(ii) No substantial basin margin, shallow-marine or terrestrial deposits have been preserved in the Savo province (within the study area). Restricted development of arkosic meta-sediment occurs within the Oravisalo assemblage (Oravisalon louhos and Kankaala lithofacies), but only in close proximity to its presumed Archaean basement source. Metapsephites indicating basement or intrabasinal provenance define, in part, the Sammallahti and Niittylahdenrantatie lithofacies along the eastern margin of the province, and the Niinikkosaari lithofacies immediately to the N of the study area. These deposits clearly belong to a proximal environment within the context of a submarine fan or distributary system and show evidence of deposition by sediment-gravity flows. The Niittylahdenrantatie lithofacies in particular appears to include deposits formed by debris flows and high-density turbidity currents, suggesting a submarine canyon or major inner fan channel system. The bulk of the monotonous Rääkkylä lithofacies is reckoned to have originated as medium-to thick-bedded turbidites in a prograding submarine fan.

(iii) The Kettämönniemicoarse-clastic lithofacies is interpreted as a proximal, partly channelized fan sequence deposited along the western margin of the Höytiäinen province, passing upwards and laterally into metapsammites less diagnostic of a suprafan - middle fan environment (Salonkylä and Suhmura

lithofacies). Similar sediment-gravity processes were responsible
for deposition of the Tikkala lithofacies, though in a setting more
distal with respect to provenance. It appears to record a
terrigenous clastic influx prograding over the fine-grained
Kukkupää lithofacies. This probably comprises metamorphosed mud
and silt turbidites for the most part but some evidence exists
for calcareous and siliceous chemical deposition. The Mulo
lithofacies is of more limited extent and believed to be almost
wholly chemogenic, with authigenic feldspar, carbonate and sulphide.
The Rauansalo lithofacies appears to be volcaniclastic, derived
from coeval mafic igneous rocks and may be in part pyroclastic.

FIGURE 40 : RESEDIMENTED CLASTIC DEPOSITIONAL MODELS

a   Envisaged permutations and development of water-sediment
    dispersions during progressive evolution and devolution of
    sediment-gravity flows, and resultant deposits, according
    to Middleton and Hampton (1976) and Walker (1979).

b   Proposed location of lithologies within a submarine canyon
    and fan distributary system, from Walker (1979).

c   Typical depositional sequences identified with channel
    migration (fining- and thinning-upward sequences) and
    prograding depositional lobes (coarsening- and thickening-
    upward sequences), according to Mutti and Ricci Lucchi
    (1974).

Walker (1979) modified after Middleton and Hampton (1976)

DEBRIS FLOWS	PEBBLY CONGLOMERATES	MASSIVE SANDSTONES	CLASSICAL TURBIDITES

TIME AND/OR SPACE

SLUMP, SLIDE

REMOLDING LIQUEFACTION — DEBRIS FLOW

TRACTION

TRACTION

TRACTION

UPWARD FLOW OF PURE FLUIDS

GRAIN INTERACTION

FLUID TURBULENCE

HIGH CONCENTRATION TURBIDITY CURRENT

RIVERS IN FLOOD

LOW CONCENTRATION TURBIDITY CURRENT

FLOW INITIATION	MAIN LONG DISTANCE TRANSPORT PROCESS	LATE STAGE MODIFICATIONS

FLUID TURBULENCE   LIQUEFACTION   DISPERSIVE GRAINFLOW   MATRIX STRENGTH

Walker (1978) fig 13

FEEDER CHANNEL

DEBRIS FLOW CONGLOM- SLUMPS

DEBRIS FLOWS CONGLOMERATES

SLOPE INTO BASIN

CONGLOMERATES INVERSE-TO-NORMALLY GRADED

GRADED-BED

GRADED STRATIFIED

UPPER FAN

TERRACE

THIN BEDDED TURBIDITES ON LEVEE

PEBBLY SANDSTONES

MASSIVE SANDSTONES

BRAIDED

MID-FAN

SUPRAFAN LOBES

SMOOTH

PROXIMAL

INCISED CHANNEL

CLASSICAL TURBIDITES

NEW SUPRAFAN LOBE

BASIN PLAIN

LOWER FAN

THIN BEDDED

NOT TO SCALE

Multi and Ricci Lucchi (1974)

THINNING UPWARD CHANNEL FILL SEQUENCE

INNER CHANNELIZED FAN (DISTRIBUTARY SYSTEM)

THICKENING UPWARD PROGRADING OUTER FAN SEQUENCE

OUTER FAN (OUTBUILDING SYSTEM)

FIGURE 41 : SCHEMATIC DEPICTION OF DEPOSITIONAL ENVIRONMENTS

Proposed reconstruction of major depositional units and environ-
ments for the Höytiäinen province and eastern margin of the Savo
province. Projection is from south to north. Crossed hammers
indicate location of Hammaslahti copper-zinc deposit.

Rääkkylä
lithofacies
metapsammites

Canyon-inner fan
deposits of Sammallahti
and Niittylahdenmäki
lithofacies

Channelized inner
fan deposits of
Kettämö lithofacies

Suprafan deposits
of Suhmura and
thinner-bedded
Kettämö lithofacies

Chemogenic deposition
of Mutö lithofacies
in ponded basin

Terrestrial and
aeolian deposition

Tikkala lithofacies
suprafan prograding
across distal and
chemogenic Kukkupää
lithofacies

Carbonate shelf deposits
and earlier Jatulian
platform cover contribute
detritus to Tikkala
lithofacies

Distal, basin floor thin-
bedded turbidites and
sulphidic, calcareous
and carbonaceous deposits
of Kukkupää lithofacies

Pyroclastic or epiclastic
deposits derived from
and flanking Tohmajärvi
igneous complex

Tohmajärvi mafic
igneous complex

eroded outlier
of Jatulian
orthoquartzite

Niva lithofacies
metapelites and
Oravisalo assemblage

5.   DISCUSSION OF SEDIMENTARY PROVENANCE

While not of value in deducing depositional environments,
sediment composition may provide information on the nature of
source-region lithologies. This may be examined directly from
relict detrital mineralogy or indirectly inferred from bulk
chemical compositions and isotopic systems given at least a
partial understanding of surficial and depositional fractionation
processes. The latter approach is obligatory in intensely
altered rocks but in the study area some lithologies preserve
sufficient detrital grains to permit direct interpretations.

5.1   DEDUCTIONS FROM PETROGRAPHICAL OBSERVATIONS: INTRODUCTION

Petrographical studies of metasediments reveal diverse
information that may be classified into four general categories.

i) Mineral growth in relation to deformational fabrics,
linking metamorphic and structural history.

ii) Metamorphic mineral assemblages and parageneses which
may be used to establish physical conditions during metamorphism.
Original lithology and composition can clearly influence
subsequent mineralogy.

iii) Relict clastic grain relations that may provide
evidence for depositional processes and environment.

iv) Composition of relict detrital grains.

It is the latter category that is considered here - the
first two were discussed in relation to deformation and micro-
structures and the third, where pertaining to lithofacies interpret-
ation, in the preceding section. Likely provenance is therefore

discussed for each lithofacies in which original detrital mineralogy is recognizable. As with structural and lithological parameters, the Savo and Höytiäinen provinces, and their respective lithological assemblages show significant and diagnostic variations in primary mineralogy and inferentially in provenance. Attention is preferentially given to coarser-grained lithofacies as larger grains tend to be more diverse and readily identifiable; smaller grains are also more subject to modification during recrystallization (Figure 42).

5.2   PETROGRAPHY AND PROVENANCE OF ASSEMBLAGES IN THE SAVO PROVINCE

5.2.i   Provenance of Oravisalo assemblage

Doubt has already been expressed as to the detrital nature of feldspar and quartz grains in the Kankaala lithofacies but if such an origin is accepted, a proximal granitoid basement provenance is logical (Figure 43a). This is demonstrable for the Oravisalo quarry lithofacies, containing as it does pebbles lithologically indistinguishable from the underlying Archaean basement. Similar detritus is evident in the Niinikkosaari lithofacies N of the study area proper (Map 2) except that fragments of orthoquartzite are locally common and angular. At the Oravisalo quarry, the unconformity itself offers incontrovertible evidence that Archaean basement was periodically exposed while the above evidence indicates the local existence and erosion of quartzites resembling Jatulian deposits, now only preserved between Kaavi and Koli to the N of the study area (Map 1).

5.2.ii Provenance of the Pyhäselkä assemblage

The mafic to calcareous metapelites of the Niva lithofacies
do not preserve any obvious detrital grain morphology and even
if they include a large volcaniclastic component this is unlikely
to be of value in determining provenance, though an intrabasinal
origin is more probable than epiclastic reworking of older
metabasites. Such a viewpoint is adopted because of the intimate
intercalation of the earlier Proterozoic mafic rocks and
Jatulian orthoquartzites: very efficient winnowing would be
necessary to separate these mixed detrital components but most
evidence indicates poor sorting during deposition.

The metapsephites of the Niittylahdenrantatie and
Sammallahti lithofacies contain intrabasinal clasts derived
from reworking of adjacent sediments and thus provide few clues
as to ultimate provenance. The Sammallahti lithofacies does
additionally contain granule-sized detritus of a quartzo-
feldspathic nature; a granitoid origin is likely for grains such
as that illustrated in Figure 43b but other examples suggest
alternative sources such as orthoquartzites (eg. Figure 43c).

The metapsammites intercalated within both the above litho-
facies and comprising the bulk of the Rääkkylä lithofacies are
extensively recrystallized but distinctive in their plagioclase
feldspar content. Of oligoclase-andesine composition, this is
an ubiquitous, frequently major component of these rocks and
their presumed correlatives throughout North Karelia, as for
example in the northern Haukivesi area (Gaál and Rauhamäki 1971)
and the Outokumpu district (Gaál and others 1975). Modal
abundances range from 5-35% and Na-cobaltinitrate staining
failed to reveal K-feldspar in thin section. In the absence of
a suitable alternative Na-bearing phase, the present plagioclase
is considered to have recrystallized from a detrital feldspar
precursor; this may have been initially somewhat more calcic

in view of the widespread occurrence of accessory garnet and
calcite.

Such abundance of plagioclase is somewhat unusual,
particularly when concomitant with a dearth of K-feldspar -
in general, relative instability of these minerals would favour
preferential concentration of the latter but the abundance of
both should diminish with respect to quartz and illite or
kaolinite with progressive weathering (Eriksson and Soegaard
1983, Folk 1974). In marked contrast to the Rääkkylä lithofacies,
this is observed in the Jatulian platform sequences throughout
Finnish and Soviet Karelia,with successively younger units
showing an increasing maturity, attributed to effective aeolian
or subaqueous reworking of earlier deposits. This culminates in
widespread, extremely pure orthoquartzite and lesser occurrences
of kaolinite-rich lenses, now represented by an abundance of
andalusite or kyanite (Metzger 1924, Hausen 1930, Ojakangas
1965, Pekkarinen 1979). Thus the Rääkkylä lithofacies is
revealed as anomalous in its detrital mineral proportions and
an explanation for this might be found in the nature of its
source regions in addition to any sedimentary fractionation
or metamorphic effects.

In an informal consensus based on available data, Folk
(1974) proffered the conclusion that the abundance of plagioclase
results from rapid deposition in a tectonically active environment
and testifies to derivation from a volcanic or plutonic and
presumably mafic or.intermediate source.

Abundant biotite in the Rääkkylä lithofacies may reflect a
chlorite-montmorillonite detrital precursor which also departs
from the clay mineral assemblage expected upon severe weathering
of the granitoids and felsic gneisses of the Archaean craton.
The weathering of mafic supracrustal rocks, such as occur in

the Archaean greenstone terranes, is a potential source for
suitable detritus. At the current erosion surface these appear
to be areally subordinate to the felsic gneisses and granitoids,
though they could have been more extensive prior to erosion.
The absence of such detritus in deposits of known basement
provenance (see below) tends to mitigate this argument but the
possibility should be considered in any detailed, quantitative
analysis concerning chemical and physical aspects of sedimentation.

If, to reiterate, the adjacent basement rocks of the
Archaean craton were not the major source for deposits of the
Rääkkylä lithofacies, and the foregoing discussion would seem
to indicate this, an alternative must be sought, capable of
having supplied abundant plagioclase-bearing detritus to a
rapidly accumulating depocenter. That provenance direction could
have been from the S and W has not been vindicated by palaeocurrent
data or precise identification of source lithologies. Nevertheless
there is some evidence consistent with, if not demonstrating
this postulate. Firstly, there is the widespread preservation
of Jatulian orthoquartzite and other platform deposits above
the basement unconformity, between Kaavi and Koli and thence
southwards through Kiihtelysvaara towards the Lake Ladoga
(Map 1). This may indicate that Archaean rocks NE of the present
outcrop of the Pyhäselkä assemblage remained blanketed by
Proterozoic cover rocks and hence were inaccessible as source
rocks unless to local graben or in deeply incised valleys.
Secondly, the Pyhäselkä assemblage is in part, if not wholly
allochthonous and the results of this study, and that of
Koistinen (1981) suggest a substantial northwards component of
translation; palinspastic reconstructions would therefore remove
the original depocenter somewhat further away from currently
juxtaposed basement. On the other hand, even a displacement
of several hundred kilometers may be relatively trivial if
consideration is given to the tortuous, seemingly capricious

courses of major distributary systems (such as the Brahmaputra and Niger rivers) and the huge volume of their associated delta-submarine fan complexes. In such cases, basement areas adjacent to the depocenter may contribute comparatively less detritus than more distal actively eroding terrains, and so exert only a minor influence on sediment supply and composition.

The so-called Svecofennidic realm (Laajoki 1983) or Svecofennian terrane (Park and others 1984) comprising the SW part of Finland presents a logical alternative source area in terms of proximality and also composition, with abundant bimodal extrusive and intermediate to basic plutonic lithologies. Though an attractive and realistic possibility, the dearth of geochronological data in the region leaves as speculative, the mutual age relationships between these and the Pyhäselkä assemblage and hence whether or not the Svecofennian rocks even existed during deposition of the latter. Only broadly bracketing constraints exist at the time of writing, with a lower age limit of 1.97 Ga for Raakkylä lithofacies correlatives at Horsmanaho, N of Outokumpu (Koistinen 1981). This determination is derived from a pre-tectonic gabbro within the Outokumpu assemblage, considered to be older than the metapsammite sequence. A conservative, reliable, minimum limit is provided by a zircon-based U-Pb concordia intercept age of 1.925 Ga, reported from Haukivesi by Gaál and Rauhamäki (1971); this refers to mafic plutonic rocks syntectonic with respect to major deformation on the Haukivesi shear zone, which is correlated with Group 4 structures of the present study area. With few exceptions, Svecofennian samples yield marginally younger ages, from 1.88 - 1.92 Ga whether they be synorogenic intrusions, metavolcanic rocks or detrital populations from intercalated metasediments (Simonen 1980, Kähkönen and Laitakari 1983).

Again in the Haukivesi area, Gaál and Rauhamäki (1971)
record diopside amphibolites of volcanogenic origin but consider
them to overlie rather than act as source for the associated
'veined gneiss' complex, here correlated with the Rääkkylä
lithofacies. It must therefore be stated that no conclusive
evidence is available concerning the provenance of the bulk of
the Pyhäselkä assemblage. Derivation from juxtaposed Archaean
basement seems less plausible than from appropriate Svecofennian
lithologies that exist to the S and W but it is equally accept-
able to envisage a more distal source with detritus transported
over great distances within a major distributary system. In this
latter context, the Lappland granulites between the Karelian
craton and Kola Peninsula is worthy of consideration, particularly
as it stabilized during orogenesis at 2.1-2.0 Ga and could thus
have actively supplied detritus to adjacent Karelia (Barbey and
others 1984).

5.2.iii Provenance of Kettämö assemblage

These lithologies contrast markedly with those of the
Pyhäselkä assemblage in composition as in the abundance of
readily identifiable detritus. The coarse, resedimented clastic
deposits near Kettämönniemi have been grouped into subfacies
on the basis of clast composition which seems to vary
sympathetically with stratigraphic position. Thus, rounded
orthoquartzite detritus is diagnostic of the lowermost
Petäjikkökallio subfacies, with feldspar-rich matrix in addition
to granitic clasts. The overlying Karvanurmi and Pitkänurmi
subfacies are defined by a predominance of granitoid-derived
detritus, with an increasing proportion of intrabasinal gray-
wacke clasts and non-diagnostic monocrystalline quartz higher
in the sequence. This trend represents an inversion of source
area stratigraphy as would be expected, with progressive
unroofing of the granitoid basement from beneath the quartzite-
dominated cover sequence.

Individual granitoid clasts closely resemble the more felsic lithologies belonging to the Archaean craton in Karelia and there is little doubt that this represents their source region. The size of some of the granitoid and quartzite clasts alone would appear to indicate proximality to source, though with some degree of resedimentation into a submarine distributary channel. Continued progradation of a submarine fan would be an obvious way in which earlier-formed deposits could be reworked, ensuring a steady, if not overwhelming supply of intra-basinal detritus in the younger part of the sequence.

In contrast to the Pyhäselkä assemblage, both plagioclase and K-feldspar are present as detritus; neither is abundant at silt size but partially rounded coarser grains are common (Figure 43f). These are frequently very turbid such that the presence of relict lamellar twinning is the only practical way to discriminate between the two feldspars yet even so, plagioclase and microcline could be confused. Na-cobaltinitrate staining has however clearly indicated the presence of K-feldspar in both thin-section and hand specimen.

Somewhat surprisingly, the stain impregnated not only discrete detrital grains but some interstitial areas as well. It is likely that this reflects authigenic feldspar cement in suitably porous deposits that lacked an abundance of phyllo-silicate matrix or labile detritus. This would be compatible with the presence of alkaline groundwaters as a consequence of severe weathering and consistent with proposed processes in the formation of the non-detrital Mulo lithofacies.

5.3   PETROGRAPHY AND PROVENANCE IN THE HÖYTIÄINEN PROVINCE

The diverse Hammaslahti assemblage contains several lithofacies lacking in distinctive detritus. The Mulo and Kukkupää lithofacies typically represent chemical and distal basin deposits and the Suhmura and Kalliojärvi lithofacies contain monominerallic detritus, chiefly quartz and consequently do not indicate any specific provenance. The notable absence of feldspar suggests a mineralogical maturity in spite of textural immaturity and is a significant discriminant between metapsammites of this and the Savo Province.

Some horizons within the Tikkala lithofacies contain a considerable diversity of clasts (Figure 37g), including granitoid, orthoquartzite, vein quartz, gneissic lithologies and a variety of intrabasinal origin. As with the Kettämö lithofacies this is compatible with a local basin-margin provenance though in this instance there is evidence of more distal redeposition by sediment-gravity flows.

Maturity inversions are also characteristic, with some poorly sorted metagraywackes showing exceptionally well-rounded mono- and polycrystalline quartz grains (Figure 43d,e,h). Angular feldspar detritus is also present and this is usually correlated with relatively rapid deposition (Folk 1974).

Opalescent blue quartz detritus is a particularly distinctive though minor component of the province. Identical quartz occur as phenocrysts in Lower Proterozoic dykes in NE Finland and also in felsic to intermediate volcanic rocks in adjacent Soviet Karelia. Thus a source region to the E would be implied, there being no appropriate lithologies currently exposed in basement inliers to the W of the Höytiäinen province.

Some of the debris flow deposits at Tikkala contain various types of sedimentary clasts., most of which probably represent reworking of penecontemporaneous basin margin or slope deposits. Pink dolomite clasts could just as equally belong to an earlier distinct phase of sedimentation, perhaps correlated with the Jatulian dolomites at Kiihtelysvaara (Pekkarinen 1979a). However from their context, an intrabasinal origin is favoured and they are perhaps the only surviving evidence for a carbonate shelf environment having existed within the basin of deposition.

5.4   CHEMICAL STUDIES AND PROVENANCE

This approach is a potentially valuable one but demands a realistic understanding of surficial and depositional fraction-ation effects as well as due consideration of diagenetic and metamorphic compositional changes. Isolation of individual components is generally more meaningful than a bulk chemical analysis which could homogenize and obscure distinctive features inherited from different source regions. That is, a typical sediment may represent the amalgamation of several distinct, fractionated end-member components and cannot be modelled so readily as, for example, linear separation of melt and residual components from a magma.

Nevertheless, some generalizations may be made from unpub-lished whole-rock major element analyses made available by Outokumpu Oy. These show Pyhäselkä assemblage metapsammites to be silica depleted and more mafic with respect to mean values for Archaean basement, based on data from Negrutsa (1974) and Pekkarinen (1979a). For comparative purposes, these results are tabulated below:

| | Archaean basement | | Savo province metapsammites | | |
| | Negrutsa (1974) | Pekkarinen (1979a) | Outokumpu Oy (unpublished data) | | |
	n = 173	n = ?	$\bar{x}$	n	σ_n
SiO_2	70.63	69.1	61.99	42	2.54
TiO_2	0.24	0.4	0.52	42	0.07
Al_2O_3	15.12	14.7	12.69	42	1.54
FeO_t	2.90	3.3	6.39	42	1.28
MnO	0.04	0.1	0.08	42	0.02
MgO	0.83	1.8	3.58	42	0.74
CaO	1.56	2.0	2.60	42	0.79
Na_2O	3.83	4.4	2.56	42	0.94
K_2O	3.45	3.0	2.80	42	0.52
P_2O_5	0.01	0.1	0.73	42	0.01

It is readily apparent that Savo province metapsammites remain distinctly less siliceous than mean Archaean basement, even at a population standard deviation of 2σ, while they are considerably enriched in MgO, FeO and CaO. This differs substantially from the composition of sediments known to have a granitoid provenance. Metzger (1924), Väyrynen (1928), Hausen (1930), Ojakangas (1965), Negrutsa (1974 and Pekkarinen (1979a) all conclude that progressive weathering and reworking of basement-derived sediments culminates in mature orthoquartzites with sericite or kaolinite matrix. Such a compositional evolution has been accorded general validity by Eriksson and Soegaard (1983) who quantify the process in terms of the Chemical Index of Alteration, defined thus:

$$CIA = Al_2O_3 / (Al_2O_3 + CaO + Na_2O + K_2O) \times 100.$$

Average mudstones have CIA values of 70 - 75 but for typical
Proterozoic or Archaean sediments they may rise to over 90,
indicating more intense chemical weathering (Eriksson and
Soegaard 1983). In contrast,values for Savo province meta-
psammmites are distributed about a mean of 61, with a modal
range between 62 and 67;lower values may reflect 'contamination'
by an authigenic component such as calcite.

However, an alternative,or additional factor that may
cause sediment composition to deviate from the more usual trend is
post-weathering loss of silica. Although quartz is apparently a
residual end-member of the weathering process, silica may be
subsequently removed from sediments: if, as is widely maintained,
the Precambrian ocean was relatively undersaturated with respect
to silica, then much of this component may have been dissolved
and remained in solution . Furthermore,solution transfer processes
during $S_2 - S_3$ and S_5 cleavage differentiation presumably liberated
silica,producing abundant quartz veins. Chemical sampling tends
to be biassed against collecting from these yet it has been
estimated that they may account for up to 10% by volume of such a
rock sequence (Beach 1976, Stephens and others 1979).

Hence, unless these factors can be accurately assessed
and modelled, quantitative use of chemical data is restricted
in its application and on this basis alone, a basement provenance
for Savo province metapsammites can neither be discounted nor
confirmed.

A somewhat different, very fruitful geochemical approach
is the analysis of REE patterns as employed by Taylor (1979).
Such an approach is based on an apparent secular trend that
can be used to discriminate between Archaean and younger deposits.

Surficial processes are assumed to efficiently homogenize
diverse upper crustal lithologies, thus representing bulk upper
crustal composition by characteristic low-variance REE abundance
patterns. With respect to chondritic abundances, Archaean and
post-Archaean sediments both show an overall greater abundance
of REE, with a relative enrichment of the lighter elements (La -
Sm). This enrichment is more marked in post-Archaean samples while
an additional, very distinctive feature of these is a negative
anomaly in Eu abundance. Taylor (1979) explains this as due to
Eu^{2+} substitution into phases preferentially retained in the
lower crust and thereby, he advocates contrasting processes of
crustal differentiation across the Archaean-Proterozoic boundary
- an assertion supported by numerous lithological and isotopic
features.

For example,McLennan and others (1979) and McLennan and
Taylor (1980) find this shift in REE patterns to be recorded
within the Lower Proterozoic Huronian succession of Canada and
the Pine Creek Geosyncline in Australia. This is interpreted as
reflecting an increased detrital input after progressive exhumation
of abundant Late Archaean granitic rocks that are known to
commonly   contain significant negative Eu-anomalies. A coincident
significant increase in the modelled radiogenic Sr content of sea
water occurs as well, attributed to large scale fractionation at
the Archaean-Proterozoic boundary (Veizer and Compston 1976).

These findings clearly have direct application to the
present study in that the major contention surrounding the origin
of Savo province metasediments is whether they derive from Archaean
basement or from penecontemporaneous lower Proterozoic sources.
That is, their age and setting make REE patterns a practical and
convenient parameter in discrimination and hence resolving the
problem. The value of results would be enhanced by corresponding
analyses from the adjacent Höytiäinen province whose derivation

from Archaean basement is not in dispute. Mixing of detrital
and authigenic components in equilibrium with ambient sea water
may alter abundances in detail, due to the secular trend towards
general REE enrichment, but should not modify the overall
morphology of the patterns. Unfortunately no REE data have yet
been published for North Karelian metasediments, though analysis
of a suite of whole-rock samples from the study area is currently
in progress.

5.5   ISOTOPIC STUDIES IN RELATION TO PROVENANCE

5.5.i   K-Ar and Rb-Sr systems

Since detrital sediments are potentially complex mixtures
of components of diverse age or origin, whole-rock isotopic
analysis is unlikely to yield a geologically meaningful result,
or if it does, this is likely to refer to a period of later
homogenization under metamorphic conditions. Even if the practical
difficulties of isolating pure, distinctive, cogenetic grain
populations are overcome, results obtained do not necessarily
correspond to either depositional or inherited source-rock ages.
Rather, they will clearly depend in part on the integrity of the
isotopic system and also on the post-depositional history of the
sample. Thus detrital muscovite may retain K-Ar source region
characteristics throughout greenschist facies metamorphism,
though this property decreases markedly in proportion to decreasing
grain size (J. Hunziker, spoken comm.). A corollary of this is
that K-Ar isotopic systems of clay-sized detritus may survive
undisturbed through diagenesis and at least early stages of
burial metamorphism, rather than effectively equilibrate with
ambient sea-water. Gebauer and Grünenfelder (1974) likewise
conclude that Rb-Sr whole-rock systems are also subject to post-
depositional isotopic exchange and unlikely to record depositional
ages. These authors also acknowledge that degree of homogenization
is size dependent, with coarse detrital micas being likely to
preserve predepositional isotopic features. Thus the final

resultant value may record a complex interaction between components having different age, Rb/Sr and initial $^{87}Sr/^{86}Sr$ and, which have not necessarily re-equilibrated to the same extent during post-depositional events. A further compounding problem with Lower Proterozoic sequences is that their deposition apparently corresponds to a period of rapid increase in $^{87}Sr/^{86}Sr$ (Veizer 1973) and this would tend to affect the reliability of model ages with respect to isotope evolution of sea water. Given these factors, Rb-Sr and K-Ar analyses were deemed inappropriate for this study.

5.5.ii U-Pb and Pb-Pb system

As an ubiquitous detrital accessory mineral, zircon has proven particularly useful in determining U-Pb ages of provenance area lithologies. Gebauer and Grünenfelder (1976, 1977) have however determined that old, highly metamict zircons may undergo substantial radiogenic lead loss when subjected to later meta-morphism; since isotopic fractionation is unlikely, it is never-theless possible that Pb-Pb ratios still preserve indications of a more ancient origin. Using this method, Vaasjoki (1981) has clearly distinguished between Archaean basement, Karelian meta-sediments and the Svecofennian lithologies of SW Finland. The Karelian data are based almost entirely on epigenetic vein samples and do not appear to significantly discriminate between particular regions: although the samples from the Outokumpu ore deposit are very distinctive in their low U and Th content, model ages fall within the diverse range encompassed by analyses from Höytiäinen province metapsammites. These latter are from 1.977 - 2.360 Ga for the Stacey-Kramers (two-stage) model or from 1.820 - 2.180 Ga for the Cummings-Richards (evolving U/Pb ratios) model; for both models an affinity with Archaean basement isotopic characteristics is evident, consistent with at least partial derivation of metasediments from this source.

The Outokumpu Pb-Pb model ages show, in contrast, less variability and according to Vaasjoki (1981), represent the primitive, older end-members of a distinctive Svecokarelian

suite of data. This evolutionary trend is recorded by a decrease
in model ages from the Outokumpu ore to Svecofennian deposits
and an increasing *Pb content and U/Pb ratio which produce a
coherent sub-linear array on various Pb-Pb diagrams. These
features are interpreted by Vaasjoki as indicating a primitive
component of mantle origin gradually modified by progressive
homogenization with mobilized lead of crustal origin. In
conclusion, Pb-Pb isotopic ratios and model ages are evidently
compatible with discrimination between the Savo and Höytiäinen
provinces, as proposed independently from other criteria.

Returning to discussion of provenance determinations based
on detrital zircon populations, only limited conclusions can
be drawn. A sample isolated from metasediments at Vuonos mine,
near Outokumpu has a Pb^{207}/Pb^{206} age of 2.245 Ga, which is
approximately mid-way between values for Archaean and Svecokarelian
intrusive rocks (Geological Survey of Finland 1976). Consideration
must be given to the possibility of this figure being other than
the primary age of zircon crystallization. The investigations
of Gebauer and Grünenfelder (1976, 1977) demonstrated that
discordant detrital zircons from Central Europe resolve an upper
concordia intercept between 1.7 and 2.0 Ga but a lower intercept
at 510 - 410 Ma. This is ascribed to substantial (80 - 95%) lead
loss during latest Precambrian (Cadomian) metamorphism, with
subsequent detrital reworking of the re-equilibrated zircons
during the early Palaeozoic. Apparently, after a considerable
period of time (at least 1.0 Ga?), highly metamict zircons are
susceptible to recrystallization and lead loss under greenschist
facies conditions. The degree of lattice bombardment must, however
depend upon U and Th content and where this is relatively low,
Archaean provenance ages have survived early-middle Proterozoic
metamorphism with zircon isotopic systems virtually undisturbed
(Page and others 1980). In North Karelia this was already
apparent in the pioneering work of Kouvo (1958), not indeed from

detrital zircon populations but from retention of Archaean U-Pb
characteristics in basement inliers while Rb-Sr whole-rock and
K-Ar systematics were reset during the Svecokarelian Orogeny.
These findings have been corroborated more recently by studies
near the basement contact in northern Sweden (Skiöld 1979b.

To supplement these data, zircons from several samples are
currently being analyzed at the Geological Survey of Finland, by
Hannu Huhma . One sample is of meta-arkose from the Kettämönniemi
lithofacies, which consists exclusively of Archaean granitoid
detritus (or of reworked sedimentary material with Archaean
basement as ultimate source). It is hoped that this result
might indicate to what extent original U-Pb characteristics have
survived Svecokarelian metamorphism.

The second sample is from the Sammallahti lithofacies near
the eastern margin of the Savo province and was selected because
of the relative abundance of zircon. These grains can be well-
zoned but with rounded apices, probably due to sedimentary
attrition rather than magmatic resorption. Archaean basement
as provenance is plausible but equivocal; the sediment is poorly
sorted and contains scattered granules of some felsic plutonic
or metasedimentary affinity. However it is probable that this
represents a local, marginal or proximal aspect of the Savo
province and an Archaean provenance need not be representative
of the major source region.

5.5.iii Nd-Sm system

Nd-Sm isotopic analyses provide yet another suitable
avenue for provenance investigations: there does not appear
to be any fractionation between these elements during sedimentary
processes (McCulloch and Wasserburg 1978) and systems are
evidently unperturbed by high-grade metamorphism. No data have
been published for Finnish metasediments but preliminary results
are available for samples from the Höytiäinen province, collected

with and analyzed by Hannu Huhma at the Geological Survey of
Finland. These are calculated model ages using Nd/Sm and Nd
isotopic ratios; analyses are whole-rock and the inability to
isolate fractions of differing Nd/Sm ratios made obtaining an
isochron impractical.

Sample	Lithofacies	Sm^{147}/Nd^{144}	T_{CHUR} (Ma)
S11A	Rauansalo	.1967	
S11B	Rauansalo	.1617	
S12	Tikkala	.1273	2329 ± 50
S13	Kukkupää	.1119	2169 ± 53
S14	Kukkupää	.1101	2156 ± 70
S15	Kettämönniemi	.1055	2322 ± 86

Discussion of samples and discussions of results:

(i) Sample S15 was the only coarse-grained metasediment
deliberately obtained. It is thereby hoped that it represented
detritus derived directly from an Archaean granitoid source (or
indirectly reworked via earlier Proterozoic sediments) and would
thus provide an indication of whether or not the isotopic system
has remianed inert, yielding provenance ages. Clast types and
proximity of basement enable an Archaean provenance to be
asserted with confidence, demanding an explanation for the younger
T_{CHUR} model age. Nothing definite can be established at this
stage but it is possible that modelling with respect to the mantle
trajectory (based on undepleted chondritic evolution), is not
appropriate. This might be particularly so if there is an
authigenic component to the sediment. Calcite is not abundant
and mica could equally be detrital; K-feldspar has been identified

as an interstitial, presumably authigenic phase and may thus
indicate mixing of several components. However since the residence
time of rare earth element in sea water is short, and REE content
in K-feldspar is small, it is more likely that the younger age
reflects   a distinct lower Proterozoic component, probably original-
ly in the form of detrital clay (M. McCulloch spoken comm.).

(ii) Sample S12 is from sulphide-bearing metapelite between
metagraywacke beds deposited in an inner fan environment developed
within the nascent Höytiäinen province. Abundant detrital grains
indicate an ultimate quartzo-feldspathic provenance, with abundant
intrabasinal clasts indicating penecontemporaneous reworking and
down-slope transport. Metapelites are believed to have originated
as mud and silt turbidites, or as the upper parts of graded
rudites and arenites, so that a detrital origin is considered
likely for mica and/or clay precursors. Thus it is not unexpected
that it yield a similar model age to that for the Kettämönniemi
lithofacies meta-arkose (S15), further suggesting that any
discrepency between this value and the true age is due to modelling.
The higher Sm^{147}/Nd^{144} ratio may be a result of an overall greater
abundance of REE in micaceous pelite as opposed to K-feldspar
arkose, particularly if it derived from a Jatulian source.

(iii) Sample S13 is from the Kukkupää lithofacies and is
a micaceous metapelite with sporadic laminae richer in fine-
grained quartz. Apart from a probable chemogenic origin for
these latter, it is uncertain whether phyllosilicates are of
terrigenous clastic or authigenic origin. If detrital, the source
need not have been exclusively Archaean as Jatulian metabasites
and others within the Höytiäinen province itself yield U-Pb zircon
ages broadly coeval with the estimated model range for this sample
(Pekkarinen 1979a, H.Huhma spoken comm.).

(iv) Sample S14 is also from the Kukkupää lithofacies, some
7 km away from the site of sample S13. It is likely that this
sample contains in part a calcareous chemogenic component and like

S13, the quartz-mica component comprising the bulk of the rock is not diagnostic of provenance. Nevertheless the samples are mutually consistent with respect to both T_{CHUR} model age and Sm^{147}/Nd^{147} ratio and clearly distinguishable from the samples of more obvious Archaean detrital origin (S12 and S15).

(v) Samples 11A and 11B indicate, by way of contrast, the heterogeneity that may exist in a single outcrop. Both samples are from hornblende-chlorite-bearing graded metagraywackes in which detrital feldspar and quartz are also evident. It was not possible to assign a model age but the high Sm^{147}/Nd^{144} ratios with respect to that of undepleted mantle (CHUR = 0.1967) indicate a short crustal residence time for the protolith. This supports the interpretation from lithological, tectonostratigraphic and petrological criteria that the Rauansalo lithofacies derives, at least in part, from the Tohmajärvi basic igneous complex. It is also consistent with Jatulian igneous rocks in general being derived from somewhat depleted mantle (H. Huhma spoken comm.).

5.6   SUMMARY OF PROVENANCE DEDUCTIONS

(i) The composition of metasediments, including recognition of detrital elements and the use of limited geochemical and iso-topic data, support the subdivision of the study area into the Savo and Höytiäinen provinces.

(ii) The Höytiäinen province contains sufficient relict detritus to identify provenance from Archaean granitoid basement and Proterozoic sediments (Tikkala, Kettämönniemi and Suhmura lithofacies) and locally from penecontemporaneous metabasites (Rauansalo lithofacies).

(iii) The Savo province records local, basal or marginal derivation from Archaean basement (Oravisalo quarry, Niinikkosaari, Sammallahti lithofacies). The bulk of the metasediments are

however, of uncertain provenance but of distinctive, biotite-
and plagioclase-rich character. They may derive from any combi-
nation of three sources - Archaean granitoid basement, igneous
Svecofennian lithologies or from the Lappland granulites to the
NE of Karelia, as intimated by Barbey and others (1984). REE and
additional isotopic data might permit resolution of these problems.
provenance ages in excess of 1.92 Ga would effectively exclude
the possibility of a Svecofennian contribution to Savo province
metasediment.

(iv) The limited Nd-Sm data available appear to affirm
the value of this isotopic system in examining provenance of
metasediments.

FIGURE 42:   RELICT AND RECRYSTALLIZED FABRICS

Scale bar represents 1 mm

a   Graded massive silt with relict rounded and angular detrital
        grains recording negligible recrystallization or growth;
        uppermost dark band is graphite-rich.   Underlying laminated
        metapelite likewise preserves depositional fabric, while
        showing thin layer-parallel veinlets and at left, pseudomorphs,
        possibly after staurolite (Onkamo 04.6   17.4). PPL.

b   Laminated depositional units are preserved, defined by graphite-
        rich laminae alternating with calcareous metapelite.   Meta-
        morphic growth of biotite, quartz and calcite has however
        obliterated individual detrital grain boundaries
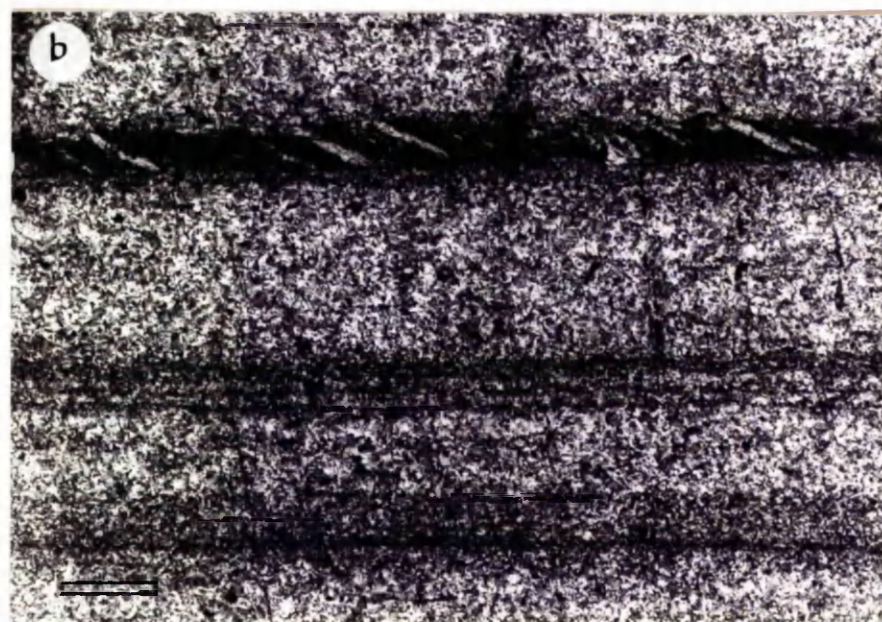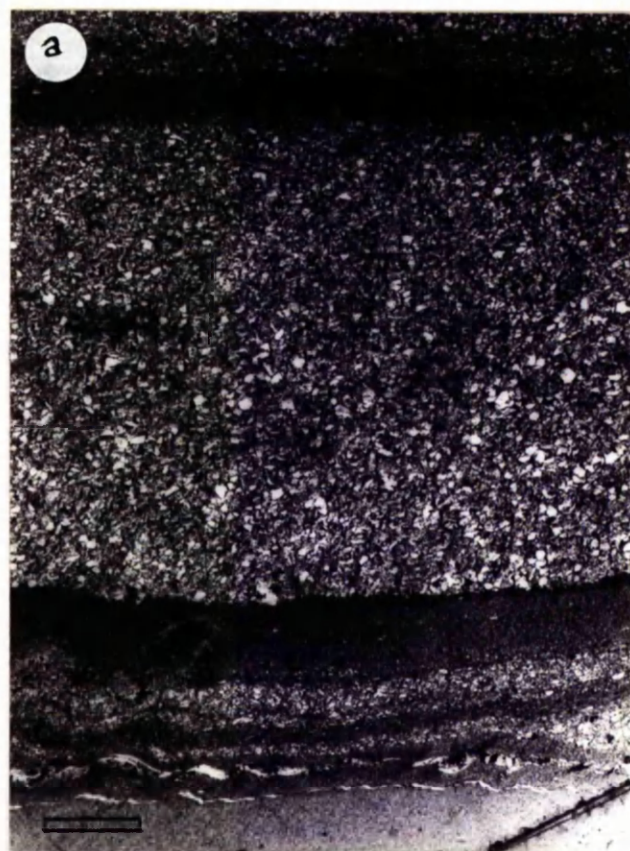   (Niittylahti 97.7   30.1). PPL.

FIGURE 43: RELICT DETRITAL MINERALOGY AND GRAIN SHAPE

Scale bar represents 1 mm

a   Deformed, matrix-poor metapsammite with microcline grains showing
     orthogonal extinction lamellae and at bottom left multiply-twinned
     plagioclase grain (Kankaala lithofacies, Oravisalo 75.3 15.2). NX.

b   Large grain of polycrystalline quartz in Sammallahti lithofacies
     metapsammite (Pyhäselkä 91.7 20.0). NX.

c   Sutured and stylolitic boundaries between detrital grains of
     monocrystalline quartz. The relative absence of matrix is typical
     of well-sorted Jatulian orthoquartzites (Viesimo 15.8 25.6). NX.

d   Well-rounded detrital grains consisting of coarser-grained quartz
     of probable granitic derivation and finer-grained polycrystalline
     aggregates of probable chert or orthoquartzite origin. Boundaries
     are well-defined by calcite (mottled grey) and incipient
     recrystallization at grain margins (Tikkala lithofacies,
     Niittylahti 98.2 35.8). NX.

e   Large rounded quartz grain in finer-grained poorly sorted meta-
   turbidite (Tikkala lithofacies, Onkamo 04.6 17.4). NX.

f   Subangular detrital grain outlines preserved in fine-grained
     quartzose metaturbidite with larger granite rock fragment at
     left and feldspar and quartz grains at right (Kettämönniemi
     lithofacies, Pyhäselkä 94.6 27.3). NX.

g   Polygonally annealed polycrystalline quartz enclosed by
     chalcopyrite. Chlorite portion at top indicates original detrital
     nature of clast, brecciated during ore mobilization (Tikkala
     lithofacies, Hammaslahti mine). NX.

h   Detrital quartz grain of probable granitic or reworked
     sedimentary origin, enclosed in a hornblende-chlorite matrix
     derived from mafic volcaniclastic sediments (Rauansalo lithofacies,
     Rauansalo 04.8 22.5). NX.
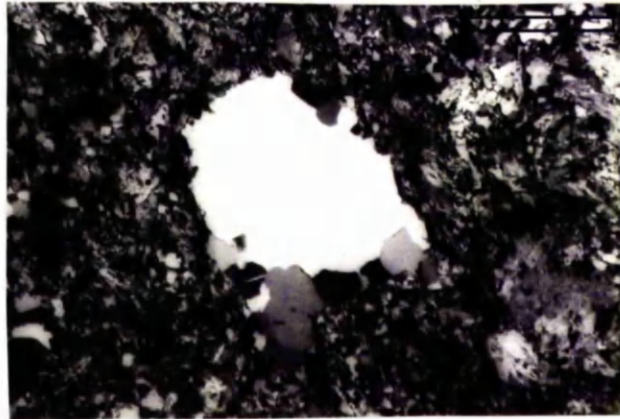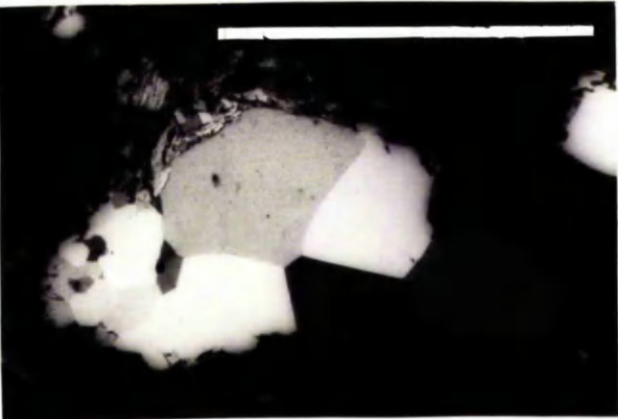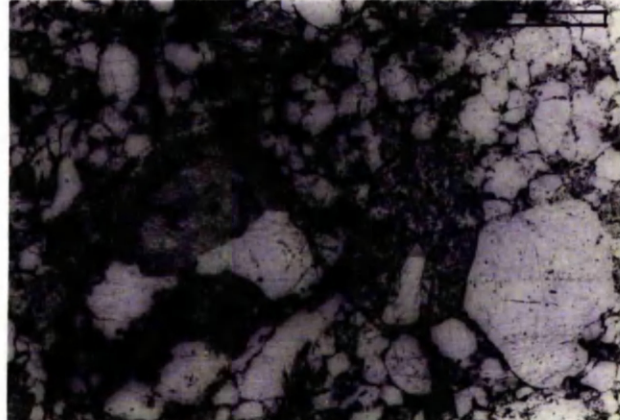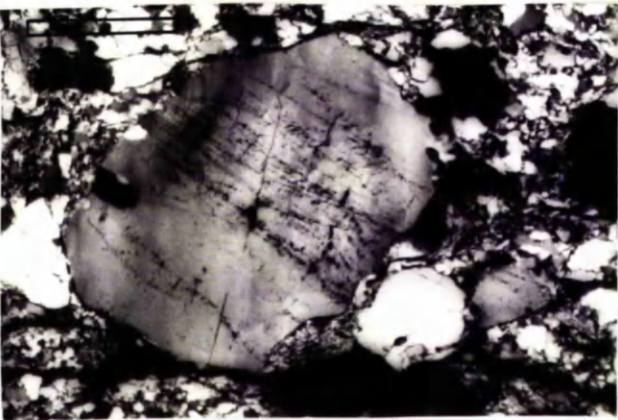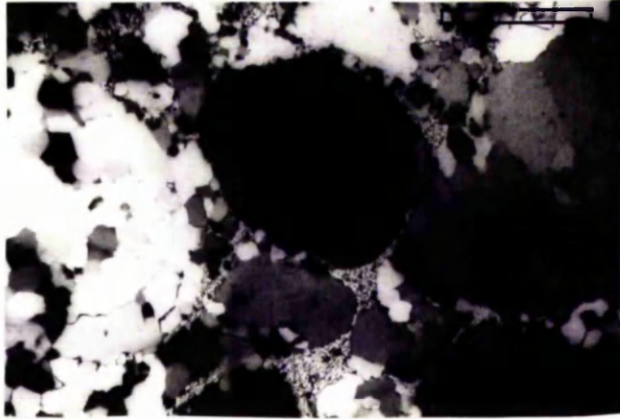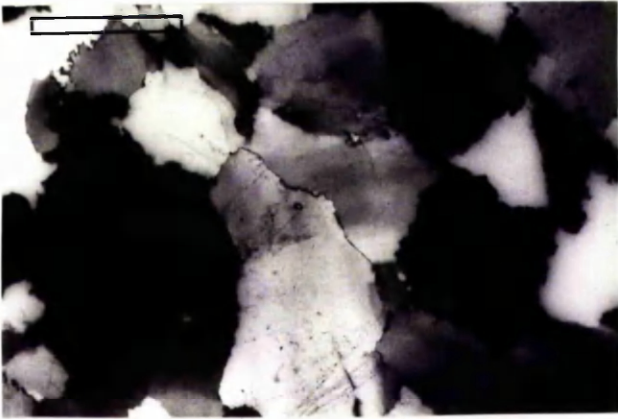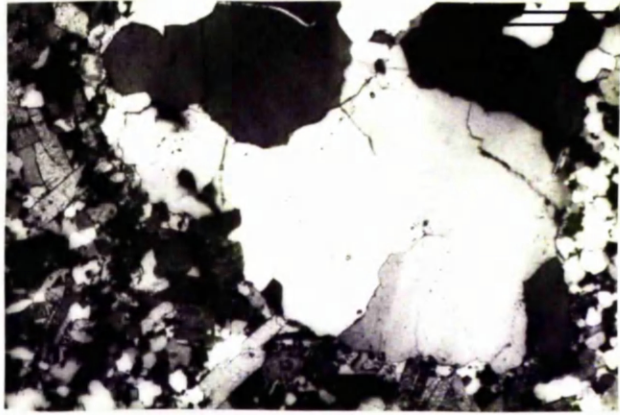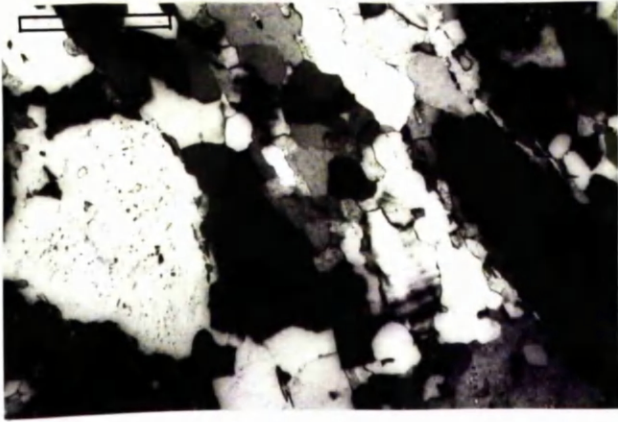
6.   INDICATIONS OF PALAEOSLOPE AND BASIN EXTENT

6.1   INTRODUCTION

At individual exposures, evidence for palaeocurrent activity may be recorded by cross-bedding, or directed scours or channels and trend of palaeoslope may be indicated by synsedimentary slumps. Only a few examples have been found in the study area - certainly too few to enable an adequate reconstruction of basin morphology, especially in view of the complex tectonic deformation. At a larger scale, transgression across underlying units, or local erosion of formerly contiguous, widespread formations may provide information concerning palaeotopography. This kind of analysis is also restricted by complex structure and lack of certainty in stratigraphical correlation.

6.2   OUTCROP SCALE PALAEOSLOPE INDICATORS

6.2.i Savo province

Only one exposure displayed clear evidence of directional sedimentary structures, within the Rääkkylä lithofacies at Nivansalo (Pyhäselkä 94.20.2). These are evident in Figure 25 as probable climbing ripples in three successive beds indicating an eastwards component of flow (right to left). An accurate current vector cannot be determined as the angular relationship betwen dune crests and the plane of exposure is unknown. Moreover, restoration to original attitude is difficult because the outcrop is apparently located in the rotated hinge of a major

F_2 or F_3 fold structure. Since this locality cannot be considered in relation to any other outcrops in the Savo province, conclusions concerning palaeoslope and regional sediment dispersal are precluded.

6.2.ii Höytiäinen province

Several widely scattered localities indicate a northerly component to the palaeoslope, which is consistent with sediment dispersal along the axis of the depositional basin. At Tikkala railway siding (Onkamo 04.6 17.4) (Figure 35), a large, rounded boulder in a presumed debris-flow deposit appears to show thicker beds on its left hand ("upslope" side than to the right, where several beds are suggestive of foreset accumulations "in front of" the obstructing boulder. These beds are on the upper limb of an F_3 structure, according to younging criteria and cleavage vergence, subsequently reoriented into a vertical Group 5 shear zone; assessing the likeliest deformation path as having been by translation and rotation, an original northwards component to palaeocurrent flow seems tenable, upon after restoring to the horizontal.

Metagraywackes on a more gently dipping F_3 upper limb at Lentosärkkä (Pyhäselkä 94.6 27.3, Figure 32a-d, map 3) also appear to record a northerly dipping palaeoslope, by the existence of a coherent slump fault, of normal sense (as opposed to the overthrust sense of tectonic structures). This is demonstrably not of tectonic origin, as it has caused a plastic deformation of adjacent beds and is itself transected by S_2 and S_3 foliations. At a somewhat lower stratigraphical level within the Kettämönniemi lithofacies (Pyhäselkä 94.4 28.4 , Figure 30c,d, map 3), a possible imbricate orientation of large granitoid clasts is indicative of a similar palaeoslope trend.

A minor discordance in mafic Rauansalo lithofacies meta-
graywackes (Figure 39a,c) likewise suggests slumping in a
northerly direction and a similar direction is indicated by
current lamination in some graded interbeds. Elsewhere, thin-
bedded turbidites or the upper parts of thicker units sporadically
show current lamination but only rarely are these unambiguous.

6.3   REGIONAL INDICATORS OF BASIN GEOMETRY

6.3.i   Savo province

Only a segment of the eastern margin is present in the
study area and even this is a major tectonic dislocation, along
the Kettämö thrust zone. Nevertheless, it is evident that
Pyhäselkä assemblage metasediments were deposited upon Archaean
basement or locally upon intervening sequences including meta-
basites and non-detrital sediments such as those at Oravisalo.
Consistent with this is the preservation of unconformities and
coarse, basement-derived detritus, at Oravisalo, Kettämönniemi
and N of the study area at Niinikkosaari and Sotkuma, all
indicating widespread exposure of Archaean basement during
deposition. If any earlier Proterozoic deposits existed on
this basement, they were subsequently removed - a viewpoint
supported by the absence of any such strata at the basement
unconformity, in spite of the presence of orthoquartzite clasts
amongst granitoid detritus in the Niinikkosaari and Kettämönniemi
lithofacies. Farther to the N however, between Kaavi and Koli,
Jatulian orthoquartzites and metabasites are extensively
preserved, and would have effectively prevented erosional access
to Archaean basement in this region (Map 1).

The palaeotopographic implications of this differential preservation of cover formations are not straightforward. It seems logical to explain it as due to blanketing of the sequence in the north, deriving sediment from basement to the south, now exposed only as inliers at Oravisalo, Kettämönniemi and Sotkuma. However, this scenario ignores the possibility that cover sequences were not uniformly developed throughout the region, and the fact that in the absence of isopach estimates for the Savo province deposits, uncertainty exists as to the location of the main basin depocenter. Indeed rather than to the north, the present, tectonically modified relationships suggest a thicker development towards the south-west, as far as Savonlinna where distinctly different lithologies and tectonic styles are encountered (Gaál and Rauhamäki 1971).

Furthermore it need not be assumed that the source terrane was necessarily adjacent to the depositional basin - as long as a major river system existed, detritus could have been transported a great distance from other areas subject to uplift and denudation. The high, presumably primary, plagioclase content of much of the Pyhäselkä assemblage does however constrain estimates of the amount of transport and weathering of detritus so that a source further than the Lappland granulite belt to the northeast of Karelia, or the Svecofennian terranes to the southwest is deemed improbable.

Only the Niittylahdenrantatie lithofacies seems to preserve evidence of a major submarine distributary system, near the present NE margin of the province. There is unfortunately no evidence of how far this has been displaced tectonically, nor any directional data to indicate sediment dispersal trends.

6.3.ii Regional indications of sediment dispersal within the Höytiäinen province

In spite of its present western boundary being delineated

by the Kettämö thrust zone, there is little doubt that this
approximates to the original basin margin, thus clearly defining
a linear depositional feature flanked to either side by Archaean
basement. The eastern margin preserves an unconformity, merely
tilted, between Kiihtelysvaara and the Soviet border (Pekkarinen
1979a) but the southwards continuation of the province, through
Jänisjärvi and Soanlahti to Ladoga shows evidence of overthrusted
relationships (Wegmann 1929, Hausen 1930). Likewise to the N,
overthrusting is prominent in the vicinity of Koli (Väyrynen
1939, Gaál 1964) though unconformable relationships are also
well-exposed (J. Marmo, spoken comm.). From inspection of Map 1
it is evident that the eastern and western margins of the province
converge northwards, inferentially into a zone of major tectonic
significance which may be contiguous with structural features
to the N, in the Kainuu metasedimentary sequence (Laajoki 1984).

At this juncture it is necessary to provide some more
precise stratigraphic meaning to the definition of the eastern
margin of the province. The diverse (Sariolan and Jatulian)
sequence consisting of meta-arkose, orthoquartzite, metabasite
and volcanogenic-chemogenic sediments (Figure 41, Map 5) is
separated by a low-angle discordance from an overlying thin,
but persistent conglomerate horizon containing clasts representing
all these lithologies (Pekkarinen 1979). This stratigraphical
sequence is consistently observed between Värtsilä on the Soviet
border and Kiihtelysvaara, and additionally in Soviet Karelia,
at Soanlahti (Hausen 1930), as well as northwards in the
Kontiolahti-Koli district (J. Marmo, spoken comm.). The discordance
at the base of this conglomerate is taken to represent both the
temporal and eastern geographical limits to Höytiäinen province
sedimentation, even though it may be diachronous. Pekkarinen
(1979a) records that units stratigraphically underlying this
discordance successively wedge out along strike so that N of
Kiihtelysvaara, metabasites and the overlying carbonate-black

slate suite are absent (Maps 1 and 2). This might be explained by southwards tilting of the land surface, with correspondingly greater erosion in the north or alternatively, that the younger sequences were originally thinner or absent in the north. In either case a depocenter in the south would be indicated for the initial stages of Höytiäinen province sedimentation.

Within the lowermost Jatulian orthoquartzites, prevailing sediment dispersal seems to have been towards the ENE (Ojakangas 1965, Heiskanen 1980, Sokolov 1980), an inference supported by the preservation of thicker pre-Jatulian arkose and glaciogenic deposits at Kontiolahti than at Kiihtelysvaara (J. Marmo, spoken comm.). However, these conclusions need not preclude sediment dispersal into the Höytiäinen province having an entirely distinct trend, induced by tectonic events that disrupted the earlier depositional patterns.

The evidence already presented for provenance of the Rauansalo lithofacies (Section 5.3) provides the only other indication of regional palaeoslope within the Höytiäinen province. From this, it would appear that sediment was dispersed in a northerly direction from the flanks of an intrabasinal volcanic edifice, now in part represented by the Tohmajärvi igneous complex (Figure 41).

7.   STRATIGRAPHICAL CORRELATIONS AND CONSIDERATIONS

7.1   GENERAL COMMENTS

Proterozoic platform cover is widely distributed across the Archaean craton of the eastern Baltic Shield. Correlation of geographically separated sequences is impeded by a number of factors, of which the following seem relevant to Karelia.

i) Lithological boundaries may be quite diachronous.

ii) Certain facies may be very impersistent, with limited lateral extent.

iii) Correlation using erosional or tectonic hiatuses and discordances, or palaeosol horizons offers no indication of the duration of the hiatus, unless lithologies both above and below the boundary are amenable to isotopic age determinations.

iv) A progressively maturing platform sequence will tend to involve partial or complete reworking of earlier deposits; certain formations may thus be absent in some areas, due to erosion, while elsewhere they are present as outliers difficult to correlate.

v) Subsequent erosion and landscape evolution, in removing thinner units or those at higher elevations, may also have removed evidence of lateral facies transitions vital to correlation.

vi) Failure to recognize tectonic effects, such as allochthonous thrust sheets, in areas of poor exposure, may lead to undetected duplication of stratigraphic sequences.

Most published work has recognized at least some of these
limitations and few recent studies actually contradict earlier
classification schemes. Important, thorough documentation of
lithological types and their distribution was first made by
Frosterus and Wilkman (1920), Metzger(1924), Väyrynen (1928,1939)
and Hausen (1930), with the presently accepted stratigraphical
sequence being first recognized by Väyrynen (1933). From a
more conceptual basis, correlation of different lithologies as
contemporaneous facies variants was attempted as early as 1924
by Metzger, at Suojärvi in Soviet Karelia, and Wegmann (1928)
interpreted different lithological groups as representing
differing tectonic environments, according to Alpine tectonic
paradigms.

· As a result of these three, essentially different approaches,
the nomenclature of lower Proterozoic sediments in Karelia has
variously acquired lithofacial, tectonostratigraphic or precise
stratigraphical connotations. Due to increasingly detailed
field investigations, and isotopic dating of igneous clasts or
intercalated metalavas, the strictly lithostratigraphical
approach is nowadays more feasible and particular reference is
made to the studies of Pekkarinen (1979a) and reviews by Sokolov
(1980) and Laajoki (1983). Principal stratigraphical units
adopted from these works are presented in Figure 44.

7.2   REGIONAL STRATIGRAPHICAL CORRELATION

The Proterozoic of Karelia has long beensubdivided into
three major units, each containing characteristic lithologies
(Eskola 1925, Väyrynen 1928). From oldest to youngest, these are
the Sariolan, Jatulian and Kalevian sequences; the Sariolan is
better developed in Soviet Karelia but there the Kalevian is
not recognized or at least is represented by different facies
(Heiskanen 1980, Sokolov 1980).

7.2.i Sariolan sequence

Sediments typically contain detritus readily identifiable as of Archaean basement provenance but a component from associated mafic to felsic volcanic rocks is also present. Volcanic intercalations themselves are most abundant in Soviet Karelia but metalavas are also present in the Kurkikylä Group in adjacent parts of Finland (Laajoki 1983). Nearer to the study area Sariolan metapsephites and meta-arkoses constitute an impersistent unit overlying Archaean basement at Kiihtelysvaara (Pekkarinen 1979a) and a more substantial and diverse sequence, including glacigenic deposits, at Kontiolahti (Marmo and Ojakangas 1983). Pekkarinen (1979a) has presented evidence that these sediments are younger than 2.34 Ga quartz veins.

7.2.ii Jatulian sequence

Throughout Finnish Karelia, Sariola-type deposits are invariably overlain, in some places disconformably, by more mature orthoquartzites which can form homogeneous sequences exceeding 500 m in thickness. Abundant metadolerite dyke swarms and sills intrude these quartzites, while metabasite lavas and tuffs are locally intercalated, as at Kiihtelysvaara (Pekkarinen 1979a). Further to the east, in Soviet Karelia, the volcanogenic component is increasingly prevalent and several magmatic and depositional cycles recognized on the Karelian craton by Sokolov (1980) and Svetov (1980) presumably correlate with major igneous activity within the then nascent Belomorian fold belt (Barbey and others, 1984). Finnish Karelia may thus record a condensed, possibly terrestrial lower Jatulian sequence consistent with aeolian and fluviatile environments deduced from the orthoquartzites by various workers (Metzger 1924, Hausen 1930, Ojakangas 1965, Pekkarinen 1979, Heiskanen 1980).

This terrigenous sequence is succeeded conformably almost everywhere by diverse, but equally distinctive lithologies, mostly of volcanogenic and chemogenic origin and originally classified as 'marine Jatulian'. Regardless of depositional environment they are characterized by massive and laminated metadolomites (including stromatolites), carbonaceous metapelites and oxide-, carbonate- and sulphide-bearing banded iron formations. Metatuffs are difficult to distinguish from other metapelites but their presence is to be expected, given continued evidence of continued mafic igneous activity in the form of metagabbro and metadolerite sills (Metzger 1924, Hausen 1930, Pekkarinen 1979a, K. Heiskanen spoken comm.).

The Jatulian sequence at Kiihtelysvaara has been thoroughly documented sedimentologically and geochemically by Pekkarinen (1979a) and is valuable as a representative reference section. Metabasites dykes in Jatulian quartzites have yielded isotopic ages between 2.13 Ga and 2.02 Ga and several Pb-Pb studies of metadolomites and iron formations give ages around 2.05 and 2.02 for later Jatulian sedimentation (Sakko and Laajoki 1975, Laajoki and Saikkonen 1977, Pekkarinen 1979).

7.2.iii Kalevian sequences

The eastern margin of the Höytiäinen Province as defined in this study corresponds to the basal Kalevian sequence at Kiihtelysvaara, in the sense of Pekkarinen (1979a). There, as through much of eastern Finland, the Jatulian is disconformably overlain by metagraywackes (many of demonstrable turbidite origin), commonly including a basal metasephite containing clasts derived in particular from the underlying orthoquartzites and metabasites (Figure 38 d,e).

Although volumetrically abundant in Finnish Karelia the
Kalevian has received little attention, mostly because of its
complex structure, dearth of sedimentary structures and meta-
morphic recrystallization. Väyrynen (1933) first demonstrated
depositional relationships to the Jatulian sequences and in
1939 also recognized two major lithological divisions corresponding
broadly to the Savo province metapsammites and Höytiäinen province
metapelites of this study. Nevertheless the differences were
ascribed principally to an increase in metamorphic grade towards
the southwest - an observation which is in general, correct.
The serpentinites, metabasites, metadolomites and other lithologies
of the Outokumpu assemblage have long been regarded as older
than the Kalevian metapsammites but formal correlation with the
Jatulian has not been attempted, in part due to their allochthonous
nature (Väyrynen 1939, Koistinen 1981).

The Kalevian is not so prevalent in Soviet Karelia but the
sequence described at Soanlahti and Jänisjärvi (Hausen 1930) is
likely to be comparable and contiguous with those from the
Höytiäinen province. In particular the Partanen conglomerate
Hausen describes is very similar in position and clast content
to the basal Kalevian conglomerate at Kiihtelysvaara. The
"Ladoga series" described by Hackman (1933) and Zaonezhye-
Suisarian sequence of Sokolov (1980) and Heiskanen (1980) also
apparently correlate, at least in part, with the Kalevian of
Finnish Karelia. However the Suisarian contains a pyroclastic,
and picritic volcanic component (Sokolov 1980) which differs
from the Kalevian of Finland, as does the overlying Vepsian
coarse-clastic sequence which may thus represent the youngest
Svecokarelian rocks preserved in Karelia (Figure 44).

7.2.iv Other lower Proterozoic sequences

None of the Svecofennian units defined inthe classical
studies of Sederholm (1897) or more recently (Kähkönen and

Laitakari 1983) are germane to the present study. However,
the area between Savonlinna and Varkaus, studied by Gaál and
Rauhamäki (1971) may contain both Svecofennian and Karelian
aspects; on Map 1, this area has been informally referred to
as the Haukivesi province. Pre-orogenic, supracrustal lithologies
include 'veined gneisses' that correlate well in composition
and general appearance with the less metamorphosed Kalevian meta-
psammites in North Karelia (Gaál and Rauhamäki 1971, Gaál and
others 1975). These appear to intercalate with diopside
amphibolites but are in tectonic juxtaposition with cordierite-
bearing metaturbidites and metavolcanic units south of Savonlinna.
Some indication of an unconformable stratigraphical relationship
may however be recorded by the Savonlinna conglomerate (Gaál
and Rauhamäki 1971).

7.3   STRATIGRAPHY OF THE STUDY AREA AND ITS REGIONAL CONTEXT

No individual units have been correlated across the boundary
between the Höytiäinen and Savo provinces, even though both
provinces contain Jatulian orthoquartzite detritus and may there-
fore be broadly coeval. Instead, the Oravisalo and Pyhäselkä
assemblages of the Savo province and the Kettämö and Hammaslahti
assemblages of the Höytiäinen province are considered to represent
two distinct stratigraphical sequences (Figures 41 and 44). The
difficulties outlined in Section 7.1 preclude all but the most
rudimentary interpretation of the Savo province, in which the
metapsammites and local metapsephites of the Pyhäselkä assemblage
were deposited on the mafic metapelites and metabasites of the
Oravisalo assemblage, and indeed the Outokumpu assemblage
(Koistinen 1981).

More evidence is available for vertical and lateral litho-
facies variations in the Höytiäinen province, though interpreta-
tions depend to some extent on the validity of correlating the

resedimented coarse clastic deposits of the Kettämö and Tikkala
lithofacies with the basal Kalevian conglomerates disconformably
overlying the 'marine Jatulian' Kiihtelysvaara carbonate-volcanite-
carbonaceous slate suite defined by Pekkarinen (1979a). This
assumption seems reasonable if all the coarse deposits are
regarded as a response to external tectonic influences and implies
that the Kukkupää and Mulo lithofacies are in the strictest sense,
'marine Jatulian' correlatives since evidence was presented in
Sections 4 and 5 that these are overlain by the Tikkala and
Suhmura lithofacies. Although a depositional hiatus is inferred
between the uppermost Jatulian and basal Kalevian conglomerate
at Kiihtelysvaara (Pekkarinen 1979a), this can result from trans-
gression at the basin margin while deposition has occurred con-
tinuously in the more distal parts of the basin. Hence, a well-
defined, disconformable boundary between the Jatulian and Kalevian
may not everywhere be present.

Finally, the Tohmajärvi volcanic complex is classified here
as a local facies developed in the early stages of Höytiäinen
province sedimentation but persisting until deposition of the
coarse clastic, basement-derived Tikkala lithofacies (Figures 41
and 44).

FIGURE 44 : STRATIGRAPHICAL RELATIONSHIPS

Major lithostratigraphical units of the study area shown in
relation to the sequence at Kiihtelysvaara (Pekkarinen 1979) and
adjacent Soviet Karelia (Sokolov 1980). Correlation has not
been attempted between the Savo and Höytiäinen provinces, which
consist of sequences traditionally classified as Kalevian
(Väyrynen 1939, 1954; Pekkarinen 1979). Early deposits in the
Höytiäinen province may also correspond in part to the 'marine'
or 'upper' Jatulian of earlier workers but lack a distinct
disconformity with the overlying basal Kalevian deposits. It is
suggested that the Vepsian and Suisarian deposits in Soviet
Karelia are younger than, if not Kalevian lateral facies equivalents.

SAVO PROVINCE

HÖYTIÄINEN PROVINCE

Rääkkylä lithofacies metapsammites

Niittylahdenranta the and Sammallahti lithofacies coarse clastic deposits

Resedimented coarse clastic deposits of Kettämö and Tikkala lithofacies and Kiihtelysvaara turbidite suite

'Marine Jatulian' volcanite, phyllite and dolomite

Vepsian and Suisarian coarse clastic, pyroclastic and porphyritic picrite lithologies in Soviet Karelia

Volcanites, serpentinites and volcanogenic and chemogenic deposits of Outokumpu and Oravisalo assemblages

Volcanoclastic Rauansalo lithofacies

Tohmajärvi igneous complex

Sariolan regoliths, glacigenic and terrestrial deposits

Jatulian mafic metavolcanite and metadolerite

Jatulian deltaic, and terrestrial orthoquartzite

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PART IV : APPLICATION OF RESULTS TO
REGIONAL FINNISH GEOLOGY

1. THE 'PROBLEM' WITH MANTLED GNEISS DOMES

    1.1 The concept of mantled gneiss domes

    Eskola's (1939) concept of basement involvement during
    deformation of cover sequences has received renewed attention and
    comment in the past decade, with several publications discussing
    mantled gneiss domes in some of the type areas of eastern Finland.
    Based upon experimental results modelling gravitational instability
    and diapiric responses (Ramberg 1967) and field observations (such
    as Stephansson 1975, Stephansson and Johnson 1976), Brûn (1980) and
    Brûn and others (1976, 1981) have advocated a diapiric origin to
    account for the morphology of mantled gneiss domes at Kuopio, in
    central Finland some 150 km north-west of the present study area.
    This interpretation was disputed by Park (1981) in presenting
    evidence that basement inliers in Karelia occur as culminations
    of interfering or non-cylindrical folds, as postulated by Hobbs,
    Means and Williams (1976); the mantled gneiss dome hypothesis was
    thereby regarded as obsolescent, if not invalid.

    Both these viewpoints embellish Eskola's original hypothesis
    to some extent, in that he did not formally propose a mechanism
    of gravity-driven diapirism but rather a general mobility induced
    by ultrametamorphism of basement granitoids. It is moreover clear
    from his summary figure 7 that although a ductile deformation of
    the basement-cover interface was implied (including foliation
    development in the basement), partial melting did not take place
    at this crustal level during Svecokarelian orogenesis in eastern
    Finland. Thus, apart from ascribing the bulk of basement fabric
    development to Proterozoic mobilization, Eskola's proposals are,
    at a descriptive level, consistent with, if less comprehensive than
    the results of subsequent studies that identify a polyphase deforma-
    tional history (such as Preston 1954, Bowes 1976, Brûn and other
    1976, 1981, Koistinen 1981, Park and Bowes 1983). The origin and

development of mantled gneiss domes remains however, contentious, and justifiably so.

1.2   Interpretation of mantled gneiss domes

Both gravitational instability and fold interference demand ductile deformation of the basement-cover interface and both Brûn and others (1976, 1981) and Park (1981) recognize that doming in Karelia was preceded by the development of two subparallel foliations and associated asymmetric folds whose enveloping surfaces are parallel to the basement-cover interface. These features are typically associated with sub-solidus diapirism (Stephansson and Johnson 1976), but do not in this case appear to be diagnostic indicators of origin.

Of considerably greater value is an assessment of the nature of synkinematic mineral lineations. According to Stephansson and Johnson (1976), radial principal elongation lineations characterize diapiric gneiss domes. It is however, patently obvious from the Karelian data compiled by Koistinen (1981, figure 9) that the dominant mineral lineation is not radially disposed about the Sotkuma and Maarianvaara domes, although it demonstrably pre-dates them. This is also the situation with regard to L_3 in the Oravisalo basement inlier in the present study area although in this case, Archaean rocks are clearly tectonically intercalated within the Proterozoic cover. Similar examples of thrusted basement slices described from the Kaavi and Koli districts (Frosterus and Wilkman 1920, Väyrynen 1939, Gaál 1964) and at Kotalahti (Gaál 1980) can be excluded from consideration as mantled gneiss domes on the grounds of their allochthanous nature.

Interpretation of the Kuopio domes remains problematical in that if basement is autochthonous, it is difficult to explain the frequently steep dips of the basement-cover interface by buckle

folding during crustal shortening. That is, unless some decollément
is identified, above which allochtonous basement and cover were
folded together, gravity-induced deformation of the interface as
evisaged by Brûn (1980) and Brûn and others (1976, 1981) is
potentially viable as an explanation of the present geometry. The
feasibility of diapirism may be assessed with respect to the rheology
of the rock units, and external boundary conditions such as depth
to the basement-cover interface and the nature of any imposed non-
hydrostatic stresses. At this stage, none of these parameters can
be realistically quantified or modelled but experimental and
theoretical models (Biot and Odé 1965, Ramberg 1972, Marsh 1979)
suggest that amplification of diapirs is promoted by

(i) increasing density contrast between overburden and source
layer,

(ii) decreasing viscosity of source layer, and

(iii) tensional stress regimes.

An example in which each of these factors has been favourable to
diapiric intrusion is given by Ollier and Pain (1980) who described
actively rising gneiss domes in southeastern Papua New Guinea. Rapid
uprise of material having considerably lower viscosity and density
than its overburden is indicated by Pliocene ages for granitoids
intruding dome cores and the presence of the overlying Papuan Ultra-
mafic Belt,while development in a tensional regime is indicated by
the proximity of Cainozoic sea-floor spreading zones, the axis of
the Papuan mountain chain and major transcurrent faults.

In contrast, evidence of partial melting or intrusion specially
into dome cores is lacking at the present erosion level in the
Kuopio district and moreover, it seems unlikely that density con-
trasts would have been substantial if Proterozoic sediments

represent reworked Archaean basement. Neither is there any direct
evidence for the Kuopio domes having developed in a tensional
setting - the strain analyses of Brûn and others (1981) indicate
extension only locally and in a vertical direction while Preston's
(1954) study indicates that the Suvasvesi wrench fault (correlated
with Group 4 structures of this study and interpreted as a response
to N-S horizontal compression), transects some of the domes. Thus,
it is still uncertain whether rheological properties and external
conditions were conducive to diapiric emplacement in the Kuopio
district. On the other hand, no quantitative constraints can
yet be applied, nor does interpretation of the domes simply as fold
interference patterns explain the underlying causes of deformation.

Limited evidence from the present study area nevertheless
suggests a rigid-block response to gravitational instability,
namely the occurrence of major Group 5 structures as ductile normal
faults. A corollary of such inferred uplift is that the basement
was in fact less dense than the overlying sedimentary rocks or else
that basement was unroofed by rapid erosion of the overthrust cover
units, with uplift accommodated along reactivated Archaean features.

2. REGIONAL TECTONIC PARADIGMS

Consideration of depositional and deformational events within the study area is independent from any interpretations of their regional tectonic context. Furthermore, the conclusions drawn do not diagnostically indicate a specific tectonic setting nor are they of inherent value in demonstrating or refuting the existence of particular environments and processes in the early Proterozoic. Nevertheless, they are of relevance in placing some constraints on existing and conceivable hypotheses concerning the tectonic development of Finland. All modern models (Hietanen 1975, Berthelson 1980, Campbell 1980, Gaál 1982a, Edelman and Jaanus-Järkkälä 1983, Park and others 1984) explicitly invoke plate-tectonic processes, including both the Wilson cycle and lateral fore-arc accretion as presently understood, so that this uniformitarian application of the plate-tectonic hypothesis will be examined first. Then follows a consideration of alternatives or modifications, which can be viewed as complementary, rather than simply antagonistic to plate-tectonic concepts.

2.1 UNIFORMITARIAN PLATE TECTONIC HYPOTHESES

2.1.i Cordilleran paradigm

Plate-tectonic hypotheses have been based on both petrogenetic and structural/geometrical analogies with extant or Phanerozoic mobile zones. Hietanen (1975) first drew attention to parallels between the evolution of magmatism in the Sierra Nevada and the Svecofennian of south-west Finland and accordingly suggested that the latter sequence formed in an island arc separated from the Karelian craton by an inter-arc or back-arc basin. Such an interpretation is logical and consistent with the available data but does not in itself demonstrate the existence of either island arcs or subduction beneath the Archaean craton. That is, the conclusions are based on temporal, rather than

coeval spatial relationships, whereas such magmatic evolution might conceivably be the inevitable result of various different processes. So far, the only other petrogenetically based deductions are those of Mäkelä (1980) who attempted to classify Finnish ore deposits in terms of marginal basin and volcanic arc environments.

Subsequent hypotheses have elaborated upon Hietanen's (1975) model, progressively introducing concepts such as accretion of several arcs and migration of subduction zones (Berthelson 1980), and integrating models with Finnish stratigraphy to a greater degree (Gaál 1982a, Edelman and Jaanus-Järkkälä 1983). Park and others (1984) have also accepted the North American Cordilleran analogue with additional emphasis on oblique collision and transcurrent displacement along the craton margin, post-dating northwards subduction. This latter is inferred from a supposed southward younging of 'arc terranes'; comparison of the limited published isotopic data in structural context (see Gaál and Rauhamäki 1971, Hopgood and others 1983, Korsman and others 1984) does suggest, however, the same order of variation along these arcuate zones as across them, so that such a proposal should be examined only after a more rigorous analysis of precisely constrained results. It is also imperative to establish polarity of proposed volcanic arcs by recognition of fore-arc stratigraphy and post-arc sequences in the manner of Dickinson and Seely (1979) or Crook and Feary (1982), rather than merely to note the presence of lithologies similar to those in contemporary arc environments.

The Cordilleran paradigm nevertheless remains a logical and attractive one although it should be noted that subduction of the Pacific plate was originally nearly perpendicular to the continental margin, and the Mesozoic volcanic arcs developed on continental crust (Hietanen 1975, Dickinson 1981). In contrast, the Karelian craton does not contain granitoids coeval with the subduction postulated by Park and others (1984), although some evidence for this may be present in northern Sweden (Skiöld 1979a,b). On the other hand, using

the Cordilleran paradigm, the Gulf of California may provide a very realistic contemporary analogue for the origin of the Höytiäinen province, involving limited rifting and volcanism at the craton margin.

2.1.ii Australia - Banda arc paradigm

At the risk of introducing another model too prematurely, Cainozoic tectonic development north of Australia is advocated here as an alternative - though not necessarily competetive - paradigm. This region in fact provides one of the only contemporary examples of island arc - continental craton collision, although analysis is complicated by the marked curvature of the subduction zone around the Banda arc, and transcurrent deformation along the leading edge of the Australia craton between Irian Jaya and the Coral Sea (Chamalaun and Grady 1978, Ollier 1981). However, these three essential elements - continent-arc collision, curvilinear subduction zones and transcurrent deformation - can be adapted to the proposed Finnish situation, after due consideration of specific features such as tectonic transport directions and fault displacements, resulting in the relationships depicted in Figure 45. Based on these relation- ships, the following tectonic development can be envisaged under a compressive stress regime directed between N-S and NNW-SSE.

(i) En echelon strike-slip basins develop within the Karelian craton as compressive stresses are accommodated along reactivated Archaean lineaments trending roughly NW-SE and N-S. Anastomozing of NW-SE trending shears also causes alternate zones of uplift and subsidence (Ballance and Reading 1980, Katz 1976) and increasing tension promotes intrusion and extrusion of mafic igneous rocks. More specifically, this extension within the Karelian craton is taken to post-date the main phase of Jatulian orthoquartzite deposition, and accounts for the intracratonic situation and base- ment or Jatulian provenance of detritus within the Höytiäinen province, as well as igneous complexes such as at Tohmajärvi. The longetivity

of these basins is not yet determined but the oldest Jatulian meta-
basites indicate initiation after about 2.13 Ga while the oldest
syn-tectonic intrusions suggest destruction before about 1.92 Ga.

(ii) At this stage, instead of all compressive stress being
resolved along NW-SE and NE-SW conjugate shears, limited subduction
towards the south or south-west takes place. If the northwards sub-
duction of the Svionian arc (Hietanen 1975) is also accepted, as it
is by Gaál (1982b), Edelman and Jaanus-Järkkälä (1983) and Park and
others (1984), then the resultant configuration would have resembled
a mirror image of Australia, Timor and the Banda arc during the
Pliocene (Chamalaun and Grady 1978). Such relationships could account
for some of the distinctive lithologies along the Bothnia-Ladoga zone,
including bimodal volcanism (Gaál and Rauhamäki 1971, Nykänen 1975,
Helovuori 1979, Huhtala 1979) and also explain the almost .coeval
isotopic ages between these and similar early intrusive rocks in
south-west Finland (Simonen 1980, Hopgood and others 1983, Korsman
and others 1984, Huhma 1985).

Observations from the study area are relevant to discussion of
limited subduction along the craton margin in that a Svecofennian
volcanic arc source has been postulated for the Kalevian deposits
of eastern Finland, including the metasediments of the Savo and
Höytiäinen provinces (Laajoki 1984, Park and others 1984). Petro-
graphical and limited isotopic results show, however, that the Höyti-
äinen province deposits derived from adjacent basement and Jatulian
detritus, and a similar provenance is indicated for at least part of
the Savo province (see Section 5). It is thus possible that the Savo
province metapsammites represent deposition on a passive cratonic
margin with the Outokumpu assemblage representing the more distal
basin floor, including a substantial component of mantle-derived
igneous material. Hence, it also follows that the Savo province need
not correspond to a prograding fore-arc or post-arc sequence, and
therefore that the Kalevian metasediments in general cannot be used
to unequivocally infer arc polarity or subduction direction.

Deformation of these deposits may however provide some evidence in
this regard, and is considered as the next stage in the development
of the orogen.

(iii) Along the Bothnia-Ladoga zone the major phase of igneous
activity culminated with intrusion of granitoids and mafic plutons during
the development of D_4 structures (Gaál and Rauhamäki 1971, Gaál 1972,
Halden 1982). This magmatism was probably initiated just prior to
Svecokarelian deformational history since gneisses interpreted as the
deformed relicts of pre-tectonic granitoids (Ward 1984) have U-Pb
zircon ages of 1.91 - 1.93 Ga, similar to those of associated felsic
volcanic rocks (Helovuori 1979, Korsman and others 1984). On this
basis, igneous activity began during abortive subduction of the
Karelian craton, and continued while the basin-fill now represented
by the Savo province was thrust along a N or NNW vector onto the
craton, as indicated by deductions of tectonic translation direction
for F_2 and F_3 in the study area. The mechanism for this thrusting may
have been either continued intensified compression against the craton
or gravitational uplift, due to the buoyancy of the buried craton
margin, or a combination of both these processes. This closely
parallels the response of the Australian continental margin during
Pliocene collision with the Banda arc north of Timor, including
regional metamorphism of the shelf sequence during deformation
(Chamalaun and Grady 1978, Berry and Grady 1981).

In addition, this interpretation is entirely consistent with
the evidence for a diachronous onset of deformation across the study
area. That is, the absence of D_1 structures, and less pervasive nature
of D_2 structures in the Höytiäinen province is compatible with
thrusting propagating from the south and progressively migrating
northwards toward the craton.

Regional metamorphism of the Savo province and underlying base-
ment can be explained by burial beneath the accretionary prism of
the volcanic arc (Figure 45). An alternative interpretation of the

Outokumpu assemblage as the allochthonous ophiolite substrate to a prograding fore-arc. sequence represented by the Savo province metapsammites also seems tenable, except that another mechanism for burial and metamorphism is required and furthermore, no counterpart to subduction complex lithogies exist beneath the allochthonous Outokumpu assemblage, which is at variance with the fore-arc stratigraphy accepted by Dickinson and Seely (1979) and Crook and Feary (1982). Farther south, in the Haukivesi area (Map 1), the Savonlinna conglomerate and Rantasalmi metaturbidites (Gaál and Rauhamäki 1971, Korsman 1977) might represent younger post-arc deposits, or else part of the accretionary prism of the Svionian arc (Figure 45). Of interest in this regard is that on the basis of . fold asymmetry and stretching lineations, Gaál and Rauhamäki (1971) proposed southwards thrusting; if this can be verified in relation to depositional and stratigraphical younging criteria, then the reconstruction by Gaál (1982a, figure 2) becomes a realistic one.

(iv). This represents the last major stage of development, and coincided with Group 4 and Group 5 structures in the study area. These have already been interpreted as products of increased vertical stresses due to thermal and gravitational instability driven by abundant magmatism and buoyancy of the buried cratonic margin (see sections 6.4 and 6.6 in structural studies). This also has an analogue in the Australia - Banda arc collision in that post-collision deformation on Timor has been dominated by vertical uplift due to the buoyancy of Australian continental crust although this has evidently been insufficient to unroof any syn-tectonic plutonic rocks (Chamalaun and Grady 1978).

2.2   ALTERNATIVE HYPOTHESES

Both the Cordilleran and Australia - Banda arc paradigms considered above can be criticized for being premature and pre-empting a better understanding of regional Finnish geology, as well as for tacitly invoking uniformitarian plate tectonic mechanism. Although some results seem to favour similar operation of the Wilson cycle and fore-arc accretion during the early Proterozoic (Taylor 1979, Hoffman and Bowring 1983, Barbey and others 1984), other workers are more guarded, in suggesting that differences existed in physical properties of the Precambrian lithosphere (see Glikson 1981).

For instance, in ascribing present-day subduction to the negative buoyancy of oceanic crust, Hargreaves (1981) considers that such buoyancy was inhibited by higher heat flow in the Archaean and early Proterozoic and concludes that subduction and lateral accretion were not at that time, the major processes leading to continental growth. However, smaller convective cells in the mantle, and lower viscosity are envisaged as having increased the viscous drag at the base of the crust, promoting underplating, sinking and vertical recycling of the crust, including down-buckling at the margins of continental crust. Such vertical recycling is consistent with palaeo-magnetic evidence for minimal mutual displacement between Archaean cratons during the early Proterozoic and also petrogenetic and isotopic evidence suggesting granitoid derivation by recycling from sources with low initial $^{87}Sr/^{86}Sr$ (see Glikson 1981).

Because of the large size of oceanic plates on the present-day earth it is difficult to find any extant example of vertical re-cycling and evolution of oceanic crust. - Iceland is perhaps the closest contemporary analogue. If spreading rates are conducive to maintainance of high heat flow and continued vertical addition of magma, partial-melting of the most deeply buried, oldest MORB-type rocks may be envisaged, producing plagiogranitoid or tonalitic magmas. If the process is permitted to continue, the evolution of successive

partial-melts (including derivation from volcanogenic sediments?) might ultimately result in K-enriched magmas, thus corresponding to the petrogenetic sequence described by Hietanen (1975) for the Svecofennides. Preliminary Nd-Sm data (H. Huhma spoken comm.) at least do not contradict such a proposal in that both Svecofennian metasediments and syn-tectonic granitoids have similar isotopic characteristics, suggesting that in each case they derived from a similar primitive, though somewhat depleted mantle source.

FIGURE 45 : REGIONAL TECTONIC PROPOSALS

a Major tectonic features of the central and eastern Baltic shield,
   accepting at least limited operation of plate-tectonic processes.
   Oblique convergence of Proterozoic crust from the south with
   Archaean continental crust from the north both initiated en
   echelon cratonic margin basins and caused their subsequent de-
   struction and deformation during the Svecokarelian orogeny. The
   onset of deformation may be linked to attempted and abortive sub-
   duction of the Karelian craton towards the south-east, combined
   with northwards consumption of Proterozoic oceanic crust. Such a
   markedly curved subduction zone trace, with concomitant trans-
   current shear and en echelon basins resembles the late Neogene
   collision of the Australia-Papua New Guinea continent with the
   Banda Arc.

b Possible accommodation of major lithostratigraphical units of
   eastern Finland to such a plate tectonic scenario. As yet however,
   insufficient evidence exists for confidently deducing arc polarity,
   sedimentary provenance or even the existence of a subduction
   complex.

BOTHNIA-LADOGA ZONE

DEPOSITION OF PUOLANKA-KAINUU SEQUENCES

NASCENT HÖYTIÄINEN PROVINCE

DEPOSITION IN PRECURSOR TO SAVO PROVINCE

b

PROGRADING FORE-ARC DEPOSITS (cf. RANTASALMI METATURBIDITES ?)

SUBDUCTION COMPLEX

MAGMATIC ARC

LAPPLAND GRANULITE & BELOMORIAN FOLD BELTS

KARELIAN CRATON

PUOLANKA-KAINUU

HÖYTIÄINEN PROVINCE

SAVO PROVINCE

BOTHNIA-LADOGA ZONE

SVIONIAN ARC

a

APPENDIX: INDEX OF LOCALITY NAMES

Names of 1:20 000 map sheets are underlined, 1:100 000 sheets in upper case.

LOCALITY	MAP 1 REFERENCE		MAP 2 REFERENCE		OTHER
Erkkilä	-		-		Map 3
Haapajärvi	-		Niittylahti	95.30	Map 3
Hammaslahti mine	KIIHTELYSVAARA	01	Rauansalo	01.28	
Hammaslahti township	JOENSUU	10	Pyhäselkä	97.24	
Haukivesi	SAVONLINNA	06	-		
Heinävaara	KIIHTELYSVAARA	03	Heinävaara	08.39	
Helsinki	-		-		Figure 1
Hypönniemi	RÄÄKKYLÄ	03	Varpasalo	68.19	
Hämeenlahti	RÄÄKKYLÄ	06	Oravisalo	76.15	Figure 2
Höytiäinen	KONTIOLAHTI	08	-		
Joensuu	JOENSUU	09	-		
Kaavi	KAAVI	01	-		
Kakkari	-		-		Figure 2
Kalliojärvi	KIIHTELYSVAARA	02	Heinävaara	03.32	
Kankaala	-		-		Figure 2
Karhunsaari	JOENSUU	04	Tutjunniemi	73.25	
Karvanurmi	-		-		Map 3
Kerimäki	KERIMÄKI	02	-		
Kettämö	JOENSUU	10	Pyhäselkä	94.26	Map 3
Kettämönniemi	-		Pyhäselkä	95.25	Map 3
Kiihtelysvaara	KIIHTELYSVAARA	05	Kiihtelysvaara	13.32	
Kilpelänkangas	-		-		Map 3
Koli	KOLI	12	-		
Kontiolahti	KONTIOLAHTI	11	-		
Kostamo	-		Onkamo	01.17	
Kukkupää	-		Niittylahti	95.34	
Kumpu	JOENSUU	11	Niittylahti	98.37	
Ladoga	LADOGA		-		
Lentosärkkä	-		-		Map 3
Liperi	JOENSUU	02	-		

	MAP 1 REFERENCE		MAP 2 REFERENCE		OTHER
Mulo	JOENSUU	08	Niittylahti	90.36	
Mulonsalo	-		Niittylahti	95.37	
Nieminen	RÄÄKKYLÄ	09	Rääkkylä	86.26	
Niinikkosaari	JOENSUU	04	Tutjunniemi	70.28	Map 3
Niittylahdenrantatie	-		Niittylahti	91.31	
Niittylahti	JOENSUU	11	Niittylahti	93.36	
Niva	JOENSUU	10	Pyhäselkä	93.24	Map 3
Nivansalo	-		Pyhäselkä	90.20	
Onkamo	TOHMAJÄRVI	03	Onkamo	03.15	
Oravisalo	RÄÄKKYLÄ	06	Oravisalo	75.15	Figure 2
Oravisalo quarry	-		-		Figure 2
Oravisalo post office	-		-		Figure 2
Orivesi	RÄÄKKYLÄ	03	-		
Outokumpu	OUTOKUMPU	07	-		
Petäjikkökallio	-		-		Map 3
Pitkänurmi	-		-		Map 3
Puhos	KERIMÄKI	12	-		
Pyhäselkä	JOENSUU	03	-		
Rasivaara	RÄÄKKYLÄ	08	Rasivaara	86.07	
Rauansalo	KIIHTELYSVAARA	01	Rauansalo	05.23	
Reijola	JOENSUU	11	Niittylahti	91.38	
Rääkkylä	RÄÄKKYLÄ	09	Rääkkylä	81.14	
Salonkylä	JOENSUU	10	-		Map 3
Sammallahti	JOENSUU	10	Pyhäselkä	91.20	
Savo	-		-		
Sotkamo	KONTIOLAHTI	04	-		
Suhmura	JOENSUU	11	Niittylahti	94.30	Map 3
Suvasvesi	HEINÄVESI	02	-		
Särkilampi	KIIHTELYSVAARA	04	Viesimo	17.24	
Tikkala	TOHMAJÄRVI	03	Onkamo	04.17	
Tohmajärvi	TOHMAJÄRVI	05	Tohmajärvi	18.02	
Ukonlahti	RÄÄKKYLÄ	06	Oravisalo	73.18	
Vannila	-		Oravisalo	75.17	Figure 2
Varpasalo	RÄÄKKYLÄ	03	Varpasalo	68.12	
Viesimo	-		Viesimo	18.12	
Viinijärvi	KONTIOLAHTI	01	-		
Värtsilä	TOHMAJÄRVI	10	-		

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ORAVISAL...
Rääkkylä l...
Oravisalon ... with granit...
Torohvenniemi ... and thick-be...
Oravisalon p...
Karhunsaari
Kankaala l...
HÖYTIÄINE...
KETTÄMÖ...
Salonkylä l...
Kettämönnie... with quartz
HAMMASL...
Kilpelänkan...
Suhmura lith...
Mulo lithofac... and sulfidi...
Kukkupää l...
Tohmajärvi l...

STRUCTURAL SYMBOLS

- gneissosity in Archaean basement

- depositional layering, frequently modified by S_1 or S_2 differentiation layering

- S_2 differentiation layering in major fold hinge, at high angle to lithological layering and S_1

- S_3 continuous cleavage and differentiation layering axial planar to generally recumbent F_3 folds

- lithological layering and foliation from published maps

- S_4 crenulation cleavage indicating sinistral (S) or dextral (Z) vergence

- S_5 crenulation cleavage indicating sinistral (s) or dextral (z) vergence

- weak disjunctive cleavages

- trace of thrust fault

- trace of high-angle fault coincident with S_5 and F_5

- L_3, defined as intersection lineation of S_0 and S_3

- L_3 as shape fabric lineation defined by elongate biotite aggregates or deformed clasts

- F_3 minor fold indicating plunge and upper limb (Z) or inverted limb (S) vergence

- lineation from published maps (Nykänen 1967, 1968; Lavikainen 1985; Pekkarinen 1979)

- F_4 or F_5 minor folds with S_4 or S_5 as axial plane

- F_3 upper limb or Z-vergence deduced from minor folds or relations between S_0 and S_3

- F_3 inverted limb or S-vergence deduced from minor folds or relations between S_0 and S_3

- F_3 hinge zone

- depositional younging direction affected only by recumbent F_3 folding

- depositional younging direction indicates inverted zone prior to recumbent F_3 folding

- unconformity

LITHOFACIES EXPLANATION

SAVO PROVINCE

OUTOKUMPU ASSEMBLAGE

- serpentinites, calcareous, grafitic, sulfidic metapelites

PYHÄSELKÄ ASSEMBLAGE

- Rääkkylä lithofacies: thick to medium bedded feldspathic metapsammite; metapelite and depositional structures rare

- Niittylahdenrantatie lithofacies: metagraywackes comprising large clasts from the Rääkkylä lithofacies

- Sammallahti lithofacies: feldspathic metapsammite with granule-sized quartzose detritus

- Niva lithofacies: massive metapelites and lesser mafic actinolite-chloritoid schists

93

493

69 20

930

94

494

69 30

Suhmura

495

496

Haapajärvi

69 29

69 29

TN

Salonkyla

Kilpelänkangas

Pitkanurmi

Karvanurmi

69 28

69 28

Lentosärkkä

Lähdenurmi

69 27

69 27

Pyhäselkä

Petäjikkökallio

69 26

69 26

1:20 000 REFERENCE DIAGRAM

outcrop and locality number corresponding
to each 1km² of map grid

lithofacies boundary, probably transitional

major stratotectonic boundary between
assemblages

Haapajärvi fault zone probably coeval in
addition to being coplanar with S_5

dextral, sinistral or neutral F_3 fold vergence

boundary between major F_3 domains

downward facing zones inverted prior to
development of recumbent F_3 folds

trace of lithological layering

access roads and tracks

Helsinki - Joensuu railway

LITHOFACIES DISTRIBUTION

Perspective projection viewed down plunge from N to S Bold locker grid coincides with 1:20000 topographical map areas
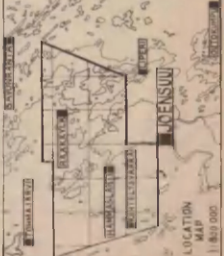
PROTEROZOIC INTRUSIVE ROCK
- Herpaselkä felsic pegmatoids

PYHASELKÄ ASSEMBLAGE
- Raskylä lithofacies
- Niittylahdenrantatie lithofacies
- Sammaslahti lithofacies
- Niva lithofacies

ORAVISALO ASSEMBLAGE
- Raakkylä lithofacies
- Oravisalo post office lithofacies
- Kankaala lithofacies
- Quarry lithofacies
- Syrjänsuu - Niinikangas lithofacies

KETTÄMO ASSEMBLAGE
- Kokkari - Hypyrichardy granitoid gneiss
- Solomkylä lithofacies
- Kettämäniemi lithofacies
- Lentaasrikkä subfacies
- Pitkäturmi subfacies
- Petajäkkallio subfacies
- Kurvanumi subfacies
- Kettämäniemi granitoid gneiss

HAMMASLAHTI ASSEMBLAGE
- Nilposkämkangas lithofacies
- Salamara lithofacies
- Mula lithofacies
- Tikkala lithofacies
- Hammaslahti copper deposit
- Raumade lithofacies

KIIHTELYSVAARA ASSEMBLAGE
successions and nomenclature of Pekkarinen (1979)

- Kaillejärvi lithofacies
- Kukkupää lithofacies

TURBIDITE SUITE
- Mica schist
- Turbidite conglomerate - quartzite

CARBONATE-BLACK SLATE SUITE
- Dolomite-carbonaceous-shale-volcanite
- Hematitic quartzite
- Dolomite - volcanite

ORTHOQUARTZITE SUITE
- Upper quartzite
- volcanite
- Lower quartzite

ARKOSITE SUITE
- Arkosite
- Breccia and weathering crust
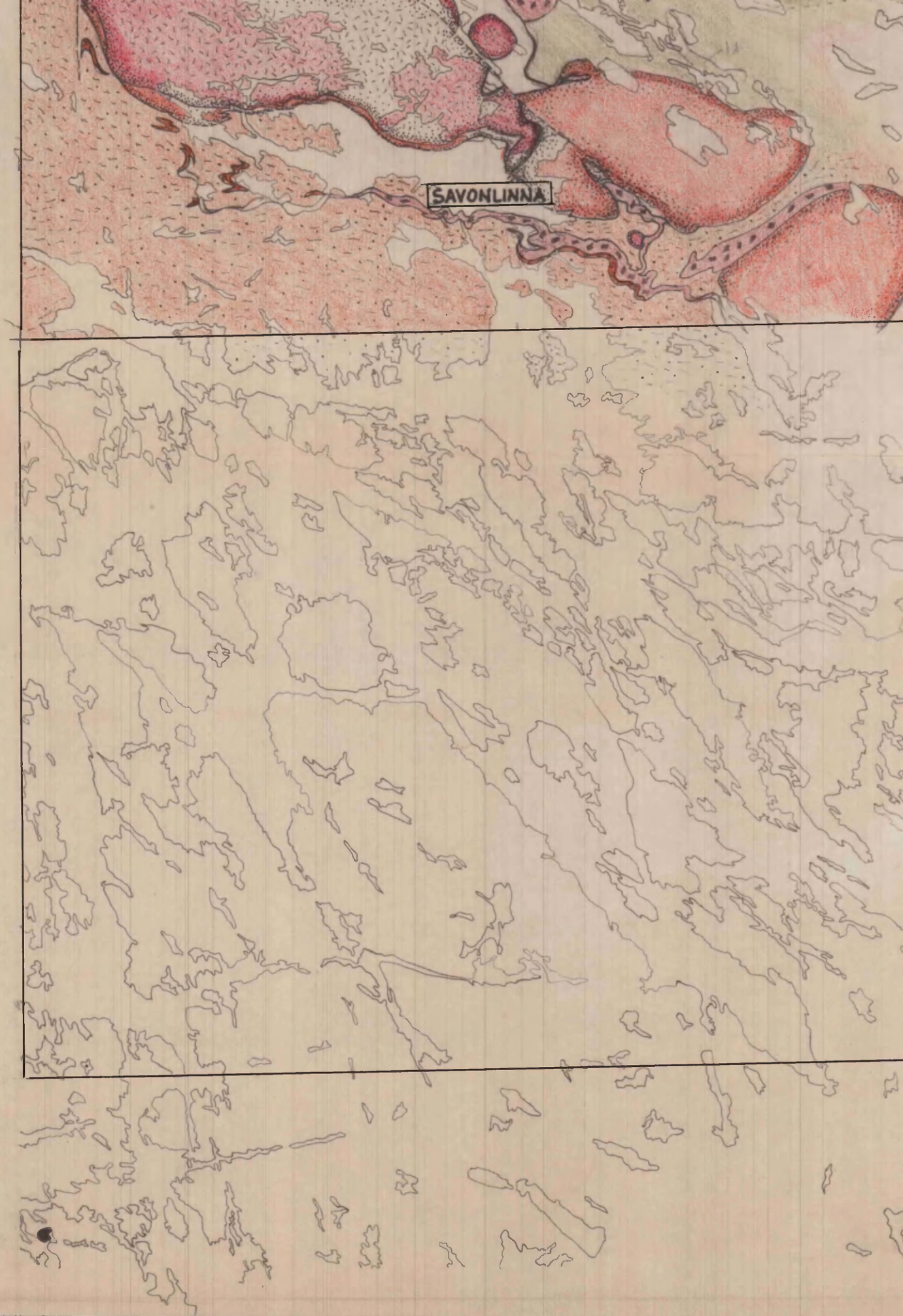- Archaean basement

SCHEMATIC LITHOFACIES RELATIONS

KETTÄMO THRUST ZONE

HAMMASLAHTI ASSEMBLAGE

KIIHTELYSVAARA ASSEMBLAGE

KETTÄMO ASSEMBLAGE   PYHASELKÄ ASSEMBLAGE   ORAVISALO ASSEMBLAGE

LOCATION MAP

JOENSUU

SAVONLINNA

STRUCTURAL DIAGRAM OF STUDY AREA
BASED ON DIMETRIC PROJECTION

ARCHAEAN GRANITOID GNEISSES

ARCHAEAN SUPRACRUSTAL
LITHOLOGIES

JATULIAN SEDIMENTS

HAMMASLAHTI
MINE

KETTANO THRUST ZONE

HAAPAJARVI
FAULT ZONE

KIIHTELYSVAARA
BASEMENT

TOHMAJARVI MAFIC IGNEOUS COMPLEX

ORIVESI
SHEAR
ZONE

SYNTECTONIC
PEGMATITES

trace of lithological layering

unconformity

Jatulian metabasite dykes

HOYTIAINEN PROVINCE

SAVO PROVINCE

ARCHAEAN BASEMENT

REFERENCE GRID FOR MAP 1 (1: 200 000 SCALE)

12	06
11	05
10	04

09	03
08	02
07	01

numbered squares correspond to 1:20 000 map sheet boundaries

total area (12 sheets) constitutes 1:100 000 sheet, denoted in upper case on Map 1.

REFERENCE GRID FOR MAP 2 (1 : 50 000 SCALE)

x. 0	x. 1	x. 2	x. 3	x. 4	x. 5	x. 6	x. 7	x. 8	x. 9
y. 9									
y. 8									
y. 7									
y. 6									
y. 5									
y. 4									
y. 3									
y. 2									
y. 1									
y. 0									

Grid area corresponds to 1: 20 000 sheet area   (10 km × 10 km)