S T U D I A G E O M O R P H O L O G I C A C A R P A T H O - B A L C A N I C A
VOL. XXXV KRAKÓW 2001 PL ISSN 0081-6434

TOMASZ SALAMON (SOSNOWIEC)

SEDIMENTARY SUCCESSION AND MORPHOLOGIC CONTROL OF GLACIOMARGINAL PALEOGEOGRAPHY IN THE FOOTHILL AREA — AN EXAMPLE FROM THE SILESIAN BESKID (POLAND)

Abstract. In the pit localized in the Łaziński Potok valley several Pleistocene sedimentary series form the succession. It starts from fluvial gravels and consists of glacigenic deposits representing a few stages of the ice-sheet advance onto the foothill area. The following series have been distinguished in stratigraphic order: glaciolacustrine series formed due to valley damming; the lower till and subglacial canal series derived from the first ice-sheet advance; ablation series of a short-term recession stage; glaciofluvial series, and the upper till of the second ice-sheet advance.

The shape of ice-sheet margin as well as glaciomarginal depositional processes were strongly influenced by morphology. A distinct glacier lobe was formed in the Vistula valley. The ice from the main valley penetrated into the secondary Łaziński Potok valley in the form of a smaller lobe. When the ice thickness increased sufficiently, the glacier advanced over the transversally oriented Rudzica Ridge. Morphology was also a dominant factor controlling glacial phenomena during the short-term retreat. In places, where glacier lobe, was still supplied with ice, the frontal deglaciation took place. On the other hand, in the zone where ice supply was limited or decayed due to morphological barrier, the ice sheet was disintegrated into dead-ice blocks.

Key words: glacimarginal zone, sedimentology, foothill area, Beskids Mts

INTRODUCTION

The Scandinavian ice sheet penetrated the Middle-European mountain ranges only during the older glaciations; the Sudetes during the Sanian II (i.e. the Elsterian) and the Odranian (i.e. the Saalian), whereas the Carpathians in the Sanian II cold period only. The mountains and high uplands created quite different glaciomarginal conditions from the ones commonly known for the lowland areas. These differences were mainly determined by a distinct morphology of bedrock, which controlled the zone character to a certain degree, depending on the local relief. On the other hand, the fact that the ice sheet advanced on a surface generally inclined in opposite direction to its movement was also very important, and have to be thoroughly considered. At last, it must

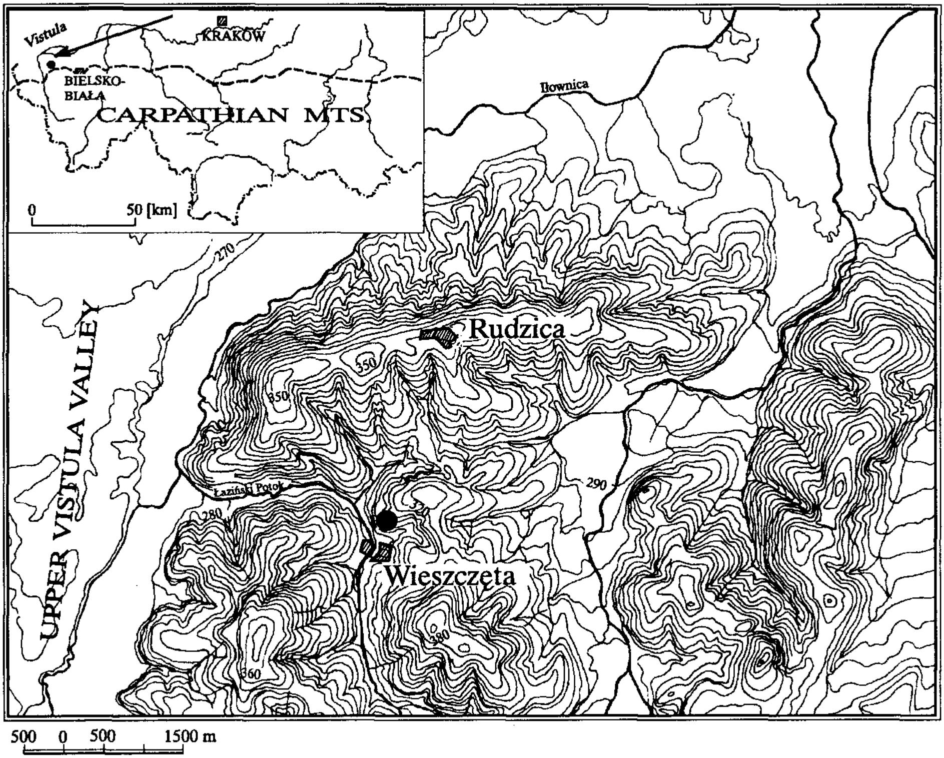


Fig. 1. The map of NW part of Bielsko Upland area and outcrop location (black spot). Maximum ice-sheet extent marked by dashed line

be kept in mind that a distance from the ice source area was distinctly longer in this case, compare to the Polish lowlands. All these factors played the important role on the ice-sheet dynamics, and the glaciomarginal zone nature in the Fore Carpathian zone (Fig. 1).

The studies of glacigenic deposits of the mountain and upland areas were already carried out in the Sudetes (Szczepankiewicz 1953; Walczak 1957, 1969, 1972; Jahn 1960, 1963, 1969; Jahn and Szczepankiewicz 1967; Szponar 1974) as well as in the Carpathians (Pawłowski 1932; Książkiewicz 1935; Jahn 1952; Klimaszewski 1936, 1948; Konior 1939; Starkel 1957; Stupnicka 1962; Łanczont and Racinowski 1994; Burtanówna et al. 1937) a few decades ago. However these researches focused on problems of the maximum ice-sheets extent and general description of sedimentary successions. No detailed sedimentological analyses have been conducted, and here is the main reason of this paper.

Thus the aim of this study is to characterise the glacigenic depositional environments in the relatively small Łaziński Potok valley, situated in the foreground of the Silesian Beskid. The valley particular location, within the foothills area, gave an opportunity to study the local conditions of the glaciomarginal zone, existing in front of the mountains.

GEOMORPHOLOGICAL AND GEOLOGICAL SETTING

This study has been carried out in the Bielsko Upland, in the Wieszczęta site, 15 km NW from Bielsko-Biała. An area is relatively narrow (6–8 km), northwards inclined plateau, in front of the Silesian Beskid. It is bordered with 500–800 m high denudation slope of the Beskids from the south, and with the Oświęcim Basin from the north. The plateau is dissected by river valleys to several meridionally oriented hill ranges. The so-called Rudzica Ridge, running from west to east, is the only morphological structure of different axis. It exists as distinct morphological element, 80 m high, above the Oświęcim Basin. Its highest summits are approx. 350 m a.s.l. The studied site is located on the southern slope of the Rudzica Ridge, on the right side of the Łaziński Potok valley. In this place the valley changes its course from N–S to W–E, and 2 km further it merges with the main Vistula valley.

The Bielsko Upland is built up of flysch series of Cieszyn Nappe and Sub-Silesian Nappe. Generally, these are unresistant shales with limestone and sandstone intercalations. The Rudzica Ridge is composed of shales and sandstones of the Upper Cieszyn Beds, sandstones and marls of the Sub-Silesian Nappe, and clays and sandstones of the Skawina Beds (Ryłko and Paul 1994).

The analyzed site is located several km in NE direction from the Ustroń Basin, the area previously investigated by M. Książkiewicz (1935) and E. Stupnicka (1962). The stratigraphic conclusions of their researchers run contrary to each other. The gravels and sands containing both Carpathian and Scandinavian material (so-called „mixed gravel") were attributed by Książkiewicz to the San Glaciation (Elsterian), while Stupnicka interpreted them as the sediments redeposited during the Odra (Saale) and Warthe Glaciations. In my opinion the idea by Stupnicka, of twofold sediment infilling of the Ustroń Basin up to 50–60 m, is of low probability. I interpret the Łaziński Potok sedimentary succession as glaciomarginal deposit of the San II Glaciation (Elsterian). At that time the ice-sheet in its maximum extent reached the Ustroń Basin and the thick covers of extraglacial and proglacial sediments were deposited there (Książkiewicz 1935).

METHODS

Field compaign was provided during the 1998 and 1999. Twelve detailed sedimentary logs were made to document the walls of a large pit. Lithofacies code of A. D. Miall (1977) with T. Zieliński's (1995) modification was used

Table 1

Lithofacies code symbols

GRAIN-SIZE CODE SYMBOLS	
G	gravel
GS	sandy gravel
SG	gravely sand
S	sand
SF	sand and silty sand
FS	sandy silt
F	silt, (silty fines)
D	diamicton
SEDIMENTARY STRUCTURE CODE SYMBOLS	
m	massive structure
h	horizontal lamination or stratification
r	ripple cross-lamination
l	low-angle cross-stratification
p	planar cross-stratification
t	trough cross-stratification
s	stratification (in till)

in description of the analysed succession (Table 1). Lithofacies, lithofacies association, and series were the three-rank division used during the sedimentary analysis of the logs. Additionally the till fabric and cross-strata orientation were measured, and the petrographic analyses of gravels (of 4–8 mm grain-size range) were prepared.

SEDIMENTARY SUCCESSION AND DEPOSITIONAL ENVIRONMENTS

The studied pit lies ca 600 m northwards from the centre of Wieszczęta village. Now the excavation is abandoned and most of its parts are just being reclaimed. The following series have been distinguished in the stratigraphic succession (Fig. 2).

FLUVIAL GRAVEL SERIES

The lowermost series consists of the Carpathian fluvial gravels (Fig. 3). The top of this package is 298–300 m a.s.l., i.e. approximate 15 m above the present-day valley floor. Deposit is very poorly sorted. Lithofacies of massive,

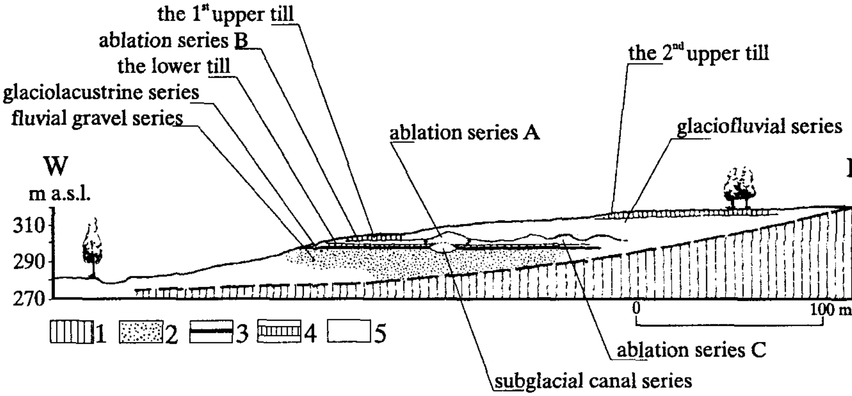


Fig. 2. Geological cross-section of the right part of Łaziński Potok Valley. 1 — flysch bedrock, 2 — gravel, 3 — silt, 4 — till, 5 — sand

clast-supported gravels with silty-sandy matrix predominate. The grain size of most abundant clasts is 5–12 cm (the maximum ones are up to 30 cm). Rare, poorly developed lags exist mainly within finer gravels. Some clayey-silty intercalations, up to 40 cm thick, are also present.

Gravels were deposited in a braided river characterized by frequent flood stages. Massive structure and very poor sorting are attributed to spontaneous deposition from highly concentrated flows during the rapid decays of flood stages (Smith 1970; Rust 1972; Miall 1996). Large velocity gradients were also typical of waning flood phenomena. Deposited material did not obtain an imbricated arrangement, i.e. the settlement typical of deposition from steady flow with repeated redeposition. The same is evidenced by a lack of well-developed lags, which are usually formed in the rivers of slowly receding floods (Klimek 1972). Silty-clayey intercalations were settled from suspension, presumably taking place in the abandoned channels, formed in the inter-riffle zones (Klimek 1972).

The depositional style mentioned above — a synergy of high river energy and sedimentary rate — suggests the lack of a dense forest cover in the drainage basin. These fluvial depositional processes are typical of tundra zone in conditions of severe climate during the ice-sheet advance (Jersak 1991; Woo 1988; Mol 1997). Additionally, the steep valley slope in the foreland of Beskids conditioned highly intensive accumulation.

GLACIOLACUSTRINE SERIES

Relatively thin series of fine-grained deposits overlies the fluvial gravels (Fig. 3). Three lithofacies associations have been divided there. The lowermost association — SFm, FSm — contains massive, fining-upward sand, which grade laterally into silty-sand lithofacies. The second association Fh,(FSh, FSr) is

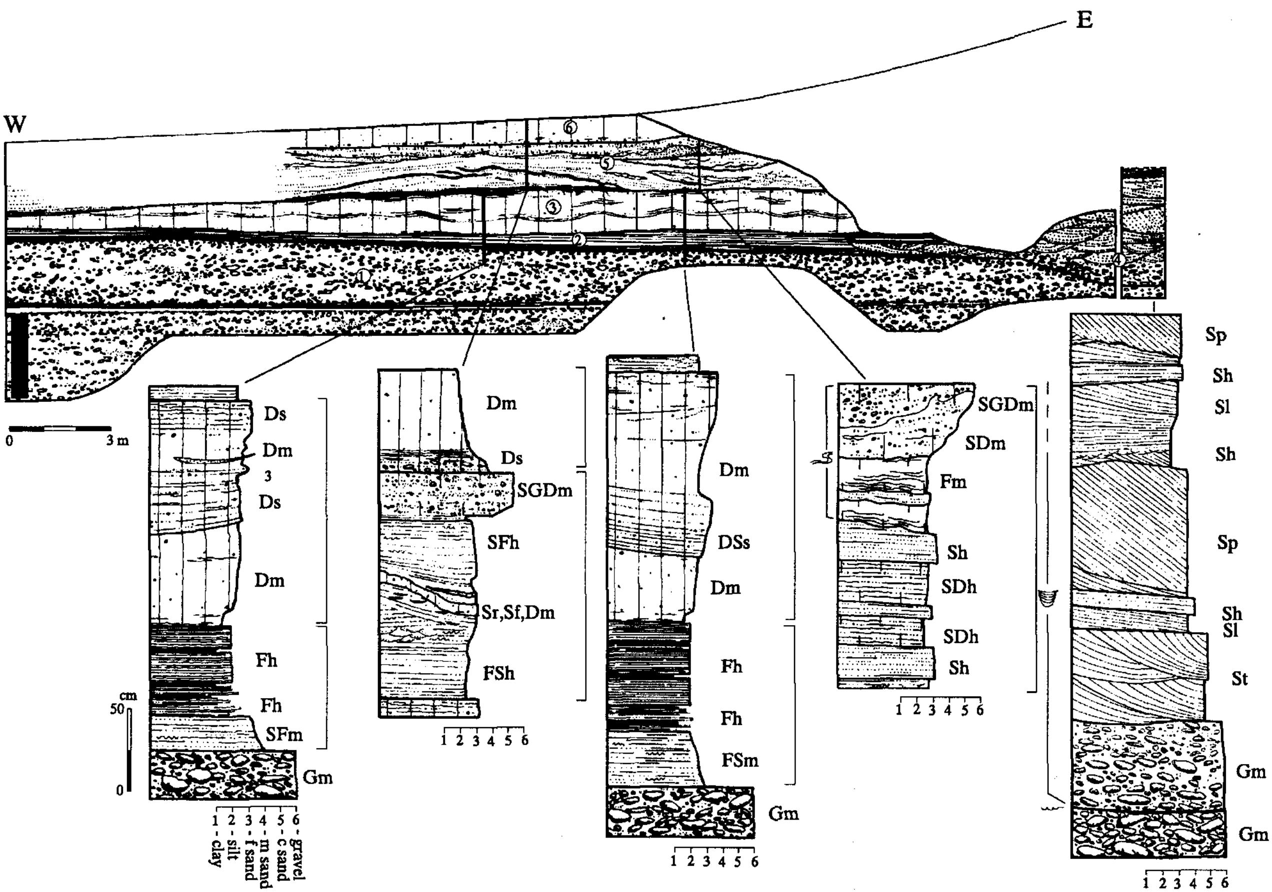


Fig. 3. Pit wall and detailed sedimentary logs. 1 — fluvial gravel series, 2 — glaciolacustrine series, 3 — the lower till, 4 — subglacial canal series, 5 — ablation series B, 6 — the 1st upper till

formed by several alternating silty-clay and silty-sand lithofacies, from among a few are cross-laminated. The upper association Fh is of rhythmic character. Two or three rhythms of clay (1.5–2 cm thick) and silt (12–14 cm) have been noted there.

The described series is an effect of deposition in the glacier-dammed lake. Three lithofacies associations where formed in succeeding lacustrine zones that rapidly migrated in upstream valley direction. The first association Sfm, FSm was deposited in a lake marginal zone, close to the mouth of the Łaziński Potok stream. The second association Fh,(FSh, FSr) is attributed to some more distal environment. Lithofacies Fh was settled down from suspension under steady water conditions, however common structures of ripple cross-lamination indicate frequent underflows distributing the sediments. The third association Fh can be explained by a further decrease of energy. Clay and silt settled down from stagnant water in a distal lake zone, far from sediment sources. Glacial nature of the lake is concluded from rhythmic, quasi-varved deposit. In this part of reservoir the sediments are to be supplied by ablation waters. Formation

Table 2

Fabric of tills

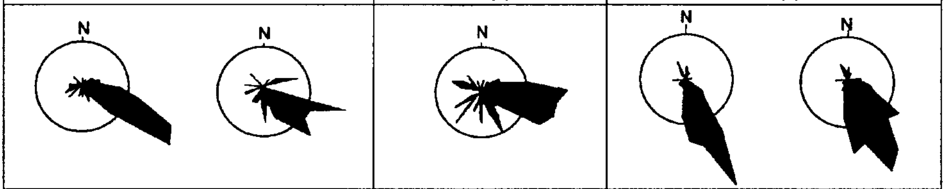
the lower till		the 1st upper till	the 2nd upper till	
				
V=125° N=120	V=124° N=65	V=120° N=85	V=160° N=80	V=158° N=135

Table 3

Petrographic composition of fine-grained gravel of some series

	Scandinavian granite and metamorphic rocks [%]	quartz [%]	flysch rocks [%]	lydite [%]	flint [%]	other [%]	number of clasts
the 2nd upper till	4.1	43.2	45.4	1.2	0.8	5.3	815
the 1st upper till	5.3	53.4	31.5	2.4	0.5	6.9	674
subglacial canal series	4.9	40.7	49.2	0.9	0.2	4.1	531
the lower till	10.6	58.1	22.2	4.5	1.6	3.0	466

of the association Fh was presumably controlled by seasonal temperature changes, which is resulted in evidently cyclic deposition (see Ashley 1989; Brodzikowski 1992). Comparable lithofacies have been cited from quite few Carpathians sites up to now (Klimaszewski 1936; Gerlach and Koszarski 1981). Glaciolacustrine rhythmites exist more abundantly in the Sudetes (Szczepankiewicz 1953; Walczak 1954; Jahn 1960).

THE LOWER TILL

0.4–1.6 m thick diamicton package lies directly on the glaciolacustrine series (Fig. 3). Two lithofacies associations have been distinguished there. The lower one is built of diamicton with massive structure and rusty brown colour passing upwards into grey shade. Its thickness varies from 20 to 60 cm. The diamicton is dominated by silt and sand with gravel admixture (ca 2–3%). A non-continuous thin layer of sediment characterized by considerably higher content of clay is observed in the base. There are no deformation structures on the contact between the diamicton and the underlying glaciolacustrine series in places, where this clay layer is present. In other places, the deformations are restricted to the zones where diamicton is underlain by silt. The scale of deformations is relatively small. The longest axes of clasts show distinct concentration in E/SE sector (Table 2). Petrographic composition of gravels is presented in Table 3.

The upper association, 60–120 cm thick, is much more structurally and texturally differentiated. Its basal surface is distinctly undulating. The diamicton beds are of different texture and reveal two kinds of structures: they are either massive or clearly laminated. Moreover some thin sandy layers have been observed there.

The lower association is interpreted as a basal till, deposited subglacially by the ice advancing into a ponded valley. The ice movement resulted mainly from sliding on a soft sediment substratum (see Boulton and Jones 1979; Boulton and Hindmarsh 1987). The upper association was recognized as a flow till. Abundant sand lenses with fine lamination indicate a meltwater agent in deposition (see Kasprzak and Kozarski 1984; Morawski 1984; Lawson 1989; Dreimanis 1989, 1990).

The polygenetic till represents the glacial environment and documents not only the ice-sheet advance into the valley area but, what is more, it reflects the phase of the ice recession. The frontal ice retreat was accompanied by the flows of moranic debris. The material was supplied onto the ice-sheet surface along thrust planes being affected by compressing shear strains in the marginal ice zone (Nye 1952; Paterson 1981; Jania 1993). Supraglacial material was gravitationally redeposited to a close foreground of the ice margin, immediately onto the previously deposited basal till.

SUBGLACIAL CANAL SERIES

The till passes laterally into a large, trough-shaped, erosional form — i.e. the canal — which is deeply incised in the lacustrine silt and the fluvial gravel as well

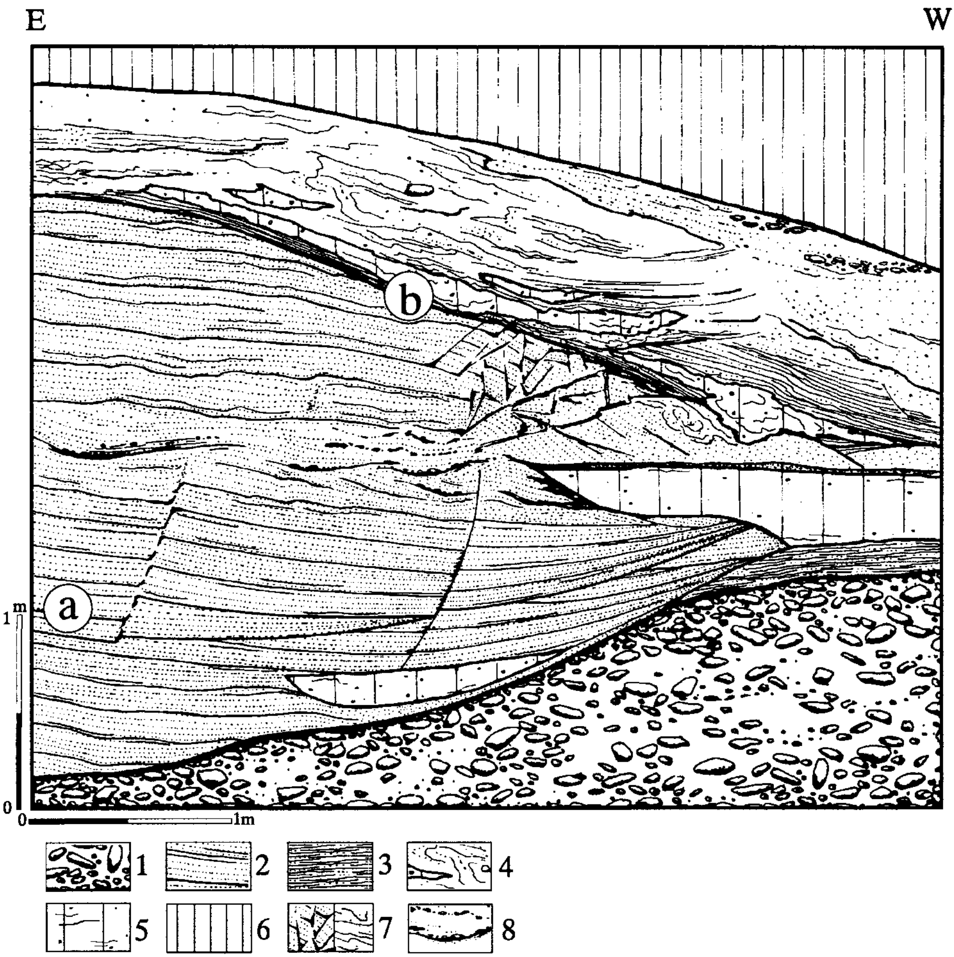


Fig. 4. Detailed sketch of subglacial canal series (a) passing upwards to ablation series of ice-crevasse (b). 1 — gravel, 2 — sand, 3 — silt, 4 — deformed heterogeneous deposit, 5 — till, 6 — slope sediments, 7 — fault and fold deformations, 8 — lag

(Figs 3, 4). In the northern wall of the pit, the massive gravel Gm overlies a broad erosional surface. The gravel passes upward into sand lithofacies: St followed by Sl, Sh,(Sp). The second fragment of this canal is observed in the opposite wall of the excavation. Medium-grained sand lithofacies Sp and Sl are the main lithofacies that fill-in the secondary troughs, and subordinately the lithofacies Sh is noted (Fig. 4). Scandinavian rocks (of pebble and granule sizes) have been found within the canal-fill deposits (Table 3). Directional data, obtained from cross-strata, indicate southward direction of palaeoflow.

The canal structure represents the high-energy subenvironment of confined flow. The canal was ca 10–12 m wide, at least 2 m deep, and at least 70 m long. The formation of such large erosional form was possible only in conditions

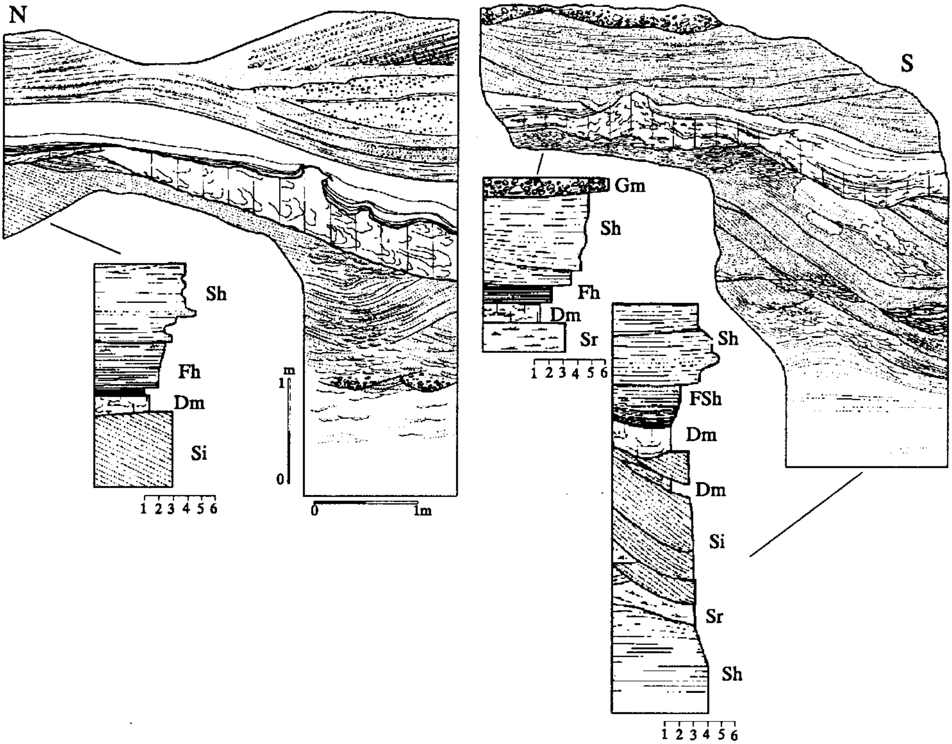


Fig. 5. Pit wall and sedimentary logs of crevasse series

of very intensive current. During the initial phase of waning stage, the incision was filled-in with gravel Gm. During succeeding decrease of flow energy, the sand lithofacies St, Sl, Sh were deposited. The canal structure orientation suggests that erosion was affected by high-pressure meltwater flow in subglacial zone. The main axes of the canal and valley are parallel, nevertheless the palaeoflow acted in upward direction, i.e. opposite to the valley slope. This phenomenon of water flowing reversely to bedrock topography can be only explained by highly increased pressure conditions in the glacier system (see Kor et al. 1991; Jania 1993). Relatively small thickness of the canal infill and quick replacement of massive gravel by the sand, indicate a rapid decrease of flow intensity, which took place after the incision episode. It can be concluded that the subglacial erosion was resulted from a sudden outburst of inglacial meltwater. In consequence of succeeding drainage of hydrological glacier system, the subglacial flow was considerably reduced, and than the canal was quickly filled-in by sediments.

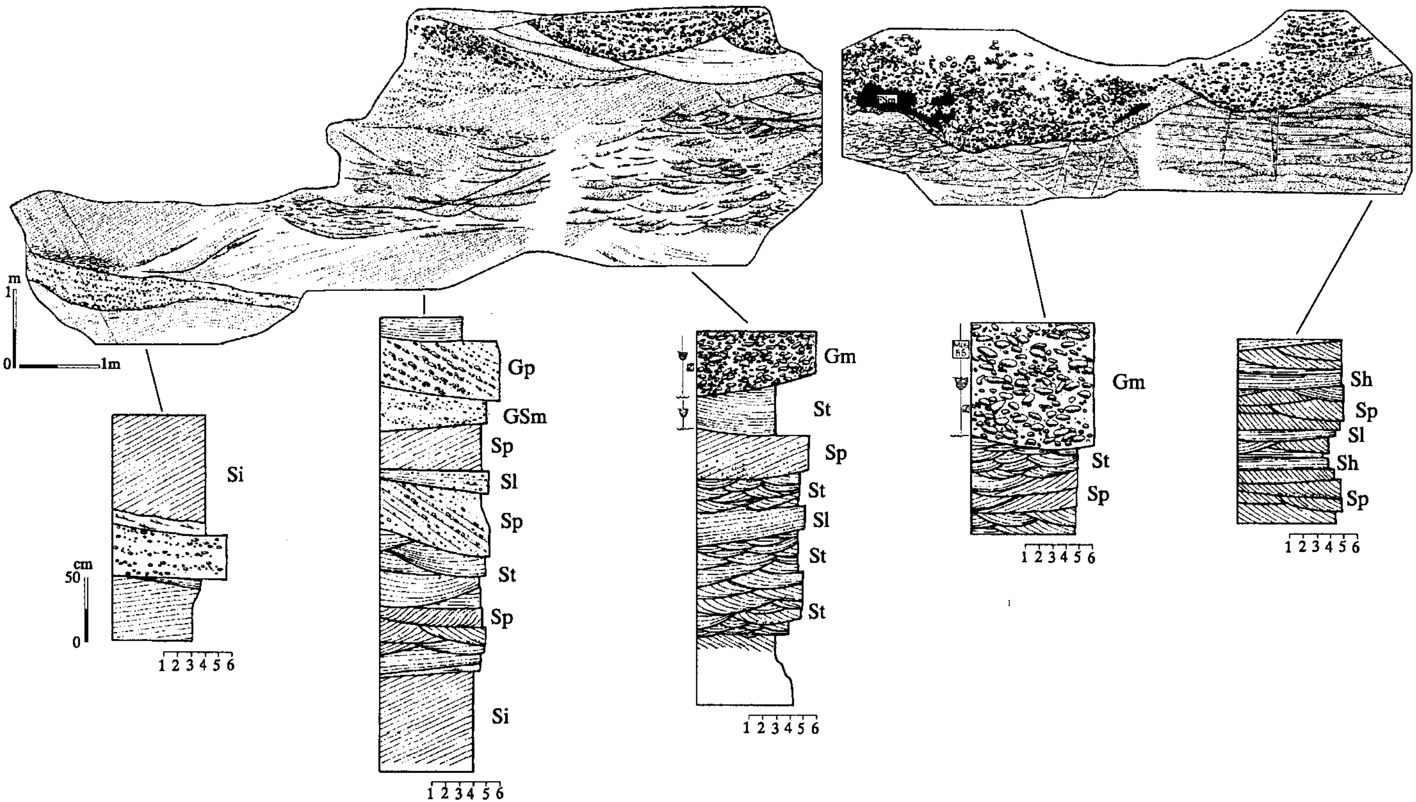


Fig. 6. Pit wall and sedimentary logs of lower part of subglacial series

ABLATION SERIES

Because of a considerable spatial differentiation of lithology, the deposits have been subdivided into three, laterally coexisting series.

ABLATION SERIES A (ICE-CREVASSE)

Ice-crevasse series has been recognized above the deposits of the subglacial canal. Till layer, existing laterally to ice-crevasse series, does not continue above it, but is erosionally pinched out (Fig. 4). It passes into a lag-like horizon of rare gravels and till clasts. Fine-grained sand lithofacies overlie the ice-crevasse series. In the lower part lithofacies Sh predominate, whereas in the upper one the lithofacies Si and Sr are common (Fig. 5). Fine-grained diamicton beds Dm with the structures of flow folds overlie the sands. Diamictons pass upwards into more fine-grained deposits: SFh, FSh, Fh. Primary horizontal lamination of silts is in inclined position, concordant with dipping diamicton. The thickness of this association is changeable, in places distinctly reduced due to erosion. The overlying sand is the result of the following depositional stage (see the fluvioglacial series). Abundant fault deformations of various scales are present in the zone above the marginal part of the ice crevasse.

This series was accumulated in the ice-crevasse evolved from the subglacial canal. This hypothesis is confirmed by the lack of till cover above the ice-crevasse deposits. Presumably the till underwent fluvial reworking, and some traces of erosional lag are evidence of this process. Low-energy depositional processes prevailed in the ice-crevasse environment. Abundant large-scale lithofacies Si (Fig. 5) represent the subenvironment of microdelta front. Some sand was additionally supplied to the microdelta crest by low-energy streams, and ripple-derived lithofacies Sr were formed there. Moranic debris flowed down from the ice-sheet surface to the crevasse, and was deposited there as the fine-grained diamicton beds Dm. Preserved depositional record above the described diamictons represents the phase of current activity cessation, and transitional conditions to suspension settling from quiet water. In this way much finer sediments were deposited in SFh, FSh, and Fh lithofacies (Fig. 6). Gravitational deformations point to progressive retreat of the ice walls and the loss of deposit support.

ABLATION SERIES B

The association predominated by silty-sand SFh, sandy-silt FSh and fine sand lithofacies Sm exists in the zone westward to ice-crevasse series, above the lower till. The aforesaid lithofacies pass laterally into the ripple-derived lithofacies SFr, Sr, and diamictons or diamictic sands in lens-shaped intercalations. The lithofacies of varigrained, massive and deformed diamictons form the upper part of the series.

Lithofacies SFh, FSh, Sr, Sm represent low-energy subenvironment of periodically waning ablation currents in a shallow pond. The uppermost beds of varigrained diamictons are interpreted as the record of superimposed mass flow deposits. Their better sorting is attributed to longer redeposition distance in comparison with the diamicton beds found on the top of the basal till; they were transported from more distant glacier areas (cf. Lawson 1989). Generally, this series is interpreted as the deposits of low-energy glaciolacustrine environment located close to the ice margin during its frontal retreat.

ABLATION SERIES C (OF DEAD ICE-BLOCKS DECAY)

Association of diamictons and silty sands, up to 1.5 m thick, exists eastwards to the ice-crevasse series (Fig. 2), and overlies the sands containing diamicton intercalations. This series was distinguished due to presence of deformations of medium or large scale. Primary horizontal beds commonly rest in steeply inclined position. These inclined beds are dissected by numerous normal faults

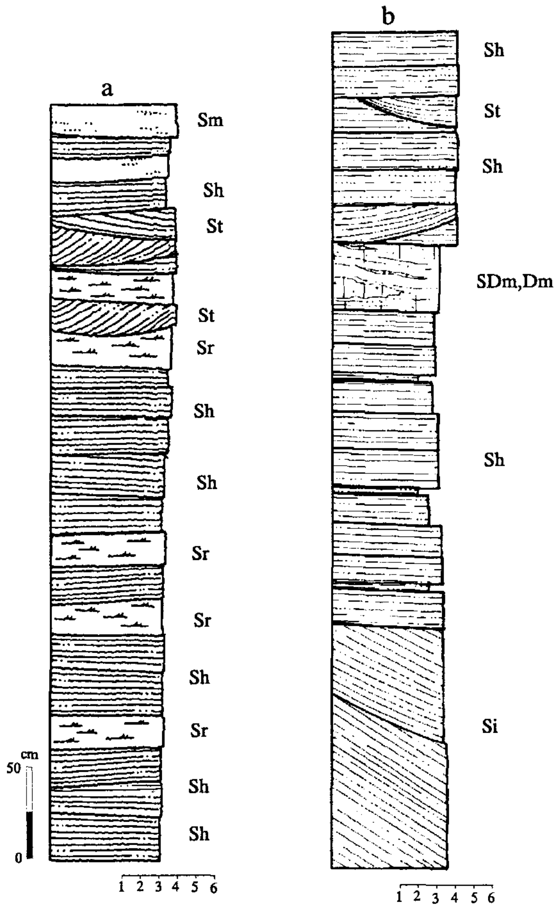


Fig. 7. Sedimentary logs of upper part of glaciofluvial series

with thrusts ca 50 cm. Locally deposits are intensively deformed and form melange zones of diamicton with silt and sand, where primary depositional structures are sparsely visible. Some diamicton beds exist as infills of the erosional troughs. The top surface of this series undulates (Fig. 2) and its character is locally either erosional or depositional.

Large scale of deposit deformations suggests that this series was formed in different conditions than other series were. It seems that deformation origin can be explained by spatially differentiated loading of sediments due to melting of underlying ice blocks and/or ice walls retreat (cf. Boulton 1968, 1970; Lawson 1989). The series described above indicates disintegration of the ice-sheet margin into dead-ice blocks isolated by crevasses.

THE 1st UPPER TILL

Massive diamicton, approximate 1 m thick, lies above the ablation series B (Fig. 3). It is characterized by a strong compaction and grey-brown colour. In the lower part diamicton is sandier, and contains a lot of distinctly oriented clasts. The main mode of their longest axes falls in E/SE range (Table 2), and is quite similar to the mean fabric azimuth obtained from the lower till. Petrographic composition of gravels is shown in Table 3.

This till could be observed only locally, thus its right genetic interpretation is impossible.

GLACIOFLUVIAL SERIES

Sand and sandy-gravel deposits overlie the ablation series (both C and partly A ones). They form 150–200 m wide terrace-like depositional plain (Fig. 2). Its surface reaches ca 12–15 m above the fluvial gravel series. Glaciofluvial series has been subdivided into two parts. In the lower one three lithofacies associations have been noted (Fig. 6). The lower association contains the lithofacies of inclined sand beds Si. Medium-scale sand lithofacies of trough and tabular cross-stratifications St and Sp predominate in the middle association. Subordinate lithofacies Sh of horizontally stratified sands, and lithofacies Sl of low-angle cross-stratified sands have been found there. The local, ripple-derived cosets Sr, have been also noted. The upper association is dominated by massive gravel filling large-scale erosional troughs. They are up to 3.5 m wide and more than 1 m deep. The base of the upper part of the series is built up of 1.5 m thick lithofacies Si (Fig. 7b). Horizontally stratified sandy lithofacies Sh exist above (Fig. 7a). These beds are of medium scale and are composed of medium- and fine-grained sands. Lithofacies Sr are of secondary importance and lithofacies St are quite infrequent there. The sands are grading laterally into more fine-grained deposits (SFm, FSm). The bed of the reworked clasts of diamicton Dm and diamictic sand SDm, has been found in one position. Another association overlies these deposits discordantly. It is built of coarse- and medium-grained sands forming abundant lithofacies Sh and some less common St ones.

Lithofacies Si of the lower association represents the subenvironment of microdelta that progrades into a shallow pond. The second association is a depositional record of relatively small, fluvial braided system developed in shallow sand-bedded channels where the sinuous megaripples (lithofacies St), and distally accreted transverse bars (lithofacies Sp) were formed. During low-water stages the channel bed took plane or nearly flattened configuration (lithofacies Sh and Sl). The ripple bed areas (Sr) originated locally, in these shallow-water conditions. The third association proved a torrential meltwater flood. Large-scale lithofacies Gm, filling-in erosional troughs, resulted due to significant increase in flow competency. Deep trough structures were eroded by high-energy currents during the flood rising stage, whereas the incisions were filled-in with massive gravel, during energy level decreasing in the waning stage. Such ablation floods with overloaded flows are considered as typical of the proximal zones of glaciomarginal alluvial fans (Zieliński 1992).

The succession of three superimposed lithofacies associations reflects the process of sediment filling-in of the shallow lake by microdeltaic progradation. The emerged depositional plain originated due to these phenomena. Sediments were distributed on the plain through the system of small braided channels, however with flows characterized by catastrophic discharges. High grain-size differentiation of deposits as well as relatively small extent of depositional units, point to subenvironment of glaciomarginal alluvial fan with flashy flows.

The upper part of the series can be interpreted as an alluvial fan dominated by sheetfloods. This is mainly evidenced by the presence of lithofacies Sh coexisting with lithofacies Sr (cf. Zieliński and Brodzikowski 1992). Microdeltaic progradation took place on marginal parts of some fans. Lithofacies

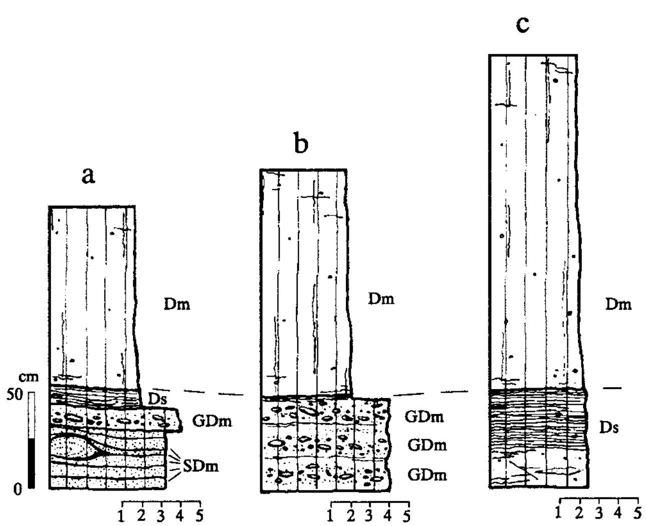


Fig. 8. Sedimentary logs of the 2nd upper till

Si were formed there, as depositional record of the fan-delta slopes. Flow tills (lithofacies Dm and SDm) were deposited on surfaces of non-active fans. Coarser-grained, unconcordantly superimposed associations were derived from younger alluvial fans of some higher energy levels.

The deposits of the upper part of depositional plain are attributed to coalescence of glaciomarginal alluvial fans. Marginal part of the plain was submerged and lacustrine accumulation prevailed herein. Such palaeoenvironmental situation was additionally determined by an increasing level of the glacial lake.

Glaciofluvial sedimentation took place after the period of the ablation series deformation, i.e. after the melt-out phase of the first glacier. It is proved by the fact that deformations typical of ablation series are lacked within deposits mentioned above. In the first stage of glaciofluvial deposition, all the depressions, developed on ablation series surface underwent the filling-in process. The intensive glaciofluvial aggradation caused rapid burying of the older series.

THE 2nd UPPER TILL

Diamicton beds, up to 2 m thick, rest on the glaciofluvial series (Fig. 8). Underlying sands are locally strongly deformed up to the depth of ca 0.5 m. The basal contact of till is sharp and horizontal. Two lithofacies associations have been noted in the till horizon. The lower one is up to 60 cm thick, and is lithologically differentiated to significant degree. In some places it is built of massive, strongly compacted sand forming well-developed layers (Fig. 8a). Specific structures of concentrically laminated, stretched sand ellipsoids, have been found between some layers. The sand passes laterally into two or three layers of gravelly-sandy diamicton, 10–20 cm thick, or into finely bedded silty--sandy diamicton (Fig. 8b,c). Gravel fabric reflects an evident mode of the longest axis dip orientation in ESE/SSE sector (Table 2). In the places, where diamicton is underlain by fine-grained sand and silty sand, the intercalations and lenses of these deposits are noted within it. The upper association, up to 1.5 m thick, is formed by the massive, silty-clayey diamicton. It contains much less gravel clasts and is characterized by a strong compaction. Thin clay lamina was locally traced in the base of diamicton. Also oval, ca 50 cm wide, erosional form has been found, filled-in by coarse-grained sand with sandy-gravelly lag layer in the base.

The lower association is interpreted as the lodgement till. The structures derived from glaciodynamic accretion (Flint 1971; Boulton 1975; Ruszczyńska-Szenajch 1983, 1998) serve the basis for this interpretation. Differentiated till lithology reflects spatially complex and temporarily changing depositional conditions in the glacier sole. Both, the varying bedrock lithology and morphology, were important factors there. The massive sand beds indicate a very short subglacial transport of frozen material. Sediment was eroded quite close to the place of deposition, because there is no lithologic difference between the deposit and substratum. The deposition took place due to a change in temperature and pressure conditions in the glacier sole (Ruszczyńska-Szenajch 1998). Stratified silty-sandy and gravelly-sandy diamictons underwent much longer subglacial transport. Additionally, the following features point out to basal accretion: high compaction ratio of diamicton (Ruszczyńska-Szenajch 1998), relatively low spread of till fabric, and significant substratum deformation, especially the presence of reversed faults (see Hart 1995; Brodzikowski 1978).

The upper association reflects a significant change of depositional environment. Strong compaction of deposit suggests that most probably it was also deposited subglacially. The presence of sand lens with marginal lag-like layer, derived from sub- or englacial water flow, points to basal origin of this till.

HOW MANY TIMES DID THE ICE-SHEET ADVANCE TAKE PLACE AND WHAT WAS THE MAIN DIRECTION OF THE ICE MOVEMENT?

The studied depositional succession was formed during the Sanian II (Elsterian) glaciation. Two till horizons originated due to the ice-sheet advance to the valley. They prove a double glacierization of the analysed area. Petrographic composition and clast fabric testify to two different directions of the ice movement.

Petrographic analyses indicate that number of the Scandinavian rocks is lower than several percent of all the gravels occurring in the till (Table 3). These rocks are strongly weathered, so their content can be underestimated. Gravel debris is dominated with quartz (in amount of 40–60%). Flysch rocks are of secondary importance. The sandstones prevail evidently in this group. Flysch rock content is highly differentiated in particular tills; in the lower till it is approximate 20%, however in the upper one reaches the value above two times higher (45.4%). Generally, the flysch clasts are rounded. This feature indicates glacial redeposition of fluvial sediments. The group of angular clasts is significant in the 2nd upper till. Most presumably they were incorporated into the ice-sheet in a way of weathered bedrock cover exaration, and underwent short glacial transport. They are believed to represent the ice-sheet stream advancing over the Rudzica Ridge. On its northern slopes the ice-sheet acted in conditions of strong compression and the friction on ice/bedrock interface was high. These conditions favoured an enrichment of the basal moraine with the local rocks, whereas quite opposite situation was in the glacier lobe advancing through the valley (cf. Lamparski 1992). This palaeogeographic hypothesis is evidenced by till fabric data. Orientation of elongated clasts in the lower and 1st upper tills (Table 2) is parallel to valley axis, whereas the mean vector obtained from the 2nd upper till is oriented in more southward direction.

PALAEOGEOGRAPHIC SCENARIO

Glaciolacustrine deposit overlying gravel alluvium is the first member connected with the ice-sheet in the studied sedimentary record. The glacial lake was resulted from damming of the side valley by the ice-sheet advancing through the meridionally oriented Vistula valley (Fig. 9a,b). Because of a small scale of the side valley, the lake rapidly enlarged its extent and sedimentation took place in several subenvironments of longitudinally shaped basin.

The lower till (Fig. 3) proves the advancing phase of ice-sheet. The ice transgreding from the Vistula valley, formed the secondary lobe in the lower reach of the Łaziński Potok valley (Fig. 9c). The signal of fast ice movement is confirmed by the fact that the till overlies the distal glaciolacustrine facies. Ice sliding over lubricating soft, clayey deposits was the important factor of this rapid advance.

In the author's opinion the second ice-sheet lobe passed the Rudzica Ridge and flew down to the valley from the north nearly in the same time. Different petrographic content of the lower till and subglacial canal series (i.e. higher frequency of angular local flysch rocks in the second one) proves this conclusion. Presumably the formation of subglacial canal took place in the boundary zone

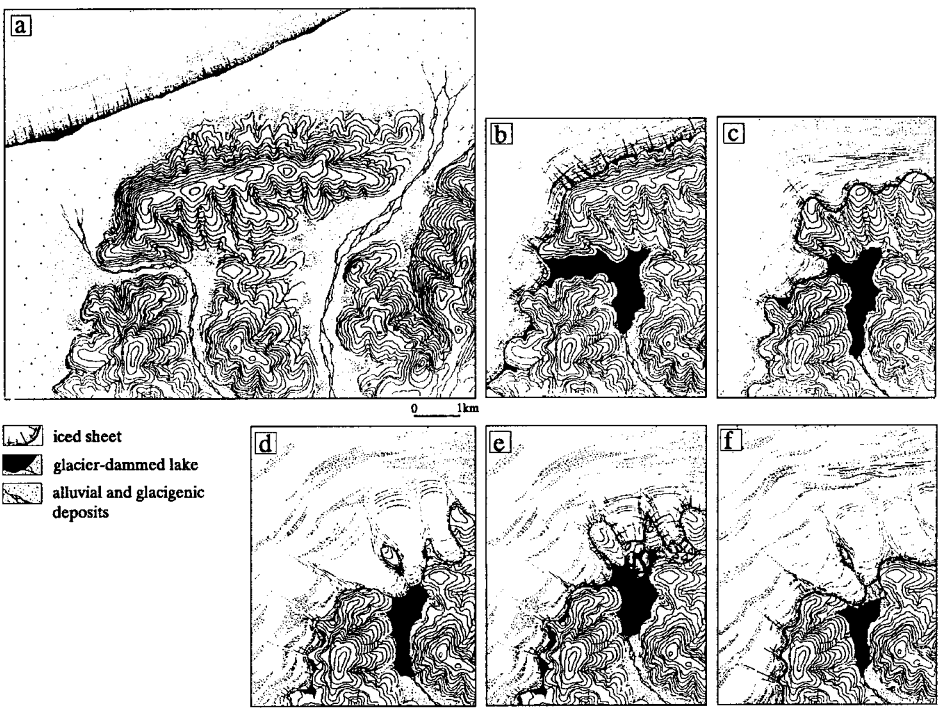


Fig. 9. Successive stages of ice-sheet advance on the NW part of Bielsko Upland area

between two lobes mentioned above (Fig. 9c), and the northern lobe was the source of material.

The ablation series evidences that the ice-sheet retreated soon. Significant lateral lithologic differentiation reflects spatially complex deglaciation picture (Fig. 9d) as well as confirms the hypothesis of two-directional ice advance. Series A originated in the crevasse formed in the place of previously existed subglacial canal. Series B, coexisting westwardly to it, indicates frontal deglaciation style. Flow till beds suggest such manner of the ice decay. Series C, existing on the opposite side of ice-crevasse series, derived from the process of glacier disintegration into dead-ice blocks. Such origin is favoured by the presence of deformations, generated close to decaying ice blocks. The connection of northern lobe with the main ice-sheet body was rather limited due to morphological barrier of the Rudzica Ridge. From this reason its fast areal deglaciation took place in conditions of increased ablation. On the other hand, the western valley lobe was still hardly connected with ice-supplying source and that is why its deglaciation occurred as frontal retreat.

The second ice-sheet advance took place after the short recession period. The 2nd upper till horizon was deposited then (Fig. 9e). The glaciofluvial series was accumulated in front of the ice-sheet margin. Depositional ratio in glaciomarginal zone was so high that ice-dammed valley lake was filled up by the prograding alluvial fans. In this way the glaciofluvial plain developed, on which the ice-sheet prograded, that is documented by the uppermost till cover.

CONCLUSIONS

The study indicates strong relationship between the shape of the ice-sheet margin, depositional processes acted there, and the bedrock morphology. During the first phase of ice-sheet advance its marginal shape was strongly determined by large-scale morphological elements (the Vistula valley and marginal slopes of foothills), but during more advanced phases this influence much weakened. The ice-sheet transgreding onto the foothills zone formed two lobes: first, the most significant lobe in the Vistula valley, and then the secondary lobe in the second-rank valley of the Łaziński Potok. After the marginal part of the ice-sheet had thickened enough the advance took also place by overriding the transverse Rudzica Ridge.

The bedrock morphology considerably influenced not only the progress, but also the character of the ice-sheet retreat. In places, where glacier lobe was still supplied with ice, the frontal deglaciation took place. On the other hand the lobe, where the ice feeding was limited or decayed by the presence of the morphological barrier (the Rudzica Ridge), underwent the areal deglaciation; its margin was disintegrated into dead-ice blocks.

Glaciomarginal depositional style was controlled by bedrock morphology as well. Sedimentation taking place in the valley floor inclined towards the ice-sheet. From this reason the glacial lakes were quite common, and the most of series derived in glaciolacustrine environment, in emerged zone close to the lake.

The northern, marginal slopes of the foothills were significant barrier for englacial hydrologic system. Both the meltwater and glacial debris were provided to the valley by the small-scale ice lobes. Increase of meltwater energy as well as sediment supply (glaciofluvial series origin) indicates that during the subsequent phases of glaciation much more of ablation waters reached the zone of the studied valley. It suggests that this change of sedimentary environment resulted from the factor of increasing ice-sheet thickness in the foothills area.

ACKNOWLEDGEMENT

I thank professor Kazimierz Klimek and professor Tomasz Zieliński for help and constructive discussion during this study.

University of Silesia
Earth Science Faculty
Department of Fundamental Geology
ul. Będzińska 60, 41–200 Sosnowiec, Poland

REFERENCES

Ashley G. M., 1989. *Classification of glaciolacustrine sediments*, [in:] *Genetic classification of glacigenic deposits*. R. P. Goldthwait, C. L. Matsch eds, A. A. Balkema, Rotterdam, Brookfield, 243–260.

Boulton G. S., 1968. *Flow tills and related deposits on some Vestspitsbergen glaciers*. Journal of Glaciology 7 (51), 391–412.

Boulton G. S., 1970. *On the deposition of subglacial and melt-out tills at the margins of certain Svalbard glacier*. Journal of Glaciology 9 (56), 231–245.

Boulton G. S., 1975. *Processes and patterns of subglacial sedimentation: a theoretical approach*, [in:] *Ice Ages: Ancient and Modern*. A. E. Wright, F. Moseley eds, Seel House Press, Liverpool, 7–42.

Boulton G. S., Hindmarsh R. C. A., 1987. *Sediment deformation beneath glacier: rheology and geological consequences*. Journal of Geophysical Research 92 (B9), 9059–9082.

Boulton G. S., Jones A. S., 1979. *Stability of temperature ice sheets resting on beds of deformable sediments*. Journal of Glaciology 24 (90), 29–43.

Brodzikowski K., 1978. *O deformacjach glacitektonicznych*. Czasopismo Geograficzne 49, 2, 137–158.

Brodzikowski K., 1992. *Sedymentacja glacilimniczna, cz.I: Procesy depozycyjne oraz charakterystyka litofacjalna*. Acta Geographica Lodziensia 62, 137 pp.

Burtanówna J., Konior K., Książkiewicz M., 1937. *Mapa geologiczna Karpat Śląskich. Wyniki badań i objaśnienia do mapy*. PAU, Kraków, 104 pp.

Dreimanis A., 1989. *Tills: their genetic terminology and classification*, [in:] *Genetic Classification of Glacigenic Deposits*. R. P. Goldthwait, C. L. Matsch eds, A. A. Balkema, Rotterdam, Brookfield, 17–84.

Dreimanis A., 1990. *Formation, deposition and identification of subglacial and supraglacial tills*, [in:] *Glacial Indicator Tracing*. R. P. Goldthwait, C. L. Matsch eds, A. A. Balkema, Rotterdam, Brookfield, 35–59.

Flint R. F., 1971. *Glacial and Quaternary geology*. John Wiley and Sons, New York, 892 pp.

Gerlach T., Koszarski A., 1981. *Pełny profil osadów zlodowacenia krakowskiego w Niebylcu na Pogórzu Dynowskim*. Sprawozdanie z posiedzeń Komisji Naukowych Polskiej Akademii Nauk, Oddz. Kraków 25, 2, 323–324.

Hart J. K., 1995. *Subglacial erosion, deposition, and deformation associated with deformable beds*. Progress in Physical Geography 19(2), London, 173–191.

Jahn A., 1952. *Profil utworów plejstoceńskich w Górach Kęckich koło Kęt*. Z badań czwartorzędu w Polsce, t. I, Biuletyn Państwowego Instytutu Geologicznego 65, 467–473.

Jahn A., 1960. *Czwartorzęd Sudetów. Geologia regionalna Polski*. Kraków, 3, 358–418.

Jahn A., 1969. *Terasy kemowe w Sudetach*. Folia Quaternaria 30, 17–21.

Jahn A, Szczepankiewicz S.,1967. *Osady i formy czwartorzędowe Sudetów i ich przedpola*, [in:] *Czwartorzęd Polski*, PWN, Warszawa, 397–430.

Jania J., 1993. *Glacjologia*. PWN, Warszawa, 356 pp.

Jersak J., 1991. *Osady rzeczne fazy pełni piętra zimnego Wisły w dolinie Wieprza między Szczebrzyszynem a Łańcuchowem*, [in:] *Less i osady dolinne*, J. Jersak ed., Uniw. Śląski, Katowice, 51–92.

Kasprzak L., Kozarski S., 1984. *Analiza facjalna osadów strefy marginalnej fazy poznańskiej ostatniego zlodowacenia w Środkowej Wielkopolsce*. UAM, Seria Geografia 29, 50 pp.

Klimaszewski M., 1936. *Zasięg maksymalnego zlodowacenia w Karpatach Zachodnich*. Wiadomości Geograficzne 14, 22–26.

Klimaszewski M., 1948. *Polskie Karpaty Zachodnie w okresie dyluwialnym*. Prace Wrocławskiego Towarzystwa Naukowego, Ser. B, 7, 225 pp.

Klimek K., 1972. *Współczesne procesy fluwialne i rzeźba równiny Skeidarársandur (Islandia)*. Prace Geograficzne IG PAN 84, 139 pp.

Konior K., 1939. *O występowaniu warstw interglacjalnych w Wilamowicach*. Starunia 18, 1–7.

Kor P., Shaw J., Sharp D., 1991. *Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: a regional view*. Canadian Journal of Earth Sciences 28, 623–642.

Książkiewicz M., 1935. *Utwory czwartorzędowe Pogórza Cieszyńskiego*. Prace Geologiczne PAU 2, Kraków, 1–13.

Lamparski Z., 1992. *Analiza składu petrograficznego*, [in:] *Czwartorzęd. Osady, Metody badań, Stratygrafia*, L. Lindner ed., Wydawnictwo PAE, Warszawa, 285–293.

Lawson D. E., 1989. *Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits*, [in:] *Genetic Classification of Glacigenic Deposits*, R. P. Goldthwait, C. L. Matsch eds, A. A. Balkema, Rotterdam, 17–84.

Łanczont M., Racinowski R., 1994. *Z badań nad osadami żwirowo-piaszczystymi okolic Pikulic koło Przemyśla*. Annales Universitatis Mariae Curie-Skłodowska, Ser. B, 49, 6, 101–122.

Miall A. D., 1977. *A review of the braided river depositional environment*. Earth Sciences Review 13, 1–62.

Miall A. D., 1996. *The Geology of Fluvial Deposits. Sedimentary Facies, Basin Analysis, and Petroleum Geology*. Springer, Berlin, Heidelberg, NewYork, 565 pp.

Mol J., 1997. *Fluvial response to Weischselin climate changes in the Niederausitz (Germany)*. Journal of Quaternary Science 12, 43–60.

Morawski W., 1984. *Osady wodnomorenowe*. Prace Instytutu Geologicznego 58, 64 pp.

Nye J. F., 1952. *The mechanics of glacier flow*. Journal of Glaciology 2, (12), 82–93.

Paterson W. S. B., 1981. *The physics of glacier*. Pergamon Press, Oxford, 388 pp.

Rust B. R., 1972. *Structure and process in a braided river*. Sedimentology 18, 221–245.

Ruszczyńska-Szenajch H., 1983. *Lodgment tills and syndepositional glacitectonic process related to subglacial thermal and hydrologic conditions*, [in:] *Tills and Related Deposits*, E. B. Evenson, Ch. Schluchter, J. Rabassa eds., A. A. Balkema, Rotterdam, 113–117.

Ruszczyńska-Szenajch H., 1998. *Struktura glin lodowcowych jako istotny wskaźnik ich genezy*, [in:] *Struktury sedymentacyjne i postsedymentacyjne w osadach czwartorzędowych i ich war-*

tość interpretacyjna. E. Mycielska-Dowgiałło ed., Wydział Geografii i Studiów Regionalnych Uniwersytetu Warszawskiego, 13–40.

Ryłko W., Paul Z., 1994. *Mapa geologiczna Polski, arkusz Cieszyn, 1:200 000, A — mapa utworów powierzchniowych*, PIG Warszawa.

Smith N. D., 1970. *The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, North-Central Appalachians*. Bulletin of Geological Society of America 81, 2993–3014.

Starkel L., 1957. *Rozwój morfologiczny progu Pogórza Karpackiego między Dębicą a Trzcianą*. Prace Geograficzne IG PAN 11, 152 pp.

Stupnicka E., 1962. *Geneza i wiek żwirów mieszanych na Pogórzu Cieszyńskim*. Acta Geologica Polonica 12, 2, 263–291.

Szczepankiewicz S., 1953. *Rozwój doliny górnego Bobru u krawędzi lądolodu w Sudetach*. Czasopismo Geograficzne 23/24, 122–137.

Szponar A., 1974. *Etapy deglacjacji w strefie przedgórskiej na przykładzie przedpola Sudetów Środkowych*. Acta Universitatis Wratislaviensis 220, 90 pp.

Walczak W., 1954. *Pradolina Nysy i pleistoceńskie zmiany hydrograficzne na przedpolu Sudetów Wschodnich*. Prace Geograficzne IG PAN 2, 51 pp.

Walczak W., 1957. *Geneza form polodowcowych na przełęczach Sudetów Kłodzkich*. Czasopismo Geograficzne 28 (3), 3–24.

Walczak W., 1969. *Terasy kemowe Gór Bardzkich*. Folia Quaternaria 30, 23–31.

Walczak W., 1972. *Sudety i Przedgórze Sudeckie*, [in:] *Geomorfologia Polski*, t. 1, M. Klimaszewski ed., PWN, Warszawa, 167–231.

Woo M. K., 1998. *Wetland runoff regime in N Canada*, [in:] Proceeding 5th International Conference on Permafrost Tapir Publ. Trondheim, 644–649.

Zieliński T., 1992. *Moreny czołowe Polski północno-wschodniej — osady i warunki sedymentacji*. Uniwersytet Śląski, Katowice, 97 pp.

Zieliński T., 1995. *Kod litofacjalny i litogenetyczny — konstrukcja i zastosowanie*, [in:] *Badania osadów czwartorzędowych. Wybrane metody i interpretacja wyników*, E. Mycielska-Dowgiałło, J. Rutkowski eds, Wydział Geografii i Studiów Regionalnych Uniwersytetu Warszawskiego, 220–235.

Zieliński T., Brodzikowski K., 1992. *Cechy przykładowych sekwencji osadów glacilimnicznych subśrodowiska przyujściowego (z obszaru Rowu Kleszczowa)*. Materiały I Szkoły Sedymentologicznej, Murzynowo k. Płocka, wrzesień 1992, tom UŁ., 143–157.

STRESZCZENIE

T. Salamon

SUKCESJA OSADÓW ORAZ WPŁYW MORFOLOGII PODŁOŻA NA PALEOGEOGRAFIĘ STREFY GLACIMARGINALNEJ W OBSZARZE PRZEDGÓRSKIM — PRZYKŁAD Z PRZEDPOLA BESKIDU ŚLĄSKIEGO

Celem artykułu jest charakterystyka środowisk sedymentacji osadów glacigenicznych w niewielkiej dolinie Łazińskiego Potoku z przedpola Beskidu Śląskiego (ryc. 1, 2). Stanowisko badawcze znajduje się w sąsiedztwie doliny Wisły, gdzie przed laty badania utworów czwartorzędowych prowadzili Książkiewicz (1935) i Stupnicka (1962). Osady w dolinie Łazińskiego Potoku wiążę ze strefą marginalną lądolodu San II, który w czasie maksymalnego zasięgu doliną Wisły wkroczył do Kotliny Ustronia (Książkiewicz 1935).

Położenie doliny za krawędziowym wzniesieniem pogórza umożliwia bezpośrednią analizę lokalnych uwarunkowań rozwoju strefy glacimarginalnej w obszarze przedgórskim. W odsłonięciu wyróżniono szereg serii osadów glacigenicznych, związanych z kolejnymi etapami nasuwania lądolodu. Powyżej żwirów rzecznych deponowanych w czasie transgresji lądolodu wyróżniono: serię zastoiskową (ryc. 3), utworzoną w wyniku podparcia doliny przez lądolód; glinę dolną (ryc. 3, tab. 2, 3) oraz serię rynny subglacjalnej (ryc. 3, 4), związane z pierwszym nasunięciem lądolodu w dolinę; serię ablacyjną krótkiej fazy recesji (ryc. 3, 5); serię fluwioglacjalną (ryc. 6, 7) i glinę górną drugiego nasunięcia lądolodu w dolinę (ryc. 8, tab. 2, 3).

Badania wykazały dużą zależność kształtu strefy glacimarginalnej i procesów sedymentacji od rzeźby podłoża. Lądolód po dotarciu do krawędzi pogórza formował najpierw wyraźny lob w strefie doliny Wisły (ryc. 9a,b). Następnie wsuwał się mniejszym jęzorem w boczną dolinę Łazińskiego Potoku (ryc. 9c). Kiedy miąższość lodu wspartego na krawędzi pogórza wzrosła, awans w strefę pogórza odbywał się także ponad równoleżnikowym Garbem Rudzicy.

Morfologia podłoża miała decydujący wpływ na przebieg i charakter zjawisk glacjalnych podczas fazy recesji, jaka nastąpiła po pierwszym nasunięciu. W strefie doliny, gdzie do aktywnego czoła stale dopływał lód lodowcowy, zachodził proces frontalnego cofania. W strefach, gdzie zasilanie lodem ograniczone było lub ustawało całkowicie, ze względu na przeszkody morfologiczne podłoża (fragment czoła transgredującego ponad Garbem Rudzicy), czoło rozpadło się na martwe bryły.