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Varve sedimentation in extraglacial ice-dammed lakes

ABSTRACT: Pleistocene extraglacial varved clays exposed in Central and northern Poland are deposits of long-time suspension currents. Their variation in the average varve thickness and the origin of varve-series is attributed to changes in the width of the depositional area in extraglacial ice-dammed lakes. The developmental sequence of sedimentary conditions leading from an extraglacial river valley to an ice-dammed lake is proposed and illustrated with examples from the investigated sections. A sharp contact between varved clays and the underlying cross-bedded sands reflects a sedimentary gap caused by the varve sedimentation encroaching upon a river terrace.

INTRODUCTION

Pleistocene extraglacial ice-dammed lakes were formed by ice-barring of a river valley and damming up the river flowing toward the ice-dam (MERTA 1978). In such a basin, three main sedimentary zones are to be distinguished: proximal, intermediate, and distal (*see* Text-fig. 14 *in* MERTA 1978).

The first of these zones is characterized by the so-called *A*-type varve sedimentation, with rather indistinct light and dark layers and the total thickness of several centimeters each varve. It is situated near the extraglacial river mouth.

The second zone bears the *B*-type varve sedimentation, with contrasting light and dark layers and sharp boundaries between individual varves. This zone is located further toward the center of the extraglacial ice-dammed lake.

The distal zone contains very thin, so-called distal, microvarves of the *C*-type. It is situated furthest away of the mouth of the extraglacial river, generally corresponding to deeper parts of the ice-dammed lake.

More detailed characteristics of all these varve types and sedimentary zones were given in an earlier publication (*vide* MERTA 1978, pp. 248 and 263, Pl. 12, and Text-fig. 14).

The similarity of varves in both extra- and proglacial deposits probably reflects the same causal mechanism, *viz.* climatic biseasonality with effective sediment supply in summer time and with stagnation in winter. The sediment supply in proglacial lakes, however, depends on the single-maximum process of icesheet ablation (DE GEER 1912, HALICKI 1932, ANTEVS 1951), whereas the rhythm of extraglacial rivers shows a few optima per year. Perhaps even more importantly, extraglacial ice-dammed lakes are rather shallow and elongated thus contrasting with huge and generally deeper proglacial basins (*see* HANSEN 1940).

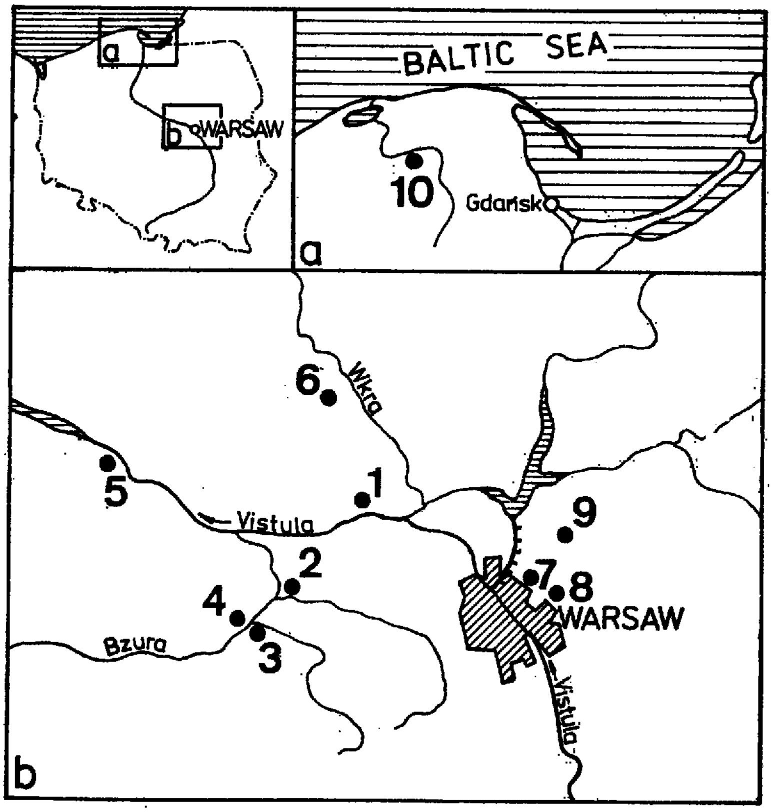


Fig. 1. Location map of varved sequences in Poland, selected for illustration of the development of an extraglacial ice-dammed lake; **a** — Lębork in northern Poland, **b** — Mazovia Lowland in Central Poland

Localities: *1* Mochty, *2* Plecewice, *3* Boryszew, *4* Kuznocin, *5* Góry, *6* Arcelin, *7* Marki, *8* Zielonka, *9* Radzymin, *10* Lębork

The present paper primarily concerns the sedimentology of Pleistocene varved deposits in the Mazovia Lowland in Central Poland (MERTA 1975, 1977, 1978, 1980). Examples from this area make up the basis for development of an extraglacial ice-dammed lake model; one example comes from a huge exposure of varved clays at Lębork in northern Poland (Text-fig. 1).

VARVE ORIGIN AND LONG-TIME TURBIDITY CURRENTS

The problem of propagation and distribution of suspended matter supplied into a proglacial basin was considered by many authors.

ANTEVS (1951) made observations on numerous present-day lakes and assumed, on this basis, that proglacial lakes had no thermal stratification in summer at all. In his view, water temperature approximated 4°C in the entire water column, from the surface down to the bottom. Hence, the lighter meltwater should have been propagated in the basin by superficial currents. A similar opinion on water temperatures had been previously expressed by DE GEER (1912), who also claimed, however, that the meltwater carrying the suspended matter must have been heavier than the water in the basin and, hence, should have been propagated by near-bottom currents.

Having considered the turbidity currents, KUENEN (1951) suggested that varves were deposited by the so-called long-time suspension currents. This interpretation of varve origin has successively gained wide acceptance for the varved clays of various geological ages (BANERJEE 1966, HARRISON 1975) as well as for the present-day varvelike deposits (BELL 1942, GOULD 1951, KLIMEK 1972). The varve, however, is not an equivalent of the turbidite (*sensu* BOUMA 1962). It originates from near-bottom clouds of suspended matter which continually supply the sedimentary material over several months, while the turbidite is a portion of material deposited during a very short period (minutes, hours, or perhaps several days but no more).

The present author (MERTA 1978) pointed to long-time turbidity currents as the cause of the varvity of Pleistocene extraglacial clays. According to this concept, the propagation of suspension currents into an extraglacial ice-dammed lake was determined by the lowermost part of the overflooded valley, *i. e.*, by the river channel area.

VARIATION IN VARVE SEQUENCES

In the investigated extraglacial varve sequences, the varve thickness is highly variable (r in Text-fig. 2). The average thickness of varves decreases upwards in the sequence, as a rule. Most frequently, it changes gradually (*e. g.*, Plecewice, Kuznocin, Radzymin, Marki, Zielonka — *see* v in Text-fig. 2, Pl. 1, Fig. 1), but rapid changes also occur here and there, in sections with varve-series (Text-fig. 3). These changes are always unidirectional, a series of thicker varves being covered by a series of thinner ones (Pl. 1, Figs 2-3).

The thickness of individual varves has been generally treated as a function of (*i*) the total quantity of sedimentary material supplied per

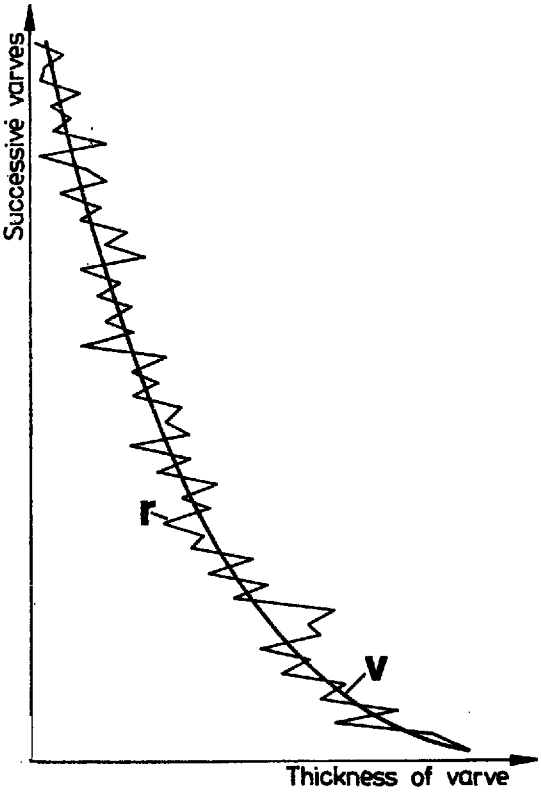


Fig. 2

Schematic varvdiagram of extraglacial varved clays
r — real thickness of varves,
v — average thickness of varves

year, and *(ii)* the position of the sedimentary area, *i. e.*, its distance from the source area. For proglacial varves, this distance is determined by the position of the icefront; for extraglacial varves it is determined by the position of the river mouth. Therefore, long-time changes in the average thickness of varves (or varve-series) have been attributed to oscillations of the icefront (SAURAMO 1929, HANSEN 1940, ANTEVS 1951, PIRRUS 1965, RINGBERG 1971). Forward and backward movements of the icefront resulted in alternating sedimentation of varve-series with thick varves and varve-series with thin ones. The variation in thickness of individual varves in proglacial lakes can, in turn, be explained by fluctuations of the ablation intensity in succesive years (DE GEER 1912, LINDBERG *in* SAURAMO 1929; *see also* HALICKI 1932).

The varve variation in extraglacial varve sequences, however, cannot be related to ablation processes or to changes of the icefront position, because the material was being primarily brought from the opposite direction. VAN SICLEN (1964) explained the changes in thickness of individual laminae by vertical oscillations of the water level and consequent changes in the extent of the basin. A simpler explanation might

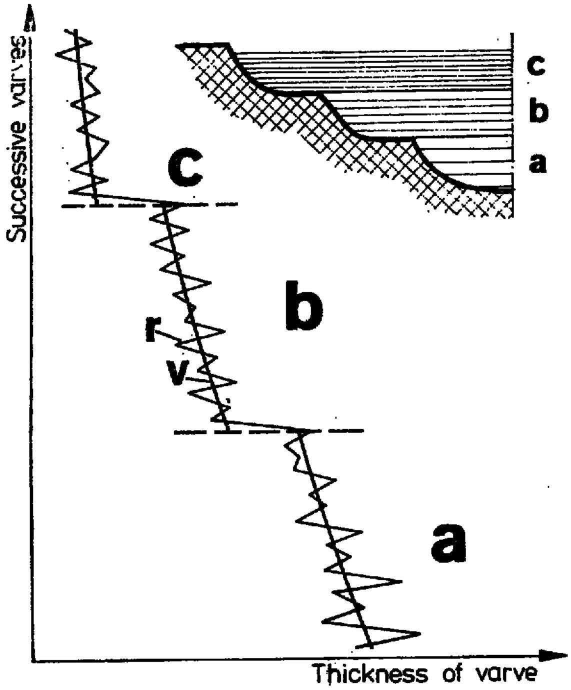
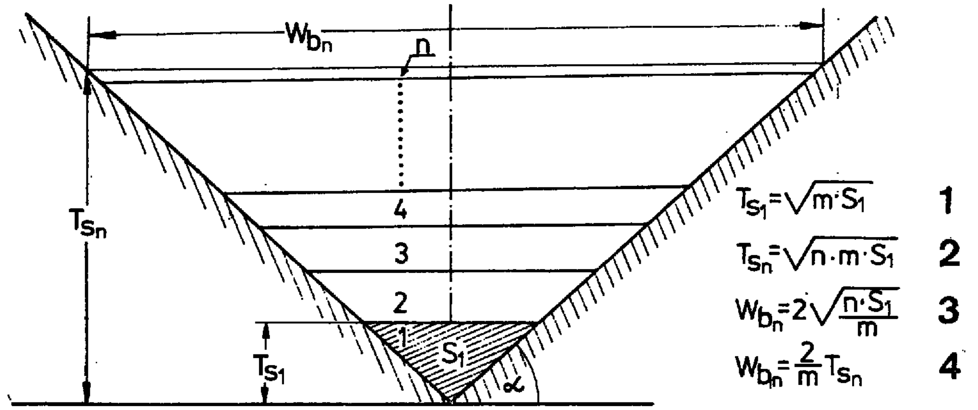


Fig. 3

Schematic diagram of a varve-series sequence and its hypothetic relation to the river-valley morphology
r — real thickness of succesive varves
v — average thickness of varves;
a, b, c — succesive varve-series (*compare* Text-fig. 8)

refer to variation in the supply of sedimentary material. Neither mechanism, however, adequately accounts for the origin of varve-series.

Extraglacial ice-dammed lakes were located in river valleys. Hence, they valley width and longitudinal profile of the valley and the angles of its slopes must have determined the width, depth, and areal extent of an extraglacial basin.



$$T_{S_1} = \sqrt{m \cdot S_1} \qquad 1$$

$$T_{S_n} = \sqrt{n \cdot m \cdot S_1} \qquad 2$$

$$W_{b_n} = 2\sqrt{\frac{n \cdot S_1}{m}} \qquad 3$$

$$W_{b_n} = \frac{2}{m} T_{S_n} \qquad 4$$

Fig. 4. Theoretical cross-section of a V-shaped river valley
α — angle of slope ($\tan \alpha = m$); 1, 2, 3... n — number of depositional acts; S_1 — cross-section area of the initial varve (numbered 1); T_{S1} — thickness of the initial varve (equation *1*); T_{Sn} — thickness of the varve sequence (equation *2*); W_{bn} — width of the bottom of the basin after *n-th* act of deposition (equations *3* and *4*)

Let us consider an extraglacial lake in *V*-shaped valley with the same angle α ($\tan \alpha = m$) of both its slopes. The first portion of deposited material (initial varve) has the cross-sectional area S_1 and thickness T_1 (Text-fig. 4). The relationship between S_1, m, and T_1 is given by the equation *1* (in Text-fig. 4). If the material is supplied in constant portions, the thickness of a varve sequence containing *n varves* will be T_{sn} (equation *2* in Text-fig. 4). At the same time, however, the width of the depositional area gradually increases from W_b = O to W_{bn} (equations *3* or *4* in Text-fig. 4). Thus, the rate of decrease in varve thickness depends upon m, or the cross-section of the river valley (Text-fig. 5). The decreasing trend of the average thickness of individual varves upwards in the sequence thus reflects an increase in depositional area rather than a decrease in supply of the material.

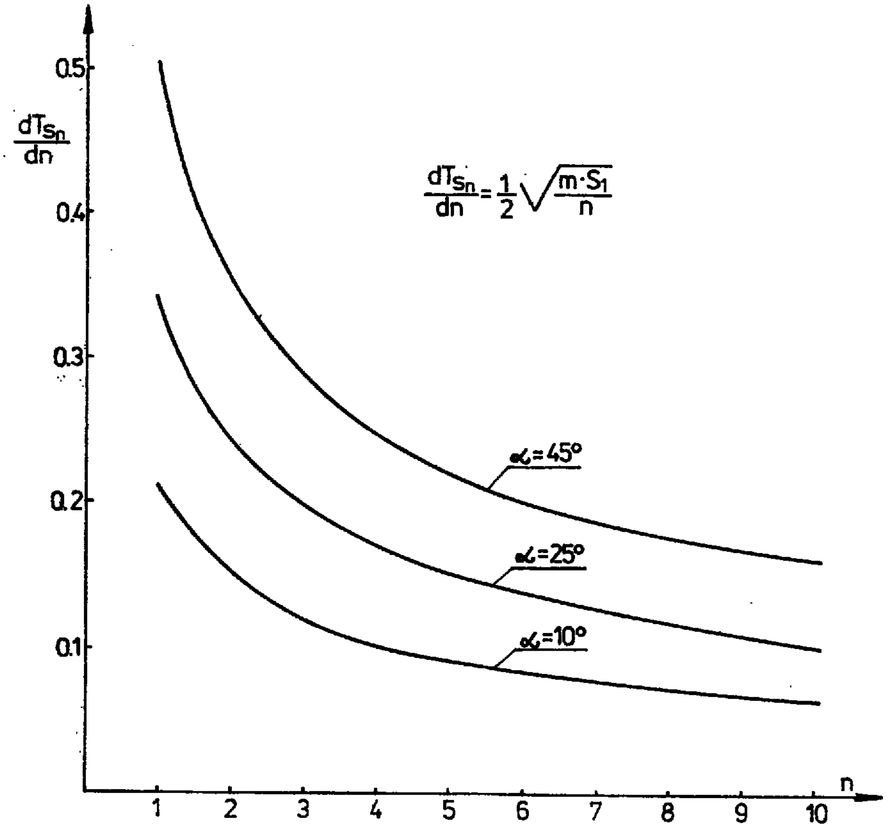
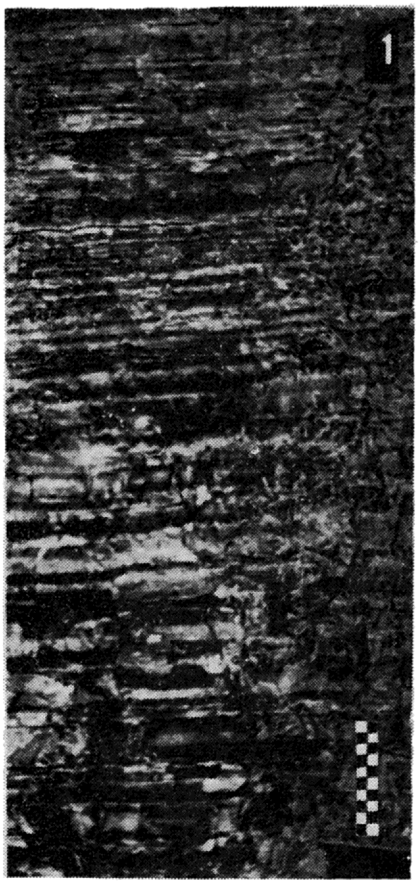
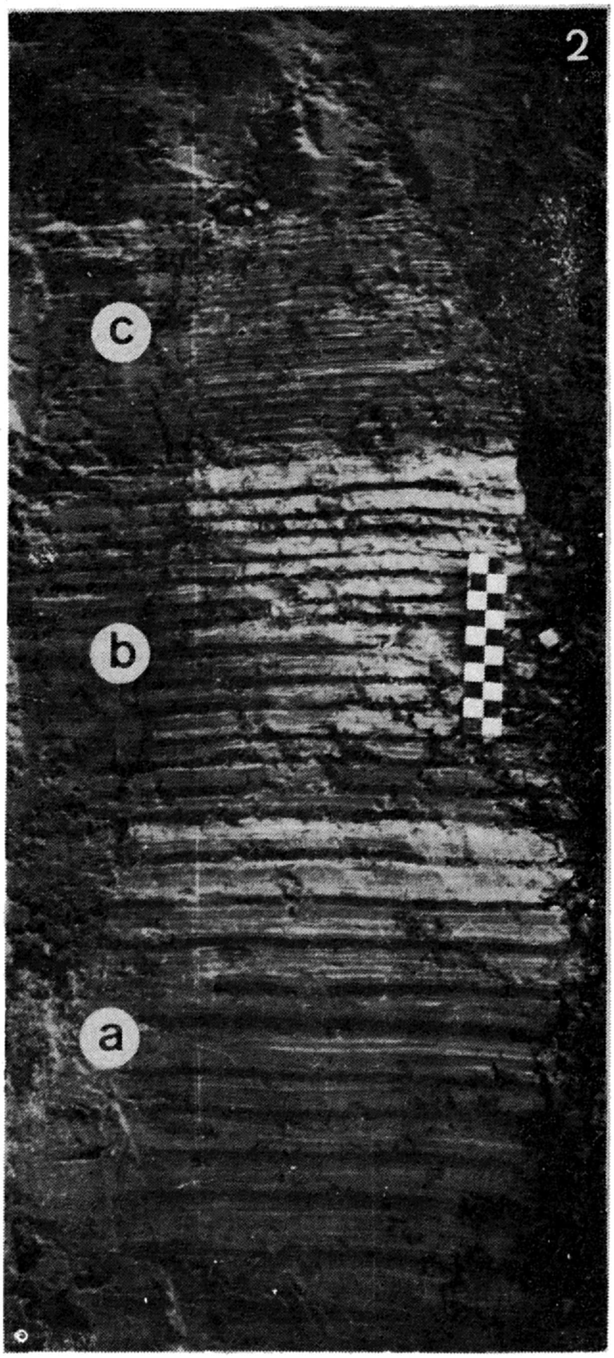
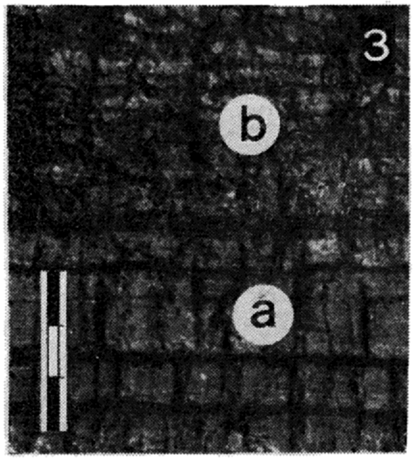


Fig. 5. Decrease in thickness of successive varves, depending on the valley-slope angle; the quantity of sediment is held constant
S_1 — cross-section area of the initial varve: value 1.0 in dimensionless unit; m = tan α; 1, 2, 3... n — number of the varve deposited after the initial one; T_{sn} — thickness of the sequence after *n* acts of deposition: the initial varve thickness is 1.0 (dimensionless unit)



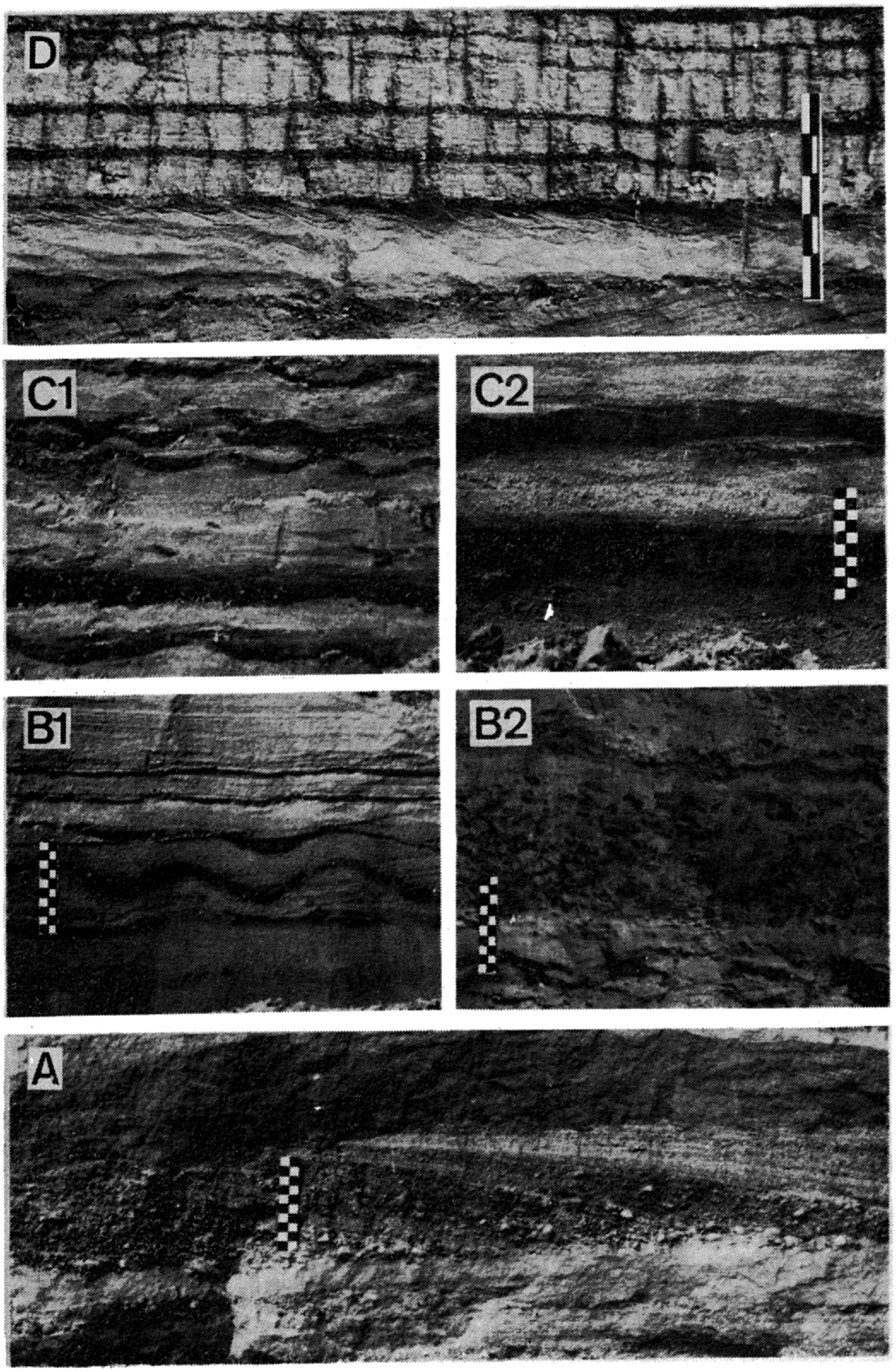




1 — Sequence of varved clays showing a gradual change in the average varve thickness; locality Marki, scale in cm

2 — Varve-series sequence: series *a*, *b*, *c* differ in the average varve thickness; locality Arcelin, scale in cm

3 — Boundary between two varve-series (*a* and *b*); locality Mochty, scale in dcm


D
C1
C2
B1
B2
A

This conclusion leads to a very simple interpretation of the origin of varve-series. The proportionality of the total thickness T_{sn} of an n-varves-bearing sequence to the width of its depositional area W_{bn} (*see* equation 4 in Text-fig. 4) indicates the possibility of a rapid change in varve thickness as a result of rapid change in width of the depositional area. For example, let us consider an overflooded valley with width of the bottom W_1 and thickness of each varve T_1 (*a* in Text-fig. 6). Hence,

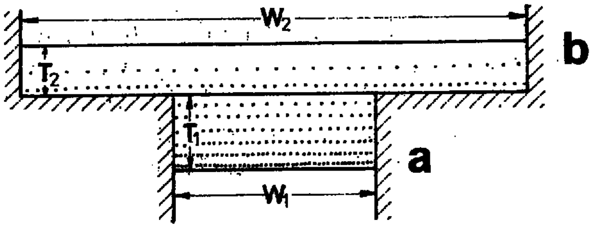


Fig. 6. Varve sedimentation within a terrace-shaped river valley
a — conditions determined by width (W_1) of the bottom and thickness (T_1) of the varve, *b* — conditions determined by a sudden change in width of the bottom (to W_2) and the resulting decrease of in varve thickness (to T_2)

the cross-sectional area of a single varve is $S_1 = W_1 \cdot T_1$. When the varves fill up the lower part of the valley and the water floods the next higher terrace, the width of the basin will rapidly change from W_1 to W_2 and, consequently, the varve thickness will also change from T_1 to T_2. The cross-sectional area of a single varve will then be $S_2 = W_2 \cdot T_2$ (*b* in Text-fig. 6). If the total quantity of the supplied material remains constant, $S_1 = S_2$, then $W_1 \cdot T_1 = W_2 \cdot T_2$. An increase in basin width must be accompanied by a decrease in varve thickness. This appears as a more plausible interpretation of varve-series origin than any allochthonous causes, that is, a rapid drop in the total quantity of sediment supply or a change in the river-mouth position. Generally, then, varve-series sequence with progressively thinner varves (*e. g.*, at Arcelin; *see* Pl. 1, Fig. 2) seem to have been deposited within terrace-shaped valleys (Text-fig. 3).

This interpretation is appropriate for extraglacial ice-dammed lakes only, while it may not be correct for proglacial lakes.

PLATE 2

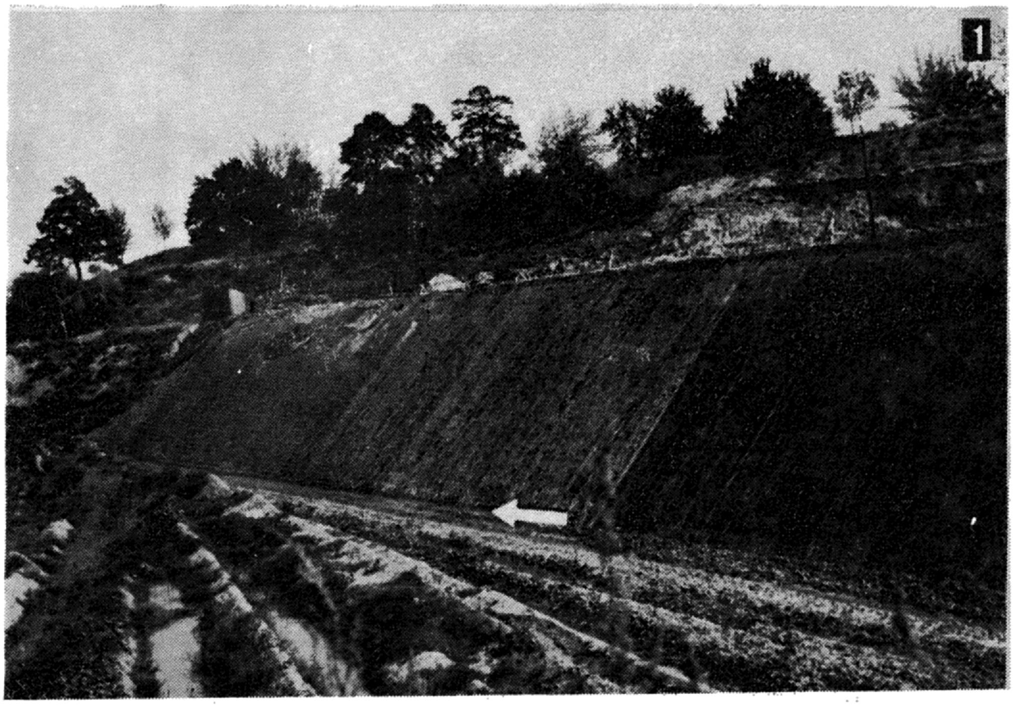
A — Cross-bedded sands and gravels underlying a varved sequence; Boryszew, scale in cm
B1 — Sandy ripples covered by clay laminae; Mochty, scale in cm
B2 — Layer of breccia consisting of redeposited mud pieces; Mochty, scale in cm
C1 — Continuous and relic clay laminae covering sandy ripples of current origin; Mochty, scale in cm
C2 — Alternation of sandy and clayey layers giving a varvelike appearance to the sediments; Zielonka, scale in cm
D — Contact between climbing-rippled sands and varved deposits; Mochty, scale in dcm

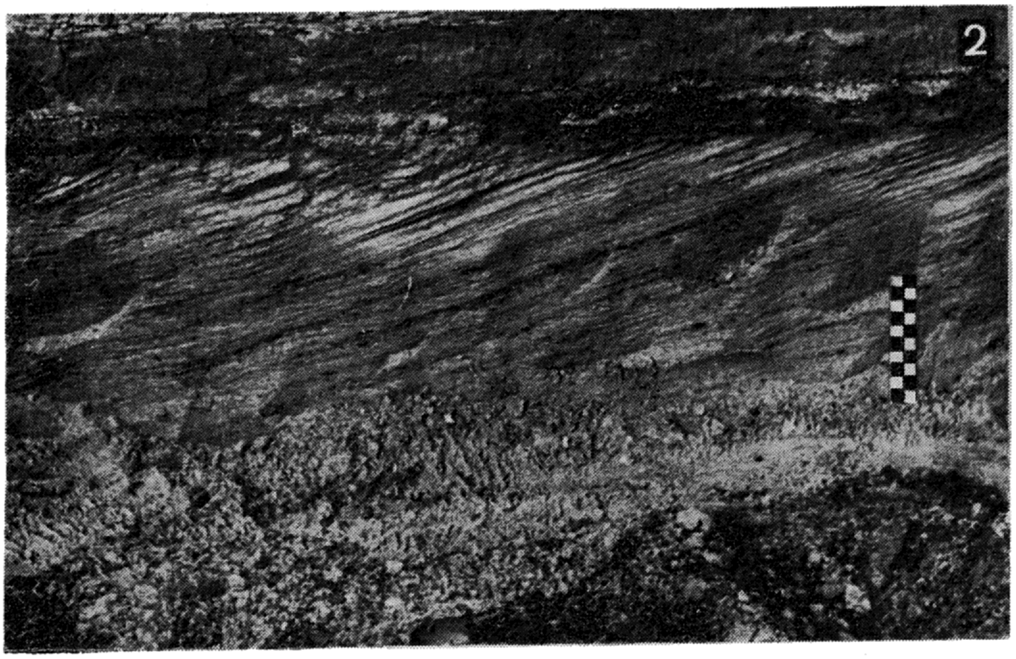
VARVED CLAYS AND THEIR RELATION TO THE UNDERLYING DEPOSITS

Sections with a gradual transition from sands to typical varved clays are scarce (Boryszew, Mochty; *see* Text-fig. 1). They illustrate a change from fluvial to lacustrine sedimentary conditions. The fluviatile phase of sedimentation is represented by tabular sets of cross-bedded sands and/or gravels (*e. g.*, at Boryszew; *see* Pl. 2, Fig. A), which gradually pass upwards into sandy beds with numerous ripples of current origin. These ripples are usually covered by clayey laminae (*e. g.*, at Mochty; *see* Pl. 2, Fig. B*1*). Sometimes, these parts of the sequence contain also beds of breccias consisting of sand and redeposited clay pieces (*e. g.*, at Mochty; *see* Pl. 2, Fig. B*2*), derived from the eroded clayey covers. Higher up in the sequence, the thickness of sandy beds gradually decreases and ripplemarks become smaller (*e. g.*, at Mochty; *see* Pl. 2, Fig. C*1*). Finally, intercalations of continuous thin clayey laminae appear between sandy beds which are several centimeters thick (*e. g.*, at Zielonka; *see* Pl. 2, Fig. C*2*). These are not yet varves but they already consist of both light and dark layers, as the proper varves do. They represent different sedimentary regimes alternating in the environment, *viz.* the dynamic-transport regime (sandy light layers) and the stagnant one (clayey dark laminae). These are the initial conditions of an ice-dammed lake, although there is still a considerable influence of the fluvial regime, *i. e.*, the material is chiefly propagated by the river currents. Such strata gradually pass upwards into typical varves, deposited from clouds of long-time suspension currents.

More commonly, the succession from sands/gravels to varves bears a very sharp contact (*e. g.*, at Góry, *see* Pl. 3, Figs 1—2; and at Lębork, *see* Pl. 4, Figs 1-2). The assemblage of sedimentary structures in the underlying sands (various cross-beddings; *see* Pl. 3, Fig. 2 and Pl. 4, Fig. 2) unmistakably defines the upper flow regime conditions of a fluvial environment. Hence, such a succession reflects a rapid change of sedimentary conditions from the dynamic-river regime (sands and gravels) to the stagnant-basin regime (varved clays). The investigated sequences lack any direct traces of such a change. The absence of a transitional interval seems to indicate the presence of a sedimentary gap between sands and varved clays.

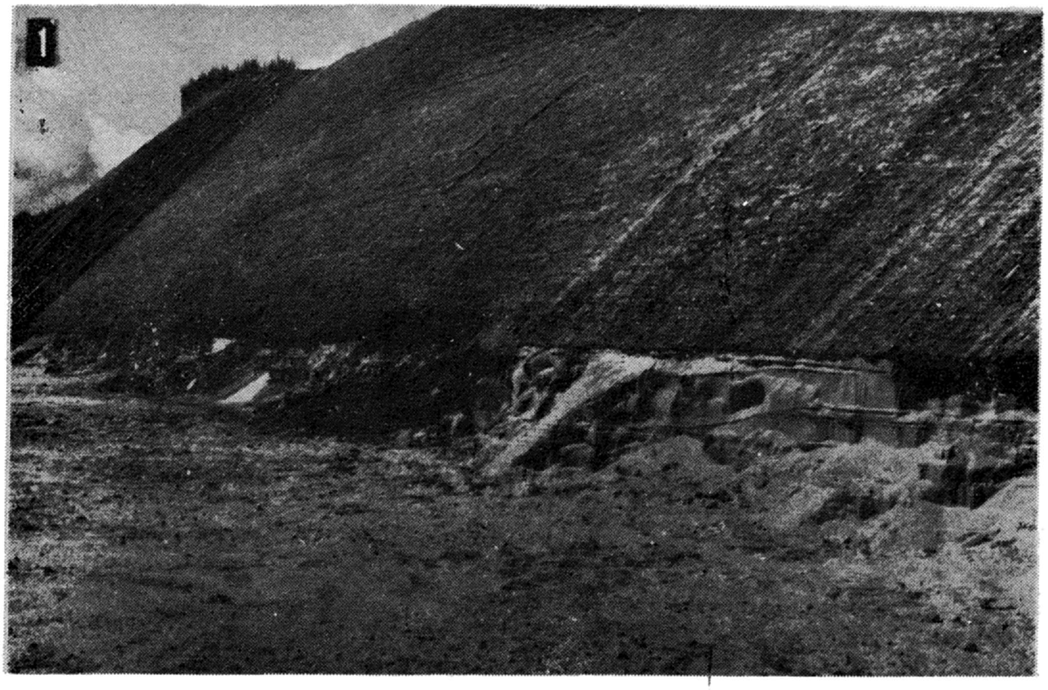
In extraglacial ice-dammed lake environments, the sedimentary material was distributed by near-bottom clouds of long-time suspension currents. At the initial phase of basin life, these currents propagated along the river-channel zone of the overflooded valley. Therefore, a sequence with gradual transition from alluvial deposits of the pre-basin stage to varved clays can only be expected to occur in this narrow zone. In the remainder of the basin, the overflooded terraces including, no

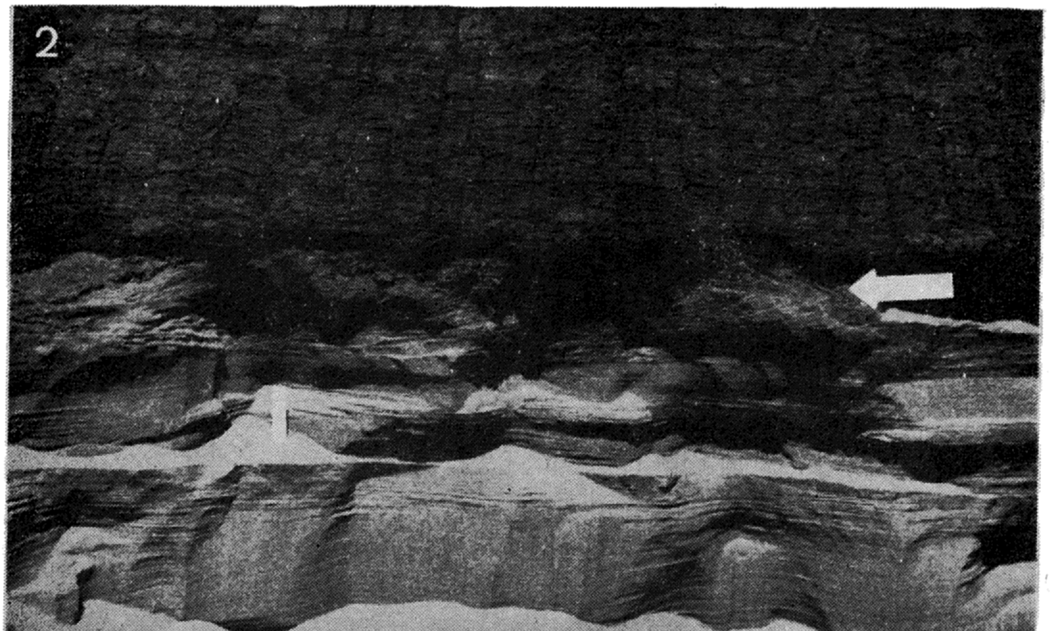




1 — General view of a varved-clay exposure at Góry; arrow indicates a sharp contact between varved clays and sands

2 — Close-up view of the contact zone at Góry: *note* cross-bedding in the sands; scale in cm





1 — Distinct sharp contact between sands and overlying varved clays at Lębork
2 — Close-up view of the contact zone at Lębork: *note* cross-bedding in the uppermost part of sands *(arrowed);* white bar is 10.0 cm long

deposition occurred at that time. The varved-clay deposition began there only after the river valley had been filled up with sediments. The flooding of each successive terrace made the depositional area wider, thus allowing for a lateral propagation of suspension currents.

The sandy deposits underlying varved-clay sequences thus represent older stages of the river-valley life. They bear no clearcut causal relationship to the subsequent existence of an extraglacial ice-dammed lake.

EXTRAGLACIAL ICE-DAMMED LAKE DEVELOPMENT

Let us consider a fragment of an extraglacial river valley. Its successive developmental phases are here schematically illustrated (*see* Text-fig. 7A-D) as well as documented by real varved-clay sequences (*see* Pl. 2, Figs A-D).

In the first phase, the valley has acquired strongly developed accumulative-erosional terraces and a narrow river-channel zone of straight type, with sedimentation of coarse sediments (*see* Text-fig. 7*A* and Pl. 2, Fig. A). When the icefront encroaches upon the valley far away from the investigated area, the flow of the extraglacial river becomes more difficult. The straight flow system is transformed into a braided one (*see* RUST 1972). At this phase (Text-fig. 7*B*), two distinct sedimentary regimes can be discerned: the dynamic-transport regime is represented by cross-bedded sandy layers with current ripples, and the stagnant regime by clay covers (B1 *in* Text-fig. 7*B* and Pl. 2, Fig. B*1*). These clay covers can later be eroded at the next-dynamic transport stage, and clayey breccias are formed (B2 *in* Text-fig. 7*B* and Pl. 2, Fig. B*2*). This sequence of sedimentary conditions may be characteristic of the transition from a fluvial to an ice-dammed lake environment.

In the next developmental phase, the icesheet completely dams the valley up and the flow of the extraglacial river water is stopped. This is an initial stage of the extraglacial ice-dammed lake life (Text-fig. 7*C*). Even if the water level of the lake is high, the material supplied into the basin by the extraglacial river migrates, along the narrow river channel, as clouds of near-bottom suspension currents. Therefore, the actual deposition is limited to a narrow zone at the very bottom of the basin; it features, first, varvelike deposits (C1-C2 *in* Text-fig. 7*C* and Pl. 2, Fig. C) and then, typical varves. Generally, no sedimentation occurs in the remainder of the overflooded area. At most, it receives only a slow deposition of colloids and eolian dust (MERTA 1980); a small quantity of sediments may also be derived from destruction of the lake banks by wave action (*see* MERTA 1975).

After the river-channel has been filled up with deposits, the lowermost terrace also becomes the area of deposition. The onset of varved-clay sedimentation on this extensive area is marked by a rapid decline of the average varve thickness. This is the beginning of a new varve-series in the axial part of the basin (D" *in* Text-fig. 7*D*). In more peripheral parts of the basin, early stages of varved-clay sedimentation begin at this moment (D' *in* Text-fig. 7*D*). These varved clays must obviously be separated from the underlying sandy deposits by a sedimentary gap (Pl. 2, Fig. D; *see also* Text-fig. 8).

CONCLUSIONS

The existence of an extraglacial ice-dammed lake is controlled by the icedam stability. When the icedam retreats, rapid erosional processes take place. The previous system of the extraglacial river flow is restored (*see* JAHN 1968) and the deposits in the axial part of the basin undergo erosion. Sometimes, however, the new river flow misses the old channel. In such instances, sequences with the transitional deposits of the river/basin stage may be preserved, as it has been recorded, for example, in the sections of Mochty and Boryszew.

Extraglacial ice-dammed lakes could have repeatedly developed in the same area (MERTA 1978). Therefore, varved sequences deposited at the same hipsometric level may nevertheless differ in age. Such a possibility has never been taken into account and heterochronous varved-clay sequences were generally interpreted as facies variants of an isochronous sedimentary series (*see* SAMSONOWICZ 1922).

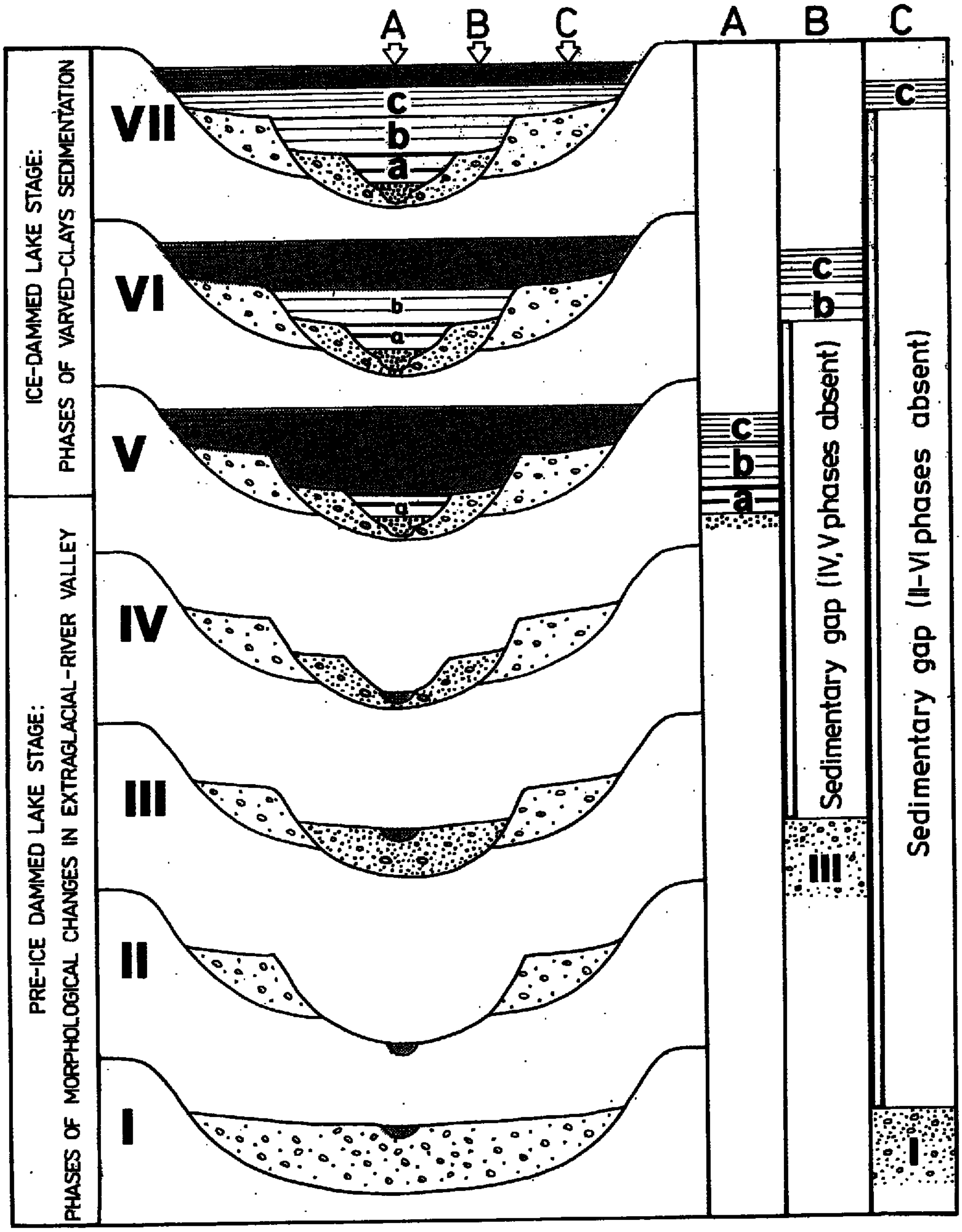
It has to be emphasized that the present interpretation of depositional history of ice-dammed lakes refers solely to extraglacial environments. It sheds some new light on the depositional mechanisms leading to a change in the average varve thickness and to origination of varve-series, the prime control being the change in areal extent of the sedimentary environment. This interpretation, however, does not exclude any annual variation in sediment supply. Consequently, the varvdiagrams of extraglacial varved sequences must be regarded as reflecting interactive effects of both these factors.

The significance of the size of the sedimentary area for the varve thickness affects also the present author's former opinion on the origin

Fig. 7

Developmental sequence leading a fluvial to an ice-dammed lake environment (detailed explanation in the text)

solid arrows — traction currents of the river type, empty arrows — near-bottom long-time suspension currents



Time-relations of a varved sequence to the underlying deposits, depending on the number of pre-lake phases and the location of the profile within the sedimentary area

I-IV — phases of pre-ice dammed lake stage, **V-VII** — phases of ice-dammed lake stage

A, **B**, **C** — position of profiles (*arrowed*) and their geological-time relations (*at right*)

a, **b**, **c** — varve-series (*compare* Text-fig. 3) and meaning of the *A*-, *B*-, and *C*-types of varves (*see* MERTA 1978). These three different types of varves must be interpreted not only as a function of distance from the river mouth, but also as a function of position in the developmental sequence of the extraglacial ice-dammed lake.

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SEDYMENTACJA WARWOWA W ZASTOISKACH EKSTRAGLACJALNYCH

(Streszczenie)

Przedmiotem pracy jest zagadnienie mechanizmu sedymentacji warwowej w ekstraglacjalnych zastoiskach dolinnych, które prawdopodobnie były dominującym typem zastoisk w plejstoceńskim krajobrazie na obszarze Polski.

Przeprowadzono, w obrębie odsłaniających się sekwencji osadów zastoiskowych Mazowsza (*patrz* fig. 1 oraz pl. 1—4), analizę zmian miąższości warw. Traktując warwy jako efekt działania długotrwałych prądów zawiesinowych (*patrz* KUENEN 1951, MERTA 1978) wskazano, iż zmiany średniej miąższości warw są głównie funkcją szerokości obszaru sedymentacji. Stopniowa zmiana średniej miąższości warw w profilu (fig. 2) wynika ze stopniowego zapełniania osadami warwowymi zastoiska rozwiniętego w dolinie V-kształtnej (*patrz* fig. 4), skokowe zaś zmiany średniej miąższości warw (fig. 3) odpowiadają sedymentacji w zbiorniku zajmującym dolinę o reliefie tarasowym (*patrz* fig. 6). Gradient zmian średniej miąższości warw jest przy tym ściśle uzależniony od nachylenia zboczy zalanej doliny (*patrz* fig. 5).

Uwzględniając powyższe zależności, przedstawiono schemat rozwoju dolinnego zbiornika zastoiskowego (fig. 7). Schemat ten daje nową interpretację profilów ukazujących ostry kontakt osadów warwowych z podścielającymi je utworami piaszczystymi (*patrz* pl. 3—4), oraz traktuje taki typ kontaktu jako efekt istnienia długoczasowych luk sedymentacyjnych (*patrz* fig. 8).