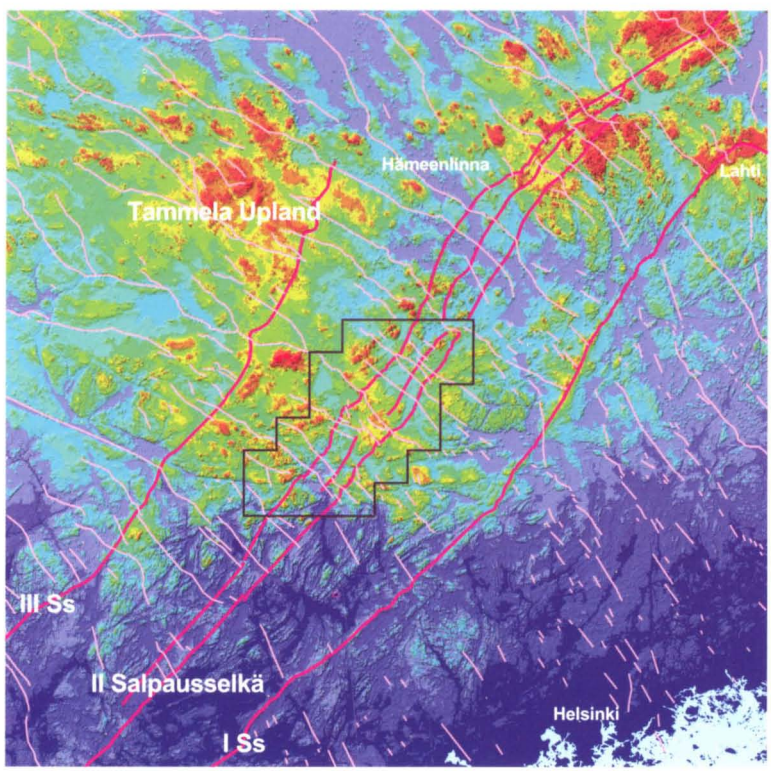
TUTKIMUSRAPORTTI
REPORT OF INVESTIGATION 148

Jukka-Pekka Palmu

Sedimentary environment of the Second Salpausselkä ice marginal deposits in the Karkkila-Loppi area in southwestern Finland



GEOLOGIAN TUTKIMUSKESKUS
GEOLOGICAL SURVEY OF FINLAND

Espoo 1999

Cover picture: Study area of Karkkila-Loppi in southwestern Finland. Topography of southwestern Finland visualised by means of height color classification and hillshading. Glaciofluvial routes and Salpausselkä-zones are delineated. Digital elevation model data © National Land Survey of Finland.

GEOLOGIAN TUTKIMUSKESKUS **GEOLOGICAL SURVEY OF FINLAND**

Tutkimusraportti 148 Report of Investigation 148

Jukka-Pekka Palmu

SEDIMENTARY ENVIRONMENT OF THE SECOND SALPAUSSELKÄ ICE MARGINAL DEPOSITS IN THE KARKKILA-LOPPI AREA IN SOUTHWESTERN FINLAND

ACADEMIC DISSERTATION

To be presented, with the permission of the Faculty of Mathematics and Natural Sciences of the University of Turku, for public criticism in Tauno Nurmela Auditorium, Main building, Yliopistonmäki on December 10th, 1999, at 12 noon

Espoo 1999

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This dissertation examines the depositional setting and sedimentology of the ice marginal deposits and landforms of the Second Salpausselkä dating to the Younger Dryas, in the Karkkila-Loppi area, southern Finland. The study area is situated in the southern part of the Tammela Upland.

Ice marginal deposits were investigated by means of observation of exposures, seismic sounding and ground penetrating radar logging. Quaternary mapping data (scale 1:20,000) of the Geological Survey of Finland was used in the field mapping that preceded studies on the internal structures of deposits.

The Second Salpausselkä in the Karkkila-Loppi area is composed of three zones that can be divided into further zones in part of the area. This study focuses on four sites representative of different types of ice marginal deposit.

The data assembled during this study indicates that it is primarily climatological factors that lie behind the exceptional geodiversity of the area and that glaciodynamics contributed to the wealth of deposits under particularly favourable conditions. Ice marginal water depth in the study area varied between 0 and 100 m, while most of the ice marginal deposits were deposited in 20-50 m deep water.

Practical applications of this study include groundwater geology and sand and gravel resources management. Methods of GIS visualisation have been elaborated in association with this study.

Key words (GeoRef Thesaurus, AGI): glacial geology, glacial features, ice-marginal features, Salpausselkä, deglaciation, depositional environment, glaciofluvial sedimentation, climate change, sedimentary structures, geographic information systems, Quaternary, Pleistocene, Younger Dryas, Karkkila, Loppi, Finland

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Tässä väitöskirjatyössä on selvitetty Nuoremman Dryas-vaiheen ikäisten II Salpausselän reunamuodostumien ja reunamoreenimuodostumien sekä -vallien syntyoloja ja sedimentologiaa Karkkilan-Lopen alueella Etelä-Suomessa. Tutkimusalue sijaitsee Tammelan ylänköalueen eteläosassa.

Jäätikön reuna-asemien muodostumia tutkittiin leikkaushavainnoinnin lisäksi koekaivannoin, seismisin luotauksin ja maatutkauksin. Muodostumien sisäisen rakenteen tutkimuksia edeltäneessä maastokartoituksessa käytettiin pohja-aineistona Geologian tutkimuskeskuksen 1:20 000 maaperäkartoitusaineistoa.

Karkkilan-Lopen alueella II Salpausselkä koostuu kolmesta osavyöhykkeestä, jotka osalla aluetta jakautuvat edelleen viideksi vyöhykkeeksi. Alueen leveys on keskimäärin 4 kilometriä. Jäätikön reunan veden syvyys on vaihdellut 0-100 metriin, suuri osa jäätikön reuna-asemien muodostumista on syntynyt syvyysvyöhykkeessä 20-50 m.

Tutkimuksessa valittiin esimerkkikohteiksi 4 muodostumaa kuvastamaan erilaisia jäätikön reunaan syntyneitä muodostumatyyppejä. Ulommassa vyöhykkeessä ovat Vuonteenmäen ja Multamäen havaintokohteet. Tupsumäen muodostuma on keskimmäisessä vyöhykkeessä ja Hirvilammin muodostuma on sisimmässä reunavyöhykkeessä.

Tutkimusalueella jäätikkösyntyisten reunamuodostumien ja reunamoreenimuodostumien sekä -vallien runsaus on poikkeuksellista eli geodiversiteetti on suuri. Jäätikön reunan alustan korkokuva ei ole aiheuttanut reuna-asemien suurta määrää. Ilmeisesti jäätikön reuna-asemien runsaus johtuu ilmaston vaihtelusta.

Tutkimuksen käytännön sovellusalueita ovat pohjavesitutkimukset, maa-ainesinventoinnit sekä luonnon monimuotoisuuden selvitykset.

Työn yhteydessä on kehitetty paikkatiedon (GIS) visualisoinnin menetelmiä. Kehitystyössä on keskitytty muodostumien korkokuvallisen sijainnin, muotojen ja eri tyyppisten aineistojen havainnolliseen esittämiseen.

Avainsanat (Fingeo-sanasto, GTK): glasiaaligeologia, glasiaalimuodostumat, reunamuodostumat, Salpausselkä, deglasiaatio, kerrostumisympäristö, glasifluviaalinen sedimentaatio, ilmastonmuutokset, sedimenttirakenteet, paikkatietojärjestelmät, kvartääri, pleistoseeni, nuorempi dryas, Karkkila, Loppi, Suomi

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CONTENTS

1. INTRODUCTION

1.1. General

This study focuses on the depositional environment at the ice margin of the Second Salpausselkä in the Karkkila-Loppi area in southwestern Finland, an area noted for the exceptional diversity of its glacial geological features.

In Finland, the Salpausselkäs form a concentric arcuate accumulation in two broad lobate zones, namely the Finnish Lake District Ice Lobe in the east and the Baltic Sea Ice Lobe in the west. The Second Salpausselkä is the innermost ice marginal accumulation in the Finnish Lake District Ice Lobe, and runs midway between the First and the Third Salpausselkä in the Baltic Sea Ice Lobe (Fig. 1).

The Salpausselkä ice marginal deposits are mainly glaciofluvial in origin. They often form long continuous ridges (Leiviskä 1920, Rainio 1991) along which individual deltas and sandurs can be recognized (Donner 1969). However, some of the deposits have been reworked by shoreline processes postdating the primary deposition (Fyfe 1990, Kujansuu 1993, 1995).

It is known that the Salpausselkä moraines (e.g.

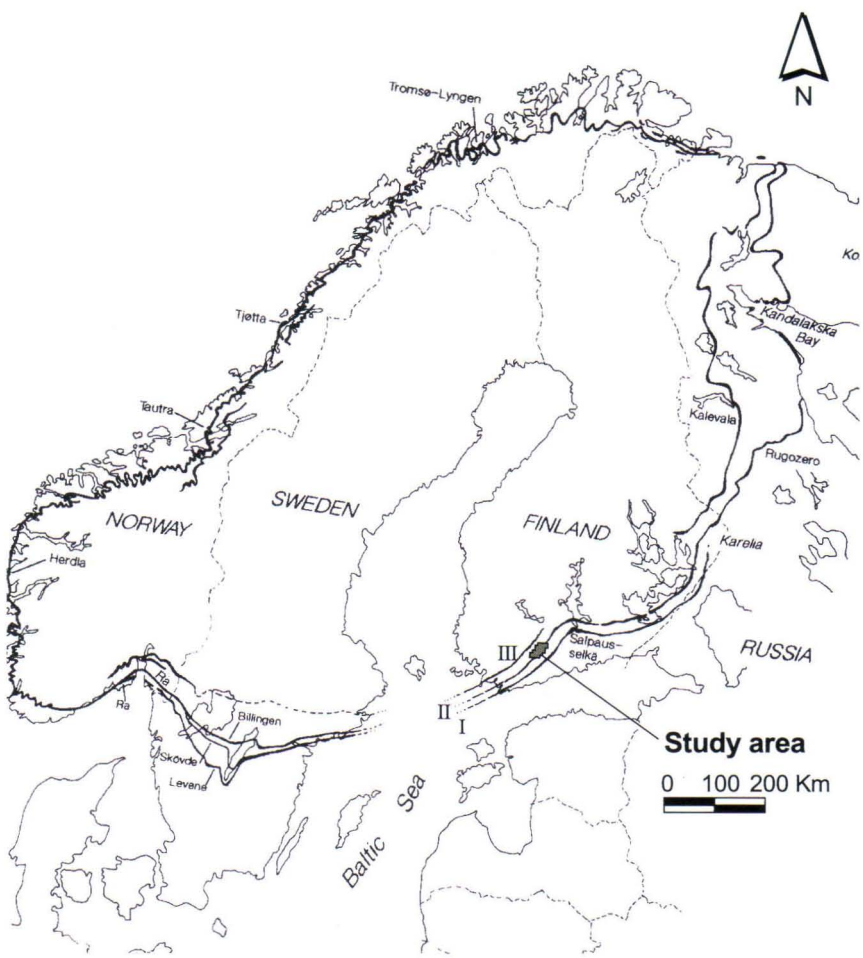

Fig.1 Location of the study area in relationship to the Younger Dryas ice marginal deposits (solid lines) of the Fennoscandian Ice Sheet (Andersen *et al.* 1995).
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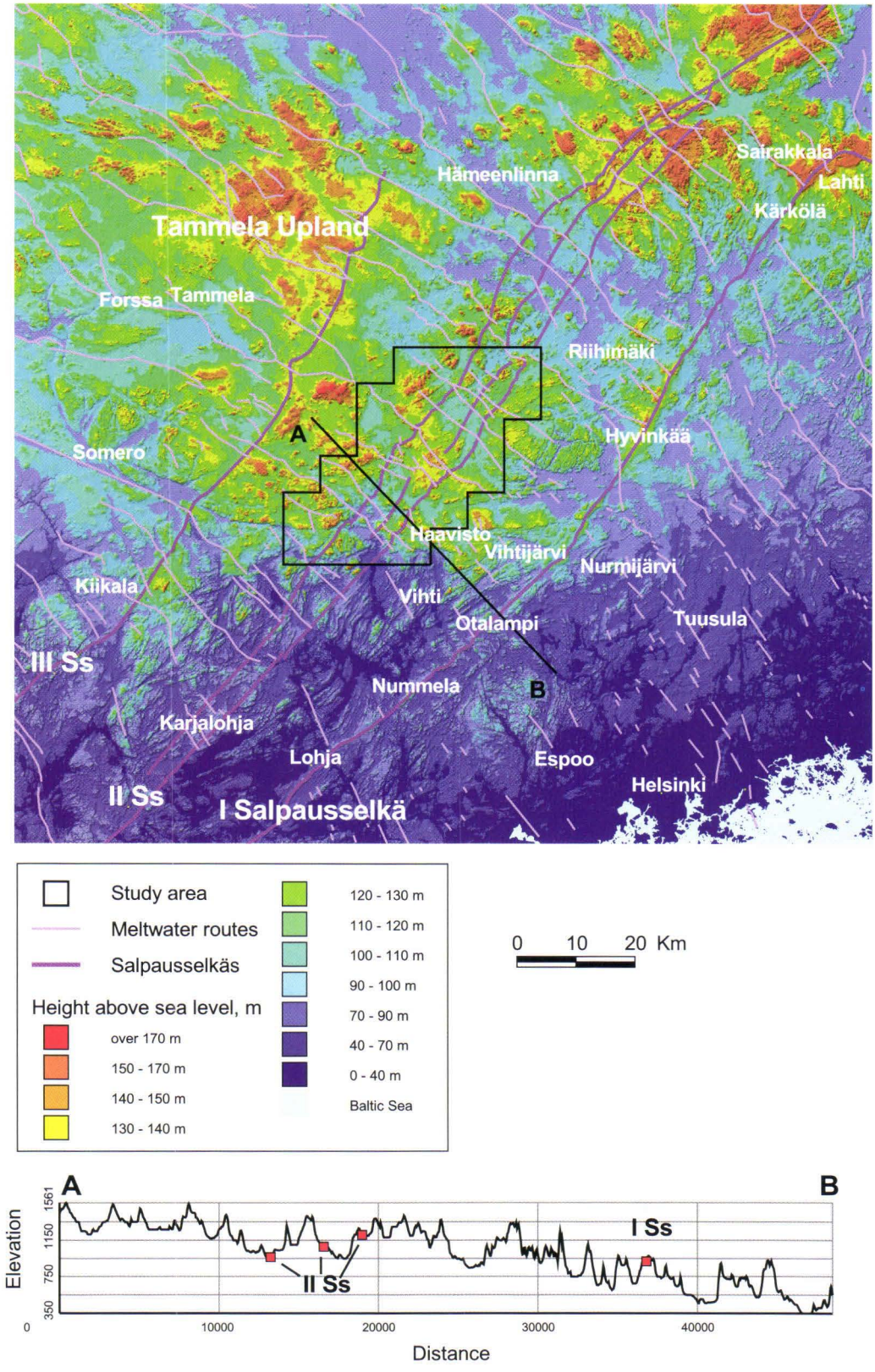

Fig. 2. Study area of Karkkila-Loppi in southwestern Finland. Topography of southwestern Finland visualised by means of height color classification and hillshading. Glaciofluvial routes and Salpausselkä-zones are delineated. Digital elevation model data © National Land Survey of Finland.

Leiviskä 1920, Rainio 1995) were formed during the deglaciation stage of the Weichselian glaciation. They are thought to have been formed in association with the cold climate event of the Younger Dryas. Ice marginal deposits of the same age are also found in eastern Karelia and the Kola Peninsu-

la of Russia, in Norway and Sweden as well. However, during the deposition of the First and Second Salpausselkä, no significant ice marginal deposition took place in the deep water of the Baltic Sea between Finland and Sweden (personal commun. B. Winterhalter 27.09.1996).

The study area is located in southwestern Finland about 80 km NNW of Helsinki, extending across the boundaries of the provinces of Uusimaa and Häme (Figs 1 and 2). Geographically, it is part of the southeastern Tammela Upland (Granö 1952, Aartolahti 1968). In the study area the Second Salpausselkä forms a 6-8 km wide zone of ice marginal deposits, mainly composed of glaciofluvial material and partly also of various diamictons.

The Karkkila-Loppi area is ideal for the study of glacial depositional environment because of its exceptional geodiversity of deglacial and sedimentary landforms and deposits. The term 'geodiversity' means here a wide range of conditions, processes, landforms and relief implying bedrock and superficial deposits leading to a wide range of natural environments and landscapes.

The area exhibits all the main features of the Finnish Quaternary geology, i.e. drumlins and eskers, clays typical of the coastal areas and an abundance of various types of ice marginal deposits that were produced at the ice margin in the Salpausselkä zone.

Previous studies on the Quaternary geology, deglaciation history, geomorphology and ice marginal deposits in the study area have been carried out by Moberg (1882, 1889), Leiviskä (1920), Sauramo (1923), Virkkala (1963, 1967), Aartolahti (1968), Niemelä (1971) and Glückert (1975). Glückert (1977, 1979) has studied isostatic uplift and shoreline displacement in the area.

The purpose of this research is to decipher and understand the behaviour of the ice margin in a shallow water environment during the Younger Dryas period. The research includes a detailed study of landforms deposited in a variety of ice marginal glaciolacustrine environments. The architecture and associations of depositional units within those landforms were examined in exposures and test pits.

The research first focused on the study of the properties of diamicton-dominated ice marginal deposits, and then on the sedimentology of sand and gravel-dominated ice marginal deposits and their structures. In particular, the Vuonteenmäki delta was studied in greater detail as it was being exploited (1990-1992). The other main sites were examined in 1992.

A secondary purpose of the research was to develop novel techniques of Quaternary geological mapping and research data visualisation. A data processing technique was developed that can be utilized in processing geological data, especially for groundwater, sand and gravel resources and for superficial deposit conservation purposes. The mapping data was processed with GIS (Geographical Information System) software, mainly Arc/Info and ArcView. The graphic material was processed using graphics software, mainly Corel Draw. The photographs were processed with image processing software, e.g. PhotoMagic.

1.2. Study area

The study area comprises the zone of the Second Salpausselkä extending across the Tammela Upland (Fig. 2), covered by map sheets 2042 01, 02C 04, 05, 06B, C, D, 08, 09, 12A and B of the Finnish Basic Maps. It is situated within the regional limits of the town of Karkkila and the municipality of Loppi.

Topography of the study area is shown in Figs 2 and 3. The highest point lies at Kaakkomäki, northeast of the parish of Loppi, at 184 m a.s.l. The lowest point lies in a river valley south of the centre of Karkkila where the elevation above sea level is 36 m. Mean elevation of the study area is 116 m a.s.l. Relative altitudinal differences are largest around the central Karkkila region, where the relief reaches a maximum of 100 m.

The bedrock of the area is Precambrian in age and crystalline in nature, with east-west trending lithological boundaries. In the south pyroxene gneisses and other pyroxene-bearing rocks include a narrow microcline granite zone, which grades northwards into a central zone of mafic intrusives. In the north mica schists and gneisses, mostly kinzigites (garnet-cordierite gneisses) and microcline granites, are the dominant rock type, with occasional remnants of gneissose granite. The central zone of mafic intrusives (gabbros and diorites) are cut by microcline granite, apparently along fracture zones. North of Lake Vaskijärvi an eastern fracture runs almost north-south, a second narrow-

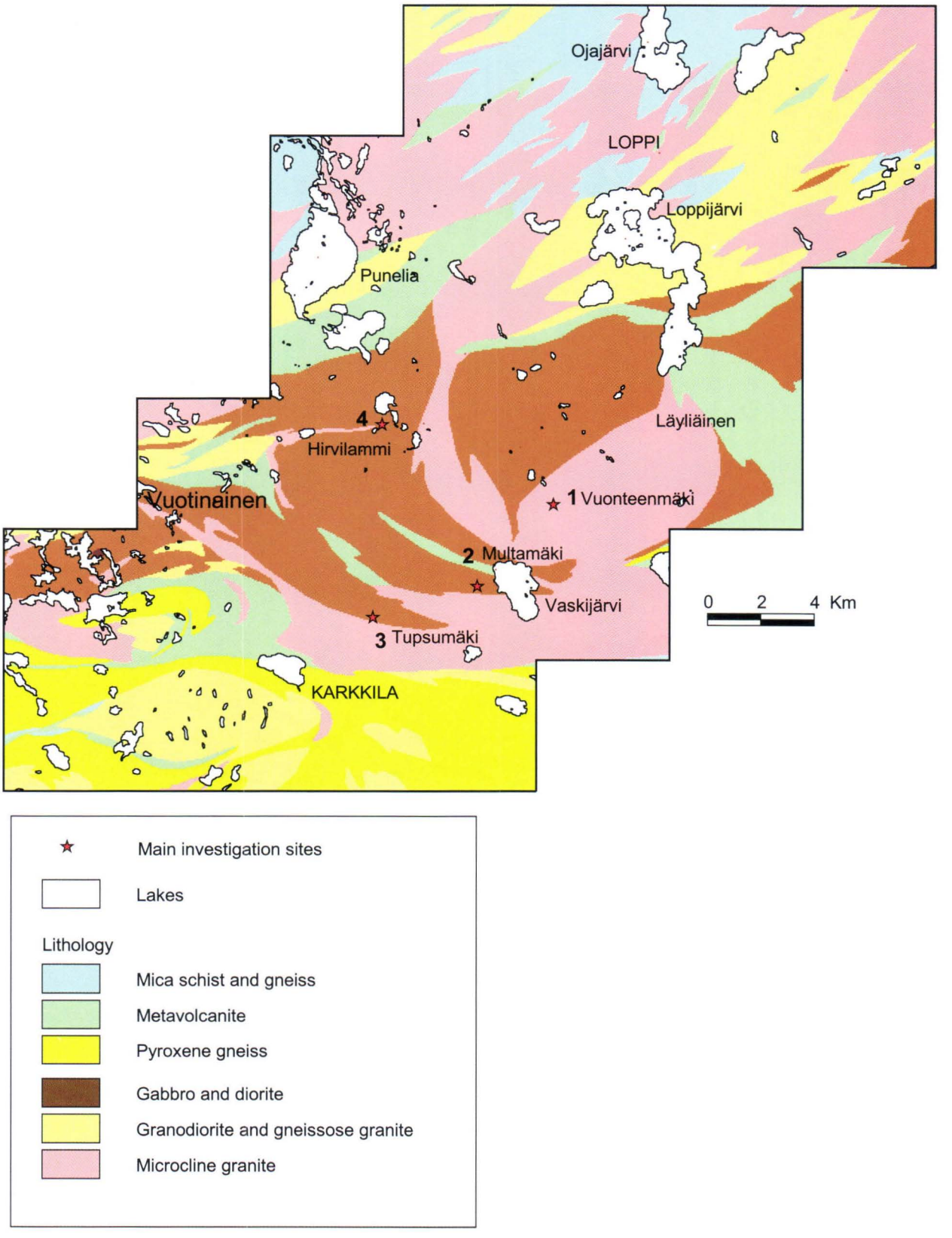

Fig. 3. Bedrock map of the study area based on the GTK bedrock map at 1:100, 000 (Härme 1954).

er fracture zone lies northeast of the centre of Karkkila, at Lake Vuotinainen (Härme 1954, p. 8). East-west fracture zones do occur in microcline granite, where they show up as a series of valleys (Härme 1954, p. 9). The thickness of the superficial deposits is greater in the granite area than in parts

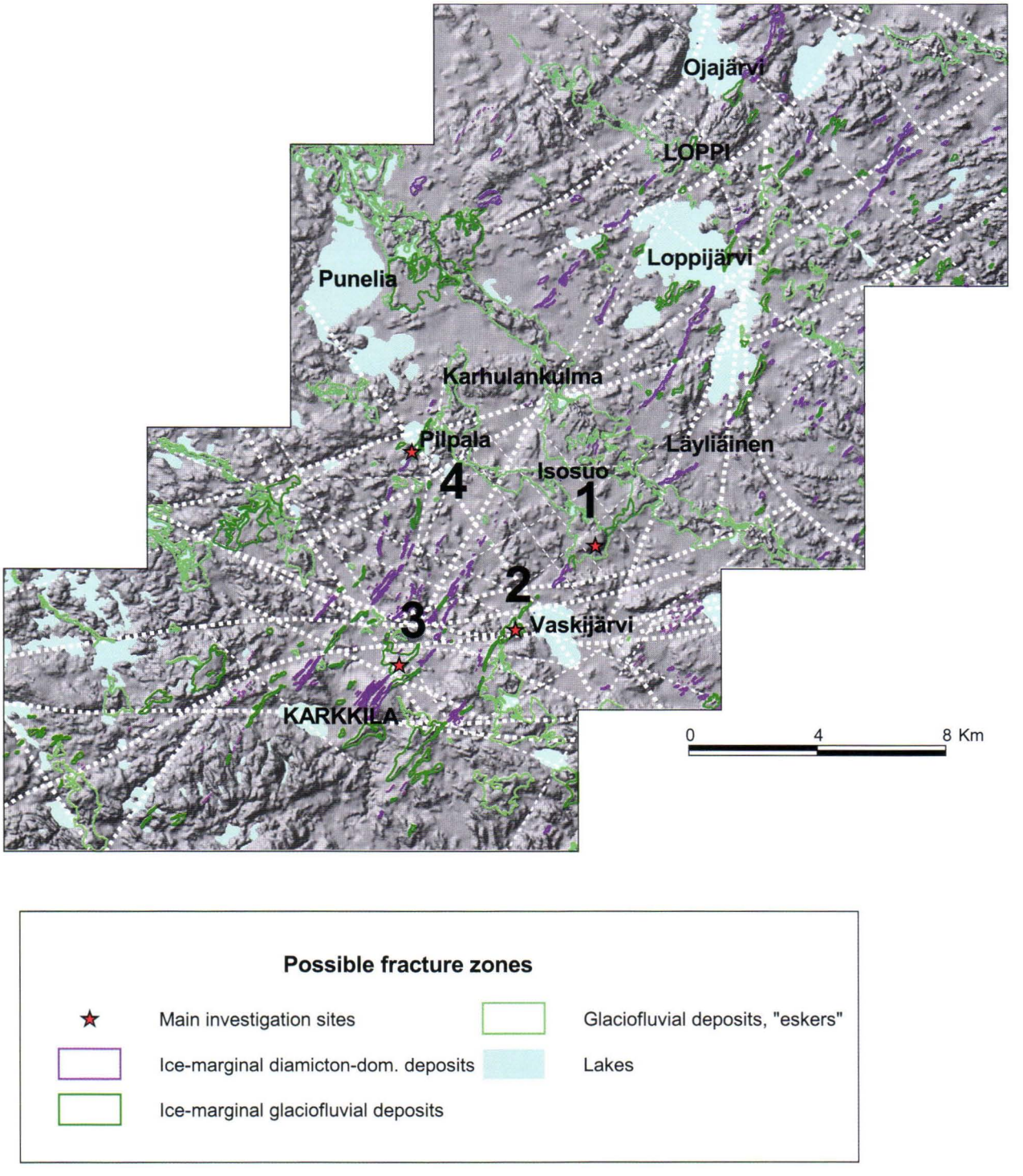

Fig. 4. Interpreted fracture zones indicated as dotted lines on a hillshaded relief map.
Elevation contour data © National Land Survey of Finland.

of the area underlain by mafic intrusive and metavolcanite (Härme 1954, p. 7)(Fig. 3).

A map of structural lineaments for the study area (Fig. 4) was prepared from the digital elevation model data and the low-altitude aeromagnetic data (GTK database). These lineaments reflect fracture zones that may influence bedrock topography. The strong fracture zones crossing the southern part of the study area in a west-east trend are considered the most important ones, as they also define to a great extent the topography of the surface in this part of the area. Other main sets are NW-SE oriented, this being the main structural trend of bedrock in southern Finland, and WSW-ENE, following the

direction of the supracrustal sequence. Moreover, WNW-ESE lineaments are particularly common especially in the area north of the centre of Karkkila (Fig. 4).

In the Esker area (Fig. 5) that constitutes the centre of the study area the location of part of the deposits seems to be associated with the bedrock fracture zones particularly around Isosuo-Karhulankulma (Figs 8 and 9). However this part of the region is covered by such a thick superficial deposit that interpretation on the basis of topography gives merely indicative information on the locations of fracture zones. Magnetic data interpretation does not reveal particularly strong fracture zones.

1.3. Superficial deposits

In southern Finland superficial deposits of Quaternary age are generally thin. The median thickness of the superficial deposits in Finland is 6.7 m (Okko 1964). This is reflected in an abundance of bedrock outcrops in the area (Kujansuu and Niemelä 1984). The small thickness of the superficial deposits is possibly due to the fact that the area may have been frequently and for long periods exposed to the erosional processes of a continental ice sheet. This speculation is based on the hypothesis that the Fennoscandian Ice Sheet's margin has frequently been positioned southeast and south of an erosional zone over southern Finland.

The study area is located centrally relative to the Baltic Sea Ice Lobe (Punkari 1980) margin, about 100 km southwest of the Finnish Lake District Ice Lobe margin lying east in the Lahti-Asikkala region (Figs 1 and 2). The fast flow of the lobe was seemingly restricted by the higher Ahvenanmaa region. Since the velocity of the ice movement is greater in deeper water, the ice margin extended further in western Uusimaa than further east (Sauramo 1923, p. 154).

Quaternary geology data is summarized as Fig. 5 where ice marginal deposits, ridges and eskers, along with striation directions and observation points are shown. Based on the observations presented here and for the purpose of this thesis, the Second Salpausselkä in the study area was divided into subzones, as shown (Fig. 5).

On the basis of striation data **the direction of ice movement** was consistently from 290-340°. The direction of ice movement lacks special features, i.e. deviating directions of ice movement older than the final stage of the deglaciation. The more westerly ice flow directions are measured in the Inner subzone where they may reflect the fact that the deglaciation took a more westwardly direction at the Pilpala valley region (Figs 5, 10, 47), where faster retreat in deeper water influenced the direction of the ice movement. The striation data has been acquired mainly during the Quaternary geology mapping program to include altogether 71 striation observations.

There are few investigations on transportation distance from the study area. Usually, debris is transported a relatively long distance before it reaches ice-marginal deposits. Perttunen (1977) has studied transport distances in the Hämeenlinna area, NNE of the study area, and calculated mean half-distance values of 3.7-5.6 km, which apply mainly to till areas.

Eskers in the study area will be presented in more detail in Chapter 3. Meltwater routes and associated eskers (Figs 5, 10 and 47) are divided into three groups on the basis of their scale and continuity. The first group is composed of large and continuous esker systems. Two such systems extend across the Second Salpausselkä in the middle of region where they form the Esker area of Fig. 5. The main esker is the Läyliäinen, or Maakylä-Porras, esker which forms a continuous ridge across the Second Salpausselkä zone (Figs 2, 10, 47). A second large esker is known as the Pilpala esker extending from Pilpala to Niinimäki (Fig. 5) The second group includes smaller and discontinuous eskers such as the esker traversing the parish of Loppi and the esker of the Siikala-Hajakka meltwater route. The third group consists of meltwater routes of intermittent deposition and stronger lateral variation of direction.

Shoreline displacement in the study area is adequately documented for the purposes of this study. According to Glückert (1977, 1979), the highest shoreline level in the study area corresponds to the B III-stage level of the Baltic Ice Lake, around 140-150 m a.s.l. The highest shoreline level dropped by 25-28 m within a short period of time when the Baltic Ice Lake drained and Yoldia phase began. This event is clearly seen in the delta levels, highest shoreline erosion levels and littoral deposits including littoral ridges.

Water depth during deglaciation is calculated by combining the data from a digital elevation model and highest shoreline observations. The calculations are based on the updated records of Glückert (1977, 1979) and my own observation records. In valley basins postglacial deposits and those depos-

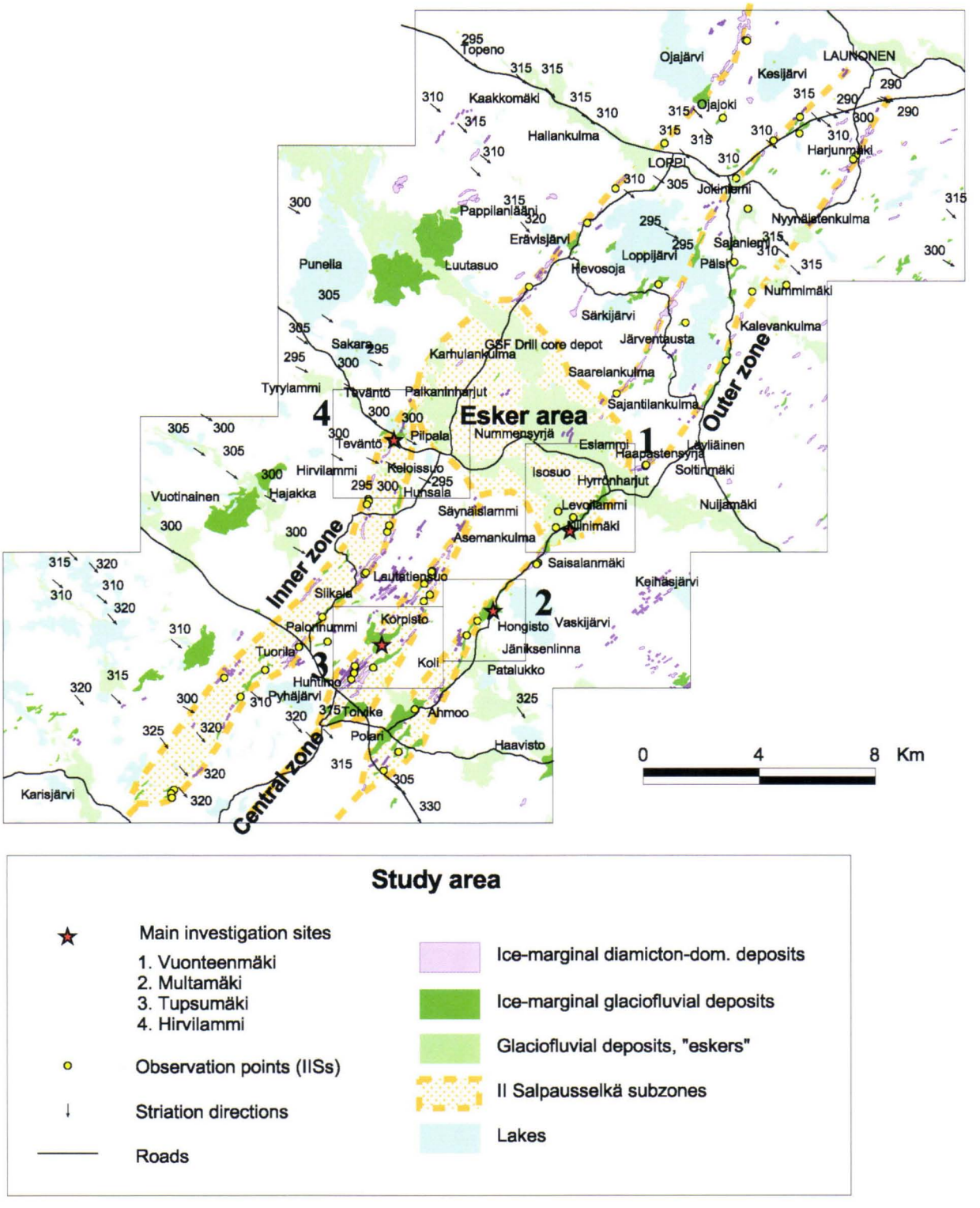

Fig. 5. General map of the study area, where the main glacigenic features, investigation and observation sites are also shown. Base map © National Land Survey of Finland.

ited during the later stage of the deglaciation constitute a source of error in the results.

Late glacial or postglacial clays and silts occur especially around the Karkkila centre and Läyliäinen (Fig. 5). Drilling records show that the thicknesses of clay and silt deposits rarely exceed 10 m in the study area. The maximum thicknesses of 10-20 m are found in the Siikala area.

Postglacial processes had only a relatively small impact on the topography of the study area. **Wave**

action reworked the area for a short period only. **Aeolian deposits** are common in the area. Dunes are found for example south of Pilpala. The study area includes also extensive areas of aeolian cover sand, thought to have been deposited during the early Yoldia stage.

1.4. Research methods

1.4.1. Mapping of the Quaternary deposits

The fieldwork for this investigation was carried out during the years 1987-1993, mainly during the field seasons of 1987, 1991 and 1992. Some of the research was carried out as part of the Quaternary geological mapping program of the area (1987) at a scale of 1 : 20 000 (Haavisto-Hyvärinen 1990 a to j) and the updating program of the sand and gravel resources database (Niemelä & Tynni 1972, Kurkinen *et al.* 1974). The Quaternary geology mapping carried out during this study concentrated on ice marginal zones, meltwater network and the continuity of ice marginal deposits as well as glaciofluvial esker systems. Bedrock outcrop data were used for the determination of depositional base level. The Quaternary geology mapping and revision were carried out according to Maaperäkartoituksen kenttätyöopas [Field work manual for Quaternary geology mapping] (1982) and Soraarvioinnin opas [Sand and gravel inventory manual] (1973).

The topographic maps (Finnish Base Maps, scale 1:20,000) used here are from the National Land Survey of Finland. Maps, mainly from the 1970's and 1980's, some from the 1960's, were used. Older maps provided geomorphological information on areas now quarried away. Topographic maps at 1:10,000 were also used for part of the area. The Orientation Society Karkki-Rasti (M. Pajuoja) kindly provided me with highly useful maps with additional information on the detailed geomorphology (e.g. contour lines) of parts of the area.

Stereographic aerial photographs from the Karkkila-Huhtimo area were used during the Quaternary geology mapping (G. Glückert, Turku University). In spring 1996, the Topographic Service of the Finnish Defence Forces provided aerial photographs on the type-deposit sites. These aerial photographs were 1:5,000 print-outs from high altitude aerial photographs, scanned and registered to a coordinate system.

All map data were digitised and transferred into GIS-format during 1992-1995. The base maps, contour line data and water routes vector data were obtained from the National Land Survey of Finland. The Quaternary geology maps were digitised by Hilkka Saastamoinen in the Geological Survey of Finland, Espoo, using the Fingis software, whereafter the data was transferred into the Arc/Info system to be processed and updated. The orientation map contour line data from around the Multamäki site was digitised.

1.4.2. Sedimentological studies

In this study, ice-marginal deposits have been classified into 1) ice marginal deposits dominated by sand and gravel i.e. glaciofluvial sediments and 2) diamicton-dominated ice marginal deposits, following the classification used by Donner (1976, p. 55) and the classification used in Quaternary geology mapping by the Geological Survey of Finland (Haavisto 1983).

Glaciofluvial ice-marginal deposits that accumulated at the highest shoreline level are called ice-marginal deltas. Incomplete deltas are called embryonic deltas. The term stratified moraine is also applicable, i.e. a linear sediment body formed parallel to the ice-margin (Ashley *et al.* 1991).

Glaciofluvial deposition took place in three (successive) sedimentary environments. 1) subglacially in ice-tunnels; 2) Subaqueously in fans that were deposited at the mouths of ice-tunnels; and 3) partly subaerially and subaqueously in deltas. In deltas, glaciofluvial traction brought about by fluvial processes caused deposition of material as (delta) bottomset, foreset, and topset deposits. In the study area, ice-tunnel deposits are found mainly in the Esker Area (Figs 5, 47).

Detailed field work included observation of basal till and diamicton exposures, examination of glaciofluvial deposits and stratigraphic successions, and detailed three-dimensional mapping of selected ice-marginal type-deposits accumulations.

Till (diamicton) deposits were investigated both within the study area and surrounding regions as well (map sheet 2042)(Palmu 1990a). In the study area, 5 exposures were examined in detail and 30 new test pits were dug in the ice marginal deposits of the Second Salpausselkä, including seven (7) sand and gravel (glaciofluvial) sites (Palmu 1990a, b). Particle size analyses were performed for 42 large (16 kg) and 13 smaller samples (Palmu 1990a, b). Careful observation and recording of diamicton units preceded the definition of its genetic type. Details observed were: nature and distribution of sorted intraclasts, fold tips (slide units), waning of

fines, variation in richness of stones, detached pieces/blocks of silt and sand transported by a glacier, and various kinds of rip-up structures.

All existing ice marginal deposit pit exposures during the summer 1991 were examined. Relative to the 1987 observation sets, there were 30 new observation sites, from which 46 sections were described. Stratigraphy was observed and the sites were documented with photographs and section drawings.

Four main ice-marginal accumulations were selected as type-deposit sites. They are 1) the glaciofluvial delta on the Esker Area at Vuonteenmäki (Fig. 5), 2) and 3) the sand and gravel dominated (glaciofluvial) ice marginal deposits at Multamäki and Tupsumäki, and 4) the Hirvilammi diamicton-dominated ice marginal deposit (Figs 5, 10).

The four type deposits sites were mapped three-dimensionally. The sections and observation sites were mapped using 1:10, 000 base maps, a Nikon-tacheometer, a field data logger and GT software. Measurements were taken on 12 occasions at Vuonteenmäki. Over 2, 500 measurement points including 2, 000 pit face measurement points were used to construct a three-dimensional data network/ system. The contacts of the sedimentary units, upper margins of pit faces, and upper and lower talus boundaries of the lower part of the face were measured. The tacheometer was used once at Multamäki and two times at Tupsumäki. At Tupsumäki a total of 124 pit face, test pit and log points were measured. Moreover, a 600 point field mapping was carried out in the eastern part of Tupsumäki. A total of 123 measurements were taken in one go at the Multamäki site.

The three-dimensional structure of the Vuonteenmäki site was measured tacheometrically repeatedly as quarrying progressed and the exposure face changed. This was accomplished by observing and measuring pit face locations and unit contact planes at given intervals of time.

The location of the sections at Hirvilammi were established using 1:10, 000 maps. Both 1:10, 000 maps and aerial photography were used to locate the western Vuonteenmäki observation points.

1.4.3. Pit exposure observations

Pit exposures were first examined for determining the overall stratigraphy and the vertical and horizontal unit relationships and recognition of large-scale structures such as erosional surfaces and channels. The exposures were photographed and the profiles sketched. Based on the large scale observations, logging points and exposures were selected for more detailed examination.

The exposures and profiles were sketched in diaries and sketching forms, some on plotting paper. Additional drawings of details were also made. Drawings, photographs laid out as mosaics and lithofacies logging data were combined after fieldwork together with particle size data presented as histograms, and palaeocurrent data presented as arrows and rose diagrams.

Most observations are from already existing exposures in the area. At Vuonteenmäki, due to progressive exploitation, observations were made repeatedly on the exposures on 25 occasions. The available exposures differed greatly.

At the Tupsumäki and Multamäki sites, most observations were carried out at one occasion. Exposures were cut back and freshened to allow for more efficient observation. In addition, test pits were dug by excavators, which normally reach down to a maximum depth of 5 m. Consequently, only the upper parts of the lithostratigraphical sequences at the test pit locations could be examined.

At Tupsumäki an earlier (1987) observation from the southwestern exposure section as well as five limited observations were made before summer 1992. In the actual sand and gravel pit at Tupsumäki there were 9 logging sites, and 5 loggings were made in test pits. At Multamäki, in addition to the exposure observations, 8 logs were made. In the Hirvilammi ridge one transverse test pit and two other test pits were dug in addition to two earlier test pits.

1.4.4. Lithofacies logging

Lithofacies logging means graphic recording of sedimentation units and their contacts in sections.

The term 'facies' has been defined (Reading 1978) in the following way: lithofacies - a body of rock with specified lithological characteristics. However, it is generally understood that facies are the product of a particular sedimentary environment, or of specific depositional processes in that environment.

As for sedimentation units, the following characteristics were defined: 1) texture (including roundness, particle shape, sorting and fabric), 2) sedimentary structures, 3) bedding relations, 4) unit or bed thickness, 5) palaeocurrent indicators, and 6) three-dimensional geometry.

Lithofacies analysis is used for examining the nature of the depositional environment (Ashley *et al.* 1991, p. 107-126). In this study, lithofacies includes e.g. ripple sand, cross-bedded gravel or massive stony gravel. The facies can be further divided into subfacies if needed, as e.g. ripple sand subtypes. The facies can be further grouped into

associations or assemblages.

Facies change in lateral and vertical successions. Variation in facies reflects changing depositional conditions. A sharp contact between facies, and in particular an erosional contact, indicates that the sedimentary environments are potentially far spaced.

The lithofacies classification and description codes used in this study are shown in Table 1. This classification is based on previous studies on water-terminating ice marginal sedimentary environments, the most important of which in Finland is the study of the First Salpausselkä by Fyfe in 1990-1991. Others include that of Powell(1990) and Ghibaudo (1992). The descriptive code is based on Miall (1977, 1978) and Eyles *et al.* (1983, p. 397). The corrections developed by Waldemarson (1986, p. 11) and Möller (1987, p. 21) have been partly taken into account (Table 1).

Particle size was determined visually in the field. Samples were taken and the visual analysis was checked and completed by particle size analysis in the laboratory. The main particle size categories used in the lithofacies logging were clay and silt, grouped as fines. Sand was subdivided into fine, medium coarse and coarse sand and gravel as granule gravel and coarse gravel. Stony gravel constituted a separate class.

In the graphic presentation of the log, the width is proportional to the particle size (Miall 1977, Eyles *et al.* 1983). As particle size increases from clay to gravel the log increases in width. The sampling depth was limited and only part of the sedimentation units were sampled. The results of the analysis are given as frequency bars for particle fractions. Sampling depths are given at left of a log. Results from the palaeocurrent measurements are shown as arrows or, in case of over five measurements, as rose diagrams.

Table 1. Lithofacies codes.

Gravel = G		
Matrix-supported	m	
Clast-supported	c	
Massive	m	Gmm, Gcm
Stratified	s	Gms, Gcs
Cross lamination, planar	p	Gp, Gmp, Gcp
Cross lamination, trough	t	Gt, Gmt, Gct
Deformation	d	Gd
Cobble gravel	Cg	Cg...
Bouldery	B	B...
Sands = S		
Massive	m	Sm
Horizontal lamination	h	Sh
Ripple lamination	r (A,B,S)	Sr(A,B,C)
Cross lamination, planar	p	Sp
Cross lamination, trough	t	St
Deformation	d	Sd
Fines = F (Silt = Si, Clay = C)		
Massive	m	Fm
Laminated	l	Fl
Diamicton = D		
Matrix-supported	m	
Clast-supported	c	
Massive	m	Dmm, Dcm
Stratified	s	Dms, Dcs
Deformation	d	Dmd, Dcd

Examination of primary sedimentary structures provides information on the transport process and fluid and energy conditions that prevailed at the time of the deposition. In addition to the primary structures, glacigenic sediments include various syn- and postdepositional structures as well and these were also noted and described.

In graphic presentation of logs, internal structure of sedimentary units is described by sedimentological symbols. There are separate symbols for massive, cross-bedded and rippled structures as well as deformed structures.

Bed boundary is a layer where there has been 1) no deposition, 2) erosion, or 3) a change in depositional conditions. A four-part classification has been developed for the presentation of the nature of contacts between sediment bodies in lithofacies logging (Waldemarson 1986, p. 11): 1) erosional contact, 2)conformable - no erosion in the contact, 3) loaded contact, and 4) interbedded contact (Eyles 1983, p. 17). The conformable contact may be sharp or transitional.

A bed is a unit with a homogeneous or heterogeneous structure and texture, cyclically varying or systematically transforming. It is a sedimentation unit deposited in quasi-stable physical conditions. From this it follows that a sedimentation unit has no minimal thickness (Otto 1938, Cambell 1967).

Sedimentation units or beds can be distinguished from the lower or upper unit or bed on the basis of their different texture and structure. The thickness of a sedimentation unit is measured vertically against the sedimentation plane.

Many of the several depositional or erosional structures that are formed when the sediment were laid down reflect the direction of water flow. Consequently, palaeocurrent data facilitates the interpretation of depositional conditions. A total of 120 palaeocurrent measurements were made, with 108 from Vuonteenmäki, 8 from Tupsumäki and 4 from Multamäki.

Most palaeocurrent measurements were taken from the foreset beds and laminae of cross-bedded structures. The orientation and angle of maximum dip of the foreset beds and laminae were measured from exposures. A dip-direction indicator makes

the measuring easier (Potter and Pettijohn, p. 99). It consists of a round plane which is placed on the foreset sedimentation plane and turned until the bubble level shows the steepest dip direction. The dip-direction was measured with a magnetic compass (and corrected for the angle between magnetic and true north). At least 10 readings were usually taken from each single measuring point. Means were calculated for the readings, and rose diagrams indicating flow direction were drawn using GeOrient and StereoNett software.

General paleocurrent indications were obtained from clast-fabric measurements and lateral grain size grading. Four two- and three-dimensional fabric measurements were taken at the Hirvilammi and Tupsumäki type-deposit sites.

Structural deformation of the deposits may indicate glacial flow direction. Since all type deposit sites showed deformation, especially that of Tupsumäki, a limited number of structural geological measurements were also taken. The glacial flow direction, i.e. the direction of stress was determined from glaciotectonic structures.

A Nikon FM 35 mm SLR camera was used to take over 2,000 photographs, ranging from large-scale pit face mosaics to detail photographs (width 30 cm). Paradox database software was used for managing the photograph data. The key photographs were scanned and saved in Kodak CD-format for further processing.

1.4.5. Processing of particle size analysis data

Samples for particle size analysis were taken from one sedimentation unit at a time, often at a specific depth. The types of constituent particles determined the sample size. Gravel samples of 16 kg were required for a reliable particle size analysis (the maximum gravel particle size to be analyzed is 64 mm). The 16-64 mm material was coarse sieved in the field (Palmu 1990 a, b). Samples of 2-4 kg with material of <16 mm were used for the actual particle size analysis. A total of 75 samples were taken, including 18 large samples of over 16 kg and 10 samples for the diamicton matrix analysis. Seven samples that had been taken in 1987 from the type deposit sites were also used.

The percentage of stones (particle size 60-600 mm) was assessed both by surface 1.0 m^2-net (10x10 cm- squares) analysis (Hörner 1944, Palmu 1990a, b) and also by visual estimates. The net analysis on normally rather loose glaciofluvial sediments presents the problem of stones falling from the surface under examination. However, the method has been successfully used by Eriksson (1960) for determining stone percentage of glaciofluvial material.

Samples for the particle size analysis were processed according to the Finnish GEO-standard (GLO-85, 1985). The guidelines presented by Head (1980) were also observed. The limit for fines was set at 0.063 mm. A sieve series of 63, 32, 16, 8, 4, 2, 1, 0.5, 0.25, 0.125, and 0.063 mm was used. The process of a particle size analysis is shown in Figure 6.

For the hydrometer test, 100 g diamicton samples with particle size less than 2 mm, and 50 g samples of fine sorted material were used. With small samples the hydrometer test was followed by wet sieving of 100 g samples in order to determine the particle size of thin diamicton or silt units. The hydrometer test dispergant was 1M pyrophosphoric acid.

Samples with less than 5% of fines were dry sieved, and the rest was wet sieved. The larger samples with gravel of 8-16 mm were sieved in as many as six smaller quotas to prevent sieve blocking.

The samples taken from coarse-grained sediments with particle size of over 16 mm were usually large, 15-20(30) kg. Total weight and proportions of 16-32 and 32-64 mm material were recorded. Water content was determined for material of <16 mm if considered necessary. Manual separation of finer materials from the surfaces of the 16-64 mm particles by brushing was one of the most laborious fieldwork operations.

For the data processing of the particle size analysis the software of The Geological Survey of Finland was used. Hydrometer data was computed first and then combined with wet sieving results. For coarse grained materials, the field sieving results were included in the final sieving data. Microsoft Excel (version 5) and Corel Draw (versions 5 and 7) software were used to produce grain size fraction histograms.

1.4.6. Seismic soundings, GPR-measurements and drilling

During the Quaternary geology mapping in the Esker Area in 1987 refraction seismic survey was conducted over a distance of roughly 200 m in total(Geological Survey of Finland, Seppo Koho). In 1987, I measured 7 ice marginal deposit sites with the 2-channel Geometrics equipment of the Department of Quaternary Geology of the Turku University. In addition, the FINNRA Uusimaa District provided seismic data for 500 line metres from Vuonteenmäki.

GPR (ground penetrating radar)-profiles were measured in July 1987 in connection with the Quaternary geology mapping and during the fieldwork for this dissertation in February 1990. GPR profiles were measured using the radar equipment of the Geological Survey of Finland, towed by a cross-country vehicle in the summer and by a

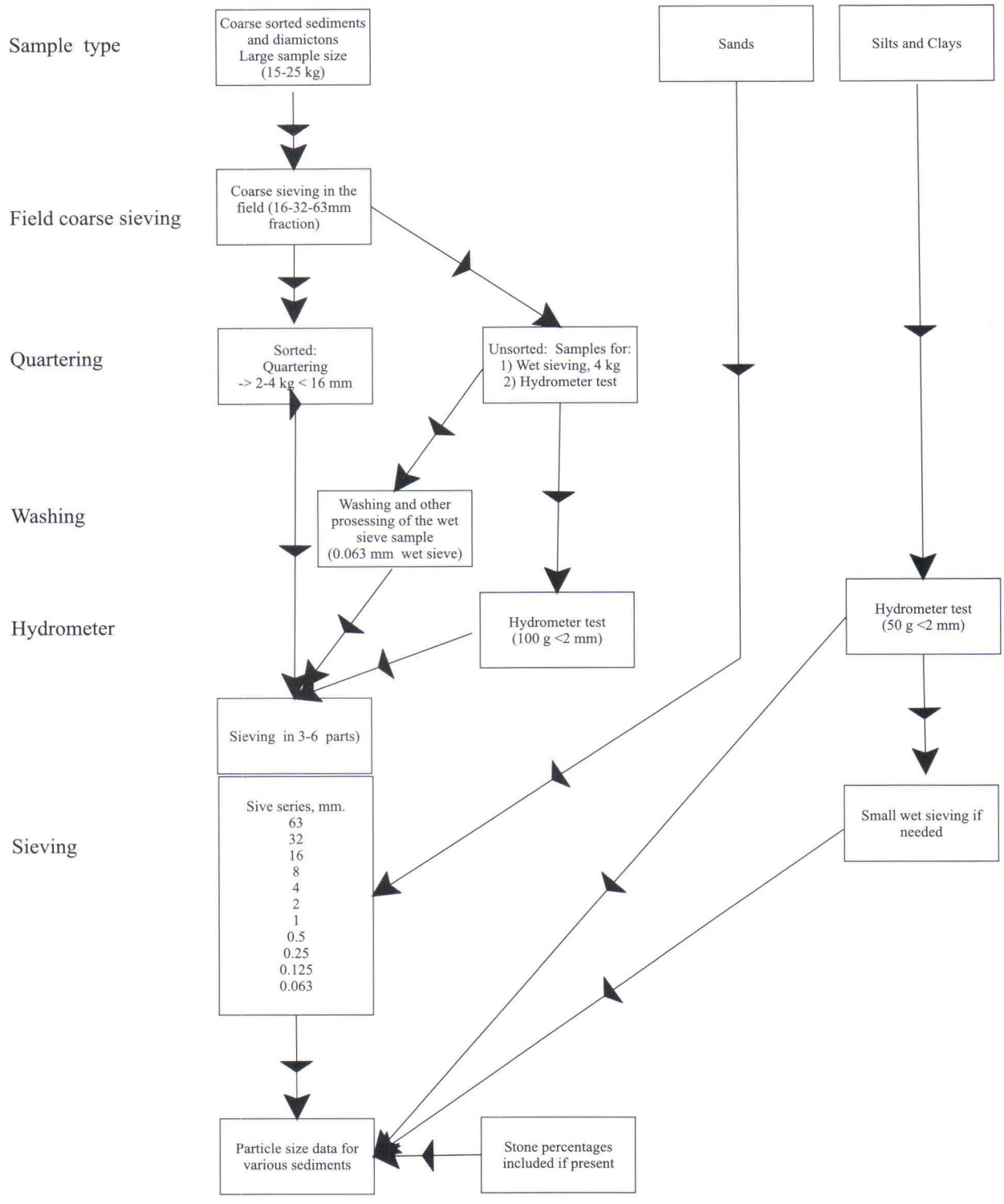

Fig.6. Particle size analysis process flow chart.

motor sled in the winter. Ice marginal deposits were GPR profiled for a total of 3 km in 1987, and for 5 km at Vuonteenmäki, including 1,300 m in a gravel pit owned by the gravel company Megasora Oy.

Most of the drilling data was acquired from the Vuonteenmäki site. The data was provided by the FINNRA Uusimaa District and the Megasora company, 22 and 8 drilling points respectively. Drillings were carried out at the Tupsumäki site for the company Tauno Valo Oy, but the data was not accessible.

1.4.7. Visualisation of data and results

One of the objects of this study was to develop geological applications of the GIS and visualisation software. The various kinds of geological data,

i.e. mapping observations, drilling data, geophysical data, pit exposure data, palaeocurrent measurements, and sampling data, is more convenient to manage and process if registered in database format. In addition, geological data is very well suited for being processed with GIS-software. Only some of the possibilities offered by the GIS-software were used in the present study, since most data has been presented in two-dimensional form. However, part of the data is presented as surfaces, i.e. in 2.5-dimensional form. In future, it will be possible to process and visualize geological data in true three-dimensional form with the so called voxel technique. A still more sophisticated level of data processing is available in the so-called four-dimensional visualisation, which combines a three-dimensional form of presentation with time to produce an animation. Visualisation of deposition processes with this method can be particularly illustrative. However, sufficiently comprehensive data rarely exists for the three-dimensional visualisation of a developmental history, let alone for the four-dimensional one. Alternative methods of visualisation are either to base the presentation on actual observation of phenomena restricted to time and place, or otherwise to adopt a more creative, free and dynamic approach with more interpretation and extrapolation to allow the larger-scale vision of the time and space aspect of the depositional events.

In this study, all the different geological data was digitised. Observation points and deposit boundary polygons are expressed as points and vectors. Fracture zones were interpreted on the basis of aerogeophysical and elevation model data, while tacheometer is the best tool for observation of sections. Aerial photographs can be scanned and registered to a coordinate system.

Partly as a result from this study, new techniques of visualising Quaternary geological and glaciogeomorphological data have been developed (Palmu 1997). Arc/Info software (version 7) was used to combine the mapping data of Quaternary deposits (updated) with other GIS data. The digital elevation model was constructed with the Topogrid component of Arc/Info software using contour data together with polygon and height data on lakes. For the visualisation of the raster data mainly the Grid module was used, and its gridcomposite hsv-technique for combining the mapping data of Quaternary deposits and the contour data (ESRI 1992).

Shoreline displacement data was processed to produce a water depth map of the area during the time of deglaciation. This was accomplished by first digitising the observation points, then by interpolating the data, and finally by combining the interpolated data with the digital elevation model to produce successive water depth maps (Figs. 10, 47). It is worth to note that the ground surface digital elevation model does not exactly correspond to the actual water depth contours in lake basins during the deglaciation. In the Karkkila-Loppi area problem areas occur mainly in the lowest-lying valleys and lake or mire basins. In these areas sediments usually not more than 5-10 m thick are found. Postglacial peat and gyttja are normally 2-4 m thick on an average and maximum thickness is 5-10 m.

The type-deposit sites were mapped by scanning the aerial photographs and registering them to a coordinate system. Topographic map and tacheometer data were combined with the aerial photographs. The tacheometer data was transferred from the field data logger into PC-format as export files. For data processing and elimination of errors GT software in PC environment was used. The data was then imported into Arc/Info in UNIX environment.

A novel technique for visualisation of elevation data was discovered (Palmu 1997). Difractive lenses allow the seemingly three-dimensional, i.e. pseudostereoscopic examination of elevation data, e.g. ground and bedrock surfaces. The impression is created through appropriate use of colour classification and hillshading effect. In this study this method is used to create relief visualisation for southwestern Finland (Fig. 2) and the four investigation sites (Figs 12, 23, 32, 41).

2. REGIONAL AND LOCAL QUATERNARY GEOLOGY

2.1. Deglaciation history

2.1.1. Deglaciation events

The deglaciation history of the region is summarized below, starting with events prior to the deposition of the First Salpausselkä.

The ice margin retreated slowly and continuously before the halt or readvance at the First Salpausselkä (Niemelä 1971). According to Holmlund and Fastook (1993), the ice margin came to a stillstand as a result of a change from a warmer to a colder

climate, with a 500 year delay in the glacier's response. Still, according to Holmlund and Fastook (1993), the delayed response was longer in Finland than in Sweden because of longer glacial flowlines and the more continental climate of Finland. Actually, the ice margin occupied a more proximal position than the model suggests (Holmlund & Fastook 1993). This may be due to more limited accumulation, greater ablation and calving, or a combination of these effects.

According to Lundqvist (1987), the lobate shape of the First Salpausselkä in its western part suggests that there probably was a readvance in the Baltic Sea Ice Lobe area; the readvance may have been triggered by a transgression from the g-level to the BI-level of the Baltic Ice Lake. However, the transgression hypothesis does not find support in the more recent studies by Fyfe (1990, 1991), which have shown that the decisive g-level deltas of the First Salpausselkä do not seem to have been deposited at the lake level after all.

Varve clay studies support the hypothesis of differential behaviour of the Finnish Lake District and Baltic Sea Ice Lobes at the time of the Salpausselkä event. Varve clay stratigraphy with a homogeneous interlayer has been observed in Koria by Sahala in 1992 indicating a readvance in the Finnish Lake District Ice Lobe. On the other hand, homogeneous varve clay stratigraphy series have been obtained from the distal part of the Baltic Sea Ice Lobe at Lappila, Kärkölä suggesting a stillstand or continuous retreat. These results suggest that the ice marginal position remained almost still in the SW, while a readvance took place in the NE (M. Okko 1962, Fogelberg 1970, Rainio 1993).On the other hand, varve clay stratigraphy (no. 16), cored by Niemelä in 1971 at Kalteva, Hyvinkää, exhibits a homogeneous interlayer indicating a possible Baltic Sea Ice Lobe's readvance or reactivation. The interlayer contains gravel clasts and intraclasts of clay suggesting re-sedimentation. Functional differences between the various parts of the Baltic Sea Ice Lobe may account for the contradictory results that exist between the stratigraphies of Niemelä (1971) and Sahala (1992).

Kärkölä is located in the more passive NE part, and Hyvinkää, on the other hand, is more central where readvance was greater although not nearly of the same order of magnitude as in the Finnish Lake District Ice Lobe. There is evidence from Hyvinkää that towards the end of the deglaciation the glacial retreat took a more westwardly direction on the proximal side of the First Salpausselkä, i.e. from a direction of 320-330° to 290-310° (Donner 1986, Kielosto *et al.* 1991).

Recent observations on fine sediments beneath the First Salpausselkä confirm that there was some functional reorganisation and readvance. Stay (1984) observed fine sediments in the Karjaa area. J. Ikäheimo observed a clay deposit in the lower parts of the First Salpausselkä at Nummela in Vihti (personal commun. 1997). There is also core data from 1998 on clay and silt layers about 20 m in thickness extending under the First Salpausselkä main ridge at Lohja (I. Ahonen and J. Ikäheimo, personal commun. 1998). Southwest of Hyvinkää at Noppo, a large intraclast of varved sediment with about 40 varves was found under the feeding esker about 20 m below the original ground surface.

North of Hyvinkää, the ice marginal readvance towards the First Salpausselkä is interpreted not to have exceeded 3 km (Palmu 1990, Kielosto *et al.* 1991). This oscillation may have caused the above mentioned varve clay interlayer observed in Hyvinkää. According to Fyfe (1991, p. 107), evidence shows that there was a readvance of roughly 1 km onto the proximal side of the First Salpausselkä in the Nummenkylä-Nummela-Ojakkala area.

On the distal side i.e. to the southeast of the First Salpausselkä, De Geer moraines are found east of Hyvinkää, trending diagonally with respect to the direction of the First Salpausselkä (Aartolahti 1972, Aartolahti *et al.* 1996). Discrepancies between the ice margin orientation and the suggested directions of ice movement within and outside the First Salpausselkä zone have usually been interpreted to be due to a glacial readvance forward to the position of the First Salpausselkä (Aartolahti 1972, Donner 1978, Donner 1986, Aartolahti *et al.* 1995). The different orientations of the De Geer moraines and the First Salpausselkä may also have been produced as a result of a change in the orientation of the ice margin during retreat due to the effect of water depth. The Nurmijärvi-Hyvinkää area lies at the SE margin of the Tammela Upland area where the ice marginal retreat may have been restricted by the relatively shallow water (Sauramo 1923, p. 134, 151). This area may have acted as a hinge, on the northeast side of which, in the deeper waters of the Hausjärvi-Kärkölä area, the ice marginal retreat was faster.

The deposition of the First Salpausselkä may have started earlier in the Baltic Sea Ice Lobe area than in the Finnish Lake District one. This hypothesis is based on the assumption that no such lengthy readvance took place in the Baltic Sea Ice Lobe as occurred in the Lake District Ice Lobe area (M. Okko 1962, Rainio 1984, Rainio 1993). According to Niemelä (1971) the deglaciation between the end of the deposition of the First Salpausselkä and the end of the deposition of the Second Salpausselkä lasted some 700 years in the Baltic Sea Ice Lobe

area. According to Brunnberg (1995), the Baltic Sea Ice Lobe ice marginal retreat rate was stable for around 600 years on the Swedish side of the Lobe before the Baltic Ice Lake/Yoldia stage time line. If there was any readvance, it must have taken place before this. According to Rainio (1993) the corresponding stage for the Finnish Lake District Lobe has taken 186 years. The deposition of the Second Salpausselkä has taken about 200 years in the study area, i.e. it was completed about 90 years before the Baltic Ice Lake/Yoldia stage time line (Niemelä 1971) which is estimated at ca 10, 740 Swedish varve years according to Brunnberg (1995) and at ca 11, 240 preliminary corrected Swedish varve years according to Wohlfarth (1996). A more detailed analysis of timing, duration and climatic controls of the deglaciation is presented in the following section.

In any case, the mass balance of the ice lobe shifted from a state of negative balance to a balanced or positive state during the deposition of the Second Salpausselkä. Only when the ice margin stood still or even advanced, could the ice marginal deposits have been produced. This was most likely due to decreased melting and its effect on glacial thickness (Paterson, 1981, Holmlund & Fastook 1993, p. 77). Other factors contributing to the stillstand are: 1) widening of the warm based zone, due to the Alleröd precipitation and thermal effects, leading to increased glacial flow velocity (Holmlund & Fastook 1993, p. 84), and 2) enhanced glacial velocity as a result of increased basal deformation, indicated by drumlinisation resulting from ice flow into the First Salpausselkä zone. Increase in flow velocity is a self-maintaining process as frictional heat induces faster flow (Paterson 1981).

There is no evidence of a larger scale ice marginal readvance in the study area or in its vicinity. The fact that there are no observations on glaciofluvial or other sediments covered by basal till that have been overrun by any extensive readvance imply that there was only a minor readvance. On the other hand in the Finnish Lake District Lobe area, basal till cover produced during the readvance of the Salpausselkä stage is widely observed.

Decreased calving produced slow ice marginal retreat followed by a standstill as the ice margin reached relatively shallow water. Evidently, the ice margin stabilises and glaciofluvial ice marginal deposits develop in shallow water provided there is abundant meltwater supply (Leiviskä 1920, Sauramo 1923, p. 151., Crossen 1991, Hunter *et al.* 1996). In the First Salpausselkä zone this phenomenon is observed e.g. in the Hanko Peninsula, Lohja, Vihti-Nurmijärvi (west of Röykkä), the Sairakkala and Hyvinkää areas (Sauramo 1923, p. 140-141), where the bedrock rises above the surrounding areas.

2.1.2. Glaciofluvial meltwater systems in the Salpausselkä zone

Baltic Sea Lobe meltwater routes in the region surrounding and including the study area are shown in Figure 2. The data is extracted from Quaternary geology maps, scale 1:100, 000 and 1:20, 000 and from digital elevation model data, hillshaded superficial deposit relief maps and Sand and Gravel Resources data. Topographical map (Finnish Basic map) interpretations and unpublished observations by the Geological Survey of Finland were also used.

Establishing the distribution of meltwater routes requires examination of the morphology and location of glaciofluvial deposits and signs of meltwater erosion, e.g potholes. Meltwater routes have been examined by Niemelä in Salo area (Niemelä *et al.* 1987) and Espoo (Niemelä 1987).

Morphological features, especially kettle hole distribution and orientation, reveal the distribution of segments of eskers deposited in tunnels (conduits). In and around the Tammela Upland area, individual contemporaneous meltwater route branches reached 10-20 km in length. The meltwater routes formed a dendritic system as the deglaciation advanced (Fig. 2).

Occasionally individual ice marginal glaciofluvial deposits would seem to follow a certain direction but were developed in fact as a result from obliquely migrating sedimentation, and are often found at the margins of higher areas. This type of formation is found for example south of the study area between Haavisto and Vihtijärvi as well as the oblique systems at Tammisaari in the First Salpausselkä forelands (Fig. 2).

In the forelands of the First Salpausselkä, the glacifluvial deposits west of the Tuusula-Vuosaari esker are characterised by individual hummocks and a lack of consistent tunnel originated eskers (Niemelä 1979). The deposits are mostly found on the lee sides of bedrock hills.

2.1.3. Glacial retreat from the First Salpausselkä

There are gaps in the First Salpausselkä in the area of the Baltic Sea Ice Lobe. These gaps are normally located in areas where transverse bedrock of high relief occurs proximal to the ice marginal position, e.g. at Otalammi and Kärkölä (Fig. 2).

Glaciofluvial sediments covered by diamicton of the basal till type are found on the proximal side

of the First Salpausselkä (Palmu 1990, Kielosto *et al.* 1991). The maximum range of oscillation associated with the First Salpausselkä in the Hyvinkää-Riihimäki-Hausjärvi area seems to have been ca 3 km.

The slow retreat of the ice margin continued from the First Salpausselkä. It seems that extensive areas of the ice sheet were at the pressure melting point, i.e. warm-based. This interpretation is based on the assumption that extensive and abundant ice marginal glaciofluvial sedimentation was possible at a warm-based ice sheet margin only. The ice margin during this deglaciation stage is also characterised by diamicton-dominated ice marginal deposits and ridges indicating ice marginal activity. Hummocky moraine fields were formed as stagnant ice melted *in situ* in higher areas. Where a supra-aquatic hill was deglaciated, diamicton-dominated ice marginal ridges graded into areas of hill top stagnant ice hummocky moraine, illustrated by the Lake Ylimmäinen area, southeast of the study area.

2.1.4. Depositional setting of the Second Salpausselkä

Distal slopes of the higher areas in southern Finland affected the ice margin behaviour, especially during stillstands. It seems that the ice marginal position of the Second Salpausselkä was influenced for example in the area of Karjalohja (Fig. 2), and that this ice marginal stabilisation had an impact as far as in the Karkkila-Loppi area.

The Second Salpausselkä forms three subzones in the study area. The maximum width of the entire zone is in the Karkkila-Loppi area (Fig. 2). Elsewhere in southwestern Finland the zone is more narrow and associated deposits are smaller in size than in the Karkkila-Loppi area.

In the Finnish Lake District Ice Lobe the Second Salpausselkä has a more consistent morphology compared to the Baltic Sea Ice Lobe occupying only one ice marginal position, which possibly results from a stronger readvance at the Second Salpausselkä there (Fogelberg 1970).

2.1.5. Final deglaciation during the Younger Dryas

The drop of water level ca 25 m that took place as the Baltic Ice Lake drained to the Yoldia phase level, is the most significant among the sedimentary environmental changes that took place during the deglaciation (Donner 1969, 1978, 1982, 1992, Glückert 1977, 1979). The ice margin became temporarily active in the northeastern part of the study area following the dramatic drop in the water level. As a result, small mainly diamicton-dominated ice marginal deposits were formed. Further in the NE, the ice margin became even more active (Okko 1957) producing a field of diamicton-dominated ice marginal ridges.

Deglaciation was mainly passive between the Second and Third Salpausselkä. Diamicton-dominated ice marginal deposits or ridges indicating ice marginal oscillation were formed only in exceptional conditions. Hummocky moraine was formed from stagnant ice in the supra-aquatic or shallow water areas after the Baltic Ice Lake – Yoldia Sea water level drop. Presumably, the ice margin was already so thin to allow large areas of stagnant ice to become sequentially detached from the active ice margin after the water level dropped. This is indicated by the extensive hummocky moraine areas in the supra-aquatic areas. Fast glacial flow during the formation of the Salpausselkä zone may have caused the ice marginal thinning.

The Third Salpausselkä is thought to have originated as a result of an individual ice lobe reactivation (Salonen 1991, Salonen & Glückert 1992) due to glaciodynamic processes. The ice lobe seems to have been governed by bed topography into more shallow water or against higher-lying bed in the Kiikala-Somero area, and into the distal slope of a more elevated area in the Tammela-Loppi area (Fig. 2).

2.2. Correlation with the Younger Dryas climate event

2.2.1. Introduction

Climatological rather than regional ice marginal dynamics may have led to the formation of the well-defined ice marginal deposition zones in the study area. Factors affecting climatic variation during the Younger Dryas are of prime significance for understanding the deglaciation of the study area.

The Younger Dryas cold period interrupted a general trend of global warming from glacial to warm interglacial stage. It has been suggested that the cold Younger Dryas stage resulted from an interruption in the North Atlantic Deep Water (NADW) flow (Broecker *et al.* 1989). This interruption was caused by a large meltwater input from

the Laurentide Ice Sheet through the river St.Lawrence into the northern Atlantic. However, the strong meltwater pulse did not take place during the Younger Dryas (Jansen & Veum 1990, de Vernal *et al.* 1996, Duplessy *et al.* 1996) but prior to (Björck *et al.* 1996, Jiang *et al.* 1998) and after the Younger Dryas event (Blanchon & Shaw 1995).

During the Younger Dryas the production of the NADW approached the present levels rather than levels produced during the last glacial maximum (LGM)(Veum *et al.* 1992). However, the production of the NADW during the Younger Dryas was less than prior and after it (Berger & Jansen 1995). The ocean current system presumably changed mode and area of action in the northern Atlantic during the Younger Dryas (Yu *et al.* 1996). According to Hugh *et al.* (1998) the production of the NADW would have been interrupted during the Younger Dryas and gradually replaced by another current system, possibly the North Atlantic Intermediate Water (NAIW). This hypothesis is based on the records of variation in the atmospheric ^{14}C concentration. However, a switch in the production of the NADW is not considered to have been a cause for the cold period (Berger & Jansen 1995, p. 89). Furthermore, the changes seem to have been produced only after the beginning of the Younger Dryas (Berger & Jansen 1995, p. 89).

Denton and Hendy (1994) consider that the ultimate factor that led to the onset of the Younger Dryas was a global climate change rather than a switch in the NADW production. The Younger Dryas ice marginal positions of the glacial readvance in New Zealand show an immediate response to global climate change, and not a delayed event inluenced by a switch in deep water formation.

Another hypothesis is the so called Heinrich event, i.e. a major iceberg discharge (e.g. Mayewski *et al.* 1994). Duplessy *et al.* (1996) have presented evidence against the latter hypothesis and demonstrated that the disintegration of the continental ice sheets cannot be the cause of the Younger Dryas.

There is also the hypothesis of floating sea ice released into the seas of Greenland, Iceland and Norway (Mercer 1969). The East Greenland Current fed large amounts of Arctic sea ice e.g. into the Sea of Norway. The suggested causes for the feeding of Arctic sea ice are 1) that the area of the Arctic Ocean expanded especially over the Chukch and Siberian Seas as the sea level rose approximately 11, 000 ^{14}C years ago (Duplessy *et al.* 1996), and 2) that the waters entering the Arctic Ocean increased in volume by about 20% as large drainage areas along the branches of the Mackenzie underwent changes (Teller 1990).

According to Blanchon and Shaw (1995) repeated collapse and growth pulses of the Laurentide Ice Sheet and associated effects on the glacier height and, consequently, location of jet streams hold the key in the interpretation of the development of the Younger Dryas. The collapse of the Laurentide Ice Sheet c. 14, 500-14, 200 calendar years BP triggered the movement of subtropical air masses and westerlies. Synchronously the northern Atlantic sea ice margin retreated and the thermohaline heat pump was activated in the area. By the time of the Younger Dryas (c. 12, 900 calendar years BP) the Laurentide Ice Sheet progressively became thicker owing to heavy precipitation. Consequently, the polar front jet stream was redispersed and its branch moved southwards. The jet stream displacement caused sea ice formation as the impact of the tropical air masses and westerly winds was limited. So, less warm water was fed into the North Atlantic and the northern hemisphere shifted quickly into the cold glacial stage. Towards the end of the Younger Dryas a new collapse of the Laurentide Ice Sheet restored the interglacial conditions. Strong insolation and possibly reduced glacio-isostatic uplift prevented the Laurentide Ice Sheet from recovering (Blanchon & Shaw 1995).

It is thought that sea ice covered extensive areas of the northern Atlantic during the cold climate event resulting from the combined effect of the Arctic sea ice breakup and the collapse and regrowth cycles of the Laurentide ice sheet. In the seas of Greenland, Iceland and Norway, however, sea ice coverage was limited during the Younger Dryas, except at its earliest stage (Mayewski *et al.* 1994). Consequently, this acted as a source of precipitation for Fennoscandia (Mayewski *et al.* 1994).

The super fjord heat pump concept is considered by Berger and Jansen (1995) to have brought about the Younger Dryas. Other significant phenomenons include the feedback effects accelerating the warming of the ocean, i.e. albedo, ocean circulation patterns and winds (Berger & Jansen 1995, p. 86). The warm and humid southwesterlies originating in part in the Icelandic Low area had an essential role in the fast decay of the Fennoscandian Ice Sheet. Berger and Jansen (1995) assumed that the Younger Dryas was initiated because meltwater supply and positive feedback stopped. The pre-Younger Dryas meltwater pulse would have owed to unstable ice masses, and the Younger Dryas is supposed to have been a break due to decreasing unstable ice masses (Berger & Jansen 1995, p. 91). This hypothesis argues that the remaining ice masses warmed during the Younger Dryas, and gave rise to a new melting pulse after an adequate warming had

taken place (Berger & Jansen 1995, p. 92).

The transition from the cold glacial stage in to the Preboreal involved a distinct rise in temperatures (Alley *et al.* 1993, Rochon *et al.* 1998). According to Taylor *et al.* (1997) the first change took place outside the Arctic (11,660 BP) and somewhat later (11,645, 11,612 BP) in the higher latitudes. It may have been a global ocean circulation that changed significantly increasing atmospheric water vapour concentration. Consequently, this caused a feedback on the retention of long-wave solar radiation establishing new stabilised climatic conditions (Taylor *et al.* 1997).

2.2.2. Timing and duration

The GISP2-core data from Greenland show that the Younger Dryas dates between 12, 890-11, 640 +-250 calendar years BP (Alley *et al.* 1993, Taylor *et al.* 1993, Mayewski *et al.* 1993, Stuiver *et al.* 1995). According to Kaspner *et al.* (1995) 11,460 calendar years BP marks the end of the event. The corresponding figure obtained from the GRIP (Greenland Icecore Project) ice core records is 12, 650-11, 550 +-90 (Johnsen *et al.* 1992, Dansgaard *et al.* 1993). The Greenland Stadial 1 is suggested to have started at 12,650 calendar years BP and ended 11,500 calendar years BP (Björck *et al.* 1998).

It seems that the climate changes were synchronous in Greenland and Europe (Wohlfarth 1996, Björck *et al.* 1996). Wohlfarth (1996) places the onset of the cold event at 12,700-12,500 calendar years BP. Parallel results from the Sea of Norway indicate that also marine environments responded immediately to the climate change (Haflidason *et al.* 1995).

At Gosciaz in Central Europe dates for the beginning of the Younger Dryas are in accordance with dates from the Greenland ice core studies, i.e 12,580 +-120 (Goslar *et al.* 1995). At Soppensee the date is 12,125 +-86 (Lotter *et al.* 1992, Hajdas *et al.* 1993). The date calculated for Holzmaar, 11, 940 calendar years BP (Hajdas *et al.* 1995), seems to be too young.

Older Swedish varve chronological interpretations indicate that the Younger Dryas started approximately at 11, 600 (Strömberg 1994) or 11, 800 (Kristiansson 1986) Swedish varve clay years BP.

On the basis of a preliminary revised dendro-calibration curve the end of the Younger Dryas is interpreted to around 11,600 BP (Wohlfarth *et al.* 1997). This date is based on the figures 11,450-11,390 BP defined by Björck *et al.* (1996) added with another 154 years. The final figure is consistent with the ice core ages obtained in Greenland (Alley *et al.* 1993, Taylor *et al.* 1993, Mayewski *et al.* 1993, Stuiver *et al.* 1995, Johnsen *et al.* 1992, Dansgaard *et al.* 1993). Former dendrochronological studies suggested that the end of the Younger Dryas was 11,045 calendar years BP (Becker *et al.* 1991, Kromer & Becker 1993).

The date obtained for Lake Holzmaar is 11,490 (Hajdas *et al.* 1995), for Lake Gosciaz 11,440 +-120 (Goslar *et al.* 1995), and for Soppensee 10,986 +-86 (Hajdas *et al.* 1993). Landmann *et al.* (1996) also present a similar date from Turkey, namely 10,920 +-132 calendar years BP.

In Sweden the end of the Younger Dryas and the beginning of the Preboreal is c. 10,200 ^{14}C-years (Berglund *et al.* 1994). The beginning of the Preboreal is dated by Brunnberg (1995) to a period somewhat prior to 10, 740 clay-varve years (1,500 BZ), equivalent to the 11,600 calendar years BP presented by Wohlfarth *et al.* (1997). 1,500 BZ signifies 1,500 years before the year 0 in the revised time scale of De Geer. The zero point in the time scale is 7, 288 clay-varve years BC or 9,238 clay-varve years before 1950 (i.e., BP) (Brunnberg 1995, p. 14). 1, 500 BZ almost coincides with the water level drop from the Baltic Ice Lake to the Yoldia Sea level. According to Strömberg (1994) the end of the Younger Dryas is at 1,700 BZ, when the ice-marginal retreat rate grew from about 10 to 50 metres per year. After around 1,500 BZ the rate grew up to 150 m per year.

Assuming that the impact of climate warming was immediate at the ice margin in southwestern Finland, then the doubled or tripled (Niemelä 1971) ice marginal retreat rate on the proximal side of the Second Salpausselkä would indicate the time when the Preboreal started in that area. Brunnberg (1995) also considers that the Baltic/Yoldia drop and the start of the Preboreal were chronologically connected and in fact only about 60 years apart.

On the basis of the ages listed above, the duration of the Younger Dryas was 1,150 years (GRIP, Johnsen *et al.* 1992, Dansgaard *et al.* 1993) or 1,300 years (GISP2, Taylor *et al.* 1993, Mayewski *et al.* 1993, Stuiver *et al.* 1995). Bard and Kromer (1995) suggested that the cold event lasted 1,000-1,300 years, while Björck *et al.* (1998) postulated that the GS-1 stadial lasted 1,150 years.

The studies at Lake Gosciaz indicate that the Younger Dryas lasted around 1, 140 years, and the date matches the one obtained in Greenland (Goslar *et al.* 1995). The Soppensee studies suggest the same duration, i.e. 1,140 ± 110 years, but at a later date (Hajdas *et al.* 1993). However, the Holzmaar studies suggest a relatively short duration of ca 450 years (Hajdas *et al.* 1995). There are other records of similar duration than that of Holzmaar from

northern Spain; according to Wansard (1996) the cold event lasted there for only ca 400 years (U/Th date at 11,900-11,500 BP). Varve clay chronological studies by Strömberg (1994) suggest that the Younger Dryas event lasted ca 660 years, while Brunnberg (1995) estimated that it lasted 800-1,000 years.

The apparent local variations in the length of the Younger Dryas can be explained by the varying temperature ranges used in the studies. It seems that during the Younger Dryas temperatures decreased less under the influence of continental conditions or in the tropical areas compared to marine realm.

Based on the Fennoscandian Ice Sheet model constructed by Holmlund and Fastook (1993, p. 83), the Younger Dryas cold event did not last more than 500 years, or there was a considerable temperature variation within the cold event (Holmlund & Fastook 1993, p. 77), of which the latter alternative seems more probable.

The Icelandic Vedde Ash observed in ice cores, marine sediments and in varved clays offers a tool for comparison of dating methods and enables the control of the dates. Ice core studies give the date of 12,207 ice layer years for the Vedde Ash, and an indirect Th/U date of 11,980 on the basis of dates obtained by ^{14}C (Grönvold *et al.* 1995). Wastegård *et al.* (1998) identified a Vedde Ash layer in southwestern Sweden in Lake Madtjärn, the age of which was interpreted as 12,045-11,975 calendar years BP. This interpretation is based on the recently observed, almost perfect chronological correlation between dendrochronological (German pines tree-ring width increase) and the ice core (GRIP $\delta^{18}O$) records. Swedish Vedde Ash studied at Lake Mullsjöen has earlier been estimated to date 11,100-11,250 Swedish varve clay years (Wohlfarth *et al.* 1993).

2.2.3. Temperature

At the beginning of the Younger Dryas event the climate cooled in less than 10 years (Mayewski *et al.* 1993). Correspondingly, at the end of the event, temperatures rose quickly; 7°C in less than three years (Alley *et al.* 1993, Koc *et al.* 1993). In the beginning of the Younger Dryas event sea surface temperatures decreased markedly especially during winter time in the Sea of Norway, and sea ice was present for 2-4 months (Rochon *et al.* 1998). At the end of the event, after the Vedde Ash eruption, sea surface temperatures, especially the summertime ones, decreased (6-11°C in August) and extensive areas were covered by sea ice up to seven months annually (Rochon *et al.* 1998).

During the Younger Dryas extreme cooling occurred in northeastern and Central Europe (Velichko 1995) where winter temperatures reached levels 20-30°C below those of today (Isarin *et al.* 1998). An anticyclone developed in the lower part of the atmosphere over the Fennoscandian Ice Sheet. Northern Atlantic jet stream increased in strength owing to a steep temperature gradient between sea ice and subtropical waters. The jet stream route also shifted southwards (Harrison *et al.* 1992, p. 294).

The Younger Dryas climate was characterised by a sequence of three cycles of expansion of the polar circulation systems (11,810, 12,220 and 12,640 +-250 calendar years BP)(Mayewski *et al.* 1993). Three such pulses occurred also during the Bölling/Alleröd and one during the Preboreal, 11,400 +-250 calendar years BP.

During the Younger Dryas, climatic variation in Greenland was 1) strong, 2) frequent and 3) brief (10-20 years). This applied to Finland as well because the areas were under the influence of the same polar circulation system (Mayewski *et al.* 1993) (Fig. 7a). The GISP2 (US Greenland Ice Sheet Project Two) five-point (100-yr) moving average data clearly indicate the three main cycles of climatic variation (Fig. 7b)(Stuiver *et al.* 1995).

In Finland, the primary effect of the climatic cooling was a decrease in the equilibrium line altitude (ELA) from 1, 500 m to 900 m (Holmlund and Fastook 1993). In Norway the ELA was 400-600 m lower than today, while during the last glacial maximum the ELA was 1, 000 m below the present altitude (Svendsen & Mangerud 1992).

According to Velichko (1995), near St Petersburg at Vuoksa, located to the southeast from Finland, the mean temperatures were ca 6°C lower and the winter temperatures ca 14°C lower at 10, 500 (^{14}C age) than at present. According to Isarin *et al.* (1998), the estimated mean temperatures in southern Finland were ca 10°C, i.e. ca 6°C colder than at present during the warmest month, and during the coldest month ca 19°C below the present temperatures, which amounts to less than -25°C. In the southeastern and eastern Finland Salpausselkä zone, frost polygons and ice wedge casts developed during the cold event (Donner *et al.* 1968, Aartolahti 1970, Donner 1995). Ice wedge casts are formed in mean annual temperature conditions of -6°C in Greenland at present. In comparison, in present-day Lappeenranta, southeastern Finland, the mean annual temperature is +4°C. The presence of ice wedge casts particularly in the eastern parts of the First and Second Salpausselkä indicates that even in southern Finland the winters became much colder eastwards. Ice wedges have not been report-

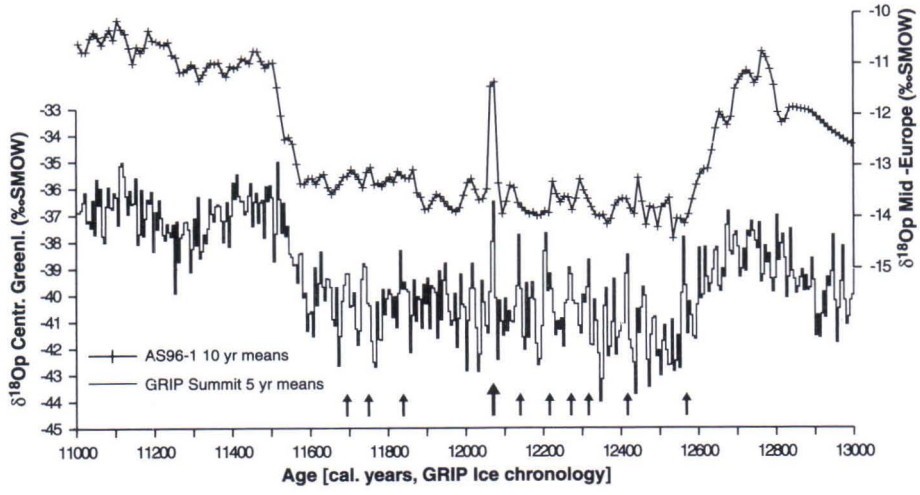

Fig. 7. a) $\delta^{18}O_p$ graphs for Central Greenland shows significantly higher variability during the Younger Dryas than during the Alleröd and Preboreal. Short subdecadal positive excursions reach almost Holocene levels (arrows). In the Mid-European record (Ammersee)(upper graph) only the highest and longest of these spikes is preserved. Reprinted with permission from von Grafenstein *et al.* 1999. ©1999 American Association for the Advancement of Science.

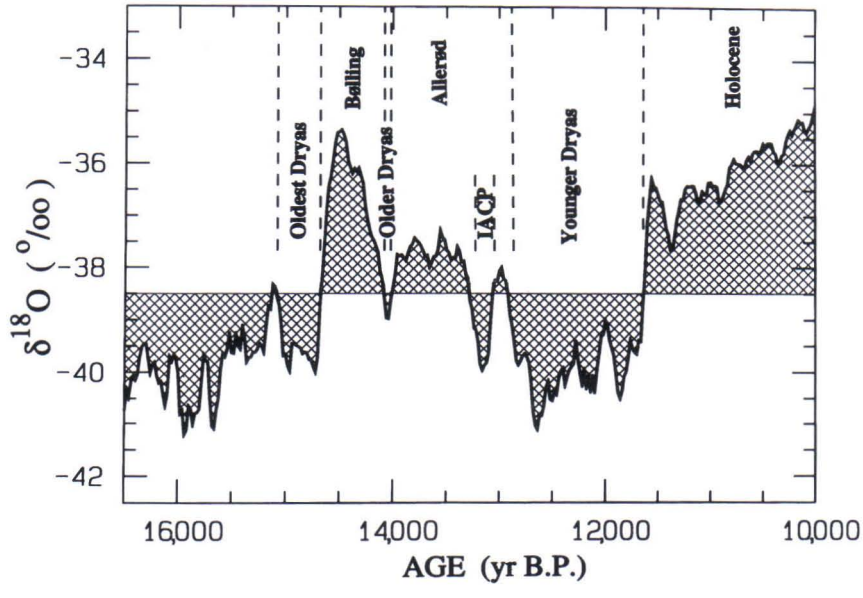

Fig. 7. b) Major $\delta^{18}O$ climate transitions (solid line represents five-point (100 yr) moving average of Figure 5. a) $\delta^{18}O$ values) and European pollen zone boundaries (Stuiver *et al.* 1995). Reprinted with permission from Stuiver et al. 1995@Academic Press.

ed in the study area or elsewhere in the Second Salpausselkä zone of the Baltic Sea Ice Lobe.

The corresponding Younger Dryas mean summer temperature in Scania was 10-13°C (Berglund *et al.* 1994). Evidence from Scania (Lemdahl 1991) indicates an exceptionally cold early part of the event (Berglund *et al.* 1994, ^{14}C date 11,000-10, 500 BP) and a somewhat warmer later part (also Lehman & Keigwin 1992, Bergsten & Nordberg 1992, Jiang & Nordberg 1996, Jiang & Klingberg 1996).

2.2.4. Precipitation

During the Younger Dryas an overall decrease in precipitation occurred in Europe (Harrison *et al.* 1992, p. 294). In Finland precipitation decreased by only c. 200 mm per year (^{14}C date 10500 BP, Velichko 1995). Results indicate that the early Younger Dryas cold climate event was relatively arid and that the end was more humid (Goslar *et al.* 1993). Similar results come from southern Sweden and Denmark (Berglund *et al.* 1994). Also accord-

ing to Hammarlund *et al.*(1999) increased precipitation was more probable than decreasing temperatures during the later part of GS-1 (Younger Dryas).

The meltwater supply from ice sheets was reduced during the Younger Dryas (Fairbanks 1989). The peak rate of sea level rise was attained after 11,500 calendar years BP, while the previous one had occurred at 14, 500 calendar years BP (Blanchon & Shaw 1995). The Baltic Ice Lake had discharged large amounts of meltwater during the Alleröd as a result of the first opening of the Billingen area (Jiang *et al.* 1998). Previously it was hypothesised that the Salpausselkäs were deposited during the Alleröd chron because the large volume of meltwater supposed to have been discharged at that time would have explained the formation of the large ice marginal deposits (Okko 1957, Okko, M. 1962, p. 139).

As the polar front shifted back to Greenland after the Younger Dryas, precipitation was doubled in that area (Alley *et al.* 1993, p. 527-529). The increase in precipitation is explained by changes in atmospheric dynamics, for example shifts in the movement and more permanent patterns of precipitating weather systems (Alley *et al.* 1993).

2.2.5. Chronology of the Salpausselkä zone

The most important single chronological reference point for deglaciation in Finland is the drop in water level of 25-28 m, i.e. the difference between the Baltic Ice Lake shoreline level (B III) and the first shoreline level of the Yoldia Sea (YI) in the history of the Baltic Basin (Sauramo 1958, Donner 1969, 1992) (Fig. 5). At this stage a connection between the Baltic Sea and the Atlantic Ocean reopened in the Billingen area (Björck 1995). The change can be clearly detected in the study and surrounding areas (Glückert 1977, 1979). Absence of intermediate levels in the double level deltas and the evidence for continued sedimentation at the lower water level indicate the drop of water level was instantaneous.

In Swedish studies the Baltic/Yoldia drop is

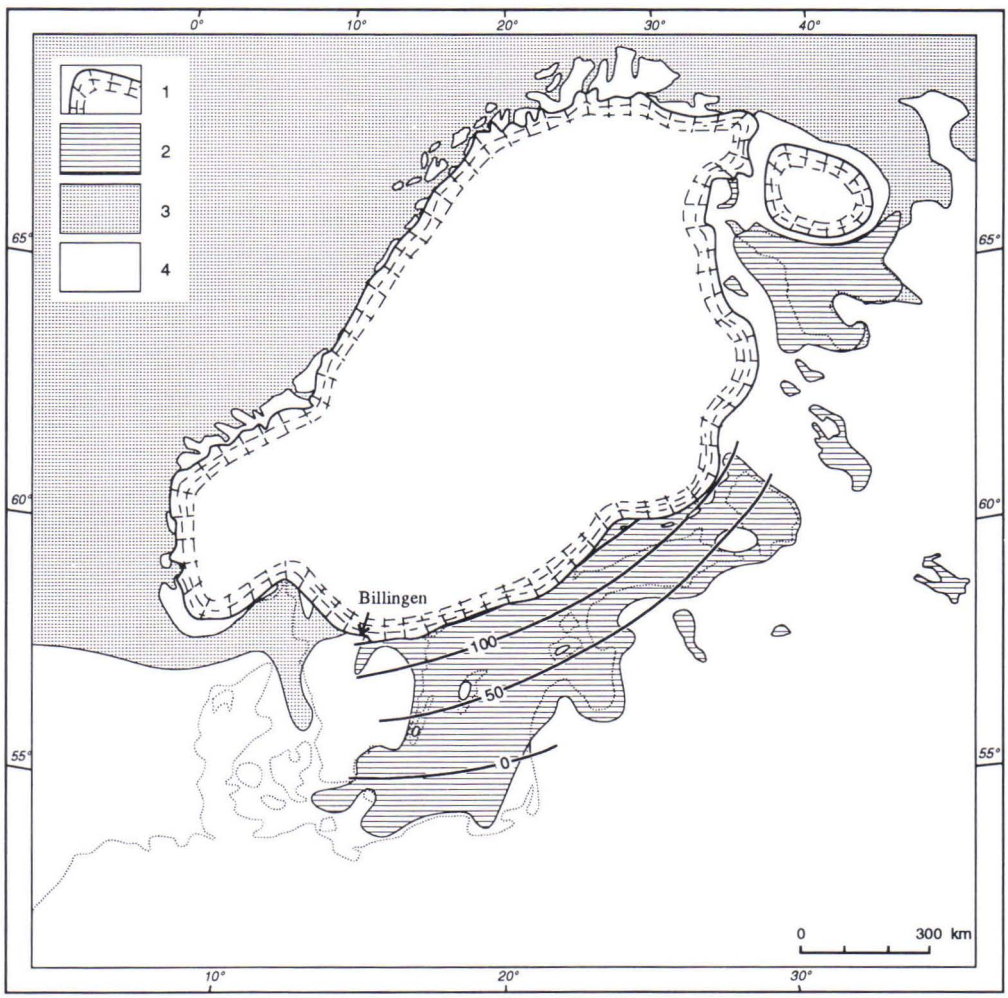

Fig. 8. The Baltic Ice Lake before its drainage at Billingen, with isobases. 1, ice; 2, lake; 3, sea; 4, dry land. (From Donner 1995). Reprinted with the permission © the Cambridge University Press.

dated to a period before the clay colour change at 1,500 BZ, possibly to c. 1,560 BZ, corresponding to 10,800 Swedish varve clay years (Brunnberg 1995).

The fact that diatactic clay found north of the Second Salpausselkä passes into symmictic clay constitutes another important chronological landmark for dating the Salpausselkä zone (Niemelä 1971). Brunnberg (1995) correlates this process with a strong marine ingression in Sweden, which took place at 1,190 BZ or 10,430 varve clay years. The date of the marine ingression (and the Preboreal Oscillation, a colder period than the early Preboreal) was placed at 11,350 calendar years BP by Wohlfarth *et al.* (1997).

In Finland varve clay observations by both Sauramo (1923, p. 111, in Somerniemi and Somero) and Niemelä (1971) show that the age of the marine ingress, i.e. the saline early stage of the Yoldia Sea, dates to 292 years after the date of the Baltic/Yoldia drop. This is when symmictic clays appear in varved clay sequences. These findings are compatible with the Swedish estimates of 300 years for the age of the Yoldia Lake (i.e. the lake that existed for a short period before the Yoldia Sea proper)(Morner 1995, Brunnberg 1995).

The Swedish and Finnish varve chronological records were earlier interpreted by Strömberg (1990) who found that there was a possible connection between the transition from the diatactic clay structure to the symmictic one on the Finnish side and the clay colour change in Sweden (1,500 BZ). Assuming that this interpretation is correct, it suggests that the marine ingress took place earlier and was more intensive in Finland than in Sweden (Brunnberg 1995).

According to Niemelä (1971) the deposition of the First Salpausselkä started approximately 1,040 years before the Baltic/Yoldia drop (9,250 B.C.) and took about 250 years. Sauramo stated that it lasted roughly 230 years (Sauramo 1923, p. 86). In the area of the First Salpausselkä deteriorating climate was reflected in the glacial retreat rate - only 10-20 metres per year. Earlier, when the ice-marginal position was southeast of the First Salpausselkä at Hynnänkorpi-Raala, the ice marginal retreat rate had already declined to about 25 m per year. This decrease in the retreat rate probably indicates the beginning of the Younger Dryas. In other areas the ice margin retreated annually around 50 m (Niemelä 1971).

The deglaciation between the First and Second Salpausselkä lasted approximately 500 years (9,000-8,500 B.C.) according to Niemelä (1971). Sauramo (1923) found that the deglaciation between the First and Second Salpausselkä lasted c. 250 years including the 100 years that the ice margin remained close to the First Salpausselkä.

The deposition of the Second Salpausselkä finished by ca 8,300 BC in the study area, i.e. about 80 years before the Baltic/Yoldia Drop in 8,213 BC (Niemelä 1971). Niemelä found that the deposition of the the Second Salpausselkä lasted about 200 years, starting at 8,500 and finishing at 8,300 BC According to Sauramo, it lasted around 180 years (8,364-8,182 BC) and was finished before the BIL/ Yoldia drop.

It is possible to calculate a preliminary estimate of when the Salpausselkä zone was deposited by considering the following data: the Swedish varve chronological date of the Baltic/Yoldia drop (1,560 BZ), the corresponding date of the Preboreal/ Younger Dryas interface (1,500 BZ), and the date of the Preboreal/Younger Dryas interface at 11, 600 calendar years BP as interpreted by Wohlfarth *et al.* (1997). Thus, on the basis of the varve clay chronological records of Niemelä (1971), the Second Salpausselkä was deposited at 11,940-11,740 calendar years BP, and the First Salpausselkä at 12,700-12,450. The corresponding dates based on varve chronological studies by Sauramo (1923, 1958) are 11,840-11,660 and 12,320-12,090 calendar years BP (Fig. 9).

Considering the varve clay chronological record of Niemelä (1971) and the chronological interpretation mentioned above, the First Salpausselkä was deposited with a 200 year delay with respect to the climatic cooling. This can be concluded providing that the early cold period of the Younger Dryas dates to 12,900-12,650 calendar years BP, as interpreted from the Greenland ice core data (Alley *et al.* 1993, Johnsen *et al.* 1992, Taylor and al 1993, Dansgaard *et al.* 1993, Mayewski *et al.* 1993, Stuiver *et al.* 1995). The deposition of the First Salpausselkä coincides with the first cold period within the Younger Dryas (12,640 calendar years BP, Mayewski *et al.* 1993).

The deposition of the Second Salpausselkä started 35-105 years after the Vedde Ash volcanic eruption. It is interpreted to have taken place at 11,975-12,045 calendar years BP (Wastegård *et al.* 1998) and the varve clay chronological studies by Niemelä (1971) indicate that the Second Salpausselkä was deposited at 11,940-11,740 calendar years BP. As dating methods involve various potential sources of error, the ash eruption may have had an impact on the development of the Second Salpausselkä (Fig. 9).

Based on chronological studies of Niemelä (1971), the Salpausselkä zone was deposited during 960 years in the area of the Baltic Sea Ice Lobe. During that time span the First and Second Sal-

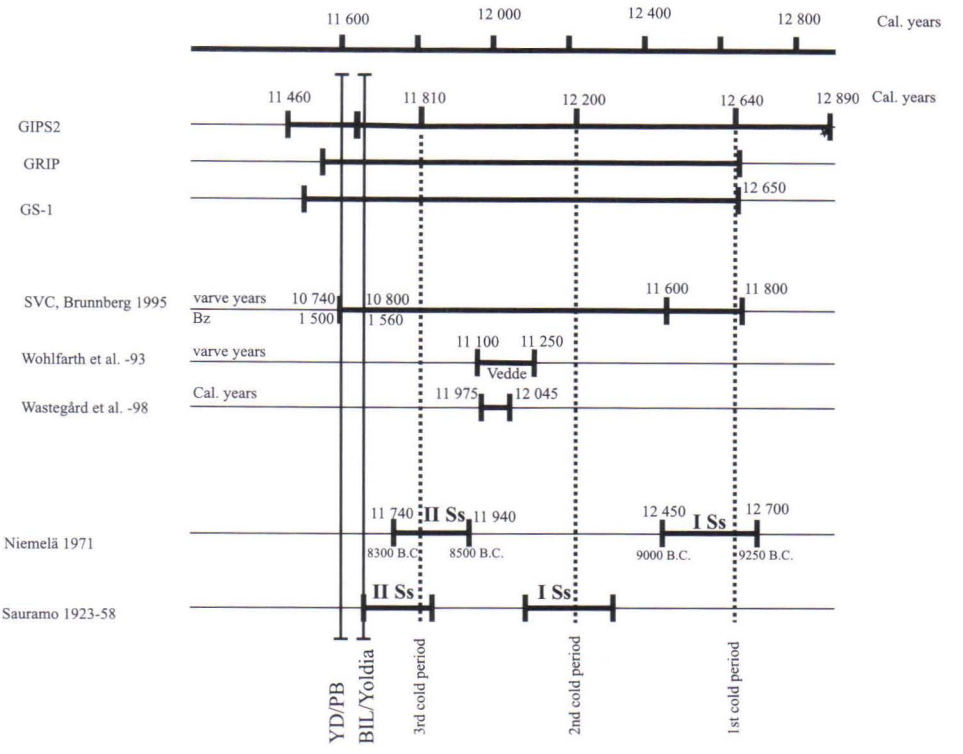

Fig. 9. Time scales for the Younger Dryas event, Salpausselkä zone and the development of the Baltic Sea. Correlation between Swedish Varve Chronology and Greenland ice core data according to Björck *et al.* 1996 and Wohlfarth 1996, Wohlfarth *et al.* 1997 (see text, p. 25). GISP2 = time scale of the Younger Dryas from Alley *et al.* 1993, Mayewski *et al.* 1993, Taylor *et al.* 1993 and Stuiver *et al.* 1995 (11,460 according to Kapsner *et al.* 1995). GRIP = time scale of the Younger Dryas from Johnsen *et al.* 1992 and Dansgaard *et al.* 1993. GS-1, Greenland stadial 1 (Björck *et al.* 1998). SVC = Swedish Varve Chronology, from Brunnberg (1995). Niemelä 1971, the latest Finnish varve chronology for SW Finland. Sauramo 1923 and 1958, Finnish varve chronology according to Sauramo. YD/PB marks the end of the Younger Dryas and the beginning of the Preboreal stage. BIL/Yoldia marks the date of the Baltic Ice Lake/Yoldia Stage water level drop (25-28 meters). Vedde gives the date of the Vedde ash event in Iceland, dated from Sweden (Wohlfarth 1993, Wastegård *et al.* 1998). The three cold periods (11,810, 12,220 and 12,640 +-250 calendar years BP) according to Mayewski *et al.* 1993.

pausselkä were formed. In the Finnish Lake District Ice Lobe area, the Salpausselkä stage (including the Heinola deglaciation) lasted at least 897 years (Rainio 1993). According to Brunnberg (1995), the period of slow deglaciation in eastern Middle Sweden (ice marginal zone) lasted at least 800 years, possibly 1, 000 years. The fact that no delay is found between the temperature rise and the faster ice marginal retreat towards the end of the Younger Dryas, and that the development of the ice marginal deposits of the Salpausselkä zone took almost as long as the 1,000-1,300-year cold climatic event, show that the ice margin responded quickly to the cold period.

3. RESULTS

This chapter presents the results from the mapping of ice marginal positions in the Second Salpausselkä and detailed sedimentological analysis on four selected type-deposits.

3.1. Ice marginal positions and meltwater routes

The Second Salpausselkä is divided into three zones (Figs 5, 10, 48) namely, the Outer Zone, the Central Zone and the Inner Zone. The terms Outer, Central and Inner describe the position of a zone in relation to the center of the ice sheet, with the Outer zone being the one futher to the southeast, and believed to be the oldest. The total width of the Second Salpausselkä is 6-8 km. The zones are discussed below in chronological order, i.e. from oldest to youngest.

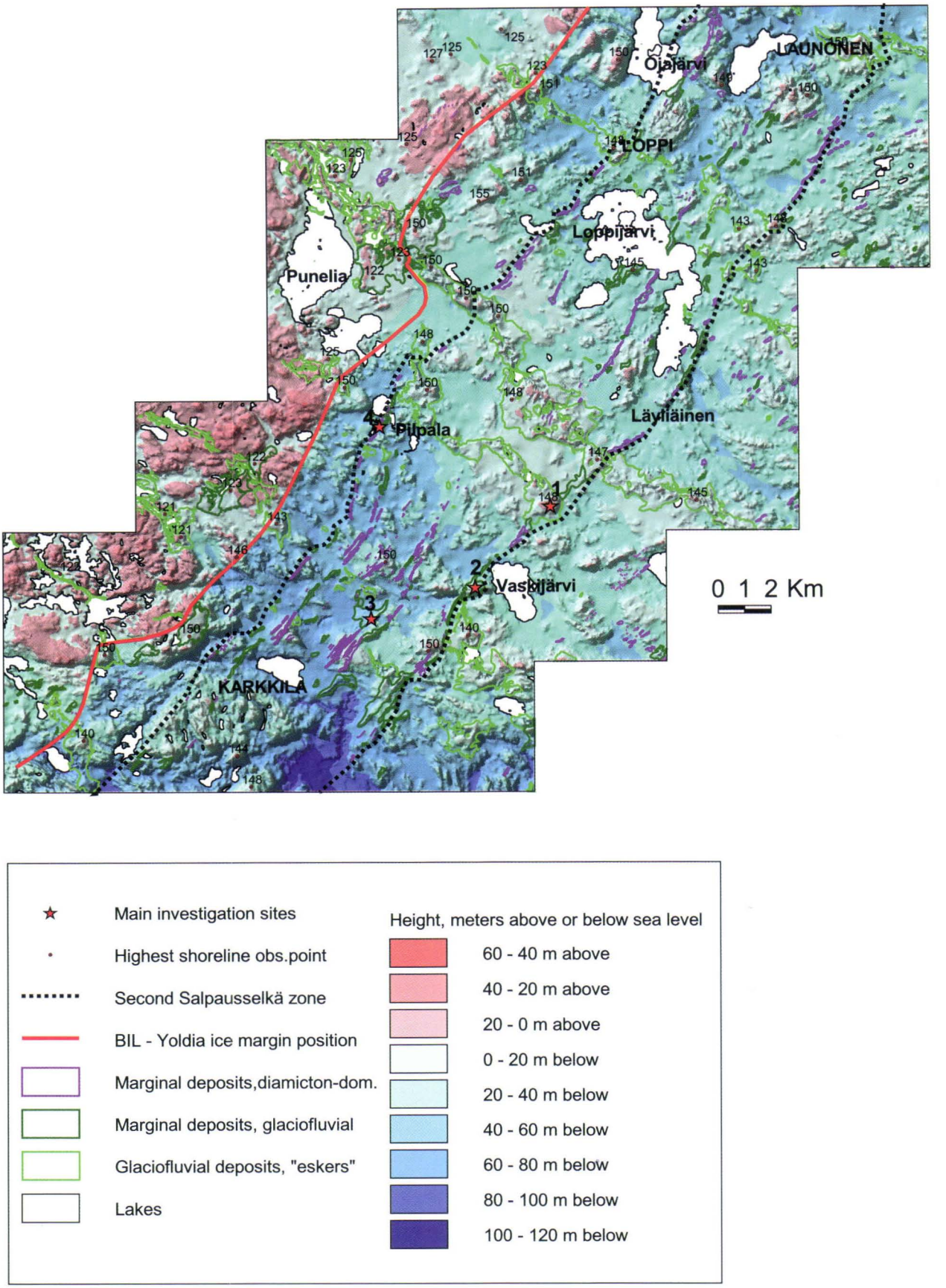

Fig. 10. Ice marginal deposits and esker systems in the study area, water depth in front of the ice marginal positions are also shown. Electron contour data © National Land Survey of Finland.

3.1.1. Outer zone

The Outer, or distal zone, is the narrowest of the three zones and northeast of the Koli hill it is usually composed of only one ridge. It widens only to the southwest of the Koli hill (Figs 5, 10). The width is ca 1 km at Karkkila (Fig. 5), where it ends short of a river valley where water depth reached 100 m during deglaciation (Fig. 10).

Ice marginal deposits composed mostly of sand and gravel dominate the Outer zone. These ice-marginal deposits can be seen at Toivike, Multamäki, the southern part of Loppijärvi, Sajaniemi and Launonen. They also form the extensive deposits of Vuonteenmäki (the Pilpala esker) and the Läyliäinen esker of Hyrrönharju in the Esker Area (Figs 5, 10, 47).

In places, the bedrock surface lies buried under 50 m of extensive glaciofluvial ice marginal deposits. In contrast, at Toivike, Multamäki and Launonen (Figs 5, 10) bedrock outcrops occur in proximal parts of deltas that were deposited at the highest shoreline level.

Diamicton dominated ice marginal deposits are found southwest of Launonen and on both sides of the Esker Area (the zone formed by the Läyliäinen and Pilpala eskers), at Vaskijärvi and Haapastensyrjä (Figs 10, 48).

The continuity of the ice marginal deposits is broken northeast of Haapastensyrjä. Other discontinuities occur mainly east of Lake Loppijärvi at Sajaniemi. East of Vuonteenmäki gaps are absent and the ice marginal deposits continue beneath the road to Läyliäinen (sections observed during roadwork) (Figs 10, 11).

3.1.2. Central zone

The Central zone ranges in width from a single ridge, northeast of Lake Loppijärvi, to an area 2 km wide, east of Karkkila in the Huhtimo-Tupsumäki area. It is largely continuous, with few notable gaps, including a 4 km wide gap southwest of the Pilpala esker. Northeast of the Läyliäinen esker the zone is continuous with diamicton-dominated sediments up to Lake Loppijärvi. The zone remains narrow northeast of Lake Loppijärvi, composed mainly of a single ridge (Figs 5, 10, 48).

Sand and gravel dominated ice marginal deposits are found in the Lake Pyhäjärvi (Karkkila centre) basin and in the southeastern end of the Siikala basin. It seems that the ice margin withdrew with an almost straight front line along the flanks of the Huhtimo hill as deep water glaciofluvial sedimen-

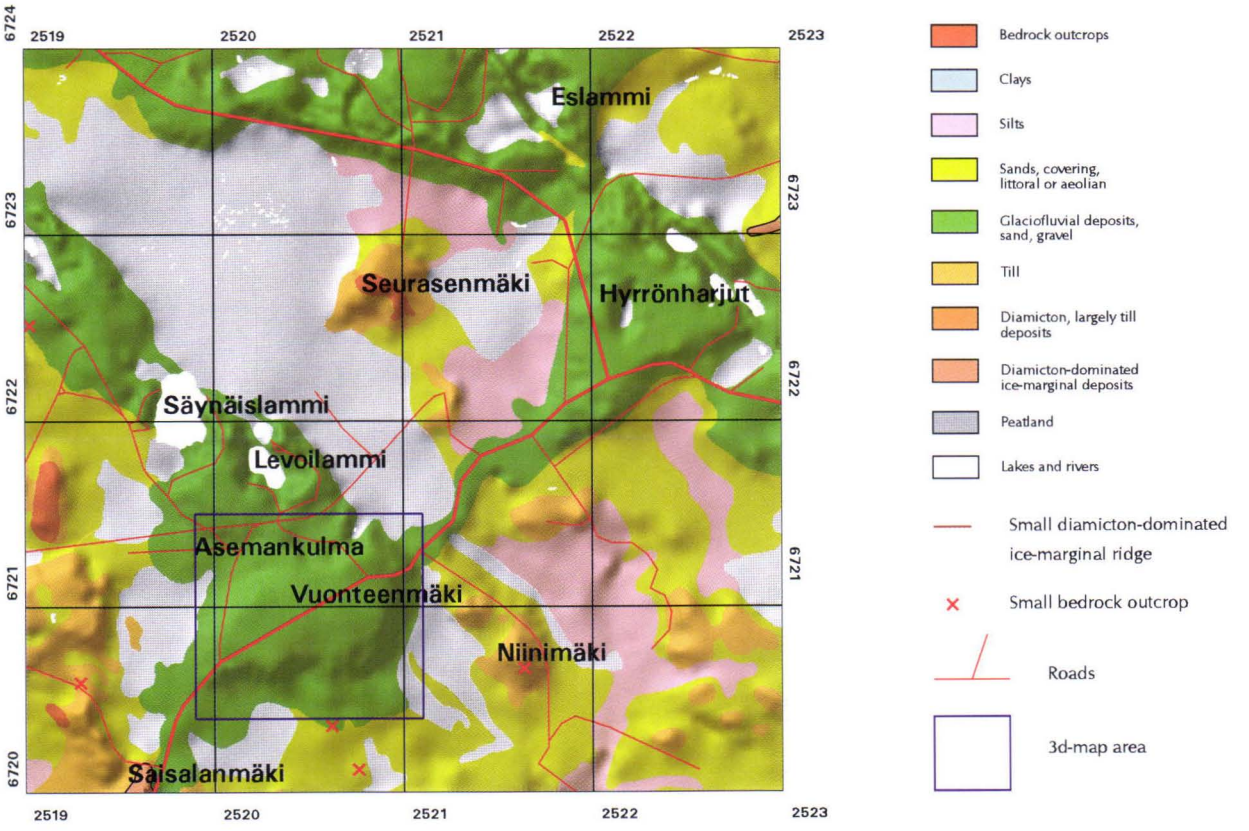

Fig. 11. Hillshaded Quaternary geology map (scale 1:40, 000) around Vuonteenmäki. Inset shows location of the hillshaded 3D map area in Figure 12. Revised version of the Geological Survey of Finland Quaternary geology map, scale 1:40, 000. Quaternary geology mapping data © Geological Survey of Finland. Elevation contour data © National Land Survey of Finland.

tation graded into the push moraine type of deposition on higher ground (Figs 10, 31).

At the Huhtimo hill, ice marginal deposits consist of diamicton and for most parts of basal till. They occupy the stoss or summital positions of hills. Elsewere, east of the Siikala basin, these are found away from depressions or from meltwater routes (Figs 10, 48). Between the Läyliäinen esker and Lake Loppijärvi, in the Sajantilankulma-Järventausta area, the ice marginal zone forms a continuous ridge that occupies a stoss side position (Fig. 10) and is mainly composed of basal till.

Northeast of Lake Loppijärvi in the Jokiniemi area, the zone exhibits a clear altitudinal contrast with sand and gravel dominated (glaciofluvial) deposits that occur in valleys while diamicton-dominated ice marginal ridges occur in higher areas. A stoss-side glaciofluvial deposit accumulated in the Jokiniemi valley. South of Lake Kesijärvi the zone is relatively wide and composed of several ridges. In that area the ridges that were deposited on distal slopes are mainly composed of sand and gravel whereas in a stoss-sloping area the ice marginal ridge is mostly composed of diamicton-dominated sediments, largely basal till (Figs 10, 48).

The largest gaps of the Central zone lie adjacent to and across the Esker Area. Southwest of the Esker Area the gap is 4 km wide and a more continuous chain of ice marginal deposits is found again at 2 km distance northeast of it. To the northeast of Lake Loppijärvi the zone is characterised by a large number of gaps and is composed of 0.5-2 km long ridges (Figs 10, 48).

3.1.3. The Inner zone

The width of the Inner ice marginal zone varies considerably. In the Pyhäjärvi-Tuorila-Siikala area up to the west side of Hunsala, the width of this zone is about 2 km. In Hunsala, in the Hirvilammi area, there is only one chain of deposits that can be clearly identified. In the Luutasuo area and east of Lake Ojajärvi the zone is approximately 1 km wide. Between Lake Erävisjärvi and the parish of Loppi the ice marginal zone forms only one chain of deposits (Figs 5, 10, 48).

The diamicton-dominated deposits are concentrated along the higher areas at Luutasuo, Lake Ojajärvi and northeast of Siikala. In addition they are found in smaller higher areas at Tuorila (Figs 5, 10, 48).

Deposits of primarily glaciofluvial origin are found mainly in association with meltwater routes, as is seen in the Pyhäjärvi-Tuorila-Siikala depressions, southwest of Hunsala and distally to the Lake Ojajärvi basins (Figs 10, 48, 49).

The proximal subzone of the Inner zone is absent in Tuorila. The Outer zone is more continuous instead, stopping 2 km short of Hunsala, where the proximal subzone is again continuous, forming the Hirvilammi ridge (Figs 10, 48).

The characteristics of the Inner zone show a similar kind of altitudinal contrast in sedimentation type as in the other zones. Diamicton-dominated ice marginal deposits are concentrated onto stoss slopes lying opposite to the direction of glacial movement, and ice marginal glaciofluvial material was deposited in adjacent valley or distally sloping lower areas. The glacial retreat seems to have been faster in the lee than in the stoss side of bedrock hills or ridges.

3.1.4. Interzonal areas

Ice marginal deposits and landforms occur in the interzonal areas as well. To the east and northeast of Lake Loppijärvi up till Lake Kesijärvi, sand and gravel dominated ice marginal deposits are closely spaced between the Outer and Central zones. Five meltwater routes are interpreted to have been active in the area (Figs 10, 48).

Between the Central and Inner zones, northeast of the Läyliäinen esker, there is a zone without ice marginal deposits. In the higher area between the esker system and the Lake Loppijärvi Basin, the deposits are diamicton-dominated. However, west of Lake Loppijärvi deposits are sand and gravel-dominated and glaciofluvial of origin. Northeast of the parish of Loppi the zone is situated distally on bedrock hills deposits of which are glaciofluvial. A concentration of sand and gravel dominated ice marginal deposits also occur along the meltwater route of the western shore of Lake Loppijärvi, close to the Central zone (Figs 10, 48, 49).

Narrow diamicton-dominated ice marginal ridges are abundant southwest of the Pilpala esker, between the Central and Inner zones. In the Siikala basin a large number of glaciofluvial ice marginal deposits exists. These deposits were most probably generated by meltwater routes that occupied the depression in the area (Figs 10, 48, 49).

3.2. The Salpausselkä type deposits

3.2.1. Vuonteenmäki, Karkkila

In the study area glaciofluvial deltas occur mainly in association with the two main esker systems, in the so-called Esker Area (Fig. 47). The Vuonteenmäki site (Figs 11, 12, 13) is a delta that forms the southeastern end of the Pilpala esker (Figs 2, 7, 8). Its size and shape is typical of the rest of glaciofluvial deltas in the study area.

Observations on the Vuonteenmäki delta were carried out on 25 occasions between 1989 and 1993. Most observations were carried out between 1990 and 1992 in the central part of the delta where foreset units were well exposed. As the quarrying at the site progressed, continuous observations were carried out every one or two months. The data was completed with tacheometer measurements (12 field projects). In 1993, the section in the western part of

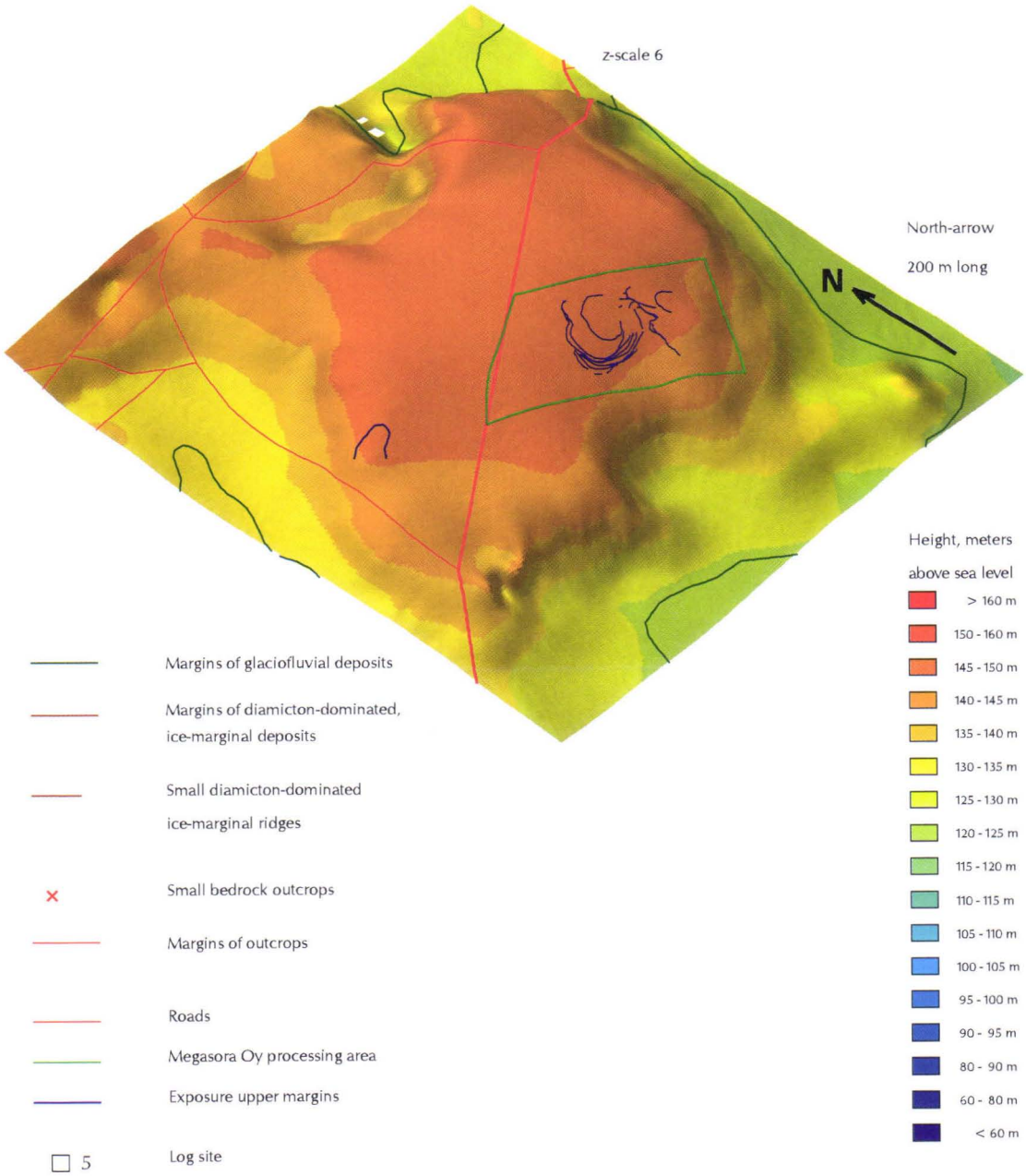

Fig. 12. Colour classified hillshaded relief map of Vuonteenmäki as seen from the NW (210° with 45° vertical angle). Elevation contour data © National Land Survey of Finland.

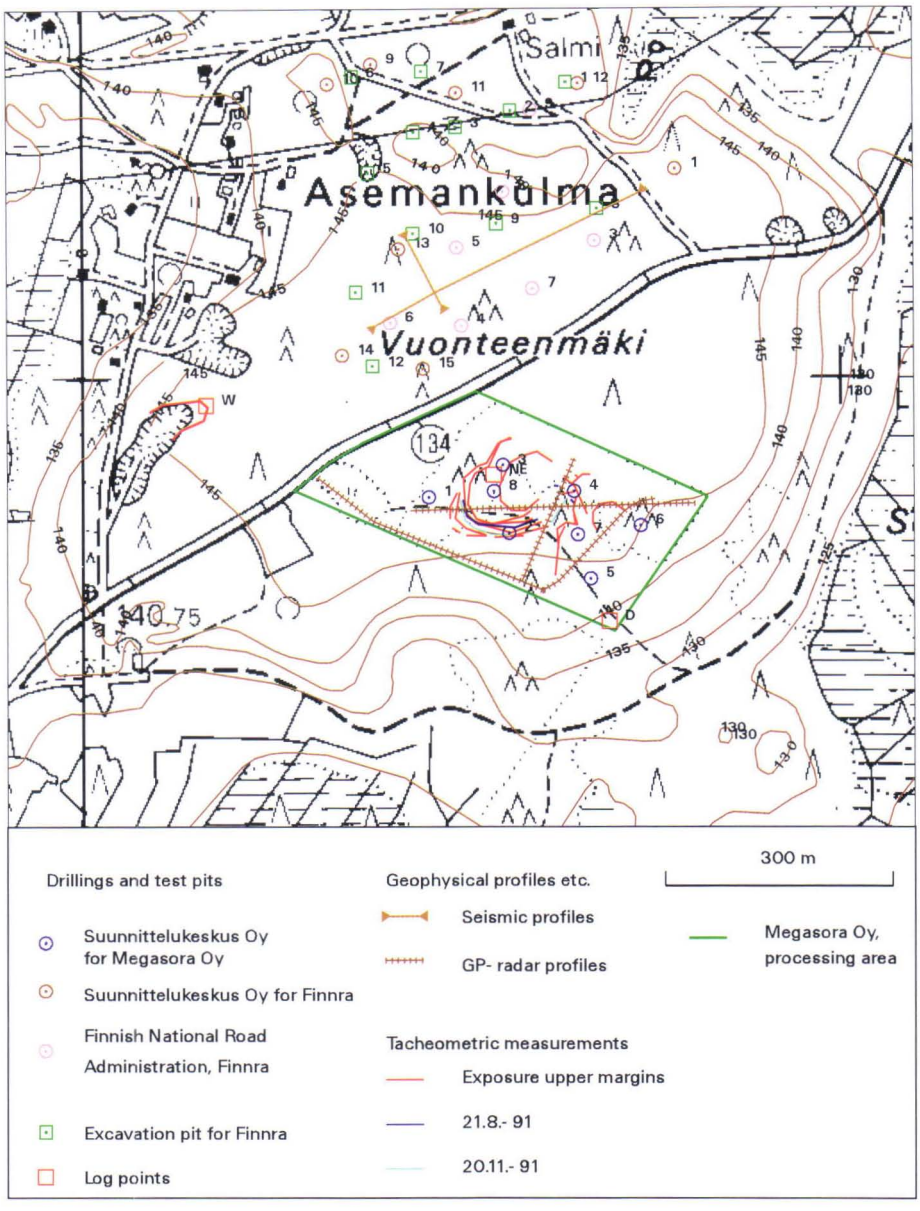

Fig. 13. Location map at 1:10, 000 scale for the sites investigated in the Vuonteenmäki area. Topographic (Basic) map and contour data © National Land Survey of Finland.

the delta was observed and monitored as well. A total number of 108 palaeocurrent direction measurements, mainly foreset unit dip directions were recorded, usually in series of at least ten measurements at a time. The company Megasora Oy and the Finnra kindly provided drilling data (30 points), seismic soundings data (500 m), ground penetrating radar data (5000 m) and test pit data (12) (Fig. 13).

3.2.1.1. Morphology

The bedrock outcrops adjacent to Vuonteenmäki are situated south of the delta, at c. 125-130 m a.s.l. Underneath the delta, in the southern part of the main exposure, the bedrock surface is at 110 m a.s.l. according to well drilling data. Seismic soundings suggest that the bedrock surface is at 110-120 m a.s.l. in the northern part of the delta. Thus, the thickness of the deposit is around 30-40 m in the delta area.

The delta top is at 147-148 m above present sea level. The top of the delta is almost at the same elevation; only at the northeastern margin of the main exposure there is a c. 100 m wide and 3-5 m deep depression running north-south. The depression may represent a relict channel. This is indicated by the fact that a sandy interlayer following the shape of the depression occurs between gravelly units in the southwestern continuation of the depression. Eastern and southern slopes are steep but less so in the northwest. A chain of kettleholes

defines the delta margin in the north. A littoral ridge about 1 m high and 20 m wide, was deposited around the distal margin of the delta.

3.2.1.2. Lithostratigraphy and sedimentary structures

The bottomset units (Fig. 17) in the lower part of the deposit are interpreted as being fine sand and coarse silt on the basis of drilling, sampling, section observations, and GPR data. The bottomset unit is about 20 m thick. The upper contact of the unit is at about 125 m a.s.l. in the central and the northwestern part of the exposure. This interpretation is based on the available drilling data (Suunnittelukeskus Oy 1989). Also in the northern part of the delta the sediment lying below the groundwater level is fine-grained: Its seismic velocity is 1400 m/s, in places only 1,300 m/s. The fine sand and silt of the bottomset unit are estimated to constitute more than 50% of the total delta material.

In the southern part of the main exposure the groundwater level is at 124-125 m a.s.l. and further north at 128-130 m a.s.l. (data from investigations of Megasora and Finnra). On a large scale these levels correspond to the upper contacts of the finer (partly silty) bottomset facies units. In the distal part of the exposure, the upper contact of the bottomset unit is at less than 125 m a.s.l.

Intermediate sedimentation units are found between the sandy/silty bottomset facies and the gravel dominated foreset facies units although they have been partly eroded at the main foreset deposition stage. The intermediate unit is mainly composed of horizontally bedded sand, ripple sand and dune cross-bedded sand (Fig. 17). The thickness of this intermediate unit is 3-7 m as interpreted on the basis of drilling results (Megasora Oy). Palaeocurrent directions vary considerably within the unit. However, directions towards SSE were most common, primarily in the dune-type sand units. Part of this low dip sedimentation facies is interpreted to belong to the lower foreset facies (Fig. 17, units IIb, IIc) and part of it to the upper bottomset unit (UB).

A coarse gravelly unit, best developed in a zone that runs from the northeast of the delta to the main observation area. Gravel (2-60 mm in diameter) forms c. 10% of total delta material and c. 20% of the material above the groundwater level. These estimates are based on drilling, core samples and seismic data. In the central part the gravelly unit is more than 20 m thick (Fig. 21). The gravel becomes finer towards the margins of the delta, and the thickness of the unit decreases to no more than 10 m on the outer border of the area.

The most important of the foreset facies units is F1, a bar which dominated the exposure's largest section since October 1990. The palaeocurrent direction for F1 is 220° on an average. The highest part, i.e. the top of the F1 is in this direction. The directions measured from this unit diverge towards both south and west by about 40° to form a fan-shaped deposit with dips between 180 and 260° (Figs 14, 15, 20). The dip directions on the southeastern side, where unit F3 partly covers subunit F1b, are 130-140°.

The thickness of F1 reaches the maximum (Figs 14, 15, 17) coincident with the subunit top area where it is over 15 metres thick as interpreted on the basis of exposure observations, ground penetrating radar and drilling data. The deepest level of the lower contact, ca 130 m a.s.l., also coincides with the barform top area (Fig. 14). Normally, foreset units are 8-10 m thick in the main observation site and the basal contact is usually at 136-137 m a.s.l. Towards the northeast the erosional lower contact of the unit F1 rises up and the thickness of the coarse part of the foreset is reduced to around 5 m.

Unit F2, lying southeast of F1, is similar, though less thick than unit F1. The GPR profile suggests that it postdates unit F1 on the margin of which it was deposited.

Between the units F1 and F2 subunits were deposited convergently together forming unit F3 (Figs 14, 15, 16). Unit F3 is maximally13 metres thick according to the GPR profiles. The sediments in F3 pass distally from coarse foreset sands and gravels into fine (horizontally bedded) sand.

Unit F1 is covered directly by unit F4 in the west. Unit F4 is interpreted as a depression infill associated with the main unit F1, infilled with material supplied mainly from the northeast. Unit F4 has a conformable and tangential lower contact with a 1.0 m bottom horizon of (horizontally bedded) sand.

The topset facies (T)(Figs 14, 17) is composed of massive horizontal sheets of stony and coarse gravel 0.5-2.0 metres thick, and of horizontally bedded sand sheets, usually 0.2-1.0 metres thick. Only estimates of the original thickness of the topset can be given since an estimated 0.5 metre of it has been stripped away. The areal extent of the topset facies depends on the height of the delta top with respect to the waterlevel. Its lower contact is at 142-144 m a.s.l. The delta topset is not usually detectable in GPR due to antenna blind angle.

The sediments become finer distally passing into sand and fine sand (partly silt) with decreasing proportion of gravel. Horizontally bedded and ripple sands characterise the bedding structure (Fig. 18, top). A 1.0 metre thick deformation horizon

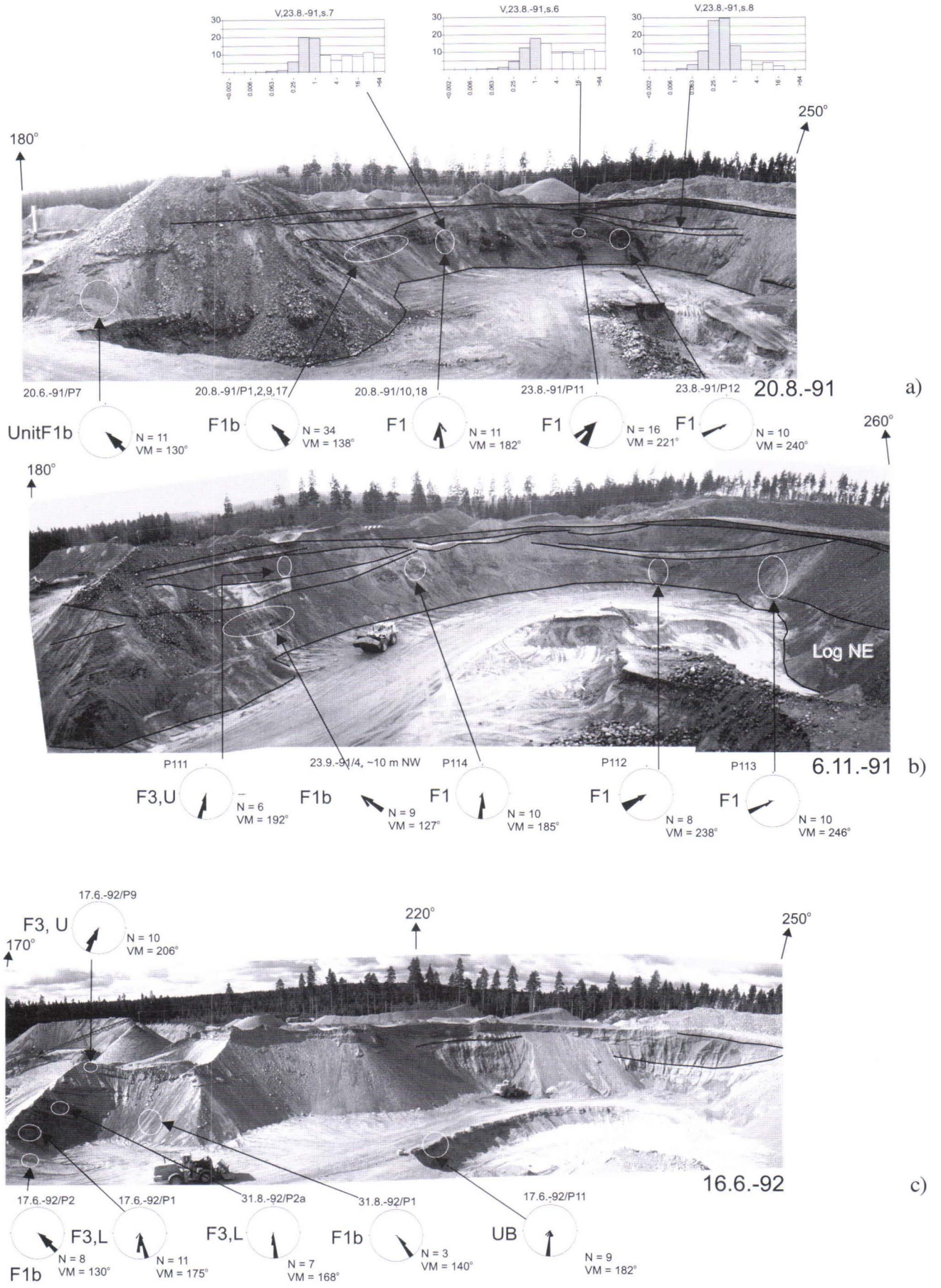

Fig. 14. a, b, c. A series of three photography mosaics from Vuonteenmäki pit main section. Pictures taken in summer 1991, autumn 1991 and summer 1992, towards WSW. Palaeoflows are represented as rose diagrams and grain size distributions as histograms. Unit F1, the main unit of the foreset facies, represents a megaforeset, interpreted to have been deposited as a barform and Unit F1b is its SE flank.

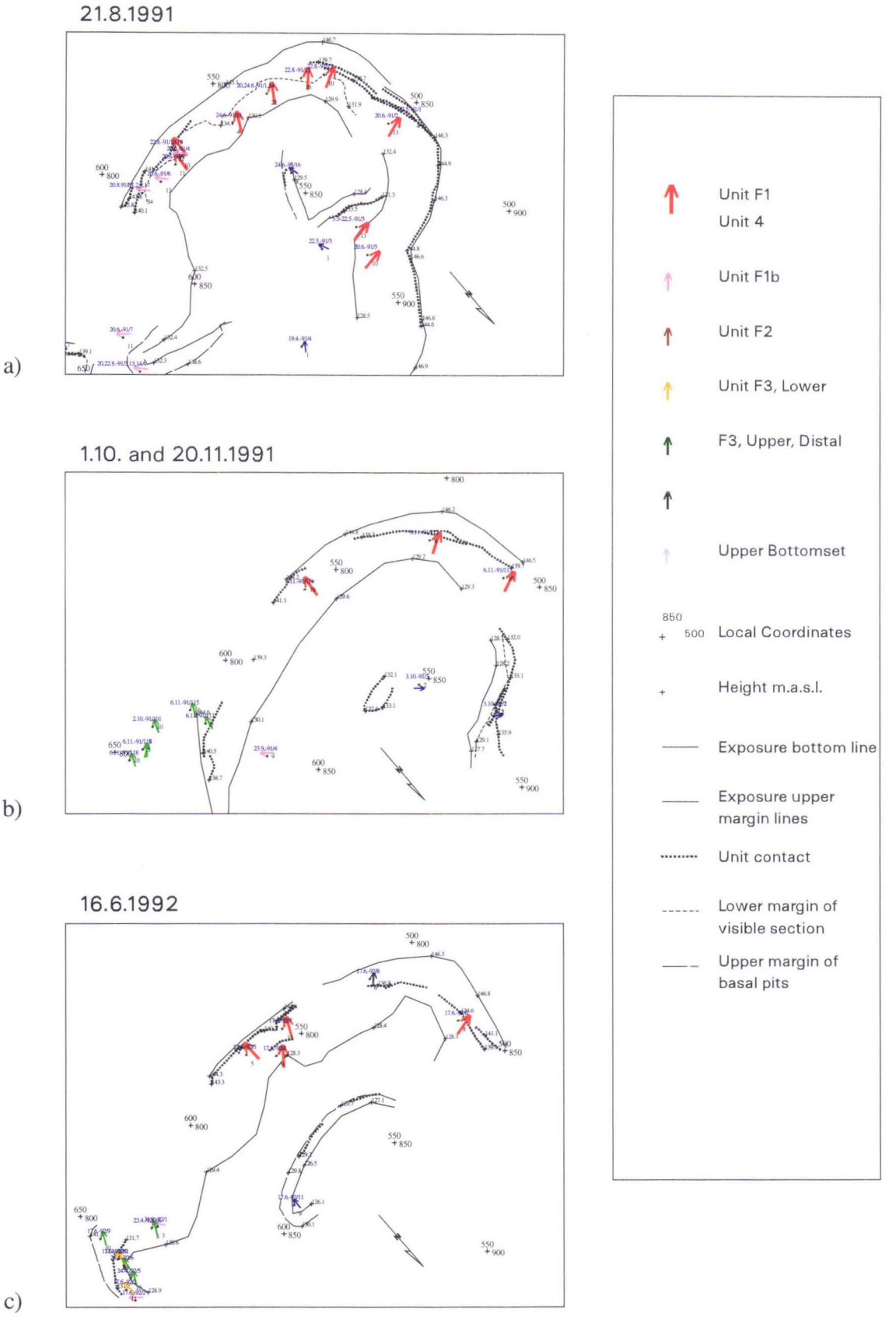

Fig. 15. a, b, c. Series of three maps related to the mosaics of Figure14. Maps have been rotated to the viewing direction of the corresponding mosaics.

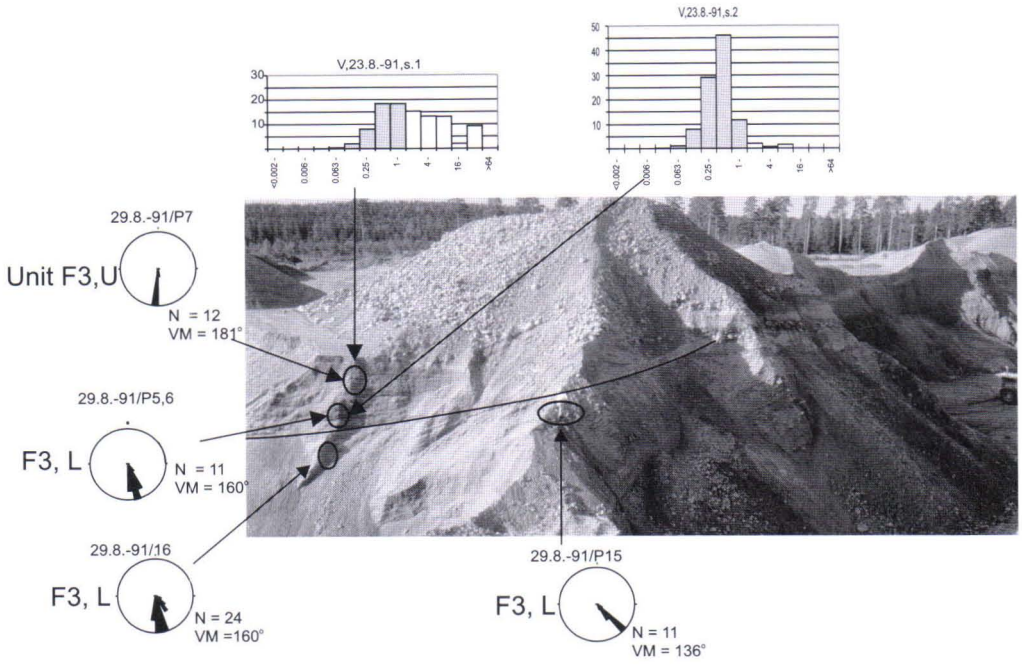

Fig. 16. Unit F3. The unit is a foreset facies unit with smaller-scale and more variable structural elements than units F1 and F2, the latter of which was deposited prior to unit F3.

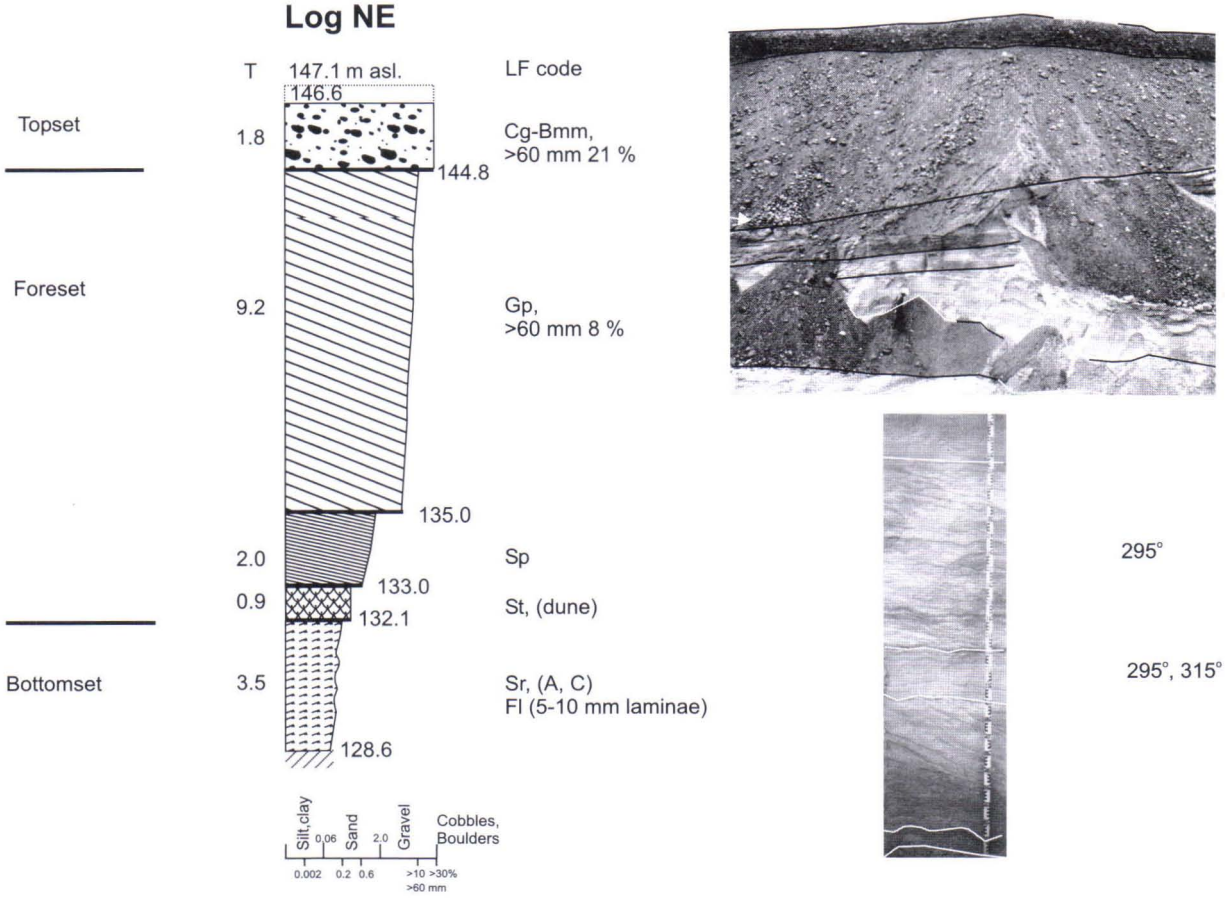

Fig. 17. log NE. Foreset facies unit F1 and associated lower contact with the bottomset facies. Foreset facies is truncated by massive gravel, interpreted as the topset facies. Lower photo is from the bottomset unit. Values 295°, 295°, 315° are measured from ripple foreset dips.

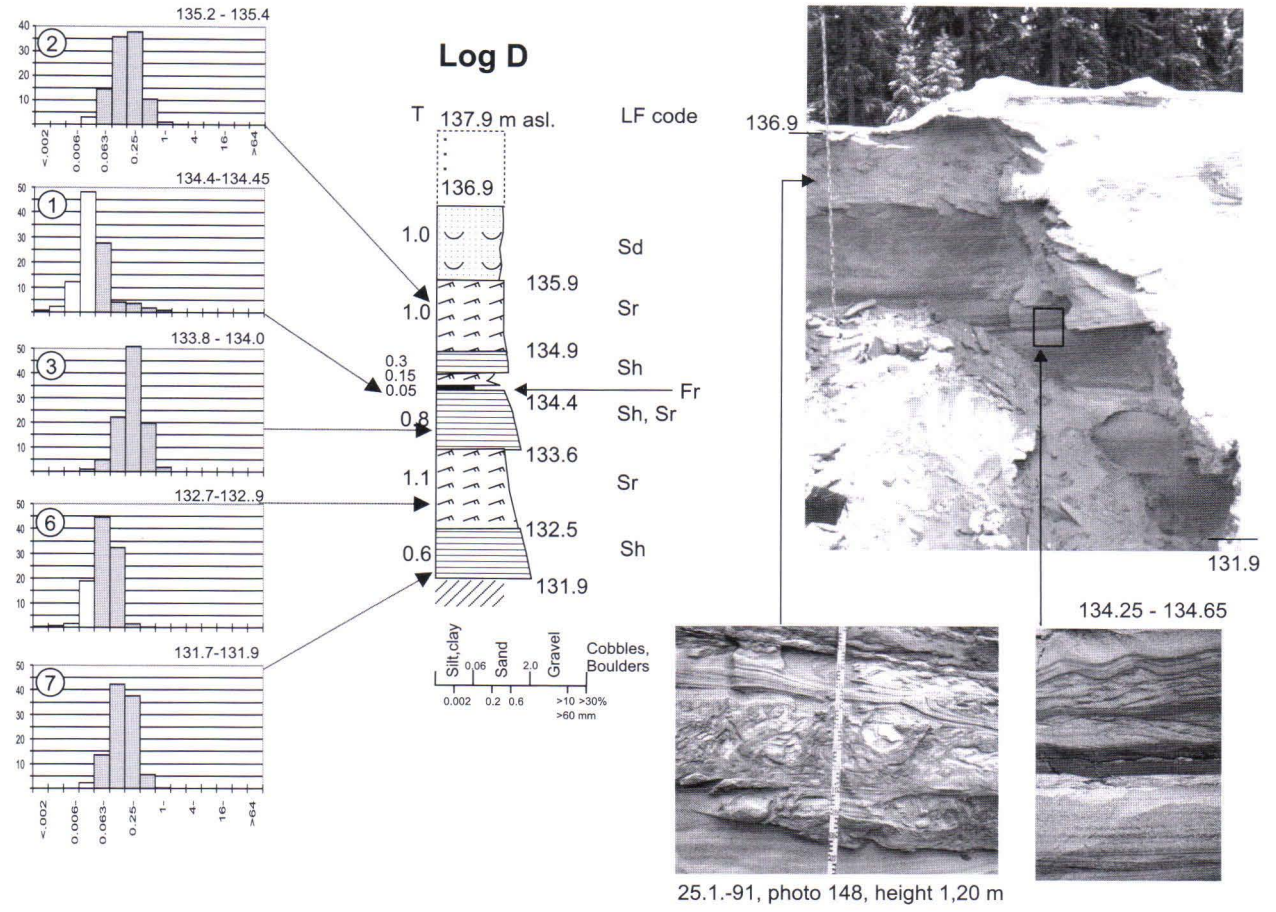

Fig. 18. Log D. Distal margin of Vuonteenmäki characterised by cyclic grain-size distributions of various units in the lower part and the deformed bed in the upper part.

with load cast or folding of thick continuous sediment beds occurs in the exposures at the southern margin of the main area (Fig. 18, bottom left).

Low dipping beds of coarse sand cover the fine sand and silt beds in the western exposure (Fig. 19). A bouldery and stony, narrow sedimentation unit occurs trending SW-NE.

3.2.1.3. Interpretation

The fact that the southwestern esker system of the Esker Area (Figs 5, 47) starts at Vuonteenmäki indicates a partial reorganisation of the meltwater routes associated with the Second Salpausselkä deposits.

The ice margin formed a delta funnel (Figs 11, 12). This is indicated by the kettlehole morphology in the northeastern margin of the delta and the bouldery diamicton in the western exposure (Fig. 19). The diamicton was formed in association with minor oscillation. A proximal push moraine ridge resulting from a readvance is absent (Donner 1976, p. 57). Kettlehole distribution reveals that the ice margin remained stable for a long period, longer than in the NNE, for example in the Hyrrönharju esker, where kettleholes suggest a two-phase ice marginal position (Figs 11, 47).

The concentration of coarse material and the palaeocurrent measurements suggest that meltwaters ran mainly from the NE, from one single supply source (Fig. 21). The distribution of the foreset units indicates that the delta was deposited as a result of delta front progradation and progressive advance of steep foreset slopes (delta fronts).

The main barform subunit F1 in foreset facies was deposited through the action of strong southwesterly meltwater currents as indicated by structural analysis (Fig. 14) and palaeocurrent data (Figs 15, 20). The fact that the bottom contact plunges towards the southwest indicates erosional effect of the current on the substrata (Fig. 17). The fan-shaped distribution of palaeocurrent directions and the cone-shaped lobe of the unit (Figs 14, 15, 20) indicate that under stable flow conditions divergent flow ran over the up-grown ellipse-shaped top of the bar (unit). The southeastern flow direction on the southeastern side of the barform (unit F1b) (Fig. 20) resulted from meltwater currents concen-

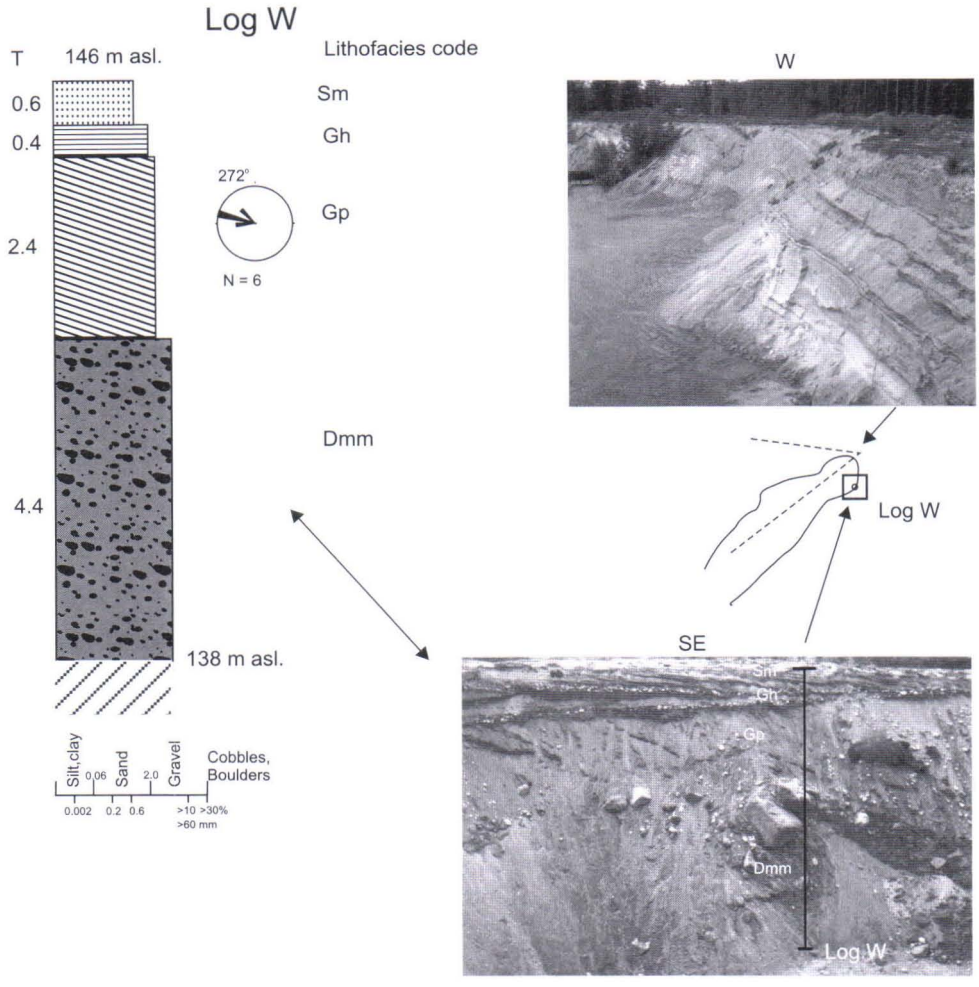

Fig. 19. Log W. Exposure on the proximal slope in the western part of Vuonteenmäki. Coarse facies of the ice contact with diamicton in its lower part and foreset facies with a westerly dip in its upper part. Ripple and horizontally bedded sands, fine sands and gravels of the upper picture occupy a higher position in the lithostratigraphy than the sedimentary units mentioned above.

trating on this side of the unit after the deposition of the main unit.

The main barform (unit) is interpreted to have been deposited mainly as bedload avalanches and mass flow (Martini 1990, p. 292). The foreset beds generally dip over 15°, usually 20-30°. Beds are 0.1-2.0 metres thick. The matrix varies; beds without matrix or supported by gravelly matrix occur in places. The absence of matrix may have resulted from a flow slide process (Colella *et al.* 1987). Beds are usually normally graded but some reverse grading also occurs. Sedimentary units with a greater proportion of fines and less sorting are present in places. These units with diffuse contacts were evidently deposited by debris flow mechanisms due to collapse events on the foreset unit upper edge.

The deposition of unit F2 (Figs 20, 21) started after the deposition of F1, and seems to have developed much the same way as F1. The material however is finer on an average. The deposition of these units was followed by the deposition of unit F3 between them (Figs 14, 15, 16), as the meltwater currents converged along the depression between the older units (Figs 20, 21). This event produced smaller scale subunits with more varying palaeocurrent directions than in F1 (Fig. 16). The depression that occurs northeast of the main exposure possibly forms part of the potential feeding route.

Unit 4 (Figs 20, 21) was deposited in a depression northwest of the F1 conformably on top of and next to it (Fig. 14). According to the GPR data, unit 4 was laid down upon finer sediments in the northern part of the area. The conformable basal contact of the unit reveals that it is basin infill sediment deposited under low energy conditions. The inner structure of the unit is composed of beds with smooth contacts indicating cohesionless debris flow deposition from the north and northeast. The lower contacts of the internal strata of the unit are tangential.

The delta topset facies units are probably bedload gravel and waning-stage sand sheets. The flatness of the hill top and the associated topset

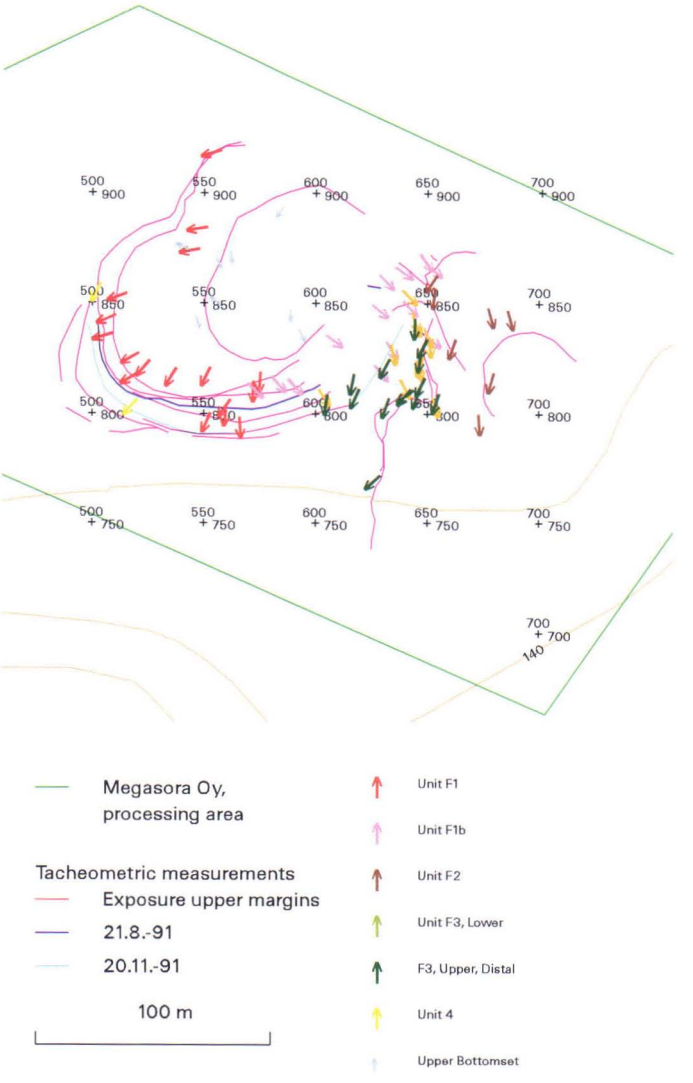

Fig. 20. Combined palaeocurrent data map from the Vuonteenmäki site. See text, pages 35, 37, 38.

units (Figs 14, 17) indicate that material continued to be deposited at the waterlevel, i.e. the highest shoreline level. Thus the deposit is a typical Gilbert delta (Gilbert 1885, Nemec & Steel 1988, Ashley *et al.* 1991, p. 117). The lack of meltwater drainage network indicates that material supply stopped during the delta stage and consequently the sandur stage never developed. The uniformity of the distal littoral ridge indicates an ice marginal retreat before the water level drop. Otherwise the erosional effect of meltwaters would have eroded the littoral ridge in part or completely. Shoreline deposits are also found in the surface of the western section (Fig. 19, top).

The distal horizontally bedded sands, ripple sands and fine sands were deposited in low energy depositional environment on the lower slopes of the delta after the coarser material had been deposited in the foreset units. Deposition took place through saltation and settling from suspension; bedload transport was not common although it is observed in places. The fine silts of the lower part of the unit were settled out from suspension. Instead, in the upper bottomset facies, climbing ripple-drift cross stratified, draped lamination and horizontally bedded sands were deposited by turbidity flows on the bottom of the basin (Gustavson *et al.* 1975, Ashley *et al.* 1991, p. 119). However, collapse of foreset slopes was not the cause of the turbidity flows, because only small-scale distal transport of coarse

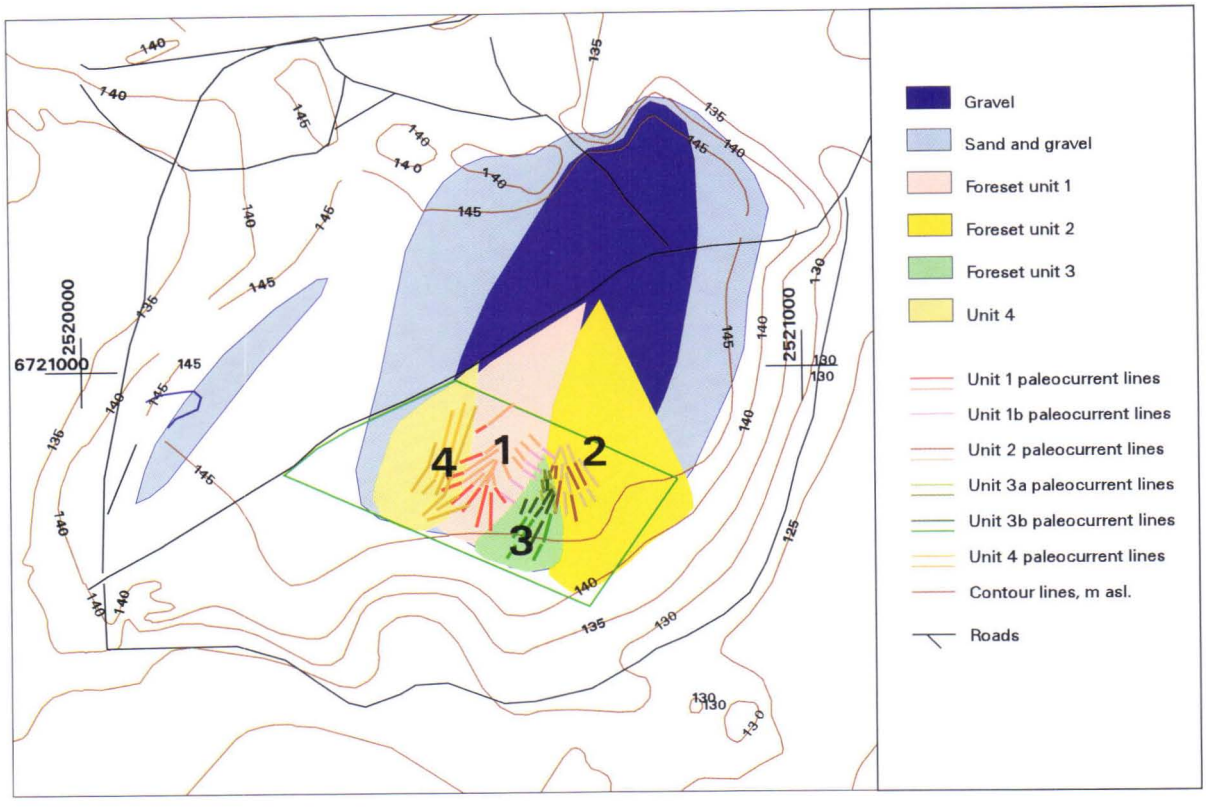

Fig. 21. Interpretation of palaeocurrent data at Vuonteenmäki. See text, pages 35-38 for explanation.

material took place and the fine material was supplied by submerged turbid (hyperpycnal) flows formed by meltwater with suspended loads (Martini 1990, p. 293). Distally, the pulsating nature of the supply is reflected in the cyclical decrease of particle size and cyclic structural variations (Fig. 18).

Fluidisation produced load structures as a result of especially rapid distal deposition (Fig. 18, bottom left). A similar deformation unit is found at Polari in Karkkila, southeast of the wells. Glückert suggested (1995, p. 54) that this deformation structure was formed in association with the rapid Baltic/Yoldia water level drop.

In the western Vuonteenmäki, on the northwestern side, sedimentary conditions were similar to those in the distal part. In consequence, horizontally bedded and rippled sandy and partly silty units also dominate at this site (Fig. 19, top). However, a stony and bouldery unit indicating ice marginal contact occurs in this area as well (Fig. 19, bottom).

From the esker system starting at Vuonteenmäki (Fig. 47) the next stable ice marginal position is in the Levoilammi-Säynäislampi area, where an ice marginal position pitted with kettleholes occurs north of the old railway line.

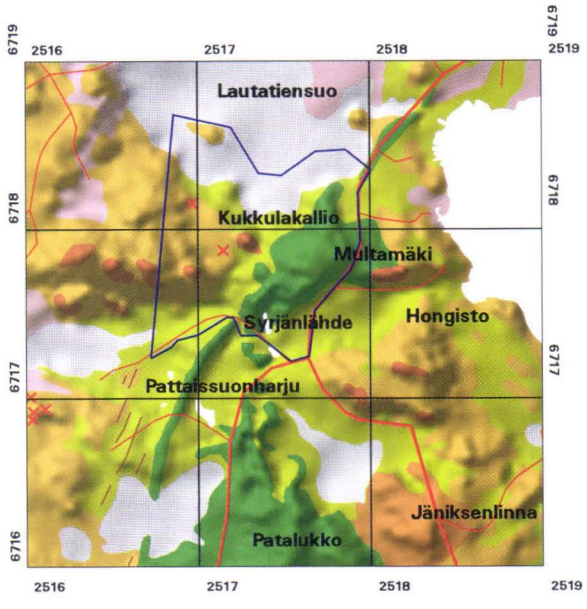

Fig. 22. Hillshaded Quaternary geology map at 1:40, 000 scale with 1x1 kilometre squares showing the location of Multamäki. Inset shows the location of the hillshaded 3D map (Fig. 23).
Elevation contour data © National Land Survey of Finland.

3.2.2. Multamäki, Karkkila

The Multamäki site is located in the Outer zone

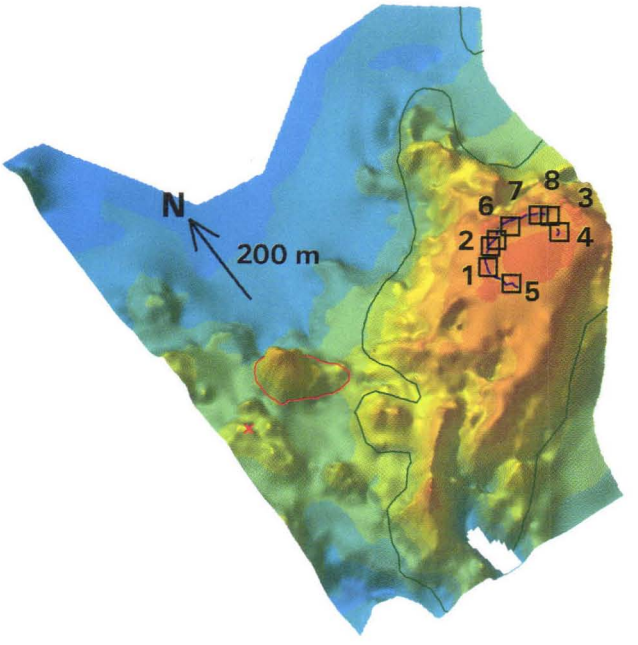

Fig. 23. Colour classified relief view of Multamäki as seen obliquely from the SW showing upper edge of the exposure and logging sites. See Fig.12 for colour classification legend. Scale: north arrow is 200 m long.
Elevation contour data © Karkki-Rasti ry 1995.

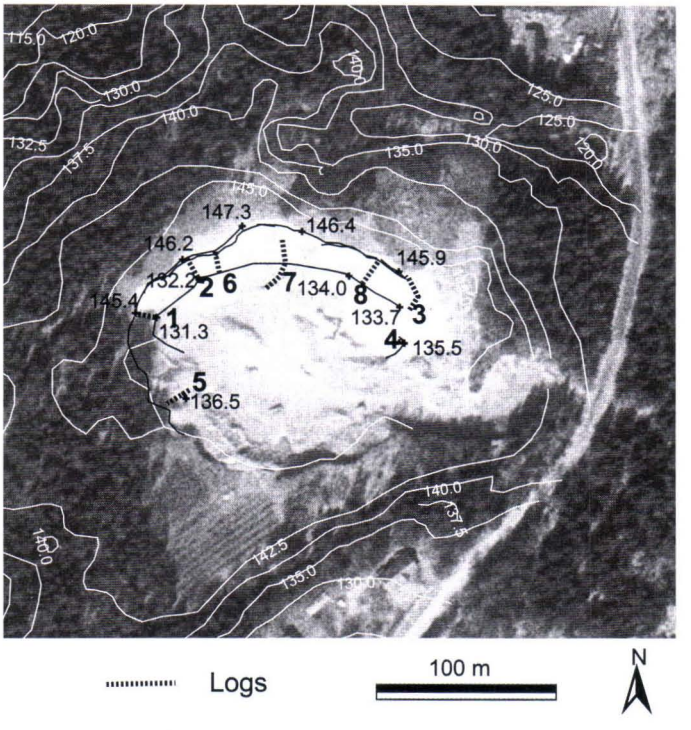

Fig. 24. Aerial photograph at 1:4, 000 scale from Multamäki. Upper and lower margins of the exposure measured with tacheometer. Numbers show log points.
Elevation contour data was digitised from an orientation map © Karkki-Rasti ry 1995. Aerial photograph © Topographic Service of the Finnish Defence Forces.

of the Second Salpausselkä (Figs 5, 10). Before the deposition of the Second Salpausselkä ice marginal deposits a meltwater drainage route had probably run through the western part of the site towards the south to the Patalukot (Fig. 22). Multamäki represents a large scale polygenic landform of the Second Salpausselkä.

Multamäki is situated at the western margin of a valley depression (Figs 10, 22). The relief of the area is 40 m (Figs 23, 24). The surrounding bedrock is composed of gabbro and diorite, and in the south microcline granite also occurs (Fig. 3) (Härme 1953, 1954). To the south of the site a fracture zone runs east-west (Fig. 4).

At the time of deposition, water depth was 30-40 m as calculated from valley bottom levels, when the highest shoreline level is estimated to have been around 145 m a.s.l. (Figs 10, 48). The top of

the ridge was deposited at the highest shoreline level (Figs 23, 24). On the eastern side of the ridge, about 70 m north of the exposure, a sharp-featured depression running east-west is interpreted to reflect the ice marginal position of the early stages of the deposition.

3.2.2.1.Lithostratigraphy and sedimentary structures

The stratigraphic sequence and lithofacies observations were made from a sand and gravel pit exposure situated at the top of the Multamäki ridge. The areal extent of the pit was 120x120 m and the main exposure was oriented between the west and north along the 120 m of the main section (Figs 23, 24, 25). The main exposure was up to 15 m high. On the east side of the exposure a face with a maximum height of 12 m (logs 3 and 8, Fig. 29) was examined. Stratigraphical logs 4 and 7 (Fig. 28) were obtained from test pits that were made in the lower part of the exposure.

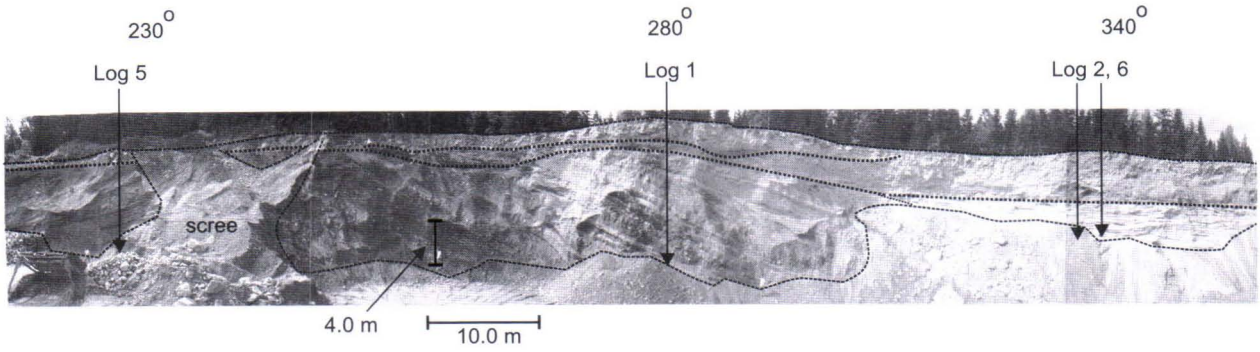

Fig. 25. Photographs laid out as a mosaic showing the main exposure of Multamäki 26.07.1991. Note the difference in composition to logs 1, 2, 5 and 6, observed in July 1992. The picture shows the contacts of diamicton, topset, foreset and distal facies.

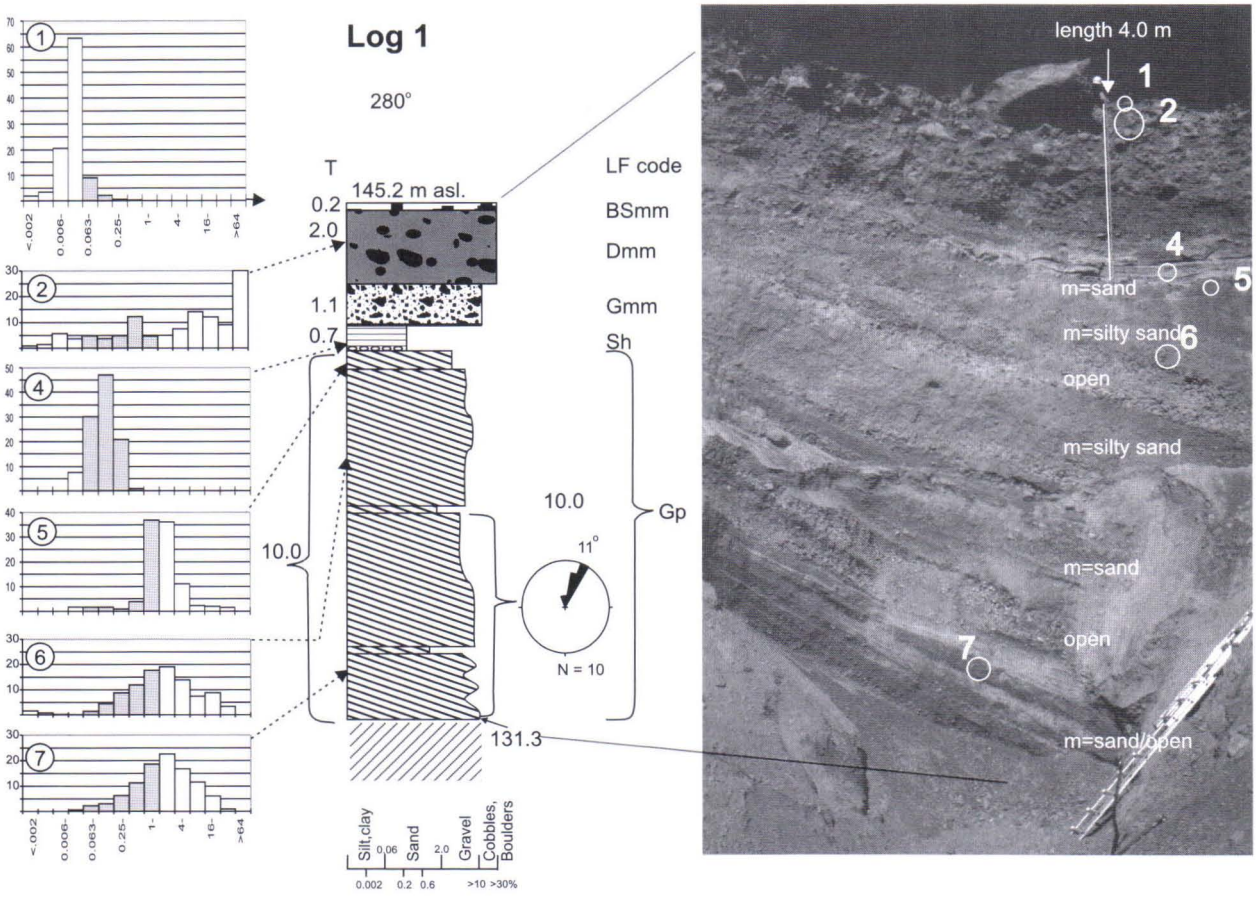

Fig. 26. Log 1 at Multamäki. Planar cross-bedded gravel (Gp) sedimentation unit is interpreted as a foreset unit, Gmm a topset and Dmm a diamicton unit. Sampling sites are marked with white circles. Varying matrix of the individual beds of the foreset are also shown in the picture.

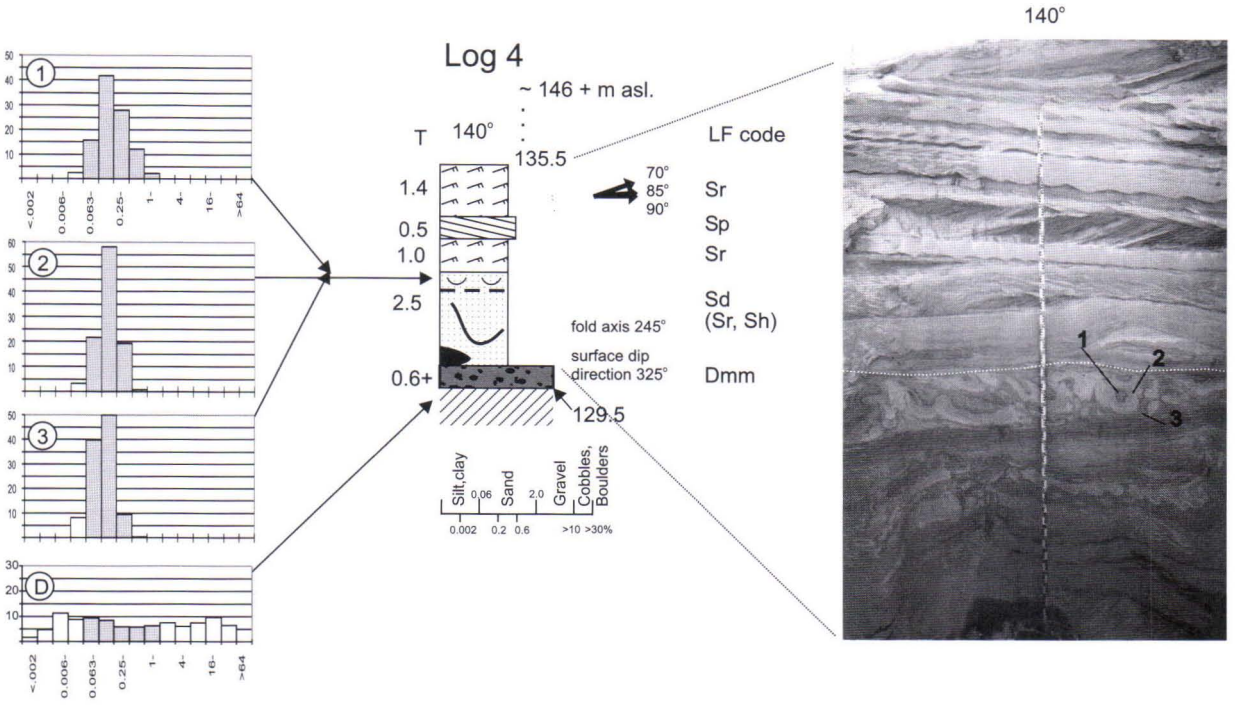

Fig. 27. Log 4 at Multamäki from the base of the exposure. Diamicton of the lowest part is not visible. Ball and pillow samples: 1 coarse, 2 medium coarse, 3 fine. Deformed sedimentation unit is interpreted to have been deposited at the bottomset stage, and overlying units at the distal stage.

Stratigraphic sequences were logged at four points in the main exposure (logs 1 (Fig. 26), 2, 5 (Fig. 28) and 6). The main sedimentary unit of log 1 is a planar cross-bedded, coarse, partly stone-rich gravel unit, in which the matrix of individual beds varies between sand, gravel and fine sand or is partly absent. The mean dip direction of the unit is 11° NNE. Massive diamicton, covered by stones and boulders is at the top of the massive gravelly sedimentary unit in the upper part of the observation point. This unit is in turn covered by a thin unit (0.2 m) with boulders and stones in a well sorted fine sand matrix (Fig. 26). The main sedimentary unit of logs 2 and 6 dips towards NNE (log 6, 14°)

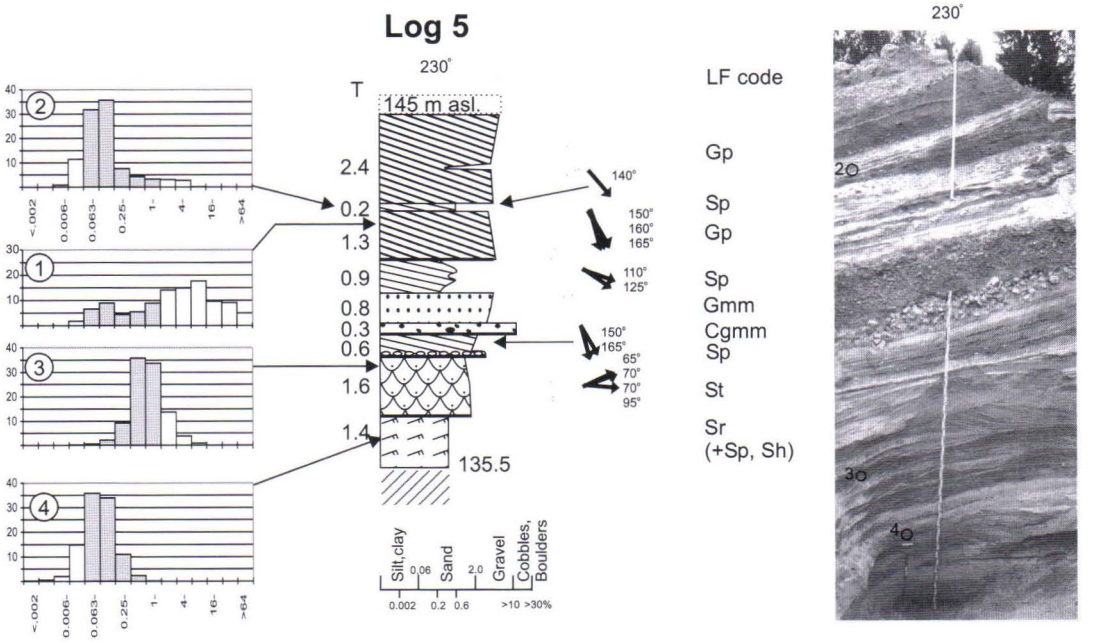

Fig. 28. Log 5 at Multamäki. In the lower part palaeocurrent direction is from the NE-ENE. In the upper part the foreset unit is over 6 metres thick with a bedding plane dip direction towards SE.

and lacks diamicton cover. An intermediate unit covering the planar cross-bedded gravel unit is primarily composed of ripple sand and becomes thicker eastwards.

A massive unit of diamicton with the upper contact dipping towards the northwest was located in the lower part of the section at log point 4. A deformed unit, originally composed mainly of ripple sand and with loading structures such as ball and pillow and flames rest upon diamicton (Fig. 27).

In the eastern part of the exposure, logs 3 and 8 show a continuous 6 m sequence where the main lithofacies is ripple sand. In its upper 2 m loading structures are common. At deeper levels, below 137 m a.s.l., horizontally bedded and trough cross-bedded sands were also observed. The upper part of the sediment sequence is composed of a surface unit of massive coarse gravel with erosional contacts that is overlain by stone-rich gravel. A sedimentary unit of planar cross-bedded gravel was observed from the log point 7 at a depth of 131-133 m a.s.l. with mean dip direction of 75° (5 measurements) (Fig. 29).

3.2.2.2. Interpretation of the development of Multamäki

Prior to the deposition of the Second Salpausselkä´s Outer zone a meltwater route ran southwards passing by the western side of the study site through Syrjänlähde towards the Patalukko delta as shown by morphological studies (Fig. 22). The interpretation is based on the identification of the sediment feeding system for the Patalukko delta on the basis of the orientation and distribution of glaciofluvial ridges and kettles.

The first step in the ice marginal deposition of the Multamäki ridge was the deposition of a small ice marginal ridge composed mainly of basal till that was observed in the basal part of the delta at log point 4 (Fig. 27).

During the main depositional stage, glacier meltwater supply came from the west via the southern side of Kukkulakallio as interpreted on the basis of glaciomorphological (ridges and kettles) (Fig. 23) and palaeocurrent studies (Figs 26, 28). Flow direction ran almost at a right angle to the meltwater supply direction that preceded the deposition of the Second Salpausselkä.

A Gilbert-type megaforeset unit with a maximum thickness of 10 m (Fig. 26 = log 1) was deposited in front of the meltwater route exit. The foreset unit consists mainly of gravel and coarse gravel, with larger clasts in places. The matrix of individual foreset beds varies from fine sand to fine gravel and is absent in part of the beds (Figs 26, 28,

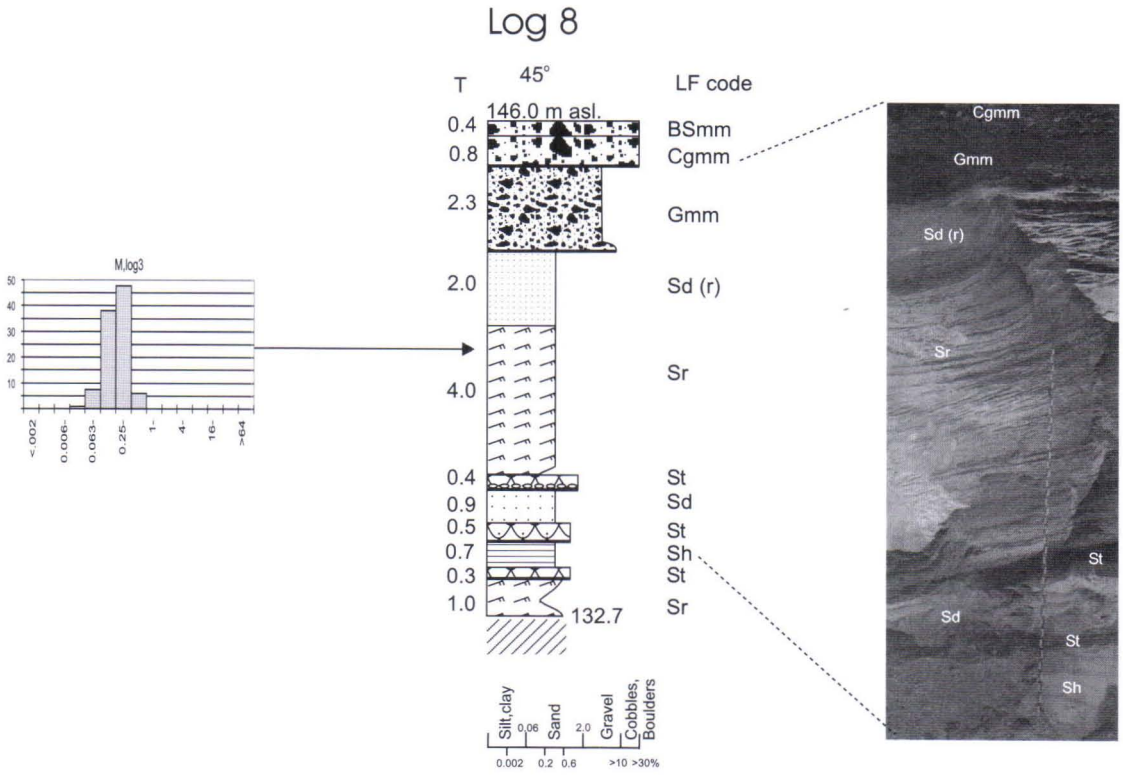

Fig. 29. Log 8 at Multamäki. The major part of the sedimentation units of the stratigraphical sequence is composed of the distal facies characterised by ripple sands deformed in their upper part which is thought to be caused by loading due to rapid deposition.

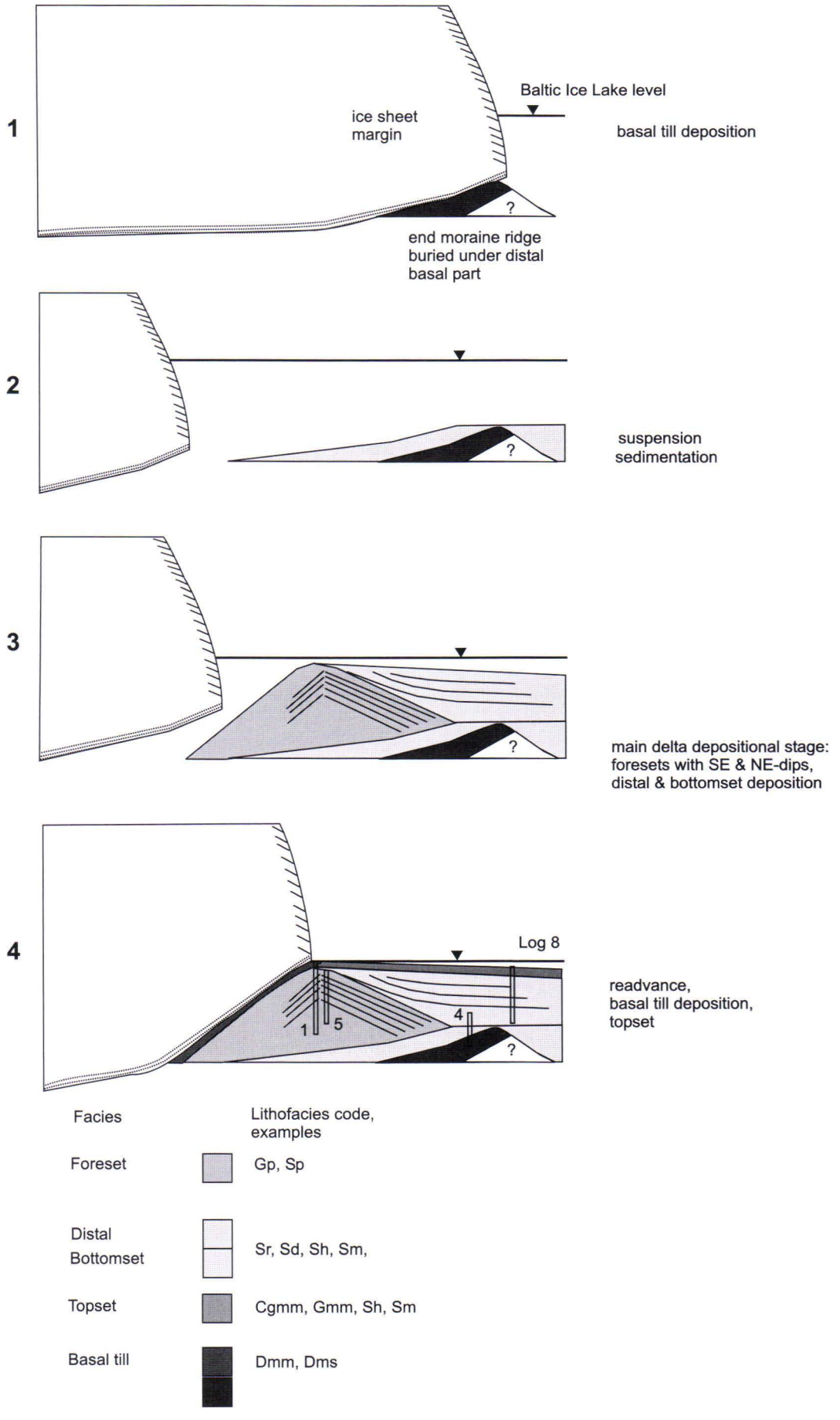

Fig. 30. Overall view of the depositional stages at Multamäki illustrated by a series of drawings: 1) diamicton-dominated ice marginal ridge was deposited, 2) ripple and horizontally bedded sands and fine sands, partly silty, were deposited from ice frontal turbidity flows and suspension, 3) the main depositional stage involved the deposition of foreset facies with bottomset sediments formed in front of them, and 4) readvance of the glacier partly on top of the deposit creating a proximal basal till type-diamicton facies.

logs 1 and 5 and corresponding photographs). On the basis of exposure face data the meltwater was initially fed eastwards (logs 4 and 5, Figs 27 and 28, lower part and log m7) and then towards the NNE (log 1, Fig. 26, also logs 2 and 6), and finally towards the southeast (log 5, Fig. 28).

A bottomset/distal unit formed beneath and in front of the foreset unit (Fig. 29). The units are mainly composed of ripple sand and partly horizontally bedded sand. The deposition took primarily place from suspension and was rapid, which is indicated by the abundance of loading structures (logs 4 and 8, Figs 27, 29, also log 3). The topset coarse and stone-rich gravel was deposited in the upper part of the delta, at the highest shoreline level (log 8, Fig. 29, also logs 2, 3, 6).

After the main stage of deposition the ice margin readvanced to override the proximal part thus creating the basal till type diamicton (Figs 25, 26). This in turn passes into the topset stone-rich gravel unit.

During the last stage of deposition the site was covered by aeolian fine sand (log1, sample 1, Fig. 26) which is similar in origin to the Lammi loess (Jauhiainen 1972). Loess was formed in association with the Baltic/Yoldia 25-28 m water level drop when large areas became exposed to aeolian activity.

To summarise (Fig. 30): Multamäki is an embryonic, i.e. incomplete delta. The main stage of deposition was preceded by the deposition of a small scale De Geer type diamicton-dominated ice marginal ridge.

During the main stage of deposition, meltwater was fed from the west. The meltwater flowed towards the east, north and southeast. The location and stucture of the megaforeset may have been influenced by 1) the ice margin that most likely was north of the exposure and 2) an older, east-west-oriented ice marginal ridge in the southern part of Multamäki. The last stage of deposition was followed by a short oscillation of the ice front against the proximal part of the ridge.

3.2.3. Tupsumäki, Karkkila

Tupsumäki is situated in the Central zone of the Second Salpausselkä. The ridge is located on the southeastern side of the Siikala valley (Figs 5, 10, 31). The Tupsumäki area consists of ice marginal ridges composed of both glaciofluvial and diamicton dominated sediments, which form a 1 km wide zone. Southwest of Tupsumäki, at Huhtimo diamicton-dominated (consisting mainly of basal till) ice marginal ridges occur occupying ten separate ice marginal positions.

The bedrock topography of Multamäki is not known. Possibly the bedrock level rises towards the SE. The present relative height is around 20 m in the southeast and 30-40 m towards the northwest (Figs 32, 33).

In the western part of Tupsumäki there is a

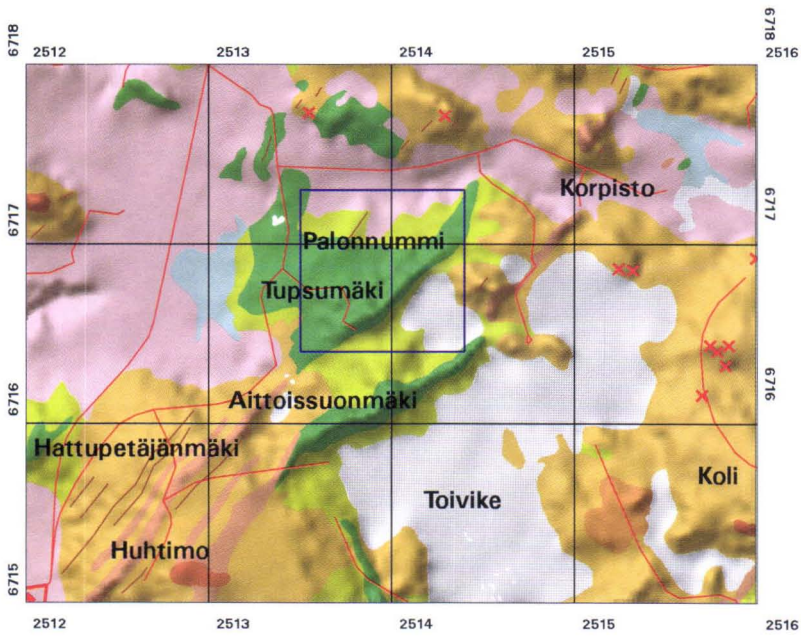

Fig. 31. Hillshaded Quaternary geology map at 1:40, 000 scale - a revised version of the Geological Survey of Finland Quaternary geology map at 1:20, 000 scale. See Fig. 11 for explanations. Inset shows location of the hillshaded 3D map.
Elevation contour data © National Land Survey of Finland.

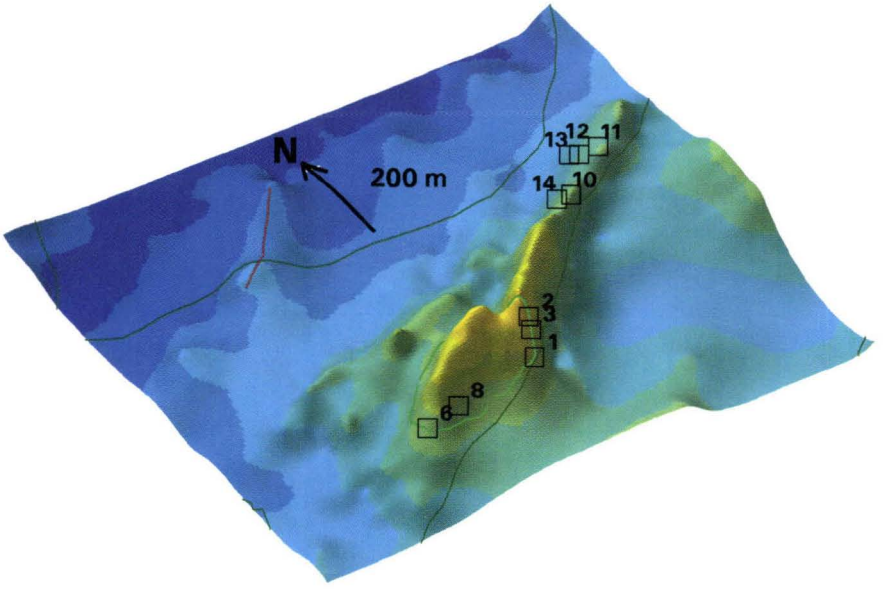

Fig. 32. Colour classified hillshaded relief view of Tupsumäki from the SW (210° with a 45° vertical angle). The relief has been processed from elevation contour data of various ages, most of which was provided by National Land Survey of Finland in numerical form.
Elevation contour data by © National Land Survey of Finland.

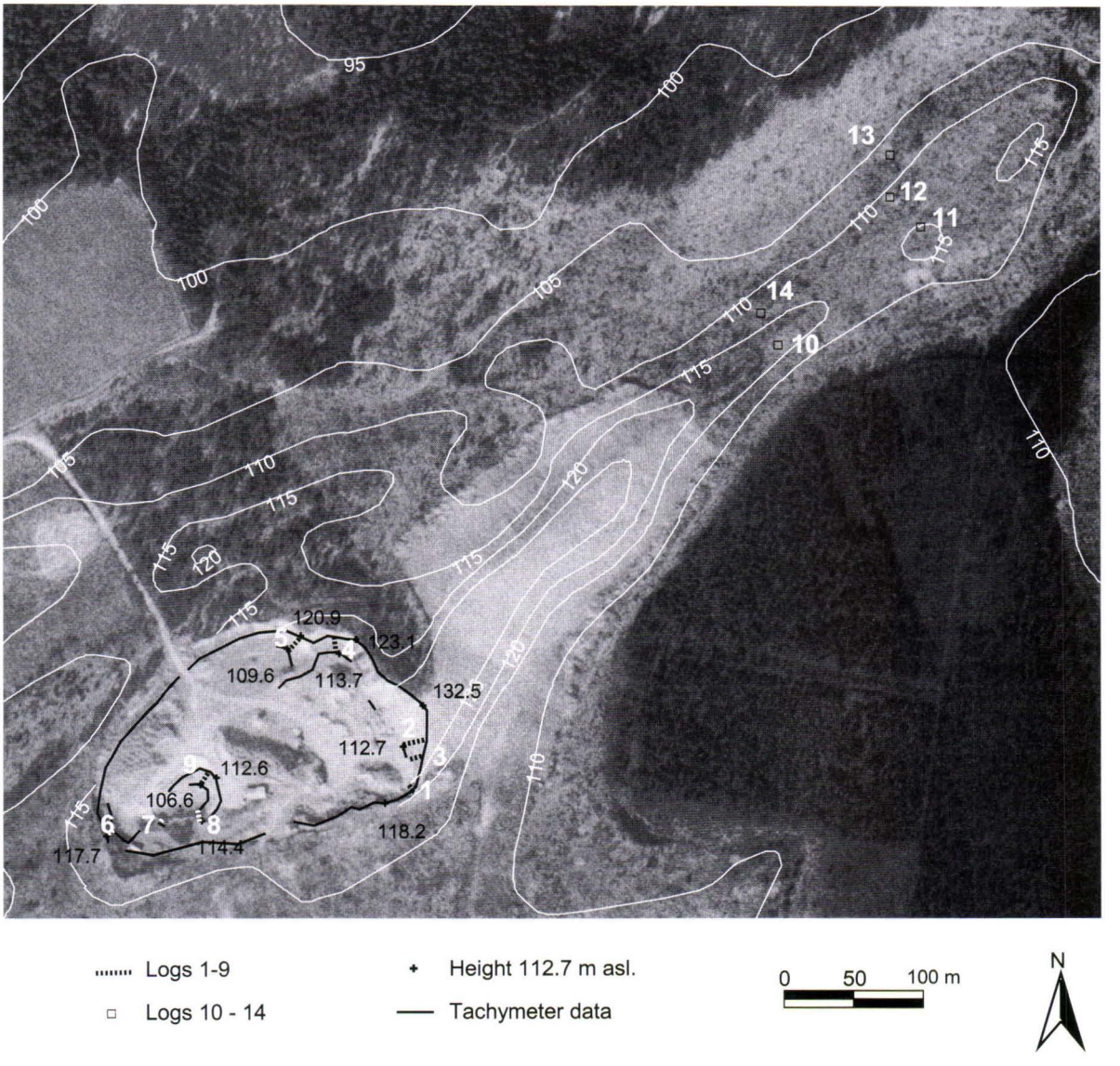

Fig. 33. Aerial photograph at 1:5, 000 scale from Tupsumäki. Tacheometer and elevation data by J.-P. Palmu. Log points and test pits and exposures denoted. Aerial photograph © Topographic Service of the Finnish Defence Forces.
Elevation contour data © National Land Survey of Finland.

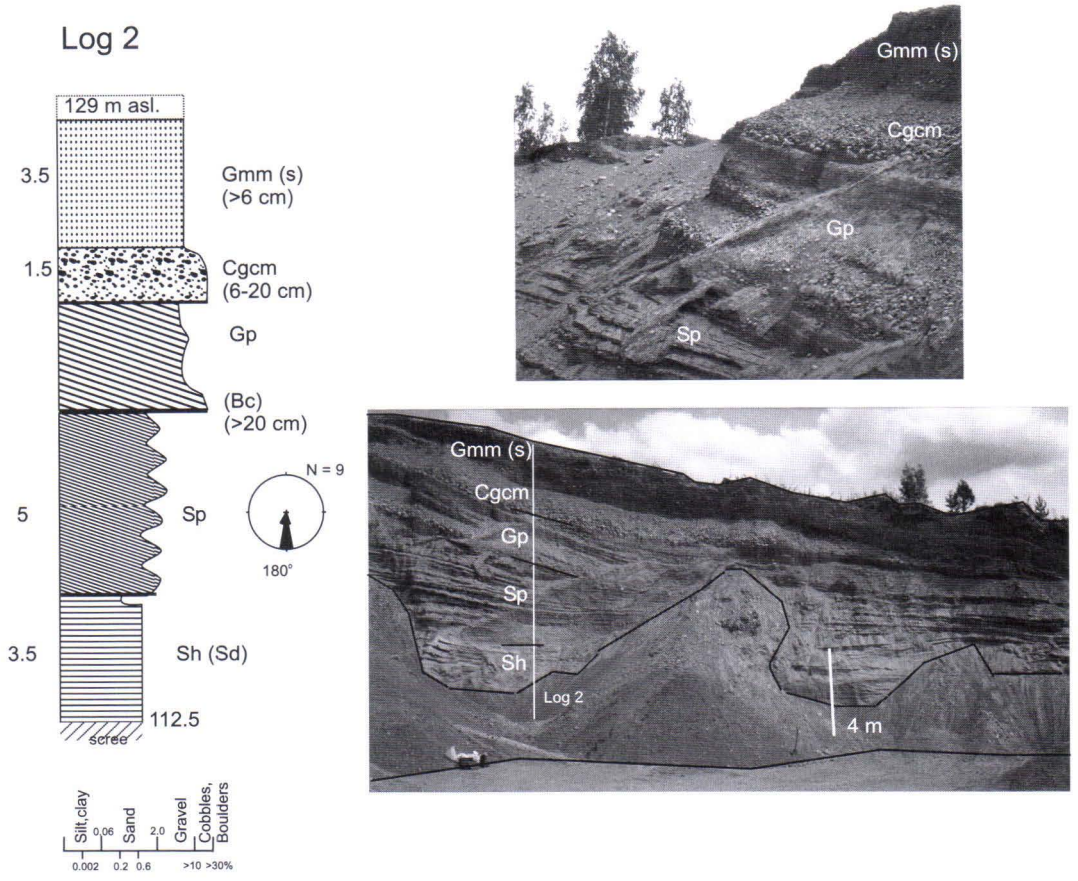

Fig. 34. Log 2. Foreset facies from Tupsumäki and its erosional contact with bottomset /distal horizontally laminated sand. Sp bottom part of the foreset is truncated distally.

proximal, smaller subsidiary ridge. According to field observations the deposit includes five successive marginal positions. Geomorphological analysis from orientation maps indicate that the original shape of the main ridge has been symmetrical. Steep slopes exist east of the proximal kettle (see below). Both the distal and proximal slopes are very steep, up to 25°. In the eastern part of the ridge north of Levoissuo only one asymmetrical low ridge occurs with a gentle (5°) proximal slope and a steep distal slope.

Tupsumäki has derived its name from the 'peg'-shaped top where pine trees grow on the hill looking like a tuft ('tupsu' in Finnish). This highest point of the hill has now been removed almost completely. The log 2 (Fig. 34) was done close to the top point. Northeast of the top, there is a kettlehole that is interpreted to indicate the eastern margin of the meltwater supply route.

The bedrock of the Tupsumäki area is composed of gabbro and diorite (Härme, 1953). A fracture zone trending east-west runs in the northern part of the site. In the western and eastern parts two fracture zones occur trending NW-SE (Fig. 4).

During the initial stage of deposition water depth was at least 40-50 m, calculated from the present elevation data, and assuming that the highest shoreline level was ca 145 m above the present sea level. To the northwest of the site water depth exceeded 60-70 m. During the final stage of deposition, water level was only around 10 m above the top of the ridge (Figs 32, 33, 10, 48).

The log points and test pit locations are shown on an aerial photograph map (Fig.33). Lithostratigraphical loggings 1-9 were performed in the exposure located in the western part of the ridge. Five test pits were dug and logs 10-14 were measured in the eastern part of the ridge.

3.2.3.1.Lithostratigraphy and sedimentary structures

A unit of massive compact diamicton occurs in the lower part of the distal margin (log 8) and represents the oldest sedimentary unit of Tupsumäki. The upper contact dips 310/36° and the till fabric trends from the north (Fig. 36). The diamicton is similar to the basal till type diamicton that occurs in the Hirvilammi site.

The lower part of Tupsumäki deposit is composed of a distal bottomset sedimentation unit. It

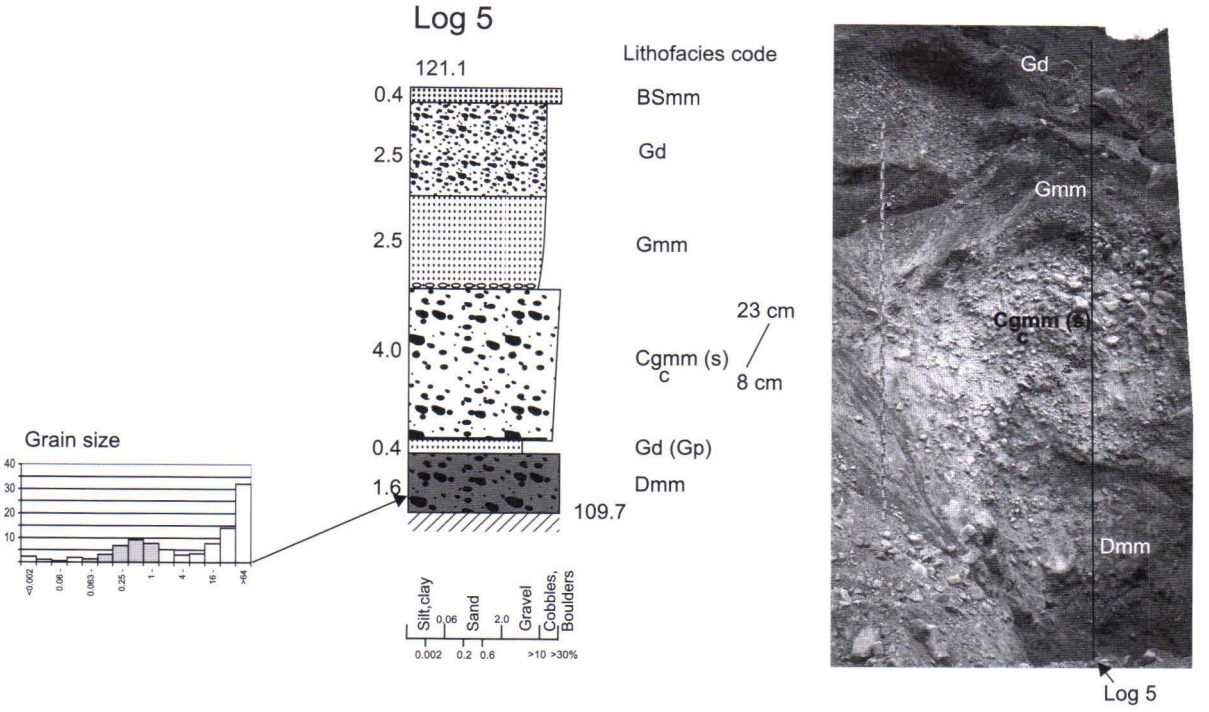

Fig. 35. Log 5. Proximal ice contact of Tupsumäki - bottom units are of debris flow origin.

consists of ripple sands, horizontally bedded sands, massive sands, and fine sands (Figs 34, 36). Gravel beds are less common, usually massive and are clast supported or supported by varying types of matrix. Horizontally laminated silt also occur, e.g. at log point 7 two 20 cm beds and at log 9. The beds are partly poorly sorted. The ice marginal ridge that occurs in the basal and distal position of the main ridge influenced the bedding plane of the sands, causing a low proximal tilt on its proximal side. A palaeocurrent direction of 186° (7 measurements) was measured for a ripple sand at log point 7.

The upper level of Tupsumäki is composed of a Gilbert-type megaforeset (Fig. 34). The maximum thickness of the unit is over 15 m. The foreset is mainly composed of well sorted material including, however, some beds with poorly sorted material and diffuse contacts. Especially the upper megaforeset is coarse and the beds dip steeply (log 2, photographs) (Fig. 34). The mean palaeocurrent direction in log 2 is 180°. From log 1 the mean palaeocurrent direction is 164° (114 a.s.l., 10 measurements) (Fig. 34). In the western part of the main exposure, at ca 116-110 m a.s.l., massive coarse, partly stone-rich gravel occur, the matrix of which is mainly gravel and coarse sand and partly clast supported (without matrix) (log.6, bed thickness at least 6 m).

In the eastern part of the ridge, at log point 10, the material is mainly composed of stone-rich massive cobble gravel. At the top of the ridge the bedding plane of horizontally bedded gravel is bent towards the distal side and parallel to the tilt of the slope surface. At log point 11 stone-rich gravel is at the depth of 2.4 m. Horizontally bedded gravel overlies stone-rich gravel with palaeocurrent direction 133° (4 measurements). A shear zone overlain by massive fine gravel runs at the depth of 1.2-1.6 m.

Basal till type diamicton was not found in the proximal part of the main section. However, coarseness of material and highly variable particle size range is typical of the proximal ice contact zone (Fig. 35, log 5). The zone also includes boulders, e.g. southeast of log 5, 5 boulders were found. The ice contact unit also includes massive diamictons interpreted to be of debris flow type (Fig. 35, log 5, lower part).

Glaciotectonic activity in the ice contact zone must have taken place in the main area of the exposure at log points 4, 5, and 6. An earlier observation on glaciotectonism was made in the western part of the exposure (Palmu 1990, p. 70-71). In this area sand and gravel beds were folded due to shearing caused by the ice oscillation. The erosional contact has a gentle proximal dip. Six reverse faults were measured at log point 4 with a dislocation of 7-60 cm, a 75-85° dip, and a dip direction of 270-330°. The reverse faults reflect glacial stress. No normal faults were observed, although they are likely to occur on steep proximal

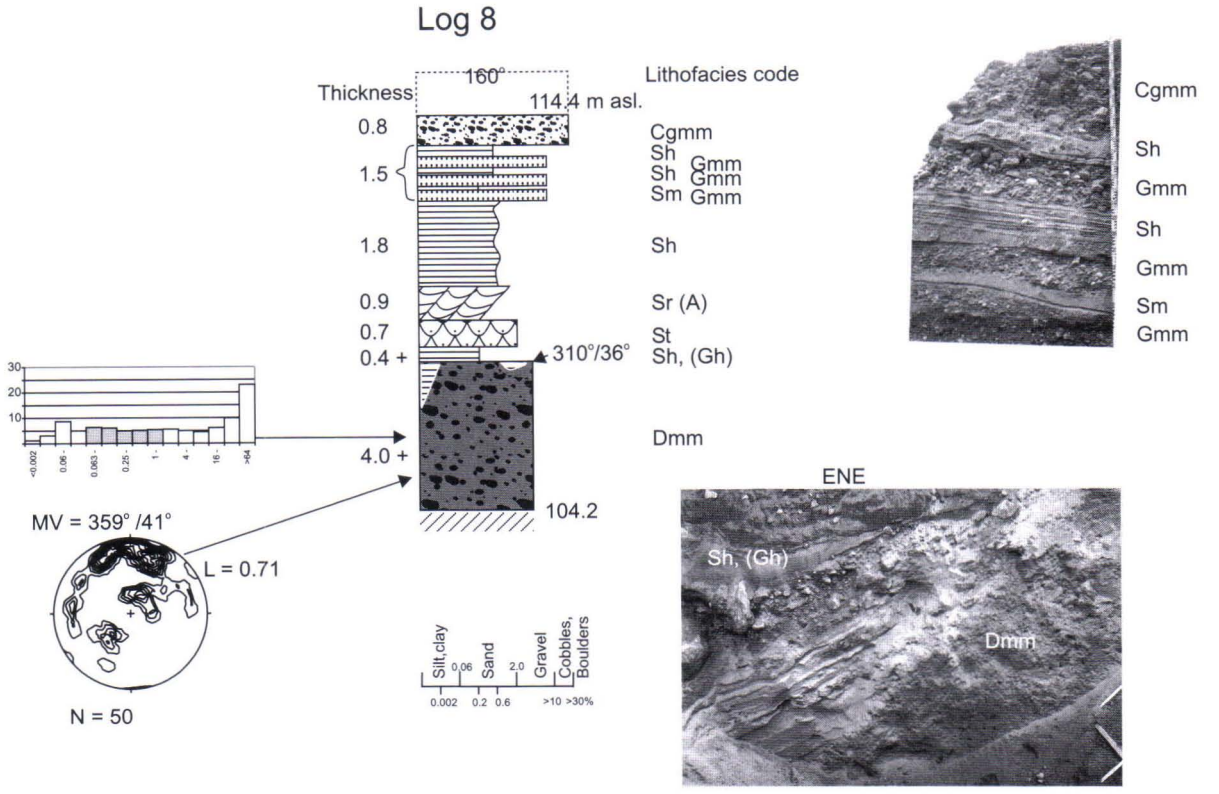

Fig. 36. Log 8 at Tupsumäki. Section in the lower part of the distal margin. Basal till-type diamicton is shown in the lower part. Gravel and sand units of the bottomset/distal facies seen in the upper part.

slopes (Lønne 1993).

In the eastern part of Tupsumäki, proximal slope surface is mainly composed of either deformed glaciofluvial material or basal till type diamicton (logs 13 and 14) (Fig. 38). At log point 11, where most material is planar cross-bedded gravel, a shear zone was discovered at the depth of 1.5 m from the ground surface, with shear planes towards 295° (2

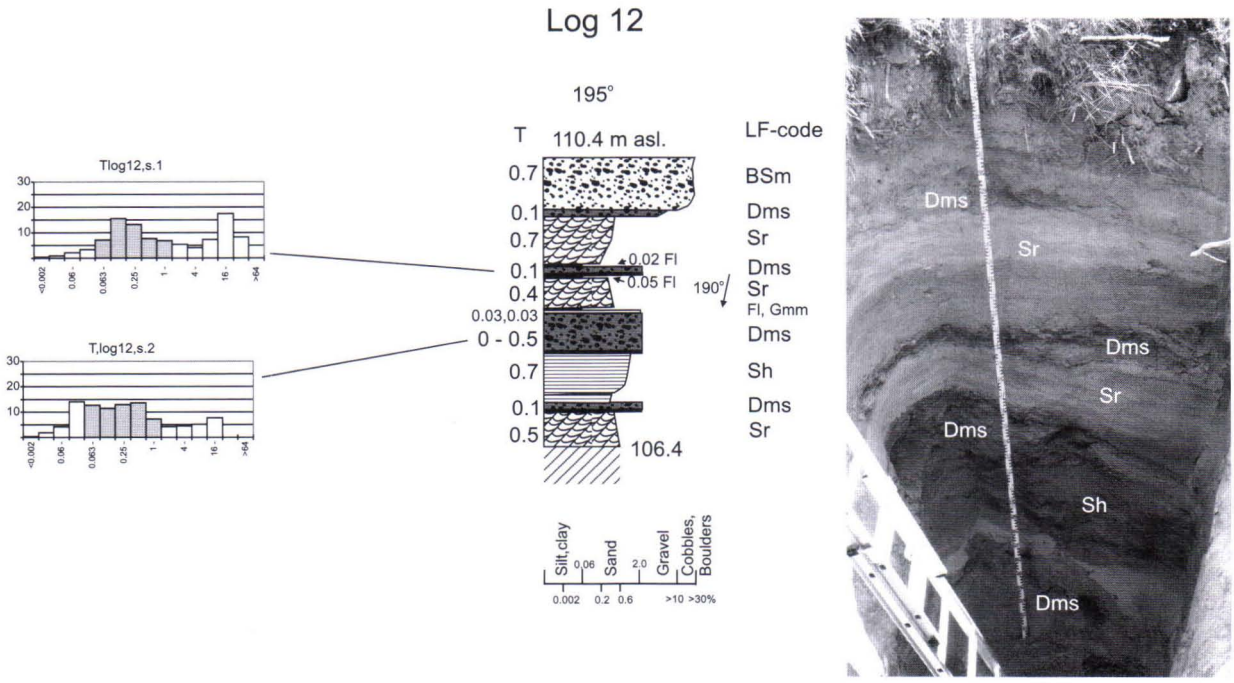

Fig. 37. Log 12 at Tupsumäki. In the eastern part, ripple sand, diamicton beds of debris flow origin and silt (laminated silt from suspension sedimentation) alternate in the test pit.

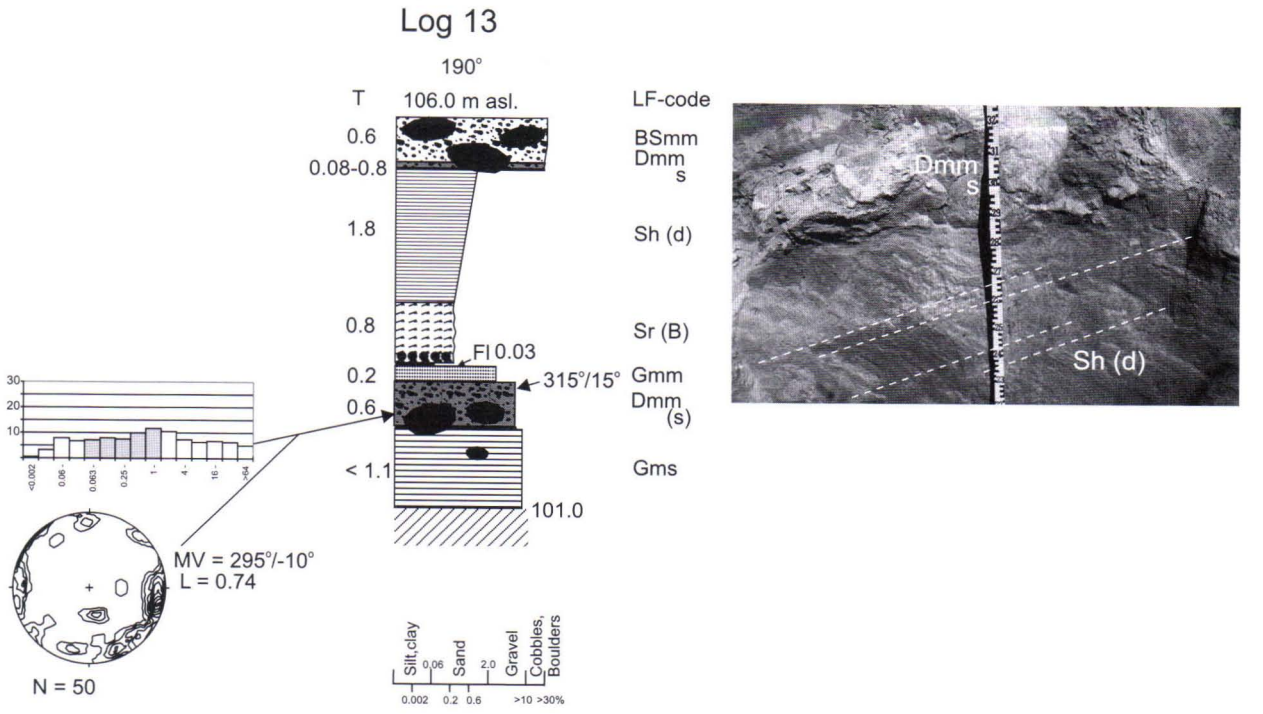

Fig. 38. Test pit located in the lower part of the eastern proximal slope at Tupsumäki incorporates two basal till-type diamicton facies interbedded with a sand unit deformed through shear that was caused by glacier oscillation.

measurements), dips of 15 and 28°, and fold axis direction of 40 and dip of 12°. The palaeocurrent direction measured above the shear zone is 133° (the mean value of 4 measurements).

At log point 13 two basal till type diamicton units were found, covered and interbedded with units of sand and gravel below and between them. The lower contacts are erosional between sand and silt. Below the upper diamicton the mean value for reverse fault dip direction is 309° with a dip mean of 21° (6 measurements) (Fig. 38, photograph). Ripple sands and debris flow type diamicton and silt were deposited from suspension upon the eastern proximal slope probably under (winter time) low flow conditions (log 12).

3.2.3.2. Interpretation

The fact that a diamicton-dominated ice marginal ridge was found in the lower part of the distal margin of the Tupsumäki ridge indicates continuous and multi-phase ice marginal oscillation. This is also indicated by the push moraine ridges that occur on the proximal slope of Tupsumäki.

A distal bottomset facies unit was deposited on top of the above mentioned small distal ridge. Massive, horizontally laminated and ripple sands as well as fine sands in the lower part of the distal margin are interpreted as being equivalent to the outer depositional zone of a plume (Hunter *et al.* 1996, p. 1033). Martini (1990) describes similar units as having been deposited by 'submerging turbid (hyperpycnal) flows' (p. 293), or 'high-density underflows formed by suspended loads' (p. 294) with reference to Postma and Roep (1985) and Colella *et al.* (1987).

In addition, coarse thin gravel units at Tupsumäki were deposited by traction currents whereas fine-matrix gravel units were products of debris flows. Ripple and horizontally bedded or massive sands are developed through a combination of traction currents and suspension settling, although varying proportion may represent turbidites (Hunter *et al.* 1996, p. 1033, Martini 1990, p. 293).

Turbidites were most likely originated from meltwater or in some cases they may result from steep upper foreset collapses. This is suggested by the erosional lower contacts that occur in some units (logs. 1, 2, 3). Silt beds settled out from suspension. Also some dune-bedded sands (log 7, log 1 lower part) indicative of stronger water flow were present. Poorly sorted debris flow type sands with fines as well as gravel are also present in both the distal bottomset and the foreset facies. No deformed sands of fluidised flow type were discovered in Tupsumäki, in contrast with Vuonteenmäki and Multamäki.

The megaforeset was probably fed from a source, where the level of the efflux jet exit rose gradually with the growth of the ice marginal ridge. The

location of the feeding apex can be determined on the basis of the peg-shaped top of the ridge. The kettlehole, northeast of the ridge top supports this interpretation. Ice was probably buried beside the meltwater supply route. The sediment supply to the eastern part of the ridge came partly from the above mentioned apex (log 10) and possibly partly from beneath the glacier (logs 11 and 12, palaeocurrents towards SE/S).

The megaforeset was deposited with well sorted material in an avalanche process. The deposition took place as a steady sliding of material through deposition from sediment gravity flow processes (Martini 1990, p. 291, Hunter *et al.* 1996, p. 1029), and upon the surface angle of repose in the grain-flow process (Ashley *et al.* 1991, p. 117). The sediment for the process was provided by bedload that was transported by a glacial meltwater jet rushing down the foreslope in the zone of detachment of the flow (Powell 1990, p. 67). The upper sedimentation units log 2 (the uppermost 8.5 m, Fig. 34) suggest a hyperconcentrated flow origin. The fact that coarse material is accumulated at unit bottoms suggests a debris fall type process (Nemec 1990, s.44). To summarise, the unit dip increases and material becomes coarser upwards and distal to proximal. This indicates continuity in depositional processes and a close connection between the depositional origins of the bottomset/distal and foreset facies.

In the western part of the main exposure (log 6) massive units of coarse gravel show a low proximal dip. This indicates that they were formed through deposition by an efflux jet rushing from beneath the glacier upon the stoss slope of the ice marginal ridge (Hunter *et al.* 1996, p. 1033).

At Tupsumäki the ice margin oscillated actively against the previously accumulated marginal deposit at different stages during its development. Consequently, a proximal ice marginal lithofacies association was produced. This facies is composed of ice contact material (log 5) with dumped and debris fall material. Also deposits of cohesive and cohesionless debris flow occur that were deposited by glaciofluvial traction deformed due to ice oscillation (logs 4, 6, 11, 13, 14) (Fig. 38). The basal till type lithofacies is common only in the eastern part of the ridge (logs 13 and 14).

The Tupsumäki ridge was deposited along the meltwater route of the Siikala valley (Figs 10, 48) that shifted to this position diagonally from the SSW (Fig. 31). The adjacent Aittoissuonmäki deposit was formed during the first stage of the meltwater route shift. A meltwater efflux jet deposited material that progressively caused the meltwater exit to move higher up. The deposit never reached the highest shoreline level as the ice margin did not remain long enough in stagnant position. The position of the ice margin altered considerably during the deposition. Consequently, push moraine ridges were formed and the proximal part of the deposit was loaded and partly overrun by the readvances of the ice margin.

3.2.4. Hirvilammi, Loppi

The Hirvilammi type site is located in the Inner Second Salpausselkä zone, on the western side of the central valley depression of the Tammela Uplands, i.e. at the western margin of the Esker Area (Figs 5, 9, 10). A fracture zone trending NW-SE is interpreted to locate at the valley margin. The bedrock at the site is gabbro and diorite (Härme 1953) (Fig. 3). The local relief is around 20 metres (Figs 39, 40).

A meltwater route ran through the northeastern part of the area (Figs 48, 49). In addition glaciofluvial deposition took place southeast of test pit 40/87 (Palmu 1990) (Figs 40-42).

At the time the ridge was deposited water depth

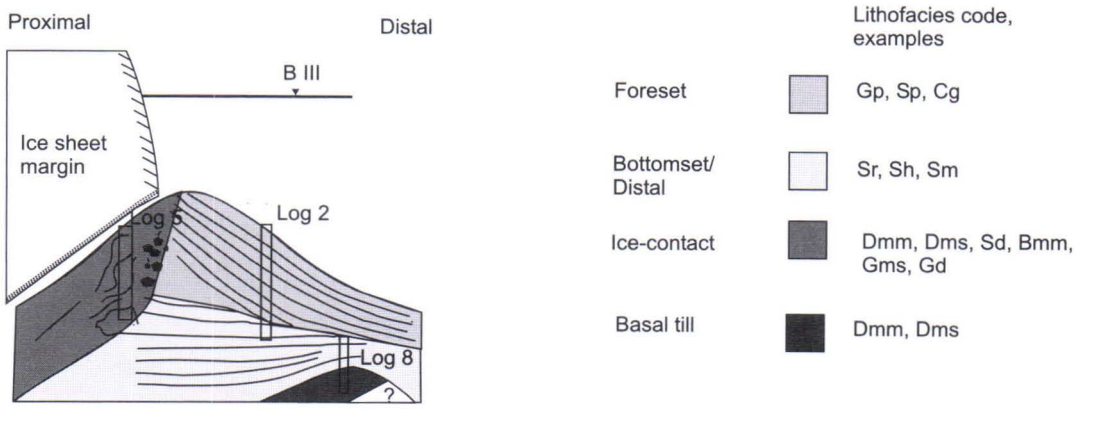

Fig. 39. Schematic drawing showing cross-section of Tupsumäki with main depositional lithofacies and an older diamicton-dominated ice marginal ridge observed in the distal lower part of the deposit.

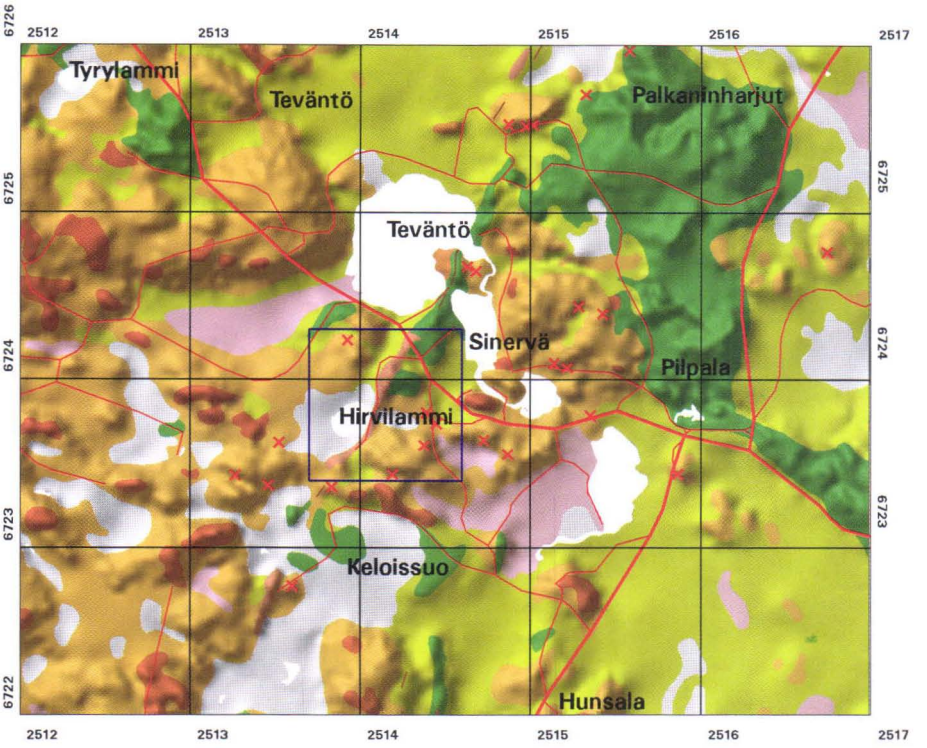

Fig. 40. Hillshaded Quaternary geology map at 1:40, 000 scale of the Hirvilammi area - an altered version of the Geological Survey of Finland Quaternary geology map at 1:20, 000 scale. See Fig.11 for explanations. Inset shows location of the hillshaded 3D map (Fig. 41).
Elevation contour data © National Land Survey of Finland.

at the ice margin, estimated from the valley bottoms (110-115 m a.s.l.), reached ca. 30-35 m, whereas the estimated highest shoreline level was at 145 m a.s.l. Correspondingly, the ridge top was deposited at the water depth of 10 m deep water (Figs 40-42).

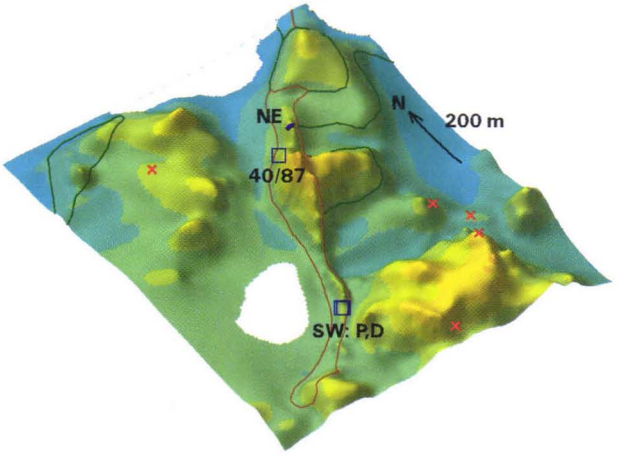

Fig. 41. Colour classified hillshaded relief view of Hirvilammi from the SW (210° with a 45° vertical angle).
Elevation contour data © National Land Survey of Finland.

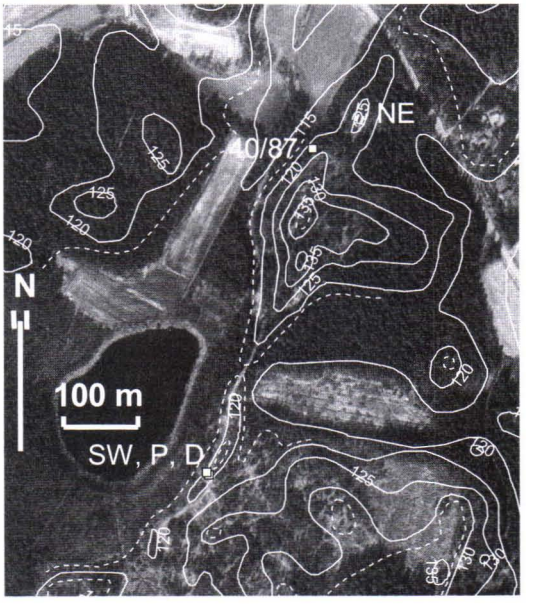

Fig. 42. Aerial photograph at 1:10, 000 scale from Hirvilammi. Log sites and test pits locations are shown.
Aerial photograph © Topographic Service of the Finnish Defence Forces. Elevation contour data © National Land Survey of Finland.

3.2.4.1. Northeastern test pit

An east-west-oriented (85°) test pit, 7.5 m long and c. 3 m deep, was cut across the northeastern ridge top (Figs 40-42), 5 m north of a forest road cutting. Previous studies have shown 4 m of horizontally bedded and ripple sand as well as minor silt in the road cutting, distally to the pit.

Three surface boulders were found at the pit site on the distal slope of the ridge. In the south and distal parts of the pit two boulders were found. Roundness of stones range between classes 2 and 4.

The pit revealed three diamicton and three sand units. The diamicton formed 2/3 of the material on the southern face of the test pit. Proximal and top basal contacts show a low proximal dip with a horizontal or distal slope in their distal parts.

The particle size distributions of the diamicton units were similar. The diamicton is sandy (GEO-classification). The particle size analysis sample was taken from unit I, below unit II sandy subunit. The percentage of stones (60-600 mm in diameter) in the diamicton units I+III was 20% as calculated using the net measurement method (measuring area 3.7-4.7).

The diamicton units were mostly massive and matrix-supported. The lowest unit (I) is heterogeneous with some sand and silt intraclasts. The unit is crossed by a contorted fine sand layer as well. The lower diamicton units have erosional contacts with underlying sorted materials. In the upper diamicton unit the contact is wavy. Loose diamicton occurred on the surface (0.5 m). Elsewhere the density of the diamicton is intermediate or high (classes 3-4). Pebble (6-60 mm) fabric was measured from the lower part of the unit I. Two-dimensional fabric analysis gave a maximum in the direction of 290° (Fig. 43).

The sandy interlayers are deformed, exhibiting faults and folding. The sand units have a high density. The upper sand unit (IV) had originally a horizontally bedded blanketing structure. The silt in the bottom unit is horizontally laminated. The basal contact was clear and conformable (Fig. 43).

A fold within the sand unit II below the middle diamicton unit has been formed against a large

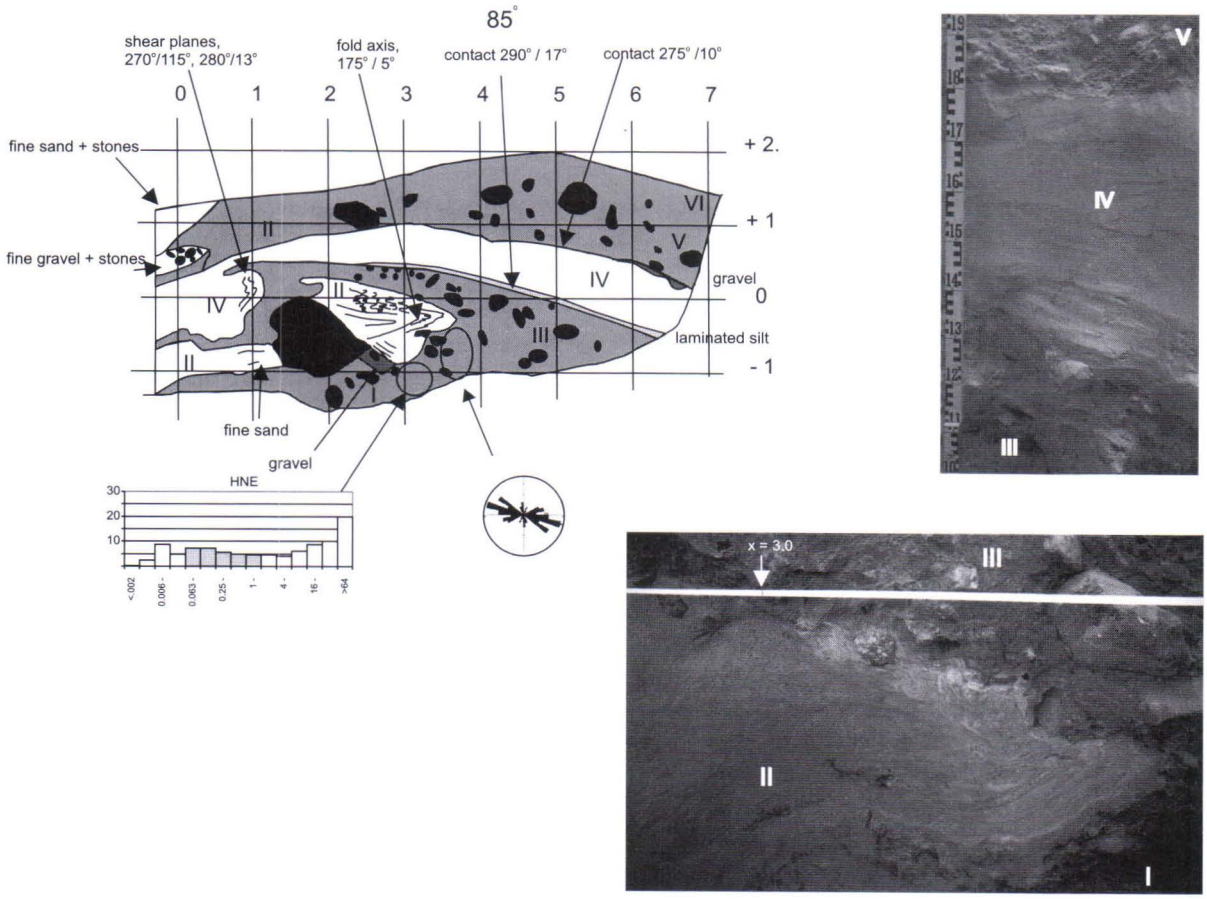

Fig. 43. NE test pit. Drawing shows S-face of the test pit marked with gravel sampling sites and also measurement points for fabric and deformation structures. Upper picture (464) is taken at 2.7-3.8 m along the reference line (RL). The width of the RL-ribbon is 17 mm. Lower picture is taken at 4.0-4.5 m (RL), immediately above the line.

boulder the fold axis of which plunge 5° towards 175°. Shear planes of 280°/13 and 270°/15° were measured from the upper sand unit (IV) at 1.0 m point along the measuring line (from the distal margin) (Fig. 43).

3.2.4.2. Southwestern test pits

In the southwestern part of the investigation site, test pits were dug on the proximal and distal slopes. On the ridge top the test pits were spaced 2 m. The sedimentary units in these test pits were similar and will be discussed below.

On the ridge top no surface boulders were found between the test pits. On the distal slope, 1-5 boulders per 100 m^2 were found. A large number of surface boulders lay scattered on the steep distal slope northeast of the pits. The ridge sediments contain boulders in great numbers; in the proximal pit 3 and in the distal pit 6 boulders were dug up, and 3 boulders were observed on the test pit faces. Especially the unit II had a large number of boulders.

The ridge top was mainly covered by loose diamicton, 0.6 m thick, with a transitional lower contact and a fine sand matrix. The distal slope surface has an abundance of stones and boulders supported by a loose sand and gravel matrix.

Underlying the surface loose diamicton (unit III, sample 1, Fig. 44) with abundant sand and gravel

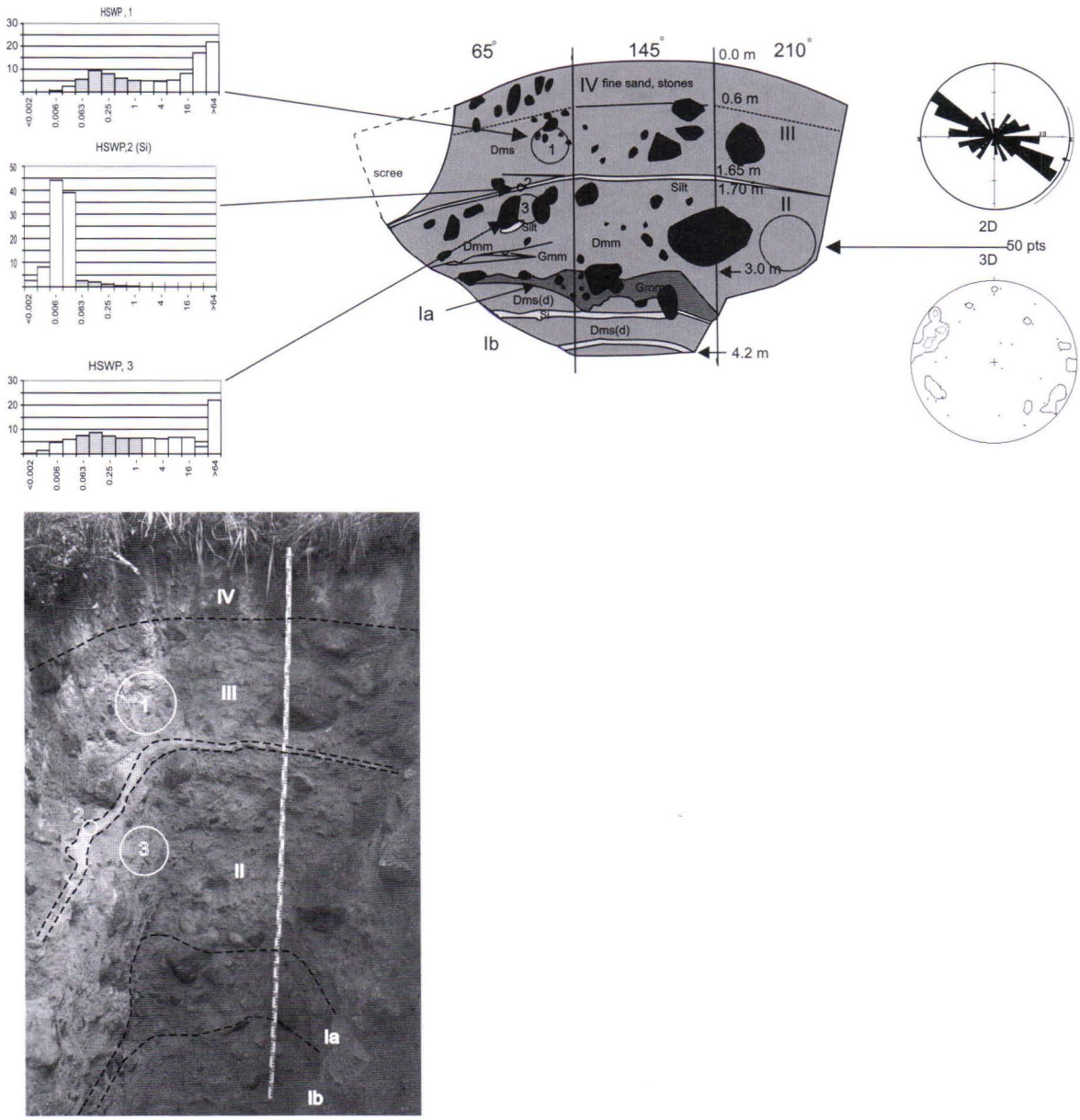

Fig. 44. SW-proximal test pit at Hirvilampi. Maximum height of section is 4.2 m. The boundaries of sedimentary units and sampling sites are also shown.

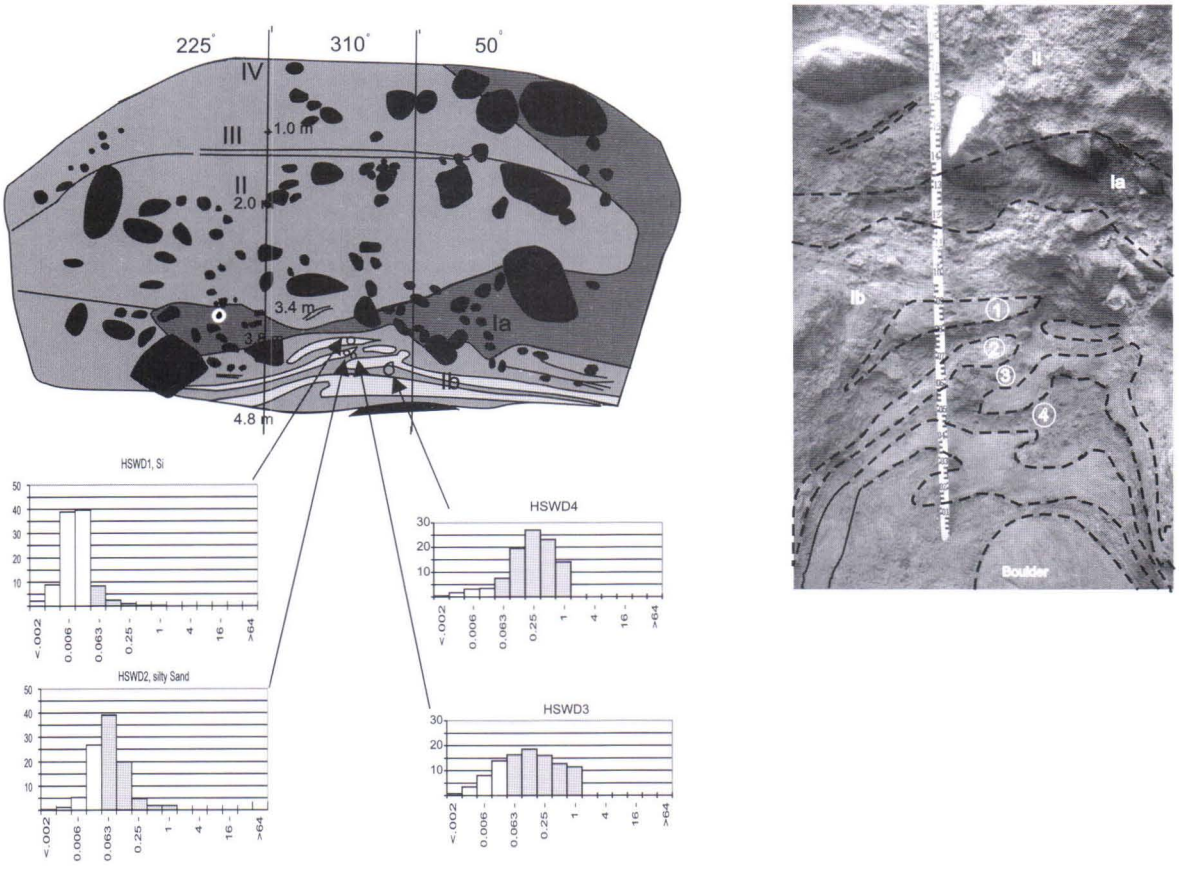

Fig. 45. SW-distal test pit at Hirvilampi. The maximum height of the section 4.8 metres. Picture (484) is taken from the basal part of the test pit. Sampling sites are marked in the picture and the drawing.

stripes and lenses occurred. Five cm of silt (sample 2) (Fig. 44) was deposited at the lower contact of the unit III and covered by a 5-10 cm layer of sand on the distal face of the proximal pit. The basal contact of the unit shows a low proximal dip. On the basis of its particle size the diamicton is classified as a gravelly sandy till (GEO standard). The proximal and distal unit thicknesses are 1.1 and 0.8 m respectively (Figs 44, 45).

Loose, coarse diamicton covers dense massive diamicton (unit II) with proximal and distal thicknesses of 1.9 and 2.1 m respectively. Estimated density falls in classes 3-4. The wavy lower contact of the unit is erosional and sharp in the proximal pit. The material is sandy till (sample 3). The stone percentage of the unit was 22% in the proximal pit. The diamicton is mainly massive. However, the proximal pit shows shear structures, revealed by a few sorted subunits and the fissility of diamicton. The shear planes show a low dip towards the WSW. Two-dimensional fabric analysis gives a mean direction of 300° for the diamicton (Figs 44, 45).

The unit of dense diamicton covers unit I. It is composed of several beds of sorted material and thin beds of diamicton (Figs 44, 45). Unit Ia is gravel dominated and 0.1-1.0 m thick. Unit Ib lies in the basal part of the test pits, with beds of silt and silty sand (2 samples) and beds of diamicton (2 samples) (Fig. 45). In the proximal part of the distal test pit the beds were folded, and in the distal part the bed planes show a low distal dip (Fig. 45).

In addition, a previous observation from a 3 metres deep test pit located on the proximal slope of the ridge is available (Palmu 1990, test pit 40/87). Down to the bottom of the test pit massive, in places feebly laminated, matrix supported diamicton was observed. The loose surface unit (0.6 m) covers material of great density (class 5), sandy diamicton on the basis of its particle size, with an estimated 15% of stones (60-600 mm) and three boulders (>600 mm). Number of surface boulders per 100 m^2 exceeds 10.

3.2.4.3. Interpretation

At the northeastern transverse pit, a series of diamicton beds of basal till type occur on the proximal ridge top indicating multiple oscillation events of the ice margin. Thick, bouldery, matrix-supported diamicton (dmm) covers the surface and forms intermediate beds. The sand units of the transverse pit and the adjacent road cutting have

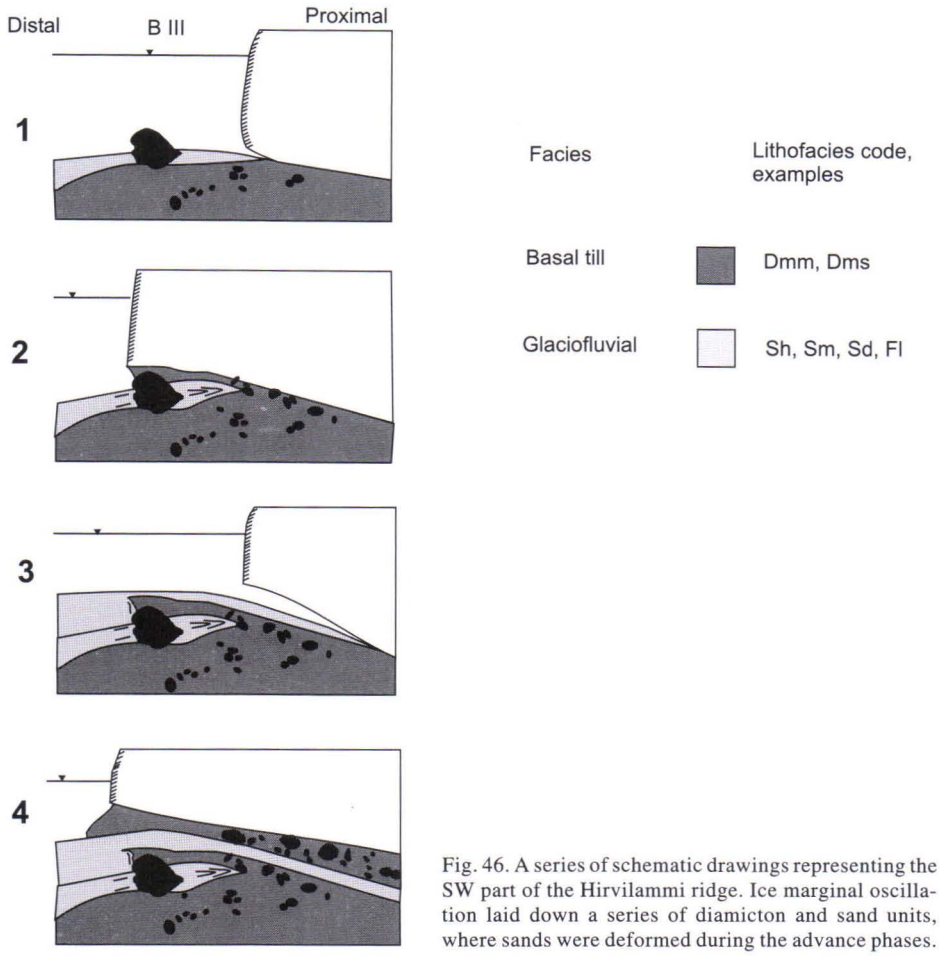

Fig. 46. A series of schematic drawings representing the SW part of the Hirvilammi ridge. Ice marginal oscillation laid down a series of diamicton and sand units, where sands were deformed during the advance phases.

been deposited by a marginal branch of the northeastern meltwater route. Ice oscillation has deformed the sand deposits. Consequently, folding and shear planes occur. The directional data of these structures reveals that oscillation came from the W-WNW.

In the southwestern corner of the site a diamicton-dominated ice marginal ridge was created by pushing. In this area only meagre meltwater supply reached the ice margin. It was this low flow that deposited the intermediate lamellae and lenses of silt and silty sand in the lowest sedimentation unit. The diamictons of this unit are mainly ice marginal sediment gravity flow deposits (cf. Powell & Domack 1995, Dowdeswell & Powell 1996). The southwestern massive dense diamicton unit is a basal till-type depositional unit formed as the ice margin advanced against the marginal deposit (cf. Dreimanis 1989). The overlying diamicton unit is a loose, partly washed sediment released into the ice marginal contact. The surface part of the ridge was also affected by the wave action near shoreline and by percolating water. The matrix of the topmost surface deposit has been interpreted as aeolian cover sand.

4. DISCUSSION

4.1. General

In this section, the ice marginal sedimentation environment of the Second Salpausselkä zone is discussed first. The following points will be focused on: 1) the effect of topography, 2) the volume and types of sediment transported into and deposited in the ice marginal environment, 3) sedimentology of the study area, and 4) morphology of ice marginal deposits. Secondly, the deglaciation of southwestern Finland in general and the study area in particular will be discussed.

Ice marginal deposits described and discussed here have been variously termed by others as fol-

Table 2. Factors controlling the ice marginal depositional environment in the study area, underlined factors are environmental variables that can be interpreted from the sedimentary record of marginal deposits, according to Ashley el al. (1991) (based on Powell 1990, Powell & Domack 1995, Fyfe 1990, Ashley *et al.* 1991, Benn & Evans 1998).

- ice flow type, compressive or extensional

- calving, dependent on: - water depth (sedimentation, dependent on sediment yield, ice marginal deposit system volume and sediment dispersal patterns)
  - thinning/thickening of ice margin (profile)
  - ice flow velocity (bed condition, ice regimen, drawdown)
  - restraining marginal deposit
  - climatology (temperature and precipitation, seasonal variability)
  - crevassing

- meltwater flow regime
  - discharge of meltwater
  - continuity of flow, annual and other cyclicity
  - depth of meltwater conduit mouth
  - meltwater jet dynamics, including trajectory relative to ice marginal deposit
  -> quantity and grain size of sediment supply

- sediment transport processes I, meltwater related
  - coarse = bedload, traction
  - fine = suspension

- sediment transport processes II, connected with diamictons transported to the ice margin
  - rate of debris supply
  - period of time for the ice margin remained in the same position
  - amount of push
  - squeezing

- ice margin activity, dependent on flow velocity and ablation/calving
  - active
    a) advancing = rate of frontal movement, readvance
    -> oscillation, push, squeeze -annual oscillation separately
    b) stabilized, flow velocity and ablation/calving balanced
    c) retreating, ablation/calving greater than flow velocity
  - passive, rate of movement small or stagnant, non-moving

- mechanisms of sedimentation
  - deposition from discharge while in contact with floor of water body
    - sediment gravity flows: (also sediment redistribution by sediment gravity flows)
      - high-density gravity flows (cohesive and cohensionless debris flows, grain flow)
      - low-density gravity flows (density underflows, turbidity currents)
        (hyperpycnal underflow straight from melt-water in glacial lakes)
    - suspension settling
    - subglacial deposition (basal till by lodgement, melt-out or deformation)

lows: for glaciofluvial, i.e. sand and gravel dominated deposits, 'transverse eskers' (Glückert 1975), 'subaqueous outwash fans' (Rust & Romanelli 1975), 'subaqueous esker deltas' (Thomas 1984), and 'glacier contact fans' (Boulton 1986) have been used. For diamicton-dominated deposits such terms as push-moraine, De Geer, and cross-valley moraines are found.

4.2. The effect of topography

The topography beneath the ice margin is the most significant factor that influences the ice marginal and glaciolacustrine sedimentary environment. Topographical factors constituting the effect are 1) water depth at the ice margin during deglaciation, 2) location with respect to depressions that trend NW-SE (meltwater stream systems, basins and hills), and 3) the effect of slope direction. Topography influenced ice marginal processes and sedimentary environment also through ice marginal calving and oscillation (Table 3).

4.2.1. Deglacial ice marginal water depth

A strong dependence exists between the type of sediment and deglacial ice marginal water depth.

Deposits dominated by glaciofluvial sand and gravel were usually deposited in over 40 m deep water. (Note that the water depth map is based on present-day ground surface data which means that the water depths given here are minimum depths. Compare with the diamicton-dominated ice marginal deposits below.) This was due to meltwater feeding routes concentrated along valleys and depressions. Only few glaciofluvial deposits, e.g. southwest of Lake Pyhäjärvi and east of Siikala, were deposited on higher ground, apparently independently from topography.

Where large scale ice marginal deposits were formed bedrock surface is at a considerable depth, in places exceeding 50 m, suggesting a relatively deep water origin. In contrast, at Toivike and Launonen, bedrock outcrops are found in proximal parts of deltas formed at the highest shoreline level, suggesting a shallow water origin. These sites are the only sites in the study area that show so-called 'shoal type' characteristics (Sauramo 1923, Crossen 1991, see also Chapter 2, p. 19).

On higher ground in the Second Salpausselkä study area, the most common ice marginal deposits are diamicton-dominated. Conversely, most of the material transported by meltwater streams was deposited along local valleys, between hills. The orientation of the low areas and the relative water depth (> 50 m) therefore controlled deposition along depressions. It is thought that shallow water depth in the elevated areas also further contributed to this.

In Maine, U.S.A., it has been observed that proximal thicknesses of ice marginal deposits are considerably smaller than distal ones (60-80 m)(Crossen 1991, p. 133). In Finland a similar phenomenon is observed in the Salpausselkä zone e.g. at Hyvinkää, the Erkylänlukot, Kouvola, Selänpää in Valkeala, and at Immolankangas (Vesinoronkangas), Imatra. In the study area no such case is known where proximal bedrock surface would lie higher than the distal one in a glaciofluvial ice marginal deposit. This is explained by the relatively narrow nature of the deposits and the effect of topography.

In the Finnish Lake District Ice Lobe, Fogelberg (1970) states that the location of deposits was not dependent on topography but that it did determine their nature in the Vääksy-Vierumäki area of the Second Salpausselkä. Sandur deltas were deposited in areas of shallow water depth with fractured bedrock topography from where the ice margin retreated at a later stage. In deeper water, only ice marginal ridges were formed due to glacial retreat at an earlier stage (Fogelberg 1970, p. 82). Probably, the intensity of calving controlled the rate of glacial retreat, or thin ice in the depression allowed a faster ice marginal retreat.

In the study area the diamicton-dominated ice marginal deposits were formed in 0-40 m deep water, most usually at 20-30 m, which is less than in glaciofluvial sites. This contrasts sharply with the results of Crossen (1991) from Maine and Okko (1957) from Jylisjärvi.

Boulton's (1986) observations on present-day diamicton-dominated ice marginal deposits in glaciomarine environment on Baffin Island show that these deposits form at water depth of 10-70 m, which parallels the results from the Karkkila-Loppi area. Also Liestol (1976) has observed in Spitsbergen that deposition takes place at the water depth of 20-60 m. Barnett and Holdsworth (1974) report in their studies from Baffin Island that ice ramps build up in water depths exceeding ca 40 m in a glaciolacustrine environment.

As more elevated areas lack fine sediment cover, ridges are well exposed for mapping. However, diamicton-dominated ice marginal ridges might have been deposited in deeper water as well, but are now buried under younger sediments. Consequently, observations on the relationship between the location of diamicton dominated ice marginal ridges and water depth at the time of deposition may give biased results. As it is known, a higher ridge (most likely glaciofluvial) is probably visible through the thickest bed of fine sediment deposits in the study area. In contrast, a low (diamicton dominated ice marginal) ridge may remain completely buried . However, on the basis of this study it seems safe to conclude that the observed abundance of diamicton-dominated ice marginal deposits particularly in higher areas is real rather than biased data.

Results from studies carried out by Heikkinen (1985) indicate that the high fractured bedrock topography of the eastern part of the Tiirismaa (Sairakkala) area and north of it governed the position of the Salpausselkä zone. Tiirismaa and Lappeenranta constitute the western and eastern margins of his study area, respectively. On the basis of trend-surface analysis it seems that bedrock topography did not exert any large scale control on the position of the Salpausselkä zone deposits in the area of the Finnish Lake District Ice Lobe.

4.2.2. Basins, valleys, higher ground and hills

The combined effect of basins and shallow water was observed by Sauramo (1923, p. 152-153) with examples from the First Salpausselkä between Hanko and Lahti, south of Lake Saimaa, and from Jaamankangas and Selkäkangas in Northern Karelia.

In deep and wide basins, parallel to the ice margin, ice marginal deposits are few and not as developed as on basin flanks or higher areas. It appears that the retreat of the ice margin was continuous and that in more extensive valley areas frontal positions were quickly shifting through rapid calving to find stability only in stoss and lee slopes and more elevated positions in general.

Virkkala (1963) and Glückert (1975) suggested that glaciofluvial ice marginal deposits would accumulate on the distal, south and east flanks of basins. In the present area such developments are not common and were observed only east of Huhtimo, at Lake Erävisjärvi, southeast of Lake Oravisjärvi and north of Läyliäinen. All of these are glaciofluvial deposits found along meltwater routes and formed between higher areas that consequently controlled the position of the ice margin. Okko (1957, p. 24) states also that glaciofluvial ice marginal deposition is independent of topography underlying ice margin in the Lake Jylisjärvi area, i.e. no (glaciofluvial) sediments are found south or southeast of depressions.

The dry land ice marginal retreat is delayed in basins in extensive supra-aquatic areas (Sauramo 1923, p. 152). The glacier retreats or melts *in situ*, last in valleys. A similar situation is found northwest of the study area in the Tammela Upland after the Baltic Ice Lake dropped to the Yoldia Sea level.

4.2.3. The effect of underlying slope orientations at the ice margin

Most diamicton-dominated ice marginal deposits in the study area were laid down on the (northwest) stoss-side slopes of hill areas, as e.g. in the Huhtimo and Tuorila area of Karkkila, and in the Sajantilankulma of Loppi, or on higher ground and on hilltops, as e.g. in the area between the Lakes Ojajärvi and Kesijärvi, between Lake Loppijärvi and Jokiniemi, and in Karkkila, east and northeast of Siikala.

Lee-side ice marginal ridges dominated by diamicton are scarce; e.g. north of Läyliäinen, and northeast of Lake Loppijärvi, at Pälsi. By contrast, glaciofluvial sand and gravel were deposited, e.g. as the relatively small-scale ridges located at Järventausta in Lake Loppijärvi and south of the Koli hill in Karkkila.

According to Virkkala (1963) the Second Salpausselkä diamicton-dominated ice marginal deposits were formed relatively independently of local topography. On the basis of his studies (1963, p. 46), 42% of deposits were formed on flat ground surface, 26% on stoss-side and 32% on lee-side slopes. In contrast, 39% of glaciofluvial ice marginal deposits (39 deposits in all) were formed on stoss-side slopes, 31.5% on flat ground surface, and 10.5% on lee-side slopes. The role of basins is particularly emphasised by Virkkala (1963). The study area of Virkkala was more extensive than the field area in this study and included areas between Bromarv and Lake Jylisjärvi where water depth at the ice margin was considerably greater in places than that in the Karkkila-Loppi area.

Table 3. The effect of topography and water depth on the location and sedimentology of ice marginal deposits.

Topographic factor	Glaciofluvial sand and gravel dominated deposits	Diamicton-dominated deposits
Depth of water during deglaciation	Usually in the beginning more than 40 m	Often less than 20 m
Position in relation to NW-SE -valleys (meltwater routes)	Mostly related to NW-SE-valleys	
Position in relation to basin areas	Partly situated on distal sides of basins	
Effect of slope azimuth	Often on lee-side slopes	Stoss-side slopes and hilltop positions

4.3. Sediment supply

Debris is transported to ice margins either 1) as well-sorted glaciofluvial material, i.e. sands and gravels (also silts and clays) through meltwater supply or 2) as poorly sorted basal till and diamictons by pushing and squeezing from beneath the ice margin. The intensity of meltwater supply deter-

mines the intensity of glaciofluvial ice marginal deposition (volumes and thicknesses of sedimentation units). Basal till is deposited on the proximal slopes of ice marginal ridges subglacially. In addition, diamicton is formed by squeezing out of subglacial sediment, pushed forward and finally deposited beyond the ice margin (Boulton 1986, Powell 1991, p. 86). The intensity of the diamicton supply to the ice margin is determined mainly by ice marginal behaviour including standstills or restricted oscillations. (Table 4.)

4.3.1. Feeding by meltwater

Analysis of meltwater routes holds a key to understanding the origin of glaciofluvial ice marginal deposition (also Aartolahti 1968, p. 65). Meltwater routes sought their way into depressions trending NW-SE or along their flanks. These meltwater routes were driven by hydrostatic pressure gradient (Shreve 1972, 1985, Fogelberg 1970, p. 74, 75). The fact that the Launonen esker was deposited at the margin of a higher area was probably due to differential glacier velocities in the esker and the depression (Figs 2, 8, 10, 48). Part of meltwater flow most likely run along shear zones.

When a subglacial meltwater efflux jet reaches the ice margin in a water body, sublacustrine or submarine fan deposits are formed first (Powell 1990). These grade into embryonic deltas parallel to the ice margin. It seems however, that due to ice marginal oscillation most of the fans will be destroyed. In some cases embryonic deltas can be preserved, e.g. in the case of restricted ice oscillation.

Ideally, a subglacial ice marginal meltwater supply process develops in a series of four stages, based on the model of Gustavson and Boothroyd (1982, 1987) starting with: 1) the tunnel stage, followed by 2) the outwash fan and bottom underflow stage. Subsequently, in 3) the delta stage, material is deposited upon the proximal part of the delta as fountain type discharge. This probably led to the emergence of the delta top above the water level marking the beginning of 4) the sandur stage (Donner 1976).

Two of these stages characterise the deposition of sorted material along meltwater routes in the present area, namely the tunnel and delta stages. Tunnel stage deposits are well preserved in the Esker Area, including the Läyliäinen (Maakylä-Porras) esker and the Pilpala esker that starts at Vuonteenmäki. Tunnel stage deposits appear as ridges in the hillshaded relief map of the eskers. They are lined by continuous, parallel series of kettleholes that are located centrally to the esker systems (Kujansuu 1967, Aartolahti 1968, Vähäsarja 1971, Banerjee & McDonald 1975, Glückert 1978, Niemelä 1979, Warren & Ashley 1994). Outside the Esker Area, tunnel stage deposits are poorly preserved, except where only limited oscillation took place. The most significant of such deposits occur between the Outer and Central zone at Ahmoo, east of central Karkkila, and at Nyytäistenkulma, east of Lake Loppijärvi (Figs 5, 10, 48).

Deltaic deposits developed at the ice margin are mainly restricted to the Esker Area, where they are identified mainly on morphological basis (Figs 11, 47) using data on ice marginal proximal contacts, associated kettleholes, and steep distal slopes (Kujansuu 1967, Aartolahti 1968, Vähäsarja 1971, Banerjee & McDonald 1975, Glückert 1978, Niemelä 1979, Warren & Ashley 1994). Bedrock fracture interpretation records were also used (Fig. 4). Extensive areas of bedrock valleys, pitted with kettleholes complicate the interpretation.

A delta (Hyrrönharju) with two main ice marginal stages was observed in the Läyliäinen esker. It was deposited contemporaneously with the Vuonteenmäki delta and the rest of the Second Salpausselkä Outer zone. Northwest of it a narrow part in the esker system reflects possibly a bedrock fracture zone. To the NW, the steep southwestern slope is believed to have formed at the contact of the ice margin. The ice margin must have retreated faster in the Esker area than where the present-day Isosuo mire is. The esker system is particularly wide northeast of the Isosuo mire. This area is interpreted as being of the same age as the Second Salpausselkä Central zone. In addition to the primary meltwater route from NNW, some meltwater was possibly fed to the western part of the esker as well.

Drillings and seismic soundings from the GSF Drill Core Depot (GTK:n Kairasydänarkisto) in the Läyliäinen esker show a over 20 m thick fine sand deposit (Haavisto-Hyvärinen *et al.* 1990d). Field observations confirm that the interpretation is valid, i.e. that the Depot location is situated at the margin of an esker delta (Figs 5, 47).

Meltwater during the tunnel stage was initially fed from NNW to the esker system that starts at Vuonteenmäki in the Outer zone. When the ice margin retreated to the Central zone, the supply direction diverted and came from WNW, from the Hunsala valley area. Again, the meltwater supply in the Inner zone came almost directly from the north (Fig. 47).

The volume of glaciofluvial ice marginal deposits in the study area is great. In Vuonteenmäki for example total deposit thickness exceeds 50 m in many places. Thicknesses of supra-aquatic fans or deltas were actually restricted by basin depths

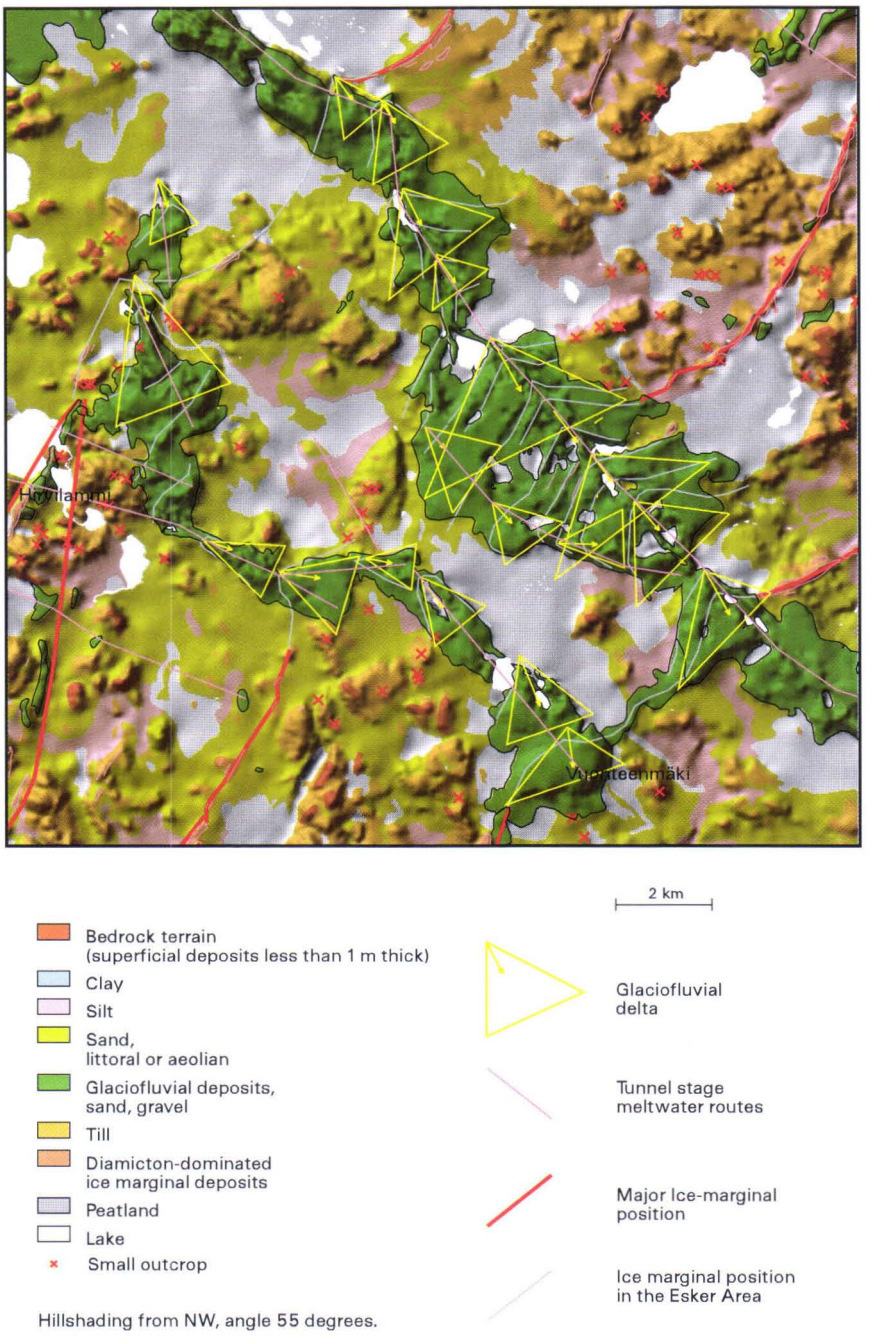

Fig. 47. Map of the Esker Area. Central sector of the Second Salpausselkä zone in the Karkkila-Loppi area is dominated by esker systems. They consist of ice tunnel stage deposits distinguished within ice marginal deltas by linear kettlehole systems. Ice contact positions interpreted from broad arrow-shaped kettlehole systems of these deltas reveal successive ice marginal stages.
Elevation contour data © National Land Survey of Finland.

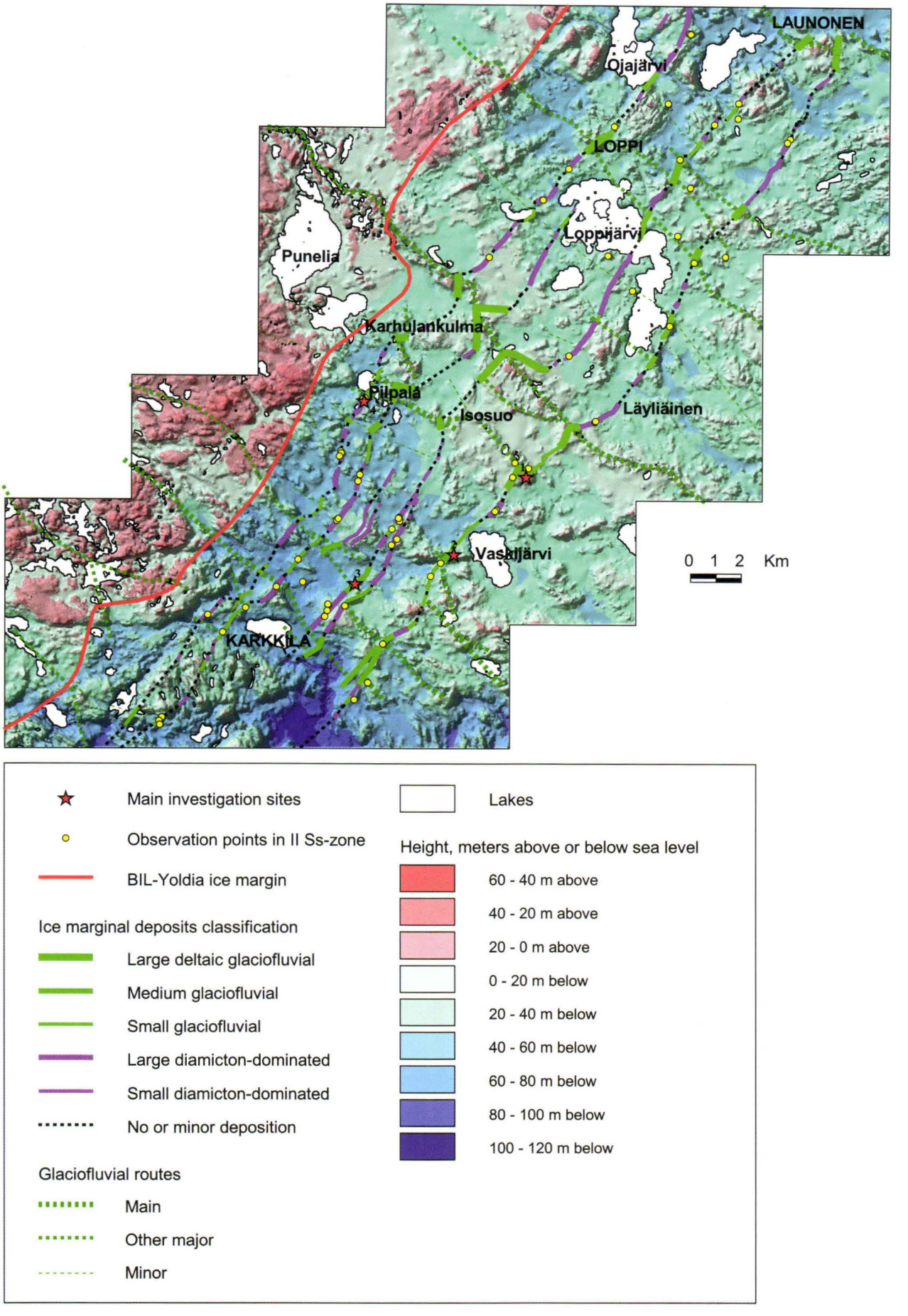

Fig. 48. Classification of the ice marginal zones of the Second Salpausselkä in the study area. Main ice marginal zones are classified into six classes. Deglacial water depth is colour coded in the map. Meltwater routes fall into three classes. Main study sites are shown with an asterisk. Elevation contour data © National Land Survey of Finland.

only. In supra-aquatic environment, however, thickness depends on the local base level (Boulton 1986).

Bottom underflow occurred as coarse sediments discharged by a powerful meltwater jet into the lake water at the ice margin were transported beyond the tunnel exit as bedload and controlled by meltwater flow velocity. The relatively short range bedload transport does not generally deliver coarse sediment far from the ice margin (Boulton 1986). Thus it is usually deposited within the detachment zone (Powell 1990, p. 67, Hunter *et al.* 1996, p. 1024). This kind of flow structure was possibly responsible for e.g. the deposition of the Tupsumäki megaforeset (Figs 34, 39). When ice marginal deposits grow into a delta and up to the water level, sandur stage begins and bringing about fluvial water flow conditions, which may result in coarse material being transported further. Consequently, in delta topset and especially in sandur conditions, coarse material is transported longer distances; this way, the material of the glaciofluvial ice marginal deposit in general becomes coarser upwards although flow velocity may not have increased at all.

Consequently, the coarsening upwards and laterally (fining away from the ice margin)indicate shallowing water conditions, and stability of the adjacent ice contact front. Another hypothesis is that the upwards coarsening of the material reflects a considerable increase in ice marginal melt water supply. Relatively low flow may have been dominating throughout the period of ice marginal position. Mainly fine and medium sands would have been deposited in such conditions. The great increase in meltwater supply would be associated with resumed glacial retreat, when the ice marginal position is abandoned owing to a climatic or glaciodynamic shifts.

Glaciofluvial deltas hold large amount of material which raises the question of the time required for their accumulation. According to Powell (1991, p. 87), the deposition of an ice-contact delta takes several decades. Powell and Molnia (1989) have reported that deposition of submarine outwash amounts to annually ca 5 m of sediments, and that fan deltas grow by ca 1 million m^3 of sediment per year (1 x 10^6 m^3/yr). According to Powell (1990) the depositional rate may be as high as 32 m within four years. On the basis of this data, the Vuonteenmäki delta may have been deposited in approximately 10 years with c. one million m^3 of annual deposits of fine sand, sand and gravel.

Studies in Alaska show that meltwater transport is responsible for the major part, i.e. 83.6-98.6%, of the material that was transported to water terminating ice margins. Large volume of deglacial meltwater were produced in the Esker Area, where an estimated 70-80% of the glaciofluvial material was deposited. In the Läyliäinen esker system there was a stable meltwater flow along the main meltwater route during the tunnel stage. Delta stages exhibit cyclicity corresponding to that of the Second Salpausselkä zones (Figs 47, 48). Earlier the Salpausselkä zone was thought to have been deposited during the Alleröd event because larger amounts of meltwater were discharged at that time (Okko 1957, p. 7, 41, M.Okko 1962, p. 139, 153, Fogelberg 1970, p. 83).

The Läyliäinen esker system is the only one that runs continuously through the Second Salpausselkä zone. Its southeastern end is at Maakylä while its northwest end extends far beyond Forssa. In the study area, extensive deltas are found only in the Esker Area. The Launonen esker continues through the entire zone as well, but at the Outer zone the meltwater route exhibits a lateral shift (Fig. 2).

Discontinuous meltwater route deposits exist e.g. on both sides of Lake Loppijärvi, between the parish of Loppi and Launonen, and around the centre of Karkkila and Siikala areas. In these areas separate ice marginal glaciofluvial deposits exist. It appears then that completely different meltwater routes were feeding the Outer zone, the most important of which terminated at Vuonteenmäki. This, esker system of Pilpala , runs continuously throughout the Second Salpausselkä zone. Some meltwater routes also terminate at the Outer subzone, e.g. east of Koli in Karkkila. Most routes terminate at roughly 10 km northwest of the Second Salpausselkä Inner subzone (Figs 2, 10, 47).

Except for the Esker Area, the Second Salpausselkä glaciofluvial ice marginal deposits are embryonic deltas by their basic structure (see also Glückert 1975, 1981). They were generated under quasi-stable conditions or conditions of ice marginal advance and follow a developmental pattern where deposition of grounding line fans is followed by that of ice-contact deltas (see also Powell 1991, p. 90). A limited proportion of the embryonic deltas were deposited at the contemporary water level or in its close proximity, e.g. Multamäki, Kotamäki and Nummimäki (Figs 5, 10, 48). For the most part, these deltas never emerged above water level due to lack of sediment or time, e.g Tupsumäki and Aittoissuonmäki in Karkkila, and Erävisharju and Soukki in Loppi.

Meltwater routes that brought material to a deposition centre broke up and shifted laterally several times during the deposition. This was caused by cyclic, and most probably annual nature of sediment input and also by glaciotectonism. As a result esker ridges and related deposits are discontinuous

and laterally dispersed. Direction of meltwater supply shifted laterally even during a melt season, i.e. when the ice marginal deposit was growing (see also Boulton 1986, p. 696).

What would happen if the ice margin uplifted periodically creating sheet flow (Powell 1991, p. 89)? Glaciofluvial ice marginal ridges of De Geer-type would be created (Beaudry & Prichonnet 1991). This type of deposits are not known to exist in the study area. This indicates that surge-type glacial advance or other factor causing rapid meltwater pressure rise did not influence the ice marginal zone.

When the ice margin occupies a certain position, lateral flow may develop between ice marginal ridges. Fogelberg (1970, p. 36-38) mentions such circumstances east of Vääksy. Except for Multamäki, deposition due to meltwater flow parallel to the ice margin has not been found in the study area.

The material in the glaciofluvial ice marginal deposits is not as rounded as in esker systems (also Virkkala 1963, p. 42). The material of the Salpausselkä ice marginal deposits has also been transported longer distances than basal till (Salonen & Glückert 1992).

Table 4. Sediment input into the ice marginal zone.

Sediment input environment	Location (typical)	Typical water depth, m	Typical depositional elements	Dominating structural units (LFAs)	Typical lithofacies
Meltwater Esker area, Läyliäinen and Pilpala esker systems	Low-lying area in central study area	In the beginning often over 40, in the end 0	Delta Ridge deposited in a tunnel	Foresets Bottomsets Topsets Tunnel-stage deposits (=Tsd)	Gp, Sp Gmm, Gms Sh, Sr Gmm, Gcm Gms, Gcm
Meltwater Other important melt water routes	Long NW-SE-valley	In the beginning often over 40	Partly developed delta Fan	Foresets, Bottomsets Proximal complex (Tsd)	Gp, Sp Gmm, Gms Sh, Sr Sc, Gc
Meltwater Separate	Lee (distal) side of a hill	0-20-40	Partly developed delta Fan	Foresets Bottomsets Proximal complex	Sp, Gp, Sh Gmm, Gms
Push, separate oscillation (winter)	Stoss (proximal) side of a hill and hilltop	0 - 20	Small marginal ridge (De Geer moraine), dominated by diamictons	Diamicton (basal till), proximal deformational unit A wedge in the lower, distal part: Mass movement diamictons + glaciofluvial sand, gravel and silt	Dmm, Dms Sm, Sh Gmm, Gms
Push, multiple oscillations	Stoss	0 - 20 (proximal) of a hill and hilltop	Large marginal ridge dominated by diamictons	Diamicton units (basal till) in the proximal part one on top another Bottom parts and intermixed: as above in the distal wedge: g.f. sediments and massmovement diamicton	Dmm, Dms Gmm, Gms Gp, Sp, Sh

In deposits with a common source of sediment, thick varved clay sediments are thought to be deposited in association with the meltwater discharge peak associated with the deposition of a large scale ice marginal deposit (Kujansuu 1995). Thick varved clay would therefore be expected to be found in the Salpausselkä zone (Lundqvist 1987). Sauramo (1923, p. 121-122) notes that an abundance of clays is found in the Salpausselkä zone and explains this to be due to the great number of varves deposited in the same area and to the thick horizons, in which coagulation has attracted large amounts of finest clay.

Varves deposited at the time of deposition of the First Salpausselkä were diatactic, thin and relatively coarse, composed of silty clay and clay (Sauramo 1923, p. 86). During faster ice marginal retreat between the First and Second Salpausselkä varve sediment were still relatively coarse (Sauramo 1923, p. 86). At the time of deposition of the Second Salpausselkä varves were thicker, finer and symmictic.

After the Second Salpausselkä was deposited thick varves continued to be formed e.g. at Vihti (Sauramo 1923, p. 121). As known, varve thickness is inversely proportional to the distance of the locus of sedimentation from the ice margin (De Geer 1940, Sauramo 1923, Ignatius 1958, Niemelä 1971, Drewry 1986). Thus it can be concluded that great amounts of fine sediments were released from beneath the glacier after the glacial retreat had begun from the Second Salpausselkä zone. Unless so it would be difficult to explain the great thickness of varved clay deposits around Vihti as the ice margin stood relatively far away. This indicates that abundant ice marginal meltwater was released. The fast deposition of varved sediments may be associated with the deposition of the Third Salpausselkä analogous to the intensive varved sediment deposition associated with the Middle Finland ice marginal deposits (Kujansuu 1995).

Erosional and depositional processes of glacial clays and glaciofluvial material may be partly independent processes during deglaciation. Glacial clay may partly originate from separate glacier bed sources. Clay fraction may have been released from ancient valley areas where pre-Weichselian sediments may have occurred like e.g. the Somero-Loimaa region (west of the Tammela Upland)(see also Gardemeister 1968, Donner & Gardemeister 1971).

4.3.2. Diamicton supply

There are three principal processes of ice marginal diamicton supply of which the formation of basal lodgement till against glacier bed is the most important as interpreted on the basis of the field observations in this study. A second process is by subglacial squeeze of debris to the ice margin (Powell & Domack 1995, Dowdeswell & Powell 1996). Other processes include deposition by glacier push, frontal-dump meltout and calve-dumping (Powell & Domack 1995). Among the investigation sites of this study, the Hirvilammi ridge deposits are dominated by basal till and squeeze-type diamicton (with debris flow origin added). In any case, the accumulation of such diamicton requires some form of frontal oscillation.

The material of smaller-scale diamicton-dominated ice marginal deposits consists mainly of basal till-type diamicton. Occasionally diamictons of debris flow origin occur in distal parts of deposits mixed with sorted sediments.

Proximal parts of larger-scale diamicton dominated ice marginal deposits are composed mainly of basal till-type diamicton that was deposited against glacier bed. The central and distal parts of these deposits are mostly dominated by diamictons of cohesive debris flow origin. These were released through squeezing from beneath the glacier margin, or in some cases even by frontal dump meltout. In addition, meltwater fed sorted material occurs in greater proportions when the size of the accumulation becomes larger.

Small diamicton-dominated ice marginal ridges (De Geer moraines) usually occur in swarms. They are thought to have been deposited during winter advances as annual push moraines (Boulton 1986, p. 681).

An ice marginal sector may lack the diamicton-dominated ice marginal ridge in the following cases: 1) younger sediments have buried the ridge, 2) there was no oscillation, 3) there was no subglacial debris, therefore oscillation carried no material (when the glacier rests on bedrock or glaciofluvial transport carries all material), or 4) ice marginal material was deformed and ridges were smoothed out.

Diamicton units do occur as well in deposits composed primarily of meltwater fed sediments. Diamictons in proximal parts are often composed of basal till that is deposited against glacier bed.

Transport within the ice mass (englacial and subglacial) carries only 1-10% of total glacial debris (Hunter *et al.* 1996). Subglacial transport by a deforming bed carries two to eight orders of magnitude more material than englacial transport (Hunter *et al.* 1996).

Studies on present-day conditions (Barnett & Holdsworth 1974) show that glacial transport requires five years to create a 3 m high ridge with a 35 m wide base. Boulton (1986, p. 682) has also observed that ridges with a maximum height and

width of 10 and 30 m respectively were deposited in front of the Aavatsmarkbreen ice margin in the Spitsbergen and ridges of 1-10 m in height, with a mean of 4 m and basal width of 2-20 m were produced in front of the 'Alpha glacier' of Baffin Island (Boulton 1986, p. 684).

Particle size distribution in some of the diamicton-dominated ice marginal ridges resembles that of the basal till of the study area. Where an abundance of sorted bed material was fed into the marginal areas of a ridge, the ridge would be composed of more sorted material. On the other hand, the character of the deposit depends on the amount of diamicton deposited as basal till on the proximal slope of the bed or released into the ice marginal contact.

Distal parts of ice marginal deposits dominated by diamicton represent an intermediate form between basal till and different types of glaciofluvial material. Material released at the ice margin is more sorted and contains more washed sediment with less fines than the basal till of the area (Palmu 1990a, b). Intensity of meltwater flow determines the degree of washing of distally deposited diamicton. An abundance of lag boulders is a special characteristic of the deposits (see also Okko 1957, Glückert 1975). Compared to glaciofluvial ice marginal deposits the fine sand fraction is more abundant. Similar observations exist from Jylisjärvi as well (Okko 1957).

4.4. Ice margin behaviour

4.4.1. Nature of ice marginal flow

Compressive flow in the ice marginal zone into shallow water is obviously a significant factor in the deposition of diamicton-dominated ice marginal deposits. This is especially the case in stoss-side topographic positions, as well as in the deposition of diamictons in proximal parts of glaciofluvial, sand and gravel dominated ice marginal deposits. In deeper water, however, the ice flow of a calving glacier margin is extensional (Powell 1991, p. 87) and nearly horizontal (Powell & Domack 1995, p. 454).

If not hampered by any ice marginal obstacle, the frontal oscillation of an ice margin produces a ridge with a low-angle proximal slope. An advance declines on the upslope direction facilitating the development of a diamicton-dominated ice marginal ridge at the terminus (Lundqvist 1987). The ridge is formed on the proximal side of the obstacle or on top of it. An intensified squeeze flow brings more debris higher up in the ice mass. Consequently more debris is released at the ice margin and the ridge will grow larger than in conditions of weak compressive flow.

4.4.2. Straightness of ice margin

The ice marginal positions in the study area indicate relatively straight ice margins. This is thought to be due to a steady state of the ice margin, which means that calving and ablation were in balance with the amount of ice reaching the ice margin.

There are however plenty of exceptions to the steady state situation. At meltwater routes the retreat of the ice margin was often more advanced and the ice margin was curved. Projections indicating oscillation of lobes are also common. Small-scale lobes of diamicton-dominated ice marginal ridges occur e.g. south of Launonen where the orientation of the ridge shows en echelon type obliqueness (Fig. 48).

The ice margin seems to have retreated with an almost straight line on both sides of the Huhtimo hill as glaciofluvial deep water deposition passes into the push moraine type of deposition in higher areas (Fig. 31).

4.4.3. Calving

Calving takes place at the ice margins that terminate in water when extensional stress generated by the glacier flow overcomes the strength of the ice. Glacial advance is counteracted by hydrostatic pressure which consequently decreases towards shallower waters. This causes the ice margin to collapse especially near the water level (Benn & Evans 1998, p. 277). The behaviour of the ice margin is largely dependent on the size of the calving ice blocks (Benn & Evans 1998, p. 279). On the basis of sedimentological observations it is difficult to establish whether calving has taken place or not in the area, but if it did, the fact that hardly any evidence of calving was observed in the study area suggests that the blocks of ice calved were probably relatively small (mostly < 2, and < 10 m in diameter at most).

Calving takes place in three ways: 1) serac fracturing causing the ice to fall into the water, 2) large sheets are broken off the glacier margin and slide or fall into the water, and 3) an ice block breaks off

subaquatically and emerges (Powell & Domack 1995, p. 458).

Calving rates have been proposed to be directly proportional to water depth. Studies by Brown *et al.* (1982) indicate a linear calving coefficient of 27. Consequently, 1) ice retreat declines towards shallower water, and 2) in order for the ice margin to maintain its position, glacier velocity must increase as water depth increases. This is valid when the ice margin is in a stable state, i.e. not advancing nor retreating rapidly. Due to absence of calving in shallow water glacial retreat depends on ablation, i.e. glacier melt. Associated more frequent annual ice marginal oscillation is observed (Crossen 1991, p. 134).

In general terms, glacier behaviour seems to depend more on changes in glacier bed conditions, especially on effective basal pressure (P_e) rather than on ice marginal conditions. Faster ice marginal retreat in valley areas may be due to faster flow resulting in thinning of the ice mass, which leads to reduction of ice marginal thickness in excess of floating (Van der Veen 1996).

Okko (1957, p. 39) states that calving takes place at thin concave ice margins. Straight and thick edges will not calve. Also Sauramo (1923, p. 146, 151) stresses the significance of ice marginal embayments, where ice flow concentrates and calving takes place. If calving and velocity rates of ice are in balance the ice margin would be expected to remain straight, however.

Zilliacus (1987) presented the hypothesis that De Geer moraines are formed sublacially as post-surge basal till thickenings. The question remains, however, how it would be possible for a glacier to become detached from the underlying bed so that the 'casts' of crevassed basal ice were preserved as ridges instead of becoming destroyed by continuous glacial activity and gradual calving. Modern glaciological studies show that basal crevassing is unlikely (Benn & Evans 1998).

Calving was an annual phenomenon in the study area. During winter calving was reduced by sea ice that accumulated in front of the ice margin, by freezing of cracks, and by reduced ice marginal velocity (Liestøl 1976, Boulton 1986, p. 696).

Calving has been mainly studied in areas where ice margin terminates into the sea. According to Warren (1991) in areas where ice margin terminates in lake water calving is only ca 10% of the total calving volume compared to the areas where an ice margin is in contact with a sea. This is due to the absence of tides, smaller influence of wave action and larger annual temperature fluctuation in lake environments.

4.4.4. Influence of ice marginal deposits on the behaviour of the ice margin

The behaviour of the ice margin is more dependent on the profile of the ice margin and also on changes in its thickness than previously believed (Van der Veen 1996). The increase in ice thickness caused by an ice marginal deposit that acts as an obstacle (Fischer & Powell 1998) influences greatly the ice margin behaviour. Sauramo also indicated in his studies that the deposition at the ice margin stabilises the ice margin when calving is no longer possible (Sauramo 1923, p. 153). As long as the ice margin is not grounded along significant ice marginal deposits or ridges water depth is a decisive factor controlling the behaviour of the ice margin (Fischer & Powell 1998, p. 37).

The restraining forces are dependent on mechanical strength properties of soil that forms the deposits in front of ice, especially the coefficient of friction, which translate into an angle of repose (Fischer and Powell 1998, p. 37). The material in the glaciofluvial ice marginal deposits of the study area is characterised by a relatively high angle of repose as the material is more or less coarse and relatively angular. It seems also that the dense and often coarse diamicton (Palmu 1990a, b) did effectively inhibit ice marginal readvance. Previous studies by Boulton (1986) show that an ice marginal deposit will deform if the material is sufficiently fine. Coarse glaciofluvial material will be more permeable to the extent that pore water pressure will not increase to levels sufficient to reduce shear strength below yield, in which case the ice marginal deposit will be more resistant (Boulton 1986).

Dropstones in varves, observed by Sauramo (1923, p. 119) indicate that, especially during the latter part of and during the period between the depositions of the First and Second Salpausselkä, calving is likely to have been considerably more frequent than during ice marginal halts.

To summarise, thinning of the ice margin depends on glacial flow which again depends on obstacles to ice marginal flow at the ice margin. Calving at the ice margin does not appear to be the cause but rather, the consequence of a process dependent on local glacial history (Fischer & Powell 1998, p. 40).

4.4.5. Causes of oscillation

Diamicton-dominated ice marginal ridges on top of glaciofluvial deposits and internal diamicton beds in the proximal parts of glaciofluvial deposits were formed as a consequence of simple or multiple oscillations of the ice margin. The occurrence

of diamicton in these is a typical feature throughout the Salpausselkä zone (Tanner 1933, Niemelä 1979). Swarms of small diamicton-dominated ice marginal ridges (De Geer moraines) also indicate oscillation.

Swarms of small diamicton-dominated ice marginal ridges occur in the most elevated points of the study area. By contrast, at Jylisjärvi they are located in the lowest lying areas (Okko 1957). Crossen (1991, p. 134) also observed that De Geer-type diamicton-dominated ice marginal ridges are found in valley areas and glaciofluvial, sand and gravel dominated ones in higher areas.

The formation of ridge swarms is thought to argue for the contemporaneous formation of these deposits (cf. Zilliacus 1987). Another possible explanation is annual repetition of oscillation. Swarms are often concentrated in relatively high and even terrain, examples are the swarms between Hyvinkää and Mäntsälä (Palmu 1990a). Virkkala (1963) states that the above is due to the fact that even areas were more exposed to glacier motion.

If for example due to climatic reasons the glacial retreat stops for a long period, a deposit composed of glaciofluvial material is usually formed at the ice margin. The glacier oscillates against this and push moraine is formed. A corresponding phenomenon is produced during halts of an ice margin that is in an advance phase (Boulton 1986, p. 695). The ice marginal deposit forms a stoss-side slope that affects the glacial advance. A similar setting is represented by bedrock hill stoss-side slopes where diamicton-dominated ice marginal deposits were very often formed.

Glacial readvance is a stable process controlled by negative feedback, eventually leading to a standstill. By contrast, ice marginal retreat is an unstable process with variable retreat velocities. Normally, restraining forces related to the ice marginal deposit control the retreat phase rate of annual oscillation. If these forces diminish below a critical value, rapid, continuous ice marginal retreat may result (Fischer & Powell 1998, p. 40). One or several strong retreat phases of the annual oscillation cycle may be adequate to start the larger-scale rapid retreat (Fischer& Powell 1998, p. 38).

4.4.6. Consequences of push or oscillation

The annual advance extends the furthest during early spring. Annual meltwater supply is most abundant in July. Depending on the intensity of the advance and fluctuations of meltwater supply multicomponent ice marginal fans with push structures on their proximal side are formed (Boulton 1986, p. 696).

It is probably a relatively common phenomenon that diamicton-dominated ice marginal ridges are hidden by glaciofluvial deposits. Okko (1957, p. 24) observed this phenomenon in the Jylisjärvi area of the Second Salpausselkä zone as well. The diamicton-dominated ice marginal deposit ridge located inside the deposit at the Multamäki site (Figs 27, 30) was deposited prior to the Second Salpausselkä Outer zone, that presents the main phase of deposition. At Tupsumäki (Figs 36, 39), the buried diamicton-dominated ice marginal ridge indicates that shifts in meltwater supply intensity or feeding point took place. Boulton (1986, p. 686) gives a present-day example of a diamicton-dominated ice marginal ridge hidden beneath glaciofluvial sediments.

4.5. Sedimentology of ice marginal deposits

The following section examines sedimentological processes and environments associated with ice marginal deposits by looking their sedimentology and internal structures. Conclusions as to the conditions under which the deposits were formed will be drawn on the basis of lithofacies units and morphology of structures and deposits.

Sedimentary processes that operate at the ice margin that terminates in water include fluvial, lodgement, meltout and calve-dumping, squeeze, and push processes (Powell & Domack 1995, p. 460)(Table 5).

4.5.1. Sedimentology associated with meltwater activity

Grounding-line fans form point-source depocentres with fan-shaped geometry that are formed at conduit (tunnel) mouths where meltwater streams discharge subaquatically (Powell & Domack 1995, p. 460). On the other hand, meltwater may be supplied to the ice margin as sheet flow or the conduit exit may migrate continuously producing long glaciofluvial ice marginal deposits.

Ice marginal fans are deposited 1) from the discharge while in contact with the floor of the water body, 2) through sediment gravity flows, and/or 3) through settling from suspension (Powell & Domack 1995, p. 460).

The megaforesets of Tupsumäki represent an ice-frontally formed jet stream detachment zone (Powell 1990, p. 58, Fig. 2C, p. 67). The foreset beds were formed as bedload transport pulsations that moved downslope. In addition the suspension load material that was carried by the jet stream was dispersed upon the floor of the water body. This caused poorly sorted sediment mixtures in the sequence. Occasionally the foreslope would have collapsed and material of debris flow origin was deposited.

As the sedimentation level of a fan-shaped delta progressively rises up to the water level at the ice margin, a fountain is formed at the ice margin/delta contact. A modern analogy is known from the Malaspina glacier in Alaska (Gustavson & Boothroyd 1982, 1987). An ice marginal restraining obstacle (ice marginal deposit or diamicton-dominated ice marginal ridge) forces the meltwater to flow upwards thus redirecting the material supply upwards from the ice marginal zone (grounding line) and giving rise to fountains (Crossen 1991, p. 133). The Tupsumäki site and particularly the Multamäki site are composed of fountain fed material. Although no direct observations were possible to effect from the proximal part of the Vuonteenmäki site, it is probable that the proximal part was also fountain fed.

Delta processes may also be classified as follows (Syvitski *et al.* 1988, Powell & Domack 1995): 1) bedload dumping, 2) hemipelagic sedimentation, and 3) by-pass and diffusion of sediment. In addition, low-density turbidity currents and large-scale slides carry material from delta environments to deeper water (Powell & Domack 1995, p. 460). Part of the material released from a fountain to be transported by meltwater flow is considerably coarse. In a fluvial environment and in shallow water, water flow has higher transportation potential and therefore it is able to transport coarse material longer distances than in a fan. This explains the coarseness of the delta topset material especially in sandurs compared with underlying foreset and bottomset deposits (Boulton 1986).

4.5.1.1. Meltwater fed ice marginal proximal units

In a majority of narrow ice marginal deposits dominated by glaciofluvial material, sand and gravel

Table 5. Depositional processes in a glaciolacustrine ice margin (based on Boulton 1986, p. 688, Ashley *et al.* 1991, Powell & Domack 1995, p. 454, 460, Dowdeswell & Powell 1996, p. 181, Syvitski *et al.* 1988.

Place of deposition	Process I	Process II	Notes
Subglacial	-subglacial basal till deposition		ramp type, diamicton-dominated ice marginal ridges - also called morainal banks
Ice margin contact debris-transportation	- frontal dump, mass movement	- sliding - slumping - sediment gravity flowage - squeezing - pushing	
From meltwater transportation	- sheet flow-deposition	-rapidly migrating conduits of subglacial meltwater discharge	
	- bedload rolling (floods control transport of coarse sediment) (=traction) Ashley *et al.* 1991	dumping (delta processes, up to small pebbles, sand up to 1-2 km)	
	-foreset	-gravity avalanches -sediment gravity flow	prograding sandy, gravelly foresets grading into bottomsets (sandy, muddy), gravel fan below
	-bottomset	-by-pass and diffusion of sediment (delta processes) -suspension settling -hemipelagic sedimentation -debris flows, mass flow -turbidity currents -(iceberg debris)	

beds dip distally from the proximal side (foreset-dip, as in the Gilbert delta model), as e.g. at Erävisharju in Loppi (Virkkala 1963, p. 236, Aartolahti 1968, p. 48). The Second Salpausselkä glaciofluvial ice marginal deposits in the Jylisjärvi area show a distal dip immediately on their proximal side as well (Okko 1957). Also Glückert (1975) states that dips are usually distal. Among the typesites this can be seen at the highest part of Tupsumäki (Fig. 34) where part of the unit becomes an ice contact sedimentation unit without a transition zone. On the other hand, a proximally dipping sedimentation unit of coarse material (coarse to stony gravel)occurs at the western end of Tupsumäki (see below).

In megaforesets sedimentation took place through a foreset slope avalanche-type sedimentation trending outwards from the ice margin. Meltwater was discharged from higher up (e.g. Tupsumäki) which may be explained by fountain origins. Meltwater must have passed via the glacier bed in order to explain the abundance of material discharged with meltwater. Previously, Aartolahti (1968) and Glückert (1975) assumed that this abundance of material was due to englacial or supraglacial transport.

Part of the deposits have dips parallel to the slope, i.e. proximal dip on proximal slope. This is the case at e.g. Polari in Karkkila (Virkkala 1963, Aartolahti 1968). Possible explanations include counterslope sedimentation by extremely strong meltwater currents or collapse of proximal support (Glückert 1975, 1981). The structure may also result from backflow, or at Multamäki, lateral infill of an ice marginal cavity. Another explanation is deposition in a crevasse (Virkkala 1963, Aartolahti 1968), which is unlikely for most deposits in the Karkkila-Loppi area. Proximal glaciotectonism, asymmetric shape of (foreset) structures, and asymmetric distribution of coarse material in the deposits evidence for sedimentation in free space. Rather, a likely explanation for a proximally dipping sedimentation unit is a jet discharged directly onto the proximal slope of an ice marginal deposit (Hunter *et al.* 1996, p. 1033).

4.5.1.2. Distal and bottomset units

Bottom underflow events outside the foreset sedimentation zone or prior to foreset deposition were responsible for transporting material to the bottomset facies in ice marginal glaciofluvial deposits. The marked difference in particle size potentially found between a gravel-dominated coarse foreset sedimentation unit and a fine distal bottomset unit is due to the fact that gravel and sand materials are transported different distances, i.e. sand but not gravel can be transported further downstream into low energy environment. The sharp transportation distance boundary depends on the proportion of sand ($0.20 < f_s < 0.40$) in the bed sediment. The sand transport rate increases greatly in sand-matrix beds (Wilcock 1998).

Sediment gravity flows are divided into cohesive and cohesionless debris flows, grain flows, and turbidity currents (low density gravity flows) (Benn & Evans 1998).

Sediment gravity flow units are common in the ice marginal deposits of the study area.

Sediments of cohesive debris flow origin are found in the Vuonteenmäki megaforeset as beds with poorly sorted material and with diffuse sedimentation unit contacts and internal transition zones, unlike in ordinary foreset beds. Further away and among earlier deposits cyclic sand units occur as sedimentation units produced by low density turbidity currents with ripple and other tranquil flow beds passing into sediments settled out from suspension (Figs 17, 18).

Detailed examination of especially the Multamäki site log 1 (Fig. 26) shows sediment gravity flow deposits. The distal bottomset unit also contains corresponding sediments as e.g. in Tupsumäki.

At Tupsumäki, considerable portion of the bottomset unit material is redeposited. In addition, the megaforeset unit includes beds of cohesive and cohesionless debris flow origin, as well as grain fall and grain flow units. Well sorted gravels and sands interbedded with sediments of cohesive debris flow origin as well as classical basal till-type diamicton (log 12, Fig. 37) occur in the gently inclined proximal slope of the eastern part of Tupsumäki.

In the southwestern part of the Hirvilammi ridge, in the distal part of the distal test pit, particularly distinct sediments of cohesive debris flow origin were observed interbedded with sand and silt units deposited by fluidal flow (Fig. 45).

In this study, the term 'suspension sediments' refers to the deposition of the most distal or weak flow silts and in general of weak flow fine sands and exceptionally clays, as sediment particles sink gradually upon the sedimentation bed. Under exceptional conditions, as in association with the particularly intensive material supply that presumably took place also in Tupsumäki, even fine gravel is known to have been carried by and settled from suspension near the zone of flow detachment (Powell 1990).

The very basal and distal deposits at Vuonteenmäki are composed of suspension sediments consisting of massive or horizontally bedded sands, fine sands, and silts covering the underlying sedi-

mentation units (Figs 17, 18).

Also at Multamäki and Tupsumäki the basal and distal units were partly deposited from suspension. It seems that occasionally deposition rates were high. The sediments are poorly sorted in places and exhibit particular particle size distributions possibly forming a combination of several individual flow units, or then matrix was deposited under completely different flow conditions among coarse material between particles (Fig. 36).

4.5.1.3. Meltwater activity at waterlevel, topset units

Glaciofluvial, deltaic shallow water deposits occur only in the broadest glaciofluvial deposits among the ice marginal deposits, the majority of which is located in the Esker Area. Topset units are known from a few other ice marginal deposits, e.g. at Multamäki.

Conclusions on the Vuonteenmäki and Multamäki topset units can be drawn as follows: At Vuonteenmäki the thickness of the topset unit at its maximum is around 3 m decreasing towards the distal margin to about 1 m. Observations indicate that the sedimentatation unit is characterised by a homogeneous series of stony gravel layers together with horizontally bedded or massive sand layers. Where examined, gravel units constitute 70-100% of total topset unit material.

At Multamäki only a limited proportion of exposures allowed the observation of the topset unit. The unit starts in the proximal part of the deposit with a basal till-type diamicton unit that passes gradually into sorted topset part.

4.5.2. Deposition of diamicton at ice margin

Diamictonic material transported by a glacier and released at the ice margin will form deposits there as follows: 1) by pushing, 2) by forming a ramp-type ridge as basal till is deposited (Holdsworth 1973a, b, Barnett & Holdsworth 1974), 3) by ice-frontal dump as the ice margin is melting and calving (Powell 1981, 1982, Syvitski & Praeg 1989), and 4) by ice marginal subglacial squeezing (Andrews & Smithson 1966).

In the study area, small-scale diamicton-type ridges running parallel to the ice margin are interpreted as ice marginal in origin (see also Virkkala (1963) and Glückert (1975)). This interpretation is based on the following facts: asymmetrical shape of ridges, dense basal till type diamicton as proximal part material, and the material of the steep distal flank frequently composed of glaciotectonised combinations of sorted beds and sediments of debris flow origin.

Proximally, subglacial deposition of basal till takes place through lodgement, melt-out and deformation (Dreimanis 1989, Benn & Evans 1998). At the ice margin, sediment is released through subglacial squeeze and push. A steady stream of material reaches the ice margin at the contact plane between bed and glacier base. Basal parts of the glacier body transport material that is released at the margin as the ice melts. Subglacial meltwaters may be released as well.

Material is deposited beyond the contact of an ice margin at a critical angle of repose by sediment gravity flows and meltwater transport. Dowdeswell and Powell (1996, p. 181) give present-day examples on steep distal slopes and their depositional conditions, i.e. on coarse material sliding downslope. Their studies are supported by photograph records obtained with remote control equipment. Previously, Boulton (1986) used sidescan sonar and high-resolution echo sounder to investigate ice marginal conditions of a water terminating glacier.

The deposition of small-scale diamicton-dominated ice marginal ridges lasted probably only 1-3 years (see also Glückert 1975). In principle, it is possible to detect annual oscillations by observing individual proximal basal till-type diamicton, and on the basis of morphology, i.e. the fact that ice marginal ridges converge to form groups of ridges.

4.5.3. Further sedimentological aspects of ice marginal deposits

Oscillation is indicated quite frequently by both surface and internal diamicton beds in the proximal parts of glaciofluvial ice marginal deposits as discussed above. Previously, Virkkala (1963) observed that basal till-type diamicton beds in the study area indicate ice oscillation. This type of sedimentation unit occur in both the Multamäki and Tupsumäki ridges (Figs 38, 43). Oscillation is also indicated there by such deformation structures as shearing, imbricate thrusting, and folding. These deformation structures indicate fragile and plastic deformation. Deformation was intensified towards the proximal side.

Loading structures, i.e. syn- or postdepositional deformation structures, are typical of the distal parts of the deposits in the Second Salpausselkä Outer zone. These were observed e.g. at Vuonteenmäki and Multamäki (Figs 18, 27) and at Polari in Karkkila. Loading structures occur in concentrated horizons of 1-3 m or rarely 5 m thick. They represent ball and pillow-type structures, the convolution type, and flame structures. These syndepositional deformation structures indicate fast deposi-

tion and an unstable depositional environment. They are also thought to have been deposited at the time of the Baltic/Yoldia water level drop and subsequent loading events(Glückert 1995).

The littoral deposits were separately identified at Vuonteenmäki. In the Multamäki, Tupsumäki and Hirvilammi ridges, littoral deposit interpreted are normally less than 1.0 m thick, most often 0.3-0.5 m thick. This unit is usually dominated by fine sand interpreted to be of aeolian origin deposited through shallow water.

At Vuonteenmäki, the littoral ridge located at the hill top margin passes into the topset unit without any distinct boundary, indicating contemporaneous deposition. The littoral ridge constitutes a distinct sedimentation unit, separate from the distal-bottomset unit and identifiable on a ground radar profile.

4.6. Morphology of ice marginal deposits

Fundamental regularity is what is looked for in the morphological study of ice marginal deposits. Morphological variables such as size, slope gradient, and shape are used to predict the most probable nature of sedimentation units and internal structure in the deposit (Table 6).

Diamicton- (i.e. basal till-) dominated ice marginal deposits are distinguished from glaciofluvial deposits by their gentle proximal and steep distal slopes. Ice marginal deposits dominated by glaciofluvial sand and gravel are symmetrical in their shape and often higher than the diamictonic ridges (Fig. 50). Also according to Virkkala (1963, p. 42) narrow glaciofluvial ice marginal deposits have steeper proximal slopes than diamicton-dominated deposits.

4.6.1. Morphology of glaciofluvial ice marginal deposits

Ice marginal glaciofluvial deposits have been categorized into three groups on morphological basis: 1) Most deposits are short, less than 300 m long, and run parallel to the ice margin. They were deposited while the position of the ice margin was quasi-stable or advancing (Powell 1991, p. 90). 2) Correspondingly, hummocks or sheet-type deposits are formed under conditions of constant glacial retreat. 3) The third group are deltas that are grown to the water level. These are mainly concentrated in the Esker Area (Figs 5, 10, 47) and Vuonteenmäki is a typical example of a deposit belonging to this group.

In the Esker Area and occasionally also in the other areas where meltwater routes occur the ice margin had a re-entrant, a funnel shaped bay-like form. It seems that meltwaters further enhanced melting at the ice margin. The funnel shape was probably produced as the tunnel stage developed into a delta. Less ice marginal oscillation took place there than elsewhere in the study area.

The Esker Area shows features indicating ice marginal contact. In the western exposure of Vuonteenmäki this is indicated by basal till-type diamicton and boulders (Fig. 19). Kettlehole distribution is a general indicator of the location of ice marginal contact zones and reveals the retreat of the ice margin (Figs 11, 23, 47).

Sickle-shaped deposits, i.e. with convex proximal and concave distal sides, also occur in the study area. Okko (1957) also describes sickle-shaped deposits from Lake Jylisjärvi. This type of deposit was probably developed through division of the ice margin into small-scale lobes that consequently moved against an ice marginal deposit not overriding it however.

Glaciofluvial ice marginal deposits that pass into a fan-shaped swarm of diamicton-dominated ice marginal ridges (Glückert 1975) was not observed in the study area. Suitable conditions for this process existed on both sides of the Huhtimo hill. The hill in itself abounds in diamicton-dominated ice marginal ridges, the largest of which may have been deposited contemporaneously with Aittoissuonmäki and Tupsumäki. Most of the ridges are straight while some smaller ridges are slightly bent around the northern slope of the hill. On the northwestern side of the hill the ridges continue down to the valley area but do not coalesce to form groups that attach to the margin of the glaciofluvial deposit but trend towards SW. No grouping occurs in the eastern parts either (Fig. 31).

The nature of the Salpausselkä deposits in the Finnish Lake District Ice Lobe provides an interesting comparison. Between Lahti and Joutseno the ice marginal terraces (often sandur deltas) remain of the same width for long distances with the material fining only towards the distal margin, but not laterally to any significant extent. Feeding eskers may occur at 20 km lateral intervals (Rainio 1995).

4.6.2. Morphology of diamicton-dominated ice marginal ridges

Diamicton-dominated ice marginal ridges form

Table 6. The relationships between morphology and sediments in ice marginal deposits.

Morphological factors	Dominating sediments and input types	
	Glaciofluvial sand and gravel Meltwater transportation	Diamicton-dominated sediments Debris in basal transportation
Size:	Large	Small
Height Width	High > 15 m Broad > 50 m	Low < 10 m Narrow < 50 m
Length (sector of margin)	Short < 200 m (fan)	Long > 200 m (oscillation of margin)
Slope: Proximal-	Steep (steepest = coarse)	Low gradient, less than 10° (indication of ice movement over deposit)
Slope: Distal-	Steep, 5-15°? Dip of foreset-unit (fans - moderately low gradient)	Steep, 5-20° Angle of repose (angle of friction + cohesion) (depending on sediment type)

a relatively uniform morphological group of features. The position and oscillations of the ice margin are the main factors contributing to the most significant feature of these deposits, i.e. great length.

Diamicton-dominated ice marginal deposits in the study area are 1-20 m high and 5-40 m wide. Dimensions of shallow water ice marginal ridges of 1-10 m and 2-20 m, respectively, have been observed by Boulton (1986). Liestøl (1976) observed heights and widths of 10 and 20-30 m.

The size of a ridge depends on the number of oscillations that have taken place. Small-scale ridges developed probably within 1-3 years, and larger scale ones within approximately 10 years. Glückert (1975) suggested that it took 20-30 years to form a large-scale ridge.

Regardless of their size the ridges are consistently shaped. The cross profile is asymmetrical. The proximal overridden slope is more gentle than the distal slope. Branching results from separate oscillations of various ranges. Ice marginal positions coalesce to form a larger composite ridge at that point and branching ridges are formed in the splitting zone. Present-day examples of similar type of ridge are described by Boulton (1986, p. 684).

The distal slope is usually steep reflecting the angle of repose, at which the subglacially squeezed debris or debris released from the ice front was deposited in front of the ice margin. Depending on its friction angle and cohesion properties, the material is deposited at the steepest possible angle. The debris flow and sorted material components produced by meltwater transported material, contribute to the depositional process. Figure 50 presents a slope inclination map showing differences in slope inclinations in deposits of glaciofluvial and diamicton-dominated material.

Boulton (1986) and Dowdeswell and Powell (1996, p. 181) present modern examples of steep distal slopes and their sedimentary environments. The small-scale, diamicton-dominated ice marginal ridges of the present study area were probably deposited under similar conditions. This can be seen in the Hirvilammi type-site (Fig. 40). Other similar sites are common in the study area, e.g. at Harjumäki, south of Launonen (Palmu 1990a).

4.7. Deglaciation of the study area

All three zones of the Second Salpausselkä are broad at Karkkila and become narrow towards the northeast short of the Pilpala esker. Northeast of the Läyliäinen esker at Loppi the zones consist mainly of a single ridge each (Figs 10, 48).

In the southwest, the Outer and Central zones end abruptly short of Karkkila. The Inner zone shows considerable weakening of expression as well, continuing as one weak row of diamicton-dominated ice marginal ridges west of Lake Pyhäjärvi. The elevated area located southwest of Karkkila is mostly characterised by bare bedrock and lacking

or virtually lacking superficial deposits. Absence of ice marginal deposits seems to result mainly from lack of sediment supply. Beyond the study area, in the NE, the Outer zone is discontinuous and has numerous gaps.

Almost all ice marginal deposits in the study area have indicative features that resulted from short frontal oscillations. This was also the case with the type-sites of Tupsumäki, Multamäki and Hirvilammi. Particularly the Tupsumäki site provides several examples of traces resulting from oscillation. Extensive areas of stagnant ice were not developed, contrary to the Lammi area (M. Okko 1972), or to the Tammela Uplands (Aartolahti 1968). Sedimentological studies as well as ice oscillation evidence do not support the hypothesis that material was deposited onto crevasses parallel to the ice margin (Virkkala 1963, p. 28, Aartolahti 1968).

4.7.1. Outer zone

On a larger scale, glacier bed topography determined the position of the Outer zone characterized by roughness and numerous protuberances. Bedrock hills include the outcrop situated near the Karkkila municipal wells southeast of Huhtimo, the hill of Koli, and Multamäki next to the ice marginal deposit, and the bedrock protuberances on the east side of Lake Loppijärvi against which the Outer zone was partly deposited (Figs 5, 10, 48).

Ice marginal deposits dominated by glaciofluvial sand and gravel are most abundant on the distal flanks of basins and depressions similarly to the Inner and Central zones.

South of Launonen in the Harjunmäki area, wavy ridges disposed 'en echelon' towards the NE suggest ice marginal oscillations of individual small-scale lobes (Fig. 10). Ice marginal deposits were also overridden as a result of small frontal oscillations, in Multamäki for example.

4.7.2. Central zone

The Central zone is widest at the western end (Karkkila town centre), becoming more narrow towards the northeast and finally less prominent several kilometers short of the Pilpala esker (Fig. 10). Probably the zone runs continuously northeast of the Läyliäinen esker but occupies a more distal position.

The Central zone shows evidence of stabilisation of the ice margin in stoss-side slope position in places, e.g. at Huhtimo in Karkkila, east of Siikala, and in Loppi in the Sajantilankulma-Järvikulma area and south of Lake Kesijärvi. In the Tupsumäki site the glacier meltwater stream network shows a lateral shift diverting its course from east of the Huhtimo hill to the Aittoissuonmäki-Tupsumäki zone thus indicating reorganisation of the meltwater stream system(Figs 10, 31, 48).

Intensive ice marginal oscillation is clearly illustrated by the complex genetical history of the Tupsumäki type-site.

4.7.3. Inner zone

The Inner zone is narrow northeast of the Läyliäinen esker. Southwest of the Pilpala esker it is broad (about 2 km) and branches out in two a few kilometers southwest of Pilpala (Figs 10, 48).

In the Inner zone, ice marginal deposits occur at the highest possible position in many places, e.g. an extensive area in Karkkila, east of Siikala, and the highest possible area in Loppi between the Lakes Ojajärvi and Kesijärvi (Figs 10, 48).

The shapes of the diamicton-dominated ice marginal ridges indicate that ice oscillation took place against bedrock hills and glaciofluvial ice marginal deposits. Overriding of the Hirvilammi ridge provides an example of this.

4.7.4. Interzonal areas

Between the Central and Inner zones, southwest of the Pilpala esker, narrow, diamicton-dominated ice marginal ridges are very common. In addition, a number of glaciofluvial ice marginal deposits were formed in the Siikala basin located in the same area. They were deposited as meltwaters concentrated along the depression area (Fig. 10).

On higher ground the ice margin has occupied an outer position longer, and, by contrast, retreated faster in valley areas. This is indicated by the fact that swarms of diamicton-dominated ice marginal ridges arch relatively far towards the southeast in the more elevated area east of the Siikala valley area. Correspondingly, northeast of the more elevated area diamicton dominated ice marginal ridges and ice marginal glaciofluvial deposits trend relatively further towards the north, which indicates that the glacier retreated faster along the southwestern valley area of Pilpala than in the higher area (Figs 10, 48).

4.7.5. Effect of the Tammela Upland

A relatively large amount of loose sediments in the Tammela Upland was available for the deposition of the Second Salpausselkä. According to Aartolahti (1968, p. 23) the thickness of diamicton

cover, mainly basal till, at the central parts of Tammela Upland is around 5-6 m. It is proposed here that relatively weak erosion took place in the uplands possibly due to frozen bed condition of the ice sheet (cf. Holmlund & Fastook 1993). This situation can have continued throughout several glaciations. The final deposition of diamicton cover took place during the deglaciation, probably largely during the drumlinisation that took place as the glacier moved to occupy the First Salpausselkä ice marginal position.

The upland basal till is washed of fines and does not contain large amounts of fines or clay (Palmu 1990a), which is possibly explained by the initial lower concentration of fines and clay in the upland source material. Superficial deposits around the upland area are mainly composed of silt and high areas are usually characterised by a thin or non-existent diamicton cover which is probably not only due to erosion by shoreline forces.

An alternative explanation for the supply of debris from the Tammela Upland is that the rock types are more erosive in the area. However, lithology, mineral grain size or any other bedrock factor does not support the idea that an increased supply of debris could have resulted from a more erosive bedrock.

4.8. Deglaciation of the Second Salpausselkä zone and its surrounding areas in southwestern Finland

The Second Salpausselkä zone in the Karkkila-Loppi area is exceptionally wide and it occupies the southeasternmost position in the Baltic Sea Ice Lobe area (Fig. 2). Consequently, the ice margin retreat seems to have been slow in shallower water. The early depositional stage of the Second Salpausselkä was analogous to that of the First Salpausselkä in the Hyvinkää area. This phase was followed by a period of steady retreat that gave rise to a large number of ice marginal deposits in the Karkkila-Loppi area.

The ice margin of the Second Salpausselkä in the Karkkila-Loppi area may have reached its most distal position as a result of a small oscillation of the ice front. There is not enough evidence to support this hypothesis, unless the diamicton-dominated ice marginal ridges at Haapastensyrjä were interpreted as outer margins of a restricted oscillation. In addition, the buried Multamäki glaciofluvial route indicates local and brief oscillation (estimated areal extent 1 km) reaching the most distal ice marginal position. This interpretation is based solely on morphological evidence.

During the retreat throughout and across the entire Second Salpausselkä zone of the Karkkila-Loppi area, the ice front remained active. Its active character is shown by traces of oscillation observed in all zones. Furhermore, extensive hummocky moraine fields do not occur in the Second Salpausselkä zone which suggests frontal instead of areal deglaciation.

The area where most of the ice marginal deposits occur is outlined in the southwest by the depression and meltwater routes running across central Karkkila, and in the northeast by the ridge running through Launonen. The higher area that is located southwest of Karkkila in the southwestern corner of the study area is characterised by an almost complete lack of ice marginal deposits, probably resulting mainly from limited ice marginal debris supply (Figs 10, 48, 49).

At Nummi-Pusula i.e. southwest of the study area in 10-km wide sector of the Second Salpausselkä is fragmented (Niemelä 1985, Niemelä *et al.* 1994) . The rugged local relief amounts to over 80 m. The ice marginal zone is narrow and dominated by narrow glaciofluvial deposits with diamicton found both on the surface and inside the deposits. Diamicton-dominated ice marginal deposits occur as well. According to Virkkala (1963, p. 46) the ice margin was stationary and mostly governed by the rugged topography of the Nummi area. On the other hand the more flat bedrock topography in the Karkkila-Loppi area allowed faster glacier motion. It seems that the volume and size of ice marginal deposits were influenced by water depth at the ice margin rather than by flatness of underlying bedrock topography. The comparatively limited number of deposits in the Nummi-Pusula area may also result from thinner erodable superficial deposits (in comparison with the Karkkila-Loppi area) and maybe also the more limited meltwater supply.

A valley depression with one large esker system and two minor meltwater routes is located at Janakkala, northeast of the study area. The fragmented Second Salpausselkä of this area is composed mainly of glaciofluvial deposits. Diamicton-dominated ice marginal deposits are mostly absent. Northeast of the valley area, the southeastern margin of the Second Salpausselkä zone was established in a zone mainly consisting of separate glaciofluvial ice marginal deposits, of fragmentary chains of

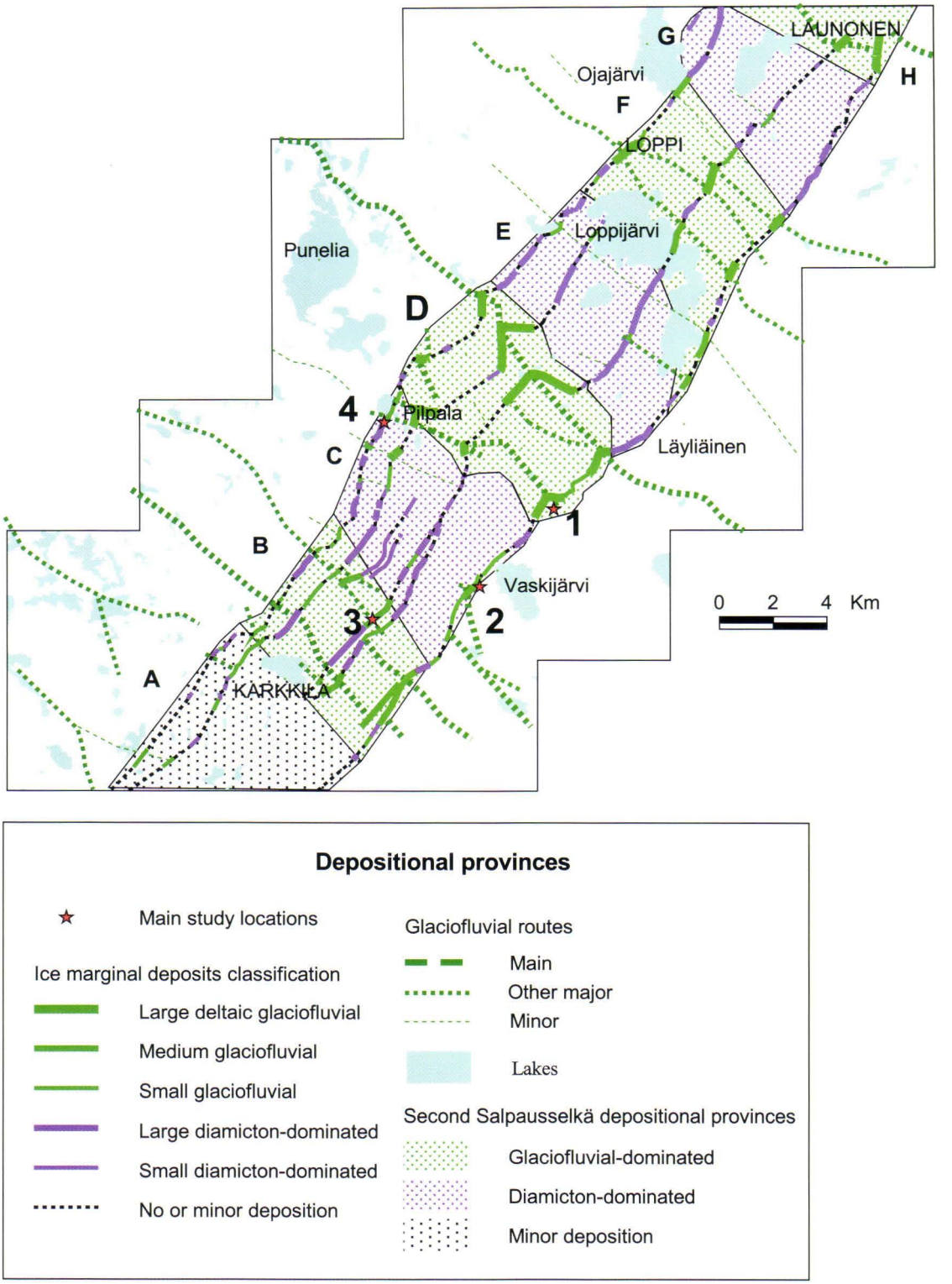

Fig. 49. Depositional sectors of the Second Salpausselkä. A: Karkkila, C: Vaskijärvi, D: The Esker Area, E: Järventausta, F: Loppi, G: Kesijärvi and H: Leinonen. Large valley systems have influenced the location of main meltwater routes. Diamicton-dominated deposits represent the main type of ice marginal deposits in higher areas between the meltwater sectors.

ridges, and partly of hummocky moraine areas (M. Okko 1968) (Fig. 2).

Water depth during deglaciation in the Karkkila-Loppi area was similar to that of the Lake Jylisjärvi area where the Second Salpausselkä has a narrower and less clear-cut expression than in the study area. Valley areas are marked by ice marginal activity with high diamicton-dominated ice marginal deposits formed by multiple oscillations. Diamicton-dominated ice marginal deposits were formed in

the first, active stage and were subsequently buried under glaciofluvial sedimentation during the second and more passive stage (Okko 1957), p. 21-26).

The setting in the Lammi area with its three main ridges is identical to that in the Esker Area of the present study area. Traces of oscillation and ice marginal diamicton-dominated deposits are absent (M. Okko 1972). Deglaciation resulted from regional stagnation which is also indicated by the presence of extensive fields of hummocky moraine. In the Karkkila-Loppi area this type of fields occurs in the southeast and northwest sides of the Second Salpausselkä zone, but not within the zone itself. It seems that their development requires that the ice margin terminate on dry land or in very shallow water.

The Second Salpausselkä zone is not prominent between Lammi and Asikkala where it consists of only a narrow discontinuous zone of deposits. In other parts of the zone, areal stagnation type deglaciation dominated (M. Okko 1972). Definite traces of ice marginal oscillation are absent.

Although ice marginal depositional conditions with respect to water depth and meltwater supplied material were analogous to those of the Karkkila-Loppi area, considerably less deposition at the ice margin took place. Apparently ice marginal oscillations were significantly weaker. There must have been considerably less material migrating to the ice margin than in the Karkkila-Loppi area which was fed from the Tammela Upland where abundant material was supposedly available. Here the proximity of the border of the Baltic Sea Ice Lobe and the Finnish Lake District Ice Lobe seems to be very significant, in contrast to the Karkkila-Loppi area which was located centrally to the lobe margin.

The behaviour of the ice margin shows irregular, overlapping cyclicity. Yearly cycles are supposed to have included the meltwater cycle and the ice marginal winter oscillation cycle reflecting temperature change. Winter oscillation may be reflected as diamicton ridges of annual moraine type. Annual cyclicity of deposition resulting from meltwater supply is rather difficult to establish. Cyclicity is also reflected in the deposition of large-scale diamicton-dominated ice marginal deposits formed during several years.

In the study area the Second Salpausselkä is composed of three zones. The inner zone is further divided into two parts. To the southwest of the study area, the Second Salpausselkä is also divided into three main zones taking into account the two main Second Salpausselkä zones in the area and that the distal (southeastern) zone exhibits a depositional break between two delta components (Kujansuu *et al.* 1993, p. 43) thus forming two parts. If each zone is regarded as a large-scale sedimentation cycle, assuming that the Second Salpausselkä stage lasted an estimated 200 years in total, one cycle would have been ca 60-70 years. According to Glückert (1975, p. 52) the zones were deposited synchronously in periods of 30 years with intervals of 50-60 years between the periods of 30 years. Previously, Virkkala (1963, p. 45) examined the duration of the deposition of the Karkkila area ice marginal positions using the varve chronological data of Sauramo (1923). Virkkala (1963)estimated (on the basis of four zones) that the completion of one zone lasted only ca 20-30 years on average.

No fundamental glaciodynamical cause for cyclicity in the behaviour of the ice margin is apparent. Neither did topography influence the fundamental genesis of the zones although within any one of them, it did influence small-scale orientation of the ice margin (see also Virkkala 1963, p. 46, 70, Glückert 1975, p. 52). Thus, location determined depositional conditions at the ice margin and consequently, the nature of ice marginal deposits.

The deposition of the First and Second Salpausselkä may reflect the cyclic climatic changes (300-500 year cycles) that took place during the Younger Dryas (Stuiver *et al.* 1995, p. 148). Short-term polar front shifts were also reflected in Finland influencing the deposition of the Salpausselkä zone. Thus, ice marginal depositional conditions seem to depend on climatological factors. Calving at the ice margin may have declined significantly due to longer winter and sea-ice season. Precipitation and glacier velocity are probably not that sensitive to climatic change. Even though glaciodynamic changes may often lie behind the deposition of ice marginal deposits, it seems safe to conclude here, on the basis of evidence presented above, that palaeoclimatological factors are primarily responsible for the diversity of ice marginal deposits that occur in the Karkkila-Loppi area.

It is known that the ice margin that terminates into a lake responds more rapidly and consistently to climatic deterioration than the ice margin that terminates into sea water (Warren 1991, p. 14).

5. CONCLUSIONS

1. The Second Salpausselkä zone is a broad zone in the Karkkila-Loppi area. The reason for this is that the retreat of the ice margin was slow since iceberg calving was restricted due to shallow water. The retreat of the ice margin was more rapid in the Second Salpausselkä zone southwest and northeast of the area because the ice margin terminated there into deeper water. (Fig. 2). On the other hand, it seems that the Tammela Upland acted as a source of abundant debris that was subsequently deposited at the ice margin in the study area.

2. The deposits of the Second Salpausselkä are divided into sectors with 1) deposits mainly composed of glaciofluvial sand and gravel and 2) diamicton-dominated ice marginal deposits and 3) sectors of weak deposition. Sectors of glaciofluvial deposits occur along continuous valley areas and their associated meltwater route systems. Sectors of diamicton-dominated deposits are located in the higher areas between them. Glaciofluvial material dominate in the deposits of the Outer zone (Fig. 49).

3. The deposition of the Second Salpausselkä zones was steered to stoss sides and tops of higher areas. The position of the Second Salpausselkä ice marginal zone in the Karkkila-Loppi area seems to reflect more southwestern conditions where the ice marginal position was steered to the upper part of distal slopes owing to declined calving in shallowing water. In the Outer zone, evidence of a short readvance can be seen in the Koli-Multamäki area of Karkkila. In other parts observations do not indicate glacial readvance.

4. Two fundamental processes influence deposition at the ice margin: 1) deposition of basal till-type diamicton associated with glacier readvance and 2) meltwater discharge as a jet-plume depositing a fan that may grow into an ice marginal delta provided that the duration and intensity of the processes are sufficient. Debris flow processes are common but less significant.

5. Deposits dominated by glaciofluvial sand and gravel were laid down in deeper water normally exceeding 40 m and situated along meltwater routes. On lee sides they were deposited possibly in shallower water. Diamicton-dominated ice marginal deposits were laid down in shallow water less than 20 m deep on stoss sides and hill tops (Fig. 48).

6. Oscillation of the ice margin was a common phenomenon in all zones. Proximal slopes were typically affected by intensive overruns of ice. Push moraine ridges are common on stoss sides of protuberances, and analogously, proximal parts of ice marginal glaciofluvial deposits have been often buried by ice oscillations. Deformation caused by ice oscillation is reflected as shear structures, folding, and imbricate thrusting in the proximal parts of the deposits (Figs 38, 43). Oscillation does not always imply deposition of diamicton onto the proximal slope as illustrated by the western end of Tupsumäki.

7. The following general rules are thought to characterise the interplay of form, internal structure and sediment in the deposits: Large and high ridges parallel to the ice margin are composed mainly of glaciofluvial sand and gravel. If a diamicton-dominated ice marginal ridge is higher it is the result of multiple oscillations to the same ice marginal position during a number of winter seasons. Small scale ice marginal ridges are usually dominated by diamicton of debris flow origin. Also small narrow ice marginal ridges of glaciofluvial origin do exist, however. On the basis of slope dip observations the material is more likely glaciofluvial sand and gravel if the ridge is symmetrical, i.e. the proximal slope is steep as well. The proximal slopes of basal till type diamicton dominated ice marginal deposits are often gentler owing to ice oscillation, but steep slopes are also found. The distal slopes of diamicton-dominated ice marginal ridges are usually steep corresponding to the angle of repose of the sediment of glaciofluvial and debris flow origin (Fig. 50).

8. The complexity (the quality of being composed of multiple elements) of glaciofluvial and diamicton-dominated ice marginal deposits results from intensive fluctuation in both time- and space-related ice marginal processes (Figs 30, 39). The interrelated fluctuations of the following elements form the main factors that contributed to the diversity of deposits: glaciofluvial deposition, glaciofluvial sediments, i.e. sands and gravels covering diamicton-dominated ice marginal ridges in the distal parts of deposits (Tupsumäki and Multamäki), debris flow, and overriding ice oscillations (e.g. Hirvilammi, Tupsumäki, and Multamäki).

9. Frequency of oscillation and scarcity of hummocky moraine indicate ice marginal activity throughout the study area and frontal, i.e. ice marginal deglaciation. The abundance of meltwater deposits indicates that the glacier was warm based, at least for a large part in the marginal zone.

10. Two components, i.e. alternating standstill and retreat phases of varying time-spans, characterised the ice marginal retreat. No property of the glacier bed explains this cyclicity. Thus, the answer may be provided by glaciodynamics. Cyclicity may have resulted from fluctuation in the basal

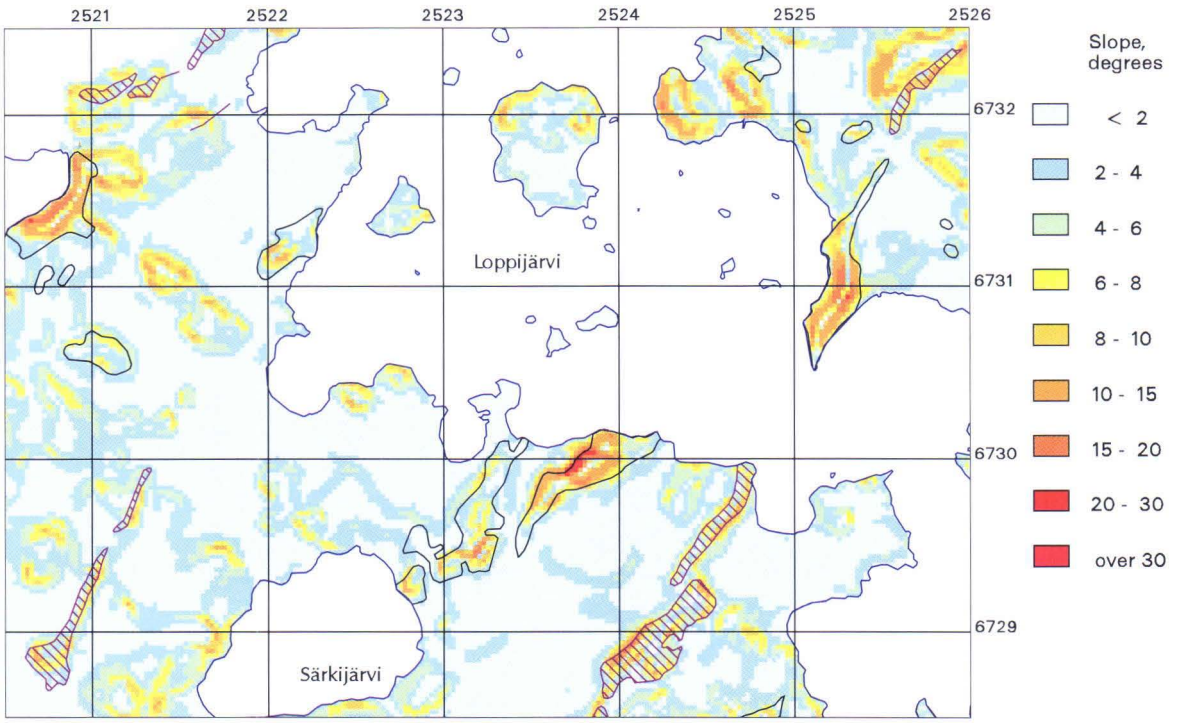

Fig. 50. Slope gradient map of the Loppijärvi area. Slope gradient processed from a digital elevation model has been divided into nine classes. The digital elevation model is based on National Land Survey of Finland (Copyright) elevation contour data. Ice marginal deposits have been classified into glaciofluvial, sand and gravel-dominated (with border) and diamicton-dominated deposits with (hatching). Diamicton-dominated deposits are very often asymmetrical, with less steep proximal slopes. Sand and gravel-dominated deposits exhibit a more or less symmetrical profile and greater relative height.

conditions of the glacier, e.g. fluctuation in glacier velocity which may result from variation in effective pressure. This again may have followed from the cyclical thinning of the glacier, influenced by changes in climatic parameters like precipitation and temperature. Reorganisation of subglacial meltwater routes at a new ice marginal position may have shown cyclicity as well. The distribution of the Second Salpausselkä zones in the study area (Fig. 48) may reflect the climatic cyclicity of the Younger Dryas, evidenced by e.g. the Greenland ice core records (Fig. 4a, b).

11. Management and visualisation of geological data offer good opportunities for the study and presentation of sedimentary environments and deglacial stages. Tacheometer equipment facilitates mapping and follow-up of exposures. The resulting numerical data is processed and visualised using GIS software. Other three-dimensional geological, geophysical and mapping data may be added. The hillshaded Quaternary geological map, also called the relief of superficial deposit map developed through the combination of Quaternary geology mapping and surface relief data (digital elevation model) is a valuable part of data visualisation. This applies also to the three-dimensional colour and height classified hillshaded map and the water depth maps constructed on the basis of shoreline studies and the digital elevation model. Examining the superficial deposit relief map pseudostereographically it is possible to identify morphogenetic features that were deposited either in glacier tunnels or ice marginally in deltas (Fig. 47).

12. Sedimentological including palaeocurrent measurement data together with visualisation of geological information, especially the superficial deposit relief map allow a fast and accurate means of estimating properties of deposits like e.g. coarseness of superficial deposit material. This kind of data is significant for effective groundwater management and protection. The superficial deposit relief map is an excellent tool for viewing and assessing geodiversity.

13. The study area is noted for its exceptional glacial geological diversity, reflected e.g. in the multitude of highly versatile shallow water deposits. The area displays a large number of exposures and has an extensive road network. It provides suitable sites for instruction and research purposes. The geological value of this area should be considered in planning land use.

6. SUMMARY

This dissertation examines the depositional conditions and sedimentology of the ice marginal deposits and landforms of the Second Salpausselkä dating to the Younger Dryas, in the Karkkila-Loppi area, southern Finland. The study area is situated in the southern part of the Tammela Upland.

Ice marginal deposits were investigated by means of observation of exposures, seismic sounding and ground penetrating radar logging. Quaternary mapping data (scale 1:20, 000) of the Geological Survey of Finland was used in the field mapping that preceded studies on the internal structures of deposits. The author participated in the Quaternary mapping revision and carried out complementary research in 1987-1993.

The Second Salpausselkä in the Karkkila-Loppi area is composed of three zones that can be divided into further zones in part of the area. The continuous zone of ice marginal positions is disrupted in the Esker Area by a broad sector of glaciofluvial deposits.

This study focuses on four sites representative of different types of ice marginal deposit. The Vuonteenmäki and Multamäki sites are situated in the Outer zone, the Tupsumäki deposit in the Central, and the Hirvilammi one in the Inner zone.

The Vuonteenmäki delta is situated in the Outer zone of the Second Salpausselkä. The ice marginal position is revealed by the shape and stratigraphical data of the glaciofluvial delta in the form of kettlehole distribution and ice marginal contact on the northwestern flank of the deposit. On a large scale the ground water level coincides with the upper contact level of the bottomset sedimentation unit. The coarser and in part gravelly material was fed to the foreset from the NW. Fanning of foreset subunits (main banks) is brought about by the meltwater supply. During the final stage the upper part of the delta was deposited with a topset unit including a littoral ridge surrounding the distal edge of the top. At the Baltic Ice Lake/Yoldia Sea interface extensive areas emerged from the Baltic Ice Lake and became exposed to aeolian activity transporting cover sand to Vuonteenmäki as well.

The embryonic delta of Multamäki was also deposited in the Second Salpausselkä Outer zone. Glacier readvance is indicated by an esker system hidden underneath the deposit. The internal structure of the deposit reveals ice marginal oscillation e.g. in the form of a diamicton-dominated ice marginal ridge buried under glaciofluvial material. During the deposition of the upper part of the deposit meltwater supply came primarily from the west as interpreted on the basis of palaeocurrent measurements and morphological data. The material was deposited in a fan-shaped form, initially towards the SE, then to the NNE, and still later towards the east. Towards the end of the deposition the ice margin oscillated upon the proximal part of the deposit.

The Tupsumäki deposit forms part of the Central zone and dates to the same period of time as the diamicton-dominated ice marginal ridges found on the stoss (N-NW) slope of Huhtimo hill. Low down in the distal part of the ridge a diamicton-dominated ice marginal ridge is buried. The main structural unit of the Tupsumäki ridge top is a large scale fan-shaped megaforeset. It was deposited as jet stream fed material generated an ice marginal deposit that grew gradually towards the water level. The proximal part displays a large number of oscillation structures and a sharp shear plane without associated basal till type diamicton in the western part. Ice marginal oscillation advanced a long distance over the eastern part of the deposit. The eastern part of the ridge is asymmetrical, the material of which is partly deformed with imbricate thrust and basal till-type diamicton.

A transverse section cut across the ridge top of the Hirvilammi diamicton-dominated site exhibited marks of multiple glacial oscillations that had partially overrun the deposit. An over 3 m thick, dense basal till type diamicton unit occurs on the proximal slope. Units of meltwater and debris flow origin are present in the distal parts of the test pits. The depositional conditions seem to correspond to the present-day ice marginal conditions described by Boulton (1986) and Powell (1996).

In the study area, sedimentary conditions at the ice margin were dependent on bed topography in particular. The component of meltwater stream concentration depends on topographical factors. Meltwater routes concentrated along sectors of extensive valley areas. On the one hand, the ice margin diamicton-dominated deposits were formed onto higher ground, especially on stoss sides or tops of bedrock hills, or, on the other, against large scale glaciofluvial ice marginal deposits.

The ice marginal retreat rate was also controlled by topography. The ice margin curved as it retreated faster in a valley area than in the higher ground. The dependence of ice marginal retreat velocity on water depth is evident in both large scale (1-10 km horizontally, 20-100 m in depth) and small scale (horizontally 0-1 km, vertically 0-20 m) settings.

Marks of oscillation in ice marginal deposits of the study area and distribution of coarse-grained sediments in their proximal parts suggest that the

ice margin was active and ice marginal deposits were formed in processes that would normally lead to deposition of material beyond or at the glacier margin. However, to the northwest of the study area, in areas of very shallow water or partly supraaquatic areas, ice margin stagnated and melted passively. This is indicated by extensive hummocky moraine fields that were mostly deposited under supraglacial, dead ice conditions.

The positions of the ice margin are exceptionally well presented in the study area. A great number of ice marginal deposits composed mainly of glaciofluvial sand and gravel and on the other hand of diamicton-dominated ice marginal deposits and ridges characterise the area. The data assembled during this study indicates that it is primarily climatological factors that lie behind the exceptional geodiversity of the area and that glaciodynamics contributed to the wealth of deposits under particularly favourable conditions. Supply of debris to the ice margin was probably influenced by the abundance of sediment being transported from the Tammela Upland. Ice marginal water depth in the study area varied between 0 and 100 m, while most of the ice marginal deposits were deposited in 20-50 m deep water. It seems that an ice margin terminating in these water depths is especially sensitive to the slightest climatic change provided that large scale glacier bed topography is relatively even.

The deposition of the Second Salpausselkä is tentatively interpreted to have taken place during the last cold period of the Younger Dryas climatic event. More accurate dating of the deposits of the study area would allow further conclusions concerning the relationship between the global climatological fluctuation during the Younger Dryas and the functional fluctuation of the Fennoscandian Ice Sheet, as well as the climatic fluctuation of eastern Fennoscandia to some degree. Varved clay studies associated with palaeomagnetic studies constitute a suitable dating method. A potential means of checking data is provided by the Vedde Ash once it is identified among the varved sediments of the Salpausselkä zone. On the basis of this investigation it is believed to be found between the First and Second Salpausselkä, close to the Second Salpausselkä in time.

Practical applications of this study include groundwater geology and sand and gravel resources inventory. The data on location, shape, size, sections, stratigraphy, and palaeocurrent measurements of a deposit is used to predict its depositional processes and internal structure, e.g. variation in permeability and coarseness in the different parts of the deposit. New possibilities of objective geological evaluation of deposits have been developed during this study, especially through combining the Quaternary geological and topographical data for geological and landscape management purposes.

Methods of GIS visualisation have been elaborated in association with this study with the focus on the visualisation of topographical location, shapes, and presentation of various types of data. The most significant means of presentation is the combination of Quaternary geology mapping data and digital elevation model data into a superficial deposit relief map. A field investigation method for periodically repeated observation of sections has been developed using tacheometer measurements. The combination of the digital elevation model and highest shoreline level data is another example of methods applying the potential of the GIS data processing in this study.

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