The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselkä I, southern Finland

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The morphology and sedimentology of the Salpausselkä I moraine in southern Finland were examined in detail, together with the distribution of associated eskers and the glacial geology of the surrounding area. Marked contrasts in the form and stratigraphy of the moraine suggest that there were differences in the style and pattern of sedimentation along the ice/lake interface. These variations were influenced by the lake water depth and the nature of the subglacial drainage system. Large individual deltas which built up to water level were the product of conduit focused sedimentation. Lower, narrower coalescing fans of finer material were formed at the ice grounding line by sediment fed from a distributed drainage system. Subglacial conduit systems were found to be unstable where marginal water depths were greatest, favouring the development of a distributed drainage system.

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It is now known that vast systems of glacial lakes lay along the margin of both the Laurentide and Scandinavian Ice Sheets during the Wisconsinan–Weichselian glaciation (Hyvärinen & Eronen 1979; Smith & Ashley 1985; Teller 1987; Eyles & Clark 1988). Moraines accumulated in the lakes and these deposits hold a key to understanding how and where sedimentation took place and thus what the ice sheets and lakes were like.

Within the last fifteen years many researchers have studied the processes and products of glacigenic subaqueous sedimentation in both contemporary (Gustavson 1975; Smith *et al.* 1982; Pickrill & Irwin 1983) and ancient environments (Rust & Romanelli 1975; Rust 1977, 1988; Cohen 1979, 1983; Clemmensen & Houmark-Neilsen 1981; Cheel & Rust 1982; Eyles & Eyles 1983; Smith & Ashley 1985; Catto 1987; Eyles & Clark 1988; McCabe & Eyles 1988; Benn 1989; Powell in press). Despite significant advances in recent research few attempts have been made to study the styles and patterns of sedimentation in a large 'continental' lake by looking at the depositional system as a whole. In the glaciolacustrine context the depositional system includes not only the lake but also the glacier supplying the water and sediment to the lake and, (where this is present), the intervening meltwater stream network.

Emphasis in the past has often been placed on proximal to distal facies transitions and vertical profile sequences rather than lateral variability, distal basin facies rather than proximal deposits and sedimentation in small alpine lakes rather than 'continental' lakes. This results in a lack of understanding of the systematic variations in the mode of sedimentation, the architecture of the basin infill and the way in which the depositional system operates.

This study therefore addresses the relationships between the style and pattern of deposition (in a predominantly ice-proximal glaciolacustrine environment), the nature of the subglacial drainage system and the bathymetry of the lake. More specifically this study examines the way in which different types of subglacial hydrological system influence how and where sediment is delivered to and deposited at the ice/lake interface. It was hypothesised that streams feeding the lake from distributed drainage systems (linked cavity or water film) and conduit based subglacial drainage systems would differ in terms of their velocity, discharge, flow depth and water density and therefore influence the mode and pattern of ice-marginal sedimentation.

In southern Finland, the Salpausselkä moraines which were deposited in the Baltic Ice Lake during the late Allerød and the Younger Dryas (*c.* 11,250–9,900 B.P.) provide an ideal field example for

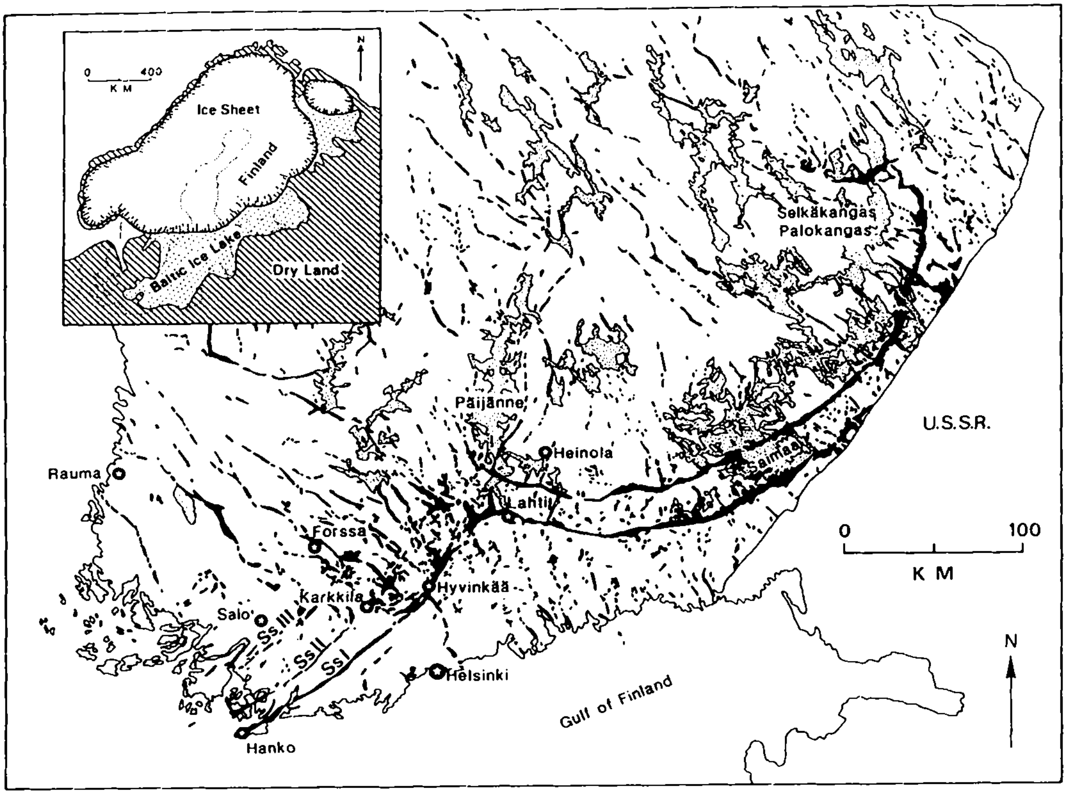

Fig. 1. Eskers and moraines in southern Finland. Ss I, II and III refer to Salpausselkä I, II and III. Inset of the Baltic Ice Lake and the Scandinavian Ice Sheet at *c.* 10,500–10,200 B.P. (redrawn from Eronen 1983).

studying large-scale lateral variations in ice-proximal glaciolacustrine sedimentation (Fig. 1). To study these moraines, fieldwork was carried out on the western arc of Salpausselkä I (Ss I), where roughly 200 km of deposits form an almost continuous ridge with a large number of exposures available for examination.

Research methods

Research was focused on the morphology and sedimentology of the western arc of Ss I and on the distribution of glacial, glaciofluvial and glaciolacustrine deposits in southwest Finland, as follows:

1. The morphology of the western arc was examined in the field, on aerial photographs (scale 1:31,000) and on topographic maps (scale 1:20,000 and 1:50,000). The long profiles of the western arc of Ss I and 20 cross profiles were constructed using 1:20,000 maps with 2.5 m contour intervals.

2. The detailed sedimentology of Ss I was examined in over 150 gravel pits along the ridge. Lithofacies and larger scale architectural 'elements' analysis was the basis of investigation (Reading 1978; Miall 1977, 1978, 1984; Walker 1984). A lithofacies code was developed during this project based on the range of facies encountered in the field with reference to existing codes for braided river deposits (Miall 1977), glacial sediments (Eyles *et al.* 1983; Eyles & Miall 1984) and glaciomarine deposits (Miall 1983; Eyles *et al.* 1985). A natural progression from facies analysis has been the development of architectural element analysis which groups individual facies into elements on the basis of depositional 'landforms' (Miall 1985). During the fieldwork it became apparent that larger scale sedimentary landforms composed of several elements and subsidiary facies characterise the glaciolacustrine environment. These larger scale sedimentary landforms are referred to here as assemblages. For example the delta topset assemblage is composed of elements (bars and channels)

which are in turn composed of facies (planar and trough cross-bedded gravels). The three-dimensional architecture of these assemblages defines the primary morphology of the moraine. The geometry of elements and assemblages was studied as faces were cut back. Sedimentary sections were levelled to nearby benchmarks to provide a three-dimensional height and location framework in which to link individual facies, elements or assemblages.

3. The distribution and nature of other Quaternary deposits in southwest Finland were examined on Quaternary geology maps of Finland at 1:1,000,000, of the field area at 1:100,000 and on 1:50,000 contour maps to put Ss I in a wider context.

Field area: the western arc of Salpausselkä I

The Salpausselkä moraines form three parallel ridges across southern Finland (Fig. 1). They were formed at the margin of the Scandinavian Ice Sheet as it stood in the waters of the Baltic Ice Lake during the late Allerød and Younger Dryas – approximately 11,250 to 9,900 B.P. (Hyvärinen & Eronen 1979; Eronen 1983; Heikkinen 1985). These moraines represent three stillstands during the retreat of the ice sheet. Ss I and II form two wide arcs across southern Finland that converge along the Lake Päijänne depression, while Ss III is found only in the west (Fig. 1). The western arc of Ss I runs from the 'bend' area just west of Lahti (Okko 1962) to Hanko on the southwest coast (Fig. 2) and can be traced under the Baltic for at least another 40 km (Häkkinen 1982).

The sedimentology and morphology of Ss I varies considerably within the 200 km of the western arc that lies above present sea level (Okko 1962; Virkkala 1963; Glückert 1986). These differences can be understood more clearly in the context of lake water depths at the time when Ss I was formed. Lake water depths can be established from present-day topography and shoreline displacement curves for the area. Raised shorelines cut into

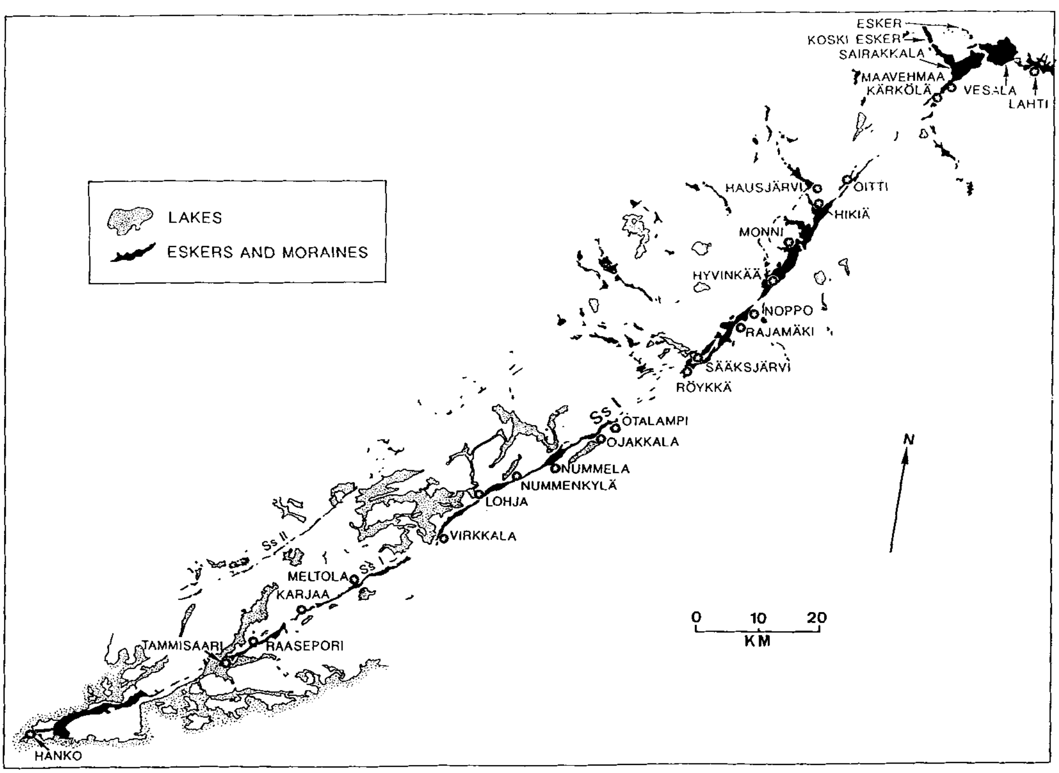

Fig. 2. Western arc of Salpausselkä I showing towns (stars) and locations referred to in the text.

the Salpausselkäs and other associated landforms have been used to construct shoreline displacement curves for southern Finland (Hyvärinen 1966; Donner 1969, 1978, 1982; Glückert 1979; Synge 1982). The shorelines on which the curves are based have been identified only on the basis of their morphology and not their internal stratigraphy; problems have arisen with this method (Donner 1982).

Despite some problems concerning the precision of the displacement curves the more general picture of water level fluctuations and ice movement is reasonably clear (Donner 1977, 1978, 1982; Glückert 1979; Hyvärinen & Eronen 1979; Synge 1982; Eronen 1983). Two different delta surfaces are recognised on Ss I; the g-level (derived from a term applied to an Arctic Ocean shoreline by Tanner (1930)), and the B I level (the highest Baltic Ice Lake shoreline (Sauramo 1958; Donner 1978)). After the oldest Baltic Ice Lake deltas (the g-level deltas) were deposited, the ice retreated and then readvanced as the water level rose by 25 m to the B I level. Since only fragments of the g-level deltas have been identified in the western arc (Donner 1977) it seems likely that most of them were destroyed by the subsequent readvance of the ice and the deposition of the B I deltas. The ice margin is thought to have stood at Ss I for about 200 years (Niemelä 1971).

The ice subsequently retreated to the position of Ss II where more deltas accumulated along the ice margin at the B III lake level (10 m below the B I level). The Baltic Ice Lake drained after the formation of Ss II to form the Yoldia Sea (Hyvärinen & Eronen 1979; Eronen 1983). This 28 m drop in water level is shown by a second set of deltas on the proximal side of Ss II (Donner 1978; Glückert 1977). It is widely stated that the drainage took place through the Billingen 'gate' in central Sweden although considerable discussion continues about the events that took place in this area (Strömberg 1977; Eronen 1983; Berglund & Mörner 1984; Björck & Digerfeldt 1984, 1986, 1989; Björck & Möller 1987; Lundqvist 1986, 1987, 1988).

The B I water level left a shoreline that is now visible at 150 m at Hikiä rising to 160 m at Kärkölä and falling again to 155 m at Vesala and 150 m at Lahti (see Fig. 2 for locations) (Donner 1977). Using the long profile of Ss I from Vesala to Hanko, and the known water levels, it is possible to work out the degree of tilt due to differences in isostatic rebound by projecting shorelines southwest along Ss I (Fig. 3). The degree of tilt on the B I shoreline for Vesala to Hanko was *c.* 0.009° as a result of *c.* 30 m of differential uplift (Vesala has been uplifted *c.* 155–157 m and Hanko only *c.* 125–127 m). Although the main isobases of isostatic rebound run roughly parallel to the western arc of Ss I the southwest end of the arc was further from the centre of ice loading (Eronen 1983). From Fig. 3 water depths at the time Ss I was deposited can be determined. The lake water was over 100 m deep at Hanko and shallowed towards the northeast until delta tops rose above lake level near Vesala.

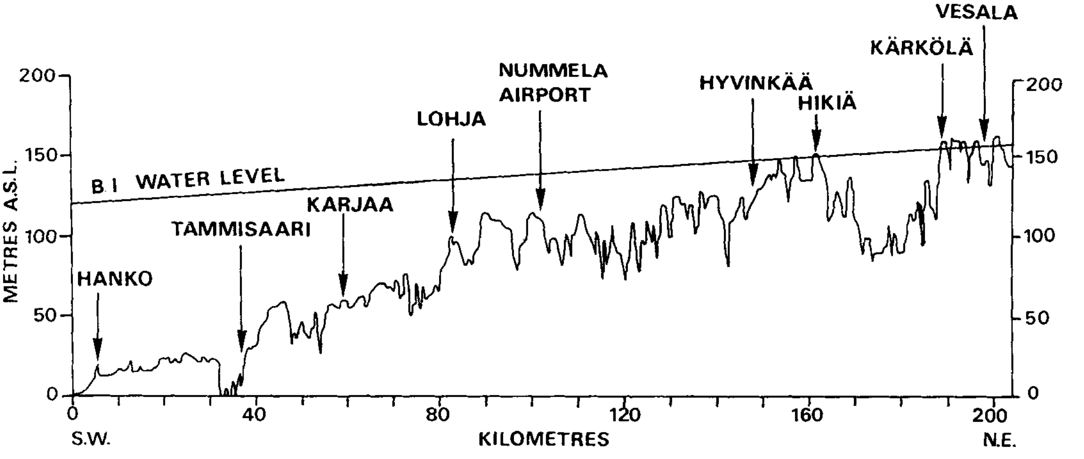

Fig. 3. Long profile of the western arc of Salpausselkä I. Locations can be found in Fig. 2.

Glacial geology and the distribution of eskers

Quaternary geology maps of Finland show a zone to the northwest of Ss I between Hyvinkää and Lahti in which eskers, drumlins and ground moraine are aligned parallel to the last direction of ice movement (NW to SE). They are associated with similarly aligned areas of eroded bedrock. Between Hyvinkää and Vesala seven eskers feed into Ss I (Fig. 2). Southwest of Hyvinkää two eskers stop short of Ss I near Sääksjärvi. Others end between Ss I and II, for example the Pori–Forssa esker (Fig. 1). Further southwest (SW of Karkkila) drumlins are virtually absent, and the eskers stop to the northwest of Ss III.

Between Rauma and Salo (Fig. 1) small ridges known as De Geer moraines run transverse to the direction of ice movement (Aartolahti 1972). De Geer moraines are argued to form at the grounding line of an ice mass standing in water (Barnett & Holdsworth 1974; Andrews & Matsch 1983). They

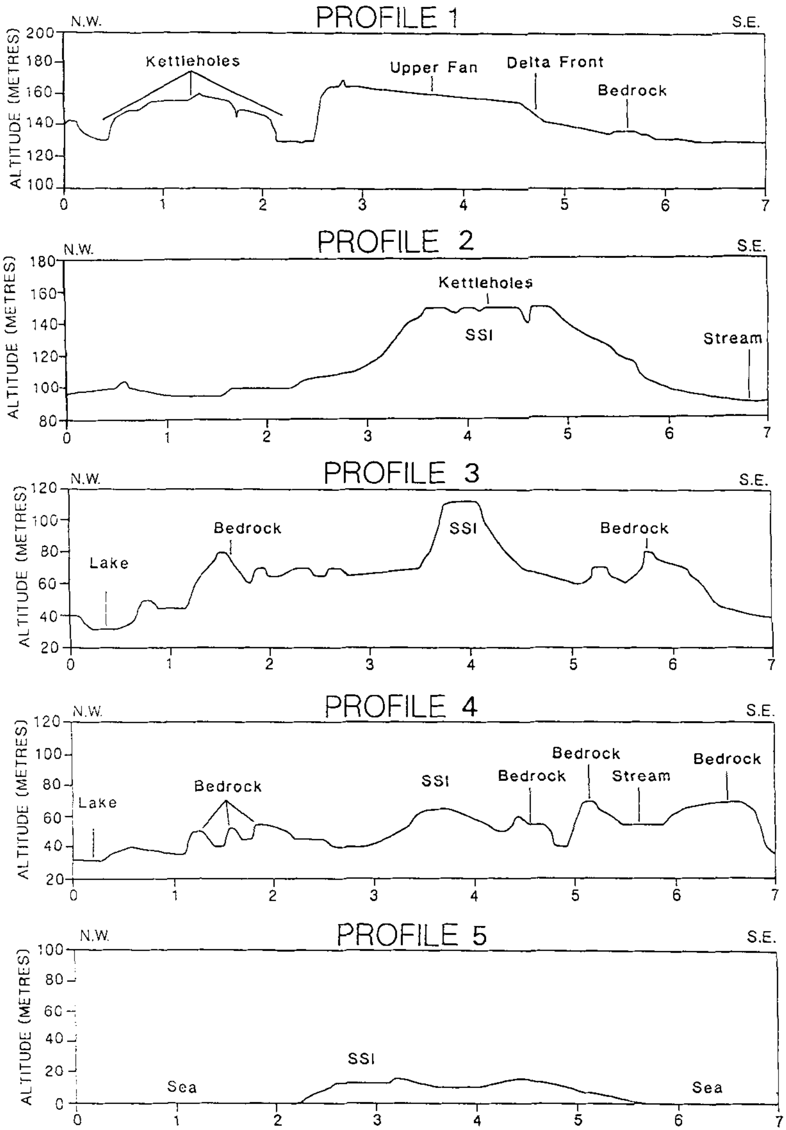

Fig. 4. Selected cross profiles through Salpausselkä I. Vertical exaggeration approximately ×16.

represent successive retreat positions of the grounding line and indicate that in the southwestern corner of Finland it retreated below water level with frequent halts. In contrast, as the ice retreated from the northeastern end of the western arc above water level, it did so steadily without halts until another stillstand in which Ss II was formed around 10,425–10,225 B.P. (Niemelä 1971; Donner 1978; Saarnisto 1982).

Sedimentology and morphology of Salpausselkä I

Vesala to Kärkölä

Where the western arc of Ss I meets the eastern arc at Lahti, the ridge is formed by four separate plateaux (Okko 1962). This study includes the two westerly ones – the Vesala and Sairakkala plateaux. Both plateaux can be classified as braid deltas (McPherson *et al.* 1987, 1988; Nemec & Steel 1988). The Sairakkala plateau is formed by the coalescence of two conduit fed deltas whereas the Vesala plateau is a separate, large arcuate delta. Both plateaux are wide (up to 3.5 km) with well-developed upper fan surfaces and short, steep delta fronts (Profile 1, Fig. 4). They resemble parts of the Selkäkangas–Palokangas formation in eastern Finland (Fig. 1), identified as a sandur delta by Eronen & Vesajoki (1988).

In contrast to areas further southwest, both the plateaux are covered by an extensive network of shallow channels and their proximal sides are pitted with kettleholes (Fig. 5). The channels, which are similar to those of contemporary sandur plains, are interpreted as relict meltwater channels (Price 1971, 1973). Larger kettleholes, which are up to 30 m deep and 400 m in diameter, are concentrated along the proximal sides of the plateaux but some smaller kettleholes (100 m in diameter) also puncture the sandur surface beyond the proximal zone.

In their proximal parts the plateaux are composed of till and boulder spreads (Okko 1962) and towards their distal margins by foreset and topset assemblages (Fig. 6). The lowest lying sediments are fine-grained sands and silts (Facies Sp, Sr, Sh, Fl, Fd). They form part of the distal foreset assemblage and are the product of traction currents and settling out from suspension. These sediments are

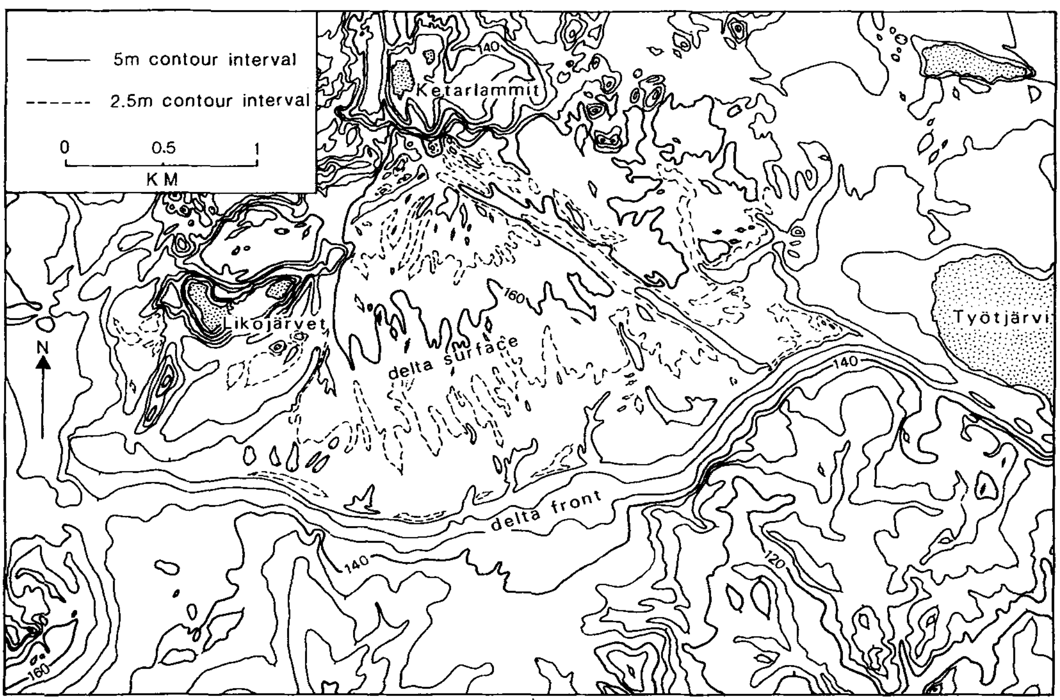

Fig. 5. Vesala fan showing the braid delta surface, the arcuate delta front and the concentration of kettleholes at the head of the fan.

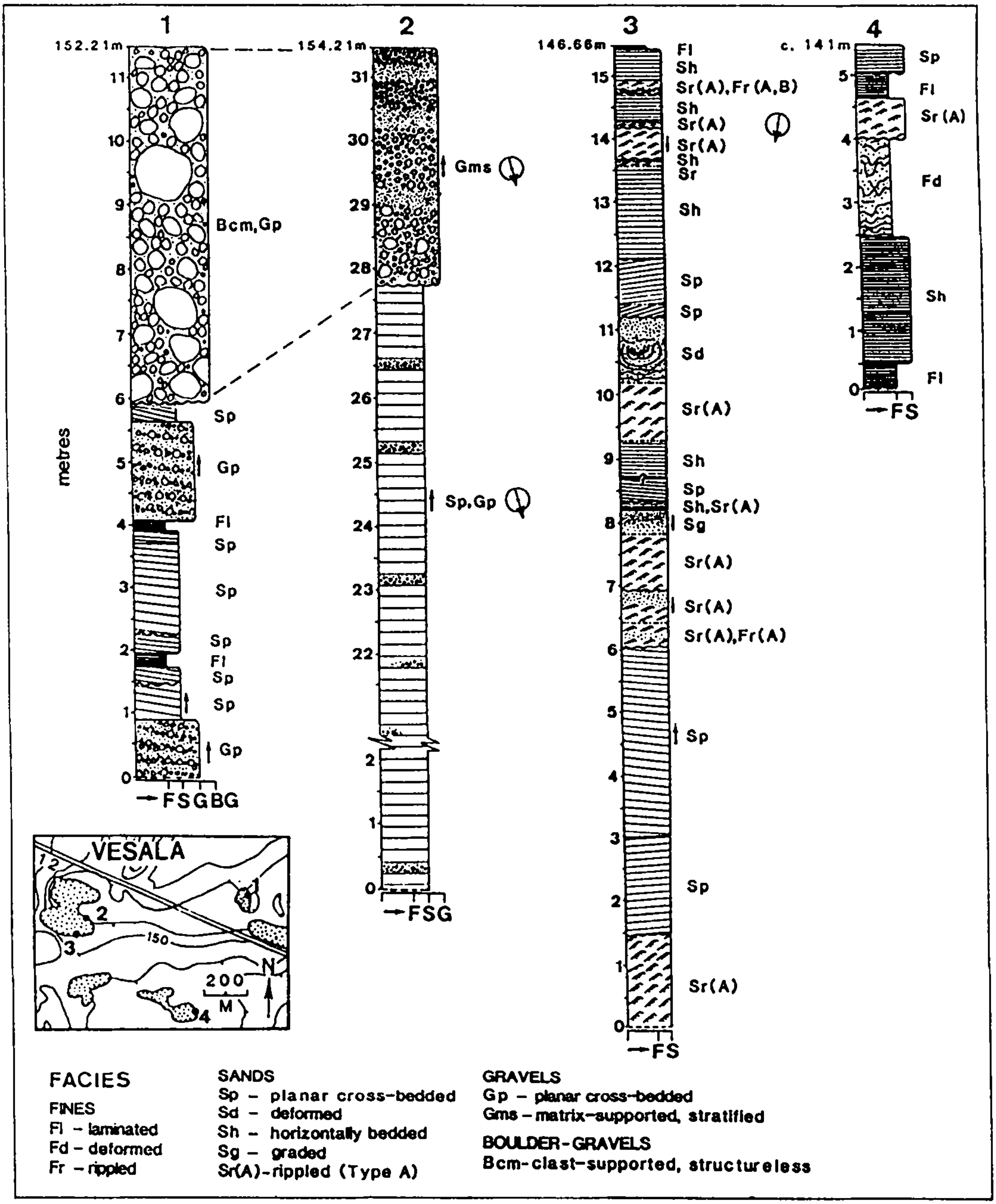

Fig. 6. Sedimentary logs from selected locations between the Vesala and Sairakkala plateaux. Numbers at top left of columns give the altitude of the top of the section above present sea level. Inset shows locations; gravel pits stippled. Dashed lines represent stratigraphical correlations.

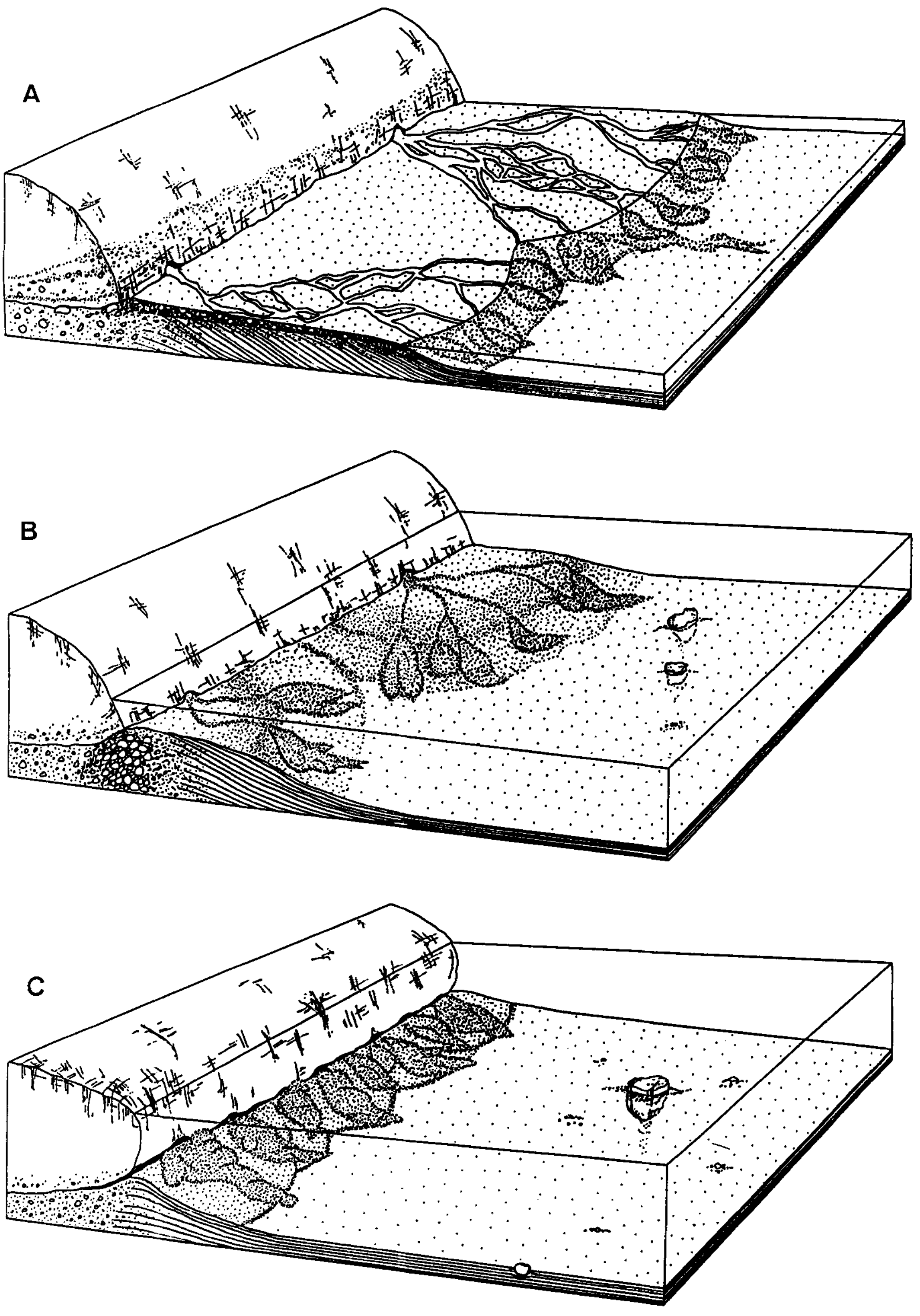
A
B
C

Fig. 8. Planar cross-bedded sands and gravels overlain with a smooth erosional contact by structureless and crudely bedded boulder-gravels. Note the sheet-like geometry of the boulder-gravels.

known to fine distally into varved clays (Okko 1962). Lying above them is a coarsening upwards sequence of distally dipping planar cross-bedded sands with a tabular geometry thought to be the product of traction sedimentation. In more proximal exposures the sands grade into matrix-supported planar cross-bedded gravels.

The delta forest assemblage is at least 20 m thick and has a thick wedge-shaped geometry (Fig. 7A). It is overlain with a smooth erosional contact by a layer of structureless or crudely bedded, clast-supported boulder-gravels (Bcm). This layer, which is typically 5 m thick and has a sheet-like geometry (Fig. 8), is interpreted as the product of high energy turbulent flows. The boulder-gravels are incised by shallow channels infilled by tabular beds of clast-supported planar cross-bedded gravels (Gp). These planar cross-bedded gravels are the product of bedload deposition and form migrating bar fronts. This sequence forms the delta topset assemblage (Fig. 7A).

The foreset and topset assemblages together form a 'Gilbert' type braid delta (Gilbert 1885; Gustavson *et al.* 1975) and there is no transition zone between the two assemblages (Colella *et al.* 1987). The Vesala and Sairakkala plateaux are interpreted as Gilbert type braid deltas (McPherson *et al.* 1988) on the basis of their sedimentary architecture and morphology. They contrast with Ss I further southwest which is known to have lain below the B I water level (Fig. 3).

Near Kärkölä, Ss I is high (up to 45 m), narrow and has a steep ice contact slope on its proximal side. The sedimentary architecture of this part of the ridge is dominated by distally dipping planar cross-bedded sands that are part of the foreset assemblage. This assemblage is overlain with an erosional contact by a 4 m thick unit of crudely

Fig. 7. □ A. Deltaic sedimentation fed by a subglacial conduit system with low water levels in relation to the altitude of the ice margin. Large arcuate fans build up and overlap where the conduit exits are close together. Braided meltwater streams feed sediment to the lake at water level. Topset bedding overlies planar cross-bedded foresets. The ice surface has a relatively steep profile (crevasse patterns schematic only). The ice front is shown in an advance phase, with debris being brought up to a supraglacial position by compression. When the ice retreats kettleholes form where supraglacial sedimentation took place. Sedimentation in the lake is dominated by underflow activity (also in B and C), but intermittent overflows and interflows may also be operating. Sedimentation took place in this way at Vesala and Sairakkala. □ B. Water level is at an intermediate position in relation to the ice grounding line. Subglacial conduits feed water to the lake below water level, and separate (but potentially overlapping) fans build up. Where sediment supply is great enough or water level low enough, individual fans may build up to the lake surface. Subglacial conduit sediments are found on the proximal side of the fans. The proximal slope is faulted due to the meltout of buried ice, or the removal of a supporting ice wall, as the ice front retreats. This model illustrates the way in which sedimentation took place along the Hyvinkää plateau. □ C. Water level is high relative to the grounding line. The ice surface profile is low, and a projecting ice ramp may overhang the grounding line (omitted here to allow the viewer to see the grounding line). Water and sediment are fed to the lake below water level through a linked cavity system or through small closely spaced conduits. Many small efflux jets create laterally overlapping subaqueous fans. Sediment dispersal into the lake is restricted because the jets are quickly mixed with the lake water. This model illustrates the style and pattern of sedimentation between Otalampi and Tammisaari.

channelised stratified sandy diamicton (Dms). The diamicton is thought to be of cohesionless debris flow origin on the basis of its channelised geometry, its structure and texture (Carter 1975; Postma 1986; Maizels 1989).

Kärkölä to Sääksjärvi

Between Kärkölä and Oitti Ss I is formed by discontinuous, low, narrow end moraine ridges (Kujansuu & Niemelä 1984) (Fig. 2). At Oitti, Ss I is formed by two small parallel ridges in which distally dipping planar cross-bedded sands and gravels coarsen upwards. These sands and gravels were deposited by traction currents and high energy turbulent flows and form a foreset assemblage.

South of Oitti the ridge expands to form the Hyvinkää plateau. Ss I is up to 2 km wide in this area and forms an almost continuous plateau as far as Sääksjärvi (Fig. 2). Near Hyvinkää four eskers join Ss I, forming substantial coalescing deltas. Three of these deltas have flat tops and are pitted with kettleholes (Profile 2 Fig. 4).

Near Monni the lowest stratigraphic units are horizontally bedded, rippled and deformed fine sands (Sh, Sr, Sd) (Log 4, Fig. 9) interpreted as the products of traction sedimentation. They are overlain by a coarsening upwards sequence of distally dipping, planar cross-bedded and rippled sands (Sp, Sr) with some planar cross-bedded gravels (Gp) laid down by traction currents. This sequence forms the foreset assemblage which is overlain with a smooth erosional contact by *c*. 5 m of clast-supported boulder-gravels (Bcm). The boulder-gravels are crudely bedded along horizontal planes and were deposited as bedload by high energy flows and form the topset assemblage.

At the western side of the Pit (Log 3, Fig. 9) there are at least 8 m of matrix-supported boulder-gravels (Bmm). These boulder-gravels have a silt or fine sand matrix and an arched bedding structure in places. Within the boulder-gravels the presence of a film of fines around clasts, the rounded shape of clasts, their high concentration and the clast support mechanism all suggest a hyperconcentrated flow origin within a subglacial conduit setting. These boulder-gravels form the subglacial full con-

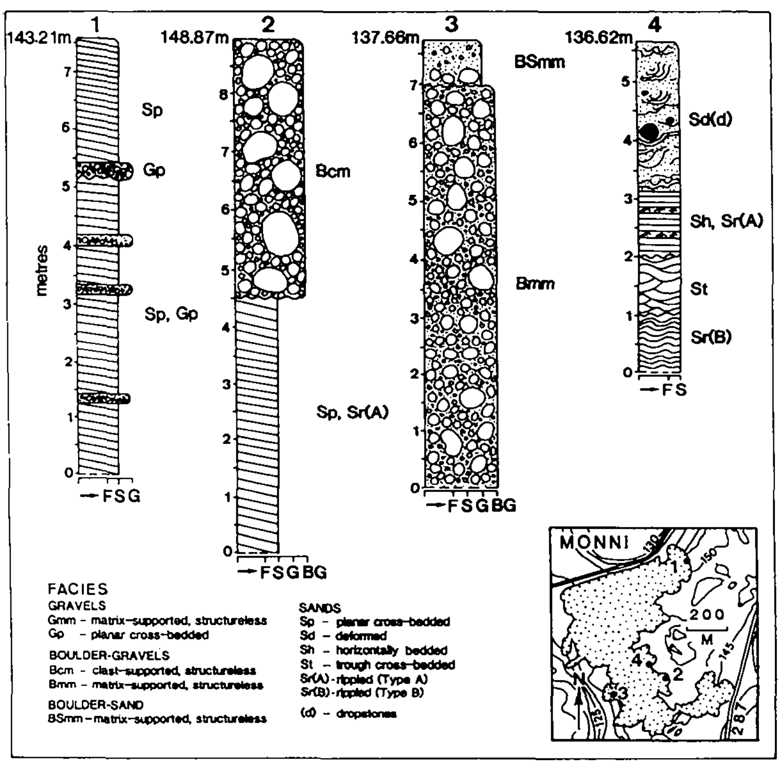

Fig. 9. Sedimentary logs from the Monni delta. See Fig. 6 for explanation.

duit assemblage. Arched bedding is found exclusively in the deposits of subglacial tunnels and requires full conduit flow (Shreve 1972, 1985). The surrounding foreset sands and gravels are heavily faulted and deformed as a result of ice meltout and loading (McDonald & Shilts 1975).

Near Noppo, Ss I is high and wide with a steep proximal slope. The sedimentary architecture of this part of the ridge is dominated by the foreset assemblage (Fig. 7B). This assemblage is cut into on the proximal slope by shallow channels filled with stratified diamicton of cohesionless debris flow origin and by proximally dipping, matrix-supported, planar cross-bedded gravels with a thin sheet-like geometry interpreted as the product of high density gravelly turbidity currents. Boulder-gravels of subglacial conduit origin grade laterally into the foreset assemblage on the proximal side of the ridge. The topset assemblage is absent in this area. Extensive normal faulting attributed to the meltout of buried ice is found particularly on the proximal side of the ridge, indicating that deposition took place in an active ice contact environment.

At Sääksjärvi, Ss I is formed by three low, parallel ridges, while further southwest it is absent as a positive relief feature as far as Otalampi (Fig. 2). Near Röykää the sequence is dominated by the foreset assemblage which is composed of distally dipping planar cross-bedded and rippled sands deposited by traction currents. The sands are occasionally interbedded with crudely channelised, stratified diamictons interpreted as the product of cohesionless debris flows.

Otalampi to Virkkala

Between Otalampi and Virkkala Ss I forms a high (45 m) continuous ridge (Profile 3, Fig. 4) but it narrows gradually towards the southwest and there are no large individual deltas. Exposures low down on the distal side of the ridge reveal horizontally or gently inclined fine sands (Sh, Sp, Sr, Sd), interpreted as distal foresets that grade into varved clays to the southeast. Above these lie sequences of distally dipping planar cross-bedded sands and gravels at least 20 m thick. In some places this foreset assemblage is eroded on the proximal side of the ridge by shallow channels filled with structureless or stratified diamictons (Dmm, Dms) thought to be the products of cohesive and cohesionless debris flows (Lowe 1982; Costa 1984; Maizels 1989). Near Ojakkala and Lohja, the top 3–5 m of sands is deformed by dewatering and loading structures (Virkkala 1963; Glückert 1986). The topset assemblage is absent in this area.

The foreset assemblage is overlain with an erosional contact near Lohja by a layer of structureless sand that includes rounded boulders (BSmm) and has a sheet-like geometry. The boulders are often concentrated at the base of the unit. This layer reaches a maximum thickness of *c.* 2.8 m near Lohja (Fig. 10) and is thought to be a littoral lag deposit. It closely resembles lag gravels described by Rust & Romanelli (1975) and Cheel & Rust (1982) formed at a littoral truncation surface on ridges composed of glaciomarine deposits near Ottawa, Canada.

Near Otalampi, Nummela and Nummenkylä

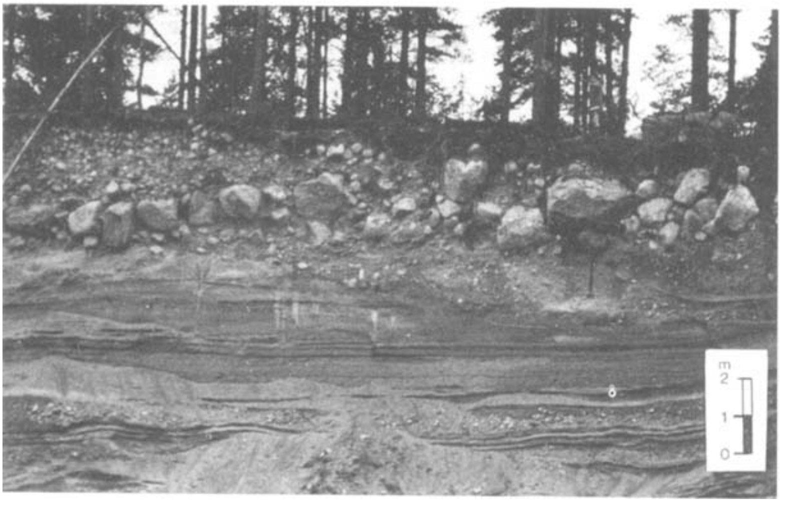

Fig. 10. Littoral lag gravel overlying planar cross-bedded sands. Larger clasts are concentrated near the base of the unit at this section. The lag layer is unusually thick at this site and is more typically 1 m thick.

exposures in small ridges to the NW of Ss I reveal structureless diamicton overlying the foreset assemblage with a smooth erosional contact. The diamicton, which has a thick sheet-like geometry and a strong clast fabric indicating movement from the NW is interpreted as subglacial lodgement till. The underlying foreset assemblage is faulted along low angle listric thrust planes towards the southeast.

Virkkala to Tammisaari

There is another short gap in Ss I between Virkkala and Meltola and then the ridge reappears and runs, almost without a break, as far as Tammisaari. Here the ridge is only up to 25 m high and quite narrow (Profile 4; Fig. 4). Horizontally bedded and rippled fine sand and silts are found in distal low lying sections. Above this is a laterally extensive sequence of distally dipping, planar cross-bedded sands which are occasionally interbedded with beds of matrix-supported gravels with a thin wedge-shaped geometry. This sequence was deposited by traction currents and gravelly high density turbidity currents (Lowe 1982) and forms the foreset assemblage (Fig. 7C). Deformation and faulting is small scale and usually confined to surface layers. The littoral lag layer is found at almost all the sites in this area and is typically 1 m thick and finer in texture than the exposure pictured in Fig. 10.

The Hanko peninsula

Ss I is low and flat and has little morphological expression on the Hanko peninsula (Prolfile 5, Fig. 4). It is distinctive both morphologically and sedimentologically and is formed by two assemblages (Fig. 11). The lower gravel assemblage has a thick sheet-like geometry and is composed of crudely stratified or planar bedded gravels (Fig. 12). Low-angle bedding planes in this assemblage dip towards the northwest (the proximal side). These gravels are well rounded, *c.* 2–20 cm in diameter and fine towards the NW across the peninsula. The gravel assemblage is thought to be the product of high energy flows in which clasts were supported by turbulence and transported as bedload. In a few places the gravels include thin (max. 20 cm) units of trough cross-bedded sands (St) thought to have been formed as subaqueous dunes moved down current creating and destroying troughs. The gravel assemblage is overlain by a sand assemblage with a thin sheet-like geometry (Fig. 11). The sand (Sm, Sh), which is fine in texture and typically 50 cm thick, is thought to be aeolian in origin (Kujansuu & Niemelä 1984; Virkkala 1963; Seppälä 1969).

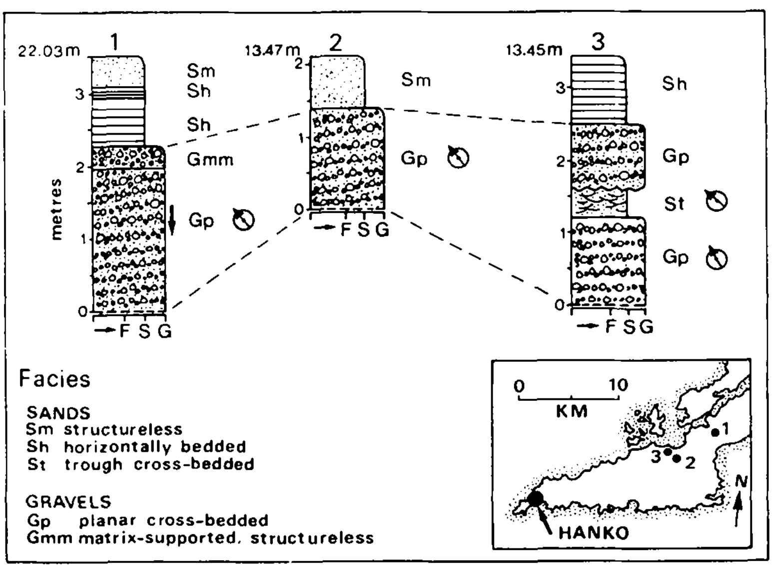

Fig. 11. Sedimentary logs from the Hanko peninsula. See Fig. 6 for explanation.

Fig. 12. Crudely-bedded well-rounded gravels on the Hanko peninsula. Section is approximately 3.5 m high.

Interpretation of the depositional setting

It was hypothesised here that the nature of the subglacial drainage system influences the style and pattern of ice-marginal subaqueous sedimentation. It is necessary to look at the way in which subglacial water is drained away to understand the influence different types of drainage system will have on ice-marginal sedimentation. Subglacial water can be drained in four different ways: (1) in a subglacial conduit system; (2) through a subglacial linked cavity system; (3) within a permeable subglacial substrate, (4) through a water sheet.

There is abundant evidence of conduit drainage systems in Finland in the form of eskers. A large network of eskers trends NW–SE, generally at right angles to the Salpausselkä moraines into which it feeds. These eskers often exhibit the arched bedding typical of full conduit flow in a subglacial tunnel (Garbutt, pers. comm.). In the NE part of the western arc of Ss I several eskers meet and terminate at the moraine, and are responsible for large accumulations of sediment in the form of deltas at these points. SW of Sääksjärvi, however, there are no eskers joining Ss I (Fig. 2), suggesting that conduits were not the primary mode of drainage. Given the high percentages of bare bedrock in the SW corner of Finland (*c.* 80%) (Kujansuu & Niemelä 1984), it is unlikely that drainage took place through a subglacial till sheet. It is therefore thought to have taken place through a distributed drainage system (linked cavities or a water film or both).

Where ice flows into water the basal shear stress is immediately reduced, resulting in extensional flow and a lowered ice surface profile (Meier & Post 1987). This, coupled with low bed slopes of generally less than 1°, implies that the Scandinavian Ice Sheet probably had a very low ice surface profile in southern Finland. Water flow beneath the ice was driven by the longitudinal pressure gradient, largely controlled by the direction and magnitude of the ice surface slope. Low ice surface slopes result in low pressure gradients and therefore tend to lead to the ponding up of water beneath the ice (Alley *et al.* 1987).

According to Fowler (1987) the stability of subglacial conduits is dependent on a parameter λ, the value of which is proportional to the ratio between glacier sliding velocity and effective pressure. As a glacier enters standing water the damming up of subglacial water flow by the lake water will lower the effective pressure and raise the sliding velocity, increasing λ. If the effect of increasing sliding velocity is to flatten the ice surface profile, this will further reduce the effective pressure and raise λ. The consequence of raising λ is to destabilise drainage conduits. Fowler's theory therefore predicts that the advance of glaciers into standing water creates conditions that favour the conversion of conduit drainage systems into linked cavity systems. For a given discharge, the steady state water pressure in a cavity system is much greater than in a conduit system (Walder 1986; Kamb 1987), so a change in the structure of the drainage system assists basal water to drain into the lake by raising the pressure in the drainage system above that of the lake. The areal extent of the cavity system will depend on the size of the zone in which lake water pressures influence conduit water pressures. This is determined by lake water depths, the discharge of basal water, ice surface profiles and bedrock topography. Larger areas of cavity drainage will be associated with higher lake water depths and lower (or reversed) bedrock and ice surface slopes.

From this it follows that different types of subglacial drainage system supply water and sediment to the ice margin or grounding line in different ways. A conduit system delivers water and sediment to the ice margin at discrete points, i.e. the conduit exit. If the conduit exit lies on dry land, water and sediment will be fed to the lake at the lake surface and deposited as in non-glacial deltas. If, on the other hand, the conduit exit lies below water level, sediment is transported into the lake in a turbulent jet (Wright 1977; Drewry 1986; Powell, in press). The higher the discharge and velocity of the incoming stream, and the higher the density contrast between it and the lake, the further it will flow into the lake before dumping its sediment load. This leads to the build-up of larger, wider fans in association with larger conduits, and with streams carrying a high suspended sediment load (Powell, in press).

A distributed drainage system delivers water and sediment to the ice grounding line via many small cavities or linking orifices distributed along the grounding line, resulting in an even distribution of sediment. The effluxes from these small exits carry lower suspended sediment concentrations and are more easily mixed with lake waters than conduit discharges, so constraining the distance of sediment dispersal (Powell, in press). Small fans will tend to overlap to form an almost continuous ridge of sediment. This is partly because of the unfocused nature of sediment supply, but also because smaller conduits or cavities are more unstable, likely to be ephemeral and therefore often switch position.

Conduit fed sedimentation of Salpausselkä I

Near Lahti the ice mass was drained by conduits and terminated on dry land. The streams issuing from the ice front bifurcated many times before reaching the lake itself (up to 2 km away) so that many shifting tributaries fed sediment to the braid delta front. 'Gilbert' deltas composed of foreset and topset assemblages developed (Fig. 7A). Meltwater channels and kettleholes on the braid delta surface formed subaerially rather than subaqueously. Small ice blocks that fell from the ice front as it retreated were buried by outwash and created kettleholes in the sandur surface as they melted. The wide zone of kettleholes along the proximal sides of the deltas indicates the extent of an ice marginal area in which supraglacial sedimentation was taking place. Compression at the ice margin was probably required for supraglacial sedimentation, indicating that the ice front was grounded above water level.

Near Kärkölä the morphology of the ridge indicates that deposition took place in close proximity to the ice front with very little sediment dispersal into the lake. The gap in Ss I between Kärkölä and Oitti can be explained by the absence of eskers joining Ss I in this area. Higher relief to the NW of Ss I between Kärkölä and Oitti may have diverted subglacial water into the Koski and Hausjärvi eskers. Where there were no subglacial tunnels delivering sediment to the ice margin (as in this area), the main source of debris was from ablation and only low, narrow morainic ridges accumulated. Deeper lake water at the ice front in this area, in contrast to the shallow water to the NE and SW, probably resulted in an unstable grounding line. Evidence of this is provided by the parallel ice-marginal ridges at Oitti (Fig. 2).

Between Hikiä and Hyvinkää the ice mass was drained by four major conduits that built up large coalescing bodies of sediment at their exits. The three more northerly conduits fed deltas that built up above water level and are pitted with kettleholes and one at least (but probably all three) is capped by topset beds. At Monni boulder-gravels with an arched bedding structure indicate that full conduit flow continued to the ice front. Sedimentation was concentrated at conduit exits, but these were close enough together to allow the deltas to coalesce into a long ridge. In this area the ice front actively influenced sedimentation, causing faulting due to the meltout of buried ice, deformation due to ice margin movements, and sediment slumping after the removal of a supporting ice wall.

Between Rajamäki and Röykkä eskers are found to the northwest of Ss I but stop *c.* 8–10 km short of it. There is no morphological continuation of these eskers (which are up to 400 m wide and 25 m high) as far as Ss I, precluding the suggestion that eskers in this area were blanketed by subsequent glaciolacustrine or glaciomarine sedimentation. The conduit system may have started to collapse close to the ice margin in this vicinity. A second line of moraines exists just behind the main ridge between Rajamäki and Röykkä (Fig. 2) and suggests that the grounding line retreated from its main position and restabilised again further to the northwest. SW of Röykkä, Ss I is formed by several small parallel ridges that mark retreat positions of the grounding line in up to 50 m of water.

Sedimentation of Salpausselkä I fed by a distributed drainage system

South of Ojakkala marginal water depths were from 40 to 100 m and the ice is thought to have been drained by a linked cavity system or closely spaced minor conduits. The ridge is narrow and virtually continuous, indicating that sediment was fed to it at many points and that sediment dispersal into the lake was restricted. As water depths increased towards the SW the area in which subglacial water pressures were affected by the lake water pressures became progressively larger. As a result the volume of sediment supplied to the grounding line, the distance it was dispersed into the lake and the size of the material transported all decreased towards the southwest. Note that Ss I on the Hanko peninsula does not retain its primary morphology or sedimentology. The underwater continuation of Ss I SW of Hanko is far narrower than Ss I on the peninsula (Häkkinen 1982).

Near Nummenkylä, Nummela, Ojakkala and Hyvinkää the ice readvanced over its own deltas, eroding them and depositing lodgement till on the proximal side of the ridge. Southwest of Lohja where the ice was merely grounded on the deltas and rarely playing an active part in sedimentation, loading or tectonism, the deposits bear less evidence of direct glacial influence.

The littoral lag gravel found at many places along the ridge is thought to have been produced during isostatic uplift (Virkkala 1963; Glückert 1986).

The effects of post-depositional uplift

It is known that Ss I continues beyond Hanko for at least 40 km (Häkkinen 1982) and that the parts of the ridge that lie in shallow water (less than 20 m) have been truncated and smoothed by wave action. On the Hanko peninsula the primary depositional feature of Ss I is also thought to have been substantially modified by wave action as this area rose above water level. As rates of isostatic rebound decrease over time, this part of the ridge was subjected to wave action for longer (*c.* 800 years according to Pyökäri (1986)) than areas to the NE, because it emerged most recently. Currents within the Gulf of Finland today move west along southern Finland, turning towards the NW into the Gulf of Bothnia around the prominent and exposed Hanko peninsula (Alalammi 1987). These current directions are compatible with the SE to NW palaeocurrent directions within the gravel assemblage on the peninsula.

Similar coarse-grained planar cross-bedded units in Denmark have been interpreted by Nielsen *et al.* (1988) as large-scale foreset sequences at the front of a prograding spit platform. Waves washing over the spit platform from the seaward side deposited and reworked material, causing avalanches of sediment down the landward side of the spit. It is concluded here that the planar bedded gravel assemblage on the Hanko peninsula was formed by waves washing over Ss I (from the SE) as it emerged from the sea. The source of the gravel size material was Ss I and the surrounding till. As the ridge rose above water level finer grained material (mainly sand) was deposited as beach ridges on the south side of the peninsula and partially reworked into dunes. The following points support this hypothesis, the well-rounded and well-sorted nature of the gravels, the SE to NW palaeocurrent directions, the fining of material towards the NW, the extent of individual beds, the unusually exposed nature of the site and the upwards facies transition from gravels to shoreline aeolian sands.

Conclusions

Subglacial drainage systems strongly influence the style and pattern of sedimentation in ice-contact glaciolacustrine and glaciomarine environments. The ice-marginal water body influences the subglacial drainage system by reducing basal shear stress at the ice margin, lowering the ice sheet profile and destabilising conduit networks. Where water and sediment is fed to the ice margin through a conduit drainage system and is then fed to the lake at the water surface (Fig. 7A), sedimentation is focused around conduit exits, sediment is dispersed far into the lake and deltas build up to water level. Where a conduit-based drainage system feeds water and sediment to the lake below water level (Fig. 7B) sedimentation is still focused around conduit exits and sediment is dispersed far out into the lake. Grounding line fans may be able to build up to water level and form deltas where water depths are shallow and there is an adequate supply of sediment. Distributed drainage on the other hand will result in unfocused sedimentation, low dispersal distances and the restriction of grounding line fan growth (Fig. 7C). In this way it can be seen that the location of the ice margin/grounding line in relation to lake water level is crucial in determining the style and pattern of ice-proximal glaciolacustrine sedimentation because of its influence on the nature of the subglacial drainage system.

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