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DEPOSITION AND DEFORMATION AT A LOWER PROTEROZOIC CRATONIC MARGIN, SOUTH-EAST FINLAND

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October 1985.
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.

Signed 7/2/1986
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 interpreted 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-beded 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 lithologies 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 intracratonic 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 precursor 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 basement and source terrain but the ultimate nature and location of provenance for the bulk of the sediment remains problematical.

Such 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 considered 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 Svecokarelian 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 effectively bisecting the Archaean craton, and marginally older than Svecokarelian 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 Lapland granulite belt, and flanked to the south-west by the somewhat younger Svecokarelian 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 orogenesis culminated at 1.9 - 1.8 Ga, since when the entire Finnish crust has behaved as a stabilized craton.
S C O P E  O F  P R E S E N T  P R O J E C T

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.
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 km² and shows 1 km² 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 km² grid having co-ordinates 97 E and 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 km² 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° to +8.0° - 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 Rauhämäki 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 displacements 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 stratigraphical 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.

Metapsammitte 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.
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).
PART II: STRUCTURAL INVESTIGATIONS
1. PROCEDURES IN STRUCTURAL ANALYSIS

1.1 GENERAL METHODS AND LIMITATIONS

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 overprinting relationships - as indicated by intrusive neosomes or mineral segregation, growth and recrystallization, - or from deformation 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 correlations 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 structures; 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 possess 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, it 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 ARCHAEOAN 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 ultrametamorphism 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 intruding 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).
2.2.1 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 ($F_{III}$), (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 ($S_{V}$) 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 syntectonic 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. Intrafolial 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 (Viesimo 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 ($S_1$). 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 $S_y$ 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 (Viesimo 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 1979a). These are in turn intruded by unfoliated oligoclase pegmatites.
2.3 CORRELATION OF STRUCTURES IN ARCHAEOAN 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 ARCHAEOAN 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 2a). 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 24c).

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 1979a). 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 (1979a), 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.
**VII**

**SY-N trending Fault**

- **Fracture**
  - Right lateral in foliation

**VI**

**SY-N trending Cleavage**

- **Foliation**
  - Biotite and muscovite

**V**

**SY-N trending Birnessite**

- **Foliation**
  - Reaction with biotite

**IV**

**SY-N trending slickensides**

- **Foliation**
  - Mitotic and biotite

**III**

**SY-N trending foliation with phyllochondrites**

- **Foliation**
  - Mitotic and biotite

**II**

**SY-N trending foliation**

- **Foliation**
  - Mitotic and biotite

**I**

**SY-N trending foliation**

- **Foliation**
  - Mitotic and biotite
FIGURE 2: ARCHAEO 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 Svecokarelían L3. 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.
Group 3 quartz-ribbon mylonites in Archaean gneiss

Proterozoic unconformity

meta-arkose of Oravisalo quarry lithofacies A

late Archaean or Jatulian mafic dykes

Group 3 thrust

Oravisalo quarry lithofacies B

Section N-S

Basement-cover unconformity

Kakkari

Vannila

Oravisalo post office

Oravisalo quarry

S3 thrust zone

L3 fold hinge

L3 mineral imbrication

S3 foliation

S3-S2 composite layering

S1-S3 banding in Archaean gneiss

100 m

400 m

200 m

0 m

TN
FIGURE 3: STRUCTURES IN ARCHAEOAN 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).
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 $S_0$. 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.

3.2.1 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 $S_0$ (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 $S_{0-1-2}$ 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).
3.2.11 Raakkyla area (Raakkyla 81. 14.)

In this area relationships amongst early structures are well established, 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 Hypönemi 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. Consistent 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 isoclinal and are consistently downward-facing. $S_9$, $S_2$ 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 ($S_6$).
<table>
<thead>
<tr>
<th>Group</th>
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<th>Rääkkylä</th>
<th>Hypölässäniemi</th>
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<th>Niittylahdenrantatie</th>
</tr>
</thead>
<tbody>
<tr>
<td>$S_0$</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
</tr>
<tr>
<td>1</td>
<td>$S_0$ differentiation layering parallel to $S_0$</td>
<td>$S_0$ differentiation layering and intrafolial $F_0$ folds</td>
<td>$S_0$ differentiation layering and axial planar $F_0$ folds</td>
<td>$F_0$ recumbent folds, $S_0$ transposing depositional layering and biotite differentiation layering</td>
<td>$F_0$ recumbent folds, $S_0$ axial planar $S_0$ with biotite growth</td>
</tr>
<tr>
<td>2</td>
<td>$F_0$ isoclinal fold with spaced axial planar layering</td>
<td>$F_0$ isoclinal folds with spaced axial planar $F_0$ layering</td>
<td>$F_0$ isoclinal folding axial planar to $F_0$ isoclinal folds</td>
<td>$F_0$ isoclinal folds with axial planar $F_0$, locally as differentiation layering</td>
<td>$F_0$ isoclinal folds with axial planar $F_0$, locally as differentiation layering</td>
</tr>
<tr>
<td>3</td>
<td>$F_0$ tight to isoclinal folds with axial planar $S_C$ and $L_C$ shape fabric</td>
<td>$F_0$ asymmetric recumbent folds with axial planar $S_C$ biotite schistosity and $L_C$ shape fabric</td>
<td>$F_0$ tight to isoclinal folds with axial planar $S_C$, locally as differentiation layering</td>
<td>$F_0$ tight to isoclinal folds with axial planar $S_C$, locally as differentiation layering</td>
<td>$F_0$ tight to isoclinal folds with axial planar $S_C$, locally as differentiation layering</td>
</tr>
<tr>
<td>4</td>
<td>Quartz veins and pegmatites</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
</tr>
<tr>
<td>5</td>
<td>Minor $F_0$ folds of dextral chirality, with NN-trend</td>
<td>Minor $F_0$ folds of dextral chirality with NN-trending $S_C$ crenulation</td>
<td>$S_0$ crenulation cleavage</td>
<td>$S_0$ crenulation cleavage</td>
<td>Steepening of earlier structures with $S_0$ crenulation cleavage</td>
</tr>
<tr>
<td>6</td>
<td>Steepening of earlier structures with minor $F_0$ folds and N-trending $S_0$ crenulation cleavage</td>
<td>$S_0$ crenulation cleavage</td>
<td>Steepening of earlier structures with $S_0$ crenulation cleavage</td>
<td>Steepening of earlier structures with $S_0$ crenulation cleavage</td>
<td>$S_0$ NE-trending crenulation cleavage</td>
</tr>
<tr>
<td>7</td>
<td>$S_0$, weak E-trending cleavage</td>
<td>E- and ESE-trending weak $S_0$; open $F_0$ folds</td>
<td>$F_0$, E-trending cleavage</td>
<td>$S_0$, E-trending cleavage</td>
<td>$S_0$, weak E-W cleavage</td>
</tr>
<tr>
<td>8</td>
<td>$S_0$, N-trending fractures</td>
<td>$S_0$, N-trending fractures</td>
<td>$S_0$, N-trending fractures</td>
<td>$S_0$, N-trending fractures</td>
<td>Conjugate fractures with E-ESE trend</td>
</tr>
<tr>
<td>9</td>
<td>Conjugate fracture arrays</td>
<td>Conjugate fractures with E-ESE trend</td>
<td>Conjugate fractures with E-ESE trend</td>
<td>Conjugate fractures with E-ESE trend</td>
<td>Conjugate fractures with E-ESE trend</td>
</tr>
</tbody>
</table>

**TABLE 2: CORRELATION OF SVECOKARELIAN STRUCTURES IN THE SAVO PROVINCE.**
3.3 SEQUENTIAL RELATIONSHIPS IN THE HOYTIAINEN PROVINCE

In addition to depositional layering being readily recognizable, except in the more schistose metapelite, 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.

3.3.1 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 metapsammite, 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 axial surface shows them to be younger, as does the deformation of Group 2 and Group 3 microstructures (Figures 11e, 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, 11f, 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 Hammalahti 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 isoclinal, 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
<table>
<thead>
<tr>
<th>Group</th>
<th>Kettämönniemi–Suhmura</th>
<th>Nuolonsalo</th>
<th>Kalliojärvi</th>
<th>Hammaslahti mine</th>
<th>Tikkala</th>
</tr>
</thead>
<tbody>
<tr>
<td>( S_0 )</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
<td>Depositional layering</td>
</tr>
<tr>
<td>1</td>
<td>Not observed</td>
<td>Not observed</td>
<td>Not observed</td>
<td>Not observed</td>
<td>Not observed</td>
</tr>
<tr>
<td>2</td>
<td>Localized slide zones and ( S_0 ) differentiation layering</td>
<td>Local ( F_b ) isoclinal folds and weak axial planar ( S_0 ) differentiation layering</td>
<td>( S_0 ) differentiation layering in metapsammites</td>
<td>( F_b ) recumbent isoclinal folds ( S_0 ) transposition layering</td>
<td>Sporadic ( S_0 ) differentiation layering in metapsammites</td>
</tr>
<tr>
<td>3</td>
<td>( F_b ) asymmetric recumbent folds with axial planar ( S_0 ) continuous cleavage and ( L_b ) shape fabric lineation</td>
<td>( F_b ) tight to isoclinal folds and ( S_0 ) transposition and crenulation cleavage</td>
<td>( F_b ) isoclinal folds and ( S_0 ) transposition and crenulation cleavage</td>
<td>( F_b ) isoclinal folds with ( S_0 ) transposition and continuous (slaty) cleavage; ( L_b ) boundins and shape fabric lineation</td>
<td>( F_b ) isoclinal folds with ( S_0 ) transposition and continuous (slaty) cleavage; ( L_b ) boundins and shape fabric lineation</td>
</tr>
<tr>
<td></td>
<td>Quartz veins</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
<td>Quartz veins</td>
</tr>
<tr>
<td>( S_1 )</td>
<td>NW-trending ( S_C ) crenulation cleavage axial planar to minor folds of dextral chirality; rarer NE-conjugate crenulations</td>
<td>NW-trending ( S_C ) crenulation cleavage</td>
<td>NW-trending ( S_C ) crenulation cleavage</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>S-SW-plunging ( F_d ) folds with axial planar ( S_d ) crenulation cleavage</td>
<td>S-SW-plunging ( F_c ) folds with axial planar ( S_C ) crenulation cleavage</td>
<td>Asymmetric ( F_d ) folds with axial planar ( S_d ) crenulation cleavage</td>
<td>Asymmetric ( F_d ) folds with steepening of earlier structures, quartz veins</td>
<td>Asymmetric ( F_c ) folds with steepening of earlier structures and ( S_d ) crenulation cleavage, quartz veins</td>
</tr>
<tr>
<td>6</td>
<td></td>
<td></td>
<td></td>
<td>NE-trending kinks and ( S_0 ) crenulation cleavage</td>
<td>NE-trending dextral faults</td>
</tr>
<tr>
<td>7</td>
<td>E-trending ( S_e ) crenulation</td>
<td>E-trending ( S_d ) crenulation</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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

**TABLE 4: CORRELATION OF SVECOKARELIAN STRUCTURES BETWEEN HOYTIAINEN 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, superimposed 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 (Gaal and Rauhamäki 1971) as well as to the north, including the Outokumpu mining district (Map 1) (Gaal and others 1975, Bowes 1976, Gaal 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 F2 folds compares well with pre-D1 and D1 structures elsewhere in Karelia, and the Group 3 lineation defined by aggregates of biotite in the study area can also be confidently correlated with L2 of Bowes and others (1984). Likewise, the form of F3 folds and the presence of Group 3 mylonites near the basement-cover interface at Oravisalo corresponds well with D2 features north of Outokumpu (Park and Bowes 1983).

Group 4 structures closely resemble those described as D2c 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.
<table>
<thead>
<tr>
<th>GROUP</th>
<th>Dn</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>GROUP 1</td>
<td>pre-D₁</td>
<td>Nappe emplacement; segregation quartz veins.</td>
</tr>
<tr>
<td>GROUP 2</td>
<td>D₁</td>
<td>Isoclinal folds (F₁); axial planar metamorphic segregation banding (S₁).</td>
</tr>
<tr>
<td>GROUP 3</td>
<td>D₂</td>
<td>Asymmetrical open to tight folds (F₂); axial planar schistosity (S₂); thrusts; mylonite-phyllonite in S₂ transposition; strong mineral lineation (L₂).</td>
</tr>
<tr>
<td>GROUP 4</td>
<td>D₂c</td>
<td>Asymmetrical folds, conjugate in places (F₂c); crenulation cleavage (S₂c); expression largely confined to major NW-SE-trending wrench-fault zones.</td>
</tr>
<tr>
<td>GROUP 5</td>
<td>D₃</td>
<td>Open, upright N-to NNW-trending folds, commonly asymmetrical (F₃); crenulation or fracture cleavage (S₃) and intersection lineation (L₃) develop locally.</td>
</tr>
<tr>
<td>GROUP 6</td>
<td>D₄</td>
<td>Open, upright NE-trending folds (F₄); axial planar cleavage (F₄); prominent crenulation or rodding (L₄) in parts.</td>
</tr>
<tr>
<td>GROUP 7</td>
<td>D₅</td>
<td>Open, upright, E-trending folds (F₅); axial planar cleavage or fractures (S₅).</td>
</tr>
<tr>
<td>GROUP 8</td>
<td>D₆</td>
<td>Open, upright, N-trending folds (F₆); axial planar healed cleavage or fractures (S₆); local pseudotachylite.</td>
</tr>
<tr>
<td>GROUP 9</td>
<td>D₇</td>
<td>Conjugate set of open, upright WNW- and WSW-trending folds (F₇) that pass into fractures (S₇).</td>
</tr>
</tbody>
</table>

**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 (Oravislalo 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 $L_3$ (Oravislalo 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 metapelitite (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 layering. Curvature of $S_5$ axial plane indicates the presence of open E-trending $F_7$ fold (Niittylahti 94.5 31.3).
4. NATURE AND EXPRESSION OF EARLY, DOMINANTLY LOW-ATTITUDE STRUCTURES IN PROTEROZOIC COVER ROCKS

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

4.1.1 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 metapsammmites (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 Gaal (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:

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

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 plagioclase 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 S₂ 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 S₃. Within both of these provinces, smaller structural domains are recognized and relatively abundant S₃-S₀ relationships and younging determinations have resulted in a better definition of geometry and vergence than for other sets of structures.

4.3.1 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 L₃ biotite lineation shows gentle ENE and WSW plunges but elsewhere it has a moderate plunge and trends consistently between SSW and WSW.

F₃ minor folds tend to be parallel to this L₃ lineation but at Oravisalo for example, they are commonly divergent, as shown in the relevant equal area plots in Figure 5. In Figure 9d, the two lineations are actually mutually perpendicular: L₃ has the same expression and trend as elsewhere, while the fold is indistinguishable from others ascribed to F₃ 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 L₃ 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-lims 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 9a,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 1:5000 scale on Map 3 has revealed the existence of a major $F_3$ fold verging to the NNW whose development 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 7c and 11d 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 metapsammit 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 9a and 10). Examples where transposition has occurred on normal-limbs of minor folds also exist (Figure 11b).

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 F₃ 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 superposed 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 12e-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 migmatites, whether as the products of in situ partial melting, metamorphic 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 representative 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 $S_3$ foliation, 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 Varpasalo 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 development 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 metapsammmites 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 relics 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 Hötylässinen 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 11f or where post-$S_3$ biotite porphyroblasts are deformed by $S_5$ (Figure 18h).

4.3.4i 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 15a,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 15c). Garnet breakdown has in places progressed even further towards total replacement, as shown in Figure 15d,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 15e).

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) porphyroblasts of chloritoid (Figure 18g). 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 thermal 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 Gaål 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 overprinting 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 Høytiäinen 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

$$\text{quartz} + \text{muscovite} + \text{biotite} + \text{feldspar} + \text{ sulphide} + \text{graphite} + \text{calcite}.$$ 

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 Hammaslhti 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 S₃ and as a later unoriented porphyroblastic growth. This again provides evidence for the thermal metamorphic peak being younger than S₃ development, but preceding S₅ formation, which is seen in general, to deform porphyroblasts (Figure 18h).

The same conclusions are supported by the presence of post-S₃ 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

\[
\text{microcline} + \text{calcite} + \text{graphite} + \text{pyrrhotite} \pm \text{chalcopryite} \\
\pm \text{pyrite} \pm \text{sphene} \pm \text{sphalerite} \pm \text{chlorite} \pm \text{quartz} \pm \\
\text{plagioclase}.
\]

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 S₂-S₃ 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 H₂O, other molecular fluid phases would have been important, such as CO, CO₂, O₂, H₂S, S₂ and CH₄, 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-S₃ 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.
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 $F_1$ 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 Metapsammitic 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 $S_3$ differentiation layering 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 discriminate 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.
schematic depiction
of reorientation
of Group 3 structures
by Group 5 structures
FIGURE 11: GROUP 3 STRUCTURES IN THE HÖYTIAINEN 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 (Pyhäsälä 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 (Pyhäsälkä 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).
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).
 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 (Pyhäsälä 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 (Pyhäsälä 94.9 28.1).

f Laminar, quartz-ribbon mylonite having completely destroyed any evidence for depositional fabric (Pyhäsälä 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 +
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

5.1.1 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 (Gaal and Rauhamaäki 1971, Gaal 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 (Gaal
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 $T = 675 \pm 25^\circ C$ 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 H"oyti"ainen 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 (Ga"al and others, 1975).

5.2.1 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 respect 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 many 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, F₅ 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 S₅ 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, S₅ 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, immediately east of which, depositional layering (where evident) and S₂ - S₃ foliations have been reoriented into an attitude almost invariably steeper than 50° (Maps 2 and 3). In such zones, and particularly between 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 S₂ - S₃ differentiation in metapsammites overprinted by an S₅ 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 F₃ upper-limbs and F₅ folds. Moreover, the frequent coaxial nature of L₃ and the L₅ intersection lineations precludes ready discrimination even in 3-dimensional exposures (Figure 37c). Examples have nevertheless been found where presumable syn-S₃ boudins are seen to pitch more steeply on S₃ planes than the intersection lineation formed with S₅.

S₅ 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 F₃ 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 S₅ crenulation cleavage domains. Some veins are located with S₅ axial planes but a more typical attitude appears to be perpendicular to this and the L₅ intersection lineation. This relationship, of E-W trending veins and S-plunging L₅ 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 F₃ fold axes and boudinaged syn-F₃ or earlier veins. Much smaller veins of similar attitude are shown in Figure 20b, recording 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 S₃ 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 12g,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 metapsammellites of the Savo province and even in the Höytiäinen province, metapsammellites tend to lack syn-$S_5$ mineral growth, apart from recrystallization within the more highly strained limbs of asymmetric microfolds.

In metapelites of the Höytiainen 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 19e.

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 19d.

Diffuse cleavage domains become more distinct as the angle between the reoriented foliation and $S_5$ axial surface decreases, as in Figure 19e. The left hand part of figure 19d 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 19f, 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 19d, b and c. Figure 19g, 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 19b, 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 8e). 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 19d, e, h).

3) Asymmetrical microfolds produced zonal crenulation cleavage
with gradational boundaries between relict microlithons and
developing cleavage domains (Figures 8f and 19e).

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 19c,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 19a,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 19c).

Metamorphic conditions during this stage of the deformational
history are not well constrained. S₅ clearly kinks or truncates
post-S₃ and post-S₄ biotite porphyroblasts, suggesting that the
regional thermal metamorphic peak occurred after S₄ formation
(Figure 18a) but prior to the S₅ crenulation cleavage (Figure 18h).
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 S₅ and relate
this to staurolite instability. Pseudomorphs, presumably after
syn- or post-S₃ 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 ac­
cording 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

5.3.1 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

5.4.1 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. F7 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 20g). They are likely to correspond to the S6 cleavage of Koistinen's (1981) classification and similar structures have been reported by Bowes (1976) and Park (1983).

Figure 20h 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 (S7?) 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 compositional 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.
FIGURE 17: GROUP 4 AND GROUP 5 STRUCTURES

a  F₄ crenulations deforming earlier composite S₀-S₃ fabric. Unless they underwent subsequent mimetic growth, it seems clear that biotite porphyroblasts were syn-tectonic with respect to S₃ and definitely pre-date S₄ (Niittylahti 91.7 35.6).

b  Isoclinal F₃ folds are overprinted by NE-trending crenulations. These could be assigned to S₆-F₆ but sinistral chirality in adjacent exposures suggests that they are antithetic conjugates to the usual NW-trending S₄. (The few examples found where S₆ is established as cutting S₅ have indicated the opposite chirality) (Niittylahti 92.4 35.2).

c  Minor S₄ shear zone containing a quartz vein and causing both distortion and compositional modification of composite S₂-S₃ fabric (Oravisalo 70.5 19.6).

d  S₅ shear zone, probably with some vertical component of displacement. Deformed fabrics at right are difficult to assign, but are either S₃ or S₄ axial planar to F₃ or F₄ dextral folds, deforming lithological layering and parallel S₂ quartz veins (Pyhäsälä 94.0 29.3).

e  Asymmetrical S₅ crenulations developed in laminated metapelite but absent from thicker massive metapsammite (Pyhäsälä 93.9 29.3).

f  Composite S₂-S₃ differentiation fabric with abundant quartz veins, folded by F₅, with no evidence of accompanying S₅ (Pyhäsälä 93.5 29.2).

g  Discrete, zonal S₅ crenulation cleavage across well-preserved depositional laminae and subparallel but weak S₃ (Rauansalo 00.7 23.9).

h  Axial planar S₅ crenulation cleavage deforming S₂ differentiation layering in metapsammite. 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 F4 microfolds that crenulate composite S2-S3 in graphite-bearing metapelite. Hexagonal grain of high relief may be staurolite, though it is not pleochroic and has sector twinning like that of cordierite (Pyhäselkä 94.7 28.6). PPL.

b Pre- or syn-S5 tremolite porphyroblasts superimposed upon S2-S3 differentiation layering in calcareous metapelite (dark bands are graphite-rich). Open crenulations are late F3 or F4 microfolds (Niittylahti 91.3 35.6). PPL.

c F3 or F4 microfolds deforming S2-S3 differentiation layering, with darker bands being graphite-rich. Disrupted pre- or syn-S3 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-S2 veinlet (Niittylahti 91.3 35.6). NX.

e Chloritic and micaceous pseudomorphs after staurolite porphyroblasts, syn- or post-kinematic with respect to S3. 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 S5, encloses a composite S2-S3 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-S3 dendritic actinolite in quartz-mica-chloritoid metapelite (chloritoid not evident in this photograph) (Pyhäselkä 93.9 24.6).

h Post-S3 biotite porphyroblasts deformed and truncated by discrete, anastomosing S5 crenulation cleavage (Niittylahti 97.7 30.5). PPL.
FIGURE 19: \( S_5 \) CRENULATION CLEAVAGE IN HÖYTÄINEN 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 (Valkeasu 17.4 18.9).

g Epidotized fracture zones (S_6), 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).
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 12a,b and 37c). 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 interpretation 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 perpendicular 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.

6.2.1 $L_3$ 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 $S_6-S_2$ 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 metapalaeophite 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 12b).

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 L₃ 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 L₃ 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 (1968a,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 shortening. Metapsephites of the study area have oblate clasts lying within S₂-S₃, consistent with this result while prolate clast long axes tend to parallel F₃ fold axes.

In contrast, Gaël 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 L₃.

Clasts in the study area are less prolate than at Savonlinna 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 perpendicular 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 21c shows granitoid and quartzose clast dimensions in a major $F_3$ fold hinge near Petäjikkö­kallio 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 14e,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 orientation usually associated with the $XY$ plane of the finite strain ellipsoid (Hobbs and others 1976). Figure 21b indicates that flattening perpendicular to this plane (as recorded by the ratio $Z/Y$) has been more substantial than elongation within it (as measured 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 $L_3$ mineral 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 intermediate 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 Gaål 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 (Gaål 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 (Gaål 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 Gaål (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 $\sigma_2$ would have been vertical, and the occurrence of such a stress system in association with syntectonic 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 impractical 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 $\sigma_1$ 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 $\sigma_1$ 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 Gaał 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, $\sigma_1$ 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 \( \sigma_1 \), as vertical. There is no direct evidence from the study area indicating the possibility of significant horizontal stress, causing substantial horizontal displacement 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 substantial 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 configurations 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 reorientation 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 Suvasvesi 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 correlated 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 Gaål and Rauhamäki (1971) and Parkkinen (1975) portray the N-S horizontal component as $\sigma_3$, with an E-W $\sigma_1$ and vertical $\sigma_2$, during the deformation 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 $L_3$ (Hammashalti 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 perpendicular to $L_3$ (ratio $Z/Y$). Granitoid clasts are commonly prolate parallel to $L_3$ (indicated by $Y/X < 1$) whereas graywacke 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.
30

+ dolomite

* intrabasinal graywacke

o granitoid: quartzite and vein quartz

20

10

granitoid Y/X

granitoid Z/Y

10

graywacke Y/X

graywacke Z/Y

60 cm

F3 normal limb and hinge zone

near Kettämö thrust zone
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 ($\sigma_1$) or decreasing vertical stress ($\sigma_3$), since shear stress

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

(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.
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, comprising metabasites or mafic metapelites.
1.4.iii Kettämö assemblage

This is represented by a well-defined stratotectonic unit between the Niittylaahdenrantatie 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 ARCHAEOAN 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 (Saxen 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 Värstilä 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 73.416.4 and at 74.217.2. 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 23a). Only rarely are there massive developments of more mafic granitoid material (Figure 23e).
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 23d). 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 23c).

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 3e).

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 23f). 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 75.715.9 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 Kettämönniemi 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: ARCHAEOAN 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

3.1.1 Oravisalo quarry lithofacies

Distribution and reference exposures:

At locality 75.8159, 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 24b 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 metapsammitite. The meta-arkoses appear very similar to the more felsic, massive granitoid against which they are tectonically juxtaposed, and clear discrimination between them is sometimes only possible where distinctly 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 (Gaål and others 1975), and Kihntelysvaara (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 24a) 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 24c, 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 28e, 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 74.815.4, 74.715.2
on the property "Kankaala" and on the Liperi-Rääkkylä road at locality 75.315.1. 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 reconnaissance maps, such as would be expected from these lithologies, if widespread.

Depositional units and structures:

At 74.814.8, 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 thrust 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 Niitylahdenrantatie 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 67.419.6, in bluffs at locality 74.217.5 and near the eastern limit of its distribution, at 94.120.2.

Depositional units and structures:

The lithofacies is remarkably monotonous, in that metamorphic recrystallization has resulted in the more massive metapsammities 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, 26c). However, impressive examples of these have been found at locality 94,1202 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 km² inland from Sammallahti (Map 2). The most representative exposures are 91.720.0 and 91.520.3.

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 91.331.7, 91.232.4 and 91.332.2.

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 metapsammitite 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 28c, 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 metapasmmites 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 KETTAMÖ ASSEMBLAGE

3.3.1 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 29a,e) with additional granitoid clasts that seldom exceed 10 cm in maximum dimension. At locality 94.928.1 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 29c-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 29c and d (locality 94.528.5) 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 Petäjikkökkallio 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 metaarkoses 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 diamicite 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 30d).

Of significance is the fact that these coarser beds are not the most basal but overlie the finer deposits depicted in Figure 29f; 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 30b.

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 Petäjikkökallo subfacies is recognized.

**Petäjikkökallo 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 (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 94.627.3 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 Niittylahdenrantatie 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 S₂-3 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.
Depositional units and structures:

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

Distribution and reference exposures:

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.

Depositional units and structures:

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 metapsammit to metapelite. Nevertheless, rhythmic alternations of these, with metapsammit 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 any 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 reconnaissance 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 33c) 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. Representative exposures are E of Haapajärvi on the Joensuu-Lappeenranta highway (95.7 33.2) and 94.8 37.3 at Mulonsalo.
Depositional units and structures:

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 34a, 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 04.6 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_{abcd,e}) 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 metapsammitite (Figure 36b). 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 36e, 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 36f). 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 36e). 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 38a, b). No granitoid or extrabasinal detritus was positively identified in the field. Where clasts appear very elongate, as above the hand lens 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, interbedded 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 37c). 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 meta­psammite frequently occur, (Figure 37a). 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 37b which may be examples of the latter. They differ from typical basal load 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 metaclasts 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 Petäjikkökkälö.

A wide region to the W and N of Tikkala contains very few bedrock exposures. 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

Distribution and reference exposures:

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 39c.

Depositional units and structures:

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 metapsammite 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 metapsammites are also typical, and the lithofacies is characterized by alternations of these two lithologies, usually 2-20 cm thick but with psammitic: 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 metapsammitic to metapelitic 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 KANKAAALA 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 lithologies 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 perpendicular 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)

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

b Tectonic distortion S₂ 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).
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. (Pyhäsälä 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. (Pyhäsälä 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. (Leppälähtö 69.6 29.5).

f Poorly sorted diamicrite 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 29: KARVANURMI AND PITKANURMI SUBFACIES

KARVANURMI 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. (Pyhäsälä 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 (Pyhäsälä 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 metapsephite in general may therefore be reworked from comminution of these pre-existing pebbly sediments rather than derive directly from an exposed granitic terrane. (Pyhäsälä 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. (Pyhäsälä 94.5 28.5).

e Poorly defined, tectonically transposed intra-basinal clast; note coarser granules near base of clast, at bottom right of photograph. (Pyhäsälä 95.4 27.6).

PITKANURMI SUBFACIES

f Stratified metapsephite layer with high matrix to clast ratio overlying fissile metapsammite or metapelite. Intra-basinal metaclasts prevail over granitoid detritus. (Pyhäsälä 94.4 28.3).

g Intercalation of irregular, lenticular poorly sorted metapsephites with better stratified metapsammites and metapelites lacking in pebbly granitoid detritus. (Pyhäsälä 94.4 28.4).
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 F₃ 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).
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. (Pyhäskelä 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. (Pyhäskelä 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. (Pyhäskelä 95.2 27.2).
FIGURE 32: LENTOSÄRKKA SUBFACIES AND SALONKYLÄ SUBFACIES

LENTOSÄRKKA 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 (L3) 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 S4 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 F2) 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).
intrafolial $F_2$ fold with axial planar $S_2$

$S_2$ well developed as crenulation cleavage and differentiation layering in micaceous metapelites
massive, sometimes concretionary metapelites

$S_2$ cleavage and quartz veins parallel to lithological layering
$S_5$ crenulation cleavage
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).
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 Ta 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 pyrte 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 abundant 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).
FIGURE 38: TIKKALA LITHOFACIES

a, b  Poorly sorted metapeseephite 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. (Niitylahti 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).

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 S3 or S4
f boudinaged metagraywacke
g S3 crenulation cleavage and accompanying minor fold
h discordant quartz vein perpendicular to S2 and F2
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 metapsammite 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 metapsammitite. 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 metapsammitite (light-coloured) and darker metapelite alternations showing disharmonic F5 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 metapsammitite dominant and displaying a well developed S2 differentiation layering, crenulated by S5 and accompanying F5 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 diamicite 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 terrestrial (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 (94.120.2) have sedimentary structures within a bed been noted. These comprise megaripples (wavelength to amplitude ratio about 4) and foreset accumulations, 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-beded 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 C₁ (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 km²; it may thus approach a volume of 10 000 km³.

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 fluidized 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 metapsammites 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 intraformational 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, B₁ and C₁ 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 A₂).

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 F₂ fold, it cannot be seen in what manner the originally overlying strata recorded cessation of progradation and fan avulsion.
4.3.4 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 S₂ 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 F₂ 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 KETTAMÖ ASSEMBLAGE

4.4.1 Kettämönämmi 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 Petäjikkökkallio 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 lithofacies. 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 turbidity 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 metapsammites 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 31ashows 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 conglomerates. 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 therefore probably represent deposition from turbidity currents, 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, inclined F3 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 metapsammites 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 recrystallization, 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 $\text{O}_2$ 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$_2$O, 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 peralkaline 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 Mullo 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:

\[
\begin{align*}
\text{KA1Si}_3\text{O}_8 + \text{H}_2\text{O} & \rightarrow \text{HA1Si}_3\text{O}_8 + \text{K}^+ + \text{OH}^- \quad (1) \\
2\text{KA1Si}_3\text{O}_8 + 3\text{H}_2\text{O} & \rightarrow \text{H}_4\text{Al}_2\text{Si}_2\text{O}_9 + 4\text{SiO}_2 + 2\text{K}^+ + 2\text{OH}^- \quad (2) \\
\text{HA1Si}_3\text{O}_8 + 4\text{OH}^- & \rightarrow \text{Al(OH)}_4^- + 3\text{H}_3\text{SiO}_4^- \quad (3)
\end{align*}
\]

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 Mulolithofacies 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 $\text{K}^+$ 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 $\text{H}_2\text{S}$ as an electron donor.

(ii) Dissimilatory sulphate reduction may have been operative, using organic carbon and $\text{H}_2\text{O}$ as electron donors. Such a mechanism would also generate $\text{CO}_2$ and in the presence of $\text{Ca}^{++}$ or $\text{Mg}^{++}$ could lead to dolomite or calcite precipitation, again yielding $\text{H}^+$. High activity of $\text{H}^+$ 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 CO₂ 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 O₂ production; accumulation of O₂ was presumably buffered by abiogenic oxidation of S²⁻ to SO₄²⁻. Biogenic sulphate reduction can occur anaerobically (dissimilatory) generating CO₂ in the absence of O₂, or aerobically (assimilatory), which yields further O₂ and S²⁻ (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 $\text{CO}_2$, reducing sulphate, yielding $\text{S}^2-$ 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), $\text{O}_2$ 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 $\text{S}^2-$ 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.
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 precipitation 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 $\text{H}_2\text{S}$ 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 lithofacies 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 diagenetically 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.

4.5.4 V 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 S₃ and S₅ foliations. A detrital origin is however inferred from microscopically observed silt-sized quartz grains, preserved in spite of mica grain growth and reorientation. 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 leved valleys. According to Piper's review, at least half of the sediment in basin plains beyond the continental 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 (Figure34b). 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 dissimilatory 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" transgressing them. It is therefore concluded that some combination of mechanisms a) and b) is more logical and appropriate.

Interpretation of deposits as thin laminit-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.
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 alternatively 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 diagenetic 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 $\delta^{34}S$ values and a constricted range of isotope ratios in the ore with respect to country rock, as well as a negative correlation between $\delta^{34}S$ 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 ($S_5$). 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 33f). If this indicates a syngenetic to diagenetic 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, lithogeochemical 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 hydrothermal 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 hydrothermal brine circulation, causing leaching of adjacent strata, including consanguineous intrusive or volcanic horizons. These processes are considered relevant for example in the Pyhäsalmen-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 metabasite 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 irregularities could be of tectonic origin but depositional units such as the coarser metagraywackes of the Tikkala lithofacies 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 A 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 \( S_5 \) 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 are some 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. Never­
theless 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 meta­sediment 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)

<table>
<thead>
<tr>
<th>FLOW INITIATION</th>
<th>MAIN LONG DISTANCE TRANSPORT PROCESS</th>
<th>LATE STAGE MODIFICATIONS</th>
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<tr>
<td>FLUID TURBULENCE</td>
<td>LIQUEFACTION</td>
<td>IMPERATIVE GRAINFLOW</td>
</tr>
</tbody>
</table>

Walker (1979) fig 13

**Feed Channel**

- Feeder Channel
  - Flume Flow
  - Debris Flow
  - Fluid Turbulence
  - Liquefaction

**Flow Initiation**

- Low Concentration Turbidity Current
- Fluid Turbulence
- High Concentration Turbidity Current
- 3D Matrix Strength

**Main Long Distance Transport Process**

- Debris Flow
- Fluid Turbulence
- Liquefaction

**Late Stage Modifications**

- Debris Flow
- Fluid Turbulence
- Liquefaction

**Not to Scale**

**Multi and Ricci Lucchi (1974)**

- Inner Channelized Fan (Distributary System)
- Outer Fan (Outbuilding System)
- Thickening Upward Channel Fill Sequence
- Thickening Upward Prograding Outer Fan Sequence
FIGURE 41: SCHEMATIC DEPICTION OF DEPOSITIONAL ENVIRONMENTS

Proposed reconstruction of major depositional units and environments 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.
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 microstructures and the third, where pertaining to lithofacies interpretation, 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

5.2.1 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 quartzofeldspathic 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 lithofacies 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 (Gaäl and Rauhamäki 1971) and the Outokumpu district (Gaäl 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 Pyhäsälkä 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 Pyhäsälkä 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 metapsammitic 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, Gaal 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 acceptable 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 Kettämönniemä 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 graywacke 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 Pyhäsälkä 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 phyllosilicate 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 Hammuslahti 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 fractionation 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 unpublished 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:
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:

\[ \text{CIA} = \frac{\text{Al}_2\text{O}_3}{(\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}) \times 100}. \]
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 metapsammmites 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 biased 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 metapsammmites 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 Eu²⁺ 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

5.5.1 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 Grünenerfelder (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 $^{87}\text{Sr}/^{86}\text{Sr}$ 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 $^{87}\text{Sr}/^{86}\text{Sr}$ (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 Grünенfelder (1976, 1977) have however determined that old, highly metamict zircons may undergo substantial radiogenic lead loss when subjected to later metamorphism; since isotopic fractionation is unlikely, it is nevertheless 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 metasediments 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 $^\ast$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 Grünenerfelder (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.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Lithofacies</th>
<th>Sm$^{147}$/Nd$^{144}$</th>
<th>TCHUR (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S11A</td>
<td>Rauansalo</td>
<td>.1967</td>
<td></td>
</tr>
<tr>
<td>S11B</td>
<td>Rauansalo</td>
<td>.1617</td>
<td></td>
</tr>
<tr>
<td>S12</td>
<td>Tikkala</td>
<td>.1273</td>
<td>2329$^\pm$50</td>
</tr>
<tr>
<td>S13</td>
<td>Kukkupää</td>
<td>.1119</td>
<td>2169$^\pm$53</td>
</tr>
<tr>
<td>S14</td>
<td>Kukkupää</td>
<td>.1101</td>
<td>2156$^\pm$70</td>
</tr>
<tr>
<td>S15</td>
<td>Kettämönniemi</td>
<td>.1055</td>
<td>2322$^\pm$86</td>
</tr>
</tbody>
</table>

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 remained 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 TCHUR 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 originally 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 discrepancy 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 $T_{\text{CHUR}}$ model age and $\text{Sm}^{147}/\text{Nd}^{147}$ 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 $\text{Sm}^{147}/\text{Nd}^{144}$ ratios with respect to that of undepleted mantle (CHUR = 0.2067) 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 isotopic 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 combination 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. Metamorphic 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 metapsammitc 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 metapsammitc (Pyhäskä 91.7 20.0). NX.
c. Sutured and stilolitic 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 metaturbidite (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, Pyhäskä 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 palaecurrent 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**

6.2.1 **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
F₂ or F₃ 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 (Orkamo 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 F₃ 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 F₃ 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 S₂ and S₃ 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

6.3.1 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 metabasites 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, Kettämönäiemi 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 (Gaål 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 Kiikhtelysvaara and the Soviet border (Pekkarinen 1979a) but the southwards continuation of the province, through Jänisjärvi and Soanlahti to Ladoga shows evidence of overthrust relationships (Wegmann 1929, Hausen 1930). Likewise to the N, overthrusting is prominent in the vicinity of Koli (Väyrynen 1939, Gaâl 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 Kiikhtelysvaara, 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 Kiikhtelysvaara, 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 been subdivided 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 38d,e).
Although volumetrically abundant in Finnish Karelia the Kalevian has received little attention, mostly because of its complex structure, dearth of sedimentary structures and metamorphic 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 in the 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 Gaâl 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 metapsammites in North Karelia (Gaâl and Rauhamäki 1971, Gaâl 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 (Gaâl 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 therefore 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 lithofacies variations in the Höytiäinen province, though interpretations 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 SuHmmura lithofacies. Although a depositional hiatus is inferred between the uppermost Jatulian and basal Kalevian conglomerate at Kiihtelysvaara (Pekkarinen 1979a), this can result from transgression at the basin margin while deposition has occurred continuously 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).
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

Rääkkylä lithofacies metapsammites

Nittylaahdenrantatie and Sammalkahti lithofacies coarse clastic deposits

Volcanites, serpentinites and volcanogenic and chemogenic deposits of Outokumpu and Oravisalo assemblages

Volcanoclastic Rauansalo lithofacies

HOWITAIEN PROVINCE

Resedimented coarse clastic deposits of Kettämö and Tikkala lithofacies and Kühtelysvaara turbidite suite

Marine Jatulian volcanite, phyllite and dolomite

Vepsian and Suisarian coarse clastic, pyroclastic and porphyritic picrite lithologies in Soviet Karelia

Tohmajärvi igneous complex

Saniolan regoliths, glacigenic and terrestrial deposits

Jatulian mafic metavolcanite and metadolerite

Jatulian deltaic, and terrestrial orthoquartzite
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), Brůn (1980) and Brůn 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 deformational history (such as Preston 1954, Bowes 1976, Brůn 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 Brün 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, Gaâl 1964) and at Kotalahti (Gaâl 1980) can be excluded from consideration as mantled gneiss domes on the grounds of their allochthonous 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 envisaged 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 Ultramafic 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 contrasts 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 Brůn 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, Berthélon 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

2.1.1 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 (Gaâl 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 Gaâl 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 competitive - 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 relationships, 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 basement 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 metabasites 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 subduction 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). Petrographical and limited isotopic results show, however, that the Höytiäinen 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 (Gaâl and Rauhamäki 1971, Gaâl 1972, Halden 1982). This magmatism was probably initiated just prior to Svecokarelian deformational history since gneisses interpreted as the deformed relics 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 basement 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 lithologies 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 palaeomagnetic 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 \( ^{87}\text{Sr}/^{86}\text{Sr} \) (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 recycling and evolution of oceanic crust. - Iceland is perhaps the closest contemporary analogue. If spreading rates are conducive to maintenance 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.
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 destruction and deformation during the Svecokarelian orogeny. The onset of deformation may be linked to attempted and abortive subduction 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 transcurrent shear and en echelon basins resembles the late Neogene collision of the Australia-Papua New Guinea continent with the Banda Arc.

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.
APPENDIX: INDEX OF LOCALITY NAMES

Names of 1:20 000 map sheets are underlined, 1:100 000 sheets in upper case.

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REFERENCES


Folk R L , 1974 : PETROLOGY OF SEDIMENTARY ROCKS. Hemphill Publishing Company, Austin, Texas, USA.


Gebauer D, Grünenerfelder M, 1974: RB-SR WHOLE-ROCK DATING OF LATE
DIAGENETIC TO ANCHIMETAMORPHIC PALAEOZOIC SEDIMENTS IN SOUTHERN
FRANCE (MONTAGNE NOIRE). Contributions to Mineralogy and Petrology
47: 113-130.

DATING OF LOW-GRADE METASEDIMENTS, EXAMPLE: MONTAGNE NOIRE
(SOUTHERN FRANCE). Contributions to Mineralogy and Petrology 59:
13-32.

ZIRCONS FROM SOME UNMETAMORPHOSED TO SLIGHTLY METAMORPHOSED SEDIMENTS
OF CENTRAL EUROPE. Contributions to Mineralogy and Petrology 65:
29-37.


Glikson AY, 1981: UNIFORMITARIAN ASSUMPTIONS, PLATE TECTONICS AND
PRECAMBRIAN EARTH. Pp 91-104 'Precambrian plate tectonics',

Goulevitch J, 1980: STRATIGRAPHY OF THE KAPALGA FORMATION NORTH
OF PINE CREEK AND ITS RELATIONSHIP TO BASE METAL MINERALIZATION.
Pp. 307-318 in 'Proceedings of an international symposium on the
Pine Creek Geosyncline', IAEA, Wien.

Gray D R, 1979: MICROSTRUCTURE OF CRENUATION CLEAVAGES : AN
INDICATOR OF CLEAVAGE ORIGIN. American Journal of Science 279:
97-128.

Hackman V, 1933: SUOMEN GEOLOGINEN YLEISKARTTA : LEHTI D2 SAVON-
LINNA, KIVILAJIKARTAN SELITYS. Suomen Geologinen Toimikunta.

Halden N M, 1982: STRUCTURAL, METAMORPHIC AND IGNEOUS HISTORY OF
MIGMATITES IN THE DEEP LEVELS OF A WRENCH-FAULT REGIME, SAVONRANTA,
EASTERN FINLAND. Transactions of the Royal Society of Edinburgh:

Hargraves R B, 1981: PRECAMBRIAN TECTONIC STYLE - A LIBERAL
UNIFORMITARIAN INTERPRETATION. Pp. 21-56 in 'Precambrian plate

Hausen H, 1930: GEOLOGIE DES SOANLAHTI-GEBEITES IM SÜDLICHEN
KARELIEN - EIN BEITRAG ZUR KENNTNIS DER STRATIGRAPHIE UND
TEKTONISCHEN VERHALTNISSE DER JATULFORMATION. Bulletin de la
Commission géologique de Finlaned 90.

Helovuori O, 1979: GEOLOGY OF THE PYHÄSALMI ORE DEPOSIT, FINLAND.


Huhma A, 1975: SUOMEN GEOLOGINEN KARTTA. 1:100 000 KALLIOPERX-KARTAN SELITYKSET 4222 OUTOKUMPU, 4224 POLVIJÄRVI, 4311 SIVAKKAVAARA KARTTA-alueiden KALLIOPERX. Geologinen Tutkimuslaitos.


Kelling G, Holroyd J, 1978: CLAST SIZE, SHAPE AND COMPOSITION IN SOME ANCIENT AND MODERN FAN GRAVELS. Ch. 11 in 'Sedimentation in submarine canyons, fans and trenches', Dowden, Hutchinson and Ross Inc. USA.


Metzger A A Th, 1924: DIE JATULISCHE BILDUNGEN VON SUOJÄRVI IN OSTFINNLAND. Bulletin de la Commission géologique de Finlannde nr. 64.

Metzger A A Th, 1925: DIE KALKSTEINLAGERSTATTEN VON RUSKEALA IN OSTFINNLAND. Bulletin de la Commission géologique de Finlannde nr. 74.


Muti E, Ricci-Lucchi F. 1975: TURBIDITE FACIES AND FACIES ASSOCIATIONS. In 'Examples of turbidite facies and facies associations from selected formations of the Northern Appenines.


Piper D J W, 1978: TURBIDITE MUDS AND SILTS ON DEEP-SEA FANS AND ABYSSAL PLAINS. Ch. 12 in 'Sedimentation in submarine canyons, fans and trenches', Dowden, Hutchinson and Ross Inc. USA.


Veizer J, 1973: SEDIMENTATION IN GEOLOGICAL HISTORY: RECYCLING VERSUS EVOLUTION OR RECYCLING WITH EVOLUTION. Contributions to Mineralogy and Petrology 38: 261-278.


Veizer J, Compston W, 1-76: $^{87}\text{Sr}/^{86}\text{Sr}$ IN PRECAMBRIAN CARBONATES AS AN INDEX OF CRUSTAL EVOLUTION. Geochimica et Cosmochimica Acta 40: 905-914.


Wegmann C E, 1928: ÜBER DIE TEKTONIK DER JUNGEREN FALTUNG IN OST-FINNLAND. Fennia 50 nr. 16.


**STRUCTURAL SYMBOLS**

- **gneissosity in Archaean basement**
- **depositional layering, frequently modified by S₁ or S₂ differentiation layering**
- **S₂ differentiation layering in major fold hinge, at high angle to lithological layering and S₁**
- **S₃ continuous cleavage and differentiation layering axial planar to generally recumbent F₃ folds**
- **lithological layering and foliation from published maps**
- **S₄ crenulation cleavage indicating sinistral (S) or dextral (Z) vergence**
- **S₅ crenulation cleavage indicating sinistral (S) or dextral (Z) vergence**
- **weak disjunctive cleavages**
- **trace of thrust fault**
- **trace of high-angle fault coincident with S₅ and F₃**
- **L₃, defined as intersection lineation of S₀ and S₃**
- **L₃ as shape fabric lineation defined by elongate biotite aggregates or deformed clasts**
- **F₃ minor fold indicating plunge and upper limb (Z) or inverted limb (S) vergence**
- **lineation from published maps**
- **F₄ or F₅ minor folds with S₄ or S₅ as axial plane**
- **F₃ upper limb or Z-vergence deduced from minor folds or relations between S₀ and S₃**
- **F₃ inverted limb or S-vergence deduced from minor folds or relations between S₀ and S₃**
- **F₃ hinge zone**
- **depositional younging direction affected only by recumbent F₃ folding**
- **depositional younging direction indicates inverted zone prior to recumbent F₃ folding**
- **unconformity**

**LITHOFACIES EXPLANATION**

**SAVO PROVINCE**

**OUTOKUMPU ASSEMBLAGE**

- serpentinites, calcareous, granitic, 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

**ORAVISALI**

- Rääkkylä lithofacies
- Oravisoa with graniitic and thick-bedded thin-bedded lithofacies
- Karhunsaaari
- Kankaala hill

**HÖYTIÄINEN**

- Kettämöklä
- Salonkylä hill
- Kettämönnyt with quartz
- HAMMASL<br>

**KILPÄLÄNMI**

- Kilpelänkan, Suhmurila lithofacies
- Mulo lithofacies and sulfidic metapelite
- Kukkupää hill
- Tohmajärvi hill
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 Ss.

- Dextral, sinistral, or neutral F₃ fold vergence.
- Boundary between major F₃ domains.
- Downward facing zones inverted prior to development of recumbent F₅ folding.
- Trace of lithological layering.
- Access roads and tracks.
- Helsinki - Joensuu railway.
## Reference Grid for Map 1 (1: 200,000 Scale)

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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.
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Grid area corresponds to 1:20 000 sheet area (10 km x 10 km)