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Development and infill of Vistulian glacial Lake Gniew (N Poland): a sedimentological analysis

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A substantial glaciolacustrine unit in northern Poland, between the valleys of the Vistula and Wierzyca rivers, was deposited in glacial Lake Gniew during the climatic amelioration of the Late Vistulian. It covers an area of 35 km^2 and has an average thickness of 7 m, but locally reaches over 20 m. Four sedimentary facies are distinguished. The silty/clayey rhythmites of facies A are interpreted as varves from the central lake bottom. They represent the initial stage of lake development. Facies B is formed by a single sand layer that is interpreted as a turbidite originating during lake shallowing due to self-drainage. Facies C consists of massive clay with dropstones and dump structures. Two hypothesis regarding its genesis are put forward: the first assumes sedimentation in a shallow basin with a high input of homogenous fine-grained suspended sediment, whereas the second explains the facies as a result of a muddy jökulhlaup, pouring into Lake Gniew and being sourced from another glacial lake; further research is required to interpret this facies reliably. Facies D consists of thick silty/clayey rhythmites that are interpreted as prodeltaic deposits.

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INTRODUCTION

Glacial lakes are one of the most characteristic elements of glaciomarginal zones and occur in various positions with respect to the ice sheet and its foreland. The geological literature provides numerous examples of lakes connected directly with an ice sheet; these include subglacial lakes (Gilbert, 1990; Shoemaker, 1991; Rovey and Borucki, 1995; Siegert, 2000; Dowdeswell and Siegert, 2002; Munro-Stasiuk, 2003), extensive terminoglacial ice-dammed lakes (Merta, 1978, 1986; Gilbert and Desloges, 1987; Livermann, 1987; Donnelly and Harris, 1989; Eyles *et al*., 2005), as well as usually much smaller lakes developed among dead-ice blocks (Ehlers, 1994; Huddart and Benett, 1997; Bennet *et al*., 2000).

Glacial lake sediments are generally easily recognisable. Their characteristic feature is lamination derived from the annual, summer/winter cyclicity. Glacial varves are therefore the deposit that is typical of all types of large glacial lakes. Nevertheless, lakes are depositional systems consisting of a variety of sedimentary environments, whereas the varves characterise only their central, most distal parts. Moreover, the depositional system of glacial lakes is very sensitive to changes in both the lake's sedimentary environment and the source area. Factors such as a lake's dimensions, depth, distance from the ice sheet or glacier, type and quantity of material supplied may all affect the dynamics of the lake water, its thermal stratification, and the type of material that is deposited. Glaciolacustrine deposits may thus develop in many sedimentary facies representing various sedimentary subenvironments such as lake bottoms, subaqueous fans and deltas.

The object of this study is a large glaciolacustrine unit in northern Poland, close to the towns of Gniew, Janiszewo and Lignowy, between the valleys of the Vistula and Wierzyca rivers (Figs. 1 and 2). It covers an area of 35 km^2 and its thickness averages 7 m, but locally reaches over 20 m. This unit, known in the literature under the name of "Gniew clays", was first described by German geologists at the end of the 19th century (Jentzsch, 1888; Sonntag, 1911, 1913, 1919). From that time on, the unit has been interpreted in several ways. Jentzsch (1888) treated it as an organic-rich surface layer overlying a till (German: "Deckton"). Sonntag (1919) interpreted it as a part of an extensive geomorphological unit together with the Gdańsk ice-dammed lake, which is located

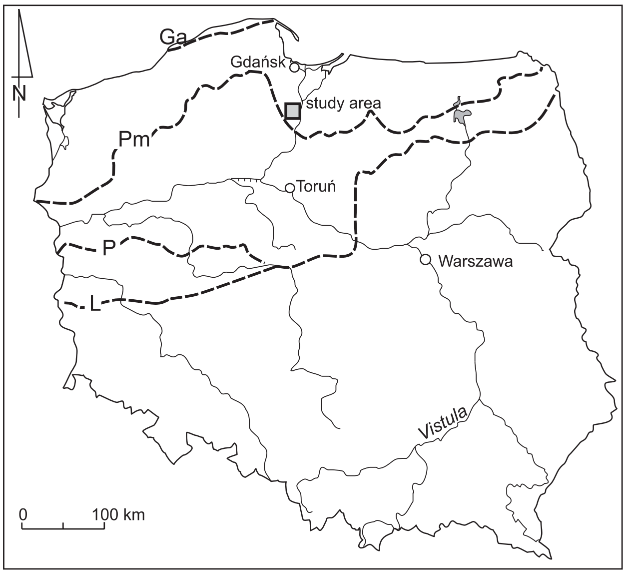

Fig. 1. Map of Poland with the location of the study area and maximum extent of the Late Vistulian ice sheet (after Kozarski, 1995)

L — Leszno phase, P — Poznań phase, Pm — Pomeranian phase, Ga — Gardno phase

north of the Gniew clays. Much later, Kotański (1956) also considered the genesis of the Gniew glaciolacustrine deposits as connected with the development of the large Gdańsk glacial Lake. According to Kotański (1956), the glacial lake originated due to damming of the meltwater from the ice sheet; the dark colour of the deposit was interpreted as a result of reducing conditions in the lake water. More recently Wrotek (1986) interpreted the Gniew clays as the deposit of an ice-dammed lake that was formed during ice retreat episode of the Pomeranian phase of the Vistulian Glaciation. Rosa (1996) presented a quite different and surprising idea, viz. that the Gniew clays are a marine or deltaic deposit transported by the ice sheet away from the bottom of the Gdańsk Bay.

METHODS

A new sedimentological analysis of the Gniew glaciolacustrine deposits was possible thanks to the existence of extensive quarries close to Gniew (the Gniew Keramzyt and Gniew Brickyard sites; Fig. 2), and to exposures in the steep slope of the Wierzyca valley (the Janiszewo site; Fig. 2) Figure 2. Geological Maps at a scale of 1:50 000, the Gniew (Wrotek, 1986) and Starogard Gdański (Rabek, 1987) sheets, were a primary source of information. The area was geologically and geomorphologically mapped, sedimentary logging was carried out at all sites, and in between the boreholes were drilled (mechanical drill *WH-5* and *Geoprobe*).

OBJECTIVES

The aim of this study was to sedimentologically characterise all the deposits in the lake basin. The most characteristic feature of this deposit is the rapid vertical changes of sedimentary facies. By means of analysis of the reasons for these rapid facies changes, the sedimentary environments were reconstructed. This analysis, which was combined with geomorphological studies of the area, form the basis for a new interpretation of the glaciolacustrine sedimentary conditions in this lake.

GEOLOGICAL AND GEOMORPHOLOGICAL SETTING

The northern and western border of the Gniew clays are morphologically barely traceable. There is no sudden change in the slope of the area and the deposits interfinger with tills that cover a plateau. In the east, the area is bordered by the slope of the Vistula valley, which is considerably lower. The southern border coincides with the lower reaches of the Wierzyca valley. In between there is the morainic plateau, which is about 10–20 m lower than the Gniew clay surface. A distinct morphological scarp is present there; no erosion is visible (Figs. 2 and 3).

The surface of the area covered by the Gniew clays is undulating, with an average height of 60–65 m a.s.l. There are many topographic highs and kettle-holes. One of the largest depressions connected with the melting of a large dead-ice block is situated north of Gniew, close to Ciepłe (Figs. 3 and 4). This basin is filled with Late Glacial and Holocene lacustrine deposits, over 16 m thick (Błaszkiewicz, 2003). The Gniew clays area also contains a deep trough, running from Gniewskie Młyny to Szprudowo, with kame hills built of sand and covered by glaciolacustrine deposits (Błaszkiewicz and Gierszewski, 1989). There is also a longitudinal hill in the central part of the Gniew clays, close to Kursztyn; it runs N–S, and shows height differences of up to 15 m. The Gniew clays reach their highest position (70 m a.s.l.) in these hills, which also are the location where the clays reach their minimum thickness (Fig. 2).

Insufficient data are available, unfortunately, to reconstruct the topography below the Gniew clays in detail. Field data indicate, however, that the morphology of the lower boundary of the clay is much more irregular than its present surface: depressions of several tens of metres in diameter and several metres deep are common.

STRATIGRAPHY

The area of the Gniew clays lies north of the line of maximum ice extent of the Pomeranian phase of the late Vistulian (Weichselian) Glaciation (Fig. 1). This line has, however, not been examined in detail in this region, and is still under discussion (Roszko, 1968; Błaszkiewicz, 2003). The age of the Pomeranian phase has not been established unambiguously either. It was dated indirectly (by comparison with the Gardno phase) and is assumed to be between 16 200 years ^{14}C BP (Kozarski, 1995) and 15 200 years ^{14}C BP (Marks, 2002). The accumulation of the glaciolacustrine deposits in glacial Lake Gniew is thought to have taken place during this time interval.

Stratigraphical subdivisions of the Vistulian Glaciation in the Gniew area (and, more generally, in the lower reaches of the Vistula) have been discussed since the end of the 19th century (Berendt, 1865; Galon, 1934; Kotański, 1956; Makowska, 1979; Drozdowski, 1986; Tomczak *et al*., 1999,

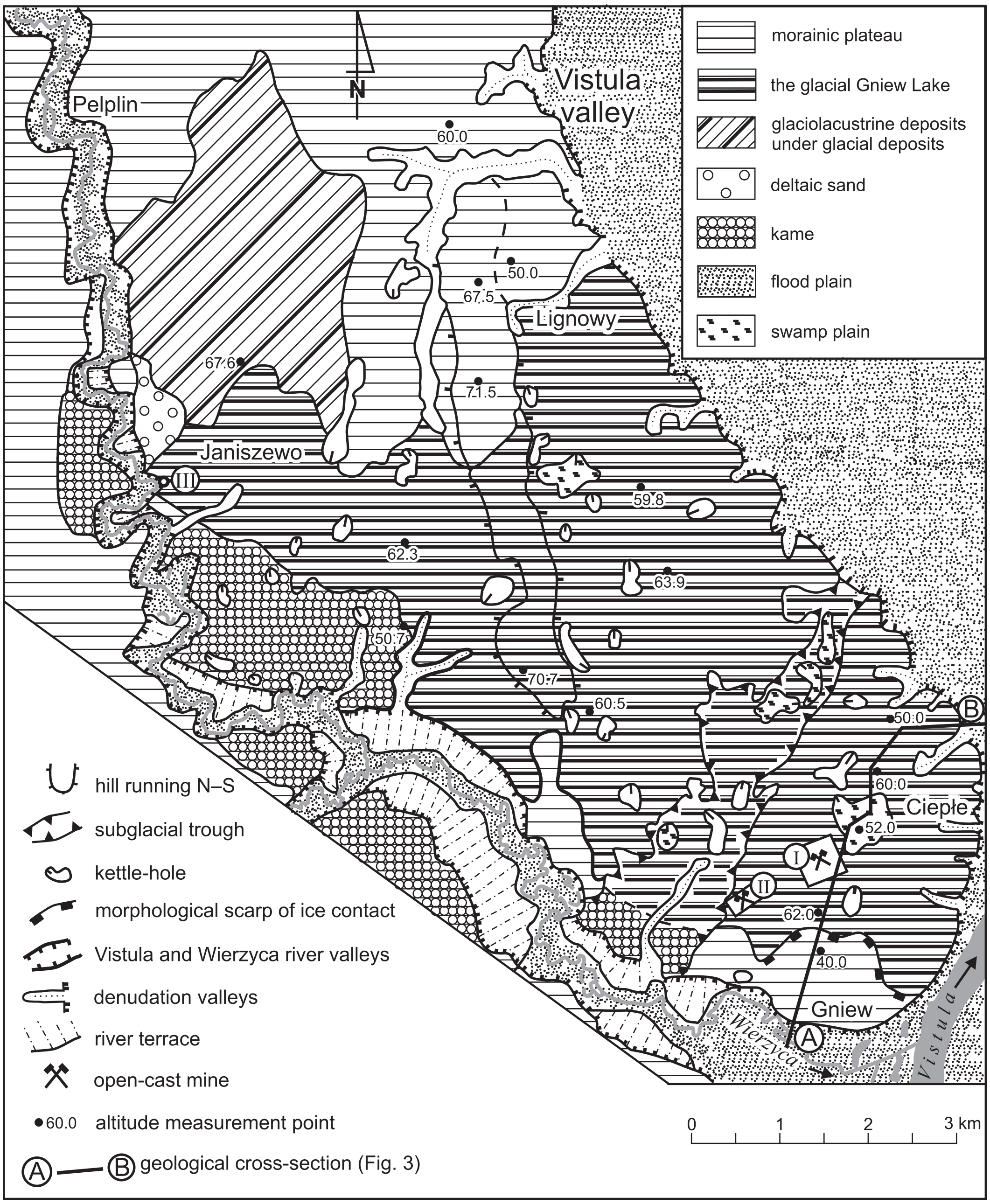

Fig. 2. Geomorphological sketch of the glacial Gniew Lake based on our investigations and on the Detailed Geological Map of Poland 1 : 50 000, the Gniew sheet (Wrotek, 1983) and the Starogard Gdański sheet (Rabek, 1987)

I — Gniew Keramzyt site; II — Gniew Brickyard site; III — Janiszewo site

Błaszkiewicz *et al*., 2002). Under a few metres of till, a thick succession of sandy/gravelly glaciofluvial deposits with a marine fauna is present (Fig. 3). This fauna used to be interpreted as Eemian, whereas the deposits in which it was found were correlated with the Krastudy Interglacial of the Early Vistulian (Makowska, 1990, 1992) or with the Grudziądz Interstadial of the Middle Vistulian (Drozdowski, 1986). Palaeoenvironmental analysis of the fauna proved that most of the species recognised did not live in the warm Eemian sea but in the cold sea of the Grudziądz Interstadial (Błaszkiewicz *et al*., 2002). The sandy/gravelly deposits containing this fauna were accumulated in the floodplain of a braided river, which represented outflow of meltwater to the SW. The marine fauna was eroded by streams from marine deposits that had been transported southwards by the ice sheet from the area now occupied by the Baltic Sea. The presence of the marine fauna from the Grudziądz Interstadial as well as the direct dating of the deposit indicate that the glaciofluvial deposits should be stratigraphically connected with the initial phase of the Late Vistulian Glaciation. The till horizon overlying the glaciofluvial succession is related to the advance of the ice sheet. The existence of a separate till unit dating from the Pomeranian phase is still an open question (Błaszkiewicz *et al*., 2002).

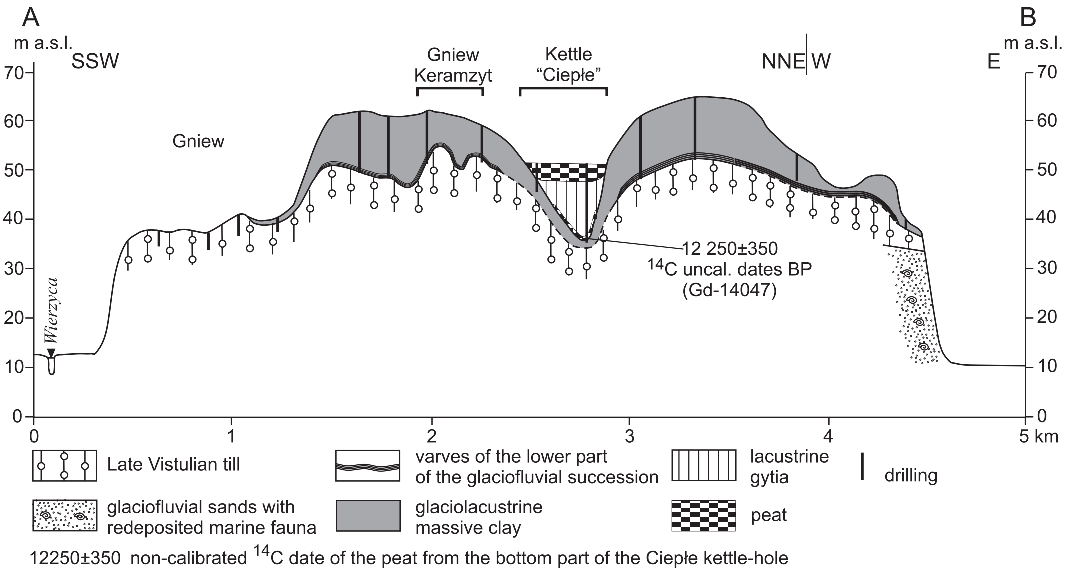

Fig. 3. Simplified geological cross-section through the southern part of glacial Lake Gniew

For location of the cross-section see Figures 2 and 4

The Gniew clays directly overlie the Late Vistulian till. Stratigraphically they belong to the youngest part of the Vistulian, and their origin is connected with the retreat of the Pomeranian ice sheet.

DESCRIPTION OF THE SEDIMENTARY FACIES

The Gniew clays show a wide variety of facies. The following can be distinguished:

Sedimentary facies A consists of laminated silts and clays, and is present throughout the area of glacial Lake Gniew. The maximum thickness is slightly over 2 m in the eastern part of the lake (Gniew Keramzyt site). The deposits occur in the form of silt/clay rhythmites (Fig. 5); in the thicker parts of the association 25 of these rhythmic couplets are present (Fig. 6). In the lower part of the facies association a few intercalations of sandy diamicton with indistinct horizontal lamination occur. The diamicton layers contain silty/clayey clasts and small pebbles. The silty parts of the silt/clay rhythms are built of silt-sized grains (50–75%) with clay admixtures of up to 25–45%, and

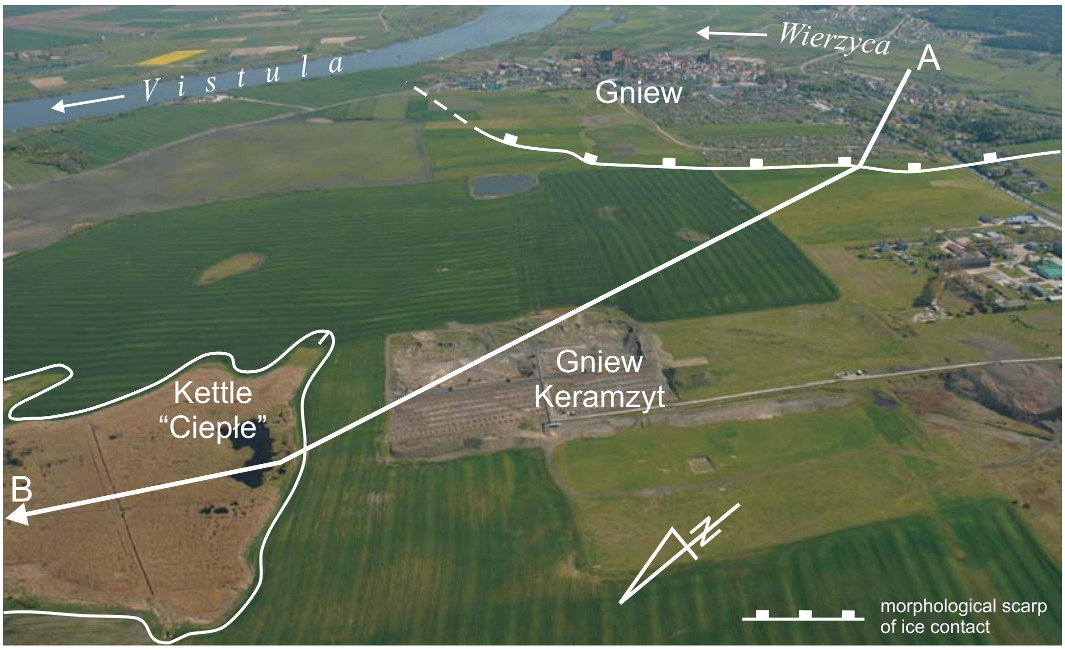

Fig. 4. Aerial view of the southern part of the glacial Gniew Lake area, showing the Ciepłe kettle-hole, the Gniew Keramzyt site and the sedimentary escarpment marking the contact with the ice block

Location of the geological cross-section A–B see Figure 3 (photo W. Stępień)

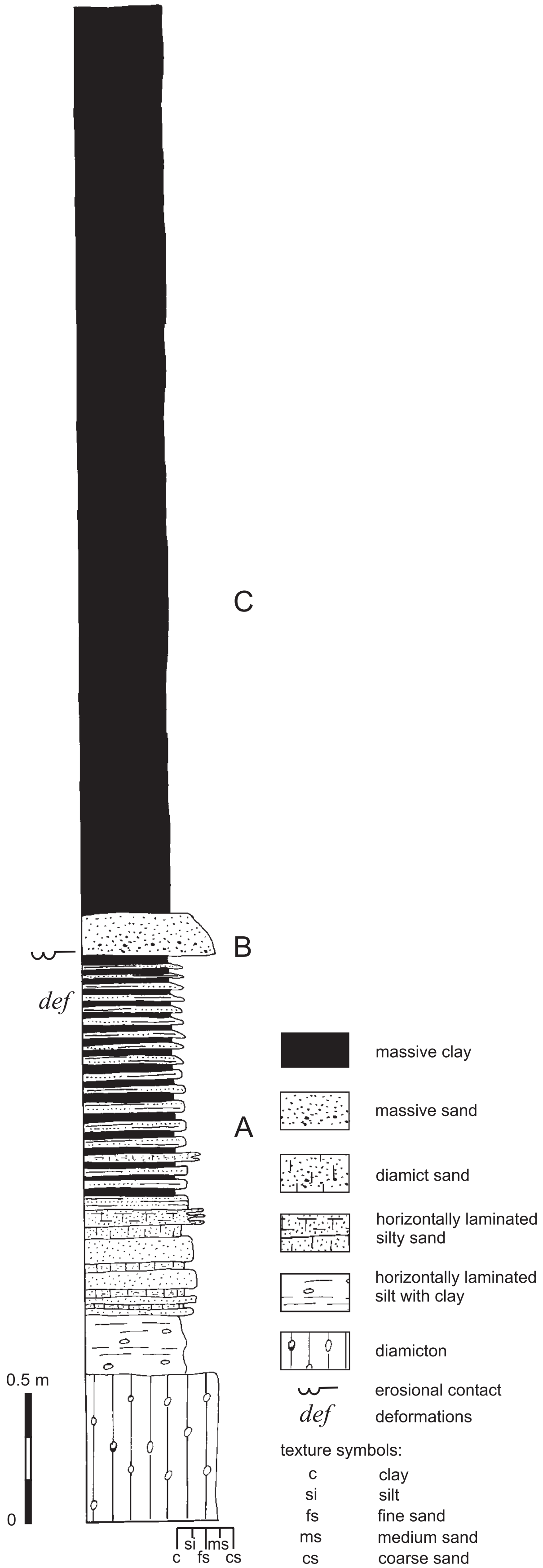

Fig. 5. Composite sedimentary log of the "Gniew clays" with succession of sedimentary facies A, B and C

Fig. 6. Clay varves of facies A; note a decrease in thickness of rhythmic units towards the top

sometimes with a minimal admixture of sand particles. Horizontal lamination is always present. The upper parts of the rhythmites are built of almost pure clay (clay particles up to 99%). The structure is massive as a rule, but the lower parts of the clay layers are sometimes laminated. Thin sandy intercalations are sometimes present (Fig. 6). The boundaries between successive rhythmites as well as those between the silty and the clayey parts of a rhythmite do not show signs of erosion. The lower and upper parts of the rhythmites differ in their content of calcium carbonate. The average carbonate content of the silt units is 20–22%, whereas it is much lower in the clay units (13–18%). The content of organic matter differs only slightly for the silt and clay units, at 3.6–4.1% and 5.9–6.5%, respectively. The average thickness of one silt/clay couplet is 10 cm in the lower part and 3 cm in the upper part of the facies. Towards the top, the thickness of the clay layers increase, while the thickness of silt layers slightly decrease. Individual pebbles up to 2 cm across occur dispersed over the entire facies; they press down the underlying sediments (Fig. 7A) and the younger sediments form drape structures over them.

The upper part of facies A is deformed. Flow folds comprising a few silty/clayey rhythmite units are common; various types of contorted lamination (with convolutions as the most frequent type) are also present (Fig. 7B). Most often the lower and upper bedding planes of the deformed beds are undisturbed. Apart from plastic deformations, normal faults are present with displacements from a few centimetres to a few decimetres (Fig. 7C). Some cut the upper part of the underlying till, run through the entire facies thickness and end in the overlying massive clays (facies C). As a rule they form conjugate systems with sets of complementary faults. In places where the substra-

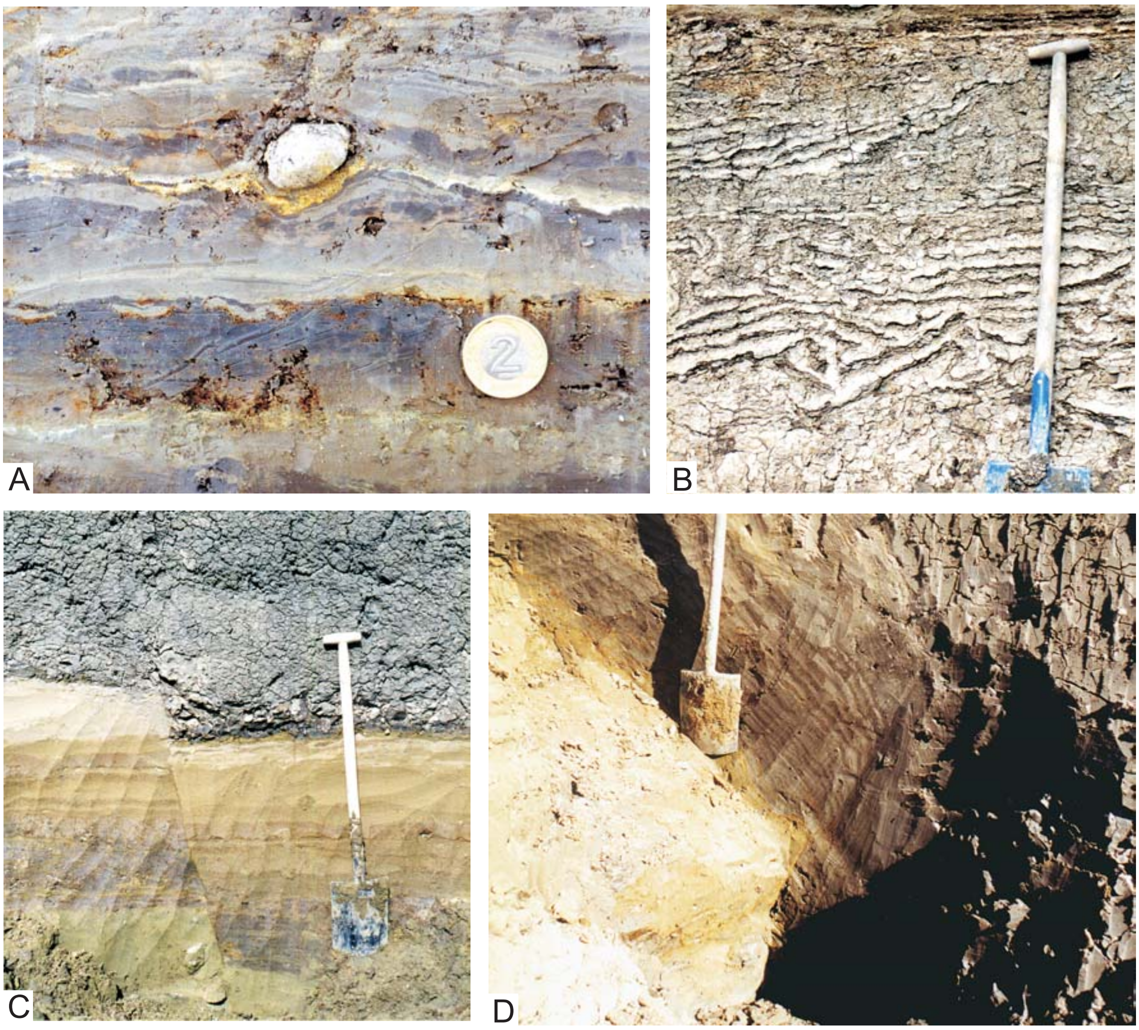

Fig. 7. Sedimentary facies A

A — dropstone in density-deformed silt/clay rhythmites; **B** — contorted lamination from the upper part of the varve series; **C** — normal fault cutting the facies A, B and C; **D** — reverse fault

tum undulates, such as in depressions within the substratum, the displacements by faults are as much as 3 m. Reverse faults are present in the marginal parts of the depressions (Fig. 7D).

Sedimentary facies B consists of a 10–30 cm thick sand bed, erosionally overlying the silt/clay rhythmites (Fig. 8). This bed occurs across the southeastern part of the Gniew clays. It has also been found in most of the boreholes that have penetrated the clay of lithofacies C. The lower part of the facies B sand bed is composed of medium-grained, massive sand, commonly with normal grading. The middle part consists of medium- to fine-grained sand with horizontal lamination or flaser structures. In the upper part, which consists of fine-grained sand, indistinct climbing ripples of A and B types are present. Locally, a completely developed five-member Bouma sequence is found.

Sedimentary facies C is generally built of a massive clay with a thickness of 4–18 m. The lower 0.5 m shows indistinct lamination (Fig. 9A), which disappears towards the top, but which is found sporadically in the middle part of the clay. The grain size does not change upwards. Clay dominates (50–97%); an admixture of silt reaches locally up to 50%,

Fig. 8. Sedimentary facies B

Sandy bed with a massive structure between the varves of facies A and the massive clay of facies C

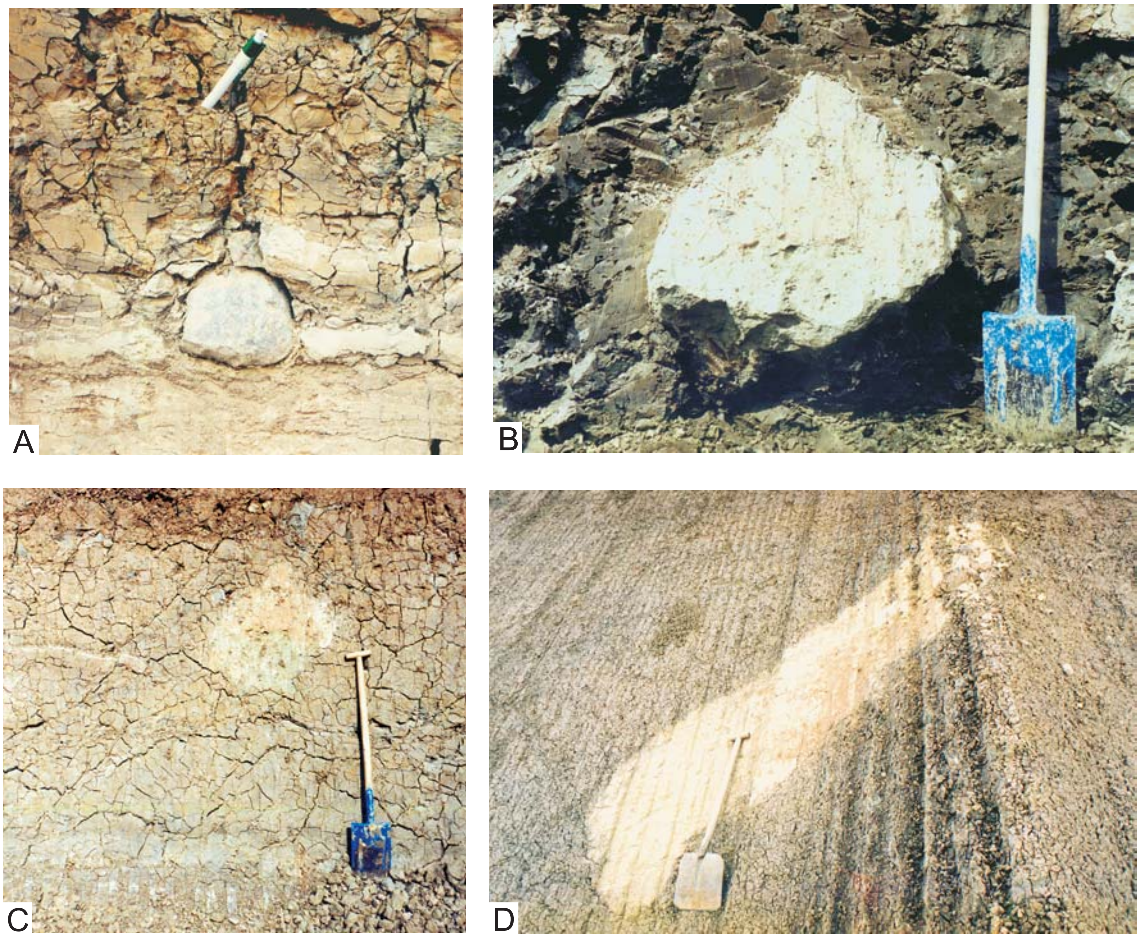

Fig. 9. Sedimentary facies C

A — lower part of facies C with indistinct lamination and numerous dropstones; **B** — diamicton dropstone in the form of a droplet; **C** — diamicton and gravel clasts in massive clay; **D** — large diamicton dump structure in the clay

whereas sand is of little importance. The average carbonate content is 16–18%, and the content of organic matter amounts to 6–7%. Randomly dispersed gravel clasts are found throughout the clay. In the lower part of the clay, above the sand bed of facies B, the concentration of clasts is higher than towards the top (Fig. 9A). The average diameter of clasts ranges from 2 to 5 cm; the largest ones are up to 20 cm across, and occur in the lower part of the clay. The most characteristic feature of the massive clay is the presence of diamicton clasts. Apart from a few small clasts (Fig. 9B, C), three unusually large diamicton clasts have been found, the largest with a long axis of 3.5 m (Fig. 9D). In the lower part of the clay, faults seem to die out that start in the silty/clayey rhythmites of facies A. The homogeneous and massive character of the clay prevents tracing the faults further upwards. Slickensides are common.

Sedimentary facies D is present in the northwestern part of the lake deposits (Janiszewo site). The deposits, which overlie the Late Vistulian till and reach a thickness of up to 8 m, consist of silts and clays in rhythmic couplets with an average thickness of approx. 40 cm (Fig. 10). The coarser-grained parts consist of silt with a massive structure or of horizontally laminated sandy silt. Layers of diamict or clayey clasts are common within the massive silt (Fig. 11A). The clasts are small (up to 1 cm across) and rounded. Small pebbles of the same size are sporadically found. Locally the silts are normally graded with clast-rich layer in their lower part, but in places the clast-rich layers are present in the upper part of the unit. Clay is present in layers 1–3 cm thick. Most frequently the clay layers are slightly laminated in their lower part and massive in their upper part.

Deformations are common in the entire facies. Pillow structures are common in some of the thicker beds. Layers with quasi-boudinage structures are also present. Normal faults with displacements of up to a few centimetres are frequently observed throughout this facies (Fig. 11B).

FACIES INTERPRETATION

Sedimentary facies A. The lower part of facies A (Fig. 5) is interpreted as the result of deposition in the transitional zone between glacial and glaciofluvial environments. Horizontally laminated sandy diamictons probably reflect fluvial resedimentation but laminar cohesive flows might be an alternative depositional agent. Such deposits of cohesive flows sometimes show indistinct lamination, especially in the lower part of beds.

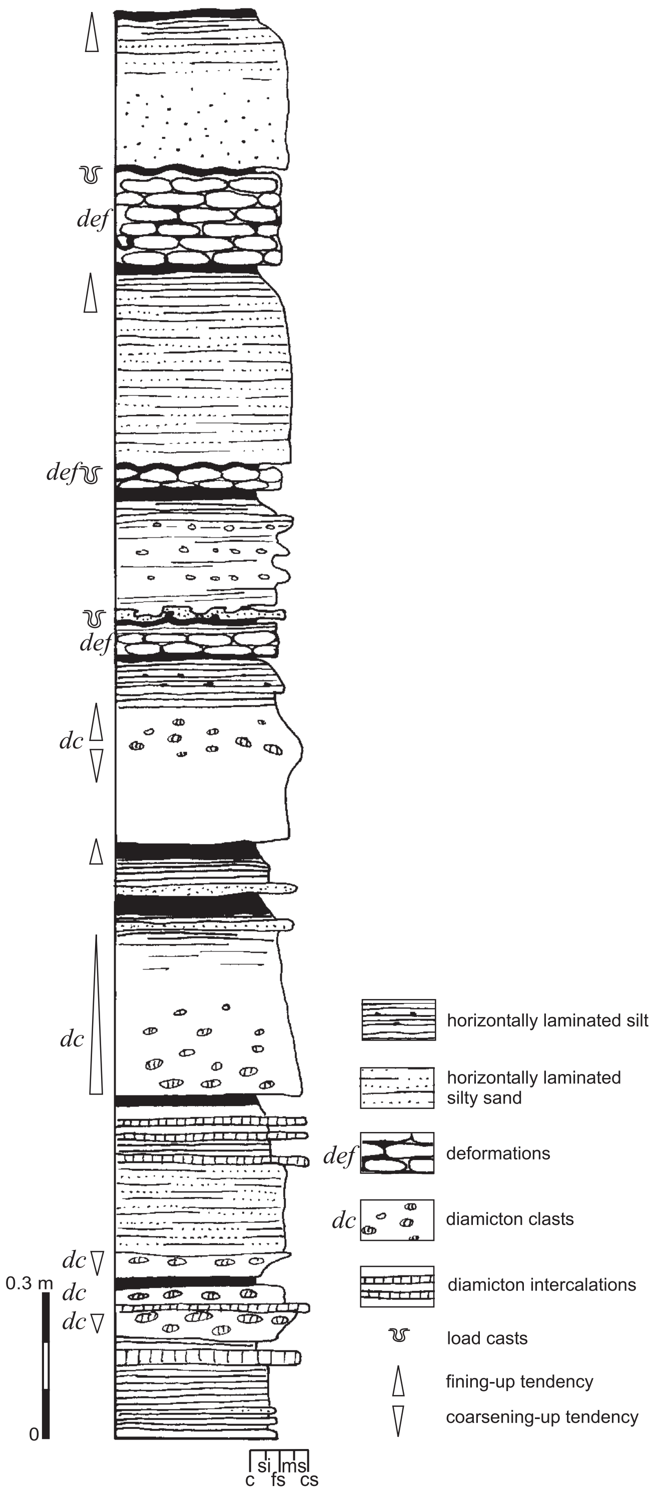

Fig. 10. Sedimentary log of facies D interpreted as prodeltaic deposits

Other explanations as on Figure 5

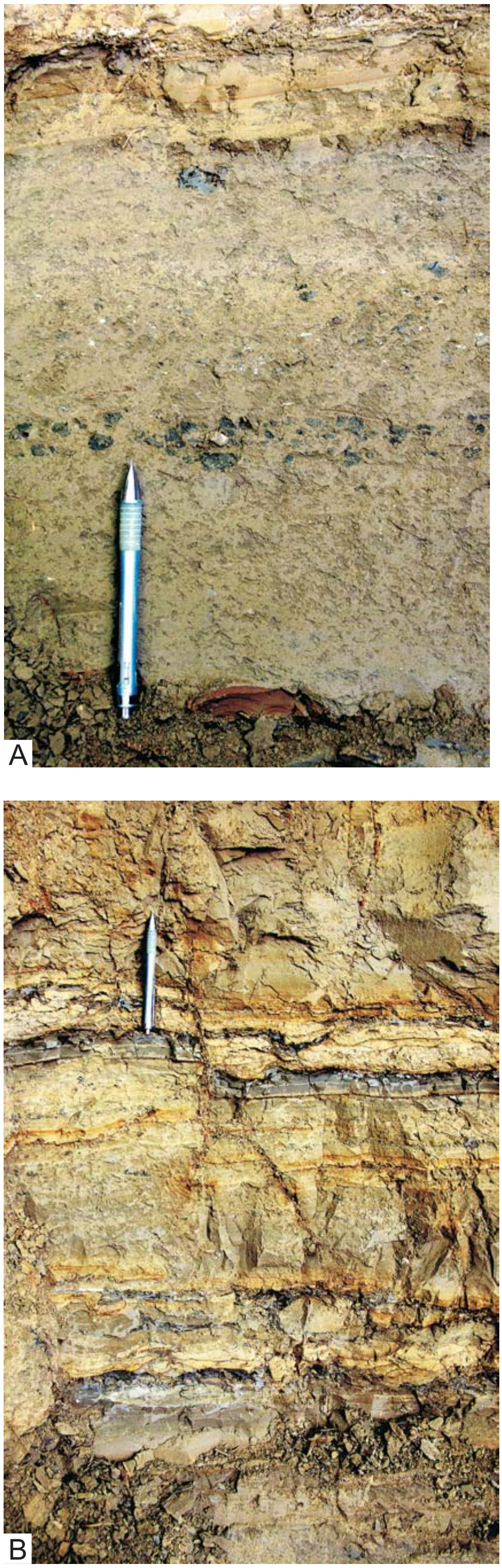

Fig. 11. Sedimentary facies D

A — horizons with small diamicton clasts in a silt bed of considerable thickness; **B** — normal fault cutting the prodeltaic deposits with accompanying set of other deformations caused by unstable density gradients

Resedimentation can be due to weak currents, which erode and wash away the deposits of the flow head (gravity winnowing; Postma, 1984). The silty deposits formed by settling from suspension. The fine-grained deposits together with intercalated mass-flow deposits represent the first sediments in the newly formed glacial lake.

The silty/clayey rhythmites above the transitional zone (Fig. 6) can be interpreted as glacial varves. The commonly gradational transition between the lower and upper parts of varves suggests deposition from turbidity currents (Kuenen, 1951). This mechanism is commonly accepted as dominating the formation of varves (Banerjee, 1973; Ashley, 1975; Merta, 1978). The varves represent distal, low-concentration, turbidites. The upper part of each varve results from suspension settling of clay in a standing water body of a most probably frozen glacial lake. The trend of an upward increase in the clay-layer thickness and a simultaneously decreasing thickness of the lower silty part can be interpreted as the result of a more distal environment. This may be either the result of a decreasing amount of material supplied to the lake during the summer season and an increasing distance from the source, or be due to an increase in the lake size, while the sediment supply remained unchanged.

The presence of gravels in the fine-grained laminated sediments (Fig. 7A) can be explained by ice rafting (Brodzikowski, 1993). The characteristic downward bending of sediment layers are the result of pebbles falling onto the soft lake bottom; the upward bends of the sediment on top of these dropstones are a result of draping by the succeeding sediment laminae (Thomas and Connell, 1985).

The deformation structures (Fig. 7) are partly syndepositional, partly postdepositional. Their origin is commonly associated with melting of dead-ice blocks under the glaciolacustrine sediments. This is indicated particularly by the faults. These displace not only the varves but also the deposits of the overlying facies, and thus they originated postdepositionally. A few metre-high displacements may be explained by either rapid melting of dead ice under the sediments, or by collapse processes on top of a cavern resulting from dead-ice melting that was filled partly with meltwater.

Sedimentary facies B. The sand bed was distinguished as a separate facies because its grain-size distribution differs remarkably from the underlying varves and the overlying massive clays (Fig. 8). The structures within the sand bed, which are diagnostic of a Bouma sequence, strongly suggest deposition by a turbidity current. The commonly erosional base supports this interpretation. This implies that the entire lithofacies was formed during one shortlived event. The considerable thickness of the sand bed and the presence of its lower Bouma interval indicate the proximal character of the current. The presence of this bed throughout the area of the Gniew clays supports the hypothesis of a large-scale event.

Sedimentary facies C. The massive clays of this lithofacies are the result of suspension settling. The homogeneous character indicates that clay particles could settle equally all over the lake. The indistinct lamination in the lower part of the clay may be due to weak bottom currents.

The clasts in the massive clay (mainly fine- and medium-grained gravels; Fig. 9A), which occur commonly, are interpreted as dropstones from melting icebergs. This supports the idea of a lake within a short distance of the ice sheet. Icebergs and ice rafts are a common source of dropstones (Brodzikowski and van Loon, 1991; Brodzikowski, 1993). Ice-rafting is widely discussed in the literature (N. Eyles *et al*., 1983, 1987, N. Eyles and Miall, 1984; N. Eyles and Clark, 1985). This indicates that the occurrence of the boulder-sized diamicton clasts (Fig. 9B–D), larger than any clasts previously found in Poland in glaciolacustrine sediments, can also be explained by ice-rafting: even much larger diamicton masses may be released from a tumbling iceberg, thus producing dump deposits (Thomas and Connell, 1985). Diamict can be released from an iceberg as completely or partly frozen blocks. It cannot be excluded that an iceberg was large enough to become grounded on the lake bottom. In such a case, the largest diamicton clasts could be interpreted as an iceberg dump till (Thomas and Connell, 1985). The homogenous character of the clay unfortunately prevents observation of any deformation structures that could be associated with iceberg grounding (*cf*. Winsemann *et al*., 2003; Mokhtari Fard and Van Loon, 2004; N. Eyles *et al*., 2005). The large scale of the diamict clasts is the only strong indication that such a process is responsible for their occurrence in this clay unit.

The clays of this facies contain relatively large amounts of calcium carbonate and organic matter. This would suggest climate warming if there were a considerable increase in these components in comparison to the underlying varves. This is not the case for the calcium carbonate, however, so the high calcium carbonate content should rather be connected with clastic input of allochthonous carbonates, derived from carbonate rocks eroded by the ice sheet (Kelts and Hsü, 1978). The concentration of organic matter is slightly higher in the massive clays than in the underlying varves. This difference is so slight, however, that it cannot prove a climate amelioration sufficient to influence the nature of sedimentation.

It must be emphasized here that the thickness of the massive clays is unusual when related to the dimensions of the lake. Such thick massive clays are well known from glaciomarine deposits, where they are formed by prolonged settling of large amounts of fine-grained suspended sediment (Miall, 1983). Such glaciomarine conditions can not, obviously, be compared to those of any continental European glacial lake. In many features the massive clay resembles the Sunnybrook Till, a pebbly mud with a thickness of up to a few tens of metres, found in the Pleistocene glaciolacustrine deposits of Lake Ontario in Canada. The Sunnybrook Till consists of clays and silts with a high concentration of dropstones composed of indurated rock or diamicton. The name "till" comes from an earlier interpretation of this deposit as of subglacial origin (Hicock and Dreimanis, 1992). N. Eyles *et al*. (2005) proposed recently, however, that the unit was formed by sedimentation in a proglacial lake reached by floating ice rafts that must have supplied the material that was deposited as dropstones and diamictons. They found a cold water ostracode fauna, which may have contributed to the destruction through bioturbation of a possible original lamination; such as interpretation may also explain the presence of organic matter in similar "cold" glaciolacustrine deposits.

Sedimentary facies D. The thick rhythmites of facies D, present in the northwestern part of the lake, are most probably a proximal equivalent of the varves of lithofacies A. Their distinct rhythmicity with clay layers separating the thicker and coarser-grained deposits, may be climate-induced: an annual rhythm cannot be excluded. The fairly constant thickness of the clay layers (Fig. 10) indicates a continuous supply of material to this part of the lake during the winter. The clay suspended in the lake water settled when the water became quiet during the winter. Coarse-grained material was supplied, probably in a few phases, during spring and summer. The clay and diamicton clasts in the silty matrix of the massive unit (Fig. 10) point to low-concentration mass flows such as turbidity currents as the transport medium. The clasts are commonly dispersed in the lower part of normally graded layers, which indicates turbulence during sedimentation. When they are concentrated in horizons in the upper or middle parts of the silt layers (Fig. 11A), a more dense transport medium is likely. The small size of the clasts and their lithology indicate, however, that they could easily have been transported by a flow with even slightly developed cohesion. The silty matrix points to a mudflow. The existence of gravity flows suggests (Gruszka and Zieliński, 1996; Gruszka, 2001) transport and deposition on an inclined surface of the lake basin, although the inclination need not have been significant.

The abundant deformation structures could have been caused indirectly by a high accumulation rate. The majority of the deformation structures developed due to loading induced by a reversed density gradient (Anketell *et al*., 1970). The presence of boudins suggests slow laminar flow of the sediment with plastic folding. This is to be expected on the slope of a lake margin. The presence of normal faults (Fig. 11B) with various dip directions suggests the melting of buried dead-ice blocks. The displacements are small, though may have occured in more then one phase. Pillow structures are present; such structures have been interpreted as a result of seismic shocks (Roep and Everts, 1992; Rodriguez-Pascua *et al*., 2000), although these structures can also result from just a reversed density gradient (Visser *et al*., 1984), which is more likely here.

MASSIVE CLAYS IN GLACIOLACUSTRINE SUCCESSIONS — A DISCUSSION

Massive clays are usually interpreted as a facies resulting from settling of homogenous fine-grained particles in a standing water body. They are frequently found in glaciolacustrine deposits as layers up to a few centimetres in thickness, being the upper parts of varves. These varves represent distinct rhythmic seasonal changes in annual cycles, which result in fluctuations of the amount and type of material supplied to the lakes. As such annual cycles are common, varves are frequently found. Thicker massive clays are, however, only rarely described from glacial lakes.

Massive clays from a glaciolacustrine environment have been interpreted in a variety of ways. According to Smith (1978) and Sturm and Matter (1978) they are formed in the most distal parts of a lake. More frequently, the massive clays are interpreted as having been deposited in a shallow lake with greatly or even entirely reduced bottom-current activity (Gilbert and Desloges, 1987; Desloges, 1994), where suspension settling dominates, and where the amount of material supplied to the lake has perceptibly diminished, for instance because of increased distance to the source area (Desloges, 1994), or because the lake had become disconnected from its source area.

Alternative interpretations exist. Contrasting with the hypothesis of reduced or absent bottom currents for the genesis of the massive clays, one may invoke a shallow lake with intense wave and current action. In combination with small-scale slumping, these factors can destroy lamination (Mathews, 1956). Owen (1996) links the massive structure of fine-grained deposits to such a high accumulation rate (particularly for mountain lakes) that seasonal influence on the lake sediments becomes of little importance. McCabe and O'Cofaigh (1995) ascribe the origin of massive clays to settling from suspended plumes, sometimes with weak currents that are responsible for locally present, though indistinct, lamination, as in the case of the Gniew clays.

These interpretations contrast sharply. None of them relate the massive clays directly to a climatic factor. However, a temperature rise may result in the absence of an ice cover on the lake during the winter, which would also result in a massive character of the fine-grained deposits (Hart, 1992). The warming might also be responsible for a decrease in sediment supply if the ice sheet, being the source of the material, would retreat far enough (Lemmen *et al*., 1987). A burrowing fauna can also disturb lamination. Larsen *et al*. (1998) carried out research on the dependence of lamination preservation on climate, lake depth and morphology. Although they discovered water chemistry to be an important factor, the data were unfortunately not sufficiently clear to provide an reliable model.

Apart from the Gniew clays, the thickest bed of massive silts and clays found so far was a unit in glaciolacustrine deposits in the Kleszczów Graben, Central Poland (Brodzikowski and Zieliński, 1992; Van Loon *et al*., 1995). Its origin is interpreted as reflecting tectonically induced resuspension. In the Lagen valley in Central Norway, a 1 m thick layer of massive silt intercalated between laminated deposits was found. Its genesis was also interpreted as resuspension, caused by rapid melting of buried ice blocks under the lake bottom (Berthling *et al*., 1999). These examples indicate the possibility of a massive clay resulting from resuspension or another type of redeposition. Such a possibility should not be excluded for the Gniew clays.

An interesting idea was put forward by Van der Meer and Warren (1997) after analysis of thin sections of fine-grained sediments (massive silts and clays) of a glacial lake. They question the correctness of macroscopic observations of structures in fine-grained sediments. In most samples examined by them, the "massive" deposit turned out to be laminated, although this lamination is invisible to the naked eye. In consequence, such deposits most often may be better interpreted as deposited from suspension in daily or seasonal cycles typical for each lake.

SEDIMENTARY CONDITIONS IN LAKE GNIEW

The characteristics of the Gniew clays indicate deposition in a glaciomarginal/terminoglacial lake. The geomorphological context of the lake area is complex, however, and does not fully suit published models of terminoglacial lakes (Brodzikowski and Van Loon, 1991; Brodzikowski, 1993). At least during the initial phase of the lake development, it must have resembled the "ice-stagnation lake network" (Ashley, 1988) of ice-contact lakes or a glaciomarginal ice-dammed lake formed over the ice bottom (Shaw and Archer, 1979). Inevitably an ice barrier must have dammed off the lake in the south, as large dead-ice blocks will have been present there, scattered over the lower reaches of the Wierzyca valley. This barrier is now expressed morphologically as a steep escarpment (Figs. 3 and 4). The huge dead-ice mass blocking the river was not the only one: the previous existence of another one is now morphologically visible in the form of the Ciepłe kettle-hole. Ice blocks existed also over a large area underneath the Gniew Lake sediments. The thickest ice masses are now reflected by the Brodzkie Młyny–Szprudowo trough, and by other vast depressions under the glaciolacustrine succession. The melting of buried dead-ice masses continued for some time, causing syndepositional subsidence of the lake bottom. Nevertheless, the main phase of dead-ice melting occurred during the Allerød (Błaszkiewicz, 2003).

In addition to the existence of dead-ice blocks, ice-rafted deposits are also indicators of an ice cap near Lake Gniew. Small dropstones do not necessarily indicate a nearby ice cap, but large diamicton clasts do. All these data point to the nearby presence of sediment-laden ice, though not necessary active ice. Contact with active ice presumably only existed during the initial phase of lake development. Diamicton intercalations in only the lower varve series of facies A demonstrates this. Shortly after the facies shifted so that the lake became in a more distal position, pure silt/clay varves accumulated. The deposits from the northwestern part of the lake (facies D) contain diamicton clasts in beds that are interpreted as mudflow deposits. This also suggests a short distance to the ice sheet, but no direct contact.

The first stage of the lake development is represented by facies A. While the sediments of facies A were deposited, the distance to the ice increased, presumably because the ice started to retreat. The basin must then have been deep enough to create thermal stratification. The somewhat less fine-grained lower parts of the varves originated from weak bottom currents, whereas the upper clay parts developed when ice covered the lake surface and the finest clay particles, kept in suspension during the summer, could settle under conditions without a thermocline (Sturm and Matter, 1978; Sturm, 1979). The number of varves indicates that this sedimentation interval was not longer than 25 years.

Although there is no direct evidence, the deposits of facies D are most probably the equivalent of facies A in the northern part of the lake, though the thicknesses of these two successions differ greatly. The number of rhythmite units is, however, more or less the same. The high accumulation rate of facies D, if compared to facies A, points to a much larger sediment supply, which was also responsible for the formation of a sandy delta in the north. Facies D thus represents prodeltaic deposits, an interpretation supported by the sedimentological evidence of a gentle slope.

The steady glaciolacustrine rhythmic sedimentation finished when a sandy turbidity current spread all over the lake bottom. The turbidite shows proximal characteristics (the massive lower part of the Bouma sequence is always present) at all sites studied. The turbidity current must have originated due to changes in the environmental conditions in the lake, perhaps reflecting fluctuations in the water level, possibly shallowing. Such shallowing could be the result of partial drainage of the lake (Gilbert and Desloges, 1987). This was accompanied by an increase in melting, as numerous dropstones are present above the sand layer.

The interpretation of the facies C remains uncertain. Although massive clays have been the subject of extensive research, no true equivalent of facies C has been found in the literature. We discuss below two options for its genesis. Both are plausible but neither is fully supported by the evidence:

— facies C is the result of sedimentation in a shallow lake with high sediment input comprising homogenously suspended material consisting almost exclusively of clay particles;

— facies C is the result of a huge muddy jökulhlaup, which carried, to Lake Gniew, water with resuspended clay deposits from a higher glacial lake.

The first interpretation explains the massive character of the deposits well. The lake was without thermal stratification, the activity of bottom currents was extremely weak or nil. Occasionally wave action destroyed any lamination that might have been present. A high sediment input must be assumed for several reasons. The presence of the steep escarpment marking the southern boundary of the dead-ice block that dammed off the river, thus forcing the lake to develop, points to a relatively short-lived lake, because the border would otherwise not have been so pronounced in its morphology. Gilbert and Desloges (1987) described a modern self-draining shallow lake, where sedimentation of massive silts and clays is less than 1 mm per year due to a low sediment input. With the same rate of sedimentation, it would have taken 20 000 years to fill Lake Gniew, but so much time was not available. This implies that the sediment input must have been high. Although the lake was shallow, syndepositional subsidence due to melting of the buried dead-ice underneath could have made it possible for a thick clay unit to accumulate.

This interpretation has, however, some drawbacks. The homogeneous character of the supplied clay contrasts with the huge ice-rafted diamicton clasts. The clay supplied indicates that the lake became disconnected from the ice sheet and that the environment became more distal, whereas the diamictons of the dump structures point to a proximal environment with a large number of debris-laden icebergs in the lake. Why would the sediment input into the lake be so steady for many years, and consist of almost pure clay, while large quantities of coarse-grained debris were deposited in the immediate surroundings?

The second interpretation adequately explains the coexistence of homogenous fine-grained deposits and large diamicton clasts. It needs, however, the presence of another lake, located at a higher altitude (probably northwards), possibly supraglacially, and filled with clay-rich deposits. This lake may have drained suddenly, for instance as a result of the undermelting of an ice dam, causing a catastrophic flood, which could have had the character of a jökulhlaup. The shock caused by a collapse due to undermelting and the sudden strong and turbulent water flow must have caused the resuspension of a significant part of the clayey bottom sediment. The water/sediment mixture flowed into Lake Gniew, which acted as a trap for much of the fine-grained sediment, although part of the clay-laden water of the jökulhlaup may have overswept the lake margins. The rapid melting of dead-ice underneath the lake and decoupling phenomena adequately explain the presence of large icebergs with debris. A large flood caused by a lake drainage event has been reconstructed in detail for Lake Agassiz (Teller *et al*., 2002). Recently three beds of massive silty clay have been interpreted as distal deposits of a huge outburst flood caused by lake drainage, but their thicknesses do not exceed a few tens of centimetres (Blais-Stevens *et al*., 2003). The large thickness of the Lake Gniew massive clay indicates the outstanding scale of this flood.

No traces of an upper drained lake have been found, however, and neither are there any traces of an input zone. The glacial environment may, however, have experienced many catastrophic phenomena during deglaciation, including jökulhlaups like the one presumed above, without leaving any traces that can still be found.

Most probably the end of Lake Gniew came rapidly. This happened when the stream into the direction of the lower Vistula reaches was no longer blocked by an ice mass. After the lake had drained, a long break in glaciolacustrine sedimentation occurred, up to the beginning of the Allerød. Then the next phase of melting of dead-ice blocks took place. Bogs and lakes with organic sediments developed in morphological lows in the area of Lake Gniew. These depressions were completely filled with organic deposits during the end of the Pleistocene and the Holocene. This phase of lacustrine sedimentation was limited to the areas of dead-ice melting and has been well documented for the Ciepłe kettle-hole (Błaszkiewicz, 2003). ^{14}C dates show that the break between the end of glaciolacustrine sedimentation and the beginning of lacustrine organic-rich sedimentation lasted at least 4 000 years. Similar examples of lacustrine deposits on top of a glaciolacustrine succession are known from north Germany (Kaiser, 2001; Krienke, 2002).

CONCLUSIONS

Lake Gniew, one of the glacial lakes known from the area in Poland that was reached by the Vistulian Glaciation, had a limited size. It existed for a relatively short time, as is typical for Vistulian glacial lakes in Central Europe (Schirrmeister, 1997; Kaiser, 2001; Krienke, 2002; Niewiarowski, 2003). The duration of one of the longest-lived proglacial lakes from the Vistulian was recently calculated (by varve series) to be not much more than 385 years (Paluszkiewicz, 2004). Lake Gniew differed, however, from typical glacial lakes of the same age in northern Poland: its glaciolacustrine filling up is not comparable to that of any other lake.

The reconstruction of the sedimentary conditions, the main objective of this study, has not been entirely achieved. Although four sedimentary facies — A, B, C and D — have been distinguished, only three of these can be reliably interpreted sedimentologically, with details of the sedimentary processes and a fair indication of the lake environment during the successive sedimentation phases.

Facies A, consisting of silt/clay rhythmites being varves, was deposited in the distal part of the lake during the first stage if its development. Facies B, a sand layer with a complete Bouma sequence, represents a large turbidity current that extended through a large part of the lake. Facies D, composed of thick rhythmites, accumulated in a prodelta environment. The origin of facies C, a massive clay, is still enigmatic, although this sediment constitutes the majority of the glaciolacustrine fill. Two options for its development are presented; we cannot decide which is the more likely. The first option explains the massive character of the deposits as the result of rapid sediment input in a shallow basin, whereas the second option involves a catastrophic flood of jökulhlaup type to explain the sedimentary characteristics.

The problem of facies C might be solved by the application of more sophisticated research methods, a macroscopic analysis according to standard geological procedures is not sufficient to unravel the origin of these apparently massive clays.

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