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GRAVITY FLOW ORIGIN OF GLACIOLACUSTRINE SEDIMENTS IN A TECTONICALLY ACTIVE BASIN (PLEISTOCENE, CENTRAL POLAND)

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Abstract. Pleistocene glaciolacustrine sediments of the Kleszczów Graben near Bełchatów – silts and clays with numerous sand intercalations – record the development and decay of a glacial lake in a subsiding basin. The sediments were classified in eight genetic facies. These facies present a wide scope of sediments of subaqueous redeposition and deposition in the basin – from dense mass flows of debris-flow type, through liquefied flows and turbidity currents, to low-energy bottom currents. The facies make the ground for identification of subenvironments and environments in the Pleistocene glacial lake. The individual complexes correspond to various environments within the lake: inclined lake bottom dominated by dense mass flows and fluidal flows, open lake with fine-grained parapelagic deposition, deepened axial part of the basin with weak bottom currents, marginal part of the lake with distal turbidity currents generated by river supply. Markov chain statistical analysis was used to determine the succession of sedimentary processes in the different lake environments. Vertical sequences of facies proved that the dense mass flows gradually evolved to more diluted Newtonian flows. The sequences typical of turbidites were noted as well as those of weak underflows generated by river inflows. The frequency and nature of the coarse-grained lake sediments formed by mass flows and turbidity currents testify to tectonically controlled style of sedimentation. Processes of violent redeposition were caused mainly by seismic tremors in the active Kleszczów Graben.

Abstrakt. Plejstoceńskie osady glacilimniczne rowu Kleszczowa (odkrywka bełchatowska) odzwierciedlają rozwój i zanik jeziora glacjalnego w warunkach subsydencji podłoża. Są to muły i iły z licznymi przewarstwieniami piaszczystymi. Wśród nich wyróżniono osiem facji genetycznych prezentujących szeroki wachlarz typów podwodnej redepozycji grawitacyjnej – od gęstych spływów masowych typu debris flow, przez spływy uwodnionego materiału i prądy turbiditowe, do niskoenergetycznych prądów dennych. Wyróżnione facje są podstawą identyfikacji subśrodowisk i środowisk w plejstoceńskim jeziorze. Kolejne analizowane kompleksy osadowe odpowiadają różnym środowiskom w zbiorniku. Podwodny skłon jeziora zdominowany był przez gęste spływy masowe i spływy uwodnionego materiału, natomiast otwarty zbiornik – przez depozycję parapelagiczną, w przegłębionej, osiowej części basenu przeważały niskoenergetyczne prądy denne, a w marginalnej części jeziora dystalne prądy zawiesinowe generowane dopływem rzecznym. Statystyczna metoda łańcuchów Markowa została zastosowana, by wyróżnić następstwo facji i procesów sedymentacyjnych w różnych środowiskach jeziora. Wyróżnione dzięki tej metodzie pionowe następstwo facji reprezentuje stopniowe przejście od gęstych spływów masowych do przepływów hydraulicznych. Notowano również sekwencje typowe dla turbiditów oraz dla słabych prądów dennych powodowanych dopływami rzecznymi. Charakter gruboziarnistych osadów jeziornych, będących rezultatem spływów masowych i prądów turbiditowych, wskazuje na tektoniczne uwarunkowania sedymentacji limnoglacjalnej. Procesy redepozycji wzbudzane były najprawdopodobniej przez wstrząsy sejsmiczne w aktywnym rowie Kleszczowa.

Key words: lithofacies analysis, glaciolacustrine sediments, gravity flow, Pleistocene, Kleszczów Graben.

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INTRODUCTION

Subaqueous sediment gravity flows have been the subject of discussion for many years. The general term refers to the sediment-fluid mixtures, in which gravity acts directly on the grains, to cause downslope movement (Middleton & Hampton, 1973). There are four main types of sediment flows: *debris flow* – in which larger grains are supported by

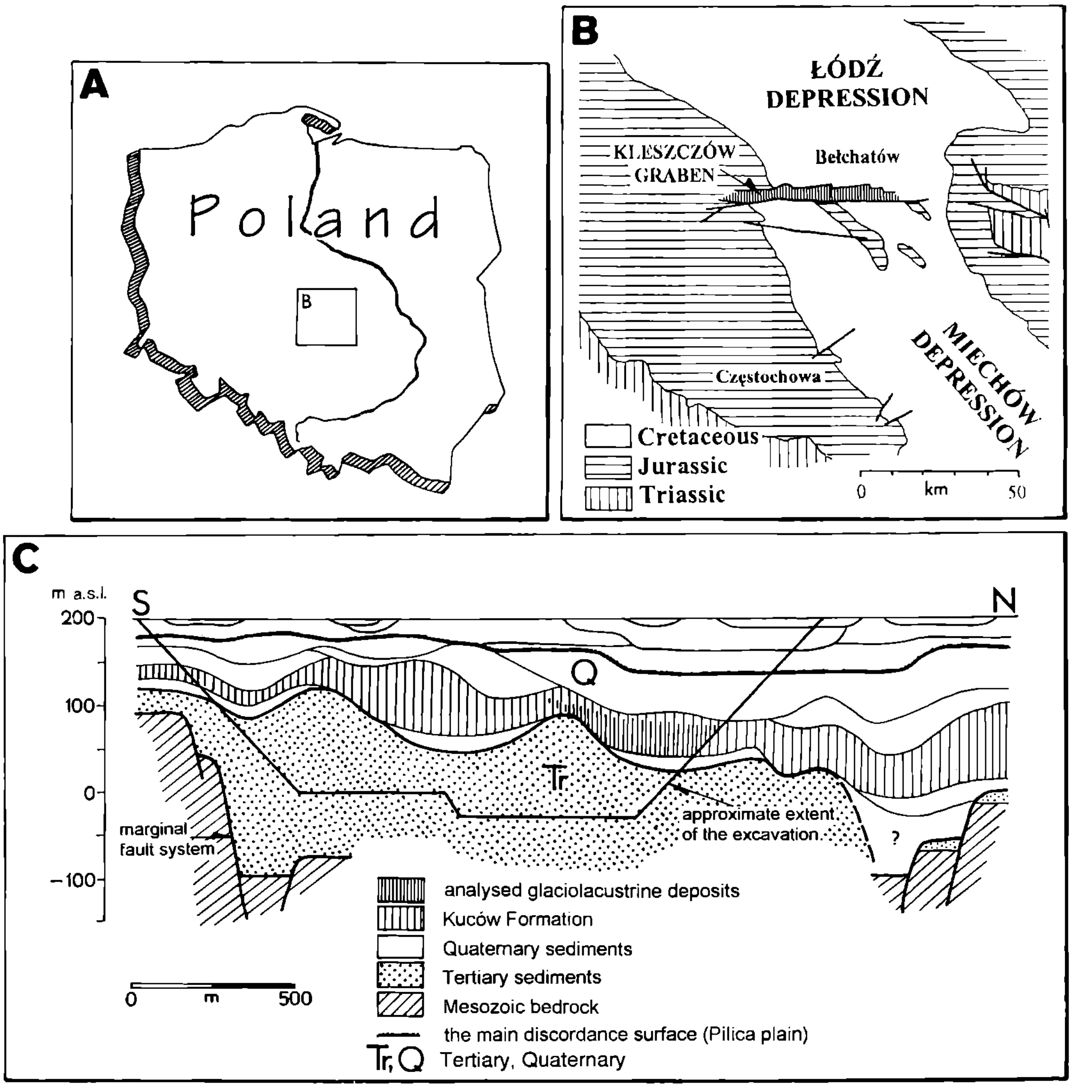

Fig. 1. Location and stratigraphy of analysed sediments. **A** – location of the sediments under study; **B** – geological setting of the Kleszczów Graben and its vicinity; **C** – schematic section of the Quaternary sediments of the Kleszczów Graben

matrix, *grain flow* – in which sediments are supported by grain-to-grain interactions, *fluidized sediment flow* – in which grains are supported by upward escape of fluid from between the grains, and finally *turbidity current* – where turbulence keeps grains in suspension. This division, based on flow rheology and mechanics, is widely accepted by geologists.

Although such sediments are most often associated with marine environment, they have been also described from glacial lakes by many authors (Harrison,1975; Cohen, 1983; Eyles & Eyles, 1983; Postma *et al.*, 1983; Eyles, 1987; Eyles *et al.*, 1987; Eriksson, 1991; Fitzsimons, 1992), having become the integral part of glaciolacustrine deposits. There are many intermediate types of flow, in which different mechanisms act simultaneously during the transport and deposition of sediments. However, it happens very rarely that all the mentioned gravity sediment types are found in sequence of one glacial lake. Quaternary tectonic activity in the Kleszczów Graben was the reason for enrichment of glaciolacustrine series in various types of gravitationally redeposited sediments. It lets us present remarkable lithological examples of the main types of subaqueous gravity sediments as well as distinguish a few transitional modifications.

GEOLOGICAL SETTING

The Bełchatów brown-coal opencast mine in central Poland presents a good opportunity for investigation of glacigenic sediments. The uppermost 200 m of Tertiary and Quaternary sediments are well exposed in the outcrop walls.

The outcrop is located in the Kleszczów graben – in the southern part of Łódź Upland, within the Szczecin-Łódź-Miechów Synclinorum (Pożaryski, 1971). The graben itself has an east-west direction, it is about 50 km long and 4-5 km wide (Fig. 1B; Biernat, 1968; Pożaryski, 1971; Ciuk, 1980). It originated due to movement of basement faults reactivated during the Alpine orogeny. The direction of the graben is almost parallel to the structural trend in the deep basement. The geology of the graben is rather complex. The

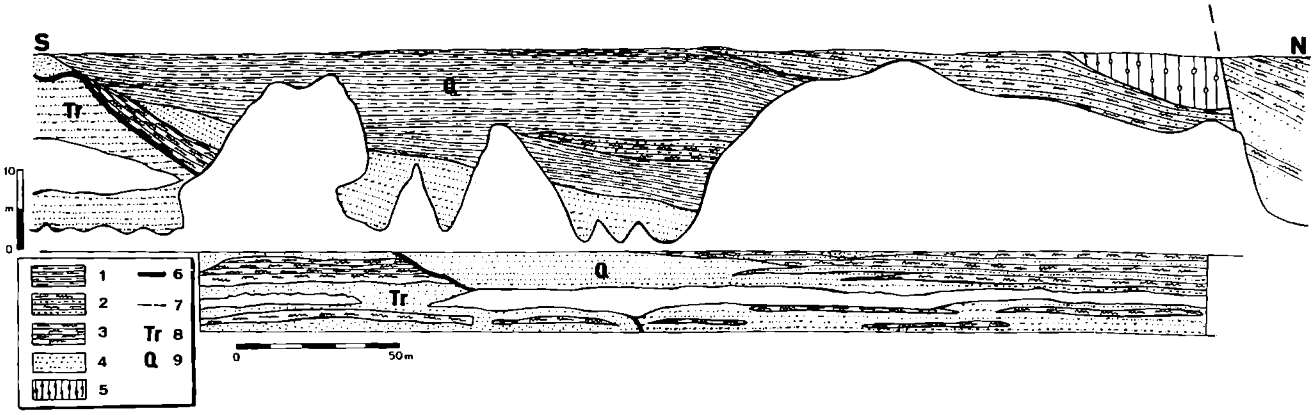

Fig. 2. A view of the two pit walls with studied glacigenic deposits. *1* – varved clays; *2* – silts with sand intercalations; *3* – silts; *4* – sands; *5* – till; *6* – Tertiary/Quaternary boundary; *7* – fault; *8* – Tertiary deposits; *9* – Quaternary deposits

basement is made of Carboniferous sandstones deformed during the Variscan orogeny, overlain by sandstones, limestones, marls and shales of Mesozoic age, which together were folded and faulted during the Laramide phase of the Alpine orogeny (Baraniecka, 1971; Kossowski, 1974; Brodzikowski *et al.*, 1987).

The Kleszczów graben is infilled with Neogene and Quaternary deposits (Fig. 1C). Most of the exposed Tertiary rocks consist of brown coal, which is being exploited. The Quaternary sequence is 50 - 250 m thick and is 3 to 5 times thicker inside than outside the graben (Krzyszkowski, 1993). In the light of former studies of many authours, there is no doubt that the trough was tectonically active through the whole Pleistocene, and continues to be so today (Baraniecka, 1971, 1975; Pożaryski, 1977; Brodzikowski *et al.*, 1987; Brodzikowski *et al.*, 1987a; Krzyszkowski & Brodzikowski, 1987; Krzyszkowski, 1989, 1993). Stratigraphically the Quaternary deposits of the graben are from Cromerian to Holocene (Krzyszkowski, 1991) and consist of glacial, glaciodeltaic, glaciolacustrine and glaciofluvial sediments. Non-glacial deposits are of restricted occurrence within the trough.

Several Scandinavian glaciations reached the Kleszczów graben area (Baraniecka & Sarnacka, 1971; Różycki, 1978; Lindner, 1982); this is documented by eight glacial cycles distinguished within the Pleistocene deposits (Krzyszkowski, 1991). Each cycle represents a separate period of glacier advance and retreat. There are two distinct structural units within the Quaternary sediments. The lower one is deformed into large-scale folds and flexures, and also displays smaller-scale deformation due to endogenic processes (Brodzikowski, 1985). The upper unit lies more or less horizontally and exhibits medium-scale glaciotectonic deformations only. These two units are separated by distinct erosional surface called the main discordance surface (Krzyszkowski, 1993) or the Pilica plain (Brodzikowski & Gotowała, 1980; see Fig. 1C) which originated during the Pilica Interstadial (Baraniecka, 1990; Krzyszkowski, 1991) – an equivalent of the Lubawski Interglacial (Lindner, 1990). Twelve formations of Quaternary age, understood as lithostratigraphic units representing particular glacial or interglacial cycles, were found and described within the Quaternary deposits exposed in the outcrop. Among them there are six glacial formations with seven till horizons. Glaciolacustrine sediments are of considerable thickness and are noted in four glacial formations: Folwark, Kuców, Ławki and Stawek (Krzyszkowski, 1992, 1993). The Kuców (of Saalian Glaciation age) and Ławki (of Odranian Glaciation) Formations display the biggest thickness of glaciolacustrine deposits.

The glaciolacustrine sediments of the Kuców Formation were investigated in 1991. They attain a thickness of 80 m and display complex internal structure with a wide variety of subaqueous sediment types. The analysed deposits represent the complete profile of a glacial lake. Cross-sections (Fig. 2) show that sedimentation took place in a narrow bay, which was a branch of a much bigger lake. The bay was about 350 m wide and was relatively deep – it was undergoing syndepositional subsidence. Large exposures gave excellent possibility of detailed sedimentological studies on different types of glaciolacustrine deposits. In particular the co-existence of varved clay and coarse sands, sometimes with gravels, of undoubtedly subaqueous origin attracted the authors' attention.

OBJECTIVES

The glaciolacustrine deposits studied in the section represent two different styles of sedimentation which adjoin vertically. These are parapelagic deposition of fine-grained (clayey) sediments – by which the authors mean pelagic deposition in lakes (Brodzikowski, 1993) – and deposition of coarse-grained (sandy) sediments by subaqueous gravity flows and bottom currents.

This paper is intended to recognize various subaqueous gravity flow facies within glaciolacustrine sediments of the Kleszczów graben. Detailed analyses of texture and structure of beds were performed to define the processes acting during the transport and deposition. These analyses let us distinguish particular types of flows. A classification of flow types, according the flow density, is presented at the end of this paper. The analysis of spatial variability of flow types

Table 1

Symbols of lithofacies used in this study

SYMBOL	GRAIN SIZE	STRUCTURE
SGt	gravelly sands	trough cross-stratification
SGm		massive structure
St	sands	trough cross-stratification
Sp		planar cross-stratification
Sl		low-angle cross-stratification
Sh		horizontal stratification
Sr		ripple cross-lamination
Src		ripple-drift cross-lamination
Sw		wavy lamination
Sm		massive structure
SFr	silty sands	ripple cross-lamination
SFw		wavy lamination
Fh	silts or clays	horizontal lamination
Fv		varved

	lithofacies	lithofacies association	lithofacies complex
description	an elementary sedimentary unit (set, coset, bed) e.g. *a single bed of ripple cross-laminated sand*	an association of sedimentary units e.g. *a package of ripple cross-laminated sands interbedded with horizontally laminated silts*	a lithosom of related lithofacies associations e.g. *a lithosome of varved clays, laminated silts, subordinately thin sand intercalations*
symbol	*Sr*	*Sr,Fh*	*c*
interpretation	the bedform type and hydrodynamic conditions of flow e.g. rippled sand bed formed by weak current	sedimentary subenvironment e.g. underflow currents activity alternatively with parapelagic settling	sedimentary environment e.g. central part of the lake with predomination of parapelagic deposition, rare phases of underflow currents

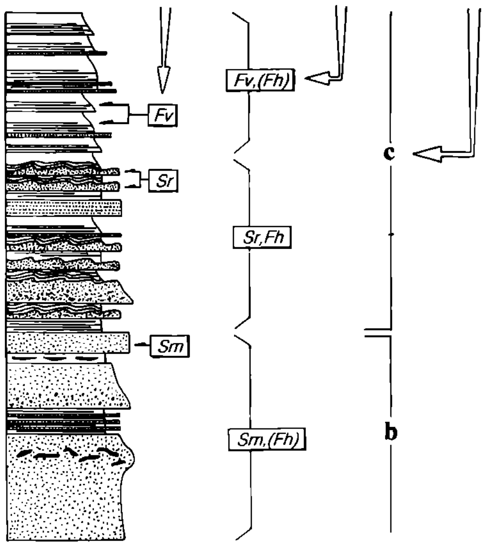

Fig. 3. Threefold lithofacies division used in study. See text for details

allowed us to evaluate the differences between the subenvironments in the lake, i.e. to decide if there was a difference in flow types between the marginal and central part of it. Markov chain analysis was carried out to define the typical glaciolacustrine depositional sequences. Well-developed lithological varieties of subaqueous gravity flows, reported from the glaciolacustrine succession of the Kleszczów Graben, can be used as a particularly good guide for identification of similar facies in other fossil sediments of uncertain nature.

METHODS

Field data underwent a three-step sedimentological analysis (Fig. 3). Description of the distribution and lithology as well as interpretation of lithofacies, i.e. the elementary depositional units (sets, cosets, beds), make up the first step of the analysis. The lithofacies are labelled using a slightly modified Miall's (1985) code – see Tab. 1. The modifications concern the textural and structural parameters. Mixed symbols of texture were introduced, e.g. *SG* (gravely sand) or *SF* (silty sand), to make the description of the lithofacies more precise. For the structural symbols, a lower-case letter – the first letter of the adjective connected with the type of depositional structure – was used. Hence all horizontal laminations, in sand as well as in silts and clays, are marked by letter *h* (horizontal), whereas letter *l* is limited to low-angle cross-stratification only. The letter *v* is used to indicate the varved rhythm of deposits (in lithofacies *Fv*).

The term "genetic facies", which appears many times in this article, is used to indicate a sediment type connected with unequivocally defined depositional process. Genetic facies are indicated by the capital letters that Miall used (1985) in his architectural-element analysis.

Analysis of the distribution of lithofacies associations is the second phase of proceeding. Each association is indicated by the symbols of the lithofacies that are most common in that association. These are index lithofacies – the most frequently noted and the most typical of the association. Lithofacies that are frequently noted, but do not dominate, are indicated in brackets, e.g. lithofacies association *Sm,(Fh)*. The lithofacies association is regarded as a product of a sedimentary subenvironment.

Lithofacies complexes were investigated in the third phase of analysis. Each complex is a collection of related lithofacies associations that originated in similar sedimentary conditions (e.g. in the central part of a glacial lake). Lithofacies complexes are indicated with lower-case letters (Fig. 3).

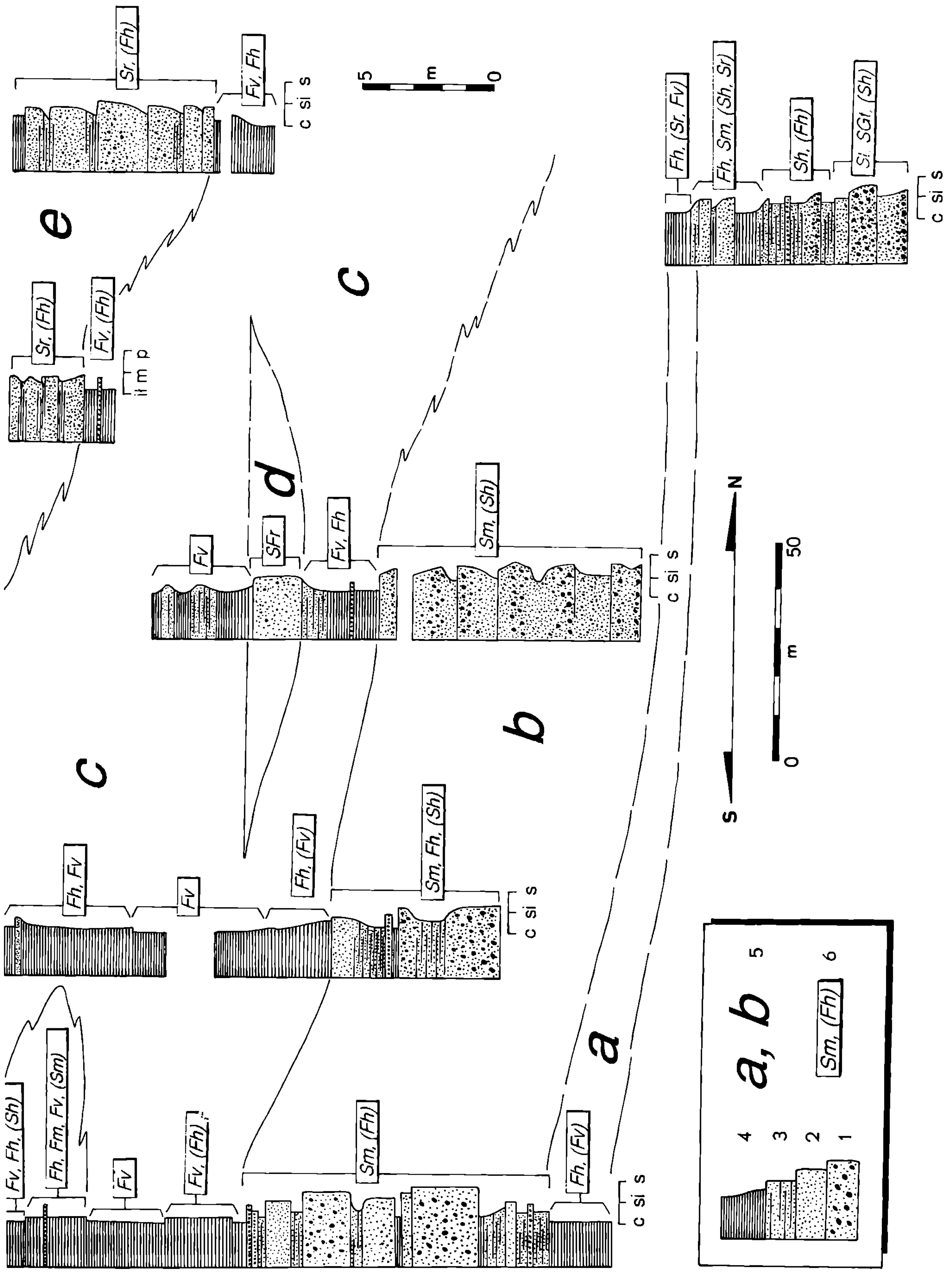

Fig. 4. Synthetic sedimentary logs of section studied. Key for shading and symbols: *1* – gravelly sand; *2* – sand; *3* – silty sand; *4* – silt and clay; *5* – lithofacies complexes; *6* – lithofacies associations

Fig. 5. The upper part of sandy complex *b* and the transition to clays of complex *c*. Note the distinct textural differences between complexes

CHARACTERISTICS OF LITHOFACIES COMPLEXES

Among all glaciolacustrine deposits there are two basic lithological types: clayey and sandy ones. The most interesting are coarse-grained sediments that cannot have been formed as the result of "normal" parapelagic deposition.

Below the glacial-lake complexes there are sandy and sandy-gravelly deposits that include the *St,SGt,(Sh)* lithofacies association at the base (Fig. 4). These are deposits of fluvial origin (proglacial or extraglacial rivers). The predominance of large-scale trough cross-stratification (channel structures) unequivocally indicates a sedimentary subenvironment of a deep channel, characterised by high-energy, unsteady flows, typical of ablation outflows. There are two lithofacies associations on top of them: *Sh,(Fh)* and *Fh,Sm,(Sh,Sr)* – see Fig. 4. Lithofacies of silts and clays with horizontal lamination coexist there with sandy lithofacies of small and average scale (of thickness). Within the lithofacies association *Sh,(Fh)*, a very characteristic trend is that both the thickness of the lithofacies and the average grain size decrease towards the top. Thus the transition from the association dominated by sandy lithofacies towards the association characterized by considerable amount of silts and clays is gradual. Both of these lithofacies associations are intermediate between typical fluvial and glaciolacustrine sediments.

Five lithofacies complexes, which distinctly differ from each other, have been specified within glaciolacustrine sediments. They represent successive phases of the lake development.

Complex *a* – the oldest one – is built of two lithofacies associations: *Fh,(Fv)* and *Fh,(Sr,Fv)* (Fig. 4). Varved clays, horizontally laminated silts interbedded with laminae of fine sand and silty sand, and ripple cross-laminated sands are the main lithofacies in this complex. In the association *Fh,(Fv)* we noted lower frequency and smaller thickness of rippled beds in comparison with the association *Fh,(Sr,Fv)*. Small-scale gradational cycles are present in both associations (they exhibit the succession from horizontally laminated silt, sometimes with intercalations of ripple laminated sands, to massive clay). In the whole complex, an upward decrease of the thickness of cycles is observed.

The deposits of complex *a* originated during the first phase of glaciolacustrine sedimentation. Conditions of deposition were relatively stable. Parapelagic sedimentation dominated in the shallow lake. Sands were supplied from the north. *Fh,(Sr,Fv)* lithofacies association represents the deposits of the marginal part of glacial lake.

Complex *b* embraces three laterally existing lithofacies associations: *Sm,(Fh); Sm,Fh,(Sh); Sm,(Sh)* – Fig. 4. It is characteristic for them that sandy and sandy-gravelly beds of *Sm* type, relatively thick, dominate in the whole complex. It is the most coarse-grained complex among all noted in the glaciolacustrine succession of the series studied (Fig. 5). Its genesis is identified with intense activity of gravity flows and bottom currents, connected with the period of increased tectonic activity of the Kleszczów graben. Detailed description and interpretation of the lithofacies are the subject of the next section.

Above the complexes described, there is the thickest and the most fine-grained complex *c* (Figs. 2, 4). All the lithofacies associations are characterised by domination of horizontally laminated silts *Fh* and clays developed as varves *Fv*. The middle part of the complex is built of lithofacies association *Fv*, where varved clays clearly dominate. They are found in 3-cm cycles (thin laminae of sand→horizontally laminated clay→massive clay). The scale of cycles increases both upwards and downwards in the association. Subordinately there are horizons of current ripple marks and horizontally laminated sands, sands displaying typical Bouma sequences, and layers of deformed clayey diamictons. At the base and at the top of the complex *c* there are some more complex associations – *Fv,Fh* and *Fv,(Fh)* – but still dominated by varved clay *Fv* and horizontally laminated silts *Fh*. Additionally there are horizons of sand ripples and wavy laminated silts. The contribution of current structures (silts and silty sands of ripple-drift cross-lamination) and thin layers of normally graded sands, increases in the associations at the top of the complex. The whole complex originated during steady, parapelagic sedimentation that was incidentally disturbed by bottom currents of various intensity. Within the complex *c*, in northern part of it, there is an indistinct marginal zone (Fig. 4). Within lithofacies associations *Fh,Fm,Fv,(Sm)* and *Fv,Fh,(Sh)*, the sandy lithofacies are of secondary importance.

Within complex *c*, in the central part of the lake sequence, there is an approximately continuous horizon of sandy-silty deposits, about 2 m thick and over 100 m wide. This is complex *d* (see Fig. 4), built of silty sands or sandy silts with flaser and lenticular laminations as well as ripple cross-laminated sands, which together make *SFr* lithofacies association. The deposition of this complex was connected most probably with bottom currents acting in the deepest axial part of the elongated bay of the lake. Taking into consideration the fact that the analysed cross-section cuts the elongated bay of the lake transversely, the isolation of sandy complex *d* within the clayey complex *c* – without any apparent connection with the marginal part of the lake – becomes clearly explicable.

In the upper part of analysed profile, complex *c* gradationally passes into sands, silts and clays of complex *e*. It is built of *Sr,(Fh)* lithofacies association (Fig. 4). Ripple cross-laminated sands dominate there over horizontally laminated silts and clays. This complex occurs in the north part of the lake only. It originated due to filling the marginal part of the basin with sandy deposits delivered by a river. Similar to the situation in complex *a*, the main inflows were coming to the lake basin from the north, i.e. from the melting glacier.

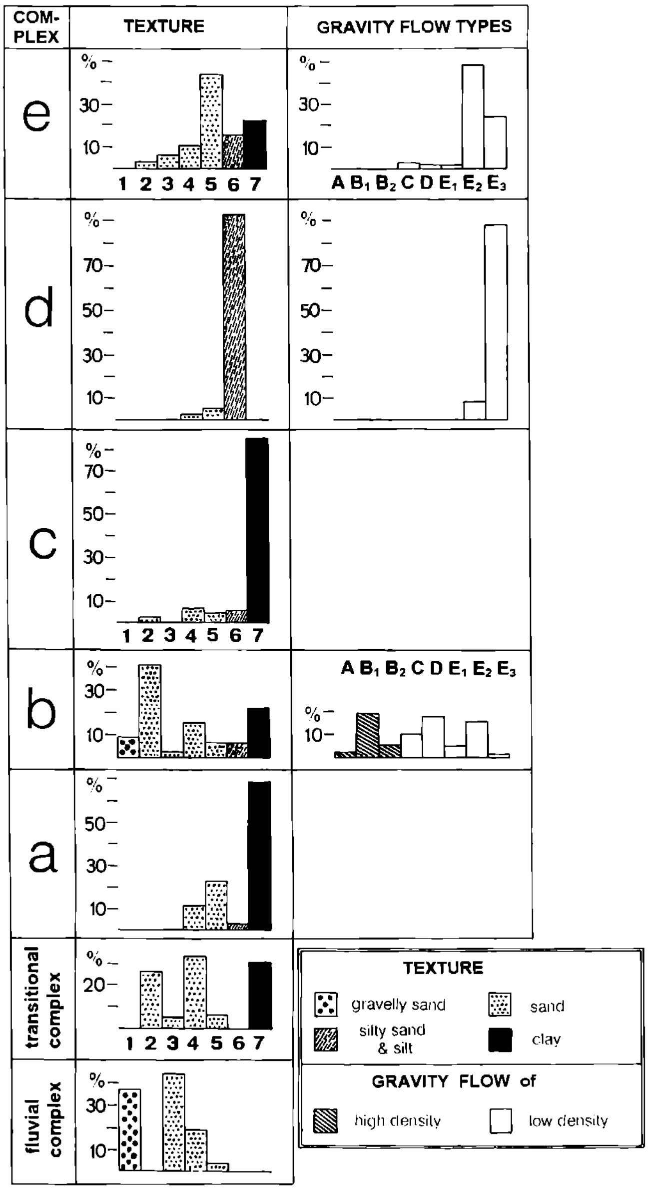

Fig. 6. Fluvial, transitional and glaciolacustrne complexes – lithological characteristics and genetic interpretation of sediments. *1* – stratified and massive gravelly sands; *2* – massive sands; *3* – cross-stratified sands; *4* – horizontally laminated sands; *5* – ripple cross-laminated sands; *6* – silty sands and silts of ripple cross-lamination; *7* – horizontally laminated, massive and varved clays

LITHOLOGY AND ORIGIN OF THE LAKE DEPOSITS

As a result of more detailed study, eight types of glaciolacustrine deposits have been distinguished within sandy complexes *b*, *d* and *e*. They can be related to particular gravity flow types or bottom currents.

A – coarse-grained diamicton (clayey breccia)

Description. This is a rare lithofacies among glaciolacustrine deposits, noted only in *Sm,(Fh); Sm,Fh,(Sh)* and *Sm,(Sh)* lithofacies associations in complex *b*, where it makes up 2 percent of all the sediments (Fig. 6). These are thin sheet beds (up to 10 cm), built of clay clasts, mainly disrupted and deformed laminae (Fig. 7). The beds are clast-supported and texturally homogenous. The matrix is sandy-silty, the structure is always massive and beds commonly have erosional bases.

Interpretation. This coarse-grained diamicton is genetically identified with *clast-rich debris flow*, i.e. *slurry flow* sensu Carter (1975). Transport of clasts in suspension could act relatively easily, because of the low density of clay in comparison with quartz (1.94 g/cm^3 and 2.65 g/cm^3 respectively). Cohesional strength and frictional strength were dominating transport factors while turbulence was distinctly limited. The flow was strongly saturated, as proved by clast-supported texture and the small bed thickness. Be-

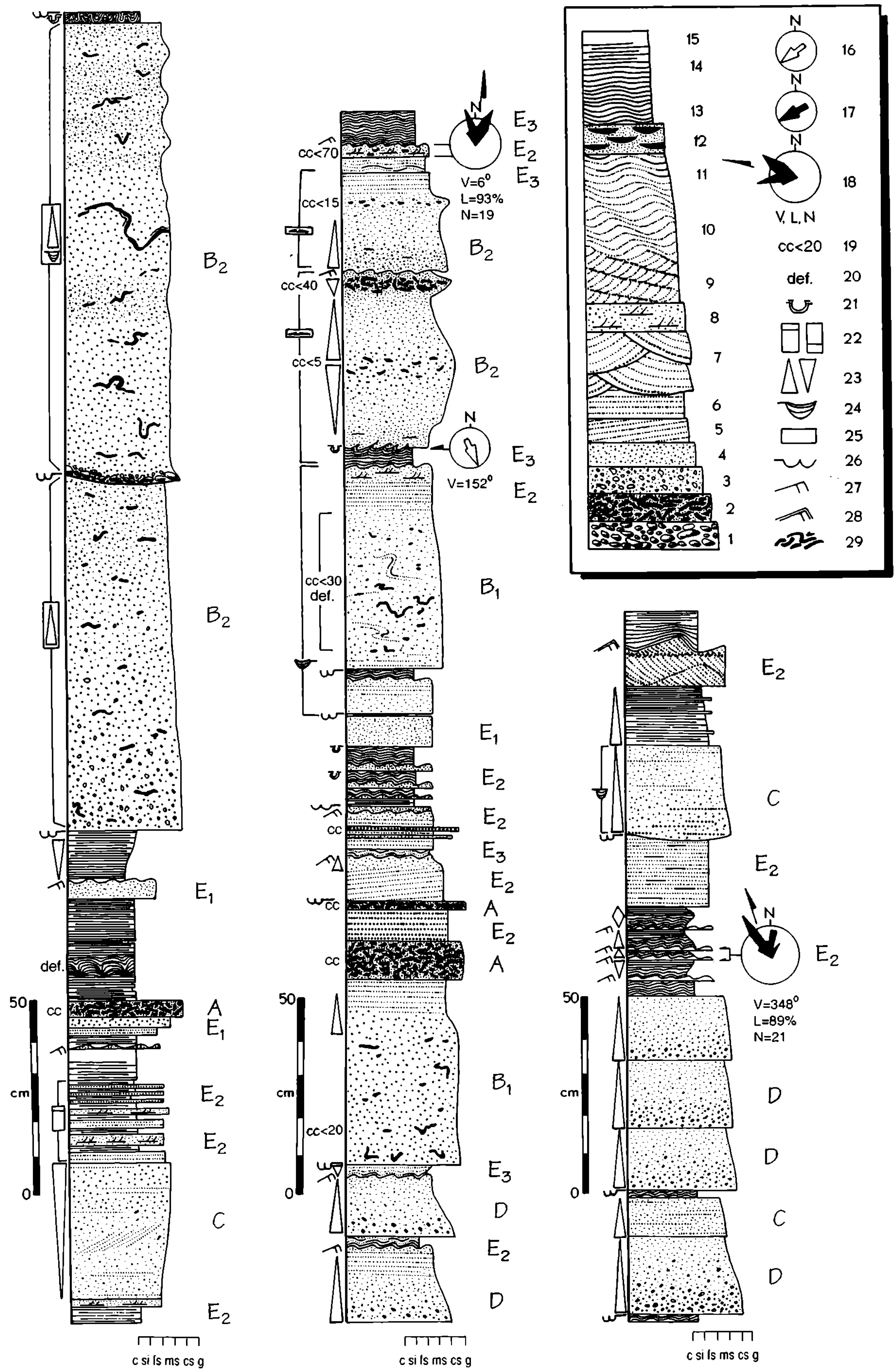
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Fig. 7. Selected sedimentary logs of complex *b*. Key for shading and symbols: *1* – massive gravels; *2* – clay clast breccia; *3* – massive gravelly sand; *4* – massive sand; *5* – sand with low-angle cross-stratification; *6* – horizontally laminated sand; *7* – sand with trough cross-stratification; *8* – sand with ripple cross-lamination; sand with ripple-drift cross-lamination: *9* – A type, *10* – B type; *11* – silty sand with wavy lamination; *12* – sand and silt/clay with flaser and lenticular lamination; *13* – silt or clay with wavy lamination; *14* – silt or clay with horizontal lamination; *15* – massive silt or clay; *16* – mean azimuth of fold plane dip without statistical significance; *17* – mean azimuth of cross-bed dip without statistical significance; *18* – directional distribution of cross-bed dip statistically significant on 95% level, where *v* – vector mean azimuth, *L* – vector magnitude (both parameters by Curray, 1958), *N* – number of readings; *19* – clay clasts (diameter in mm); *20* – deformed sediments; *21* – load casts; *22* – thinning (left) and thickening (right) upward sequences; *23* – fining (left) and coarsening (right) upward cycles; *24* – channel-fill structure; *25* – sheet bed; *26* – erosional contact; *27* – fossil ripples; *28* – fossil dunes; *29* – clay clasts

cause of this, the slurry spread on quite long distances. It was deposited by "freezing" (i.e. very sudden, "en masse" deposition).

This lithofacies derived from the densest type of mass flow noted within the studied sediments. Such flows originated during sporadically occurring processes of disintegration of clayey lake bed, caused most probably by earthquakes in the active Kleszczów graben. Slumping in the near-shore zone of the lake was the direct source of wide, sheet-like debris flows. Analogous thin beds connected with subaqueous debris flow were noted in glaciolacustrine deposits by Cohen (1979) and Visser (1983) and in lacustrine ones by Eriksson (1991). It is interesting that glaciolacustrine facies *A* is quite similar to the facies of contorted and brecciated mudstone clasts within sandy matrix noted by Shanmugam *et al.* (1994) in the Cretaceous marine debris flows "frozen" after short-distance transport.

B_l – matrix-supported massive sand with clayey clasts

Description. This is the most common lithofacies in complex *b*, where it forms 20% of all deposits (Fig. 6). Coarse-grained sands and coarse-grained sands with dispersed pebbles make up this lithofacies. The sediments are poorly sorted ($0.8 < \delta_I < 3.1$ phi). Uniformly dispersed clayey clasts within matrix are characteristic feature of facies B_l (Fig. 7). The clasts are most often of disrupted and deformed laminae of clay (Fig. 8). The beds are usually texturally homogeneous, but in some cases sandy matrix and gravels show gentle coarsening-up grading. The structure is always massive (Fig. 9). In some beds of B_l lithofacies, indistinct flowage folds were noted. The bases of beds are distinct, sometimes lithofacies B_l passes gradationally upwards into horizontally laminated sands (Fig. 7).

Lithofacies B_l was noted in two variants. Large-scale beds (up to 1.2 m) with erosional bases (Fig. 9) make the first variant. In places they were found in large channels up to 12 m wide, in which case they form thin layers of pavement at the bases of them. Beds of medium scale – in sheets 10 - 60 cm thick – make the second variant, which is more fine-grained (Fig. 10).

Interpretation. Facies B_l is an equivalent of the pseudoplastic debris flow of Shultz (1984). The significant thickness of the beds, the grain-size homogeneity, massive structure and presence of flowage folds at the base, testify to that conclusion. The presence of angular clayey clasts, uniformly dispersed within beds, unequivocally indicates relatively high density of flow. Clay clasts as well as rock granules and pebbles were suspended by the buoyancy force

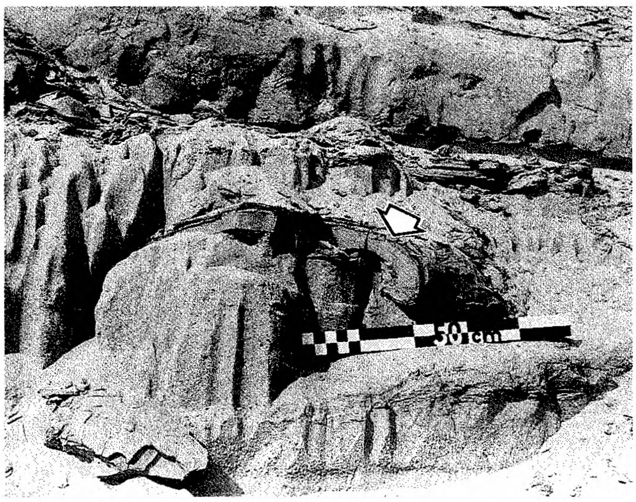

Fig. 8. Facies B_l (debris flow deposit) – large deformed clay clast in massive sand

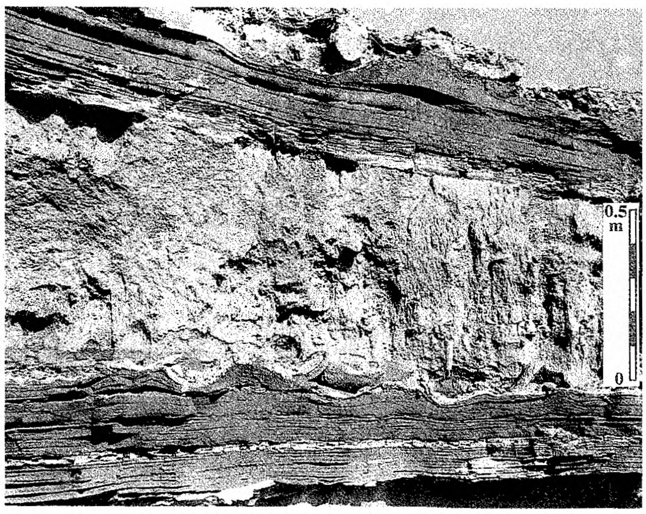

Fig. 9. Facies B_l (debris flow deposit). Note the eroded and deformed clay clasts at the base and the massive structure of sand

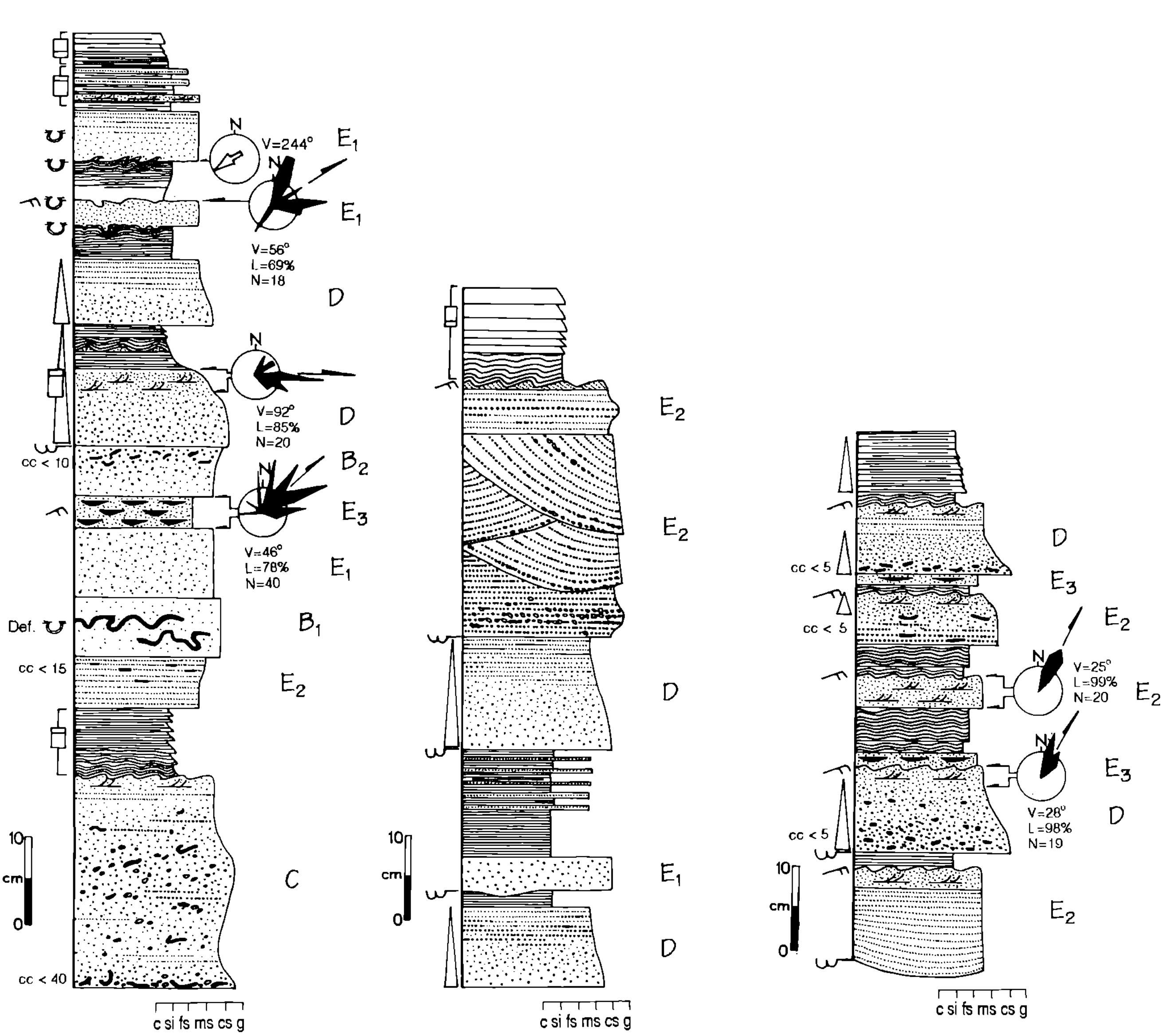

Fig. 10. Selected sedimentological logs of complex *b*. For explanation see Fig. 7

of liquefied sand and by pore water escape. Turbulence played a significant role only where fining-up tendency of granules is noted now. These flows were more diluted than the *A* type ones. On the other hand, the presence of flowage folds proves the still dense character of flow – these beds were deposited by flow characterised by high viscosity (*viscous flow*). The presence of distinct top surfaces proves that the beds did not undergo syndepositional reworking by water turbulence. There is nothing unusual that such deposits have been found in erosional channels. Postma *et al.* (1983), Eyles & Clark (1988), Myrow & Hiscott (1991), described analogous examples where sediments of debris flow filled plugged troughs. Subaqueous diamictons of debris flow origin, showing similar lithology were noted by Broster & Hicock (1985), Eyles & Clark (1988), Liverman (1991),

***B2* – massive sands with clayey clasts horizon**

Description. Facies *B2* is noted in complex *b* and it is a lithological variant of facies *B1* (Fig. 7). In principle there are no textural differences between *B1* and *B2* beds. On the Passega diagram, these two facies occupy a common area – the most peculiar one in relation to all the other lithofacies (Fig. 11). Analogous dependence is seen in size-frequency distributions. The graphs show the existence of some modes in gravel-sand range (see *SGm* and *Sm* lithofacies on Fig. 12). The shape of plots is similar only to the sand-range part of distributions of clayey-sandy diamictons (subaqueous tills) from complex *c* (Fig. 12). Clayey clasts are smaller and distinctly rounded. The clasts are most frequently concentrated about 5 - 10 cm below the top of the beds (Figs. 7, 13). The *B2* beds are mostly about 25 - 60 cm thick. The thicker ones are usually found in erosional channels, while thinner beds are noted as sheets, with drag-deformed load structures (small diapir folds of the same vergence).

Interpretation. Gravity flow depositing facies *B2* was less dense than that depositing lithofacies *B1*. Sandy matrix was much more fluidised, so that clasts could easily migrate up to the top of the flow. Moreover, minor thickness of beds and lack of flow folds testify to lower density of debris flow in comparison to that of facies B_1. However that flow was not mobile enough for intergranular collisions to play the main role in transportation. Thus the flow had not the rheological nature of a grain flow. Unresistant clayey clasts would have been much more crumbled and rounded during such intensive flow.

To check our hypothesis about the density differences between B_1 and B_2 flows, the K coefficient (yield strength), i.e. resistance to deformation, was calculated. The parameter K is proportional to flow density ρ. Johnson's (1970) formula was used:

- for sheet debris flow: $K = d\,g\,S$ (1)
- for channelled debris flow: $K = d\,g\,\sin S / (2d / w)^2 + 1$ (2)

where: K – resistance to deformation [dn/cm^2], d – depth of flow, i.e. the bed thickness [cm], g – density of clasts [g/cm^3], S – slope inclination [degrees], w – width of channel flow [cm].

The average values of K, calculated from the two lithofacies are as follows: facies B_1 – channelled type - 510 dn/cm2, facies B1 – sheet type - The values obse of some post-depositional steepening of dip (due to subsidence). Fortunately, the relative differences of K parameter between facies are free of systematic error, thus they can be compared. The channellized flows of B_1 type were denser and were characterised by 20 percent higher resistance to deformation than the sheet flows of the same lithofacies. What is more, the value of parameter K calculated for the flows of B_1 type was 35 percent higher than that for the flows of B_2 type. Thus, the hypotheses based on lithologic features of sediments were fully confirmed by palaeorheologic calculations.

Both facies – B_1 and B_2 – are identified with liquefied flow, i.e. the gravity flow of rheological nature typical of density-modified grain flow *sensu* Lowe (1976) or with plastic-viscous grain flow *sensu* Carter & Norris (1977). The sediments were deposited by "freezing".

C – sands displaying indistinct cross-stratification

This is a secondary lithofacies which forms 10 percent of sediments in complex *b*. It occurs also in complex *e*, where it makes up 2 percent of sediments (Fig. 6).

Description. This lithofacies is formed of fine, medium and coarse-grained sands, sometimes with granules and well-rounded clayey clasts. Poorly developed horizontal lamination or cross-stratification existing locally among massive sands is a characteristic feature of this lithofacies (Figs. 7, 10). There is a hint that cross-stratification becomes more distinct towards the top of the beds. The beds are of medium thickness (15 - 75 cm) and they have erosional bases. Most often they are noted as sheets.

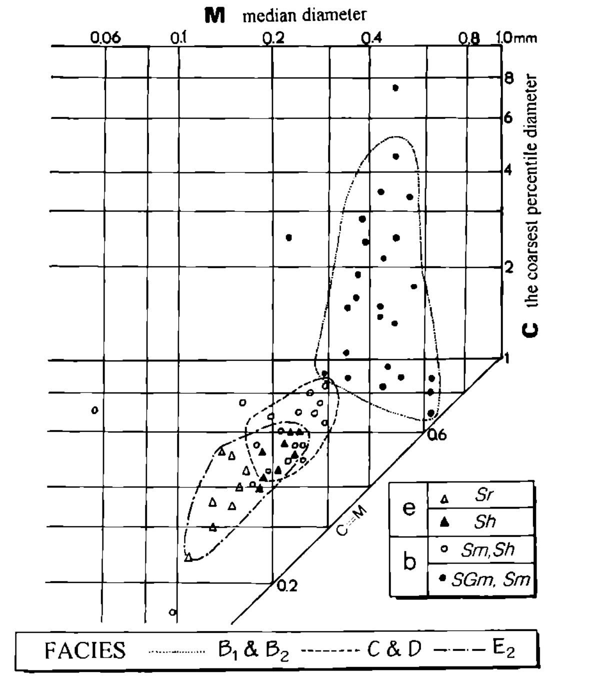

Fig. 11. Passega CM diagram. Three fields of facies originated in different dynamic conditions

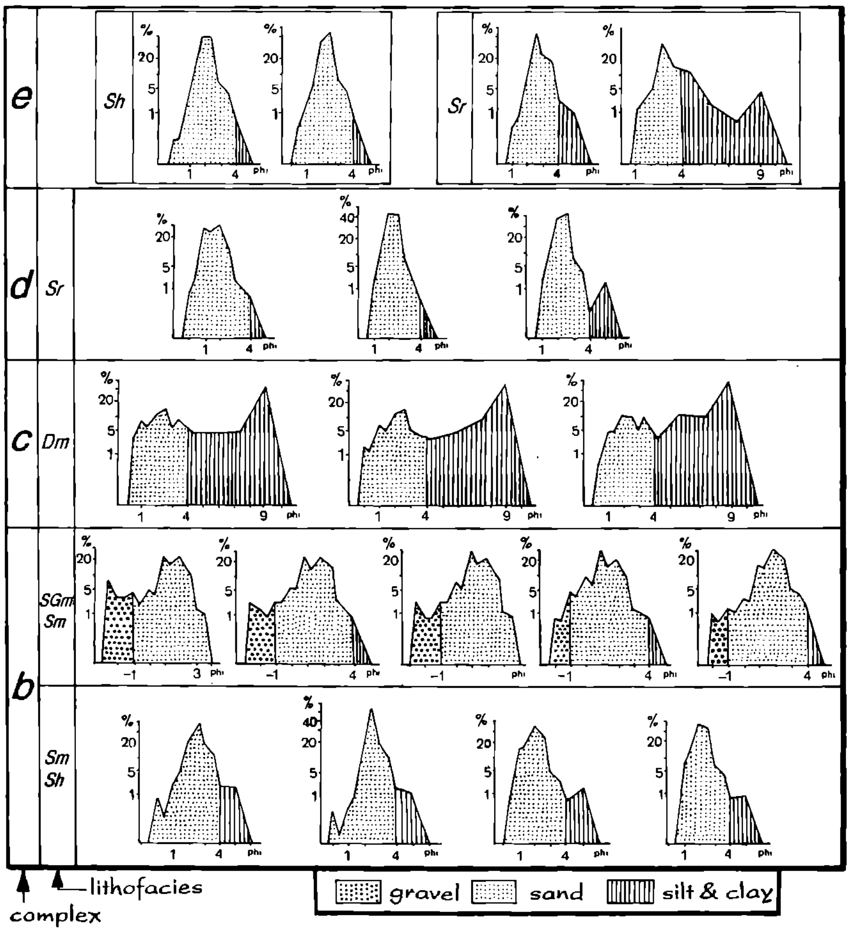

Fig. 12. Grain-size frequency distributions of deposits originated from mass and stream flows

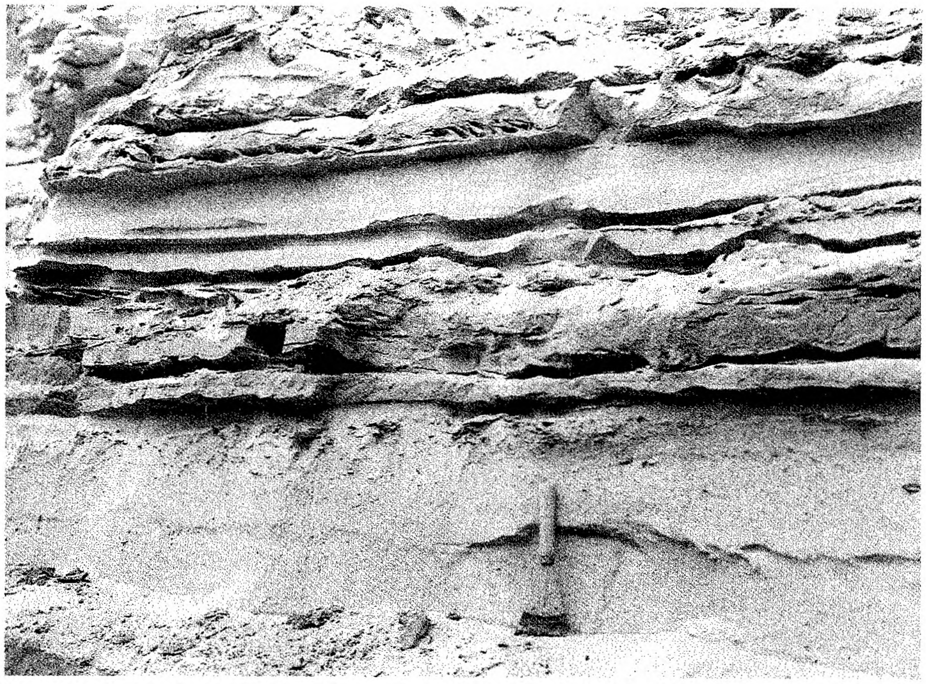

Fig. 13. Facies B_2 (density-modified grain flow). Sand bed with clay clasts in its upper part. Shovel is 20 cm long

Interpretation. Facies *C* is identified with intermediate flow (Bull, 1964), i.e. transitional between debris flow and stream flow (Nemec & Steel, 1984). This genetic interpretation is confirmed by the position of samples on the Passega diagram (Fig. 11). Facies *C* and *D* occupy a common area, which is transitional between facies B_1 and B_2 (treated as deposits of mass flows), and facies *E* (treated as deposits of stream flows). Lack of grain-size grading is the important feature of facies *C*. It proves that some yield strength compensated the gravity force. The uplifting forces keeping the coarse grains in suspension, could be a result of turbulence or buoyancy of matrix. Probably both above mentioned forces were operating simultaneously during transportation, and turbulence distinctly predominated over the buoyancy. Pierson & Scott (1985) proved that intensive turbulence was able to keep even the granules in suspension. Locally in that moderately dense and strongly turbulent environment, much more fluidised zones appeared, where stratified sediments were deposited, mainly in the conditions of upper flow regime.

Facies *C* is genetically identified with hyperconcentrated flow, that is with transportation in conditions of sliding bed *sensu* Saunderson (1977). Analogous indistinct cross-stratification sediments were noted by Eyles & Eyles (1989) and Myrow & Hiscott (1991), who came to similar genetic conclusion.

D – sands of normal grain-size grading

Description. This is a very common lithofacies in complex *b*, where it forms 20 percent of deposits. Additionally it was noted in complex *e* (Fig. 6). These are sheet-like beds displaying normal grading, 5 - 30 cm thick, commonly with erosional bases (Figs. 7, 10). Lower parts of the beds consist of coarse-grained sands, coarse-grained sands with gravels and/or clayey clasts (Fig. 14), and medium-grained sands. Clayey clasts are usually rounded and are, as a rule, concentrated close to the base of the bed. This part is always massive. It gradationally passes upwards into massive or horizontally laminated fine-grained sands (Fig. 7). They are commonly covered with ripple cross-laminated sands or with a thin horizon of current ripples (Figs. 10, 14).

Interpretation. Facies *D* is typical of turbidite flow

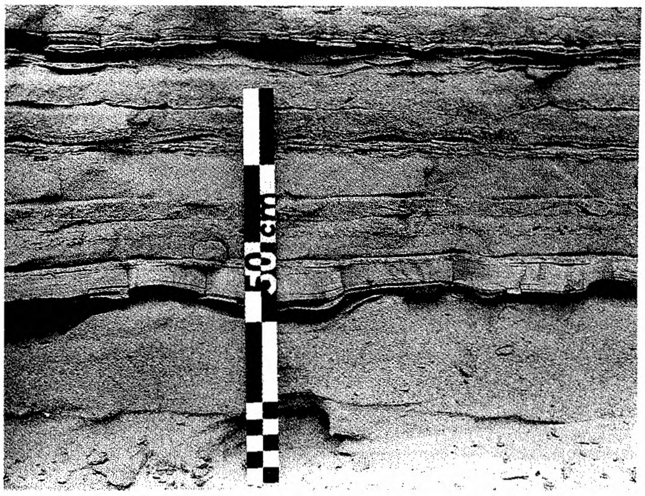

Fig. 14. Facies *D* at the base. Note massive sands with thinning upward trend of clasts and ripples forms at the top of the bed. The bed is capped with wavy laminated silts

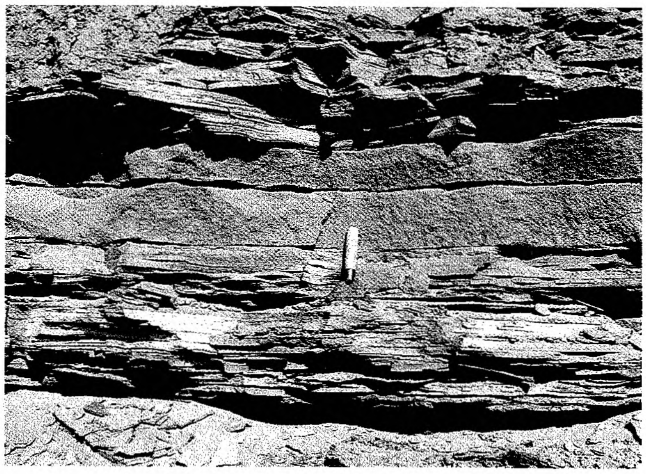

Fig. 15. Two beds of massive sands (facies E_1) between horizontally laminted silts. Shovel is 15 cm long

deposits, i.e. true turbidite flow where the main mechanism of grain transportation was controlled by turbulence (opposed to dense turbidity current *sensu* Lowe, 1982; Eyles *et al.*, 1987; Ghibaudo, 1992). Facies *D* usually displays typical Bouma (1962) sequence: *Sm*→*Sh*→*Sr*→*F*. At the base of the flow, the grains were transported as a traction carpet, i.e. by a flow similar to hyperconcentrated flow (Todd, 1989), that gave origin to *Sm* member. A little higher, or further, transport acted in conditions of typical stream flow and the grains were deposited in the upper plane bed (*Sh* member). Finally, low-energy accretion of rippled bed, in condition of lower part of lower flow regime, resulted in the origin of *Sr* member. Such turbidite flows spread over the lake bottom as subaqueous sheet flows.

Well-developed turbidite flows, in this basin, should be compared to proximal underflows. Their erosional bases (Lambert & Hsü, 1979; Teisseyre, 1983; Eyles, 1987; Syvitski *et al.*, 1988) as well as preserved lower members of the sequences (Dżułyński & Walton, 1965; Walker, 1967; Malik & Olszewska, 1984) testify to this conclusion. These sediments are much better developed than typical Quaternary glaciolacustrine turbidites (Donnelley & Harris, 1989; Huff, 1989; Liverman, 1991; Krzyszkowski, 1993). On the other hand, they are similar to Carboniferous glaciolacustrine turbidites described by Banerjee (1966). The beds described were not generated by river inflows (thickness of bed criterion – Pharo & Carmack, 1979). Most probably, facies *D* is the depositional effect of distal, dispersed gravity flows generated tectonically.

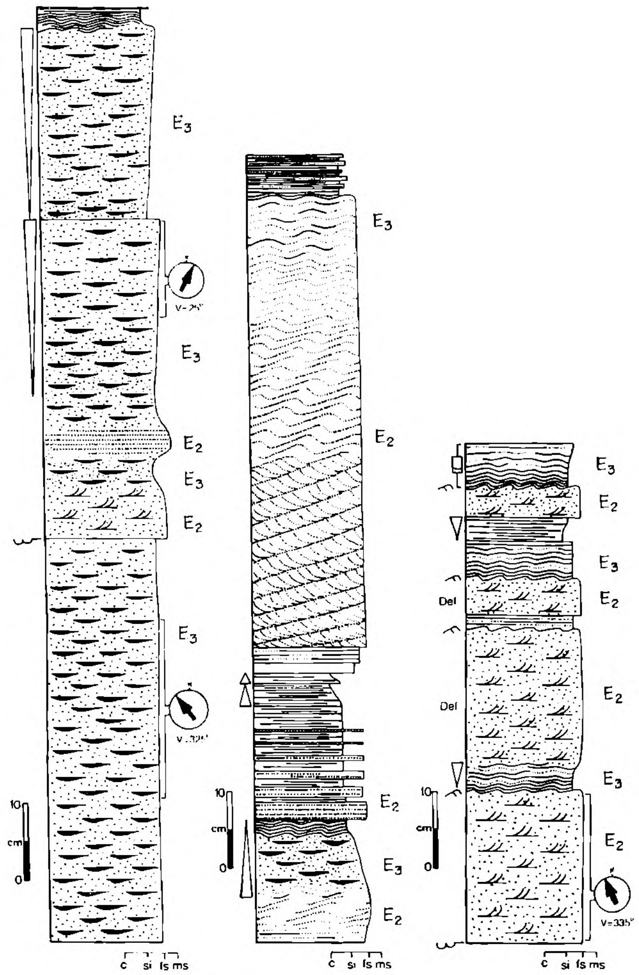

Fig. 16. Selected sedimentological logs of comlex *d*. For explanation see Fig. 7

E1 – *massive sands*

Description. This is rather an uncommon lithofacies in complex *b*, making up only about 5 percent of sediments (Fig. 6). It is found rarely in complex *e*. Two fundamental features differ this lithofacies from others, analysed earlier, *Sm* beds: its fine-grained character and small bed thickness. Facies *E1* is formed of fine-grained sands, sometimes of medium-grained sands with admixture of granules. The beds are texturally homogenous, always massive (Fig. 10). They are noted as sheet-like beds, about 5 - 10 cm thick (Fig. 15). At the base, they are usually loaded into underlying clays and silts (Fig. 10). These beds are most often noted as intercalations within horizontally laminated silts and clays (Fig. 14). Occasionally they are covered with horizontally laminated sands or the horizon of fossil current ripples (Fig. 10).

Interpretation. Lithofacies *E1* is the depositional effect of subaqueous stream flow of highly agradational character. It was de-

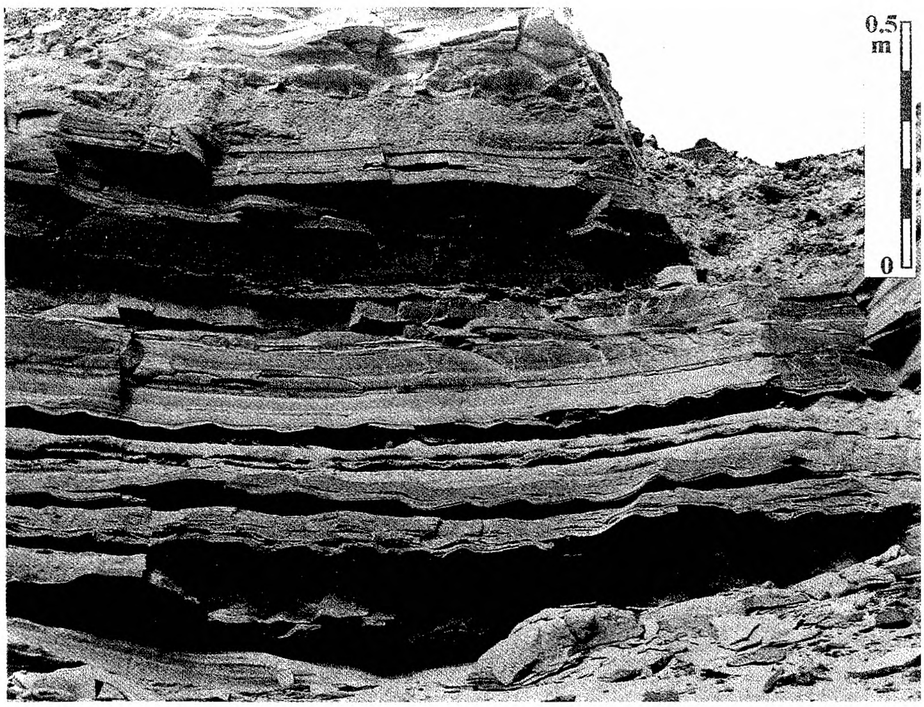

Fig. 17. A package of sandy rippled horizons (facies *E2*) interbedded with wavy laminated silts (facies *E3*)

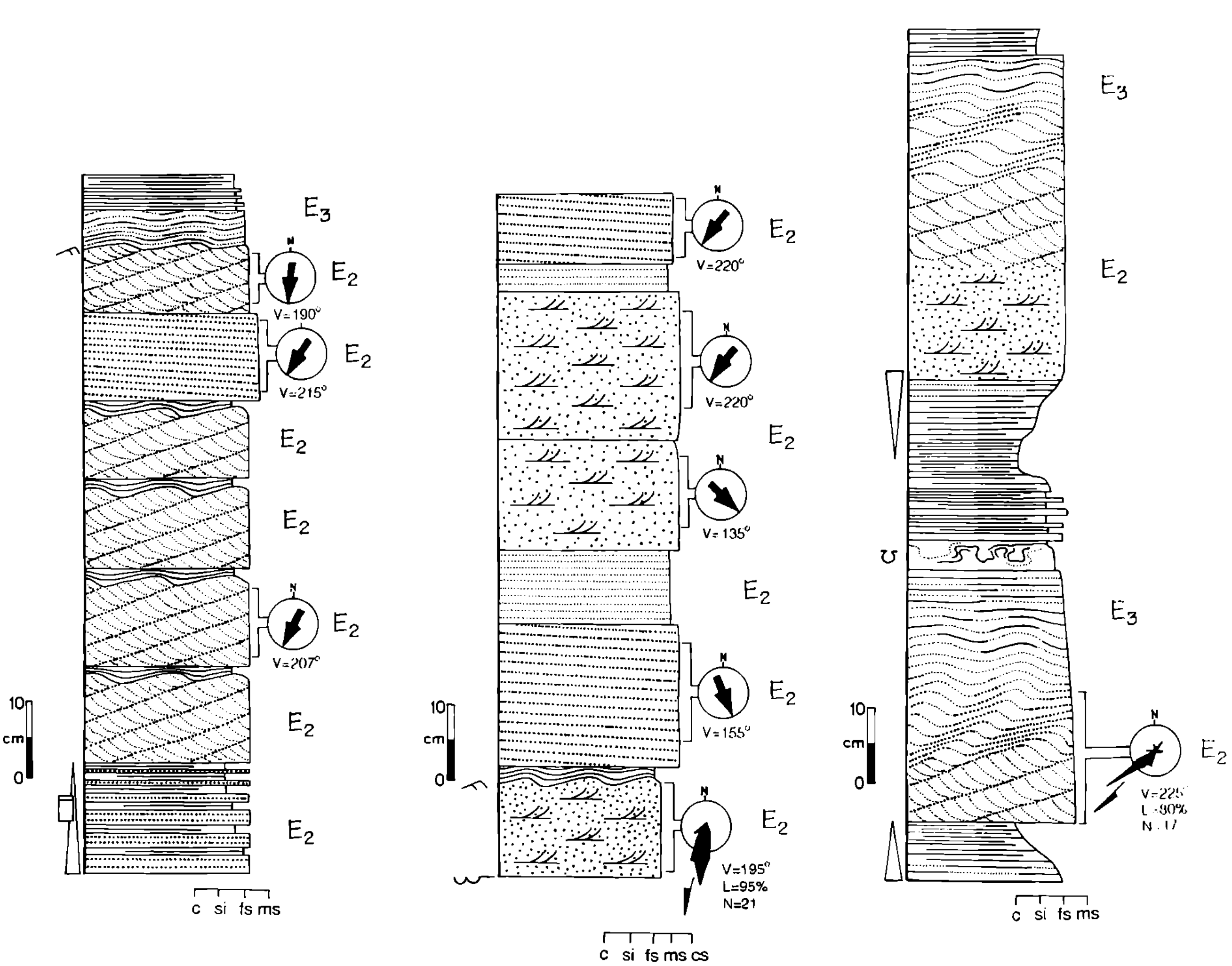

Fig. 18. Selected sedimentary logs of complex *e*. For explanation see Fig. 7

posited by strongly turbulent flows overloaded by sediment. In such conditions, even a small decrease in flow energy caused sudden, intensive accretion without syndepositional sorting (textural homogeneity, lack of stratification). Sometimes the phase of aggradation was followed by quite different, low-energy deposition.

Such sediment-overloaded stream flow was not the effect of erosional incorporation of sands by underflows, but was connected with mass flows representing the outermost zone influenced by gravity flows.

E2 – stratified sands

Description. This is the dominant facies in complex *e*; it is common in *b* and secondary in *d* (Fig. 6). This lithofacies is represented by several lithological modifications:

• Lithofacies *Sl, Sp, St* – cross-stratified sands. The most common are sands with low-angle cross-stratification *Sl* (Figs. 7, 10). These are sands of all sizes, that form beds 10 - 20 cm thick. Their thickness is proportional to grain size. Lithofacies *St* is noted as the infill of single channels up to 0.4 m deep and about 5.5 m wide, or 10 - 15 cm thick trough cross-bedded cosets of tabular shape. Lithofacies *Sp* is mostly noted as horizons of straight-crested dunes of amplitude about 10 cm.

• Lithofacies *Sh* – horizontally laminated sands. These are usually fine-grained, rarely medium-grained sands. They form thin sheets up to 5 cm and also thicker ones – up to 15 cm (Fig. 7).

• Lithofacies *Sr* – ripple cross-laminated sands. These are usually fine-grained, rarely medium-grained sands formed as thick (up to 40 cm) cosets of ripple cross-lamination or as horizons of current ripple forms (Figs. 10, 16, 17). These are mostly 1 - 3 cm high ripples, overlying *Sr* cosets, or noted as intercalations within wavy laminated silts (Fig. 7).

• Lithofacies *Src* – fine sands with climbing ripple cross-lamination of two types. Type A is noted in cosets up to 10 cm, and type B only as the middle member of gradational sequences A→B→S, which are from 30 to 60 cm (Fig. 18).

Interpretation. Facies *E2* is interpreted as sediment deposited by underflow currents, where sand material volume was balanced by flow capacity, so the flow displayed the character of an equilibrium flow. Aggradation rate was relatively low and hydraulic sorting of grains acted well during deposition. Sometimes (as for instance of *St* and *Sr* lithofacies) the flow caused local, repeated erosion pulses. Sands were transported by saltation in conditions of lower

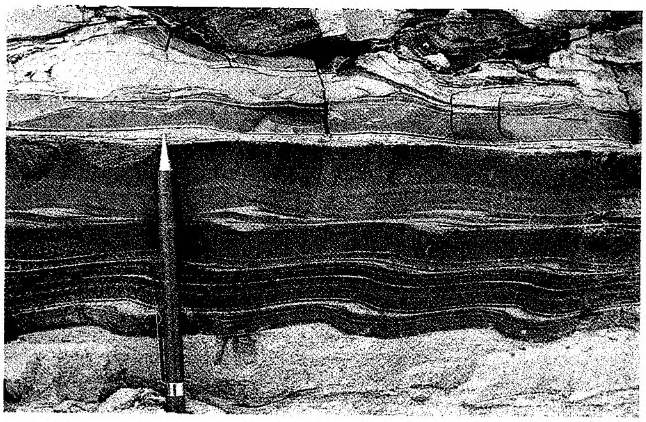

Fig. 19. Silts and fine sands with wavy lamination – facies E_3. Horizon of ripples at the base – the uppermost part of facies *D*

flow regime *(St, Sp, Sr)* or as a traction carpet in conditions of upper flow regime *(Sh)*. Analogous current-derived lithofacies have been noted within glaciolacustrine sediments (Shaw & Archer, 1978; Huff, 1989; Krzyszkowski, 1993).

E3 – laminated sands and silts

Description. This is the dominant facies in complex *d*, common in *e*, and sporadically met in *b* (Fig. 6). This lithofacies is formed of fine-grained sands, sandy silts and silty sands. It is noted in two lithological variants:

• Sands and silts of flaser or lenticular lamination *SFr*. They are noted (Fig. 16) in cosets of small scale (up to 10 cm) and medium or large scale (15 - 80 cm).

• Wavy laminated sands and silts *Sw, SFw*. They build the uppermost parts of the climbing ripple sequences A→B →S. Sometimes thet are noted as individual cosets 10 cm thick (Figs 7, 18, 19).

Interpretation. Lithofacies E_3 originated in the lowest energy conditions of all those discussed above. Silty sands and sandy silts were deposited by weak currents. Grain-size curves show the existence of a distinct silt peak. Such curves prove the simultaneous deposition from bedload (sand mode) and suspension (silt mode). Sometimes short periods of flow decay took place, and then the episodic deposition of silt drapes occurred (the origin of the wavy lamination). Lithofacies E_3 is an equivalent of distal bottom current or distal turbidity current (compare D member of the classical Bouma sequence).

Facies E_3 is genetically linked with weak subaqueous stream flow. Analogous facies were noted by Kerr (1987) , Donnelly & Harris (1989), Huff (1989) and described as quite common within glaciolacustrine sediments.

GLACIAL LAKE SUBENVIRONMENTS IN THE LIGHT OF THE GRAVITY FLOW FACIES FREQUENCY ANALYSIS

All the lithofacies mentioned and described above have been presented in a special order. It starts with the facies *A* as the deposit of the densest flow of debris flow type, and finishes with the facies E_3 – depositional effect of weak subaqueous stream flow. In that way we have obtained a sequence of physical properties of flows (Fig. 6). Generally the sequence may be divided into:

• flows of high density *(mass flows)* and

• flows of low-density *(fluidal flows)*.

Facies *A*, B_1 and B_2 are genetically connected with the first group of flows while the facies *C, D*, E_1, E_2, and E_3 with the second one. The last group can be subdivided into the flows dominated by suspension transport (*C, D* facies) and the flows with dominance of saltation transport (E_1, E_2, E_3 facies).

The analysis of sandy complexes, which considers the role of the mass and stream flows, unequivocally indicates palaeoenvironmental differences between complex *b* (with high frequency of dense gravity flows) and complexes *d* and *e* (with stream flow predomination) – Fig. 20. Complexes *d* and *e* also differ greatly from each other. In complex *e*, stream flows with predominant suspension transport are sporadically noted (origin of facies *C, D)* whereas complex *d* is completely deprived of them and dominated by low-energy bottom currents (facies E_3). The authors made an effort to answer the question: does any spatial differentiation of gravity sediment facies exist within particular complexes? In other words, whether the various gravity mass flows and currents are typical of different glacial lake subenvironments. In spite of the fact that the analysed cross-section represents a rather narrow zone of the basin, a picture of distinct spatial variability of flow types is seen from the marginal zone towards the central part of the glacial lake. In the proximal zone, in the lithofacies association *Sm,(Fh)*, mass flows (facies *A*, B_1, B_2) decidedly dominate. Closer to the centre of the lake, the flows seemed less dense. This process was controlled by energy decrease (decrease in bottom slope with distance), loss of sediment load (due to intensive aggradation) and flow dilution (dispersion) in the lake waters. Hydraulic suspension transport was dominant in this zone. For this reason in the most marginal lithofacies association *Sm,Fh,(Sh)*, stream-flow derived facies *C* and *D*,

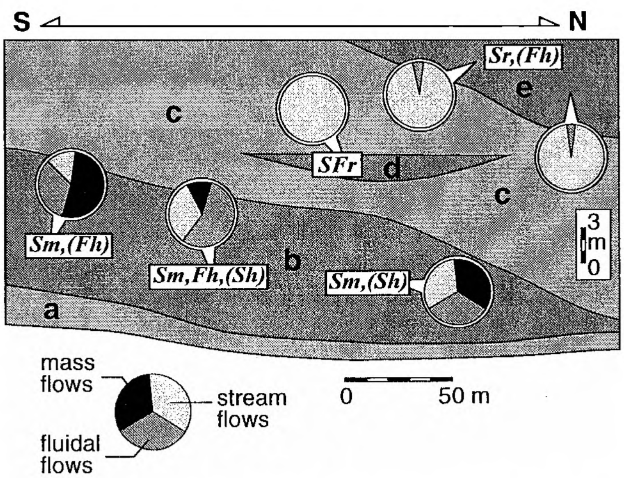

Fig. 20. Genetic differentiation of glaciolacustrine sandy complexes in the light of mass-, fluidal- and stream-flow content. Note also the lateral variability (from proximal to distal zone) within the complex *b*

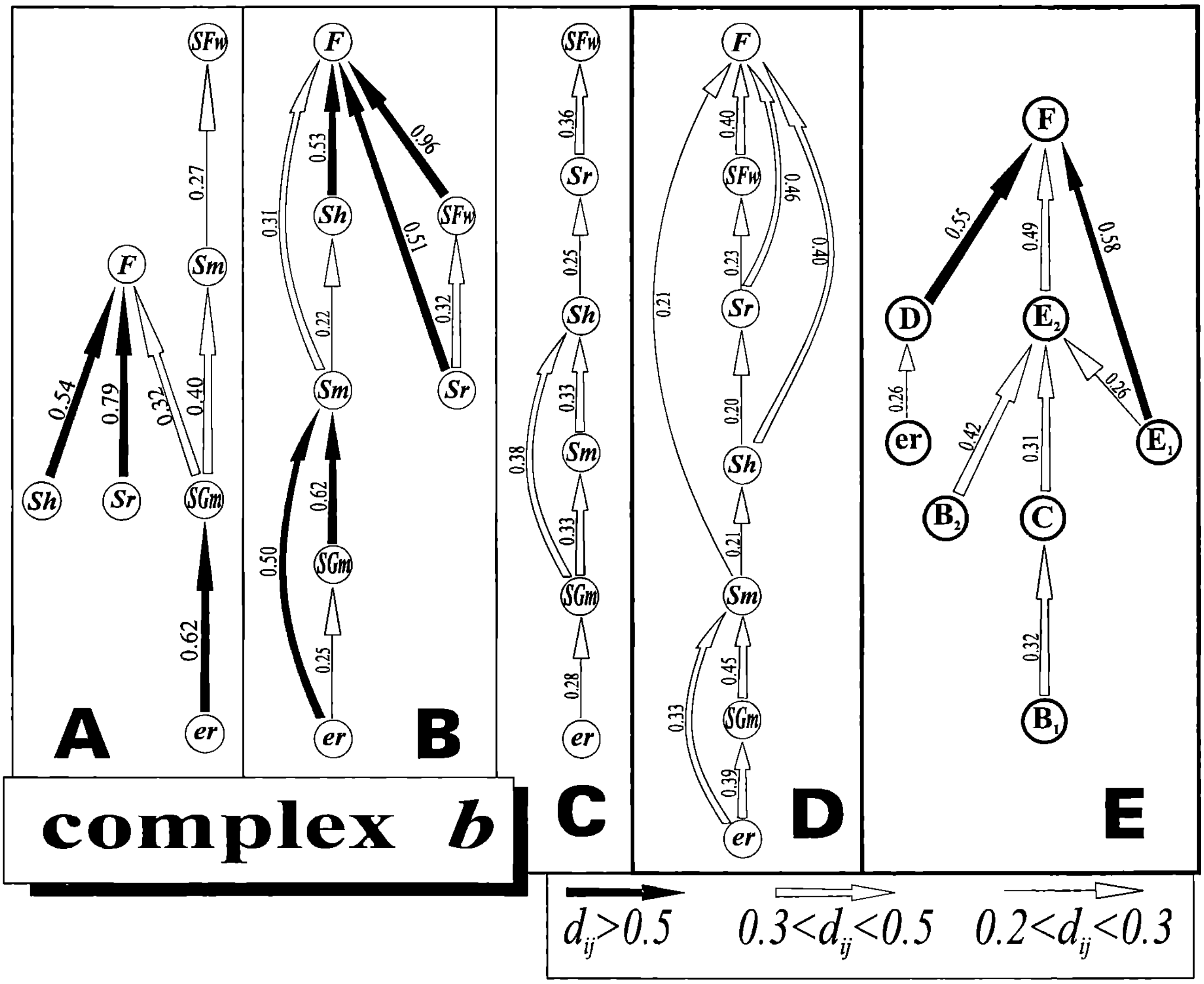

Fig. 21. Complex *b* – average lithofacies sequences obtained from Markov chain analysis: **A, B, C** – derived from three lithofacies associations; **D** – average lithofacies sequence for complex *b*; **E** – average lithofacies sequence of genetic facies for complex *b*

are most common. Near the centre, in the *Sm,Sh* association – all the three groups of facies show an almost equal contribution, which means that the low-energy bottom currents gained in frequency, in contrast to dense flow types. Such distinct spatial changeability of flow facies was not noted in complex *e*, where there was no lateral variation of the lithofacies associations.

VERTICAL SEDIMENTARY SEQUENCES ANALYSIS

A considerable, frequently sudden variability of lithofacies in the succession of glaciolacustrine sediments, gives the impression of chaotic arrangement of vertical lithofacies succession. Thus the Markov chain method was used in order to decide if there was any specified, non-random vertical lithofacies superposition. The analysis was carried out in two ways, using lithofacies sequences and genetic facies sequences. Such a procedure lets us avoid some mistakes and adds some interpretative details, which is not possible after lithofacies sequences analysis only. For instance *Sh* lithofacies was noted both in facies *D* and *E*, *Sm* lithofacies is present in facies *C, D, E,* etc.

It has been shown in the foregoing chapters that sediments genetically connected with gravity flow activity are noted in three lithofacies complexes: *b*, *d* and *e*. The quantitative contribution of lithofacies decidedly modifies the complexes. Different dynamic processes played the dominant role in sedimentation of particular complex. The authors decided to check this hypothesis by statistical analysis. Complex *d* was not analysed because of its small thickness and lack of sufficient data.

PROCEDURE AND RESULTS OF MARKOV CHAIN ANALYSIS

The following states were taken into account in lithofacies sequence analysis: *er* – erosional surface; *SGm* – sands with gravels and/or with clayey clasts, massive; *Sm* – massive sands; *Sh* – horizontally laminated sands; *Sr* – trough cross-laminated sands; *Src* – sands with climbing ripple cross-lamination; *SFr* – sands and silts with lenticular lami-

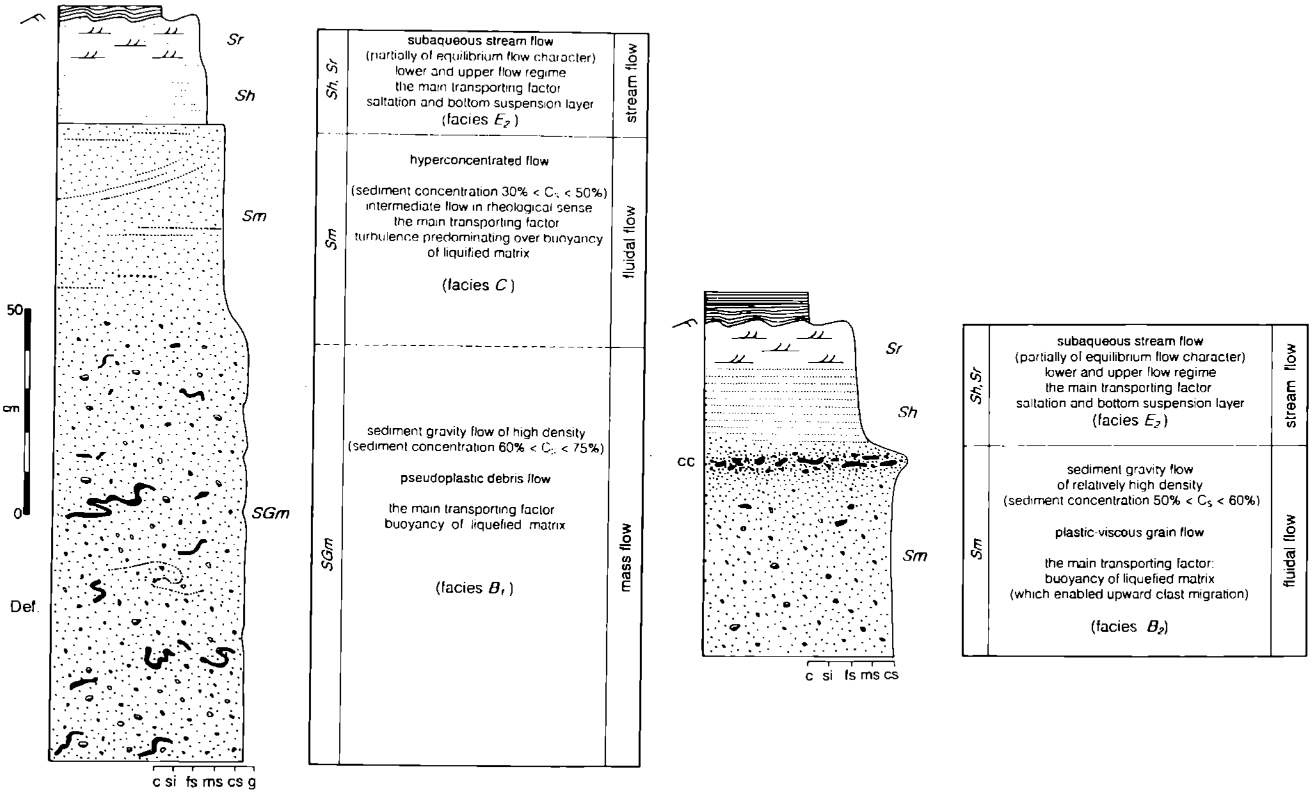

Fig. 22. Two typical sedimentary sequences of dense gravity flows obtained as the effect of Markov chain analysis of complex *b*

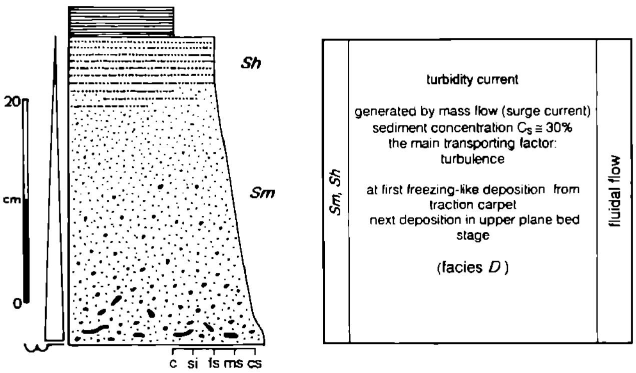

Fig. 23. Typical sedimentary sequence of turbidity current obtained as the effect of Markov chain analysis of complex *b*

nation; *SFw* – wavy laminated silty sands and silts; *F* – massive silts and clays. In instance of flow facies sequence, the following states were considered: B_1, B_2, C, D, E_1, E_2, E_3 and F – settling of suspended fines. Facies *A* was not included because of its too rare occurrence in the analysed sections.

The Markov chain analysis was done following the usual course (c.f. Nemec, 1981; Casshyap & Khan, 1982; Casshyap & Tewari, 1982; Doktor & Gradziński, 1985). At the last stage, a difference matrix d was calculated, which was taken into consideration in further conclusions. The highest values of d indicate the most probable transitions of states and are independent of state frequency. The difference matrix d was the basis for ascertaining the most typical vertical sequences. All the sequences presented are statistically significant.

Typical vertical lithofacies sequences were found for three succeeding lithofacies associations in complex *b* (Fig. 21 A, B, C). As one can see, the most logical, normally-graded sequence has been obtained for the most distal lithofacies association *Sm,(Sh)* of complex *b* (Fig. 21 C). Towards the more proximal (i.e. near-shore) part of the lake, the resultant sequence becomes more complex (Fig. 21 A, B). This fact proves the conclusion that in the zone of more

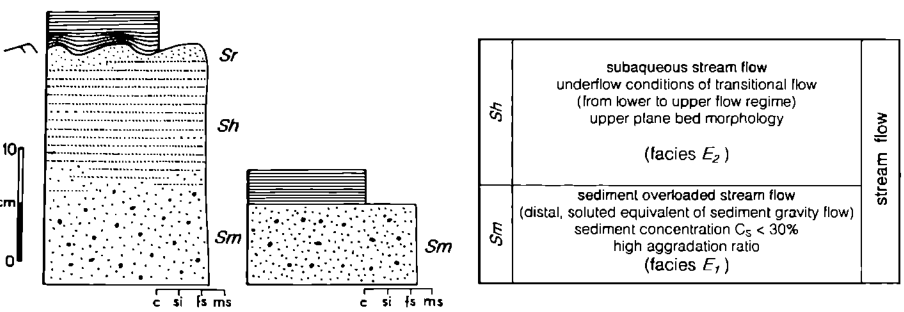

Fig. 24. Typical sedimentary sequence of density-driven hyperconcentrated flow obtained as the effect of Markov chain analysis of complex *b*

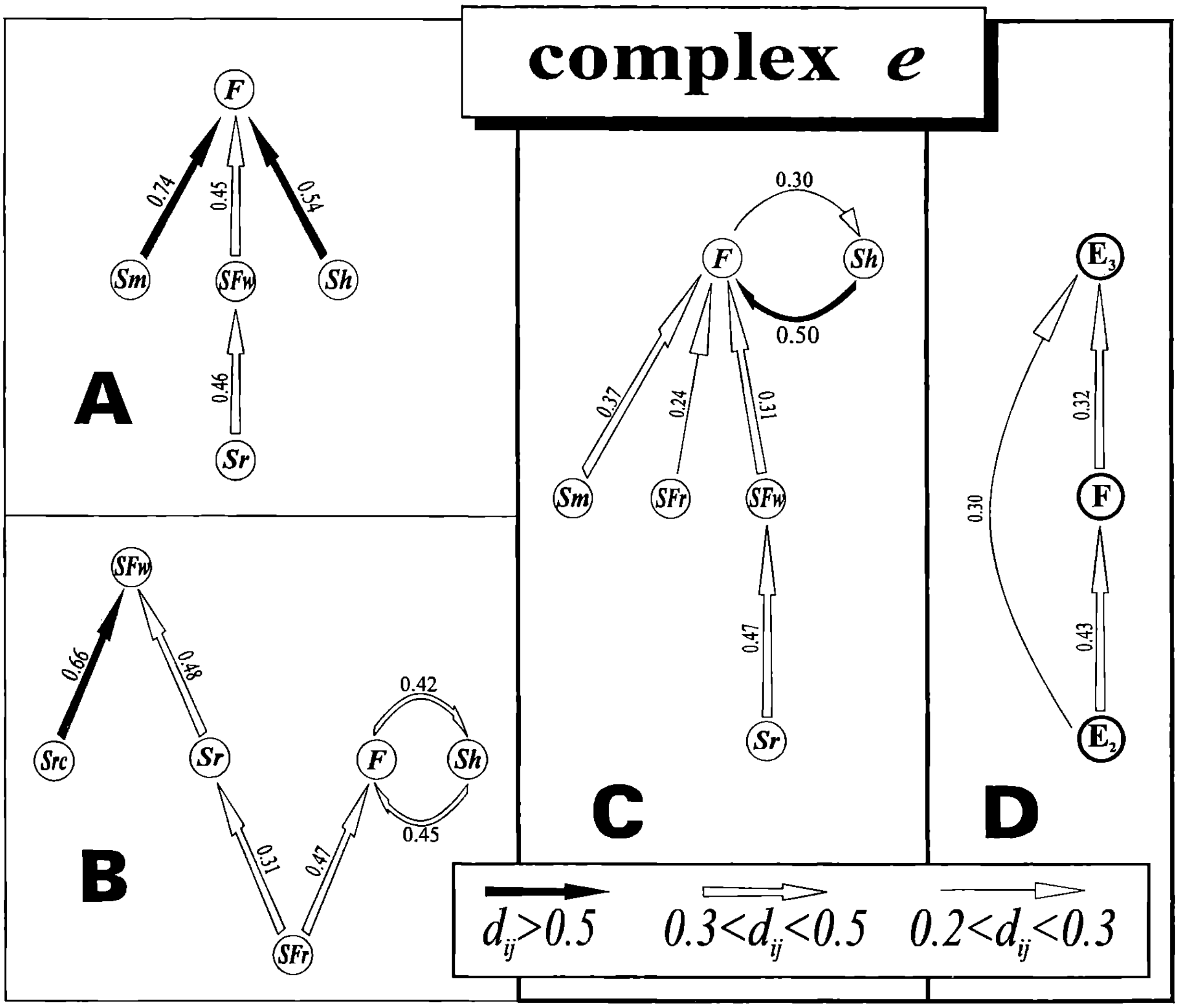

Fig. 25. Complex *e* – results of Markov chain analysis. Typical lithofacies sequences of more proximal (**A**) and distal (**B**) zones; **C** – representative lithofacies sequence of complex *e*; **D** – representative genetic facies sequence of complex *e*

proximal (i.e. more dense) gravity flows activity, the sedimentary cycles were not evident – because of the variety of depositional processes involved. The resultant lithofacies sequence for complex *b* is a normally graded cyclothem with erosional base: *er→SGm→Sm→Sh→Sr→SFw→F* (Fig. 21 D). What kind of dynamic processes did take part in the sequence origin? The genetic succession indicates coexistence of two basic flow types: (1) dense, liquefied flow which transformed with time into the flow of high sediment concentration and further into stream flow (Fig. 22), and (2) turbidity currents (Fig. 23) and sediment-overloaded bottom currents (Fig. 24) acted independently of the first type.

In complex *e* (Fig. 25) there are a few simple lithofacies sequences, which distinctly contrast in a degree of development with the sequences in complex *b*. These are direct, two-member transitions: sand to silt (or clay), or their somewhat more complex modifications (Fig. 25 A, B, C). Lack of basal erosional surfaces is characteristic of these sequences. Sometimes reverse-graded transitions occur: *F→Sh* (Fig. 25 C), which were not noted in the complex *b*. The resultant lithofacies sequence in complex *e* is as follows: *Sr→SFw→F*. It is the depositional effect of low-energy bottom currents activity (for details see Fig. 26).

A comparison of the vertical lithofacies sequences and the dynamic processes in both analysed lithofacies complexes indicate distinct palaeoenvironmental differences. During the deposition of complex *b*, true mass flows were common; with diminishing of low energy they transformed into less dense flow types.

The examples of such processes and their sediments were frequently noted in glaciolacustrine and lacustrine environments (Smith, 1978; Broster & Hicock, 1985; Eyles *et al.*, 1987; Brodzikowski & Van Loon, 1991; Eriksson, 1991; Liverman, 1991). The authors are of the opinion that seismic activity in the graben was the most probable trigger mechanism for such common gravity flow activity. Increased subsidence within the graben resulted most probably in deepening of most local lakes and in steepening of their slopes. Even in condition of varied bottom morphology it is difficult to find well-grounded explanation for numerous gravity sediment facies as found in complex *b* (Fig. 20). This interpretation is even more probable so as within glaciolacustrine

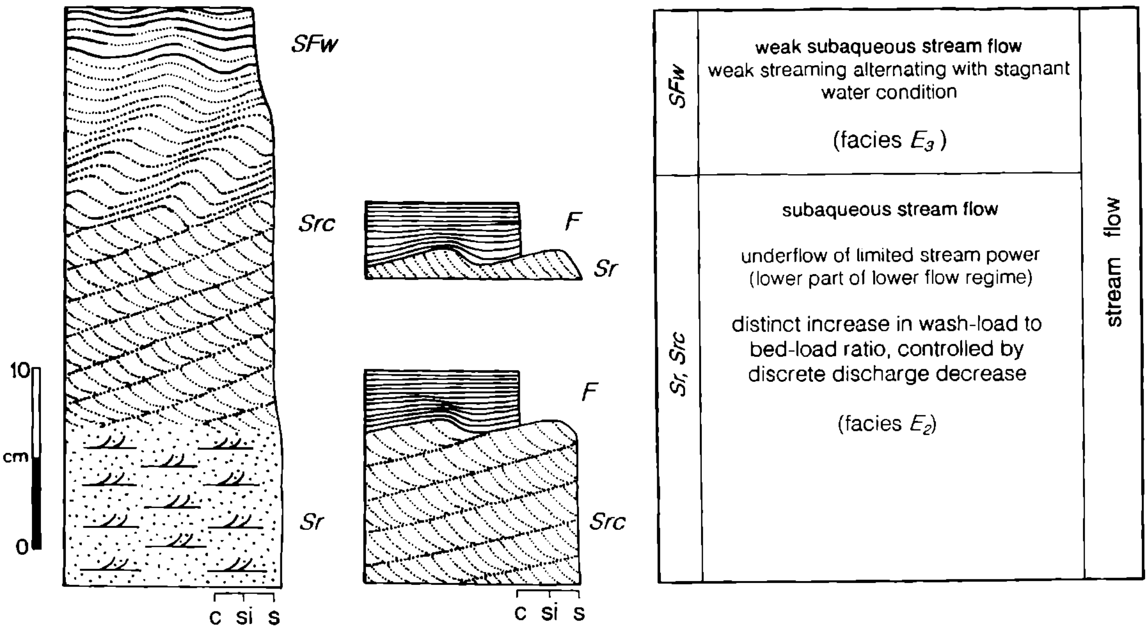

Fig. 26. Typical sedimentary sequence of weak underflow and its reduced variations (complex *e*)

sediments of the Kuców Formation the deltaic lithosomes, which could be a source of intensive gravity supply, have not been noted so far. The Kleszczów graben had been a tectonically active geological structure for the whole Tertiary and Pleistocene (Baraniecka, 1971, 1975; Pożaryski, 1977; Brodzikowski *et al.*, 1987; Brodzikowski *et al.*, 1987a; Krzyszkowski & Brodzikowski, 1987; Krzyszkowski, 1989, 1993). Gravity flow facies were described by Krzyszkowski (1993) from the Bełchatów outcrop and the flows were attributed to tectonically active faults surrounding deeper parts of the lake basin. The examples of mass-flow sediments formed in glaciolacustrine basin as the result of seismic shocks were noted from the Kleszczów Graben (Van Loon *et al.*, 1995; Brodzikowski *et al.*, in press) as well as from other glaciolacustrine and lacustrine environments (Eyles, 1987; Eyles & Clark, 1988; Doig, 1991; Van Loon *et al.*, 1995). In the analysed palaeolake, the dense gravity flows transformed into turbidity currents (so-called surge currents), and followed towards the distal, i.e. central part of the basin.

FINAL REMARKS

Coarse-grained glaciolacustrine deposits in Bełchatów are an integral part of the glacial lake deposits of the tectonically active area. The authors present evidence that all the deposits coarser than clay and silt are of subaqueous gravity flow origin.

Within the complete sedimentary succession of glacial lake the authors distinguished eight types of subaqueous gravity flow facies – from the most dense debris flow to weak subaqueous stream flow. A spatial changeability of flow types in the direction from the near-shore zone to the central part of the lake is clearly visible.

The following model can be provided. True mass flows (i.e. debris flows) originated in the marginal part of the deep lake, on a subaqueous slope. They spread downslope on relatively long distances changing their density, energy of particle movement, and the rheological character. The flows became more fluidised with distance travelled. From the near-shore zone to the central basin the coarse-grained diamicton, made up of clayey breccia, passed to massive sands with floating clay clasts, and finally to massive sands with clay clasts horizon just below the tops of beds. At this stage of sediment movement frictional and cohesional strengths were the main forces keeping the grains in suspension. The slope decreased towards the central part of the lake and caused a decrease in flow energy. Some yield strength was still operating but turbulence played much bigger role. These hyperconcentrated flows formed beds of sand with indistinct cross-stratification and finally, where turbulence became the most important transport factor – turbidity currents resulted in normally-graded sands. When flows reached the flat central area of the lake, their energy much decreased, which resulted in aggradation in nearly quiescent conditions. In the distal subenvironment bottom currents were rare, causing cross-lamination in fine-grained sands. Sometimes there were periods of flow decay and then sedimentation was from parapelagic suspension. That was the reason for lenticular, flaser and wavy lamination in fine sands and silts.

To summarise: the proximal parts of the lake were dominated by true mass flows whose concentration decreased with the distance from dense slurry to more diluted flows transitional to hydraulic medium. More distal subenvironments connected with central part of the lake were dominated by much less dense flows, i.e. bottom stream flows of various energy. In this way different gravity flow sediments can be sometimes helpful in identification of a deep lake subenvironment.

The model mentioned above was supported by Markov chain analysis. Differences between subenvironments of the lake are exhibited by the typical lithofacies sequences. The most complex vertical sequences, consisting of six to seven members, normally-graded and always with erosional bases, were characteristic of the most proximal part of the lake. In the more distal zone, simple two or three-member transitions predominate, without erosional bases, mainly normally-graded.

Described facies of gravity flows originated most probably due to slumping processes in the near-shore zone of the lake. Lack of deltaic facies or wave action structures indicate a relatively deep lake. Hence the wave activity or changes in water level could not be the origin of flows. The gravity flow sediments are distributed very irregularly in the whole glaciolacustrine succession, and concentrate only in three complexes *b, d* and *e*. Moreover the dense gravity flows (mass flows) are found only in complex *b*, where all the flow sediment types described in the paper are noted, rarely intercalated by sediments from suspension. In the authors' opinion the most probable trigger mechanism for setting the sediments in motion were earthquakes in the active Kleszczów Graben. Even weak shocks connected with the graben subsidence could result in gravity flows, on a probably quite steep subaqueous slope.

Geological facts mentioned in the paper give substance to the thesis that the analysed palaeolake was narrow and deep. In all probability such lakes were typical of the active graben areas.

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Streszczenie

Grawitacyjnie redeponowane osady glacilimniczne rowu Kleszczowa (pleistocen, centralna Polska)

Beata Gruszka & Tomasz Zieliński

Analizie sedymentologicznej poddano osady kompletnej sukcesji jeziora glacjalnego, które istniało w rowie Kleszczowa (odsłonięcie kopalni "Bełchatów") podczas zlodowacenia Sanu (Fig. 1). Wśród typowych glacilimnicznych iłów i mułów występują tam liczne pakiety piasków (Fig. 2, 4, 5), które stanowiły główny przedmiot opracowania. Zastosowano trójstopniową gradację wydzieleń litofacjalnych: od litofacji, poprzez zespoły litofacji do kompleksów litofacjalnych (Fig. 3, 4; Tab. I).

Interpretacja sposobu depozycji względnie gruboziarnistych osadów jeziornych umożliwiła wydzielenie ośmiu facji genetycznych, tj. osadów powstałych w efekcie różnych odmian spływów i prądów grawitacyjnych (Fig. 6, 11, 12).

Facja *A* – gruboziarnisty diamikton (brekcja ilasta) to bardzo charakterystyczny osad tworzący cienkie, masywne ławice o pokroju taflowym, gdzie klasty ilaste (często w formie rozerwanych i zdeformowanych lamin) są gęsto i bezładnie upakowane w matriksie piaszczystym (Fig. 7). Jest to osad gęstych spływów kohezyjnych migrujących po stromych skłonach dna jeziora.

Gruboziarniste piaski masywne z rozproszonymi klastami ilastymi (facja B_1) i piaski masywne z klastami skoncentrowanymi w stropie (B_2) to osady występujące w miąższych ławicach o erozyjnych spągach, pokroju taflowego lub stanowiące wypełnienia kanałów (Fig. 7-10, 13). Obie te facje interpretowane są jako dwie odmiany (B_1 – bardziej gęsta i B_2 – mniej gęsta) spływów uwodnionego materiału, w których czynnikami transportującymi były zarówno siły wyporu matriksu, naprężenia kolizji międzyziarnowych, jak i turbulencja.

Różnoziarniste piaski o niewyraźnych strukturach warstwowych (*C*) zawierają często domieszki żwirów lub/i obtoczonych klastów ilastych (Fig. 7). Powstały one w efekcie depozycji z wysokoenergetycznego przepływu o dużej koncentracji osadu – pośredniego między spływem masowym a prądem hydraulicznym.

Piaski o normalnym uziarnieniu frakcjonalnym (*D*) reprezentują sekwencje pionowego następstwa strukturalnego (struktura masywna → laminacja pozioma → przekątna laminacja riplemarkowa) typowego dla turbidytów (Fig. 7, 14).

Najbardziej drobnoziarniste osady piaszczyste reprezentowane są przez grupę facji *E*: E_1 – cienkie ławice piasków masywnych (efekt nagłej agradacji z intensywnych prądów dennych); E_2 – piaski o warstwowaniach przekątnych średniej skali, laminacji poziomej i przekątnej laminacji riplemarkowej; E_3 – piaski mułowe o laminacji soczewkowej i falistej (Fig. 10, 15-19). Wszystkie te facje utożsamiamy z prądami dennymi o malejącej mocy strumienia: od intensywnych prądów cechujących się wzmożoną agradacją (E_1), przez przepływy formujące megariplemarki i górne płaskie dno (E_2), przepływy kształtujące dno riplemarkowe, aż po niskoenergetyczne, okresowo zamierające prądy (E_3).

Rozpatrzono czasową i przestrzenną zmienność jakości i ilości wyróżnionych facji w obrębie osadów jeziornych. Największa frekwencja działania gęstych spływów mas przypadała na czas powstawania kompleksu *b*, natomiast piaszczyste osady młodszych kompleksów (*d, e*) powstały głównie w efekcie hydraulicznych prądów dennych (Fig. 20). Dowiedziono również, że w przekroju kopalnego jeziora występuje nieprzypadkowa oboczna zmienność poszczególnych facji. W przybrzeżnej (proksymalnej) strefie dominują wyraźnie osady spływów mas (facje *A*, B_1, B_2). W kierunku dystalnym (tj. ku centrum zbiornika) pojawiają się kolejno facje *C* i *D* (osady przepływów zawiesinowych), a następnie facje grupy *E* (związane z coraz słabszymi prądami dennymi). Takie oboczne następstwo facji dowodzi, że w strefach stromych skłonów dna powstawały osuwiska, z których materiał spełzał jako gęste spływy masowe. Z czasem tej podwodnej redepozycji osad ulegał stopniowemu rozproszeniu w wodach zbiornikowych, tracił na gęstości i przeradzał się w szybki spływ uwodnionego materiału. W efekcie postępującej depozycji koncentracja osadu w spływie malała, aż dochodziło do przejścia w prąd zawiesinowy. W strefach bardziej połogich skłonów i przejścia do płaskiego dna otwartego zbiornika prądy turbiditowe wytracały swą prędkość i obciążenie osadem i ewoluowały do prądów dennych deponujących osad z transportu saltacyjnego. W najbardziej zewnętrznych strefach zamierały one i tam trakcyjna depozycja piaszczysta przechodziła w parapelagiczną depozycję aleurytowych zawiesin z wód stojących.

W celu określenia szczegółowego następstwa procesów depozycyjnych w trakcie powstawania miąższych pakietów piaszczystych dokonano analizy tych osadów metodą łańcuchów Markowa. Najbardziej kompletną sekwencję wychwycono w kompleksie *b*, w którym istnieje największe bogactwo facji spływowych (Fig. 21). Wypadkowa sekwencja reprezentowana jest pakietem o normalnym uziarnieniu frakcjonalnym. Rozpoczyna się ona facją gęstego spływu, która ku górze przechodzi w odmianę o mniejszej gęstości, a wyżej nadbudowana jest osadem

prądu trakcyjnego (Fig. 22). Z tego kompleksu uzyskano też inne, krótsze sekwencje związane z działaniem prądów zawiesinowych (Fig. 23, 24). Natomiast w piaszczystym kompleksie *e*, znaczącym etap finalnego zapelniania jeziora, stwierdzono obecność prostszych sekwencji (niekiedy wręcz rytmów osadowych) utworzonych przez drobnoziarniste facje słabych prądów dennych (Fig. 25, 26).

Autorzy sądzą, że wyjątkowe bogactwo osadów spływów i prądów grawitacyjnych w rozpatrywanej serii plejstoceńskiego wypełnienia rowu Kleszczowa ma związek z tektoniczną aktywnością tego obszaru. Częste powstawanie osuwisk i gęstych spływów mas wynikało najprawdopodobniej z licznych wstrząsów związanych z subsydencją podłoża rowu. Jezioro będące środowiskiem sedymentacyjnym analizowanych osadów było z pewnością głębokie, a urozmaicona morfologia dna zbiornika ułatwiała rozwój szerokiej gamy procesów redepozycji grawitacyjnej.