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STRUCTURAL AND SEDIMENTOLOGICAL CHARACTERISTICS OF A WEICHSELIAN KAME TERRACE AT SØNDERBY KLINT, FUNEN, DENMARK[1]

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ABSTRACT

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At a coastal cliff near Sønderby, SW Funen, the internal structure of a Weichselian kame terrace is well exposed. It exhibits, though in a rather deformed state, a sequence of lodgement till overlain by stratified glaciofluvial beds plus a capping of flow till. An elongated ridge, interpreted as an intraglacial crevasse-infilling, is perched right on top of the kame terrace, whereas laterally the terrace body passes into a small sandur.

The kame terrace and the overlying kame ridge are interpreted to be the result of two subsequent oscillations of the Belt glacierization stage.

The effect of static load diapirism is demonstrated and the presence of slump structures in a seemingly anomalous spatial attitude is explained.

By combining geomorphological, sedimentological and tectonic data, the depositional and deformational history of the kame terrace and associated landforms is reconstructed.

INTRODUCTION

In the southwestern part of the island of Funen the Sønderby cliff extends along the Little Belt. Although it has not the impressive size of classical cliff sites such as Røgle Klint and Ristinge Klint (both on the Funen island group, too), this cliff is exposed over a considerable length (500 m) and height (up to 25 m). For an exact location reference is made to figures 1 and 2.

According to Danish sources the sediments exposed in the cliff must date from the Belt stage which should be one of the last phases of the Weichselian glaciation (SMED, 1962; HANSEN, 1965). Although in Denmark the various stages of the Weichselian glaciation are still a matter of debate (ANDERSEN, 1963; BERTHELSEN, 1973; SJÖRRING, 1977; PETERSEN, 1978), it is clear that the Belt stage with glacier tongues moving into the two Belts can hardly be questioned. As is the case in the plains of northern Europe, the reconstruction of the glacial history of the Danish archipelago is based to a considerable extent on glaciomorphological evidence. Of course, other data have

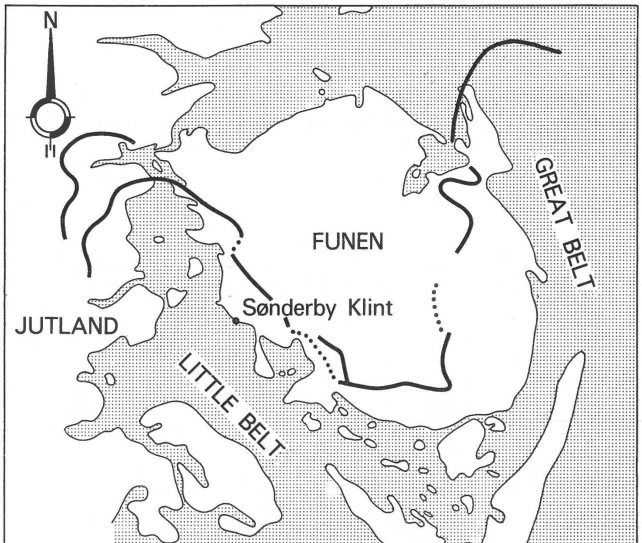

Fig. 1
Location of Sønderby Klint. The borderline of the Belt-stage glacierization is indicated by a solid line where confirmed, and with a dotted line where hypothetical. After Hansen (1965).

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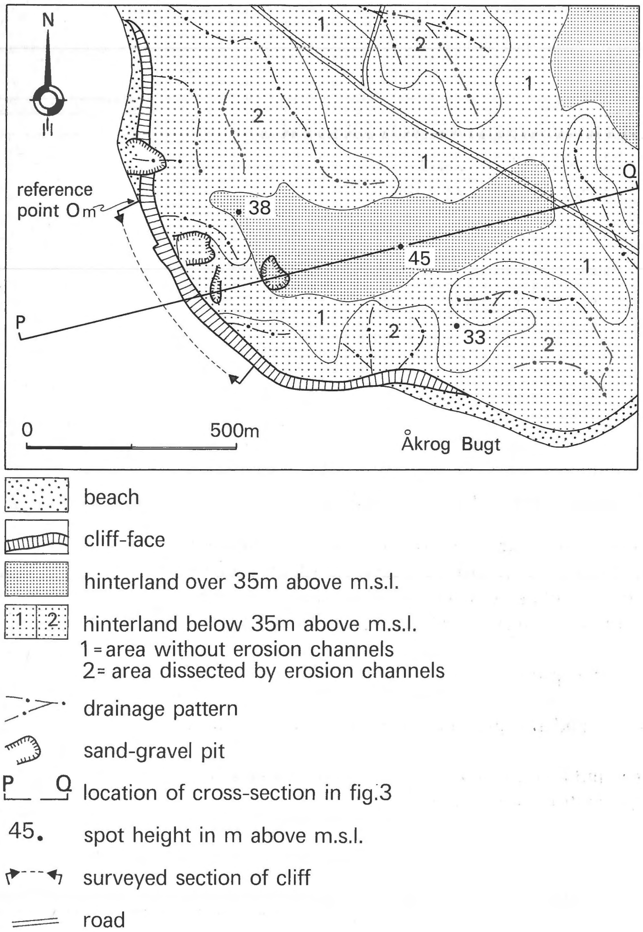


Fig. 2
Geomorphological sketch map of Sønderby Klint and its hinterland. The line P-Q is the location of the cross section in figure 3.

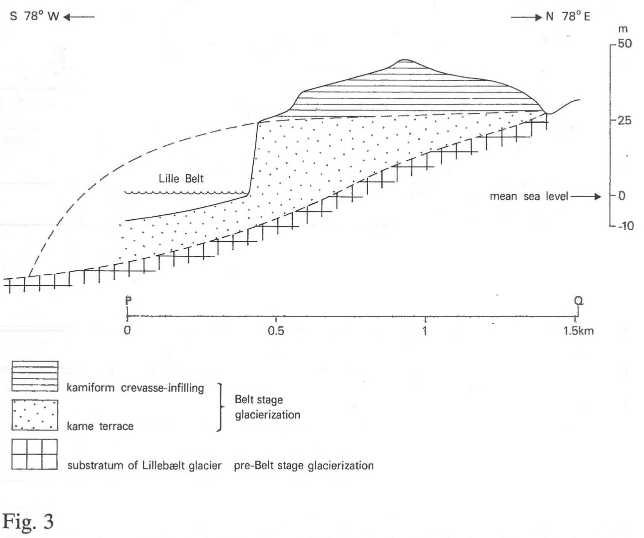


Fig. 3
Cross section of the Sønderby cliff (see figure 2 for location). The broken line shows a conjectured reconstruction of the kame terrace and associated landforms prior to the recession of the cliff.

also been used: especially the analysis of erratic boulders and glacitectonic structures. It remains true, however, that the Danish concept of the Weichselian glaciation bears a strong geomorphological imprint. In this concept at least three regional ice advances in Denmark are recognized during the last ice age plus a somewhat confusing number of minor oscillations which have only local significance.

In our opinion there can be little doubt about the activities of a glacier tongue which penetrated the valley of the present Little Belt, moving into a northerly direction (Lillebælt glacier). The landscape in SW Funen bears its clear imprints. Therefore we assume that the sediments of Sønderby Klint were deposited during the Belt stage of the Weichselian glaciation. Yet, the stratigraphical position of the deposits is not the main concern of this paper. Primary attention has been directed towards both the way of sedimentation and to the subsequent deformation of the deposits.

DEPOSITIONAL ENVIRONMENT

We basically agree with SMED (1962) who – for geomorphological reasons – defined the sediment body at Sønderby Klint as an ice-marginal kame. If it is true that a glacier tongue moved into the former valley of the Little Belt, the presence (during a stagnant-ice phase) of kamiform sediment bodies can be expected in the lateral zone of the valley. In glacial times the sedimentary surface along the lateral margin of the Lillebælt glacier was inclined, and deposits formed here in a stagnant-ice environment can most appropriately be called a kame terrace. It should be kept in mind, however, that Sønderby Klint is probably only a last remnant of an originally larger body which has been eroded away over an unknown distance since the postglacial sea-level rise. Even a few days of rough weather (with a sea level higher than normal) are sufficient to demonstrate that recession of the cliff still goes on.

As is shown in figure 2, an elongated hill occurs at a small distance inland from the cliff. This hill overlies the highest part of the cliff. It contains only one abandoned gravel pit, heavily overgrown with scrub; as far as could be detected, the hill consists entirely of undisturbed, well-stratified sands and gravels. It is therefore assumed that this hill represents yet another kame deposit on top of the one exposed in the cliff. The elongated shape of the hill, the orientation of its long axis roughly perpendicular to the direction of the glacier's movement, its location right on top of the cliff's highest part, and finally its internal structure seem to suggest a kame-like crevasse-infilling.

If this interpretation is correct, the picture arises of a kame terrace overlain by a kamiform crevasse-infilling (tentatively reconstructed in Fig. 3).

In the cliff two till beds are present, separated by several metres of glaciofluvial sand and gravel. As will be discussed

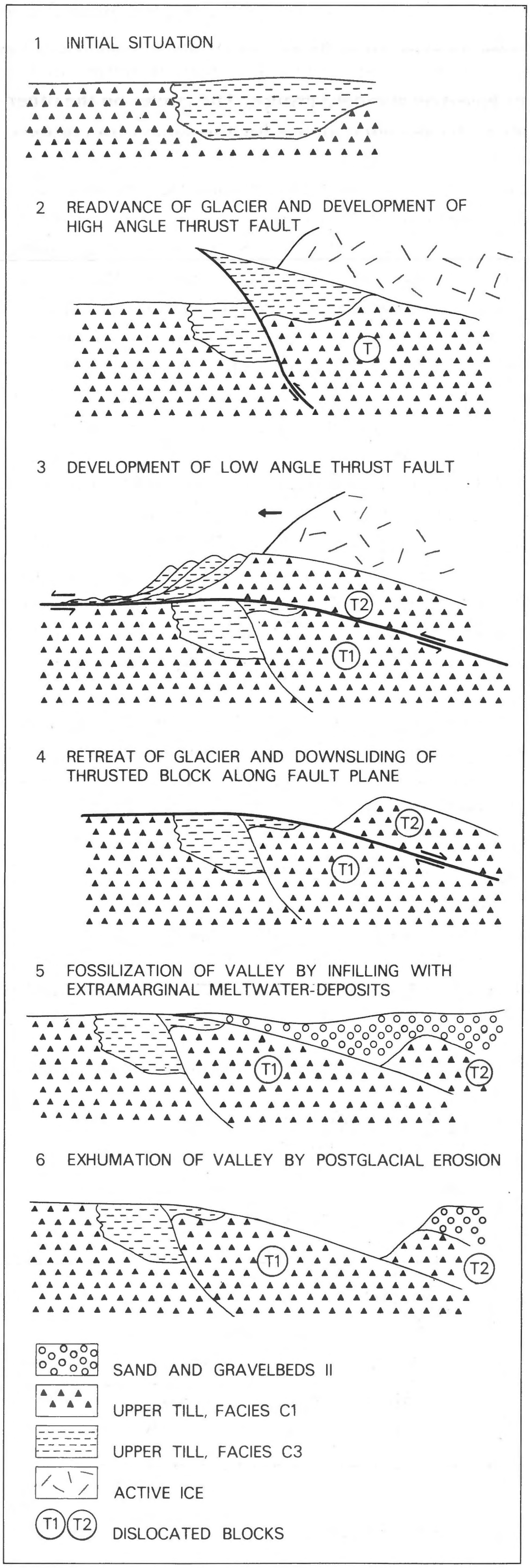


Fig. 4
Genesis of the sliding plane and assymmetrical valley near r.p. 490.

below, we interpret the lower till as a lodgement till and the upper one as a flow till. This means that the latter is supposed to be deposited in a supraglacial and subaerial environment during a waning stage of the Lillebælt glacier. The kame ridge (crevasse infilling) overlying the cliff sequence must therefore have been deposited during a renewed advance of the glacier. This implies that in this cliff two advance stages of the glacier can be recognized.

Work done elsewhere in western Funen has supported the idea of two oscillations. A relevant site where indications are found for this hypothesis is called Hjerupgyde and lies some 14.8 km north of Sønderby Klint. In a forthcoming paper it will be dealt with separately.

In figures 5 and 16 it can be seen how at r.p. 490[4] the height of the cliff face decreases to approx. 5 m. Here the upper (flow) till disappears and is laterally replaced by horizontally bedded sands and gravels. The contact between the two deposits is absent because of a narrow and asymmetrical valley, the possible genesis of which is shown in figure 4.

Beyond the valley, the horizontally bedded sands and gravels overly the flow till; since they show a glaciofluvial character, they are probably associated with the younger oscillation of the Lillebælt glacier. The corresponding sedimentary body is thought to be deposited extramarginally and as such to form a minor sandur.

Lastly, attention is drawn to a number of erosional channels (see Fig. 2) which dissect the kame-terrace surface and the adjacent area. In a later chapter (the deformational structures) their role in the genesis of superficial slump structures along the cliff face will be explained.

The major geomorphological features of Sønderby Klint and hinterland are schematically shown in figure 5.

THE LITHOSTRATIGRAPHIC UNITS

In figure 16 the vertical and lateral relationships between the various sediments are shown in detail. Four lithostratigraphic units have been distinguished, viz.:
(4) Sand and gravel beds II (unit D)
(3) Upper till (unit C with facies C1, C2 and C3)
(2) Sand and gravel beds I (unit B)
(1) Basal till (unit A).

Basal till

This till (unit A) consists of a calcareous, bluish or light-brownish, very compact moraine clay with a generally low content of coarse clasts. Occasionally reworked fragments of the Eemian shell *Arctica islandica* (previously known as *Cyprina islandica*) are found, and this is considered characteristic for

[4] In the text we will use the abbreviation r.p. for 'reference point'. The number indicates the total distance (in m) along the cliff, referring to the point O in figure 16.

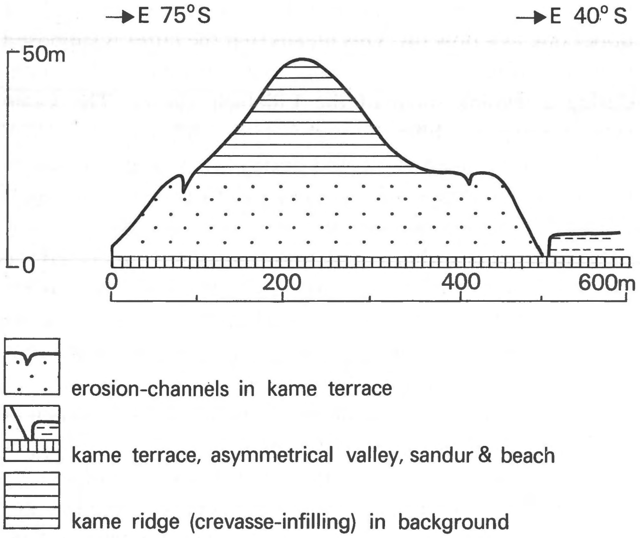


Fig. 5
Schematized frontal view of Sønderby Klint and its hinterland, showing the major geomorphological features.

glacial tills along the Little Belt coast (MADSEN ET AL., 1908; MILTHERS, 1959, p. 93). According to HANSEN (1965, p. 39) glacial sediments which have incorporated Eemian marine elements were all deposited in the younger stages of the Weichselian glaciation.

Boulder counts – MILTHERS (1942) published three indicator-boulder counts from the direct vicinity of Sønderby Klint. Although his counts do not refer specifically to our rock unit A (two counts in beach shingle which immediately borders the basal till; one count in a gravel pit which must overly the basal till), they give a rough impression of the petrological conditions at this site. Milther's results have been plotted in a ternary diagram (Fig. 6), in which each of the counts is shown twice: in crosses and in circles. The former represent assemblages based on standard groups. The circles on the other hand refer to recalculated values for the same observations. Aaland-indicators and 'unspecified Dalecarlian porphyries' are included in the standard ø-group (Baltic components) and in the standard s-group (Dalecarlian components), respectively.

As it is generally assumed that indicator assemblages along the Funen South coast should be dominated by Baltic components (MILTHERS, 1942, p. 133, 1959, p. 99), the standard ø-group (crosses in Fig. 6) is unexpectedly low. When the standard Baltic component and Aaland indicators are added up into one group (circles in Fig. 6), the assemblage acquires a more typically Baltic imprint. It is uncertain, however, whether these Aaland indicators may be considered as an additional element of the standard ø-group. For a critical treatment of Milthers' indicator-counting method reference is made to MARCUSSEN (1974) and OVERWEEL (1977). The latter author has given his own interpretation of Milthers' original porphyry-distribution map, and concludes that the Danish area can be divided into three NE-SW running zones, in each of which one of Milthers' standard groups should be preponderant.

In this (somewhat oversimplified) picture by Overweel the Funen West coast should be located right in the middle of the zone with predominance of Dalecarlian indicators, but Overweel admits that his Dalecarlian zone has a more mixed nature than the other two. Therefore, Milthers' Sønderby Klint counts fit neatly into current concepts about Danish indicator-boulder distribution: marked predominance of Dalecarlian porphyries on the basis of standard groups (Fig. 6, crosses), but with a distinct shift towards a purely Baltic composition when the so-called 'semiprecious' indicators are included (Fig. 6, circles).

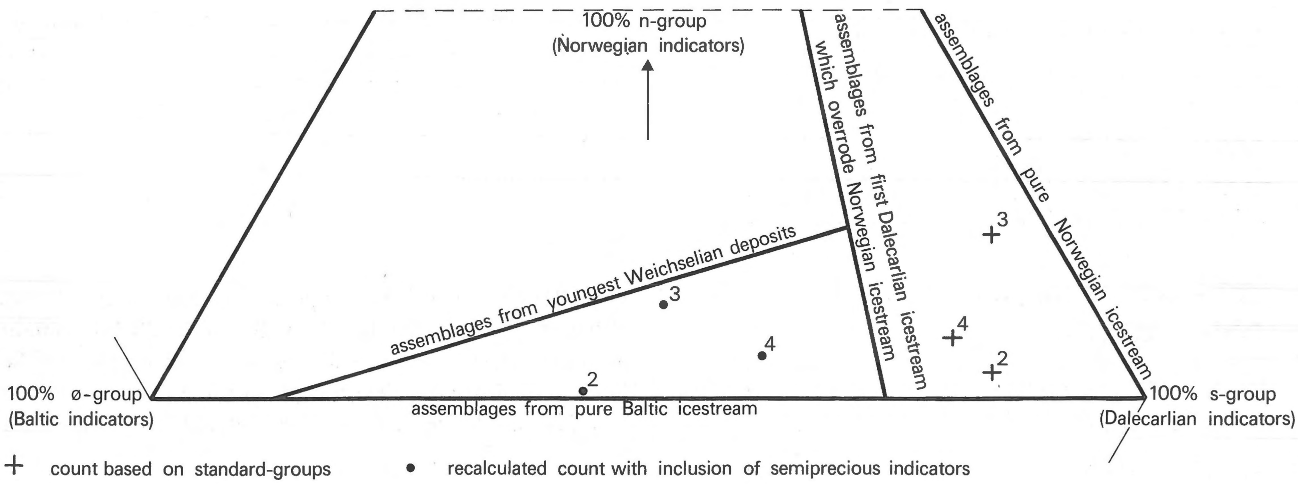


Fig. 6
Results of three indicator-boulder counts near Sønderby Klint.

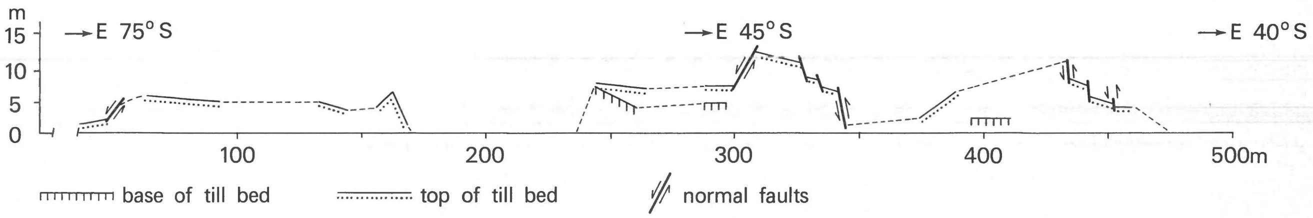


Fig. 7
Surface topography of the Basal till at Sønderby Klint.

The results of counts which cover both the Basal till and the overlying rock units are quite un-baltic. From 13 samples (fraction 5-8 mm; average sample counts 300 clasts) the mean Palaeozoic-limestone content was calculated to be 9.6% with a standard deviation of 5.5%.

Fabric – A till-fabric histogram (N = 48) in the Basal till shows a preferred a-axis orientation[5] of 150°-330°, though not a very strong one (circular variance = 0.2554). Therefore this result has only restricted significance, but a number of other structural measurements (to be presented further on) also support an ice movement with roughly N-S orientation.

Conclusions – In combining all observations (shells, petrology, fabric) it must be concluded that unit A was deposited as a lodgement till by a presumably Baltic ice stream, regionally moving towards the north, i.e. by the Belt-stage advance. This is in agreement with recent Danish work (SJÖRRING, 1977, p. 10; PETERSEN, 1978, p. 130). During or very shortly after its emplacement the till was subjected to considerable glacitectonic deformation. Since subglacial deposition and tectonism normally require mutually exclusive conditions (e.g. BOULTON, 1975), a postdepositional (but subglacial) deformation seems the most likely.

Additional evidence for this interpretation can be found in the cliff face:

(1) The top of the Basal till at Sønderby Klint rises to an unusually high level (the cliff owes its very existence to this height) and moreover its topography is quite irregular (Fig. 7). It could be argued that any lodgement till, even in its primary undeformed position, may have an irregular surface topography. Yet, it turns out that in both coastal cliffs and inland exposures, till bodies in W. Funen generally exhibit a very gently undulating or even flat and horizontal topography – with an exception only in places where ice push is evident. So there is hardly any doubt that glacitectonism is involved wherever a till body abruptly rises to an extraordinary height. Figure 7 shows how the till body has been reshaped into a system of two ridges, whereas in a more northwestern direction the till thins out to disappear over a distance of 65 m. In this trajectory, a floe of till has presumably been incorporated by overriding ice: a minor case of large-scale block inclusion (MORAN, 1971). The series of normal faults in the two ice-pushed till ridges suggest instability following the ultimate retreat of the glacier. In any case, no indications were found for faulting as a result of a collapse of buried dead ice.

(2) Between r.p. 405 and 450 (Fig. 16) the top of the till bed is intensely brecciated or mylonitized to use two very appropriate terms, proposed by LAVRUSHIN (1971). Figure 8 gives an impression of the highly complex meso-structures involved, but their interpretation is beyond the scope of this paper. To a lesser extent, comparable signs of brecciation are also found at the base of unit A.

It is striking that the sands and gravels of unit B by no means exhibit such a degree of tectonization, even where they immediately overly the brecciated parts. They form a subhorizontal sediment body, interspersed by a series of steeply dipping normal faults. This proves that glacitectonic deformation of the Basal till took place *before* the deposition of unit B.

(3) Since syndepositional and penecontemporaneous deformation took place, the till body may be treated as a tectonite with endiamict primary structures (BANHAM, 1975). Therefore, the attitudes of various planar structures (contact between till and overlying beds, shear planes, unconformities) have been measured. The result is depicted in figure 9A, where the girdle indicates ice push from the south towards the north.

This additional evidence (1-3) forms sufficient arguments to interpret unit A as a lodgement till, subglacially tectonized during or very shortly after its deposition by an ice lobe moving in a northerly direction. In the nomenclature of DREIMANIS (1976) it belongs to the class of deformation tills.

Sand and gravel beds I

The gravels, sands and occasional loams of unit B are nearly all well sorted and well stratified. Their characteristics suggest the following genesis:

(1) Deposition occurred in a stagnant-ice environment.

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

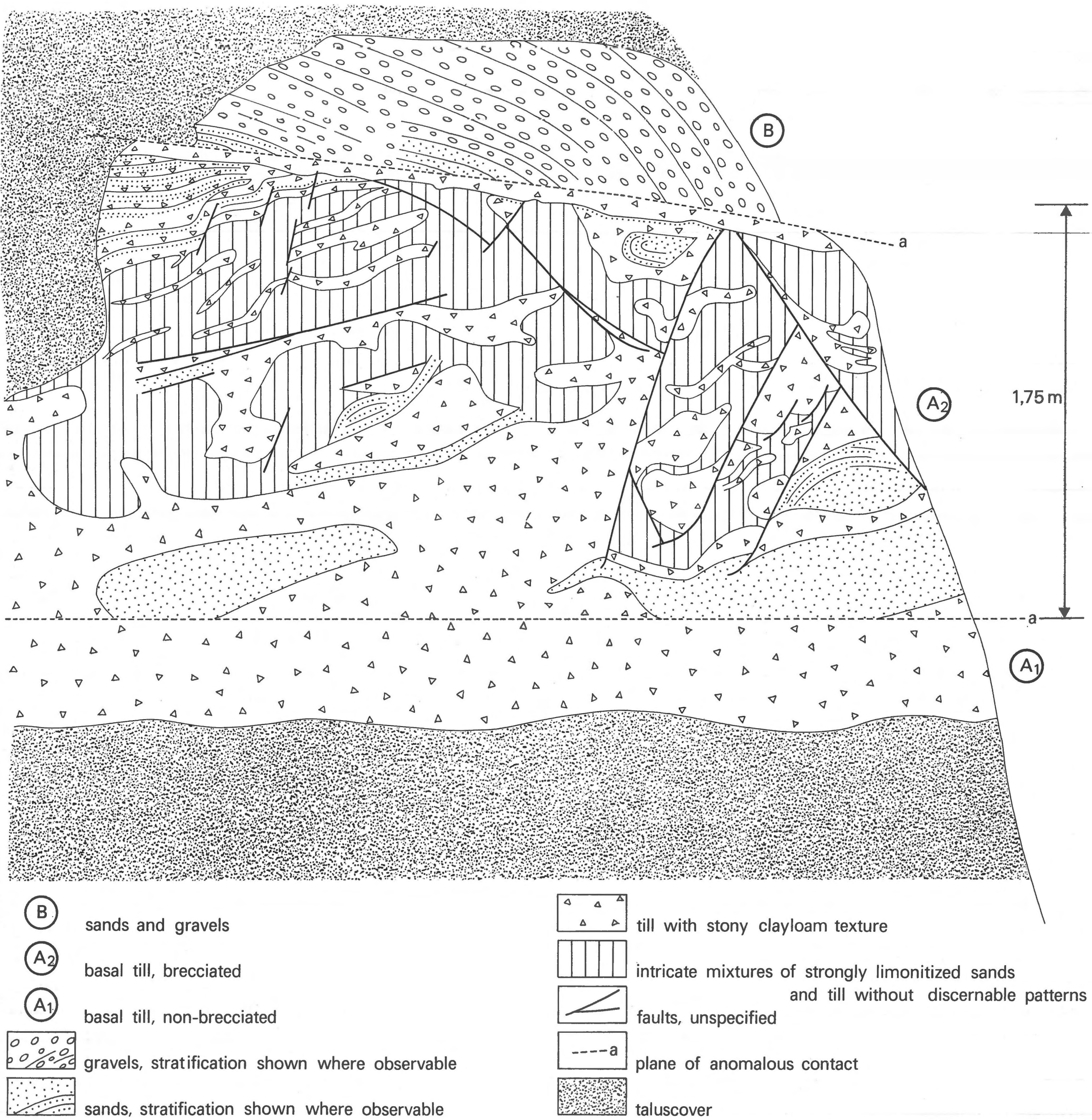


Fig. 8
Details of the brecciated till near r.p. 450.

Both vertically and horizontally the strata exhibit abrupt changes in grain-size distribution. This is to be expected under conditions of intraglacial or supraglacial sedimentation rather than in an environment of laterally unconfined aquatic deposition. Generally speaking, frequent and sharp vertical (and horizontal) discontinuities are likely to occur where hollows and crevasses in stagnant ice are filled up with sediment; in contrast, deposition of an unconfined sandur would produce more gradual and systematic transitions in the beds' lithology.

The base of the Upper till which overlies unit B closely approaches the top of unit A near both ends of the section. So unit B forms a huge lense-shaped body of sands and gravels trapped in between units A and C. This position is still another indication for sedimentation in a supra- or intraglacial basin, followed by reversion of the original relief due to melting of adjacent and/or buried ice. The margins of the former ice-walled or ice-floored basin are then represented by the places

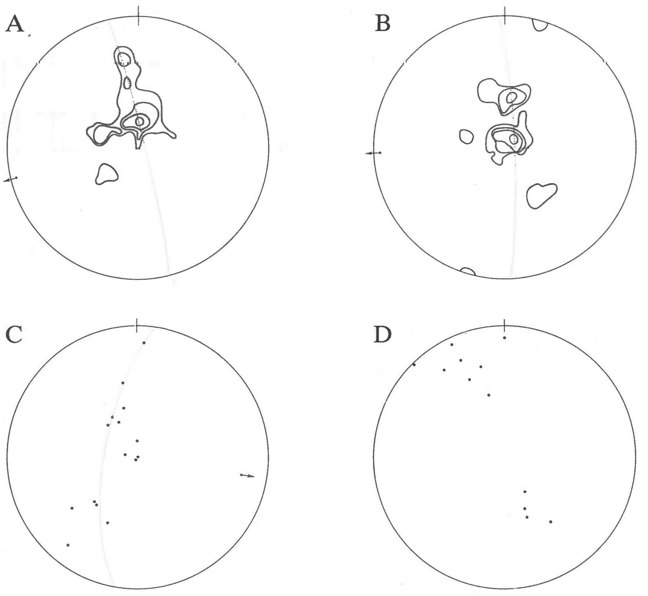


Fig. 9
A: contour diagram. Poles of planar structures in the Basal till. N = 31. Contours: 1.5, 2.5, 3.5 and 4.5 points per 1% area.
B: contour diagram. Poles of bedding planes in the Sand and gravel beds I. N = 31. Contours: 1.5, 2.5, 3.5 and 4.5 points per 1% area.
C: scatter diagram. Poles of sliding planes and other anomalous contacts in units A, B and C. N = 15. Rectified position of poles is based on mean tilt of the stratified beds.
D: scatter diagram. Poles of normal-fault planes in units A, B and C. N = 12.

where units A and C are in contact with each other.

(2) Deposition of unit B took place intraglacially in an ice-walled rather than in an ice-floored basin system. True collapse structures were not observed, although normal faults and folds do occur, but these features are not believed to be the result of underground deglaciation, i.e. of ice melting from beneath the cover of stratified sands and gravels. The faults extend downwards right into the Basal till (Fig. 7) and it seems very unlikely that the wasting ice mass could cause fracturing of its own substratum. Nor would the faults, in the case of buried ice, assume the shape of slightly concave but continuous planes undisturbedly reaching close to the cliff's top.

One more argument in favour of intraglacial rather than supraglacial sedimentation comes from the locally distinctly erosive contact between units A and B. In places the base of unit B even contains pebbles and boulders consisting of material from unit A (r.p. 170, Fig. 10). Apparently the erosion took place during the infilling of the ice-walled basin. In the case of supraglacial sedimentation and subsequent underground deglaciation such erosion is much less likely to occur.

The strata of unit B lie horizontal or gently dip over long distances; they are, however, tilted and folded between r.p. 335 and 365 (Fig. 16). This deformation is clearly associated with the normal faults below. Normal faulting and subsequent tilting and folding seem to have affected the sands and gravels especially where these overly the unstable highest parts of the Basal-till surface. As is suggested by the contour diagram (Fig. 9B) tilting and folding in unit B must be associated with ice movement along a 177°-357° axis.

In some places (r.p. 255, 420-450) the base of unit B is in anomalous contact with the subjacent sediment (Basal till). Probably these contacts have a tectonic origin, but their genesis is not precisely understood.

At r.p. 220 and 370, and also N of r.p. 0, there are some squeeze-up structures. These occur at the base of the cliff face as thin (about 0.5 m) subvertical plates with a slightly curved shield-like shape. The scutiform plates (Fig. 11) are in unconformable contact with the beds of unit B and consist of the same moraine clay as unit A. These plates were formed when water-saturated till was squeezed upwards into basal crevasses of unit B. By subsequent recession of the cliff face the 'shields' became exposed in their present position. Differential vertical pressures are the essential cause of the genetic process which is called 'subglacial squeeze-up', 'static load diapirism' (BANHAM, 1975) or 'subglacial till flowage' (SUGDEN & JOHN, 1976). It is believed that the necessary conditions for this process were satisfied at the end of the first glacierization stage (Table II). The coexistence of wasting ice masses, side by side with heaps of mineral sediment, may at that moment easily have produced wide-spread differential pressures. Moreover, deglacial conditions then had lasted long enough for the Basal till to attain an unfrozen and water-saturated state.

Upper till

In this unit C three facies (C1, C2 and C3) have been distinguished.

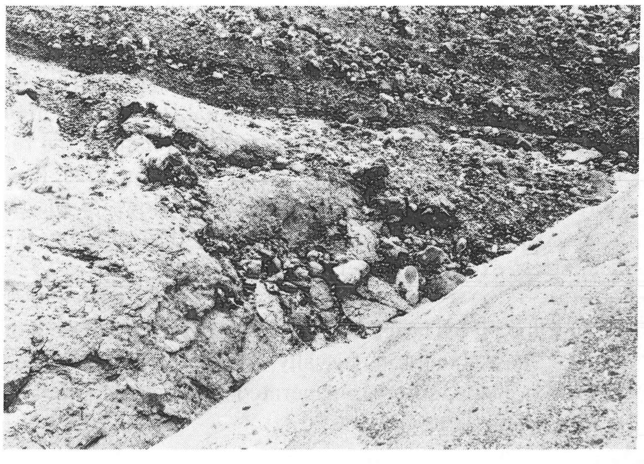


Fig. 10
Pebbles and boulders of moraine clay at the base of unit B near r.p. 170. The clasts are derived from the underlying Basal till (lower left).

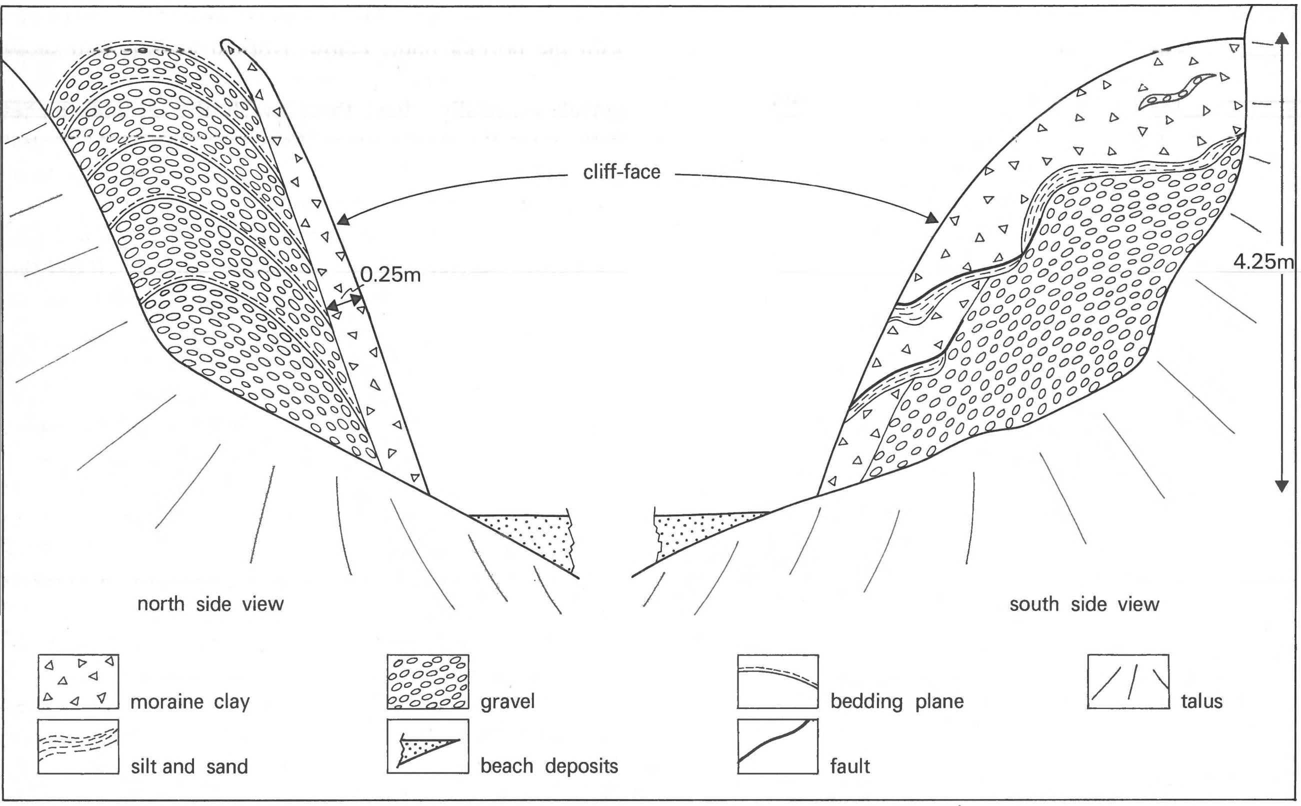


Fig. 11
Side views of squeeze-up structures near r.p. 220.

Facies C1 – This unit consists of a diamictic sediment with variable grain-size distribution. From r.p. 0-90 the unit is roughly built up by sandy loam with floating pebbles, while beyond r.p. 150 the facies becomes clayey instead of loamy. This finer-grained facies is compact and stiff, thus resembling unit A (Basal till), but in some other respects it is significantly different, showing the following features:

(1) Though essentially massive, both the coarse and the fine variety exhibit in several places an irregular and rather vague bedding, sometimes plane and horizontal, but more often with sedimentary folds. These structures never show the recumbent character described by EVENSON ET AL. (1977). Probably this layering results from successive flows of plastic mud creeping downslope, subsequently distorted by the shear stress exerted by later flows.

(2) In some places facies C1 shows isolated bodies of sand or gravel. These inclusions usually have an elongated and drawn-out shape, so creep-generated shear-stress may also be responsible here.

Both phenomena (1) and (2) are thought to be indicative of subaerial gravity flow in a plastic state and this is one reason why facies C1 is interpreted as a supraglacial flow till. As will be seen below, the characteristics of facies C2 and C3 yield another argument in favour of this interpretation.

Facies C2 – This laterally discontinuous unit overlies facies C1. It consists of irregularly stratified sandy loams and loamy sands which in general are poorly sorted. Bedding is sufficiently distinct to show a number of large open folds which are the most conspicuous characteristic of this sediment body. Apart from this undulating appearance, there are some erosional surfaces and small-scale contortions.

The lower boundary in places is quite irregular and texturally it passes gradually into the subjacent facies C1 (Fig. 12). Therefore a close relationship between both facies must be assumed.

The thickness of the beds, the numerous internal deformations (Fig. 13) and the lateral discontinuity (possibly determined by palaeotopographic lows in the substratum) all seem to point towards redeposition in a (semi-)plastic state. The resulting folds, which seem to be gravity-induced, are a response to continuous changes in topography of the ablating glacier. For these reasons facies C2 is interpreted as a flow till with higher mobility than its more massive and finer-grained counterpart C1. During the deformational process sedimentation by settling from currents continued.

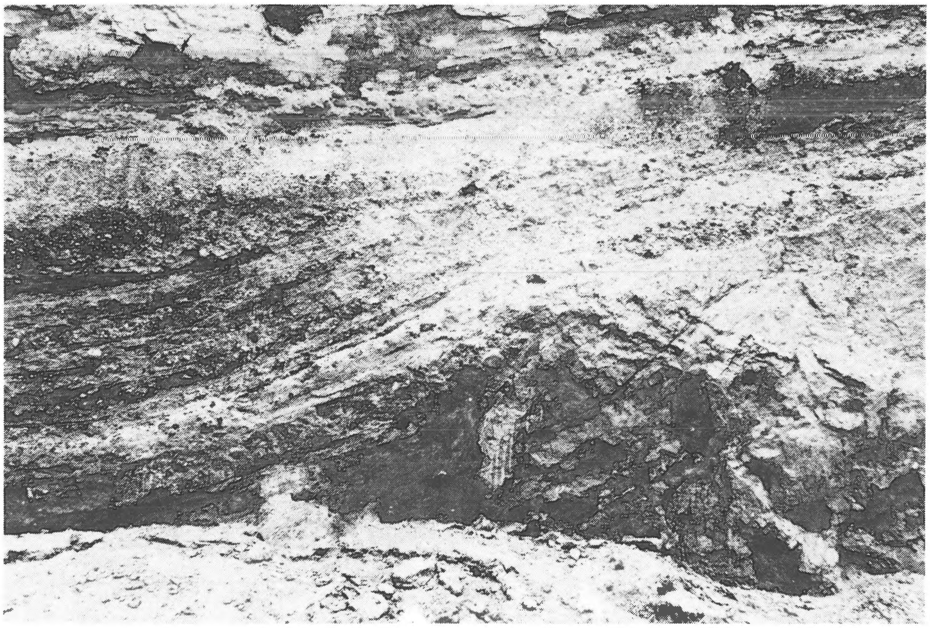


Fig. 12
Two facies of the Upper till near r.p. 23. Facies C with crude stratification overlies the massive facies C1 (lower right).

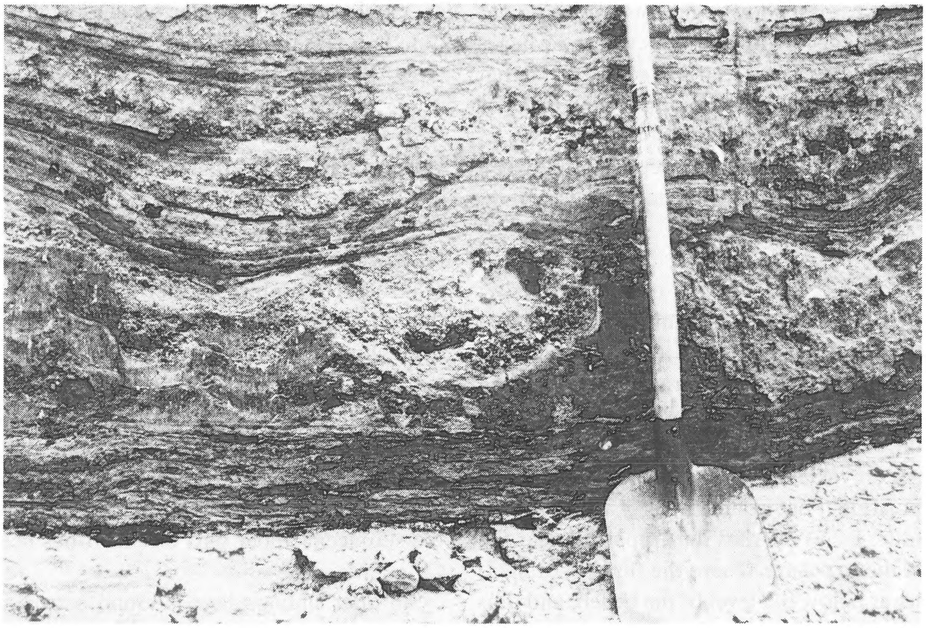


Fig. 13
Facies C2 of the Upper till near r.p. 16. The poor sorting and undulating bedding are characteristic. A contorted sandy inclusion is visible in the middle.

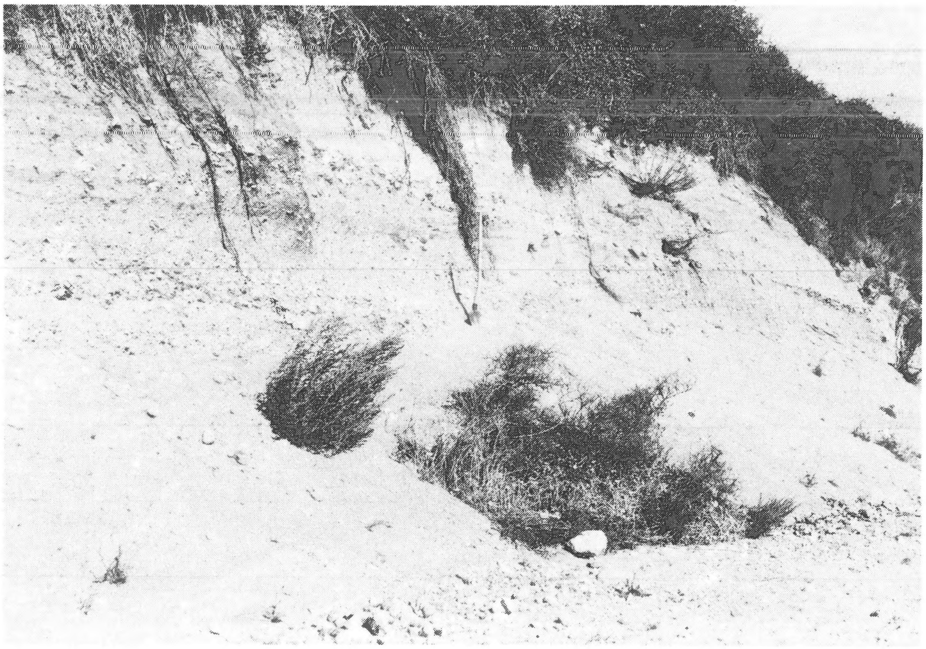


Fig. 14
Sandur facies of unit D near r.p. 535. The bedding is horizontal and not disturbed by faulting. Abrupt vertical and lateral changes in grain size are absent.

Facies C3 – This facies consists of stratified well-sorted silts and clays interfingering with and partly underlain by facies C1. It only occurs near the southeastern end of the Sønderby Klint exposure. The nature of the contact between facies C3 into C1 is taken as proof of their genetic relationship.

Most probably facies C3 is a body of glaciolimnic sediment deposited contemporaneously with the upper part of facies C1. Apparently the mobility of the latter was so restricted that – at least locally – a depression containing stagnant water could form upon its surface. During its infilling influxes of plastic, slowly moving till material were deposited between the strata settling from aqueous suspension.

Sand and gravel beds II

The sediments of this lithostratigraphic unit D were formed in two different environments, viz. (1) as a thick intraglacial crevasse-infilling on top of the kame terrace (Figs. 2, 3 and 5), and (2) as an extra-marginally deposited small sandur beyond r.p. 490 (Figs. 5 and 16). Arguments for interpretation of the first facies as a crevasse deposit have been given in the chapter on the depositional environment.

The second (sandur) facies overlies the Upper till near the southeastern end of the exposure. There the flow till (unit C) dips down to disappear below the level of the beach, and thus the sandur facies of unit D is restricted to the lowermost part of the cliff. It shows a virtually undisturbed and distinct horizontal bedding (Fig. 14). Its appearance is much more fluviatile than that of unit B (Sand and gravel beds I). Abrupt vertical and lateral changes in grain size do not occur and well-developed current ripples are rather common. Due to both the geomorphological expression and the sedimentological characteristics the sediments of unit D beyond r.p. 490 were most likely formed as a sandur. Whereas units A, B and C are attributed to an older oscillation of the Belt glacierization, unit D is thought to be deposited during a later readvance of the same glacier. This was argued previously.

THE DEFORMATIONAL STRUCTURES

Attention was paid above to some structural features insofar as they were associated with a specific rock unit. Three classes of structures require further attention: (1) sliding planes in the top of the kame terrace; (2) normal faults; and (3) superficial slump structures.

(1) The large and subhorizontal sliding planes occur in the cliff's southeastern portion (Fig. 15). Between r.p. 355 and 425 a sliding plane is visible just below the top of the kame terrace. Figure 16 shows how a block consisting of unit C plus the subjacent upper part of unit B has been thrust over unit C.

Another sliding plane is found between r.p. 478 and 488, where it coincides with the cliff's surface. Its genesis is suggested in figure 4. According to this tentative reconstruction a slab of unit C was displaced along a sliding plane due to subglacial shearing. At a later stage (when the ice had vanished) the block began to slide back (dual or reversed movement) under the influence of gravity. For this reason this sliding plane does not fit in either class of normal or reversed faults. Their spatial position suggests that these sliding planes have formed during the active-ice phase of the younger Belt-readvance, i.e. the same oscillation during which unit D was deposited.

In the scatterdiagram (Fig. 9C) the dip directions of sliding planes and other planes of anomalous contact exhibit a weak preferential orientation along a 9°-189° axis. This line is supposed to represent the axis of glacier movement at the time of formation of the sliding planes.

(2) Frequent normal faulting – locally accompanied by subsequent folding – can be seen. The largest of these faults start in the Basal till (Figs. 7 and 16) and extend right up into the Upper till.

This normal faulting is not evenly distributed, but tends to concentrate in and above the till ridges of unit A. Furthermore, from the intersecting faults between r.p. 420 and 450 it can be inferred that normal faulting occurred after the formation of the planes of anomalous contact and therefore must be relatively young. These observations suggest that the system of normal faults developed in response to the ultimate recession of the 'cold shoulder', i.e. at the end of the Belt-glacierization's younger oscillation. The ridges were then subjected to differential sagging and hence normal faulting concentrated in and above them.

As can be seen in the scatterdiagram (Fig. 9D), the dip directions of the normal-fault planes show a preferred orientation of approximately 155°-335°. This value probably corresponds to the axis of glacier movement during the period of final deglaciation.

In table I five groups of structural measurements are compiled. They are all thought to possess directional significance with relation to glacier movement although, of course, their meaning is restricted both in space and time. Moreover, the given values refer to two separate advances of the Belt glacier as suggested in this paper. Despite considerable variation a roughly 167°-347° axis of movement is suggested for both oscillations of the Belt glacierization.

(3) The slump structures referred to are found near r.p. 175 (Fig. 17) and between r.p. 275 and 305. As the presence of these particular slumps does not fit at all into the general structural make-up, these structures are thought to be the result of slumping down the face of the cliff. In this way non-glacial slump structures are superimposed as foreign bodies on the glacially-shaped internal structure of the kame terrace.

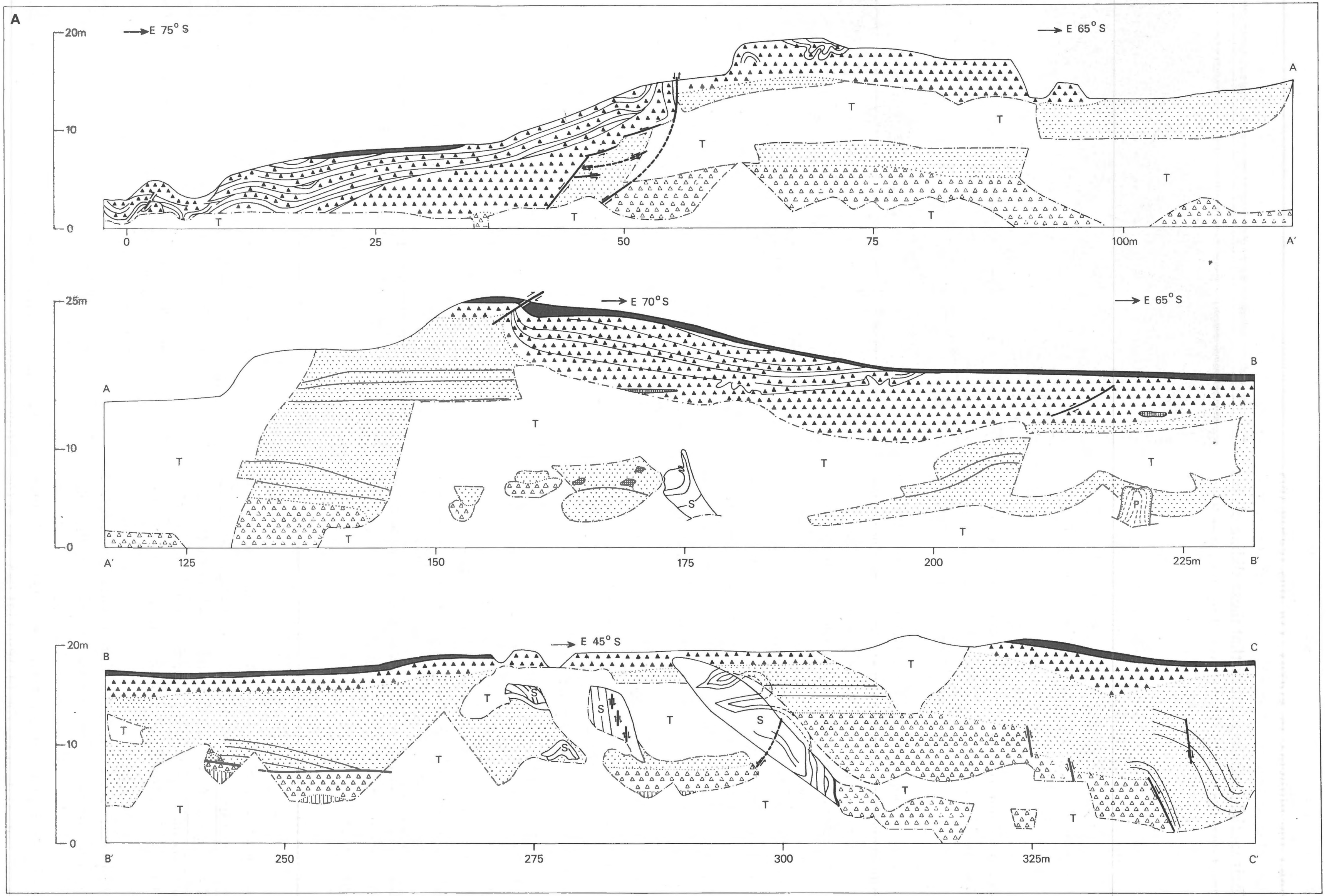
It could be expected that these slump folds would strike parallel to the cliff face. Yet, the mean strikes of the folds (n = 22 and circular variance = 0.1382) and the cliff face differ by 45°. To explain this anomaly it is suggested that the slumping did not occur along the present cliff face, but at the

Fig. 15
Subhorizontal sliding plane near the top of the kame terrace at r.p. 360. The plane is marked by the contrasting lithologies. Compare with inset in figure 16.

Table I
Glacidynamic indicators.

rock unit	structural element	number of measurements	figure in text	inferred axis of glacier movement
A+B+C	dip direction of normal-fault planes	12	9 D	155°-335°
A+B+C	dip direction of sliding planes and other anomalous contacts	15	9 C	9°-189°
B	dip direction of tilted bedding planes	31	9 B	177°-357°
A	dip direction of various planar structures	31	9 A	167°-347°[x)]
A	a-axes of prolate clasts	48	–	150°330°

x) Maximum in corresponding contour diagram indicates ice push from the South.


A
20m
10
0
E 75° S
E 65° S
A
A′
T
0
25
50
75
100m
25m
10
0
E 70° S
E 65° S
A
A′
B
B′
S
P
125
150
175
200
225m
20m
10
0
E 45° S
B
B′
C
C′
250
275
300
325m

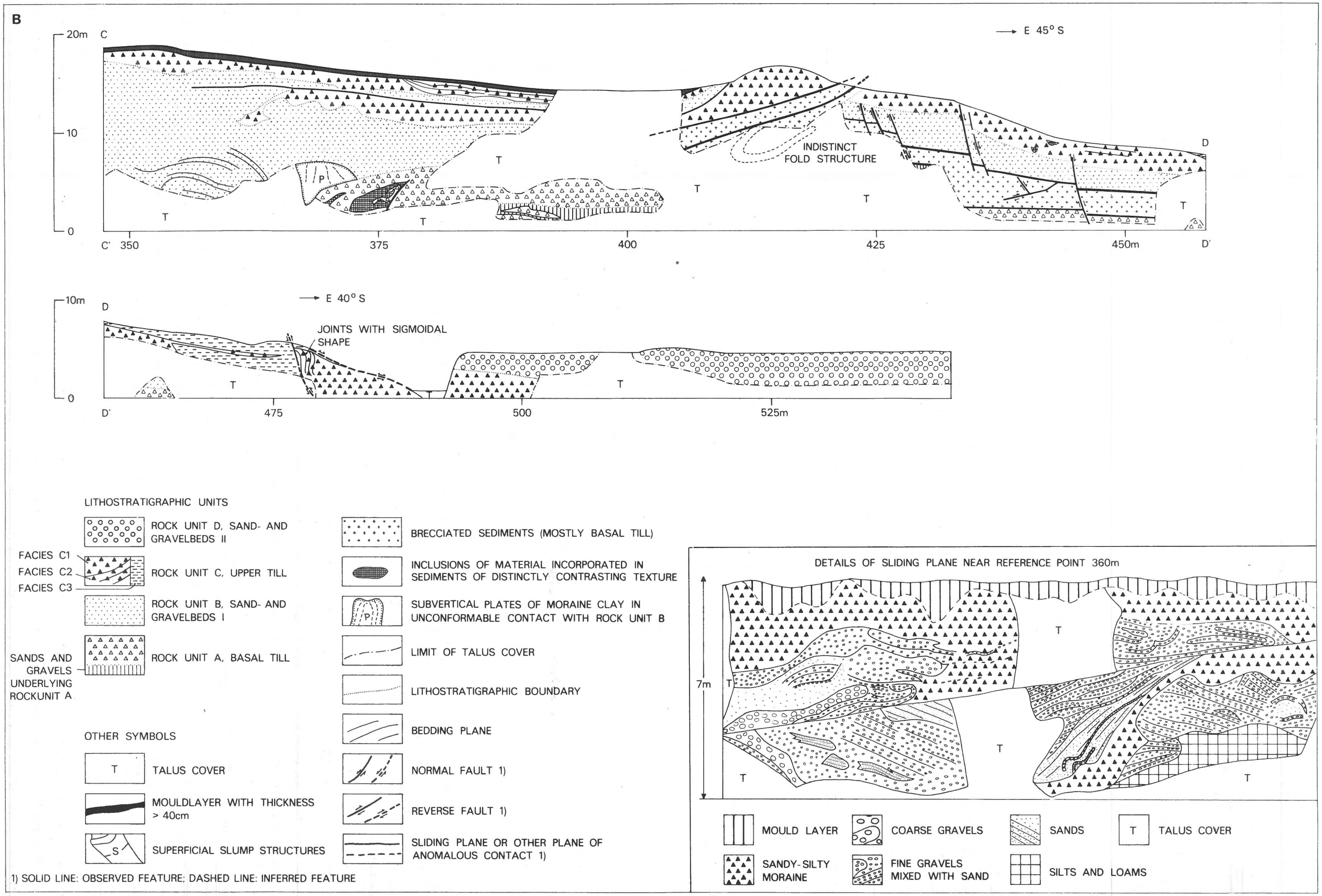


Fig. 16
Section of Sønderby Klint, 0-543 m.

Table II
Reconstructed history of Sønderby Klint

<table>
<tr><th colspan="2">Environmental conditions</th><th colspan="2">Processes</th></tr>
<tr><td colspan="2">Postglacial rise of sea-level and amelioration of climate</td><td colspan="2">Cliff-recession and partial destruction of kame terrace. Incision of erosion-channels on terrace surface and subsequent development of superficial slumpstructures.</td></tr>
<tr><td rowspan="2">Younger oscillation of Belt-glacierization</td><td>Final deglaciation</td><td>Deposition of sands and gravels (unit D) in two different sub-environments viz.
a as a thick intraglacial crevasse-infilling on top of kame terrace,
b as an extra-marginally deposited minor sandur beyond ref. pnt. 490m.</td><td>Widespread normal faulting and subsequent folding due to retreat of "cold shoulder". This process primarily affects the beds which overly the tops of the unstable till ridges. Dual movement along sliding plane produces asymmetrical valley near ref. pnt. 490m.</td></tr>
<tr><td>Active ice phase</td><td colspan="2">Development of sliding planes in top of kame terrace by sub-glacial shearing of readvancing ice.</td></tr>
<tr><td rowspan="5">Older oscillation of Belt-glacierization</td><td rowspan="3">Stagnant ice phase</td><td colspan="2">Static load diapirism produces squeeze-out structures in lodgement till underneath stagnant-ice deposits.</td></tr>
<tr><td colspan="2">Deposition of supraglacial flow till (unit C); basic structure of kame terrace completed.</td></tr>
<tr><td colspan="2">Deposition of stratified sands and gravels (unit B).</td></tr>
<tr><td rowspan="2">Active ice phase</td><td colspan="2">Till sheet remodelled into system of unstable ice-pushed ridges. Glacitectonic brecciation of till-material.</td></tr>
<tr><td colspan="2">Deposition of lodgement till (unit A).</td></tr>
</table>

sides of valleys in the surface of the kame terrace. Incision of these valleys and subsequent slumping must have taken place during the climatic amelioration of Late Glacial or post-glacial times.

Due to the postglacial rise of the sea level, the cliff face receded over a considerable distance. It is therefore quite probable that the numerous erosion gullies depicted in figure 2 are merely the headwater portions of formerly much larger valleys. Consequently, only scattered remnants are left of the slump structures which developed on the walls of these old valleys.

THE EVOLUTION OF SØNDERBY KLINT AND HINTERLAND

In the preceding chapters landforms, sediments and deformations associated with Sønderby Klint were described and interpreted. The observed facts plus their environmental and chronological interpretations are summarized in table II. In this table the life history of the cliff and its immediate environment has been tentatively reconstructed.

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Drs. J. Schijf, a geologist, carried out a large number of structural measurements. His tentative ideas on deformational processes turned out to be fruitful.

Ms. S. Kars, a draughtswoman, prepared figure 16, the overall picture of the surveyed section, with skill and devotion.

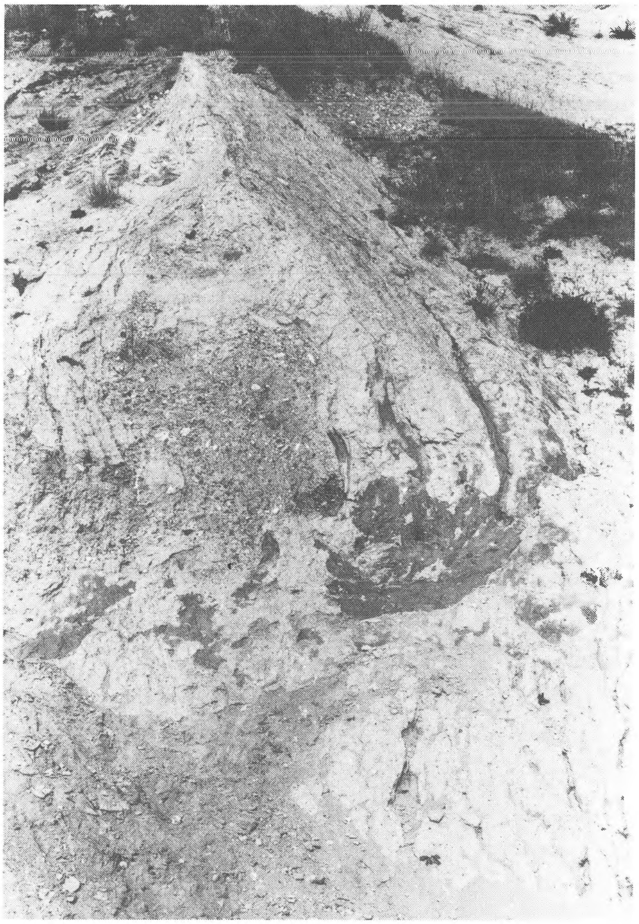

Fig. 17
Superficial slump structure near r.p. 175.

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