THE SEDIMENTARY SEQUENCE OF A WEICHSELIAN INTRAGLACIAL LAKE AT ORMEHØJ (FUNEN, DENMARK)[1]

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ABSTRACT

Schwan, J., A. J. van Loon, P. G. van der Gaauw & R. Steenbeek 1980. The sedimentary sequence of a Weichselian intraglacial lake at Ormehøj (Funen, Denmark) – Geol. Mijnbouw 59: 129-138.

Weichselian deposits have been investigated in a high kamiform hill of the Vissenbjerg dead-ice landscape. The sediments were formed by the gradual infilling of an intraglacial lake underlain by a basal till. Normal deposition by a prograding delta was interrupted by the catastrophic bursting of a supraglacial lake, resulting in a boulder bed.

Various facies can be distinguished, though vertical and lateral relationships often are obscured by diapiric activity of a subjacent clay.

The meaning of the terms 'supraglacial' and 'intraglacial' (often used in a confusing way) is shortly discussed.

INTRODUCTION

The exposure at Ormehøj (island of Funen, Fig. 1) is situated in a large circular hill, South East of the township of Århus. A cross section through this hill is shown in figure 2. MARCUSSEN (1975) refers to this site and emphasizes the peculiar shape of the hillslope which is interrupted by an almost perfectly horizontal 'terrace'. The altitude of this 'terrace' coincides with the toplevel of the lower plateau hills surrounding Ormehøj.

MARCUSSEN (1977), in a paper concerning the genesis of Weichselian deglaciation-landscapes in Denmark, states that such plains with steplike arrangements represent one major element of such landscapes. Apparently he considers the Ormehøj hill plus surroundings as a prototypal model for his concept.

Indeed, the Ormehøj hill forms, together with a group of similar hills –many of them with a pronounced flat top–, a prominent feature of the Vissenbjerg dead-ice landscape (so named by SMED, 1962). Due to its characteristic landforms and sediments, this area might be considered a classical example of a dead-ice landscape of the 'uncontrolled' type.

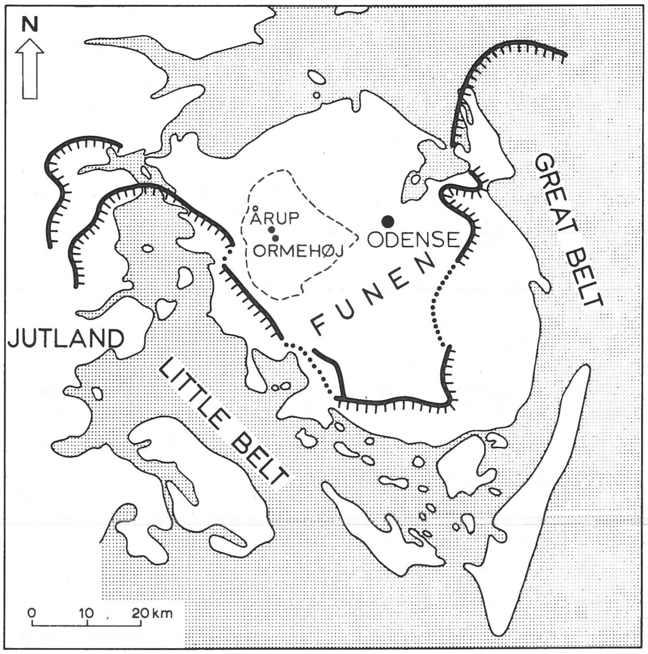

Fig. 1
Location of Ormehøj. The borderline of the Young Baltic Ice Advance, entering from the South, is indicated by a solid line where confirmed, and by a dotted line where hypothetical. No dentation is indicated unter the letter F of Funen, since it is not known to which phase of ice advance the borderline there belongs. The dashed line shows the Vissenbjerg area. After Hansen (1965) and Smed (1962).

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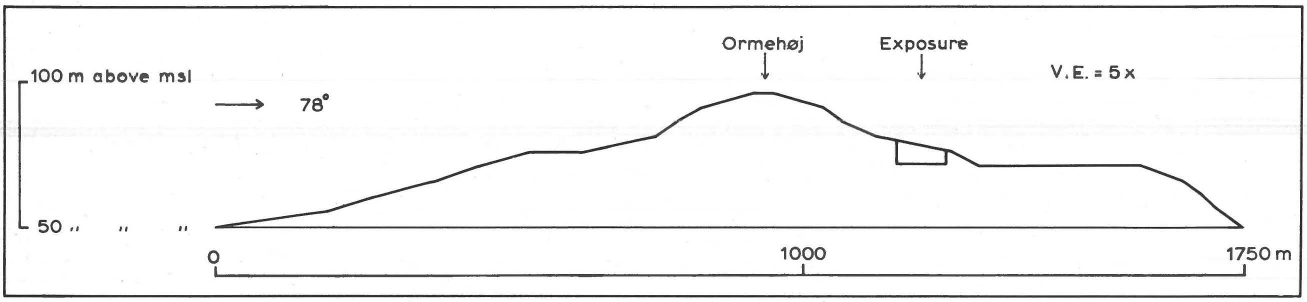

Fig. 2
Cross section of the kamiform hill E. of Årup with location of the investigated exposure. The horizontal levels in the slope are at 70 and 75 m above mean sea level.

GEOLOGICAL SETTING

According to HANSEN (1965) the youngest glacial drift of the Funen-mainland N of the borderline (Fig. 1) was deposited during the penultimate Weichselian advance (the so-called NE Ice). He also states that the glacial surface deposits S of the borderline are associated with the last Weichselian glacierization phase in Denmark (the Young Baltic Advance). Although this interpretation is still valid, a different view may emerge in the near future since the Weichselian glacial stratigraphy of Denmark now is reassessed in a very thorough way by BERTHELSEN (various papers on this subject from 1973-1979) and his associates.

SMED (1962, p. 41) made a detailed glaciomorphological analysis of the Funen islandgroup (supported by an equally detailed survey of the pertinent Danish litterature). He suggests that his Vissenbjerg area (Fig. 1) might even predate the NE Ice Advance. This tentative opinion is based i.a. on the already earlier known fact (MILTHERS. 1932) that this region shows an exceptional richness in Norwegian indicator boulders. This would implicate that the Vissenbjerg area represents a kineto-stratigraphic window in the sense of BERTHELSEN (1973): overridden but preserved by the younger NE Ice.

Even when the Vissenbjerg area chronostratigraphically belongs to the NE Ice Advance (a likely case) and not to any of the older main advances, then still it would form part of the extensive Central Funen region which remained unglacierized during the youngest readvance.

According to Hansen, Smed and their predecessors, the desintegrating ice masses, left behind by the ablating NE Ice, prevented the Young Baltic Advance to reach Funen's centre (i.e. the present area N of the borderline).

In this concept the Central Funen dead-ice from the NE Ice Advance should have survived the Asnæs interstadial which BERTHELSEN (1975) has proposed for southeastern Denmark. Whether or not this concept is right, it can be observed in the Vissenbjerg area that the higher parts of the ice-lake deposits (as exposed in excavated plateauhills) show precious little signs of glacitectonic disturbance or coverage by basal till.

As a third criterion for a non-destructive glacier transgression BERTHELSEN (1979, p. 126) mentions the presence of scattered erratics on the surface. Indeed it happens that the dark, humus-containing mould-layer holds more Nordic blocks than the subjacent unweathered beds which may be entirely devoid of them (e.g. in the case of meltwater clay). Yet, the significance of this is doubtful, since the mould-layer results from a complex of mostly non-glacial processes like solifluction, washing, pedogenesis, bioturbation and tillage

THE SEDIMENTARY SEQUENCE

The sediments at Ormehøj are intruded by and intercalated with diapiric clays. Since this diapirism is an interesting phenomenon more attention will be paid to it in another paper (SCHWAN ET AL., 1980). Here, we will neglect the influence of diapiric activity in order to show more clearly the essentially simple sequence.

Since no generally accepted lithostratigraphic nomenclature for the sediments under study is available, we will use informal names, too. There are four lithostratigraphic units to be distinguished, which are from top to bottom:

(4) Upper stratified beds;
(3) Boulder bed;
(2) Lower stratified beds;
(1) Basal till.

The distribution of these four units is shown schematically in figure 3, where it can be seen that the Ormehøj exposure comprises two separate areas: a large southern sandpit (with a mean depth of 10 m) and a smaller northern pit (with a depth of only 2.5 m).

Unit 1: the Basal till

The poor exposure (only at sites E1, E2 and F: see Figs. 3 and 4) and the moderate to strong affection by diapirism makes the establishing of the stratigraphic position tentative. The till underlies stratified silts (unit 2?) at site F and the Boulder bed (unit 3) at site E2 (Fig. 4). In the latter case this is thought to be the result of temporary non-deposition (of unit 2) as a result of local uplift by diapirism.

The Basal till is a bluish to light-brownish diamicton with stiff and plastic consistency. Its matrix (i.e. the fraction smaller than 2 mm) should be called a clayloam (SOIL SURVEY STAFF,

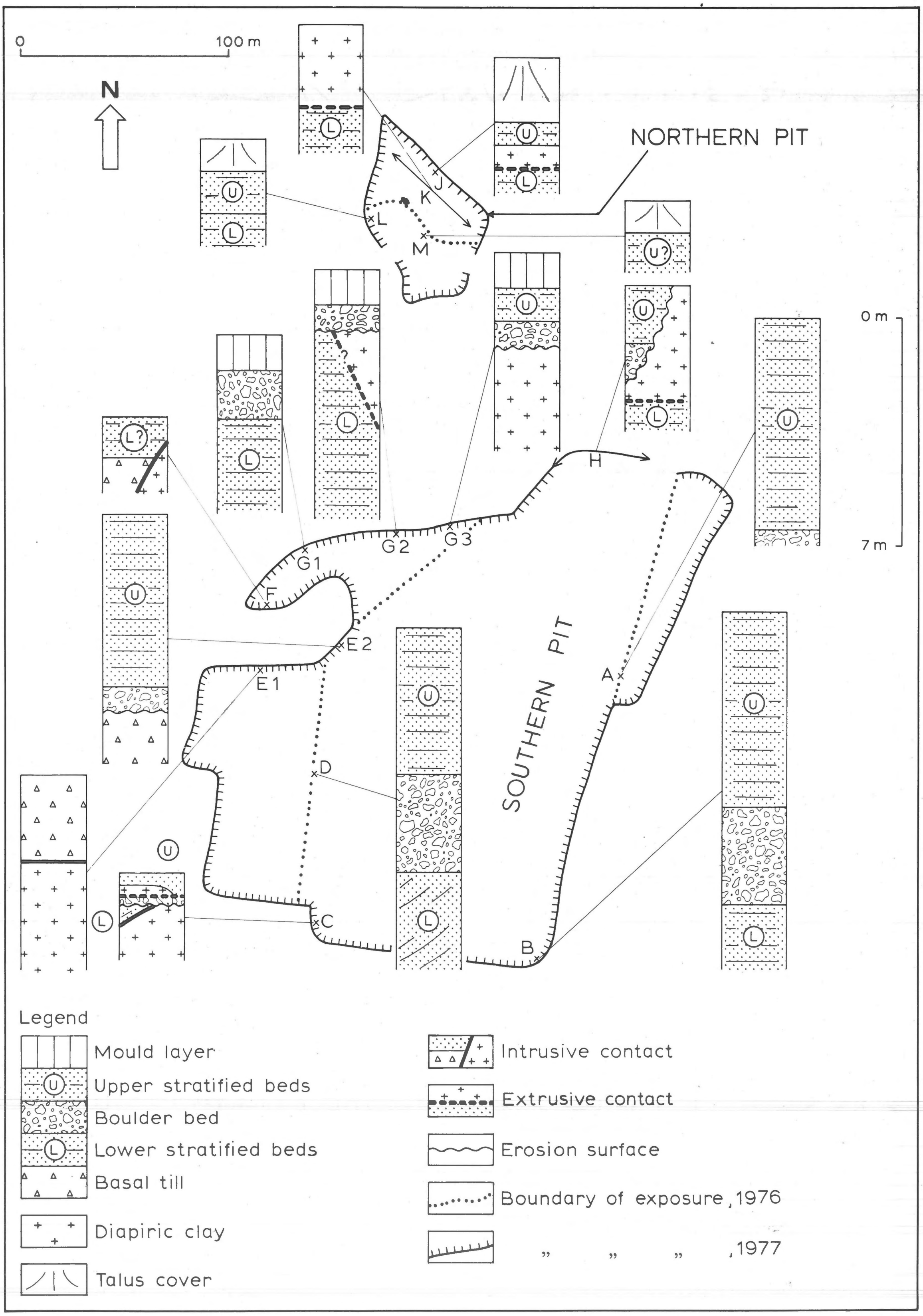

Fig. 3
Plan of the northern and southern pits at Ormehøj, indicating well-exposed sites with schematic profiles.

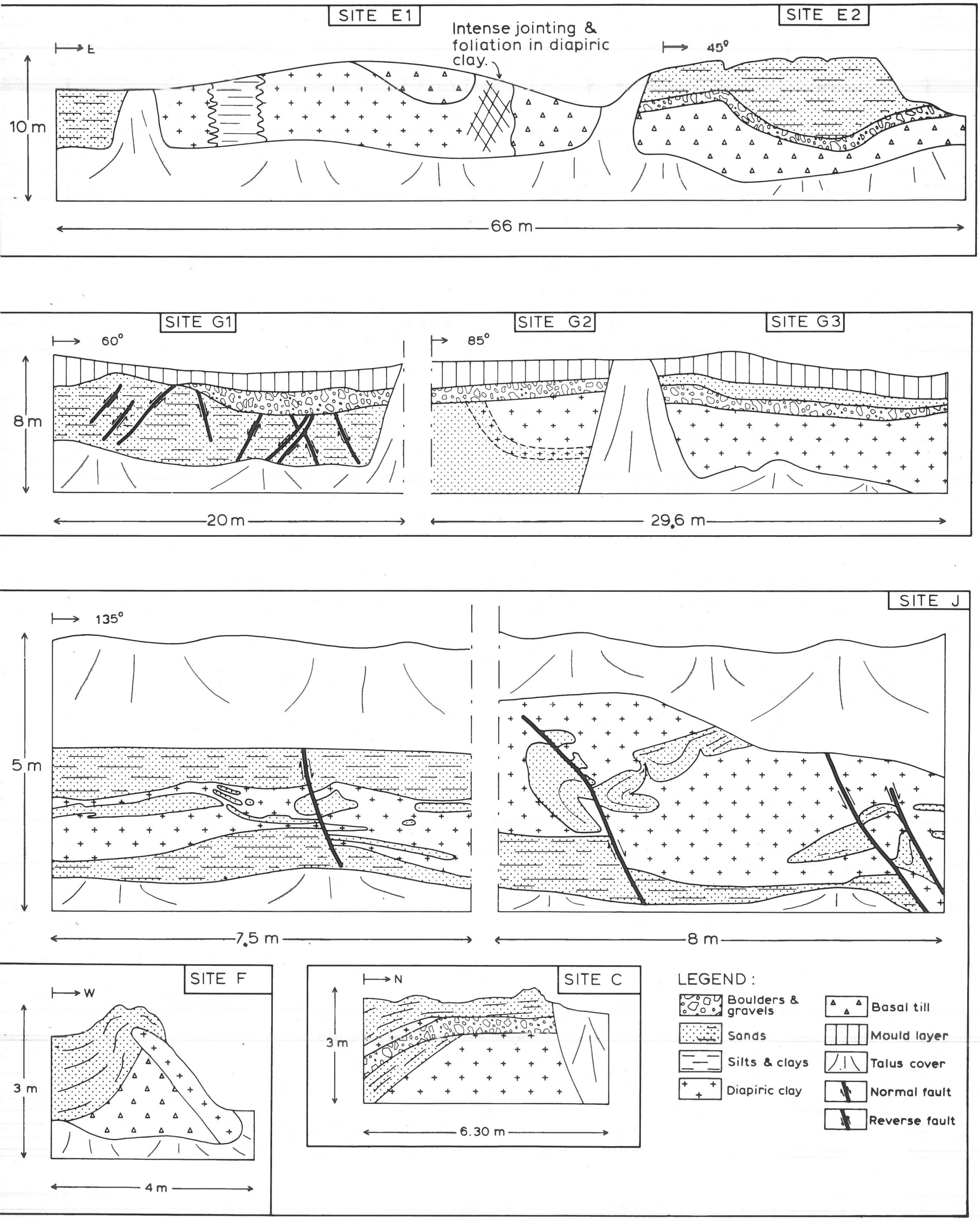

Fig. 4
Details of various sites. For location see figure 3.

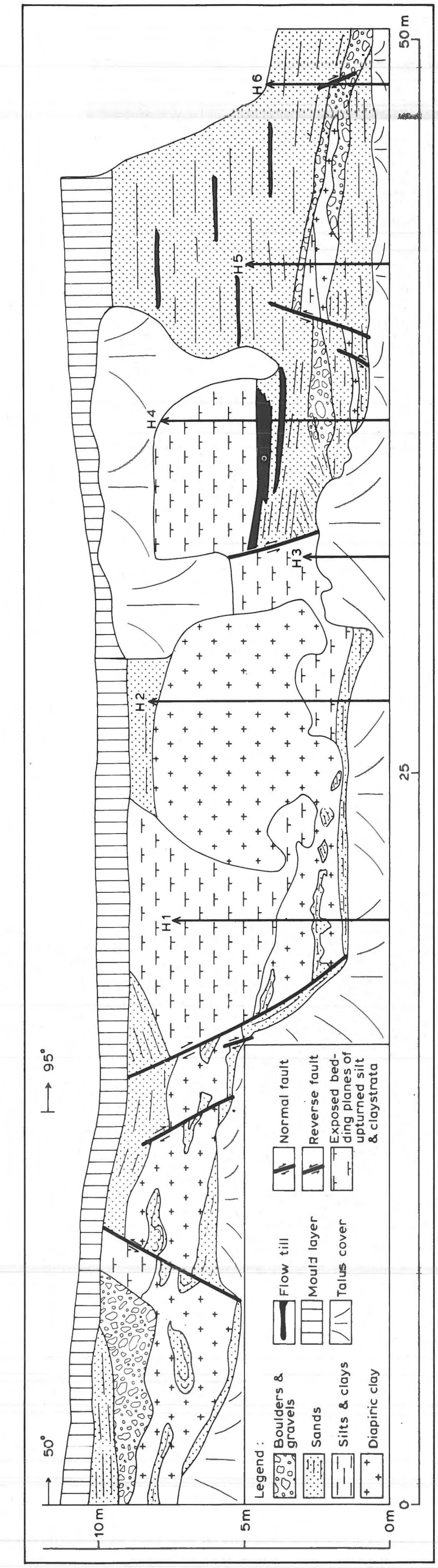

Fig. 5
Detailed section of site H. For location see figure 3. The vertical arrows indicate positions of perpendicular cross sections H1-H6 (see Schwan et al., 1980, Fig. 8).

1951).

Based on both own experience elsewhere (SCHWAN & VAN LOON, 1979) and on information given by MILTHERS (1940) and SMED (1962) this sediment is interpreted as a lodgement till. The poor exposure and the disturbed nature, however, make an application of the more subtle genetic criteria (like those proposed by KRÜGER, 1979) impossible. Support to this interpretation is given by SMED (1962) who states that this till is thought to underlie all the glaciolacustrine sediments in the Vissenbjerg area; he also mentions the very rough palaeosurface topography of this unit. This latter feature may, however, not only be the result of the mode of deposition, but also be influenced by subsequent diapiric activity.

Unit 2: Lower stratified beds

In most of the southern pit stratified beds can be distinguished both underlying and overlying the Boulder bed (Figs. 3-6). In the northern pit the Boulder bed (unit 3) is absent, but still the bipartition into two stratified units can be made. This may be due either to intercalated diapiric clay strata which conform the rule of superposition (SCHWAN ET AL., 1980) or to rather abrupt changes in lithology (Fig. 7). Only in a few cases (e.g. at site M, Fig. 12) the distinction may be impossible. Nevertheless it seems justified to consider these two stratified beds as separate units.

Within the Lower stratified beds four facies can be distinguished, which are described as facies LSB-1 to LSB-4. As will

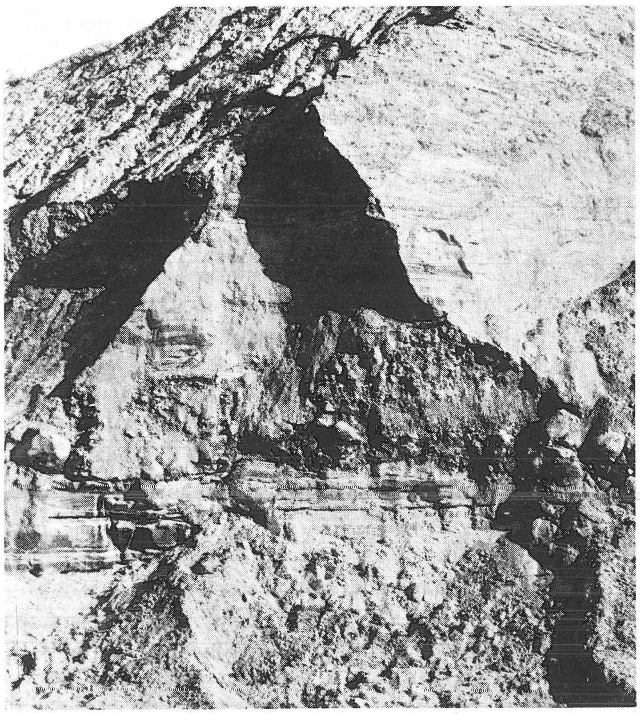

Fig. 6
Upper and Lower stratified beds, both represented by finely laminated silts and silty clayloams, with the Boulder bed in between. Thickness of the profile approx. 1.75 m.

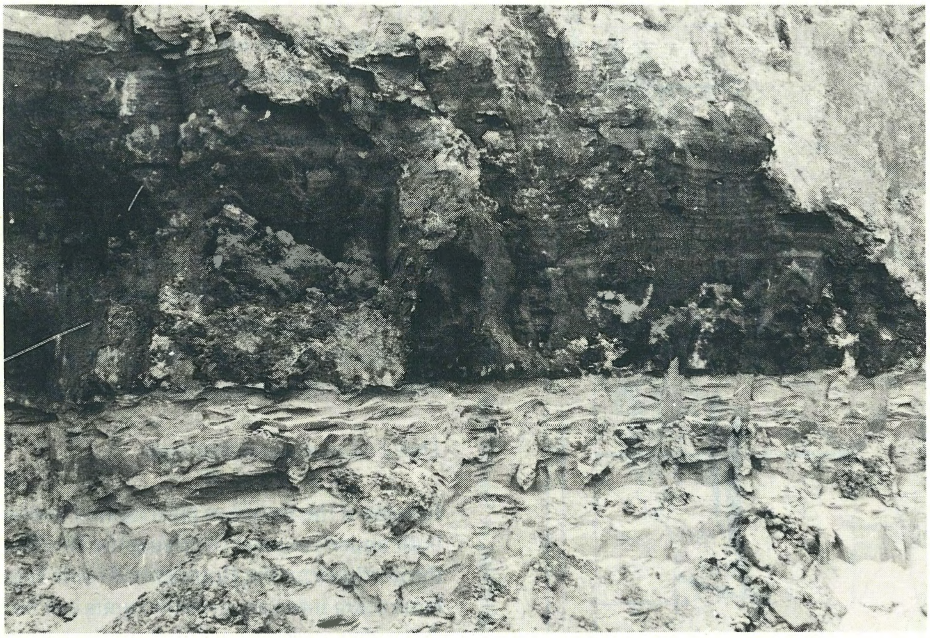

Fig. 7
Stratified beds in the northern pit. Probably the light-coloured sands with climbing ripples belong to the Lower stratified beds, while the overlying even-laminated dark clayloam represents the Upper stratified beds. Thickness of the profile approx. 2.30 m.

be shown, these four facies are highly cogenetic in that they typify various parts of an environment in which meltwater distributaries debouch into an intraglacial lake.

Facies LSB-1 – In this facies horizontally bedded or massive loam, sand and gravel strata prevail. Frequently abrupt changes in grain size occur. The coarser a layer is, the poorer the sorting and the bedding; this must be due to rapid, collective deposition under high-energy conditions. So this facies may represent lateral-accretion deposits of shifting distributary channels in a glaciofluvial environment (cf. SHAW, 1975). This facies is found along the eastern face of the southern pit (e.g. site B, see Fig. 3).

Facies LSB-2 – The second facies is at least 3 m thick and consists of sands with foresets dipping some 20° (Fig. 8). This fits quite well in the picture of a Gilbert-type delta which has been built out in an intraglacial lake by shifting meltwater streams. The absence of the topsets of this small delta may be due to truncation by the overlying coarse Boulder bed. This facies was found near site D in the western face of the southern pit as it existed in 1976.

Facies LSB-3 – Silt, sand and fine gravel are predominant in this facies, in which a fining-upward trend is visible. Towards the top the beds become thinner. In the lower part graded bedding is present, but winter-clay layers (as could be expected with varves) seem to lack. The upper part primarily is characterized by ripple-foreset crosslaminae and complete rippleform laminae (due to climbing ripples: Fig. 9). This facies can be found in the northern and western faces of the southern pit.

It therefore looks that the low competency was decreasing and/or that the depth of flow was diminishing. The graded beds suggest deposition by density currents, a rather common phenomenon in this type of deposits (see, e.g. SHAW & ARCHER, 1978). Such characteristics indicate sedimentation on a glaciolacustrine prodelta slope (GUSTAVSON, 1975; GUSTAVSON ET AL., 1975; SHAW, 1975).

Fig. 8
Foreset beds of a deltaic deposit in the Lower stratified beds. Mean dip value is approx. 20°. View to the north.

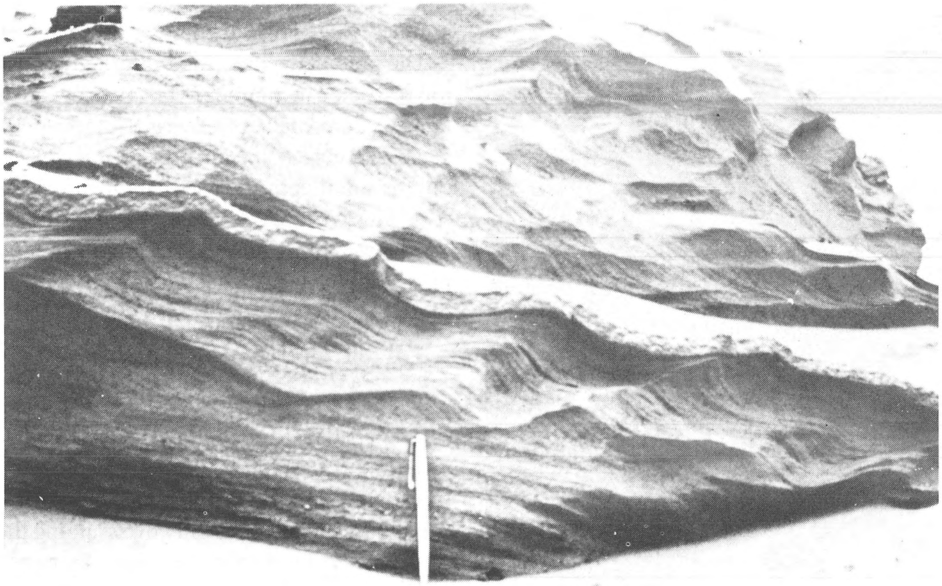

Fig. 9
Climbing ripples in sands of the Lower stratified beds. Ripple-foreset cross-laminae near pencil; complete rippleform laminae with low angle of climb directly above pencil. Terminology after Hunter (1977).

Facies LSB-4 – Fine sand to silty clay (loam) is characteristic for this facies which is present in the northern pit. Sedimentary structures are rippleform laminae, draped laminae (GUSTAVSON ET AL., 1975) and graded bedding. Grading is present in sandy beds of more than 15 cm thick, but also in thin silty to clayey rhythmites of about one cm each. In these rhythmites bounding strata of clay (which may settle in winter from suspension) are not distinctly visible.

The characteristics of this facies indicate primarily deposition by density currents and –to a lesser degree– by settling from suspension. The stream power was rather low. Most probably the sedimentation took place (again) on a glaciolacustrine prodelta slope: the rhythmites may have been deposited in front of it in the proximal part of the lake bottom.

Unit 3: the Boulder bed

In the northern pit the Boulder bed is absent, but in the southern pit it can be found between the two stratified units, reaching a maximum thickness of 3 m (between sites B and D). Towards the N and W this unit thins, which may be the result of contemporaneous diapiric uplift. This local uplift may also be responsible for the relative strong erosion, e.g. at sites C and G3 (Fig. 10).

There is a skeleton of boulders mutually supporting each other with interstices filled with gravel and sand; so it should be named a coarse-grained framework gravel (RUST, 1975; SAUNDERSON, 1975). In its upstream part boulders of up to 1 m are present, but both the maximum and the mean size diminish towards the downstream direction (300°[4]).

[4] All orientations are given in degrees, starting from the North, and going clockwise.

The characteristics indicate fluvial transport. The size of the blocks and the often considerable erosion indicate sedimentation during a flooding stage of exceptional magnitude. One might imagine the rapid and catastrophic bursting of a supraglacial lake, flowing out into the intraglacial lake. When such a temporary waterbody is perched on top of the ice that surrounds the lower level intraglacial lake, the necessary conditions for catastrophic flooding are fulfilled.

Unit 4: Upper stratified beds

The spatial distribution of this unit is more or less similar to that of unit 2. Granular composition and sedimentary structures are different, however; they allow the recognition of two facies types (USB-1 and USB-2).

Facies USB-1 – Horizontally bedded strata of loam, sand and pebbly gravel are characteristic, thus resembling facies LSB-1 except for a higher clay content. Besides, there is a grading-upward tendency, combined with gradually poorer sorting. In the upper part some layers of flow till alternate and interfinger with the parallel-bedded glaciofluvial strata (Fig. 11). This facies is found in the southern pit.

It seems that infilling of the intraglacial lake proceeded rapidly under high stream-power conditions, until at the end an attenuation of the flow regime took place (otherwise the fragile flow tills would have been eroded).

Facies USB-2 – This facies, which can be found in the northern pit, virtually is identical with facies LSB-4. Characteristic is site M, some details of which will be discussed (see Fig. 12 and next section sub small-scale structures). The succession of

Fig. 10
Erosional contact between light-coloured diapiric clay and overlying Boulder bed at site G3. Thickness of profile approx. 7 m.

figure 12 is 120 cm thick and comprises graded sands (layers B and G), fine sands to silts with mainly complete rippleform lamination (layers D and E) or draped lamination (layers A and E, upper part), and plane-bedded graded rhythmites of silt and clayey loam (layer C). Layer F, contorted by load casting, is silty and graded lower down, but horizontally bedded and sandy in its upper part.

The relative chronology of the sedimentary, deformational and morphogenic events at Ormehøj-exposure is tentatively summarized by SCHWAN ET AL., 1980 (their Table I).

SOME SEDIMENTOLOGICAL FEATURES

Although the first aim of this study was to establish the sequence and to recognize the various facies, some attention has also been paid to other sedimentological aspects. Interesting features appeared to be the palaeocurrent data, the thickness of the Boulder bed, and various small-scale structures in the Upper stratified beds.

Palaeocurrents

Palaeocurrent directions as measured in the exposure were entirely in the nature of scattered point-estimates. The six readings collected in the Lower stratified beds all fluctuate around 270°, i.e. palaeoflow from East to West. Included in this set is the average of twelve measurements in the dipping foreset beds of site D. Two Boulder bed-sites produced mean palaeocurrent directions of 310° and 300°. These values were averaged from readings on imbrications, a-axes of prolate clasts and scour-and-fill crossbedding structures in the fine-grained framework of the boulders.

The obtained values look internally consistent but they are too restricted in number to be taken as conclusive evidence for unidirectional palaeoflow. As the Ormehøj-exposure is only a small section of an ancient ice-lake with roughly circular outline unidirectional flow can only have occurred on a local scale unless some outlet had existed which kept the water at a fixed level. In the absence of any further evidence for such a natural spillway, a radial pattern of inflow for the former lake in its entirety is far more probable.

Thickness of the Boulder bed

The coarse texture of the Boulder bed facilitates a quick recognition at isolated places. The absolute altitude fluctuates considerably, as does its thickness. Both facts can be explained satisfactorily only by assuming that the lake bottom was affected by diapiric activity. Apparently a more or less continuous process of diapirism manifested itself in the form of local upheaval. When the catastrophic deposition of the Boulder bed began, the lake bottom probably had an irregular topography which was only incompletely levelled due to the short duration of this sedimentation process.

A systematic trend in the thickness or upheaval was not really found. Only a rough domeshape could be deduced from the isopachs in the western face of the southern pit (see SCHWAN ET AL., 1980, Fig. 2).

Fig. 11
Intercalation of flow till in glaciofluvial sands of Upper stratified beds at site A.

Small-scale structures

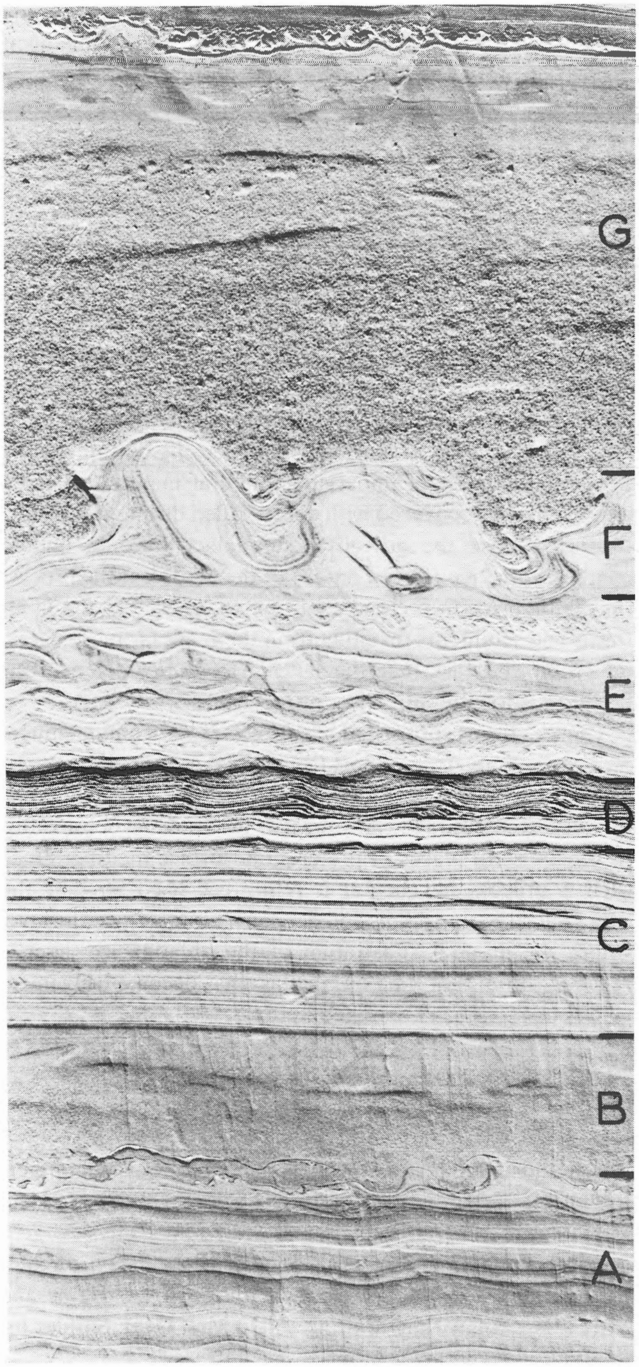

Fig. 12
Profile (approx. 122 cm high) at site M. Sedimentary structures explained in text. Lacquer peel by S. Bijlsma.

Well-exposed parts of the northern pit showed beautiful sedimentary structures, especially in facies USB-2 of the Upper stratified beds. Some of them are shown in figure 12 and will be commented upon here.

(1) Load structures at the base of units B and G have been tilted systematically by fluid drag and have resulted in well-developed flame structures in between. Also attributable to fluid shear stress are the roll-up structures (GUSTAVSON ET AL., 1975) at the base of unit B.

(2) A distinct erosion surface occurs at the top of unit E, as is evidenced by the truncated small-scale load structures. Since no traces of local scouring are found, the truncation might be due to the action of wind-induced waves. It cannot be completely excluded, however, that the levelling was caused by the stream from which the graded part of layer F settled later.

(3) Both the deformation of the flame structures and the lee side of current ripples show a consistent flow towards the West. This is in accordance with other palaeocurrent directions (see above). Apart from a few minor ripples only layer D shows an almost completely reversed current direction. As has been explained by SAUNDERSON (1975) this is not uncommon for meltwater streams debouching into glacial lakes.

(4) The poor sorting and the graded nature of layer G make it probable that it formed by settling from a density current. This glaciolacustrine turbidite is not the only phenomenon indicative of this type of mass movement. Moreover, the environment (sediment-laden meltwater streams debouching in the lake) seems favourable for this process. One might wonder why various other characteristics of turbidites are lacking in this layer, but as is known from flysch areas, fine-grained material is not very suitable for the development of the classical intervals (BOUMA, 1962) of turbidites (e.g. VAN LOON, 1972).

(5) An interesting phenomenon is shown by the sequence from layer B up to layer E. Layer B consists of graded sands and shows load casting at its base. Layer C shows a grain size which is more or less similar to that of layer B, whilst the entire layer consists of parallel laminae (only one wedges out, right part of photo). Layer D starts with the same grain size, suddenly becomes coarser, and then rapidly decreases again to the 'original' grain size; the sudden coarsening in layer D is accompanied by the already mentioned reversed current direction. In layer E the current direction is 'normal' again as shown by the ripple-drift cross-lamination; the grain size still gradually decreases and ends with a mean size of about silt. The entire sequence (layer B to E) closely resembles the Bouma intervals for one turbidite and we indeed assume that the material was deposited in this way. The suddenly reversed current and the coarsening in layer D might be caused by the mixture of two density currents with opposed flow directions as one of them flowed out from the facing lake border. Both currents should be characterized by coarse material, high stream power and short duration. The shortly reversed flow retook its original direction when the power of the 'opposite'

density current was brought to a stand by the overwhelming mass of the 'normal' density current. This leads to a mixing of the two currents as would happen with currents from the same source but having different flow velocities. In such a case one might speak of a *double-source turbidite.*

SUPRAGLACIAL AND INTRAGLACIAL

Since the usage of the adjectives 'supraglacial' and 'intraglacial' is not very consistent in litterature some clarification seems desirable. Intentionally, the ancient ice-lake we are concerned with in this text has been referred to as an intraglacial lake. By this we mean a large body of still water bordered all around by walls of stagnant ice. As to its bottom, this is thought to consist of impervious mineral material to some unknown extent covered with debris-laden disintegrating ice. In such a case the adjective 'supraglacial' would unduly emphasize the *ice-floored* nature of the lake. Yet the processes taking place in such an environment as well as the landforms and sediments resulting from them are far more determined by the fact that the lake is *ice-walled.* Hence our preference for the adjective 'intraglacial'.

The two terms contrasted here are as it were end members on a continuous scale. Apart from ice-lakes formed in marginal and subglacial positions, most other ice-lakes will start their life cycle supraglacially, be it as thermokarst depressions or otherwise in places of preferential surface melting. Once such a process has begun it reinforces itself so that the supraglacial basin widens and deepens: ultimately to the level where the mineral substratum of the ablating ice-sheet is reached. From that time on, the lake has become truly intraglacial in the sense envisaged here.

ACKNOWLEDGEMENTS

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