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The deformational structures of the deposits on the Höfdabrekkujökull forefield, Mýrdalsjökull, Iceland*

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Abstract: The deformational structures of the developing subglacial (a) substratum deposits and the near Höfdabrekkujökull forefield (b) are the characteristic dynamic parts of the glacio-sedimentological system. These occur in different geomorphological situations: a) under sandur deposits, on the periphery of the fossil embankment of the frontal moraine, on both its distal and proximal slopes, and in the backside depression of this form and b) within the dead ice kettle which was formed in the surface part of the fluvioglacial deposits of the VI sandur level. The older, sub-sandur glaciotectonic discordances, which occur below sandur deposits, represent dynamic structures of two separate glacial advances. Independently of these, deformations of gravitational type also occur. In contrast, the deformations of melt-denudational deposits of the dead ice kettle, surrounded by fluvioglacial deposits, belong to younger (at least several tens years) distortions of gravitational type.

Key words: Iceland, glacier, deformational structures, sub-sandur and dead-ice kettle deposits

Introduction

The dynamics of the Höfdabrekkujökull on the south-east edge of the Mýrdalsjökull (Björnsson *et al.*, 1978; Björnsson, 1979; Jónsson, 1982; Humlum, 1985; Krüger, 1994), and the nature and relief of the bedrock determine the spatial distribution of the moraine and glaciofluvial deposits and the scale of the deformation structures. The deposits of various glacial and glaciofluvial facies display numerous types of deformations in the southern part, close to the forefield of the Höfdabrekkujökull. Structural-lithofacial analysis of these may help to elucidate the glacial processes which control the dynamics of the development and decay of the Höfdabrekku Glacier in the forefield of the main frontal moraines (Heim, 1983). An exposure in the steep slope of the sandur level (VI) revealed below fluvioglacial deposits, a fossil frontal moraine ridge, parts of its disturbed forefield and a thick development of backfield deposits as well as the glaciotectonically-deformed deposits of a proglacial basin represented by numerous sedimentary facies. Also, the younger (at least more than a decade) deposits of a kettle in the proximal part of the sandur (VI) were analysed in structural-lithofacial terms. This enabled the authors to determine the glaciodynamic history of the advances and decay of the frontal part of the Höfdabrekkujökull, with details of the hydrological and morphogenetic changes in the marginal area of the glacier.

Due to the degradation of the so called main moraines, their forefield was free from glacial forms (Heim, 1983; Krüger, 1994). It was completely covered by the levels connected with fluvioglacial activity. Therefore, by means of geological (lithofacial, structural) investigation, it is important to determine the dynamics and position of the front at its older limits rather than in the line of the end moraines.

Glacial and fluvioglacial relief of the south-eastern part of the closer glacier forefield

Jóhannesson (1985) determined the position of so-called outer ("Y") frontal moraines associated with the Older Dryas. As measured in the 1980s, these were situated about 8–18 km from the glacier edge. This

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shows that the Höfdabrekkujökull snout was then situated SE of the Hafursey. Þórarinsson (1959) cited Sigurdsson's sketch which showed the approximately 1943-position of the glacier edge NW of Hafursey, i.e. much closer to this massif than its present position. In its southernmost extent, the glacier snout of that time reached the easternmost resistant block of the Rjúpnagil. Heim (1983) determined the position of the southern and middle parts of the glacier edge in the years 1945, 1955, 1960, 1975 and 1980. In these years the edge of the Höfdabrekkujökull always occurred within the so-called main frontal/end moraines (Hauptendmoränen), which were situated in the median strip between the NW promontory of Hafursey and the Moldheidi rocks. Similar information for the whole Höfdabrekkujökull forefield was provided by Krüger (1994) who also determined the position of Heim's (op. cit.) end moraines for the year 1900. Is it possible that this is a morphological expression of the limit of the Höfdabrekkujökull at the end of the Little Ice Age? Certainly, Krüger (1994) critically considered the possibility that a 4 km-long glacier advance took place during the Little Ice Age.

Slightly more detailed, but nevertheless still generalised geomorphological features (Fig. 1), were presented by Heim (1983) and Maizels (1992), while a further elaboration of the middle-northern part of the marginal zone of Höfdabrekkujökull was provided by Krüger (1994, Fig. 96).

Heim (1983) mapped three courses of moraines in the forefield of this glacier, i.e. in the triangle cornered by the Hafursey, Rjúpnagil and Moldheidi Massifs (Fig. 1). Two of these belong to the principal zone of moraines. The third marks the changes in the frontal moraine from 1945, then from 1960 (when the glacier advanced and overran its previous position) and from 1980 (when the glacier overran its limit of 20 years previously). Other landforms distinguished by Heim (op. cit.) are: ground moraine, a zone of marginal supraglacial moraine of the older and younger glacier edge, and three categories of sandurs: (a) older – sand-gravel covers; (b) middle – which originated after the catastrophic "jökulhlaup" of 1918 and after 1945 and before 1966–1975 (a period when smaller "jökulhlaups" were occasionally generated); this is the so-called generation of "flat sandurs"; (c) younger – marginal sandur cones.

Maizels (1991, 1992) investigated the sandur surface and the lithofacial development of their sediments, as related to the changing hydrodynamic situations. She also divided the western part of the Mýrdalssandur into three relative age categories: a "pre-1918" sandur, a "jökulhlaup 1918" sandur and an active sandur. In her geomorphological map (1992, Fig. 1), upper supra-flood outwash levels, which are densely overgrown by plants, also occur in the zone of active sandur; however, she did not specifically designate them as such (e.g. the sandur in the shadow of small western moraine island – Figs. 1, 2). Also, the designation of "fluted terminal moraine" is not precise, because it associates forms of different genesis including e.g. the Moldheidi rock prominence.

During geomorphological mapping in 1996, the authors identified two moraine patches of different size in the forefield of the Höfdabrekkujökull (Fig. 1), the larger of which is located in the shadow of Moldheidi obstacle. These suggest a distinct duality of glacial relief. The NE side represents directional fluted moraine. In contrast, the SW side represents undulating ground moraine, the orientation of which is similar to that of the edge of the glacier. Two ramparts of end moraine are preserved here. This patch is surrounded by two older sandur levels (II and III) on the distal side (Hafursey) and one of the younger levels on the north. On the side proximal to the glacier, i.e. the western side, the moraine was dissected by various melt water systems. From the NE to the SE they created: (a) a planar zone of outwash, eroded by meltwater, which narrows towards the SE, (b) a younger, intra-moraine level of sandur, mainly of N-S orientation but which curves towards the SW, and (c) a cover of gravel-sand which represents the present-day system of alluvial cones.

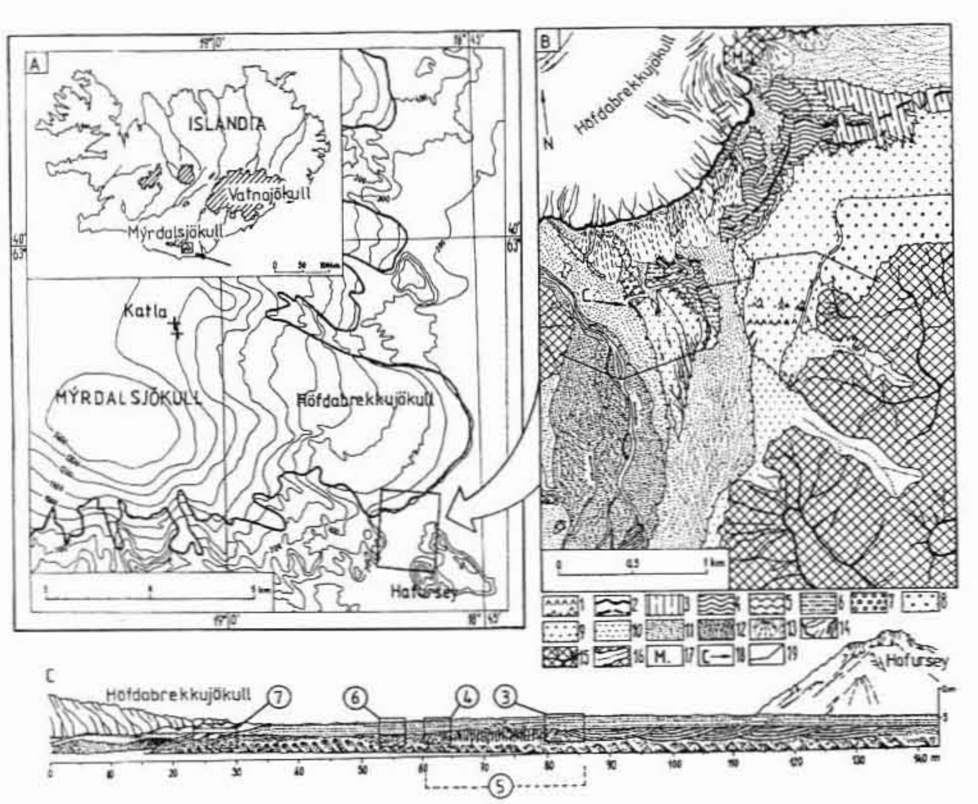

Fig. 1. The Höfdabrekkujökull (A), geomorphological map of the SE part of its proximal forefield (B) and the panoramic profile of the sandur area (C) with the location of the outcrops (Figs. 4–7).
1 – residual concentrations of boulders; 2 – courses of end moraines; 3 – flat ground moraine; 4 – undulating ground moraine; 5 – hummocky moraine surrounded by end moraines; 6 – erosional planes of melt water; 7 – "pitted" sandur; 8–10 – older sandur levels; 11 – modern upper sandur level; 12 – modern lower sandur level; 13 – marginal sandur cones; 14 – the glacier surface with crevasses; 15 – rock massifs; 16 – valley cuts of the rock massif; 17 – Moldheidi rock massif; 18 – location of the outcrops (Figs. 4–7); 19 – line of altimetric profile.

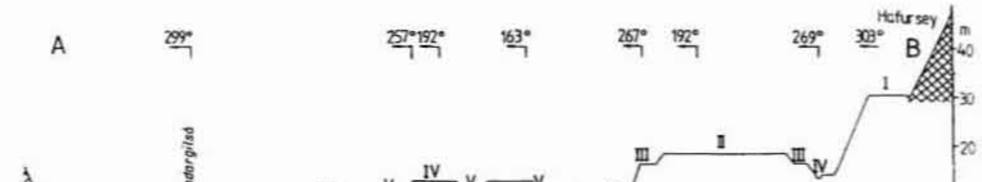

Fig. 2. Altimetric profile of the sandur surfaces (I–X) and morainic surface, west of Hafursey with the morphological units as determined on Fig. 2B (4, 6, 8–12).

A second, smaller ground moraine patch is situated west of the first. This is elongated towards the N-S triangle and it is also composite in relief. The distal part, which narrows towards the S, represents an undulating ground moraine. The widened proximal (northern) part represents hummocky moraine with fragments of the same ramparts of end moraines. These two parts are separated by a system of valley incisions. The western edge of the moraine island is surrounded by an erosional plane formed by melt water.

At present, the youngest end moraine is developing along the entire edge of the glacier. It has the form of an ice-core moraine rampart. On the proximal forefield of this moraine, a 300 m-wide zone of laterally overlapping or eroded alluvial cones is being formed. This is the equivalent of Heim's (1983) "high sandurs".

The main part of the Höfdabrekkujökull forefield is also occupied by sandurs. Their vertical range, relative to the River Remundargilsá, is about 30 m (Figs. 1, 2). The largest cut between the highest levels (I and II), which were designated by Maizels (1992) as "pre-1918" (I) and "1918-jökulhlaup sandur" (II), were developed in the eastern arm of the NS fluvioglacial system. The Authors have distinguished five sets of sandur levels here. These are: the highest level (I), higher levels (II–III), middle levels (IV–VII), lower levels (VIII–IX) and floor (X) which is a modern, active sandur. The Katla subglacial volcano and catastrophic outflows of a large amount of volcanic-pyroclastic material from below the ice result in vast fluvioglacial surfaces of Mýrdalssandur being black (I–II), black-brown or black-brown-grey (e.g. VII–X) in colour. The bright green colour of the dense plant cover present on the VI and V sandur levels distinguishes these from the others. These levels may therefore be called "green sandurs".

The "green" sandur level (VI) is characterised by a smooth, rather flat surface which slopes gently southwards. The distal part (i.e. the S) is cut by shallow channels of melt water (at present mainly nival). The proximal part (N) is distinctly less elevated and contains numerous dead ice kettles and a deep valley of melt water. The kettle-part of the sandur is developed on the eastern edge of the hummocky moraine patch (Fig. 1). Both parts of this level (VI): the proximal part of the "pitted" outwash and the more elevated main flat, decline westwards along a fresh, very steep erosional slope to the sequence of lower sandur terraces (VIII–IX). The main part of the "green" sandur (VI) rises 5.5 m above the VIII level (Olszewski, Weckwerth, in press). This wall shows the sandur structure. The sediments which underlie the sandur may also be seen; these include the deposits of fossil and eroded older end moraine and the deposits of the proglacial basin which were formed on the proximal side of this older end moraine slope.

The sub-sandur deposits and their deformations in the contact zone of the distal part of the fossil end moraine

The oldest deposits of the first of the outcrops studied represent three lithofacial sets (Fig. 3). These are: **morainic deposit – till (1)**, **basic diamicton (2)**, deposits surrounding the fossil morainic rampart distal base which underlie the channel facies of the younger sandur and **bedded gravel-sandy sediments (3)** which contain numerous deformational structures.

The olive-coloured sandy diamicton with an admixture of gravels (2) developed in the proglacial environment, probably in the ice-contact zone. It has a massive structure. It has an erosional contact with the overlying pavement and gravel channel facies. Apart from the overlying fluvioglacial series, the diamicton intrudes from beneath into the deformational structures of the sandy-gravely deposits (Fig. 3, a_1). These

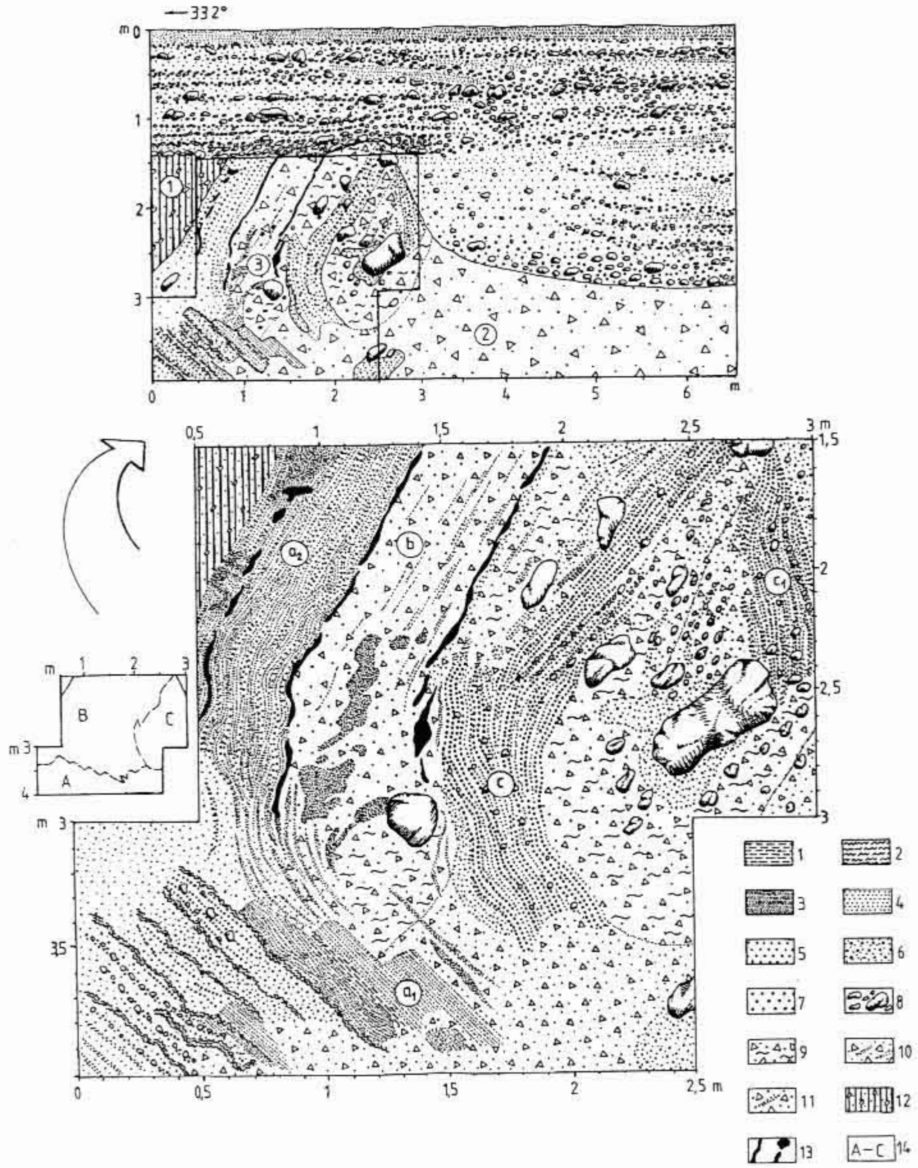

Fig. 3. Geological structure of the sandur plane (VI) and disturbed sub-sandur deposits – the contact of the distal part of the moraine rampart with the deposits of deformational structures.
1 – black sand homogeneously inclined; 2 – significantly disturbed light brown silt; 3 – silty and fine-grained black sand; 4 – medium-grained black sand; 5 – coarse-grained black sand; 6 – heterogeneous black sand; 7 – black gravel; 8 – boulders and pebbles of different size; 9 – sandy-clay grey-brown diamicton; 10 – sandy dark olive diamicton; 11 – sandy-gravel olive diamicton; 12 – glacial grey-olive diamicton (till); 13 – layers and lenses of black organic substance (fragments of old plant cover, now in the disturbed position); 14 – lithofacial complexes.

deformed deposits (3) fill the space between this diamicton (2) and the morainic deposit – till (1) of the distal and deformed slope of the fossil end moraine (Fig. 9: 1).

In the zone of **disturbed deposits (3)**, which are located between till and lower diamicton of the distal base of fossil moraine rampart, three different deformation elements may be distinguished in the sediments, namely:

(a) **lower (A)** – black, laminated fine-, medium- and coarse-grained sands (Fig. 3, a_1) with gravels. Part of this structure is built from rhythmically laminated sands with thin beds of light brown silt. The whole structure is disposed isoclinally to the south. The degree of the destruction in this structure (a_1) increases in two directions: upwards and towards the moraine (i.e. to the NW). The deformational structure is represented here by the surfaces of sudden disruption to bed continuity, their rafting and the small folds in the silt layers.

(b) **upper-proximal (B)** – at the till margin, it is built from the structures of uni-directional inclination of the following layers: sands (a_2), diamicton (b) and sands with gravel (c). These all belong to the western limb of the disturbed anticline. The secondary structures of fold micro-disturbances are a characteristic feature of the sand layer. These occur in the roof zone of the sands, at the contact with the till. In places where they directly touch the floor of the morainic deposit, their sharp, flat structural deformation is visible. Other contorted layers which do not contact with the overlying till and which plunge into the sandy matrix show a more ovate form. The olive-coloured sandy diamicton (b) is 0.5 m thick and shows a complex internal structure. Various, amoeba-like structures of fine-grained black sands are present.

Downwards, the sandy and diamictonic deposits (a_2, b) merge into sands of variable grain size and create structures which open to the north-west, the layers being twisted towards the south-east (a_1/ a_2, b, c). Also a diversion/compression structure of black sands of different grain size occurs here. Deformed laminae of the original sedimentary structures between (a_1–a_2) are clearly visible here.

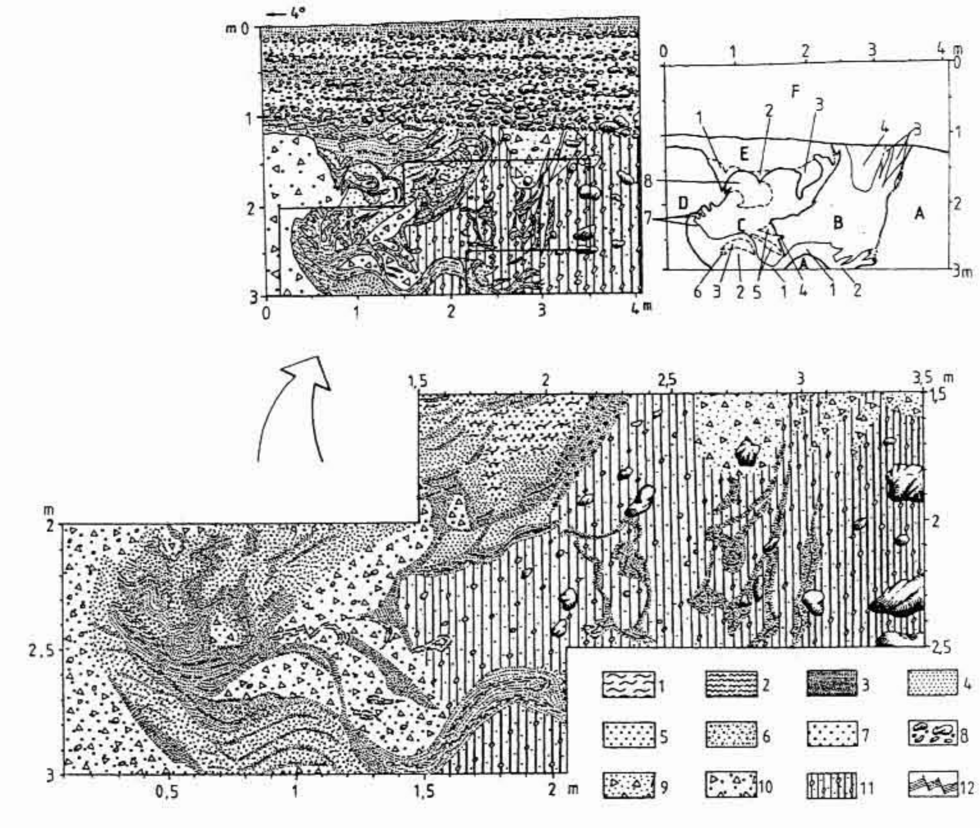

Fig. 4. Geological structure of the sandur plane (VI) and disturbed sub-sandur sediments – the contact of the proximal part of moraine rampart sediments with the deformed, younger sediments.
1 – olive-grey clay; 2 – disturbed black silt; 3 – silty and fine-grained black sand; 4 – medium-grained black sand; 5 – coarse-grained black sand; 6 – heterogeneous black sand; 7 – black gravel; 8 – boulders and pebbles of different size; 9 – sandy olive diamicton; 10 – gravelly olive diamicton; 11 – glacial olive-grey diamicton (B) – thin lines, and till (A) – bold lines; 12 – faults.

There are also three thin, upright and inclined horizons of organic matter (Fig. 3). They are, of course, equivalents of the old subaerial morphological surface (or surfaces). If the age of this material was determined, it may have offered a more reliable control to the interpretation of the deformation process which took place on the distal side of the buried and now-destroyed moraine rampart. Unfortunately, during the next expedition in 1997, this part of the outcrop was found to have been eroded away.

(c) **upper-distal (C)** – built from the upright fold structure (Fig. 3, c–c_1) of the laminated coarse-grained sands and gravels. These surround the gravel-sand core and the internal diamicton casing. To the SE, this structure is broken by an erosional surface which is the base of the much younger fluvioglacial channel sedimentation. The axis of this fold (which can only be traced discontinuously on the southern side) has approximate deep 14,3° on NW (strike – N 329,3°; Fig. 5B).

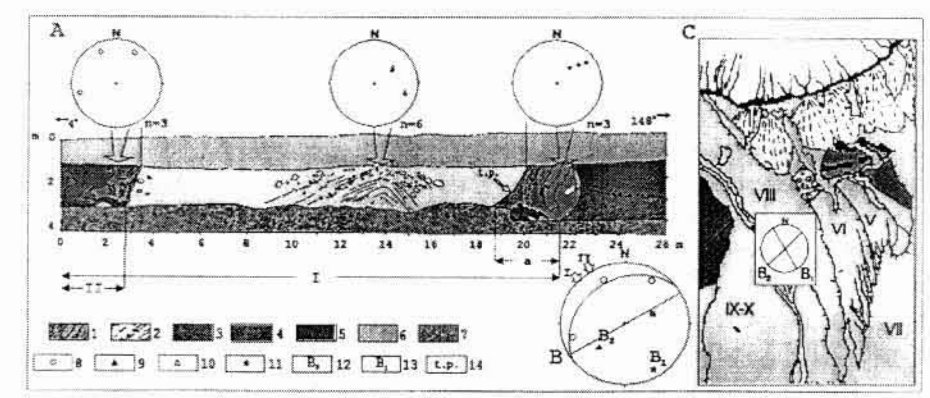

Fig. 5. The distribution of deformation structures within the central part of the fossil end-moraine and its distal and proximal margins.
1 – sand, gravel and diamicton of the fold on the distal side of the fossil end moraine; 2 – till of the fossil end moraine; 3 – basin sands and silty sands; 4 – sand and gravel diamicton of sedimentation in the ice-contact zone; 5 – bed sands and gravels of the younger fluvioglacial period; 6 – sands and gravels from the sandur level VI; 7 – ridge;
A: Elements of structural deformations of the core of the fossil end moraine and its environment: a – zone of static dislocations preceding recession and thrust (I), I – zone of dynamic structures of a thrust (I), II – zone of dynamic and static structures of a thrust (II);
B: Structural elements of deformations of the fossil end moraine and its surrounding (Schmidt's stereographic projection, upper hemisphere): a course of the breaking edge of the Höfdabrekku Glacier front before a thrust (I), during deformation of the zone (a), I – average direction of the glacier thrust during deformation of the zone (II);
C: Geomorphological sketch of a part of the Höfdabrekkujökull forefield, upper right corner of the rectangle with orientation of folding structures B_1 and B_2 showing the location of the exposure in Fig. 3.

Sub-sandur deposits and their deformation in the contact zone of the proximal part of the fossil end moraine

The deposits which are present under the youngest part of the fluvioglacial series (F), on the proximal side of the mentioned fossil end moraine, originated on the floor and in the forefield of the retreating glacier edge (Fig. 4). Their dynamic deformations started then and finally developed after a longer period of glacier retreat from the line of fossil end moraine during a renewed advance. The next lithofacial sets of this part of the geological cross-section are:

(a) **till (A)** – developed as an olive-grey glacial diamicton (Figs. 4, 9: 1, 2). This is very rich in gravel. The numerous boulders present are as much as 30–40 cm in diameter. Many of them are placed obliquely, parallel with the line of the proximal wing of the arched anticlinal structure of morainic matrix described above fossil end moraine (Fig. 3: 9). The mentioned fold structure has the axis which inclines 43,2° on NE (strike – N 45,7°). Its crestal line is underlined by boulders aligned with the till lamination anticlinal structure, which in the roof part is cut by the floor of the upper sandur series (F). Above this shear surface, there are larger, residual concentrations of the boulders in the sandur (Fig. 5: A).

(b) **glacial diamicton (B)** – is olive-grey in colour and contains boulders which are generally smaller than those in (A) up to 18 cm in diameter. This set is also more richly endowed with structural forms (Figs. 4, 9: 4). There are numerous fold structures, especially in the floor, at the contact with the underlying till (Fig. 9: 5). These are: upright anticlines in fine-grained sands (B1), which thin out towards the glacial thrust, and a syncline which rises towards the (A) set and which is irregularly disrupted at the till contact. There are numerous streaks and inserts of silt, sand and sandy diamicton. Folds developed at the contact with till merge into deformations of glacial "penetrations" (Różycki, 1970; Olszewski, 1974; Stankowski, 1977), where they are accompanied by uprising inclusions of fine-grained sand and silt, which indicate a phase of glacial push on till. These were shown in the form of poles in Schmidt's stereographic projection (Fig. 5: A, B). The mean direction of their course is N 329,6° (Fig. 5 B–II).

(c) **multi-grained sands with sandy diamicton (C)** – In terms of structure and texture in this part of the series, these represent the most differentiated lithofacial set (Figs. 4, 9: 4). This results from their indirect location between the lithofacial assemblages of diamictons (B) and (D). In the lower part there is a fold structure, the central part of which – the anticline (C2) of multi-grained sands with inclusions of sands and fine silts – became thickened (0.5 m), a response to a moderate glacial push (Fig. 9: 5). Its fold radius permitted the continuity of the constituent layers and their primary lamination. In the axis of the anticline, elements of "overtaking" inside the layer pattern may be seen (C2). Above these fold zones, a lens of olive-coloured diamicton (C3) was deposited here within the decline, distally, towards the thickened anticline. A distal tilt of the diamicton lens then took place, its proximal part becoming underlain by a translocated multigrained sand (C6). The translocation is also associated with the development of the structures of sandy layer disruption, wedge deformations and lenticular penetration of silty bodies inside the diamicton (C4, C5). Deformational microstructures (C7), which developed where this series is contiguous with the diamicton (D) on the proximal side, are quite similar.

(d) **gravel-sandy diamicton (D)** – this is olive-coloured and belongs to the northern limb of the exposure (Fig. 4). It has a form of an inner, homogenous, compact sedimentary block (Fig. 9: 3). Its almost vertical contact with the series (C) has been deformed into an S-shaped fold along a prominent thrust surface (Fig. 9: 5).

Those sediments in the lower part of the exposure described above (Fig. 4) represent a succession of the primary sedimentary environments. They developed from the marginal-glacial environment – till (A) and the proximal-marginal environment – glacial diamicton (B) to the ablation environment (two types, i.e. submarginal and ice-contact) – gravel-sand diamicton (D) and to the proglacial basin environment – laminated gravel, sands and sand-silty layers (C). During glacial thrusting, a complex system of glaciotectonic structures developed in these sediments. This occurs in the deposits which were formed on the proximal slope of the fossil end moraine.

(e) The **lithofacial set (E)** is variable in thickness and represents a unit quite different in relation to those underlying deposits which have glaciotectonic structures (A–D). The nature of its deposition, i.e. horizontal above the diamicton (D), and the normal discontinuity structures which occur there, suggest a post-sedimentary and gravitational genesis for the deformational structures. There are three main structures:

– (E1) – a wedge-like sand-silty fill with an admixture of clay, about 30 cm thick with inner normal faults which throw about 2 cm. The overlying curved sandy layers are not in sedimentary continuity – rather, there are normal faults and pocket-like lenses;

– (E2) – a wedge structure, 24 cm deep, filled with black fine-grained sand. This penetrates fluidal structures (C8) of the underlying unit;

– (E3) – a curved pocket-like structure filled with grey clayey gravel and sand. This is disrupted by normal faults which have throws of about 1.5 cm.

The authors regard these structures as due to the melting of small lumps of dead ice which were left in the deformed material after the glacier advanced (Fig. 9: 6). The gaps which developed during the ablation of these dead ice masses were simultaneously occupied by black sandy diamicton which introduced overlying layers of silt and sand.

The lithological character of this series (E) and its depositional structures and gravitational deformations suggest that accumulation took place in a shallow basin which existed periodically on the dead ice and that the destruction of these deposits had a relaxation character (Rotnicki, 1977; Dadlez & Jaroszewski, 1994). The existence of this small basin was associated with ablation of ice which earlier, in its dynamic phase, marked the former base of the glacier and its contact zone with disturbed sediments of the near substratum (Fig. 9: 5).

Deformed sediments of the proglacial basin and associated facies on the back of the fossil morainic ridge

Sediments immediately adjacent to the sandur cover (F) (Fig. 4) are represented by older fluvioglacial deposits (A), glacial deposits (B) and semi-glacial sediments of the proximal ice-contact environment (D) and deposits of a proglacial water basin (E). All these facies are structurally deformed. The deformational structures, differentiated according to type and size, are probably associated with an older deformation phase at present of the fossil morainic ridge (the older deformation structures) and with a post-recessional new readvance of the Höfdabrekkujökull (the younger deformation structures) (Figs. 5, 6, 9). The following lithofacial sets have been distinguished:

(a) Laminated and disturbed **fluvioglacial gravels and sands (A)** – present in the lower part of the outcrop (Figs. 6, 8–I). A distension of the lower fold occurs to the north. This comprises the lens of the diamicton core (A2) in its disrupted gravel-silt hinge. The dynamic surface of this fold is horizontal and the Authors associate this with the horizontal, older glacial advance (Fig. 9: advance I). Above, occurs a black silty sand (A3) which creates a zone of glaciotectonic shear structures which is inclined towards the NNW. This is certainly younger than the fold structure and it probably originated after the recession, during the next advance of the glacier, which banked up the already deformed deposits of the substratum. The later glacier advance also created the small drag folds which were formed below the shear plane (Rotnicki, 1971).

The upper part of this set is built from a small glacial slice or raft (A5). Its front is distinctively raised, the floor is well-defined and it shears the sandy sub-

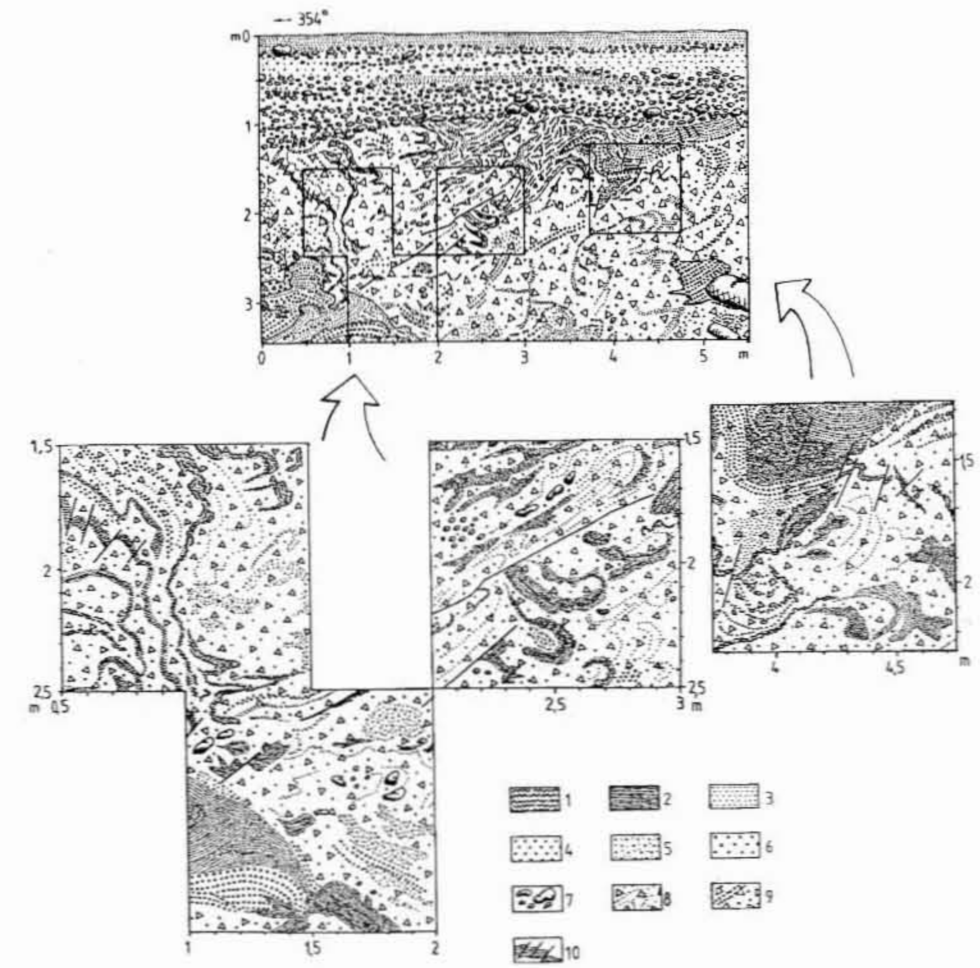

Fig. 6. Geological structure of the sandur plain (VI) and disturbed sub-sandur sediments of the fossil proglacial basin. 1 – dark brown silt; 2 – silty and fine-grained black sand; 3 – medium-grained black sand; 4 – coarse-grained black sand; 5 – heterogeneous grey-blue sand; 6 – black gravel; 7 – boulders and pebbles of various size; 8 – sandy dark olive diamicton; 9 – pebbly dark-olive diamicton; 10 – faults.

stratum (A6). The secondary disturbed structures of the discontinuous and broken folds and fluidal lenses (A6) probably originated as the products of the younger glaciodynamic generation, during the next advance of the glacier (Fig. 9: advance II).

The next part of the deposits of the (A) complex, which is located 1 m south, consists of a sand and gravel diapir (A7) which is markedly inclined towards a glacial thrust. The hinge of this structure is surrounded by curved layers of gravel and coarse-grained sand (A9) and, locally, silts (A8), which create microstructures of wedges and folds (Figs. 6, 8: I). The scale of disturbances is clearly gradational. Further south, two small gravel-sandy glacial rafts occur (A10) with primary deposit lamination indicative of flowing water.

(b) **Morainic diamicton (B)** – gravel, locally sandy-gravel, sporadically with boulders up to 10 cm in diameter. This occurs about 1 m below the floor of the sandur layer (F). Its roof rises gently in the distal direction. The structures of the diapir penetrate it from below and parcels of older fluvioglacial sediment also occur there; these have form of isolated folds. In the lower part of this set, the axial planes of these folds show horizontal disposition or are gently inclined (B1). Upwards, some of them are sharply intersected by slip planes which are inclined at about 30° N/NW (B2).

The morainic diamicton (B) represents a subglacial (submarginal) deposit, which originated here during an earlier glacial advance (Fig. 9: 3, advance I) on the deposits of the older fluvioglacial facies (A).

(c) **Diamicton with an admixture of gravel and sand (C)**; in textural terms, this represents a heterogeneous lithofacial assemblage (Figs. 6, 8: I). It is the deposit of a close-to-ice contact environment. Gravel-mud landslips, typical of the stagnation and disintegration of the glacial mass in its recession stage participated in its creation (Fig. 9: 4). After recession, the next glacier advance occurred and this deposit entered into the active submarginal environment (i.e. the subglacial environment close to the glacier edge). This caused an increased glaciotectonic disturbance in this sediment (Fig. 9, advance II).

Some structures in the distally-inclined and disrupted layers of sand in the lower part of this complex may represent fragments of recumbent fold flanks (C1), which are associated with the underlying fold (A1) which was disrupted in the next glaciotectonic cycle. Small-scale folding in the fine-grained sands (C2) which overlie these deposits show horizontal or gently-inclined disposition of the axial movement planes. This suggests that the deeper part of the complex might have originated in a glacial crevasse, the overlying, proximal part of which still showed significant input of energy. Thus, these structures must be associated with the older generation of dynamic dislocation during its termination.

It should be emphasised that this set (a) has an oblique disposition, its proximal (N/NW) inclination being considerable, from 30–40° in the lower part, above (A1–4), to about 20–30° in the upper part, whereas, on the roof section, at the contact with the (E) complex, the inclination increases to 40–50° and (b) is located in the production of the intrafold, slip structure of sands (A3). These two superior structural features should be associated with the second, post-recessional, generation of glaciotectonic dislocations. This generation is also represented by the following structures (though some are secondary): planes of slip faults with the main surface of dislocation inclined towards NNW (C3); "blind" dislocations (C4) which occur somewhat deeper in the profile (Chrzanowski & Kotowski, 1977) and numerous reverse faults and shear faults which occur below basin sediments (EII).

The termination of glacial thrust, which initiated the second generation of glaciotectonic deformations, took place in the zone where the "ice-contact" sediments (C) are associated with the upper part of the subglacial deposits (B). Here, they overlapped the older deformations and formed multiple superimposed glaciotectonic structures (Fig. 8–Ib). With respect to its occurrence in the contact area of the upper parts of the lithofacial assemblages (A) and (B), a "transitional" zone may be distinguished where all these deformations overlap (Rotnicki, 1976a, 1976b).

(d) Dark olive-coloured, **bedded gravel diamicton (D)** (which is sometimes massive) occurs above the lithofacial set (C). In terms of its environment, it represents a transitional facies between (B) and (C) sediments. It comprises numerous layers of gravel (B1), especially in its middle part, where it is associated with short and numerous reverse faults, prolixities and rafts of fine sand and silt in the distal part (D2). There is a characteristic wedge-shaped narrowing of the olive and partly massive diamicton which thrusts between the blocks (EI) and (EII) of the overlying complex.

The lithofacial, textural and pre-deformational character of the depositional structures suggests a transitional sedimentary environment, which developed at the boundary zone of the removing ice-contact in the proximal part of the water basin.

(e) **Fine-grained basin sediments (E)** occur above the lithofacial assemblage (C) and the proximal basin deposits of the nearby ice-contact environment (D). Dark-brown silts and black, fine-grained silty sands occur in the northern packet (EI). The southern packet (EII) consists of black, fine- and medium-grained silty sands underlain by gravel.

The proximal-floor part of the (EI) sediments, which were deformed by thrusting of the edge of the diamicton packet (D), developed into horizontal "wedges", and small overlaps. In the middle part of the packet, stream-blurred structures occur with small secondary folds (E2).

Dislocation of the central parts of these structures was horizontal, whereas the floors were distal and elevated. This suggests a directional association with a nearby deeper slip structure which comprises the (A3–C) set.

A group of upright, slightly inclined folds appears inside the (EII) packet in the propagation of a wide drag structure (E3) and on the contact with (C). Their edges are distally dislocated. The scale of dislocation increases upright from (C) towards (EII in E4), which is an indicator of the directional structure. Comparing the limits of this structure, the static and dynamic elements of the thrust may be estimated (Ruszczyńska-Szenajch, 1979). A slightly deformed gravel layer (E5), with a gently folded roof, occurs towards the south. The overlying, laminated, black, fine-grained sands are cut by a series of small-scale reverse faults (E6). Both these structures suggest a glaciodynamic genesis – a small-scale retrograde slip (slip-hinge) movement in this part of the sediment is indicated which is located below the main axial part of the inverted and rended fold.

The principal features of this lithofacial set (predomination of the laminated fine-grained deposits of an environment of slow-moving and/or stagnant water) suggest a local break in the glacial episodes, on the salients from the fossil end moraine rampart. Analysis of the structures in the outcrop suggests the presence of two glaciotectonic cycles in this part of the forefield of the Höfdabrekkujökull (Figs. 8–I, 9), both predating the deposition of the surface sandur cover (F).

Dead ice kettle sediments of the proximal part of the sandur (VI) and their deformation structures

The forth exposure is placed in the dead ice kettle of the sandur level (VI). This is situated at the contact of the different (in geomorphological terms) parts of sandur: the proximal, which is relatively low-lying and has numerous dead ice kettle depressions, and the central and distal parts which are flat and more elevated (Fig. 1). The melt-out-denudational lithofacial complex of the dead ice kettle is located here. This is placed in the lithofacial structure of fluvioglacial deposits and the substratum and diamicton sands (Figs. 7, 8–II).

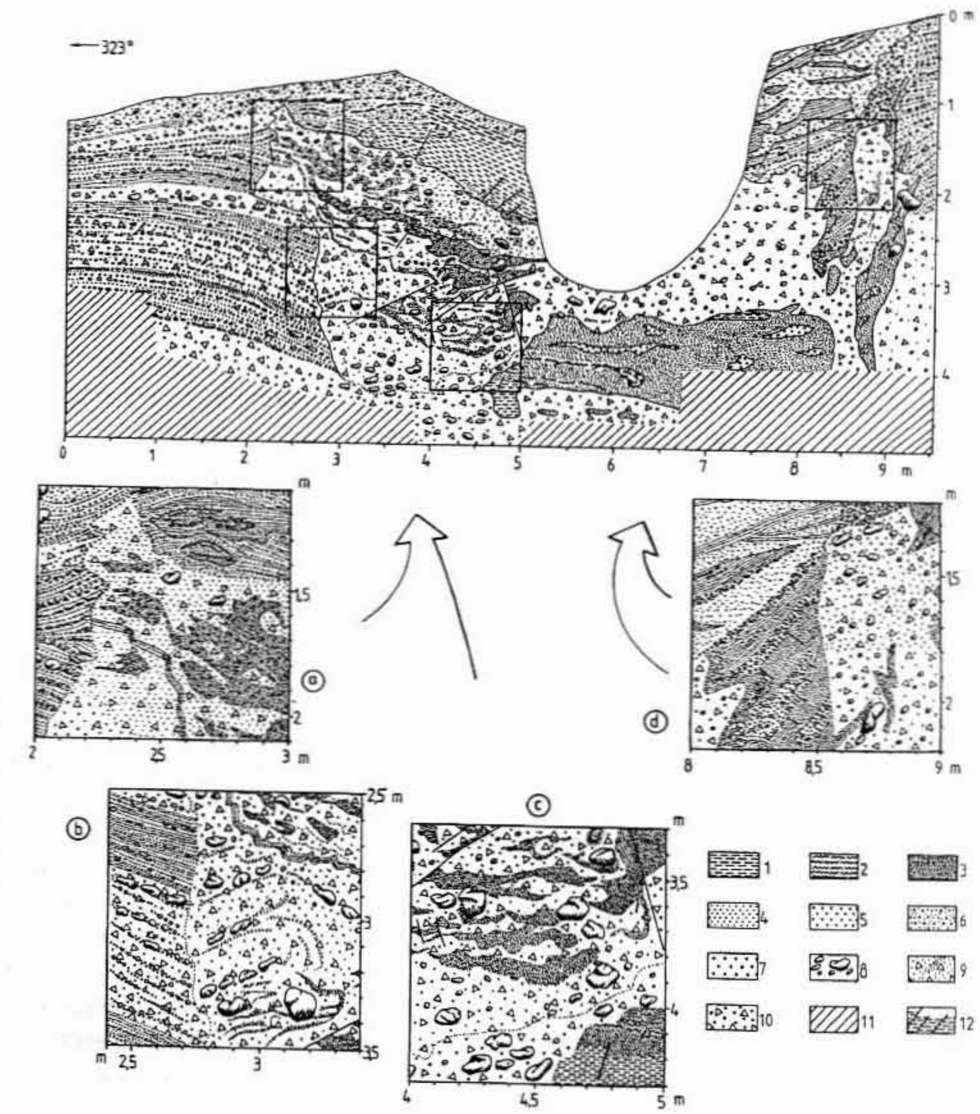

Fig. 7. Geological structure of the sandur "pitted" outwash level with the kettle filled with melt-out-denudational sediments. 1 – black silt homogeneously inclined; 2 – disturbed black silt; 3 – silty and fine-grained black sand; 4 – medium-grained black sand; 5 – coarse-grained black sand; 6 – various-grained black sand; 7 – black gravel; 8 – boulders of various size; 9 – sandy grey diamicton; 10 – pebbly olive diamicton; 11 – embankment; 12 – faults.

(1) **Bedrock deposits** are represented by two lithofacial sets:

(a) olive, **morainic diamicton (A)** is massive, gravely, and locally contains lenses of black, fine-grained sand. The diamicton roof is erosionally lowered below the melt-out and denudational lithofacial series and it merges upwards below the fluvioglacial deposits to the N and S. The thickness of the substratum below the dead ice kettle and sandur ranges from 0.8 to 2 m. The elevated, southern festoon of the diamicton was the reason for the more elevated position of the sandur here. It might also have blocked the transport of dead ice during the jökulhlaup flood.

(b) black, **massive fine-grained sands (B)** with lenses and streaks of olive-coloured silt and coarse-grained sand. The original lamination of the silt is cut by a few small normal faults with throws of about several centimetres. The direction of their inclination (N/NW) is similar to that of the faults of the overlying lithofacial set (G). The textural features of the deposit, its slightly disturbed structure, the occurrence of long, horizontal and the thin lenses of coarser sand and silt laminae all suggest a lithofacial environment of stagnant water with episodic flow. It might also be the trace of older basin sedimentation from the glacier recession period which took place before the development of fluvioglacial deposition (C–E; Figs. 7, 8–II).

(2) **Fluvioglacial sediments (C–E)** represent the environment surrounding the dead ice kettle. They may be correlated with the youngest (roof) lithofacial complex in the model of structural composition of Iceland sandurs described by Maizels (1991, 1992). It is also very similar (Fig. 7), particularly in terms of the structure, to the material deposited during declining phase of the jökulhlaup, by the "Katla flow of 1918, resting on undisturbed soil at Núpar" (Jónsson, 1982, p. 64), south of the River Múlakvisl water gap. In the fluvioglacial uppermost part of the complex, three lithofacial sets have been distinguished:

(a) **lower (C)** – 1.4–2.2 m thick, consists of horizontally bedded gravel and sands with small boulders on the northern part of the dead-ice kettle. This is inclined slightly towards its fossil floor. It was also noted that there is a change in the inclination of both the laminae and the layers in the roof of the lower set (C) in the contact zone with the melt-out-denudational complex (3). Within the middle set (D), these are inclined upwards, which appears to be a structural record of a major change in the sandur sedimentation at the contact with melt-out kettle deposits.

(b) **middle (D)** – which consists of diverse sands and gravels. This part of the sandur was filled with a considerable volume of dead ice which was emplaced at the beginning of the deposition of this fluvioglacial set.

(c) **upper (E)** is a mixture of sand, gravel and boulders of diameter <15 cm. In the floor, the lamination of the deposit is only locally visible. It also occurs laterally above the youngest lithofacial set of the kettle (J) but this does not occur on the upper level of the flat sandur, south of the kettle.

(3) **Melt-out-denudational deposits of the kettle and their depositional-deformation structures (F–J/E).** The sediments which fill the dead ice kettle of the sandur comprise a complex of six lithofacial sets. These are as follows (Fig. 8–II):

(a) **outwash and melt-out diamicton deposits (F)** consist of olive-grey gravels and boulders up to 20 cm in diameter. The oldest deposits fill the channel depression between the fluvioglacial and the substratum deposits and reach 4.4 m below the surface of the sandur edge. This lithofacial set of the diamicton is a grain-like mass of the massive packing skeleton. The main concentration of boulders is situated in the lowest part of the channel depression, which is 1 m wide. In relation to its floor, the inversion course of the roof of these deposits has been created by bilateral tension of the ice and overlying deposits, which were shifted gravitationally. During the bilateral deformation of this part of the set, a narrow diamicton diapir (A) of the substratum was injected between the channel diamicton (F) and the substratum sands (B).

(b) **fine and medium sands with diamicton (G)**, light brown, with boulders in the middle part of the lithofacial set and with layers of laminated black volcanic dust. This sediment represents the lithofacies of a small depression which was permanently filled with water. It was formed below and between the melting wall of the ice with the diamicton, which flowed from the upper part of the ice mass and sedimentary edge of the sandur. The complex was deposited in a relatively low-energy water environment. The silt layers, which were, of course, originally deposited horizontally, were then significantly deformed by the gravitational thrust of the overlying deposits (H) and sinking ice block. The tension was undoubtedly bilateral, though it was probably stronger from the N/NW than from the southern sector, where the substratum sands (B) absorbed much of the thrust. The depositional structure of the highly saturated sediment became folded and dissected by shear faults. Small fragments of this facies, which are in contact with the substratum in the south part of the kettle (Figs. 7, 8–II), show several reversed faults in a high-angled, rotational position.

(c) **basic diamicton (H)** – brown and olive-grey with numerous boulders and coarse gravel. This is the main lithofacial set which fills the kettle. It derives from ablation of the morainic load from the ice block which had been deposited on the sandur surface. It extends over the entire (and maximal) width of the kettle. It also has the greatest thickness (2.1 m) on both sides (northern and southern) of the kettle. The diamicton mass pinches out to the centre of the kettle. The con-

centric thrust of the deposits of this lithofacial set was directly responsible for the development of deformational structures in the underlying sediment (G).

That the highest elevation of its roof on the fossil edge of the kettle coincides with the maximum width of the kettle is another characteristic and notable feature of the basic diamicton (H). On the northern side, this part of the diamicton occurs on the erosional fluvioglacial shelf (C). In the line of the maximal external range, the basic diamicton forms prominent summits. In the spatial pattern, these deposits show palaeomorphic structure of the **rimmed mound (rim) of the diamicton**. These surround the middle fragment of the fossil part of the dead ice kettle depression. The fossil rimmed structure also shows their association with the basic diamicton of the kettle fill and not with any secondary facies. Thus, the main internal fill of the kettle bottom and **rimmed mound** palaeoform create two associated parts of the basic and central set in the depositional series of the kettle depression.

In the cross-section of the rimmed mound, both slopes – external and internal – are steep. The external slope shows a morphological negative of the eroded and deposited with diamicton sandur wall. The internal slope was formed by deformational processes during dislocation of the younger fill deposits.

The root of the rimmed mound also shows deformational structures. On the northern side (Fig. 7b), above the basal shear fault, the following elements occur: (a) traces of the bedding in respect of the small pebbles in the changed position of the gravitationally inclined root structure of the rim, which is the converse of the neighbouring part of the sandur; this occurs near the northern erosional wall of the fluvioglacial mass; (b) fold structures in fine sands in the casing of the diamicton mass, situated further from the fluvioglacial wall. These are the result of the internal deformation of the whole of the diamicton, and they developed concomitantly with the basal shear plane. On the southern side (Fig. 7d), inside the basic diamicton (H) in the root of the rim, some thin lenses of fine sands show vertical (rotational) orientation of inverted microfolds.

Thus, the internal structure of the rimmed ridges shows that they are not just gravitational forms of the ablated glacial sediment. They are also the product of both continuous and discontinuous deformation in their root zone.

(d) The lithofacial set of **black multi-grained sands and black diamicton (I)** has a tripartite form. This consists of the following two parts on the northern side: (a) **lower (I_1)** which comprises sandy diamicton in the base, which contains wave-disturbed continuous layers of volcanic dust; (b) **middle (I_2)** which is the main packet of this sediment and consists of a sandy diamicton with numerous lenses of fine- and medium-grained sands of a stream-involution structure (Grzybowski, 1970). The reversed fault of a southern vergention (36°S) cuts across the folded structures of dark brown sands (**I_2**) and the roof of the lower part; (c) **upper (I_3)** which consists of non-bedded, medium-grained, black-brown sands, gravels and pebbles. This packet increases in thickness towards the kettle centre. For the first time in the vertical arrangement of the kettle deposits, the development is the converse of the lithofacial sets in the lower part of the melt-out complex.

On the southern side, the sands and gravels of the lithofacial set (I) cover a smaller area. They are (as in the case of the northern side) facially associated with the sandur horizon. The packet (I_1) was formed as a narrow, zig-zag wedge structure with a set of small thrust faults. These are situated in a vertical series which have been rotated from their original position. The parts of (I_2) and (I_3) are deposited above the basic diamicton of the rimmed mound. They are clearly an integral extension of the fluvioglacial sediments (C, D) with a dislodgement structure of the grain packing.

In terms of position and texture, the whole lithofacial set (I) of the melt-out fill correlates with the surrounding sandur deposits. Therefore, this accumulation probably filled part of the kettle in which the underground melting of ice was still taking place. This caused subsidence of the deposit and slow gravitational flow on the northern side and movement from an *in situ* position or fall on the southern side of the kettle.

(e) **fine-grained deposits of basinal sedimentation (J)** consist of black silts and fine-grained sands which appear to be ice-dammed lake sediments. This deposit locally preserves its primary lamination structure, e.g. to the north, in its middle part (**J_2**). To the south of the melt-out depression, within the oldest packet (**J_1**), deformational structures of the sudden, block-gravitational movements occur. This is also the reason for its "columnar" position. It resulted from part of the detached sandur being placed in the fissure of the basic diamicton.

The **upper** packet (**J_3**), which is well-preserved on the northern side, shows secondary relocation of the concentrically inclined layers and a series of normal, collapse faults. These occur in its basal part in two concentrations: an upper, the inclination of the fault planes of which decrease towards the centre of the basin (28–42°S) and a lower, which is more central and in which there is increasing inclination of these planes (35–50°S). This is the result of variable intensity and range of the sedimentary slips in different parts of the basin floor. The dip-slip amplitude is small. As stressed by Jewtuchowicz (1970) in the case of the disturbance of the deposits of the end-moraine zone (supposedly a diagnostic feature of the underground decay of dead ice), it did not cause intralamination deformation of the grain packing. A small textural and structural change of the packet (J_3) on the both sides of the kettle axes shows the nature of lithofacial conversion. This is dictated by: a differentiated geometry of the basin floor, the lithology of the direct source surrounding of the facies and the intensity of the delivery process.

The lithofacial set (**J**) in the final period of sedimentation and resedimentation was deposited in the deepening melt-out thermokarst depression, which was, at the same time, morphologically shallowed due to depositionary processes. A slight deformation of the deposits in this stage of its structural deposition took place in accordance with the developing surface of the kettle which became progressively more horizontal.

(f) The **youngest** packet of the **fluvioglacial deposits (E)** was deposited before the final decay of the buried ice. This is confirmed by the inclination of the layer (E) only above the central part of the kettle (Fig. 8–IIa).

Discussion

Lithofacial and structural analysis of sub-sandur deposits of the level (VI) of melt water outflow from the Höfdabrekkujökull in the southernmost part of its forefield has revealed: (1) a hitherto unknown range of the Höfdabrekkujökull at the fossil line of the end moraine; (2) a spatial succession of deposits and their temporal interrelationship which suggests the nature of deglaciation in the area and (3) the frontal and subglacial development of glaciotectonic structures.

In terms of the model of the recession of environmental succession (Brodzikowski, 1982; Eyles, 1983), the spatial chain of the deposits formed during the Höfdabrekkujökull ought to be different in respect of the lithofacies developed (Olszewski, 1976). Locally at least, a more stable morphological surface was preserved, especially on the distal slope of the moraine, its existence being confirmed by the layers which contain organic matter (Fig. 3). Deposits of the distal margin of the fossil end moraine form a fold the axis of which is parallel with the direction of the glacier movement (Fig. 5A, B–B_1, C). Its formation may be associated with soil removal from the fissure which is perpendicular to the glacier front along the line "a" in Fig. 5 (Kotowski, 1989). Similar anticlines, almost perpendicular to the edges of glaciotectonic ice rafts, were determined by Kraiński (1989) in the Dalkowskie Hills. Less likely, owing to the openness of the situation, seems to be the concept of "edge glaciotectonic" as proposed by Brykczyński (1982). The analysed fault would have been a deformation in the peripheral zone of the lobe, i.e. an effect of its frontal dynamics. However, in this case, the glacier must have been moving from the south as shown by the course of the fold axis B_1 (Fig. 5B, C), even if, later, it was further disturbed. Hence, the fold structure is the oldest glaciotectonic deformation (a) in the area investigated. It is not associated with the active glacier and is slightly older than the formation of the fossil (at present) end moraine ridge, which, at that time, was not disturbed (Figs. 5A, 9).

The rhythmically bedded sands and tills deposited below the fold (on the distal side of the fossil structure of the moraine ridge) (Fig. 3A) abruptly lose their continuity upwards and towards the moraine. They are distinctly separated and their internal structure shows no sign of disturbances. This may suggest frost penetration during the deformation process. Abrupt shearing of the deposits may indicate the existence of an old, rough surface of separation and slip formed by the pushing of the fossil end moraine deposits (Owen & Derbyshire, 1988). The formation of an anticline (B_2) in the fossil end moraine which eventually led to its thrusting (Fig. 5 A–t.p., B, C), was the result of such deformation (I), which was connected with frontal dynamics of the Höfdabrekkujökull (NW thrust). If the contact of tills and sands was more horizontal and advanced glacial shearing occurred, the development of a thrust plane would have resulted in the initiation of deformations leading to a possible mylonitization of the ceiling (a_2), as described by Kozarski and Kasprzak (1992).

The structure of the declined fold in the central part of the fossil frontal moraine may be associated with deformation caused by bending and with overthrusting (I) of the deposits from the glacier forefield (Fig. 9) (Möbus, 1989; Dadlez & Jaroszewski, 1994; Wiśniewski *et al.*, 1996).

On the proximal side of the frontal moraine, a deformational series (B) has been formed. It is a glaciodynamically – *in statu nascendi* – rejuvenated, facially-immature basal moraine (Croot & Sims, 1995), braided on the proximal slope of the anticlinal arch of the older frontal moraine (Figs. 4, 5, 9). Within the series B of wedges structures and glacial marks reflect the glacial pressure on the bedrock (B) during the next thrust of the glacier (advance II), i.e. from the NW direction. At the contact with the proximal part of the fossil moraine, they indicate the zone of the greatest possible deformation (Rotnicki, 1971, 1976a, 1976b; Brodzikowski, 1978). Glacial pressure was not as strong under the glacier sole and the transported complex (B) was such that no significant zone or shear plane was formed on the border (B/A). Only a few deposits were added at the contact. That deformation episode (II) must be regarded as postdating the fossil deformation of the end moraine (I). This is confirmed by a comparison of the anticline and the fossil end moraine with the B_2 axis; certainly, there is a mutual meridian of poles of this structure of marks on a stereogeographic grid (northern hemisphere, Schmidt's projection, Fig. 5B). Even if it was

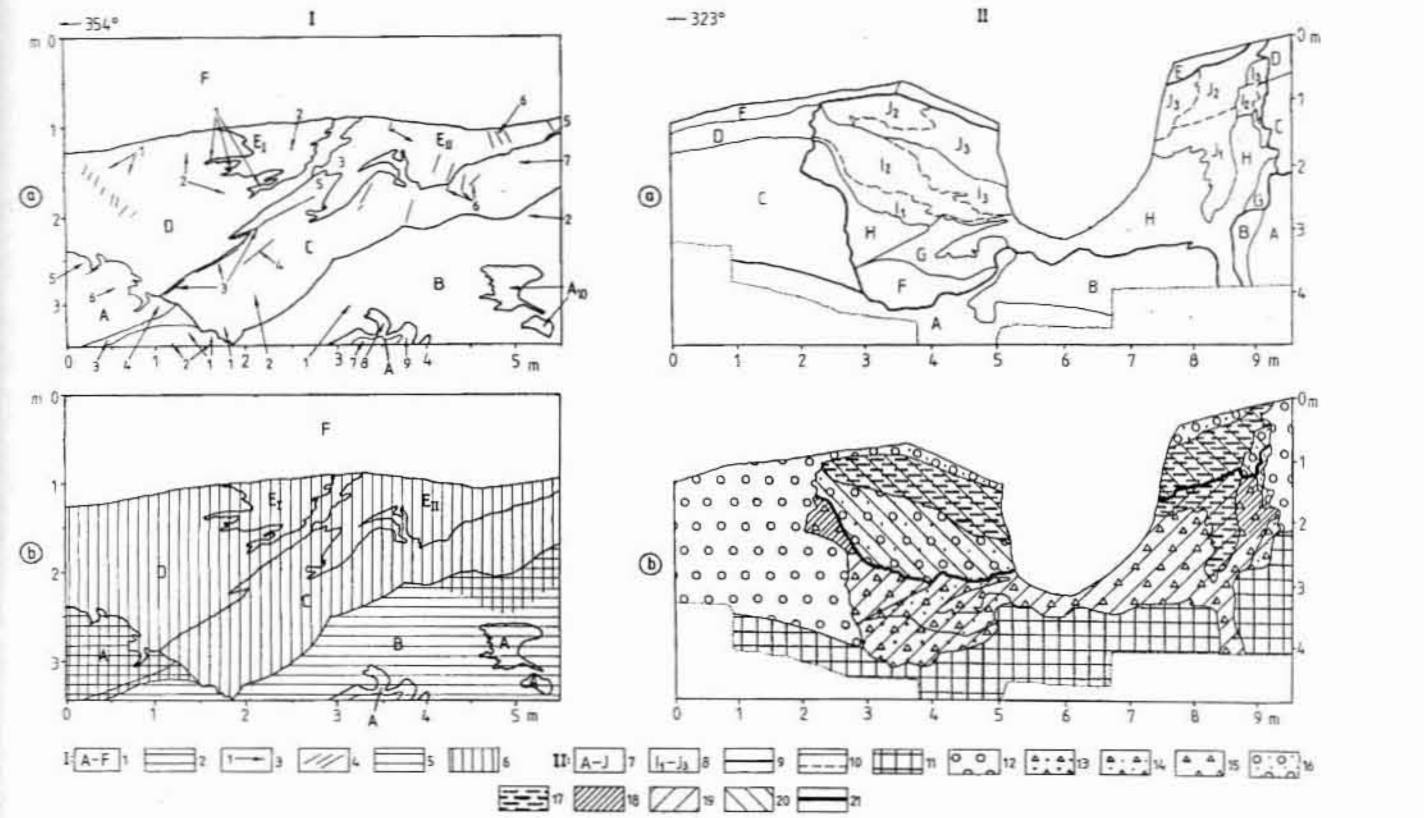

Fig. 8. I: Lithofacial sets and series (a) and generations of the sub-sandur deposits disturbances (b) in the third outcrop (Fig. 6); II: Lithofacial complexes and sets (a: A–J) and generations of the deposit disturbances and their lithology above the substratum sediments (b) in the fourth outcrop (Fig. 7).

I: (a): 1 – lithofacial sets; 2 – boundaries of the lithofacial sets and sub-sets; 3 – location of selected deformation structures; 4 – faults; (b): 5 – the zone of the glaciotectonic disturbances of the older generation (I); 6 – the zone of the glaciotectonic disturbances of the younger generation (II);
II: (a): 7 – lithofacial complexes and sets; 8 – lithofacial sub-sets; 9 – boundaries of the sedimentary series; 10 – boundaries of the lithofacial sets and sub-sets; (b): 11 – older sub-sand substratum; 12 – sandur deposits; 13 – sandur-melt-out diamicton (F); 14 – fine- and medium-grained sands with diamicton (G); 15 – basic diamicton (H); 16 – heterogeneous sands with diamicton (I); 17 – fine-grained sediments of the basin sedimentation (J); 18 – rimmed structures of the basic diamicton lithofacial set; 19 – the zone of the disturbances of the older generation within the sediments where diamicton predominates; 20 – the zone of the disturbances of younger generation within sandy and silty sediments of the upper part of the kettle; 21 – litho- and morphogenetic boundary and generation of the deformational structures within the sediments which fill the rim-normal kettle.

a sedimentation feature of the till, its projection along the line of the big arch, oblique to the B_2 axis, demonstrated the existence of two generations (I, II) of glaciotectonic dislocations (Dadlez & Jaroszewski, 1994), i.e. the formation of younger glacial marks (II).

In terms of the type and shape, there is a great similarity between the deformed sediments on the opposite slopes of the fossil moraine rampart and the thrust end moraines present in SW Sweden which are controversially termed "oscillatory crossed glaciotectonic ridges" by Fernlund (1988). However, the deformational structure of the central part of these two supposedly comparable forms, i.e. the fossil end moraine rampart in Iceland and the Halland coastal moraines in Sweden (Fernlund, op. cit.; Figs. 2, 3, 10) is very different. Both the lithology and the inner texture of these forms are different. Other essential factors are as follows:
– the pressure of the thrust of the glacier or ice sheet (a);
– the distance of the thrusting pressure/mass in relation to the object of thrusting (b);
– the environment of the glacial contact (c) which is subglacial in Sweden and marginal-lateral in Iceland; the latter influenced the deformations of both fossil slopes of the end moraine, which is now buried under the sandur in the forefield of the Höfdabrekkujökul.

In the area of the third exposure (Figs. 6, 8–I) an older generation of glaciotectonic disturbances of the diamicton (B) and the older fluvioglacial deposits (A) can be identified with the thrust (I). It is manifested by the horizontal axial planes of the folds. Small-scale drag folds, formed below the shear planes and the secondary structures of disturbed, disrupted folds and fluidal lenses (Rotnicki, 1971) are associated with the restarted movement (A). Structures similar to the above are interpreted by Banham (1988) as an effect of a younger phase of glaciotectonic deformations in the Contorted Drift of Norfolk.

The zone of maximal and youngest (II) glaciotectonic discordances contains the proximal-floor part of the southern packet (EII) of the basinal deposits

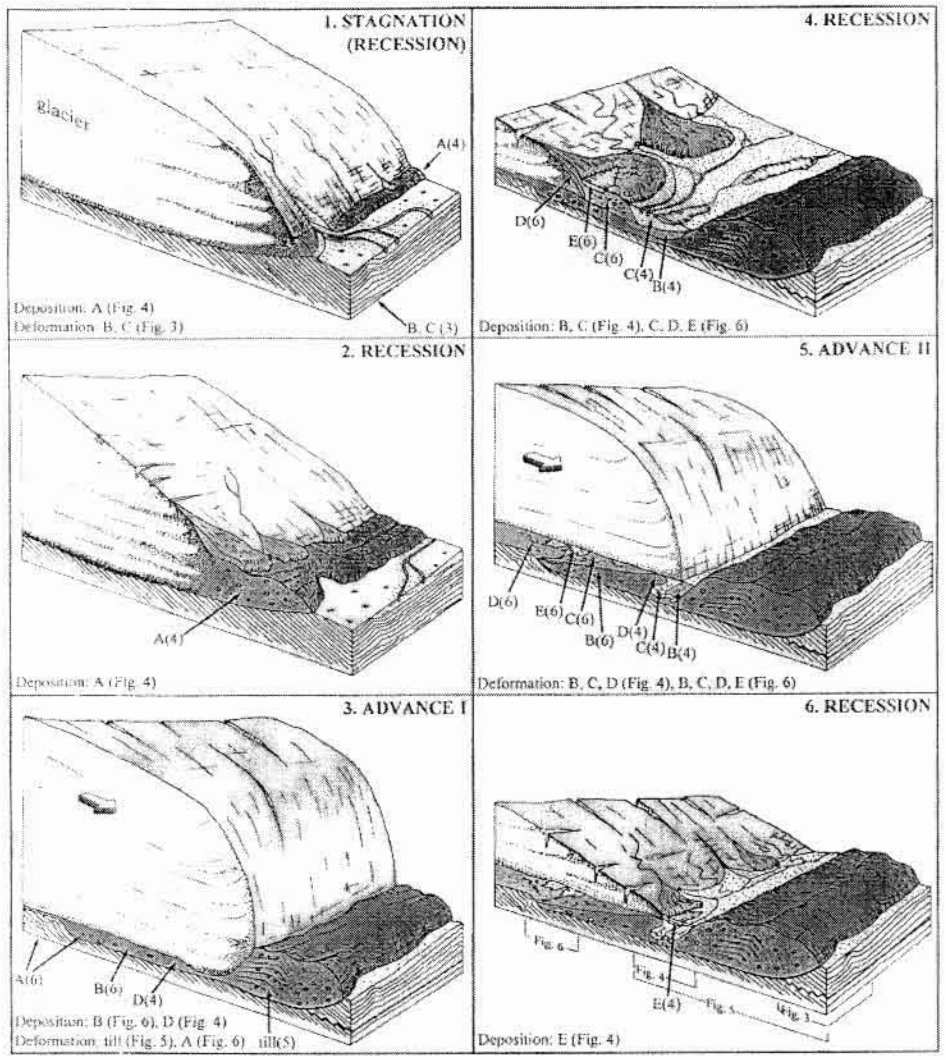

Fig. 9. The scheme of development of deposits and deformation structures formed before deposition of surface sandur deposits including older phases of recession and thrusting of the Höfdabrekkujökull (letters A–E mark deposits and their structures; digits 3–6 mark their location in particular figures).

(Fig. 8–I, EII). This confirms the relative incompetence of the soft basin deposits (E) in respect of glacial erosion and deformation. Such processes and their structural results were described by Ingólfsson (1987) in Late Vistulian glacio-marine sediments which occur under "deformation till" in western Iceland, by Motyka (1995) in Alaska and by Wiśniewski *et al.* (1996) in modern lacustrine-glacial sediments in south-eastern Iceland, where they have been deformed under the sole of the Skeidarár Glacier. These processes were also discussed by Brodzikowski (1978, 1982). The separation of (EI) and (EII) packets and the inclined support of the disturbed part (E3) by the underlined slip assemblage of fold structures (C5) of the thrust sediments indicate the considerable strength of the thrust (Figs. 8, 9).

Boulton *et al.* (1982) and Björnsson (1996) emphasise the considerable erosional role of the edge of Breidamerkur Glacier and substratum erosion in its submarginal zone during the Little Ice Age advance. The pattern of the end moraines of the maximal limit and older moraines (especially the recessional forms, which developed after the edge retreat of 1890) shows that before and during the time when depositional processes began to operate in the marginal part, the erosional process transformed into a sedimentation process which was capable of producing deformational structures. There can be no doubt that the glacier produced such structures in the proglacial zone of the terminal depression; this is certainly the case where the glacier edge overtook its limit of 1965 by some 200 m as compared to its position in 1974 on the northern shore of the crypto-depressional Jökulsárlón Basin (Boulton *et al.*, 1982). A similar situation, where new glaciotectonic structures were produced where wet lacustrine deposits were overridden, was described by Wiśniewski *et al.* (1996) in the marginal zone of Skeidarájökul. The same principle (of dynamic glaciogenic cause and glaciotectonic effect of the deformation of lacustrine sediments) is valid in the case of the deformational structures of the Höfdabrekkujökull forefield (Figs. 5, 6, 8–I). Also, when taking this principle into account, structural development of a complex "transitional" zone where two generations of glaciodynamic deformations overlap (Fig. 8–Ib, C) (the younger of which, i.e. the plane cutting structure was initiated within and beyond the convex parts of the earlier fold structures; Fig. 8–Ib, A). It is also obvious that, within soft

sediments, if similar, spatial deformation of the basal zone of the glacier is possible *in statu nascendi*, apart from the roche moutonnee convex part of the bedrock with the cover of glacio-morainic floor layers of the glacier. This was observed by Olszewski (1991) in respect of the Casement Glacier in Alaska.

The model (Fig. 9) presents the location and relation of the main processes and environmental situations of depositions and deformations (the oldest marginal one and two post-recession glacial thrusts I and II) which took place before the deposition of surface sandur material. According to Heim (1983) and Krüger (1994), from age considerations of the so called main moraines (a), their spatial location in respect of the fossil form of the end moraine (b) and the nature of glaciotectonic dislocations of older deposits from various environments (c), it may be assumed that the fossil marginal form was formed at least before the end of the Little Ice Age. Together with the deformation structures discussed here, it could denete particular episodes of the glacier front behaviour during the earlier phases of the Little Ice Age.

Lithofacial analysis of the deposits filling up the younger melt-out form leads to a realisation of the palaeomorphological structure of the rimmed mound (rim) of the diamicton. Such stony-diamicton rim microforms on the surface of the higher sandur from the jökulhlaup of 1918 were described by Maizels (1991, 1992). This type of form was earlier described by Fuller (1914), after Fairbridge (1968), who in New York State, distinguished three lithological types of rim on sandurs: **"outwash rim"**, **"till rim"**, and **"boulder rim"**. Meanwhile, Thwaites (1935) called this type of form a **"kettle rim"**, and emphasised the importance of debris, moraine and boulders, melted out from ice, in its structure. However, their fossil, residual preservation under the smooth sandur surface at a depth of about 1 m has not previously been postulated (Figs. 7, 8–IIb) in the form of palaeomorphic structure of diamicton circular rim.

Conclusions

In the outcrop of the "green" sandur (VI), a differentiated palaeomorphological and lithogenetic situation has been recognised under the horizontal surface fluvioglacial deposits. Based on field investigations of the palaeorelief and sub-sandur deposits, the following palaeoforms have been distinguished: (a) palaeoforms of the proglacial (marginal) channels of the meltwater network, (b) palaeoform of the end moraine rampart and (c) palaeoform of the broad end depression filled by basin water from the earlier glacier recession period.

Conclusions concerning the glaciotectonically deformed deposits which accompany the fossil end moraine rampart are as follows:

1. Before the origin and geomorphological development of the end moraine ridge (the present-day fossil position) which is built mainly from till, on its forefield deformational structure of the fold type had developed (Figs. 3, 5: B–a, A–a) within sands (and also partly within silts) of substratum (B–C). We associate its origin with bottom squeezing of ground into the marginal and transverse crevasse's of the glacier edge.

2. Post-recessional advance (I) of the Höfdabrekkujökull (Fig. 9: advance I), which reached to the earlier end moraine ridge caused its folded deformation (anticline, fairly symmetrical, in the central part of the rampart) as a result of the pushing. The thrust plane and the shear plane developed at that time in the zone of the distal slope of the fossil end moraine (Fig. 5).

3. During the second advance of the glacier (II) on the proximal slope, the till was more resistant. This caused collision and contact structures at the boundary of the glacially transported mass and the resistant mass of the end moraine (Fig. 9: advance II). The deformations which took place in several stages of the successive process created fold structures and then rounded them and dislocated them horizontally in the distal direction. These processes involved the deposits of the submarginal glacial contact zone.

The following conclusions may be drawn concerning the deposits which filled the proglacial depression and associated facies:

1. Within the soft, fine-grained basin deposits and associated coarser diamicton facies, a clear succession of deformational and glaciotectonic structures may be identified. There is also a **transitional zone where these structures overlap** which belongs to two glacio-deformational phases (I, II).

2. The characteristic features of glaciotectonic structure creation are a chronological succession (stage by stage) and a hierarchy arrangement (primary and secondary). A deformation system developed which consists of: shear zones – shear faults – fold-drag deformations which accompany faults – small reverse faults. A similar vertical succession of these structures may also be identified.

3. The depth range of the sediment deformation where overridden by the glacier is marked by shear faulting which is accompanied by underlying folds. The depth range of glaciotectonic deformations involves glacial and semi-glacial sediments and also those deposits of the deeper substratum which were not deformed earlier.

Within the sandur with holes, the deformation structures of the sediments which fill the kettle have been determined. These include the following:

1. The sediments of the kettle fill represent different lithofacial sets: melt-out-denudational, basinal and decaying flow of melt water.

2. The origin of the deformation structures, especially those of gravitational nature, should be associat-

ed with: (a) the different rates of ice and morainic loading (as located in the ice mass) melting; (b) the different scales of the ice mass burial by wash-out material; (c) gravitational lability of the melt-out material and instability of the ice mass itself in soft, saturated sediment; (d) solar exposure of the melting ice-sediment walls and (e) zonal character of the denudation processes inside the kettle.

3. The rimmed-mound forms of the kettles originated in the conditions of local restraint or decay of the sandur flow and this is why they were preserved as palaeomorphological structures of rimmed mounds (rims) of diamicton. This takes place despite the resumption of any flow of glaciofluvial water; the flow is not able to destroy them and can only bury them.

4. The rimmed mound develops in stages as a slide-dropping and polygenetic microform (appropriately in the lower and upper part of the rimmed form). There is also an evolution of this form in respect that it changes from the phase of the subareal and partly sublacustrine form to the fossil form of the rimmed mound. This takes place owing to the successive stages of the development of melt-out and even sandur deposits, the youngest, local accumulation of which changes non-destroyed mound forms into fossil structures. This newly recognised type of kettle may be called the **rimmed-normal kettle**.

5. The genetic complexity of the rimmed-normal kettle is also reflected in the vertical arrangement of the fill deposits which are oriented in different, often opposing directions and are completely dissociated from the frontal thrust of a single tension. This changeability is associated with increase in the thickness of lithofacial sets; it starts from the eccentric growth towards the peripheries of the negative form in the lower part of the series and progresses to the concentric increase of the thickness in the younger complexes of the series (Fig. 8–IIb).

6. The extremely small range of the glaciostatic, melt-out and subsidence and also denudation-gravitational deformations (which comprise the sedimentary series which fill the kettle) and, indirectly associated with them, the limited structural modification of the sandur margins is not in genetic, spatial or stratigraphic accordance with the model of a dynamic, frontal-zonal glaciotectonic deformation of the older sub-sandur sediments.

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The north-eastern boundary of the Baikal rift zone

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Abstract: The Baikal rift zone is defined on its NE side by the Olyekma-Amur system of NW/SE-striking transverse lineaments. Transverse faults also separate the north-eastern part of the rift zone into the Chara and Tokko sections. The former shows a typical range of neotectonic forms which include (from NW to SE): the inclined horsts of the Kodar ridge; an axial system of grabens and interbasin faults and a marginal dome. The Tokko section is located within a sector of the Olyekma-Amur system of lineaments. Here, the neotectonic forms are smaller and a degradation of the NW flank of the rift zone has taken place. A large marginal step defines the south-eastern flank. It is considered that these structural modifications relate to changes in the anomalous mantle protrusion near the north-eastern boundary of the rift zone.

Key words: rift, Baikal rift zone, tectonic relief

Introduction

The problem of the extension and closure of the Baikal rift zone has been controversial for many years. Commonly, owing to the general lack of detail concerning the relationships of the different structural elements present in the rift zone, the various interpretations which have been proposed have all been inconclusive. This paper defines the boundary of the rift zone and describes the structural transformations in the regions of closure; we have selected the north-easternmost link in the rift zone, the Chara-Olyekma link, as the field model for our reinvestigation of this problem (Fig.1).

The tectonic relief of the NE part of the rift zone

A study of topographic maps, together with geological and geophysical surveys, permits the elucidation of recent tectonics in this region (Fig. 2). The tectonic relief reflects the form of the neotectonic structures; the patterns of neotectonic forms reflect the spatial relationships of the rift zone elements and also those in adjacent areas. Also, for a full structural analysis of the rift zone, it is necessary to erect a model of the tectonic relief (a summit-incidence surface); the mountain uplifts, with their complex and variable relief patterns, extend alongside the rift valleys which are the main elements of the rift zone structure.

In the Olyekma and Nyukzha valleys, the topography of tectonic relief clearly reflects the north-eastern limit of the rift zone (Figs. 1 and 2). Here lies the sharp change between the complex tectonic relief of the rift zone and the plains which are marginal to the Siberian Platform. In the NE, the rift zone is bounded by NW-striking transverse faults which are part of the extensive Olyekma-Amur lineament zone (Grishkyan *et al.*, 1977; Ufimtsev, 1984). In the Chara-Olyekma region, these can be either transverse faults, as in the Olyekma and Nyukzka valleys, or more complex combinations of tectonic relief forms (Fig. 3). As shown by the pronounced southward displacement of the eastern part of the Kalar gabbro-anorthosite massif (Zorin *et al.*, 1988), cumulative dextral displacements along these, reactivated many times during the Cainozoic, may be as much as 15 km. The westernmost components of this system are the faults which bound the Chara Basin to the NE and which also dissect the Tokko Basin. We refer to these as the Khani-Sulumat and Evonokit-Tokko transverse faults, respectively (Fig. 3). Close to their intersection, the structure of the rift zone is much transformed.

The tectonic relief provides a straightforward answer to the discussed question: does the rift zone continue to the E towards the limits of the Stanovoy Ridge.