Pleistocene sandur deposits represent braidplains, not alluvial fans

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Weichselian sandar in NE Poland show characteristics that are inconsistent with the commonly accepted alluvial-fan-like model for outwash deposition and sandur formation. Analysis of the lithofacies and their vertical and lateral transitions indicates that the Polish sandar developed as braidplains, not as alluvial fans. Analysis of the geomorphic conditions under which modern sandar form, indicates that these conditions (which are characterized by deposition in a narrow belt between ice-covered mountain ranges and the sea) cannot be considered representative of those that prevailed in the geological past when sandar developed as braidplains in confined valleys, to end up in a lowland area where the deposits could spread out further in lateral directions. The latter conditions have been found consistently for all Polish Weichselian sandar that were investigated in much sedimentological detail. This raises the question whether sandar are alluvial fans or not. Because the development of the sandar in NE Poland seems to be much more representative for outwash deposition than the present-day sandar in Iceland and elsewhere, the current alluvial-fan-like sandur model – based on the fairly exceptional present-day situation with deposition in a narrow belt – should therefore be replaced by the braidplain-like sandur model – based on deposition in a valley and in a wider lowland area in front – that has been established for the Polish examples.

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A sedimentary body deposited by proglacial waters, i.e. ablation streams flowing away from melting glaciers (Fig. 1), is commonly named ‘sandur’. This term stems from Iceland, where it means ‘sandy-gravelly area formed by proglacial streams’. The Icelandic plural form is ‘sandar’, which is the term we use in the following as advised by the Editor-in-Chief, Professor Jan Piotrowski, and by Dr. Andrew Russell, who kindly reviewed the original manuscript. The sediments building up a sandur are commonly called ‘outwash deposits’.

Outwash deposits cover large areas in the mid-European lowlands, commonly in combination with glacial till. Where till is absent, sandar may be used for a reconstruction of phases in the development of glaciations. As a rule, however, they are lithologically fairly monotonous and of less help for palaeogeographic reconstruction than other glacigenic facies.

The years from 1972 to 1987 can be considered as a ‘golden period’ for the study of both modern Arctic and ‘fossil’ Pleistocene sandar (see, among others, Klimek 1972; Gustavson 1974; Rust 1975; Ruegg 1977; Allen 1982; Fraser 1982; Fraser & Cobb 1982; Hammer & Smith 1983; Cherven 1984; Landvik & Mangerud 1985; Smith 1985; Dawson & Bryant 1987). Sedimentological models of ‘wet’ alluvial fans and braided fluvial systems were established at that time (Church 1972; Boothroyd & Ashley 1975; Church & Gilbert 1975; Boothroyd & Nummedal 1978), and they are still in use. Some interesting studies on outwash facies were also published in the 1990s (Fraser 1993; Krzyszkowski 1993; Olsen & Andreasen 1995; Aitken 1998); of particular interest is Maizels’ (1993) subdivision of sandar in accordance with their hydrological regimes. The occurrence of catastrophic meltwater floods – the jökulhlaups in the foreland of Icelandic glaciers, which commonly strongly influence the characteristics of sandar, also received much attention (Maizels 1997; Russell & Marren 1999; Russell & Knudsen 1999). The fan-like outwash models that emerged from studies of present-day Icelandic and Alaskan sandar are still very popular among earth scientists.

Weichselian (= Vistulian = Wisconsinan) sandar in Poland (Fig. 2) show characteristics – among others the spatial distribution of lithofacies and, in particular, the vertical successions – that reflect much less well developed proximal-to-distal facies transitions than

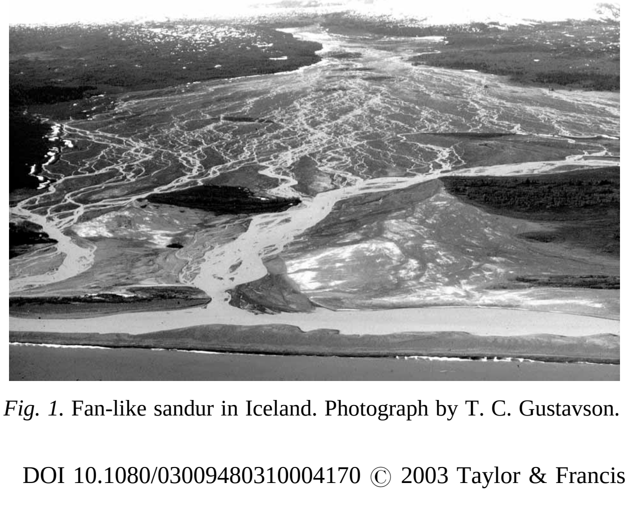

Fig. 1. Fan-like sandur in Iceland. Photograph by T. C. Gustavson.

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fans do. Consequently, they do not fit in the sedimentary models that are based on analysis of recent sandar. The apparent discrepancy between the Polish Weichselian sandar and other recent sandar underlines that relatively little is known about the formation of these sedimentary bodies. A better understanding of the sedimentary nature of Pleistocene lowland sandar has been the main reason for carrying out this study. One of the questions to be addressed is therefore how regular – and how representative – the configuration of the present-day 'model' sandur is.

Aims and objectives

The aim of the present contribution is to analyse and model the sedimentological characteristics of outwash deposits in NE Poland. In this region, these deposits form bodies that are morphologically well preserved and – due to quarrying – well exposed. The region was covered by an ice sheet during the Weichselian glaciation. During the ablation of this ice sheet, a system of sandar formed. These are the largest Pleistocene outwash plains in Poland, and their deposits have been studied in 16 outcrops (Fig. 2).

In more detail the present study had the following objectives:

- recognition of the most typical lithofacies in the outwash deposits, the interpretation of their depositional mechanisms, and the reconstruction of their depositional environment;
- analysis of the cyclicity that these deposits show, i.e. the distinction of the most common vertical sequences and interpretation of their depositional parameters;
- inventory of the most common vertical facies transitions within thick successions;
- the construction of a general sedimentological model for the Pleistocene sandar in NE Poland;
- comparison of these sandar with other glaciomarginal landforms addressing the question whether these sandar are alluvial fans or not.

Outwash lithofacies

We distinguished lithofacies on the basis of qualitative

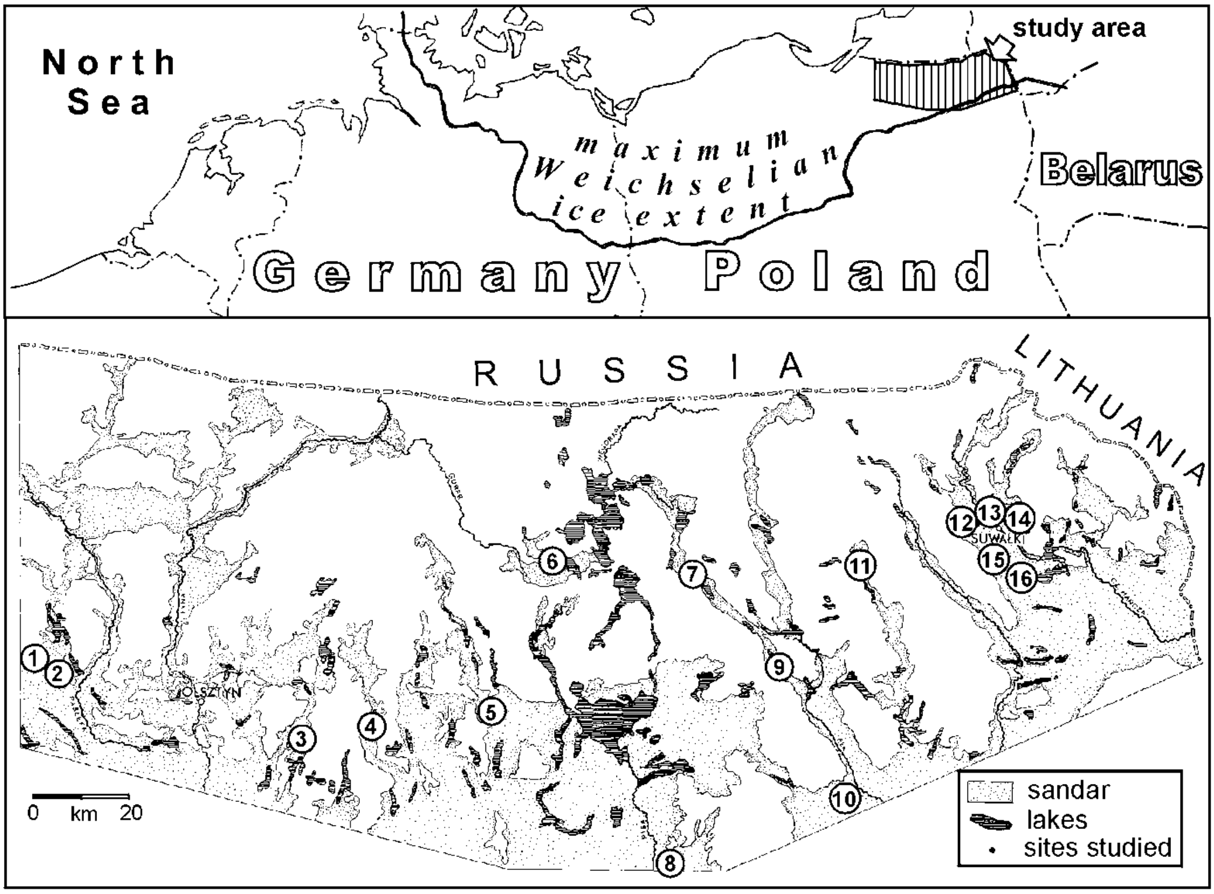


Fig. 2. Study area showing the extent of the Weichselian ice sheet and the location of the investigated sites (1–16).

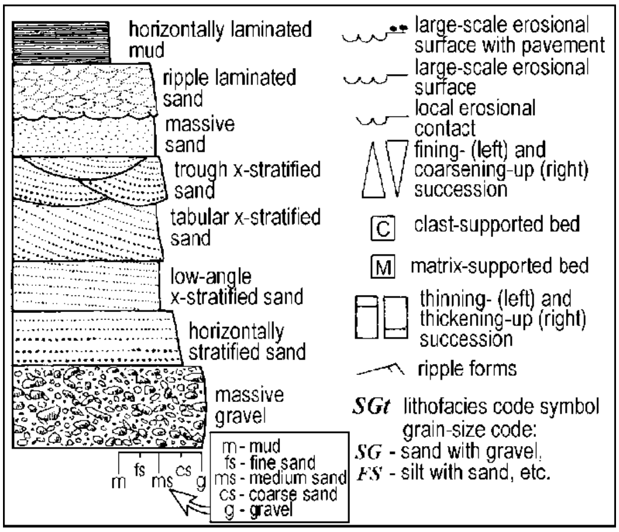


Fig. 3. Symbols used in Figs 4, 5 and 7.

and quantitative sedimentological analyses of all the investigated sites. Two main groups of facies are present: deposits of gravel-bed (proximal) streams (coded *P*) and deposits of sand-bed (distal) streams (coded *D*).

These two groups are considered by us as facies associations. Together, they comprise nine lithofacies. The individual facies are described and interpreted in the following sections. In order to avoid long descriptions of lithofacies, only the most prominent features are mentioned. More details are provided in the related figures, which show ‘idealized’ successions, based on observations of all exposures. The various symbols used in drawings are explained in Fig. 3. As a consequence, the facies sections are almost entirely devoted to interpretations of the depositional conditions, and on discussions of their depositional (sub)environment on the basis of the geological context in which the various facies occur.

Facies association *P*

This facies association comprises four lithofacies, coded *P-1* through *P-4*, respectively. Distinguishing between these lithofacies is fairly easy on the basis of grain-size characteristics; these will be detailed in the following sections.

Facies *P-1*

Facies *P-1* is the coarsest of all sandur deposits and contains the thickest beds of all deposits in the study area. It comprises cobble beds with large-scale tabular cross-stratification intercalated with gravel/boulder sheet-like beds (Fig. 4).

The massive, coarser-grained (cobble and boulder) sheet-like beds *Gm* are interpreted as channel lags or diffuse gravel sheets in the sense of Hein & Walker (1977). They must have formed during catastrophic floods. The flow intensity was so high during the maximum of these floods that gravelly sediment was deposited from a traction carpet under upper plane bed conditions (so-called high-concentration bedload *sensu* Sohn 1997). In addition, a few other structures were found that resemble those produced experimentally by Alexander *et al.* (2001: fig. 9) from antidunes. This underlines the high flow intensity.

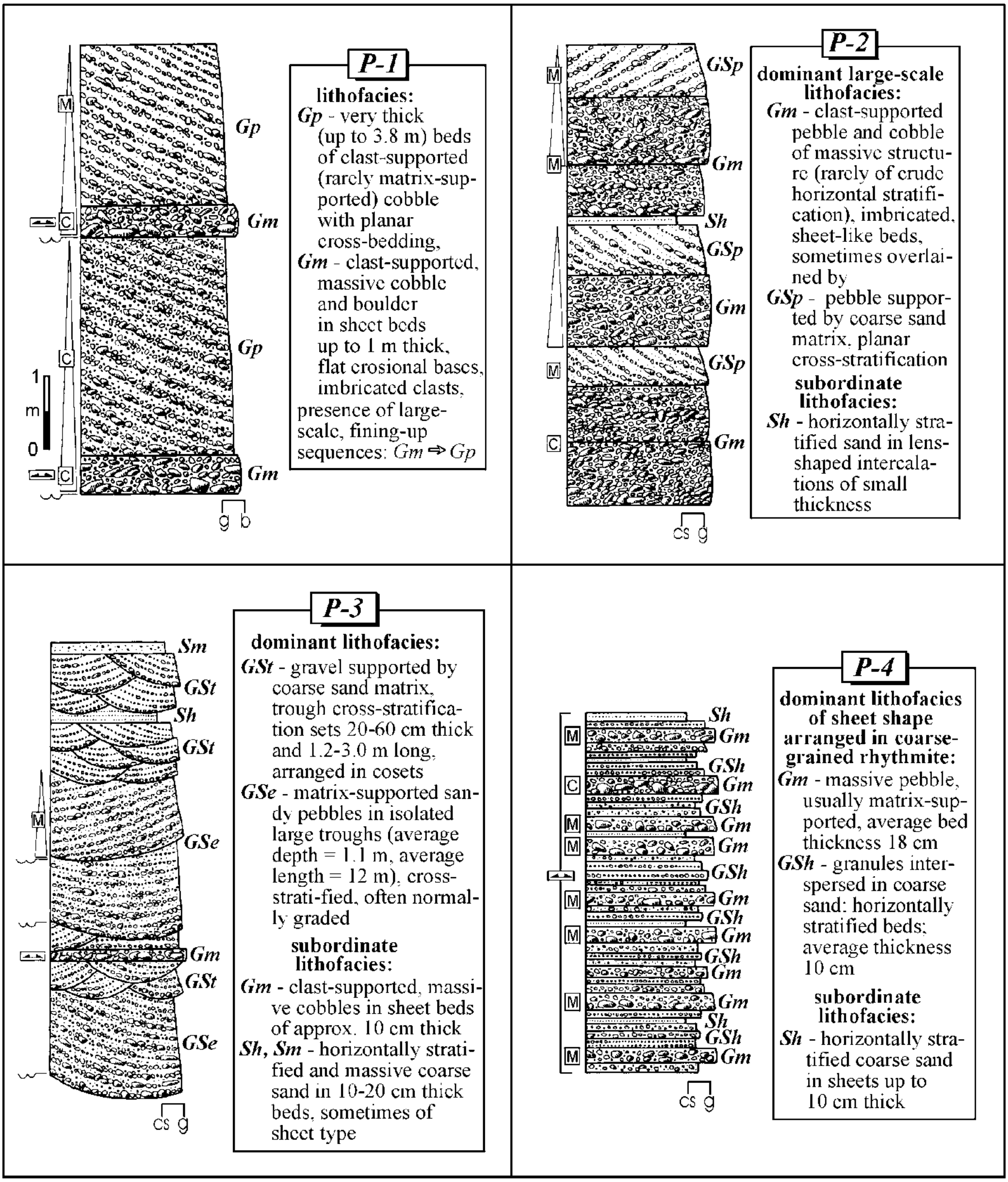
The coarse-grained cross-bedded sets *Gp* must result from down-current progradation of large gravelly foreset bars. Owing to the considerable depth of the channels (cross-stratified gravel units, up to 4 m thick, represent bars that, by definition, were covered by water), lower flow-regime conditions prevailed and bars formed. Similar large-scale gravelly foreset bars have been described by Rudoy & Baker (1993); they resulted from a catastrophic flood after sudden drainage of a glacier-dammed Pleistocene lake. Bars due to catastrophic flows, so-called ‘expansion bars’ (Baker 1973), contain an identical lithofacies *Gp* (O’Connor 1993).

Taking into account the geological context, it must be deduced that the deep channels with high-competence currents were situated close to melting ice, and were shaped during catastrophic ablation floods following the drainage of supraglacial, englacial or subglacial reservoirs; the deposits thus must be attributed to high discharge events, possibly (but certainly not necessarily) jökulhlaups. Such torrential, short-lived, high-discharge flows are characteristic of the proglacial outflow zone (cf. Church 1988; Maizels 1989, 1993; Russell & Knudsen 1999); values of 400–600 m^3 s^{-1} have been mentioned for the Skeidarar, and values of over 1000 m^3 s^{-1} occur in braided systems in Iceland (A. Russell, pers. comm. 2002). Comparable sediments of catastrophic ablation floods have been described earlier for Pleistocene terminoglacial fans (Zielinski & Van Loon 1999). On the basis of the sedimentological characteristics, facies *P-1* is a typical deposit of the most proximal zone of an outwash plain and represents the subenvironment with the highest energy level and aggradation rate.

Facies *P-2*

Facies *P-2* (Fig. 4) is a common constituent of the coarse-grained outwash deposits. It is dominated by massive gravel (*Gm*) intercalated with tabular cross-beds of sandy gravel (*GSp*).

The massive gravels are typical of coarse, longitudinal bars. They are comparable with the ‘bar core facies’ as distinguished by McDonald & Banerjee (1971) and with the ‘channel-bar sheets’ described by Nemec & Postma (1993). The massive structure and the



P-1

lithofacies:

Gp - very thick (up to 3.8 m) beds of clast-supported (rarely matrix-supported) cobble with planar cross-bedding,

Gm - clast-supported, massive cobble and boulder in sheet beds up to 1 m thick, flat erosional bases, imbricated clasts,

presence of large-scale, fining-up sequences: *Gm* ⇒ *Gp*

P-2

dominant large-scale lithofacies:

Gm - clast-supported pebble and cobble of massive structure (rarely of crude horizontal stratification), imbricated, sheet-like beds, sometimes overlained by

GSp - pebble supported by coarse sand matrix, planar cross-stratification

subordinate lithofacies:

Sh - horizontally stratified sand in lens-shaped intercalations of small thickness

P-3

dominant lithofacies:

GSt - gravel supported by coarse sand matrix, trough cross-stratification sets 20-60 cm thick and 1.2-3.0 m long, arranged in cosets

GSe - matrix-supported sandy pebbles in isolated large troughs (average depth = 1.1 m, average length = 12 m), cross-strati-fied, often normally graded

subordinate lithofacies:

Gm - clast-supported, massive cobbles in sheet beds of approx. 10 cm thick

Sh, Sm - horizontally stratified and massive coarse sand in 10-20 cm thick beds, sometimes of sheet type

P-4

dominant lithofacies of sheet shape arranged in coarse-grained rhythmite:

Gm - massive pebble, usually matrix-supported, average bed thickness 18 cm

GSh - granules interspersed in coarse sand; horizontally stratified beds; average thickness 10 cm

subordinate lithofacies:

Sh - horizontally stratified coarse sand in sheets up to 10 cm thick

Fig. 4. Idealized sections through the facies *P-1* through *P-4*.

crude horizontal stratification of the gravel point to deposition from supercritical or transitional flow (under upper-stage plane-bed conditions). The channels must have been deep, *c.* 2 m, as can be assessed from the bed thickness. The cross-stratified gravelly-sandy beds, *GSp*, are thought to represent deposition during waning floods, when lower-energy conditions resulted in more sinuous flows that favoured lateral growth of the finer-grained slip faces of longitudinal bars (cf. Smith 1974; Rust 1978; Fraser 1993).

The frequency of erosional contacts between the various beds in this lithofacies is low, which is interpreted as a depositional record of streams overloaded with sediment, typical of the sandur zone close to the ice-sheet margin. The combination predominantly of gravels derived from longitudinal bars, the large thickness of the individual beds, and the scarcity of sand layers makes this facies comparable with classical successions of proximal braided-river alluvium: G_{II} sensu Rust (1978) and Scott River (Miall 1977, 1978). The depositional environment of this facies must have been similar to that of type-2 braided rivers in Miall's (1985) classification, i.e. a high-energy gravel-bed alluvial channel dominated by accumulation of longitudinal bars.

Facies *P-3*

Facies *P-3* consists of sandy gravel with trough cross-stratification (Fig. 4). It shows more abundant erosional contacts between successive sets, however, than do facies *P-1* and *P-2*.

The matrix-supported sandy gravel of the trough cross-stratified beds (*GSt*) is typically a channel deposit representing sinuous dunes. Presumably these bedforms covered the entire bed of the channels during mean discharge. During torrential floods, however, gravelly dunes could form (cf. Khadkikar 1999), but large troughs also developed in the channels and became filled with sandy/pebbly beds (*GSe*) under these conditions (Fraser & Bleuer 1988). This indicates deep, high-energy channels. The troughs show the same characteristics as the so-called 'scour pools' of Siegenthaler & Huggenberger (1993).

The abundant presence of erosional contacts suggests that deposition took place by streams of which the sediment load was in equilibrium with the flow energy. Since the adjacent sedimentary units show that this facies still forms part of the proximal zone, the diminished sediment load (compared to the load in facies *P-1*) indicates that this facies must have formed at the distal end of the proximal zone, where the meltwater streams were no longer overloaded with particles because much material had already been deposited upstream; a downstream decrease of the sediment load is a well-known feature of proglacial streams (see, among others, Klimek 1972 and Maizels 1983b). The facies resembles the 'White Channel' lithotype dominated by thick *Gt* units, as described by Morison & Hein (1987), which they interpreted as a high-energy, torrential braided-river alluvium.

Facies *P-4*

Facies *P-4*, consisting of massive pebbly beds alternating with horizontally stratified sandy gravel (Fig. 4), differs distinctly from the other gravel-dominated facies for the following two reasons: (1) all beds are distinctly thinner (less than 20 cm), and (2) pebbly and sandy sheets alternate. The dominant lithotypes are *Gm* and *GSh*, whereas *Sh* is of minor importance.

The relatively thin but laterally extended beds are indicative of shallow sheetflows (Nemec & Muszynski 1982; Blair 1987). The alternation of more and less coarse-grained beds points to streams that were characterized by rapid variations in discharge. This feature is characteristic of proglacial currents fed by ablation of a nearby ice mass. It is most likely that the glaciomarginal currents occurred almost exclusively during summer and that they lasted no longer than one day to a few days (cf. Klimek 1972; Hammer & Smith 1983; Russell & Marren 1999). Although rhythmites from deep-water settings have been described in the literature, e.g. from the 1996 jökulhlaup on the Skeidararsandur (Russell & Knudsen 1999), we did not find any indication of a 'deep' environment. We therefore conclude that the strong floods, which were mainly supercritical during the short time of occurrence, were fairly shallow.

Facies association *D*

This facies association comprises five lithofacies, coded *D-1* through *D-5*, respectively. The distinction between these lithofacies is less simple than the distinction between the proximal facies. This is because of the spatially less well separated positions, indicating closer interrelationships.

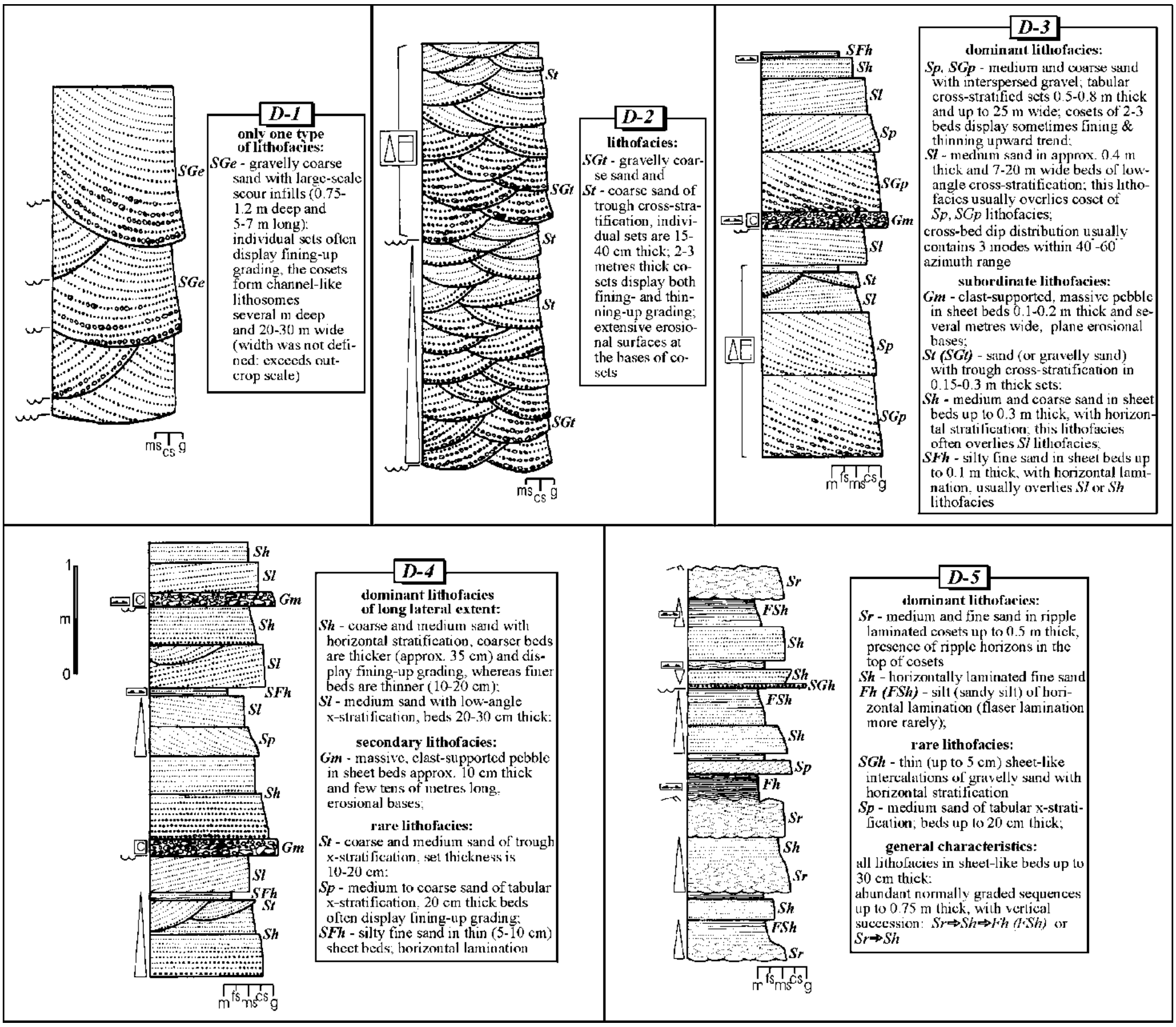
Facies *D-1*

Facies *D-1* (Fig. 5) consists exclusively of coarse, gravelly sands, as fillings of large-scale troughs (*SGe*). It usually coexists with facies *D-2*.

The deep troughs must have been eroded by strong currents during peaks in ablation. This facies formed in deep (i.e. first-rank) channels of a proglacial braided-river system. Braided channels of this type are, as a rule, 1–6 m deep and several tens of metres wide (McDonald & Banerjee 1971; Cant & Walker 1978) and their sediments are most commonly dominated by large-scale trough cross-sets (Fielding & Webb 1996; Hjellbakk 1997). As pointed out in the next section, the palaeohydraulic parameters of facies *D-1* conform the data fairly well.

Facies *D-2*

Facies *D-2* consists of gravelly sand and sands with trough cross-stratification (Fig. 5) and resembles facies *D-1* in several aspects. Both comprise coarse sands in



D-1

only one type of lithofacies:

SGe - gravelly coarse sand with large-scale scour infills (0.75-1.2 m deep and 5-7 m long): individual sets often display fining-up grading, the cosets form channel-like lithosomes several m deep and 20-30 m wide (width was not defined: exceeds outcrop scale)

D-2

lithofacies:

SGt - gravelly coarse sand and

St - coarse sand of trough cross-stratification, individual sets are 15-40 cm thick; 2-3 metres thick cosets display both fining- and thinning-up grading; extensive erosional surfaces at the bases of cosets

D-3

dominant lithofacies:

Sp, SGp - medium and coarse sand with interspersed gravel; tabular cross-stratified sets 0.5-0.8 m thick and up to 25 m wide; cosets of 2-3 beds display sometimes fining & thinning upward trend;

Sl - medium sand in approx. 0.4 m thick and 7-20 m wide beds of low-angle cross-stratification; this lithofacies usually overlies coset of *Sp, SGp* lithofacies;

cross-bed dip distribution usually contains 3 modes within 40°-60° azimuth range

subordinate lithofacies:

Gm - clast-supported, massive pebble in sheet beds 0.1-0.2 m thick and several metres wide, plane erosional bases;

St (SGt) - sand (or gravelly sand) with trough cross-stratification in 0.15-0.3 m thick sets;

Sh - medium and coarse sand in sheet beds up to 0.3 m thick, with horizontal stratification; this lithofacies often overlies *Sl* lithofacies;

SFh - silty fine sand in sheet beds up to 0.1 m thick, with horizontal lamination, usually overlies *Sl* or *Sh* lithofacies

D-4

dominant lithofacies of long lateral extent:

Sh - coarse and medium sand with horizontal stratification, coarser beds are thicker (approx. 35 cm) and display fining-up grading, whereas finer beds are thinner (10-20 cm);

Sl - medium sand with low-angle x-stratification, beds 20-30 cm thick;

secondary lithofacies:

Gm - massive, clast-supported pebble in sheet beds approx. 10 cm thick and few tens of metres long, erosional bases;

rare lithofacies:

St - coarse and medium sand of trough x-stratification, set thickness is 10-20 cm;

Sp - medium to coarse sand of tabular x-stratification, 20 cm thick beds often display fining-up grading;

SFh - silty fine sand in thin (5-10 cm) sheet beds; horizontal lamination

D-5

dominant lithofacies:

Sr - medium and fine sand in ripple laminated cosets up to 0.5 m thick, presence of ripple horizons in the top of cosets

Sh - horizontally laminated fine sand

Fh (FSh) - silt (sandy silt) of horizontal lamination (flaser lamination more rarely);

rare lithofacies:

SGh - thin (up to 5 cm) sheet-like intercalations of gravelly sand with horizontal stratification

Sp - medium sand of tabular x-stratification; beds up to 20 cm thick;

general characteristics:

all lithofacies in sheet-like beds up to 30 cm thick;

abundant normally graded sequences up to 0.75 m thick, with vertical succession: *Sr⇒Sh⇒Fh (FSh)* or *Sr➔Sh*

Fig. 5. Idealized sections through the facies *D-1* through *D-5*.

trough-type cross-stratified sets. An important difference, however, is that the troughs in facies *D-2* are significantly smaller (Fig. 5). They are characteristic deposits formed as sinuous sand dunes in channels. The height of the dunes can be estimated to 25–80 cm. Such high dunes can only have developed in relatively deep channels, although these must have been significantly shallower (about half as deep) than those of facies *D-1* (see Table 1). We therefore interpret the facies *D-2* channels as secondary channels in a braided fluvial system.

Facies *D-1* and *D-2* are commonly found together in the same outcrop. This association supports the interpretation of the depositional environment as a braided fluvial system with channels of different size, as well known from both present-day and fossil-braided fluvial systems (cf. Hein & Walker 1977; Bristow 1996).

Some lithologic and palaeoenvironmental relationships between facies *D-1, D-2* and *P-3* exist: they all developed in a similar way, resulting in trough cross-stratification (however of different sizes). The relatively coarse-grained character of facies *P-3* indicates that this gravel-dominated facies originated more proximally with respect to the water/sediment source than the two sand-dominated facies. It is only logical that the competence of the meltwater currents decreased downcurrent the sandur system, but the overall depositional conditions (i.e. channel bed configuration) remained the same over a fairly large distance, as observed also on present-day sandar.

Table 1. Parameters affecting transport and deposition in facies *P-1* through *D-5*.

		Palaeohydraulic parameters				Flow regime				Sediment transport	
						Subcritical					
	Facies	Depth d [m]	Velocity v [ms^{-1}]	Shear stress τ [Nm^{-2}]	Unit power ω [Nm^{-1}s^{-1}]	Lower part $Fr < 0.3$	Upper part $0.3 < Fr < 0.7$	Transitional $0.7 < Fr < 1.0$	Supercritical $Fr > 1.0$	Sand	Gravel
Gravel facies	***P-1***	3–4	1–5	75–250	100–300		■	■	■	➰	→ ↘ ➰
	P-2	1.4–2.4	3.0–3.9	42–73	126–282		■	■	■	↘ ➰ ➰	→ ↘ ➰
	P-3	1–3	1.9–2.7	24–46	45–124		■	■		↘ ➰ ➰	→ ↘
	P-4	0.3–0.5	1.7–2.2	15–20	26–44			■	■	↘ ➰	→ ↘ ➰
Sand facies	***D-1***	1.0–2.3	1.8–2.1	36	68		■			↘ ➰	→ ↘
	D-2	0.8–1.5	1.6–1.8	24–27	40		■			→ ↘	→
	D-3	0.8–1.0	0.8–2.1	12–23	10–46		■	■		→ ↘	→
	D-4	≤ 0.4	< 2.0	9–21	8–42			■		→ ↘	→
	D-5	0.1–0.2	< 1.2	< 4	< 5	■	■	■		→ ↘	

Transport modes: → rolling ↘ saltation ➰ suspension.

Facies *D-3*

Facies *D-3* consists of sand with both high-angle and low-angle tabular cross-stratification (Fig. 5). The units *Sp* are interpreted as resulting from migration and distal accretion of transverse bars – the macroforms typical of sand-bed aggradation in braided channels. These foreset bars must have covered most of the channel bed. They apparently coalesced, resulting in sand flats.

The sand beds with low-angle cross-stratification (*Sl*), which are less common in this facies, also fit well into the framework with braid bars. As a rule, braided channels are shallow and during phases of waning flood shoaling takes place. The foreset bars develop less high under such conditions and the bed configuration evolves consistently towards the upper plane-bed stage. The *Sl* beds must have formed in this way.

Another secondary facies, *St*, is seen to be related to sinuous dunes in interbar channels, whereas the *Sh* facies is a consequence of the upper plane-bed areas in shallow near-bank flows. Sandy-silty intercalations (*SFh*) developed during low-water stages following periods of limited ablation.

On the other hand, high-discharge flood events are recorded in this facies as well. These short-lived high-discharge events formed diffuse gravel sheets (*Gm*).

A comparison with the literature data shows distinct similarities between facies *D-3* and 'classical' sand-bed braided rivers. Particularly striking are the similarities with the Platte River succession (Smith 1970; Miall 1978): a high frequency of *Sp* lithofacies, secondary contributions of *St* and *Gm* facies and sporadic presence of silt beds. The channels of facies *D-3* are also analogous with types 9 and 10 of Miall's (1985) classification of braided rivers. Two other fluvial successions – the South Saskatchewan River (Cant & Walker 1976) and the Donjek River (Miall 1977) – have been attributed to similar palaeoenvironmental conditions. We interpret facies *D-3* therefore as aggrading braided channels dominated by foreset bar accumulation.

Facies *D-4*

Facies *D-4* occurs in association with several other lithofacies and consists of sands displaying horizontal stratification *Sh* and low-angle cross-stratification *Sl*. Most beds of facies *D-4* (Fig. 5) are 10–30 cm thick. Many have a wide lateral extent (sheet-like shape) and show fining-upward grading. A few fining-upward sequences have also been found; these commonly form vertical series of successively: *Sp, Sl* and *SFh* units.

The major types of deposits formed under similar hydrodynamic conditions. Both the sand and gravel (i.e. *Sh, Sl* and *Gm*) were deposited from supercritical or transitional flows.

The depositional conditions of facies *D-4* thus represent a lower energy level than existed during the formation of facies *D-3*; deposition most likely took place in the more distal zone of the outwash plain. The transport capacity of the meltwater streams was relatively low, as indicated by the grain size. The aggradation ratio was presumably also lower, as suggested by relatively small bed thicknesses.

The shallow depositional channels support the suggestion that facies *D-4* developed farther away from the ice margin than facies *D-1*, *D-2* and *D-3*. This is consistent with the extensive literature data (among them Abdullatif 1989; Blair & McPherson 1994; Miall 1996) indicating that sand-bed braided channels frequently evolve into shallow and wide streams (comparable to sheetflows) in a distal direction. Sediments similar to facies *D-4* are abundant also on the Middle Polish Lowland; they are found particularly at sites where relatively thin outwash series overlie melt-out till.

The most important characteristics of facies *D-4* are comparable with those of the Bijou Creek sedimentary succession (McKee *et al.* 1967; Miall 1977). Both are dominated by horizontally stratified sands. These characteristics are consistent with those of instantaneous, quasi-periodic flood events. Some analogies exist also with the S_{II} succession sensu Rust (1978).

Facies *D-5*

Facies *D-5* comprises ripple-laminated sand *Sr* and horizontally laminated sandy silt *FSh*, and has the highest content of silt and the lowest content of gravelly intercalations of all facies. A few erosional contacts exist. Ripple-derived structures are abundant (Fig. 5). Most beds were originally formed under conditions of the lower part of the lower flow regime, i.e. in weak currents. A gradual but ongoing shallowing of the currents made the bedding change from rippled to planar (the upper stage plane bed). The fining-upward successions (i.e. upward change from deposition from saltation to settling from quiet suspension) suggest that accumulation took place in successive phases of waning discharge.

Facies *D-5* lacks the cross-bedded units (of the trough type in particular) that are indicative of deep and intense channel flows. The sediments were deposited by low-energy, shallow and widespread, periodically fluctuating currents, presumably flowing through secondary, ephemeral channels located in the marginal parts of a braided fluvial system and in between the first-rank, perennially active waterways.

Facies *D-5* represents the lowest energy level among all deposits studied. Together with facies *D-4*, facies *D-5* can be considered as analogous to types 11 and 12 of braided river alluvium in Miall's (1985) classification, i.e. sandy, sheet-like beds deposited in shallow channels with fine-grained intercalations representing overbank conditions.

Overview

The question to be addressed here is whether the above facies are more typical of an alluvial-fan or of a braidplain environment. Six of the nine facies distinguished (i.e. *P-1*, *P-2*, *P-3*, *D-1*, *D-2* and *D-3*) are typical of deep gravel-bed or sand-bed channels of low sinuosity, most probably braided ones. This is obvious from the abundant presence of trough cross-strata, both of large and medium scale (facies *P-3*, *D-1* and *D-2*) and of planar cross-strata thicker than 0.5–0.7 m (facies *P-1*, *P-2* and *D-3*). It cannot be ruled out that some gravel-bed channels existed on an alluvial fan, as such deep channels ('trunk channels') are quite common on proximal fans. In the case of terminoglacial fans, however, the channels are frequently filled by initially reworked glacial debris, i.e. by very poorly sorted, massive or crudely stratified diamictic sands or gravels (Zielinski & Van Loon 1999). These types of deposits were not found in the Polish Weichselian outwash successions.

Only three facies (i.e. *P-4*, *D-4* and *D-5*) contain features that indicate deposition by the shallow, broad flows that characterize alluvial fans; these facies are also common in Iceland nowadays, i.e. in quite-phase surge landscapes (A. Russell, pers. comm. 2002). Moreover, the mass-flow deposits that are typical of proximal terminoglacial fans (see Krüger 1994; Zielinski & Van Loon 1996) and arctic alluvial fans (Catto 1993) have never been described from outwash facies.

Terminoglacial fans usually have a limited length (up to 1–2 km). The wide variety of conditions present on the relatively small surface area of such fans therefore results in regular lateral facies transitions, which can often be observed in large pits. Although the scale of sandur excavations is frequently large, no such distinct lateral facies passages have been noted in any of the exposures that were investigated. This supports the conclusion that the outwash deposits presented in this study are a record of broad braidplains, where spatial variability in conditions (and thus in subenvironments) was rare and where facies transitions were rapid and accidental, resulting in deposits displaying characteristics that have only limited resemblance to the characteristics shown by most present-day sandurs.

To summarize, the qualitative character of the facies strongly suggests that the deposits settled mainly in channelized, relatively deep flows. Consequently, outwash deposits are thought to represent the environment of a braided-river system (braidplain) rather than the environment of a terminoglacial fan where shallow sheetflows tend to prevail.

Palaeohydraulics

We tried to reconstruct the most relevant palaeohydraulic parameters, because this provides a deeper insight into the environmental conditions under which the various facies could build up both laterally and vertically. The flow depth (*d*) was reconstructed from the thickness of cross-bedded units. Other quantitatively estimated parameters of the palaeocurrents are: velocity (*v*), Froude number (*Fr*), shear stress (τ) and power per unit width (ω). The latter two parameters are particularly important for a comparison of the prevailing energy levels in the various fluvial subenvironments. It is stressed that the accuracy of the following palaeohydraulic calculations is limited (see also Maizels 1983), but all facies underwent a uniform procedure of calculations. The findings may therefore be considered reliable at least when comparing these data for one of the facies or subfacies with those of another. In addition, we consider outcomes of our calculations for the various facies only as different from one another if these outcomes differ by at least a factor of 2.

Input parameters for calculations

The following features were used as input data.

(1) Thickness of cross-beds, t [m].
(2) Manning roughness coefficient, n [dimensionless]. In the case of sand deposits, bedforms interpreted on the basis of sedimentary structures made reconstruction of the flow regime possible, and therefore also of the roughness coefficient (cf. Simons & Richardson 1961; Albertson & Simons 1964; Pathridge & Baker 1987). In the case of horizontally stratified or massive gravels, the grain-roughness coefficient, n_g, was calculated on the basis of the grain-size diameter, D, according to Strickler's (1923) formula.
(3) Grain-size diameter, D [m]. Because sedimentological studies indicate that the flow dynamics is reflected in the coarsest sizes of deposited material, the maximum particle size MPS, i.e. the average diameter of 10 maximum-size grains in the sample, was taken into account (cf. Maizels 1983a; Steer & Abbott 1984). The intermediate axis, b, is taken in this case as the relevant diameter.
(4) The slope of the palaeochannel bed, S [dimensionless]; this slope is assumed to be the same as the slope of the terrace only in the uppermost deposits. This assessment was made only when the terrace in the field still shows a genuine surface of sandur accumulation, i.e. without traces of later erosional processes, and without an eolian cover.
(5) Specific weight of water, γ [N m^{-3}], which is taken as 9820, on the basis of measurements in a proglacial stream of 4°C and with 2 kg m^{-3} of suspended load (Boothroyd & Ashley 1975).

Palaeohydraulic results

In the case of cross-bedded sediments, medium-scale trough cross-stratification is a proxy for the height of sinuous dunes; height has been calculated as the average from the values that have been obtained by applying the formulas given by Simons & Richardson (1962), Harms & Fahnestock (1965) and Cant (1978). The height of dunes has been used to estimate the mean depth of the palaeocurrents following Simons & Richardson (1962), Harms & Fahnestock (1965) and Goodwin & Masters (1983).

Medium- and large-scale tabular cross-bedded sands that formed due to progradation of foreset bars may also be treated as a good indicator of depth, according to Klimek (1972), Eynon & Walker (1974), Saunderson & Jopling (1980), Carling (1990) and Leclair & Bridge (2001). The mean velocity of the palaeocurrents has been estimated by using the Manning formula: $v = d^{0.67}S^{0.5}n^{-1}$. Where palaeoslope data were lacking, the velocity parameter was estimated from the average of the empirical equations for gravel deposits given by Miller *et al.* (1977), Koster (1978), Costa (1983) and Williams (1983).

The shear stress, τ [N m^{-2}], of the palaeocurrent has been derived from Shields' formula plot (see for instance Church & Gilbert 1975: fig. 17b) and from empirical formulas. These have been calculated separately for sand deposits from the formula given by Collins & Ringler (1982) and for gravel deposits from the average outcomes of the formulas given by Baker & Ritter (1975), Carling (1983), Maizels (1983a) and Williams (1983).

The values of the parameters d, v and τ presented in Table 1 are averages of the results obtained by applying the various equations.

Power per unit width (unit power), ω [N (ms)$^{-1}$], is the best measure for the energy level of a palaeocurrent. We applied a different procedure to estimate the above values for massive and horizontally stratified gravels. They lack cross-bedding, so that the depth of the palaeocurrents has therefore been calculated from the well-known expression described first by Paul du Boys (1847–1927) in his treatise on the hydraulics of the Rhine river. The velocity of the palaeocurrents that deposited gravels has been calculated using Manning's formula.

The prevailing unit stream power in the palaeochannels has been checked to find out whether it is consistent with the above-presented division into two facies associations, i.e. the gravel-dominated, proximal facies (coded *P*) and sand-rich, distal facies (*D*). The energy of the currents in these associations decreased from *P-1* to *P-4* and from *D-1* to *D-5* (Table 1). We consider it a result of sound sedimentary analysis that our distinction between outwash facies and facies associations, as carried out in the 'traditional' (qualitative) way, resulted in the same outcome as did our quantitative palaeohydraulic method.

The succession (*P-1* to *D-5*) of outwash facies in Table 1 shows roughly decreasing Froude numbers. The gravel-dominated, proximal facies are characterized by the highest *Fr* values and were most commonly deposited by supercritical flows. On the other hand, the sand-dominated, distal facies of low-energy origin are distinguished by significantly lower *Fr* values. Upper-flow regime bedforms are uncommon in this latter association.

The gravel-rich facies *P-1*, *P-2* and *P-3* are of particular interest in terms of their high values of stream power ($50 < \omega < 300$ N m^{-1} s^{-1}). These values reflect the conditions during torrential ablation floods. Hydraulic data presented by Ashworth & Ferguson (1986) also identify sediments with stream-power values comparable to those obtained for facies *P-1* and *P-2* as results of strong ablation floods. Magilligan (1992) defined a value of $\omega \cong 300$ N m^{-1} s^{-1} as a threshold value for catastrophic flow. However, this threshold value can be an underestimation (cf. Nanson & Croke 1992). In our opinion, facies *P-1* and *P-2* thus might represent jökulhlaup deposits. It must be emphasized, however, that the hydraulic parameters of present-day jökulhlaups controlled by the sudden outbursts of lakes due to volcanic eruptions are significantly higher (with respect to shear stress, current velocity, and – particularly – stream power) than those estimated for 'normal' Pleistocene ablation floods in NE Poland. On the other hand, maximum values of ω for facies *P-1* are over 1000 N m^{-1} s^{-1}, which are comparable to those of jökulhlaups (cf. Baker & Costa 1987; Maizels 1993; Russell & Marren 1999).

Palaeocurrent data

Cross-bedding dip azimuths were measured at each site. Their frequency of occurrence was so high that the calculated palaeocurrent directions are statistically significant at the 95% confidence level. With respect to the mean vector magnitude, *L*, all deposits are considered to have been formed in channels of low sinuosity. The average value for *L* is 68% ($54\% < L < 80\%$), both for gravelly and sandy facies. The parameter of sinuosity, *sn*, of outwash channels has been estimated according to Langbein & Leopold's (1966) formula $sn = 1 / [1 - (V_d / 252)^2]$, where V_d is the maximum difference between mean azimuths obtained from several packages superimposed in one vertical section. As a rule, our *sn* values are extremely low, ranging between 1.0 and 1.1 only. The above values indicate that the channels on the Polish sandar had very low sinuosity (i.e. most probably were braided).

The fairly constant orientation of foresets is, obviously, one of the consequences of the specific hydrological conditions determining the discharge of the meltwater streams. Low-stage currents follow a more sinuous pattern than do strong currents. In ephemeral streams – and sandur streams belong to this type – low-stage conditions exist for a relatively short time, so that insufficient time is available for channel-bed morphology to adapt. This fact is important because most beds of the various outwash facies were formed during floods.

Directional distributions contain usually 1–3 modes. Both the relatively low number of modes (cf. Bluck 1974; Steel & Thompson 1983; Mader & Teyssen 1985; Aitken 1998) and the low spreading shown by the roses (Casshyap 1973; Jones 1979; Fielding & Webb 1996) indicate deposition in a braided river environment.

The directional characteristics of various types of cross-stratification developed in the sands have been analysed (Fig. 6). Significant differences appear to exist between the distributions of planar cross-stratification developed in transverse bars, in contrast to trough cross-stratification developed in highly sinuous or linguoid dunes. Planar cross-beds *Sp* and *SGp* are characterized by a lower spreading in the roses with 1–3 modes. Distributions of trough cross-stratified beds *St* and *SGt* cover broader azimuth ranges and are formed of 3–4 modes, which are sometimes oriented in opposite directions (Fig. 6). These differences are also expressed by the magnitude of their vectors: L is on an average 44% higher for planar than for trough cross-stratification. A similar relationship has been described for Jurassic fluvial deposits of Nova Scotia (Tanner & Hubert 1992) and for Pleistocene proglacial alluvium of southern Poland (Zielinski 1992). This phenomenon is due to the bars being large, aggradational depositional forms conditioned by the main channel currents. On the other hand, the dunes (especially the sinuous and linguoid ones) can prograde in a wide range of directions, even in a straight channel, because their migration may be affected by the presence of larger forms such as bars. We agree with Bluck (1974) and others that sand beds with large-scale planar cross-stratification provide the most accurate and reliable palaeocurrent direction of braided rivers.

Sequences and cycles

In some of the thick outwash successions studied, till horizons are present together with proglacial alluvium. The tills prove the local presence of an ice sheet. Proglacial sediments underlain by glacial diamict should be considered as representing a retreating glacier (regressive outwash), whereas outwash sediments with till on top should be considered as transgressive. This interpretation of transgressive and regressive outwash holds only when no significant hiatuses or erosion horizons between the glaciofluvial and glacial units occur.

Both transgressive and regressive outwash sediments are, at least theoretically, characterized by specific sequences (the classical sandur model is an example) as a result of ice advance and retreat, respectively. However, a single, characteristic sequence cannot be

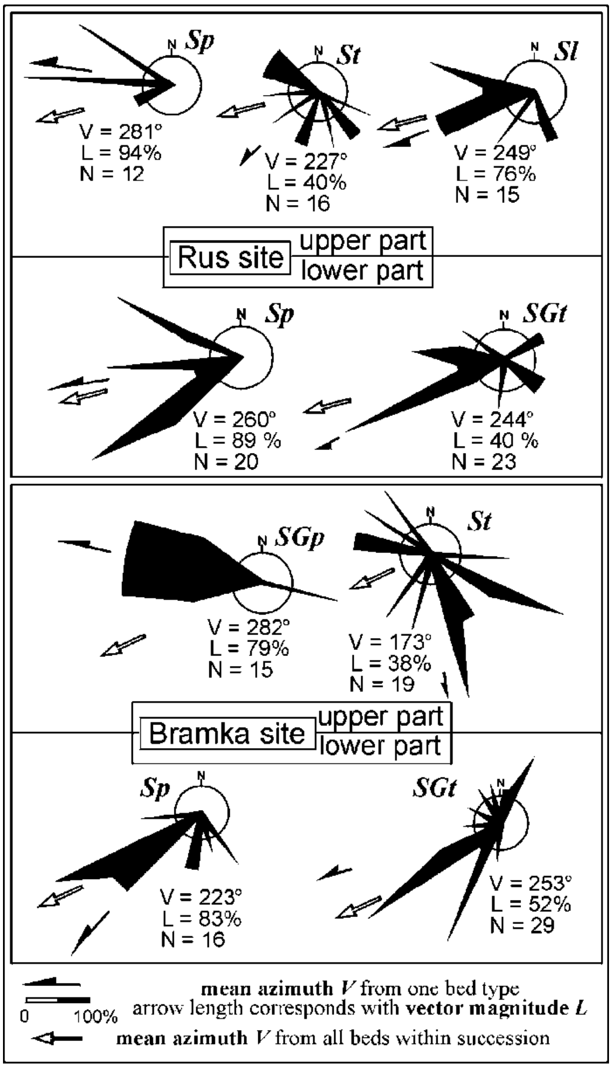


Fig. 6. Palaeocurrent data derived from cross-stratified units in two of the investigated sites.

found in the Polish outwash deposits. Nevertheless, sequences are numerous. Sometimes they give the impression of cycles, but a true cyclic succession is rare. The interpretation of true sedimentary cycles is still one of the unsolved problems in sandur deposits. The sequences – either or not in cycles – are more common in braided-river alluvium (see, for example, Bluck 1979; Cant & Walker 1978; Mader 1985; Godin 1991; Bridge 1993; Viseras & Fernández 1995).

Types of sequences and cycles

The coarsest cyclic sequences (Fig. 7, cycle A) formed in high-energy, deep proglacial channels close to the ice-sheet margin. The two-member sequence closely resembles that described by Middleton & Trujillo (1984) from a Proterozoic proximal braided river alluvium. A sequence in which the two units occur in reversed vertical order has been interpreted by Russell & Marren (1999) as present-day jökulhlaup deposits with a lower unit deposited by a high-discharge flood, and an upper unit as due to surficial reworking (cf. Petch & Whittaker 1997) during waning of the flood.

Another gravelly, sometimes cyclic sequence consists, from bottom to top, of a thick *Gm* bed (or beds), a *GSp* unit, and an *Sh* unit (Fig. 7, cycle B). It is interpreted as successive accretion of longitudinal bars within an aggrading channel during the peak, followed by waning stage (first with lateral growth of a bar, then with a shallow, supercritical flow above the bar). A similar sequence has been described by Blakey & Gubitosa (1984) from the Triassic Chinle Formation. Some similarities exist also to a sequence in a Palaeogene braided river (Evans 1991).

Repetitive, two-unit couplets (Fig. 7, cycle C) reflect short-term ablation floods that occurred in a cyclic mode on the sandur. Gravelly/sandy alternations of similar scale have been described by Blair (1987), Nemec & Muszynski (1982), Zielinski (1992) and Aitken (1998), who all interpreted them as effects of periodic, shallow floods.

A fining-upward coset of trough cross-stratified *Gt* and *St* units (Fig. 7, cycle D) is interpreted as a deposit formed in a deep channel where sinuous or linguoid dunes prevailed; when the competence and depth of the current decreased, both the grain size and the thickness of the cross-bedding units diminished. Cyclic fluvial packages of similar lithological character have been noted by Williams (1971) in deposits of ephemeral streams, and by Singh & Bhardwaj (1991) in deposits of the braided Ganga River.

Another sequence is characteristic of outwash sedimentation. The base-to-top succession is: erosional surface, sheet-like *Gm* bed, two to four beds in which the *SGp* units diminish in grain size to *Sp* and *Sl* and subsequently to *Sh* (Fig. 7, cycle E). This compound sequence is thought to represent a single flood event in a braided channel dominated by sandy foreset bars. The origin of the basal gravel sheet reflects the episode of peak discharge. The main depositional phase took place then; sandy transverse bars grew when the flood became less torrential. The final units represent shallow flows. The bars became lower and the bed morphology evolved according to the upper plane-bed stage at the end. This channel cycle has been described frequently by other investigators (Smith 1970, 1974; Williams 1971; Boothroyd & Ashley 1975; Crowley 1983; Brierley 1991; Bristow 1993) from a wide variety of environments, ranging from proglacial to semi-arid ephemeral streams.

Four varieties of two-part repetitive units have been distinguished (Fig. 7, cycle F). Their lithological features point out that they were deposited by shallow, short-lived supercritical (or transitional) flows, the

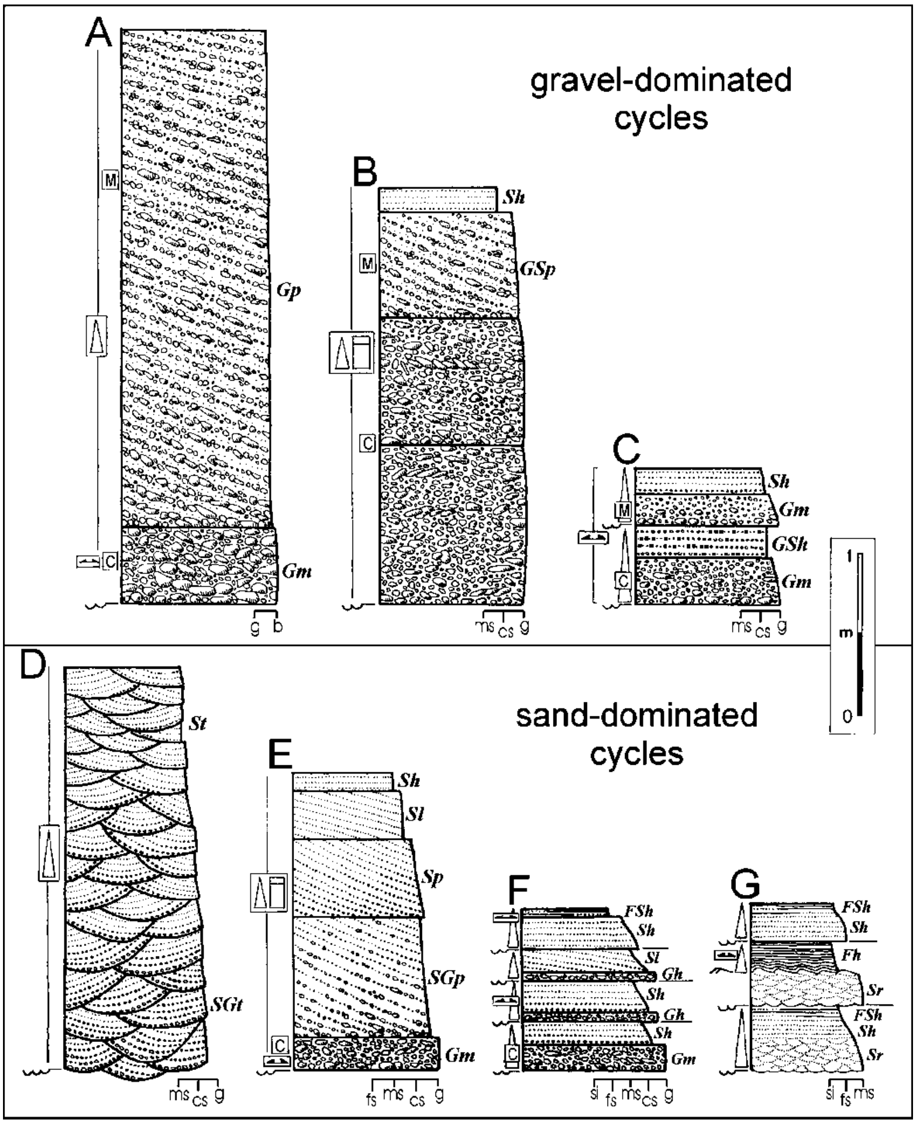


Fig. 7. The seven types (A–G) of cyclic sedimentation in the sandur deposits under investigation.

current activity of which stopped abruptly. Fining-up repetitive two-part units have been genetically connected with flashy, shallow streams by Harvey (1984), Tunbridge (1984) and Dreyer (1993).

Low-energy, fining-upward sequences are commonly composed of sand and silt (Fig. 7, cycle G). Ablation floods apparently reached the marginal channels of the braided system only as shallow currents of low velocity. Each such flood cycle ended in a stagnant-water phase, when settlement of fines resulted in the upper unit (*Fh* or *FSh*).

A model of the sequences and cycles

The studied outwash sediments are characterized by more frequent sequences and better developed cycles than are present in non-glacigenic braided-river alluvium. This observation must be ascribed to the high aggradation rate in sandur streams, resulting from the abundance of clasts provided by the nearby ice mass. Such a wealth of clasts is fairly common in outwash deposts but is less common in other ephemeral streams (for instance in arid or semi-arid environments). The highly variable ablation-induced discharges are the second reason for the better development of sequences and cycles on sandar than in other, non-glacigenic fluvial environments. All sequences presented here (maybe except those from facies *D*-5) were formed in channels.

The thickness of the various sequences is a function of their grain size. This relationship is due to the decrease of both energy of sandur currents and aggradation ratio in a downcurrent direction. The duration of the individual floods, however, was independent of the distance from the ice-sheet margin.

The discharge of outwash streams, depending on a combination of insolation and ablation, follows three cycles: day-night, summer-winter and an interseasonal one. The diurnal rhythm leads in summertime, particularly when no clouds are present, to considerable daily cycles in discharges (cf. Sambrook Smith 2000); the main character of the currents, however, does not change drastically, largely because of the lag-time associated with ablation as insolation increases. Larger cyclic changes in discharge occur – and this is important particularly for high discharges – between summer and winter (Gurnell & Fenn 1984; Chikita *et al.* 1991). The most extreme situations (commonly in the form of a few distinct, multi-day floods) occur as interseasonal phenomena (Hammer & Smith 1983; Cowan & Powell 1991). These interseasonal cycles with large ablation (known also from present-day glaciers: Hodgkins 1996) seem to be the prime governing agent determining the outwash sequences presented here.

The thick, coarse-grained successions in facies *P-1* cannot be explained by the 'normal' rhythms. They must be ascribed to meltwater outburst floods. Such floods take place commonly in front of the glacier during spring, when the above zero temperature of rain causes superficial ice masses to melt, and thus contributes to the destruction of ice-dams of supra- and englacial reservoirs (Church 1988; Collins 1991); such englacial reservoirs may be formed under conditions of fast-moving ice, closing the outlet of tunnels (Mayo 1989). The floods resulting from the sudden discharge of englacial lakes thus formed induce surges, which may also have another origin, however (Björnsson 1992). Such flood events are non-cyclic phenomena and presumably occur at a specific site at irregular intervals. The resulting sequences therefore cannot be considered to form part of cycles (cf. Björnsson 1992, 1998; Tweed & Russell 1999).

Discussion: braided-river sedimentation versus outwash-plain sedimentation

Sandar are commonly regarded as a special type of alluvial fan (Boothroyd & Ashley 1975; Church & Gilbert 1975; Cherven 1984; Landvik & Mangerud 1985; Smith 1985; Dawson & Bryant 1987; Maizels 1993; Rice & Edgett 1997). In contrast, we found several fundamental aspects in which the deposits constituting the Polish sandar differ from those forming alluvial fans, but are consistent with braided-system deposits. The two most important characteristics in this context are (1) the vertical successions (sequences), and (2) the palaeohydraulic characteristics.

Vertical successions

If the outwash deposits were consistent with alluvial-fan deposits, each vertical succession would, taking into account that the deposits were formed during slow ice retreat, show a 'regressive' sequence, i.e. a gradual fining upward (Fig. 8A). Only in one case, however, was such a character found, i.e. at the Grajewo site (Fig. 2, site 10), where the coarse-grained facies *P-1* directly overlies glacial diamict (Fig. 8B). All other profiles show a complex succession, as also found sometimes in modern situations (e.g. Van Tatenhove 1996), that needs some discussion. Good examples of complex successions (Fig. 8B) are the two sites of Suwalki and Rus (sites 2 and 15 in Fig. 2). The sandur morphology has not been affected, and they are therefore considered to represent 'normal' sandar, formed during ice retreat (or possibly a phase of stillstand). Both sites show vertical sections, however, that start with sandy distal facies, passing upwards into more proximal sediments deposited in deep channels with high-energy currents. These successions thus show quite 'ice-transgressive' character. Such 'reversed' regressive sections are not restricted to our study area: they have also been described by Costello & Walker (1972).

Ideal 'transgressive' sandur successions (i.e. successions formed in front of an advancing ice sheet) would show up as the opposite development of the 'regressive' development presented in Fig. 8A (cf. Miall 1980; Ehlers & Grube 1983). However, even the sections that are capped with a till – and thus might point to ice advance – do not show such a succession (Fig. 8C). Sections similar to those at Barcikowo and Bramka (sites 1 and 3 in Fig. 2) have been described by Vincent (1984) in Canada and by Kasprzak (1997) in other regions in Poland.

No ideal superposition of facies has been noted either in the profiles representing 'transgressive-to-regressive' sedimentary records (Fig. 8D), as found at Kruklin and Brejdyny (sites 5 and 7 in Fig. 2). At the Kruklin site, the lower, 'transgressive' part of the sandur deposits shows the opposite lithological succession to what is to be expected, with a gravelly succession of proximal facies in the lower part of the section, and a sandy facies capped with glacial diamict in the upper part. At the Brejdyny site, the 'transgressive' part of the outwash succession (below the till) starts with facies *D-4*, which is followed by facies *D-3*. An increase in the palaeocurrent energy is recorded in this succession; indeed, but there are no upper sediments that could be identified as proximal facies. The Jeze site (site 8 in Fig. 2) is even more complex, because till is succeeded by a single bed of massive gravel. Although this unit was deposited by a high-energy current, it cannot be regarded as an equivalent of a proximal sandur facies, i.e. the facies with which the upper, 'regressive' part of succession starts. This high-energy gravel possibly reflects only a single flood event. We consider the fine-grained

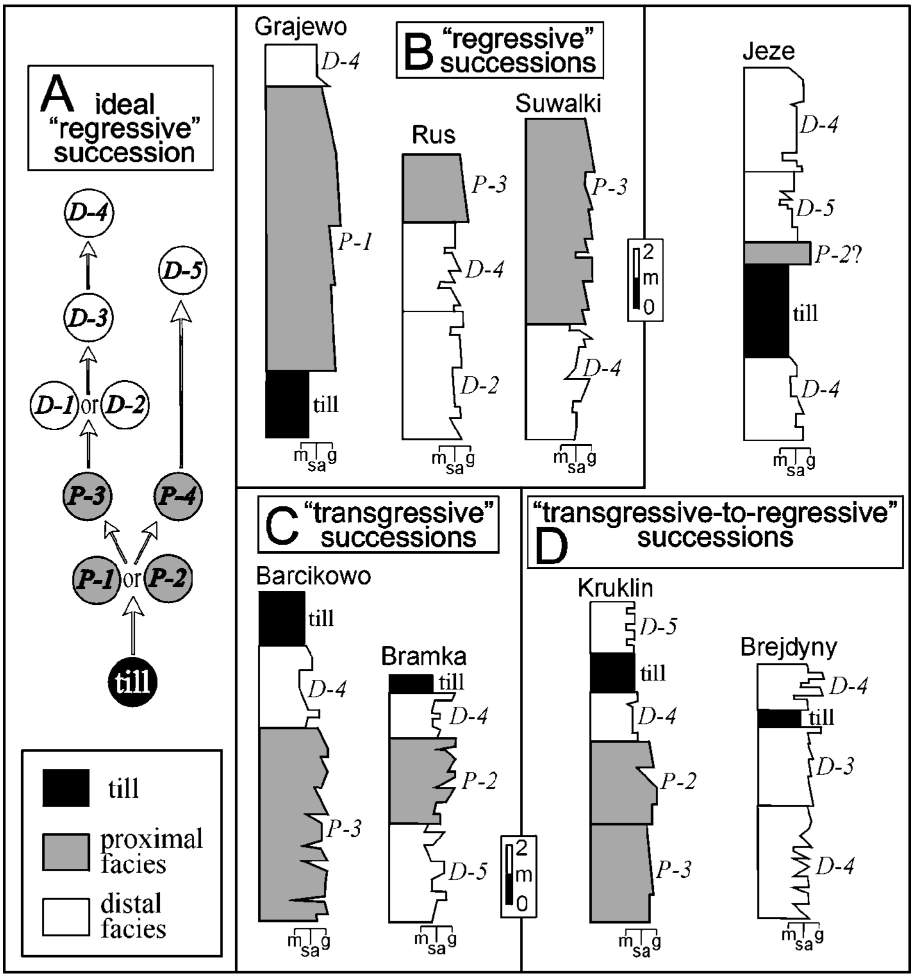


Fig. 8. Facies alternations in some of the investigated sites (B–D), showing distinct differences with the ideal 'regressive' succession, as well as the ideal (reversely developed) 'transgressive' succession based on the model of present-day sandur development (A).

package above the gravel as part of the low-energy *D-5* and *D-4* facies.

To summarize, typical vertical successions reflecting consequent and logical shifting of alluvial-fan subenvironments (from distal to proximal, or vice versa) are extremely rare within Polish sandar.

Palaeohydraulic characteristics

In our opinion, two palaeohydraulic parameters play significant roles in the distinction between fans and braidplains. The depth of palaeoflows is the first one: as a rule, a river environment is characterized by relatively deep channels, whereas shallow sheetflows dominate on alluvial fans (Table 2). Our palaeohydraulic analysis pointed out that five outwash lithofacies represent deep (1–4 m) channellized flows, and that only three facies represent sheetflows (see Table 1).

The Froude number is the second important parameter. As detailed by Blair & McPherson (1994), the steep slopes of fans result in supercritical sedimentation on alluvial fans. On the other hand, relatively deep channellized river flows are mostly characterized by subcritical ($Fr < 1.0$) hydraulic conditions (see Table 2). Our analysis indicates that only three gravel facies were deposited under conditions of prevailing supercritical flows. A great majority of the Polish sandar were dominated by flows with $0.3 < Fr < 1.0$ (Table 1), i.e. conditions which can readily be identified with braided-river hydrodynamics.

The discrepancy problem

Present-day models of outwash sedimentation (Fig. 9A–C) are based on regular accumulation, giving rise to a sequence of three groups of facies (see Boothroyd & Ashley 1975, and their 'classical' figure 25). These models all seem to take for granted that sandar form in a relatively narrow belt between glaciers and the sea. The models are therefore based on the assumption that the energy of ablation streams decreases progressively in an outward direction from the glacier margin. As a consequence, gravels with a dominantly massive

Table 2. Palaeohydraulic characteristics of sandar, braidplains and fans. Partly based on data from Gustavson (1974), Boothroyd & Nummedal (1978), Boothroyd & Ashley (1975), Middleton & Trujillo (1984), Ashworth & Ferguson (1986), Blair & McPherson (1994), and Perez-Arlucea *et al.* (2000).

Environment		Channel depth *d* [m]	Channel width *w* [m]	Flow velocity *v* [m s^{-1}]	Froude number *Fr*	Stream power ω [W m^{-2}]	Discharge Q_{max}/Q_{min}
Braid plains	**Gravel-bed channels**	≅3 (max = 5)	10–100 (max = 200)	<2.5	0.4–1.1	35–300	5–20
	Sand-bed channels	≤1	≤100	≤1.5	0.4–1.0		
Polish sandar	**Gravel-bed channels**	≅1.7 (max = 4)	No data available	Average ≅3.0	0.3–1.2	≅130	5–40
	Sand-bed channels	≅0.8		Average ≅1.3	0.1–1.0	≅40	
'Classical' sandar	**Upper**	≅1 (max ≅ 2)	From 20 (mean stage) to 1000 (flood stafe)	1.5–3.0 (max >5)	0.75–1.25	≅400	10–100
	Middle	0.2–0.7		1.0–2.0	0.25–0.9	5–50	
	Lower	No data		≅0.5	No data	No data	
'Wet' alluvial fans	**Upper**	0.5–1.5	No data	From 1.1–2.2	>1.4		
	Middle	≤0.5	10–300	decrease	>1.0	decrease (Quantitative data are lacking)	Up to 100
	Lower	<0.3	>100	to ≤0.5	≥1.0		

character (deposited as longitudinal bars) are – according to the currently accepted models – formed in the proximal zone. In a more downcurrent direction, sand-bed braided channels dominate in an intermediate position, with transverse foreset bars and sand beds with (mainly) planar cross-stratification. In the most distal zone, where the energy of the currents is minimal, the channels sometimes become sinuous, and the style of accumulation corresponds with transitional channel sedimentation and even with that in meandering ones, so that point-bar successions and sand-to-silt cycles are occasionally present.

This fairly simple but convincing sedimentary model, established on the basis of studies of recent outwash settings in Canada, Alaska and Iceland, has also frequently been taken as a basis for sedimentological analyses of fossil sandur deposits (Casshyap & Tewari 1982; Fraser & Cobb 1982; Houmark-Nielsen 1983; Cherven 1984; Olsen & Andreasen 1995). When testing this model against the Weichselian sandar in NE Poland, it turned out, however, that all Polish sandar are inconsistent with the model (Fig. 9D): the vertical facies successions are chaotic and many profiles show successions that are rather the opposite than the equivalent of previously presented fan-like sandur models (Fig. 8). This discrepancy needs an explanation.

One might question whether the Polish sandur architecture could be incidental. This is highly improbable, as our observations at 16 sites are consistent. Since the situation in Poland during the Weichselian ice retreat was not different from elsewhere in the lowland areas of Eurasia and North America, it thus seems that the Weichselian sandar present worldwide are not alluvial fans in most cases. Since evidence is overwhelming that present-day sandar on Iceland and elsewhere are alluvial fans, there is an intriguing difference between the Weichselian and the present-day sandar.

We have addressed this discrepancy problem in a fundamental way (see also Zielinski & Van Loon 2002), by comparing the geographical depositional conditions of the Polish and the present-day sandar that have been taken as examples to model sandur formation. The great majority of modern sandar (Fig. 9A–C) are built up in narrow belts between mountain massifs and the shoreline. They thus form small, relatively steep alluvial fans (see also Fig. 1) that may extend maximally 10–30 km in a downcurrent direction. This limited space is incidental (as a result of the imbalance between present-day shoreline and ice extent), so that recent sandar must be considered as not representative of the geological past, when much larger sandar could develop in front of the ice sheets. The short longitudinal section of modern sandar results necessarily in slopes that are significantly steeper than is commonly found for Pleistocene examples, and different hydrodynamic conditions must therefore exist. These different conditions are, obviously, reflected in the relative importance

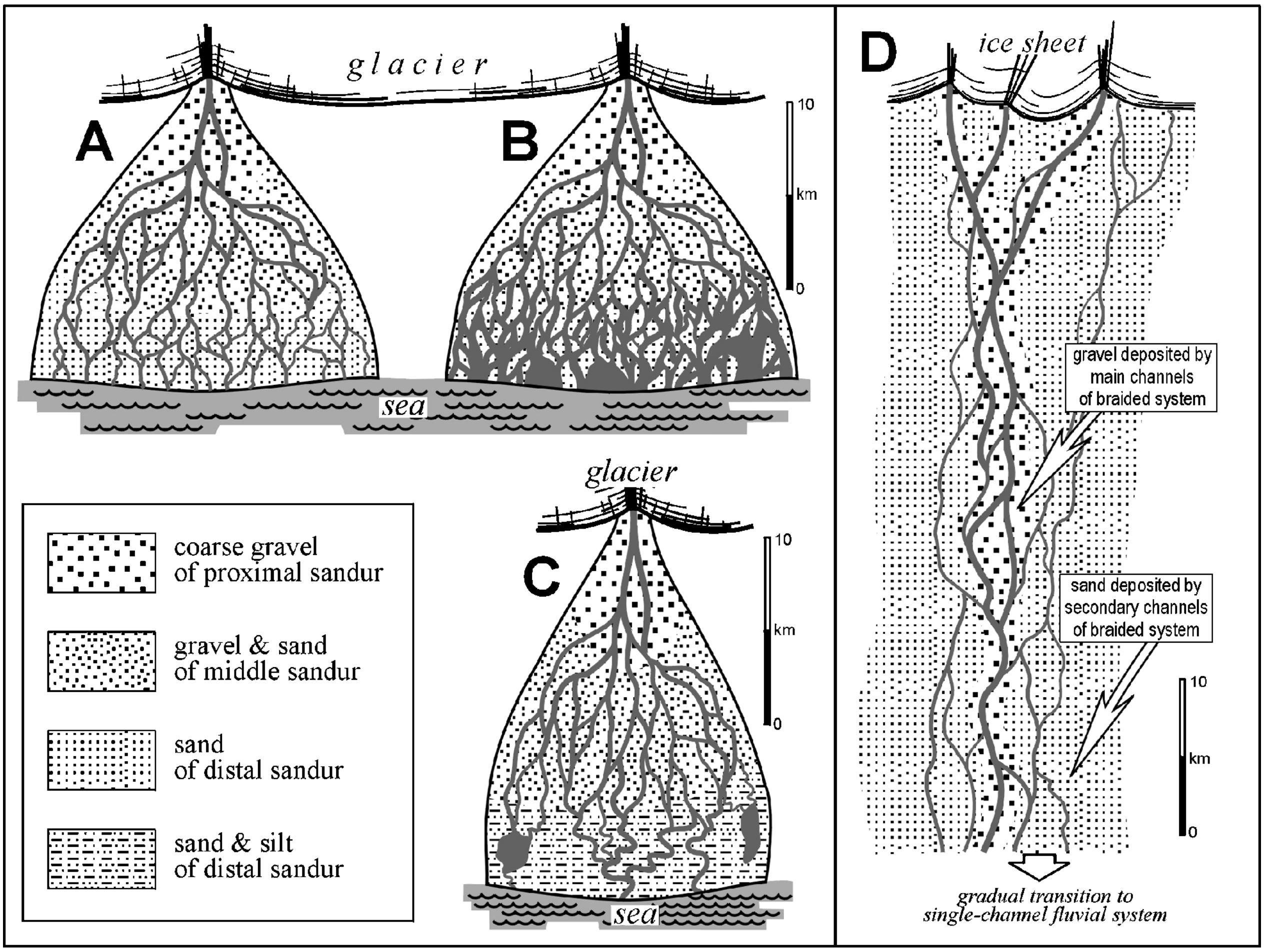


Fig. 9. Models of present-day and Weichselian sandar. A (based on Bothroyd & Nummedal 1978) and B (based on Krigström 1962): models of present-day sandar from Iceland; C (based on Boothroyd & Ashley 1975): model of present-day Alaskan sandar; D: model of the Weichselian sandar in NE Poland. Modified from Zielinski & Van Loon (2002).

of erosional and aggradational phases and thus in the types of deposit and in their lateral and vertical transitions.

The palaeogeomorphic conditions determining the Weichselian outwash formation in the Polish lowlands were probably much more representative for the geological past. The sandar were formed by proglacial rivers that flowed in long (50 km and more) tracts, where – in the more proximal parts – valley outwash accumulated, changing towards the south into wide outwash plains (Fig. 2). It is obvious that the Pleistocene sandar cannot be considered as alluvial fans, neither from a geomorphological nor from a sedimentological point of view. In fact they are broad, flat plains accumulated by proglacial rivers (braidplains).

In this context is should be emphasized that the Polish meltwater streams, though confined in straight, fairly long valleys, should not be considered as being situated in a landscape that was fundamentally different (because narrower) from the landscapes in which sandar develop nowadays. The 'valleys' developed on till plains are very shallow and wide. They are lows where erosional incision of meltwater was unimportant or absent. The mean width of the Polish valleys is almost 7 km, so the proglacial channels could shift without much spatial restriction. One could therefore say that – if the Polish streams were confined at all – the fluvial systems were only confined with regard to the depositional processes. The outwash plains even reached widths of some 45 km and fluvial processes acted there in a quite unconfined way. To conclude, sandur fans could theoretically develop easily in Poland, but practically they did not form there.

The braidplain conditions

The facies distribution within braidplains depends mainly on the morphology of the proglacial fluvial system. A specific feature of braided river systems is the

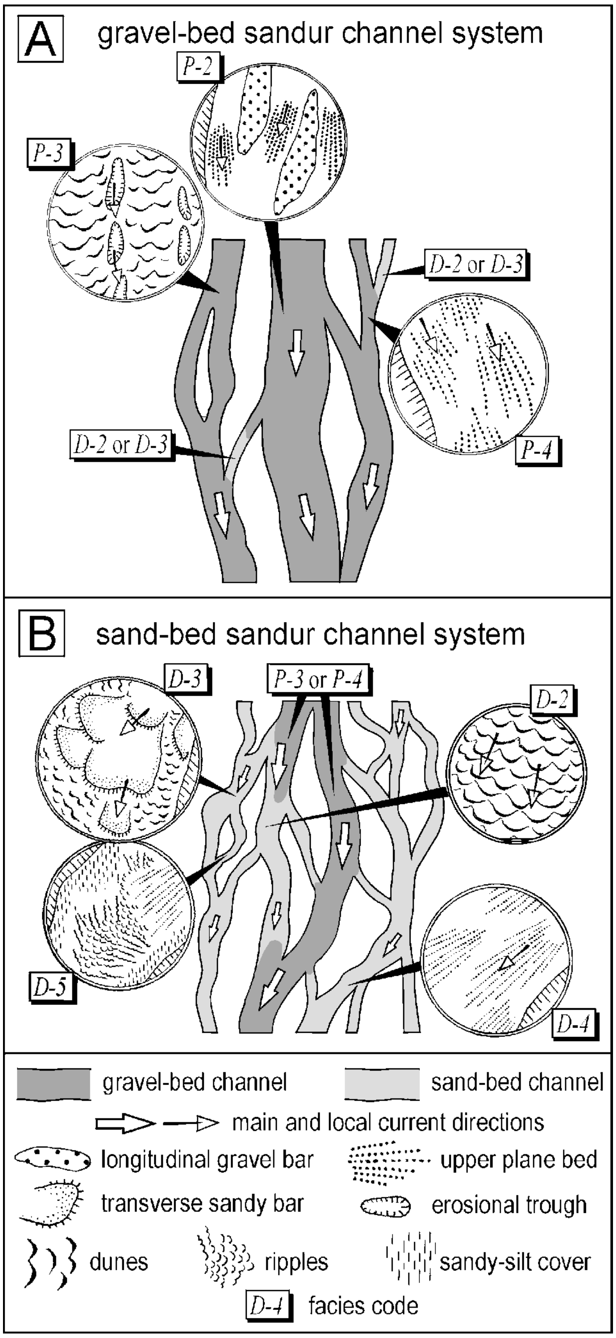


Fig. 10. Differences in sandur channel architectural elements for the various facies.

presence of main and secondary channels, with different hydrodynamic conditions and style of sedimentation (Bristow 1996), bedload transport and channel stability (Nicholas & Sambrook Smith 1998). Frequent, strongly fluctuating ablation discharges lead to intense, abrupt shifting of channels – by avulsion. The braided alluvium is therefore characterized by a high variability of textural and structural features, both vertically and laterally, and most of the lithofacies develop simultaneously. The gravelly facies prevail in a more proximal position to the ice-sheet margin (Fig. 10A) and pass distally into sandy facies (Fig. 10B). In braidplains this distally fining tendency is less evident and stretches out over longer distances than in alluvial fans. The extent of the proximal coarse-grained reach of braided rivers makes up to 15–50 km (Williams & Rust 1969), while the same zone is not longer than a few kilometres on alluvial fans (Nemec & Postma 1993; Owen *et al.* 1997). This overall picture, however, is complicated by the presence of main and secondary channels. For example, some sand facies (*D-2* and *D-3*) can develop in secondary channels situated in a proximal zone where gravels dominate (Fig. 10A) and coarse-grained (*P-3* and *P-4*) facies can develop during strong floods in distal sandur (Fig. 10B). The picture is further complicated because originally sandy (distal) facies can undergo a drastic change into gravelly ones during occasional episodes of extremely high discharge events. On the other hand, proximal sandar, close to marginal moraine chains, in our study area were frequently built up of sand facies only, so that lithologically they give the impression of 'distal' outwash.

Conclusions

Most sedimentologists and Quaternary geologists consider sandar as equivalents of alluvial fans (cf. Smith 1985). Indeed, sometimes their shapes on maps resemble big fans. However, a detailed analysis of the sedimentary architecture, in combination with a reconstruction of the palaeohydraulics, indicates that the well-developed Weichselian sandar in NE Poland are braidplains of proglacial rivers, and not alluvial fans.

This finding is consistent with the evaluation by Blair & McPherson (1994) regarding the distinction between alluvial-fan and braided-river environments; they state that the main alluvial-fan features that make them different from braided rivers are: a non-linear shape and relatively short extent, a steep slope, predominance of sheet flows over channellized streams, abundant presence of shallow and fast (i.e. supercritical) flows, an ephemeral character, and a considerable frequency of debris-flow deposits.

Whereas Blair & McPherson's (1994) study was based mainly on recent or subfossil forms, our study concentrated on sandur facies. It was found that, in contrast to what is a rule on alluvial fans, true debris-flow deposits are absent in the studied outwash. In addition, although all facies can be divided into gravelly and sandy and thus allow logical sequences, no regular vertical facies successions (neither those indicative of advancing ice, nor those reflecting – what should be common – retreating ice; cf. Möller *et al.* 1994) were present in the study area. This is another strong argument against a fan-related character of sandar,

Table 3. Comparison of the most relevant characteristics of sandar, braidplains and fans.

	Size, form & slope	Hydrology	Depositional processes and facies	Vertical facies successions	Distal grain-size variation
Braid plains	10–>100 km long; 100 m–10 km wide; up to 20–30 m thick; elongate form (so-called braidbelt); downstream slope from 0.005 to 0.00005 m/m	System of anabranching channels of low (<1.15) sinuosity, including main and secondary channels; highly irregular discharge	*Gravelly proximal system* up to 20 km from source area: in channels: gravel longitudinal bars (*Gm*) and gravel dunes (*GSt*), in overbank zones: sand splays (*Sl,Sr*), rarely silts (*Fh*) in flood basins; *sandy, distal system:* in main channels: sinuous dunes (*St*), in secondary channels: transverse bars (*Sp*) and upper plane bed (*Sh*), in overbank areas or in abandoned channels: fine-grained suspension settling (*Src*, *Fh*)	In sandy system possible local cyclicity derived from: floods, filling and abandoning of channels; in large aggrading rivers presence of long-term (several ka) megacycles	Distal decrease of grain size (from gravel & sand to sand) only on large scale
Polish sandar	Up to >50 km long; up to 10 km wide; usually 10–20 m thick; elongate artery: braidplain or valley fill; downstream slope from 0.002 to 0.0005 m/m	System of braided channels of different scale: large, gravel-bed, of higher energy; and smaller, sand-bed, of lower energy; gravel-bed channels prevail close to ice sheet; interseasonal cycles, catastrophic jökulhlaups	*Gravel-bed channels:* longitudinal bars (*Gm*), sinuous dunes (*GSt*) and upper plane bed (*Gh*); *sand-bed channels:* transverse bars (*Sp*), upper plane bed (*Sh*) and dunes (*St*); ripples (*Sr*) and silt drapes (*Fh*) in abandoned channels or in overbank areas with low-energy flows	Local cyclicity derived from channel upfilling and large floods	Poorly developed
'Classical' sandar	Up to 25 km long; up to 10 m thick; conical form; segmented slope, slope commonly from 0.008 (max = 0.02) in proximal zone to 0.002 m/m in distal zone	1-3 low-sinuous channels in proximal zone; braided channel system in the middle and distal zones (sinuous channels possible in distal zone); tendency of distally shallowing channels; daily and seasonal cycles in discharge; hyperconcentrated flows in flood peaks	*Proximal part:* longitudinal bars (*Gm,Gh*) in gravel-bed channels; *middle part:* transverse bars (*Sp*), gravel sheets (*GSm*) and upper plane bed (*Sh*) in sand-bed braided channels or in sheetflows; *distal part:* channel side bars, sometimes point bars (*St,Sr*), well-developed overbank facies (*Sr,Fm,C*)	Typical fluvial cycles only in distal zone (meandering-river cycles)	Clear distally fining tendency: from gravel & boulder to sand & silt or from coarse gravel to sand
'Wet' (humid) alluvial fans	Up to several km long; up to 20 m thick; conical form; slope clearly segmented, slope from approx. 0.03 (max = 0.08) in proximal zone to 0.006 m/m in distal zone	Common hyperconcentrated flows in early stage of fan evolution; frequent large floods; abundant channel avulsion; downstream shallowing and widening of channels; braided channels or sheetflows in middle zone; sheetflows and ephemeral ponds in distal zone	*Proximal part:* deep channel deposition (*Gt,Gp*) with mass-flow diamict intrabeds; *middle part:* shallow braided distributary channel deposition (*Sp,Sh,St*) or sheetflow deposition (*Sh*); *distal part:* low-energy sheetflow deposition (*Sh,SFh*), point bars (*Sr,St*) in sinuous channels, overbank and pond facies (*FSh,Fm*)	Clear fining-up successions (10–20 m thick) of proximal ⇨ middle ⇨ distal facies superimposition; frequent flood rhythms	Clear transition from gravel (or gravelly diamict) to sand or silty sand

because regular vertical and lateral facies transitions are typical features of terminoglacial fans (Zielinski & Van Loon 2000).

The conclusion thus should be that the Pleistocene sandar in NE Poland cannot be regarded as alluvial fans. On the other hand, they show all characteristics that are known from braidplains (see Table 3). The fact that recent sandar resemble alluvial fans rather than braidplains must be attributed to the fact that the present-day situation where sandar have insufficient space to develop between the (commonly mountain-bound) ice front and the sea; in addition, this temporary situation forces accumulation on a surface that is, particularly in the more proximal part, much steeper than was commonly the case in the more remote geological past. The sedimentological model of sandar should, as a consequence, be adapted to account for more representative situations, such as those that existed when the sandar in NE Poland developed.

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