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# Environmental changes in Poland and Sweden after the maximum of the last glaciation

## Proceedings of the First Polish–Swedish Seminar Poznań, Poland, October 9–15, 1986

Edited by

**STEFAN KOZARSKI**

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CONTENTS

# THE FIRST POLISH–SWEDISH SEMINAR ON "ENVIRONMENTAL CHANGES IN POLAND AND SWEDEN AFTER THE MAXIMUM OF THE LAST GLACIATION", 9–15 OCTOBER 1986, POZNAŃ, POLAND

## INITIATIVE AND ORGANIZATION

In the last three decades Polish–Swedish scientific contacts have developed in physical geography and Quaternary research. They were initiated by Professor Hoppe and Professor Dylik who collaborated on the organization of the Abisko Symposium at the IGU Congress, NORDEN'60 to continue this at the INQUA Congress that was held in Poland in 1961. Afterwards, there were frequent exchanges of young scientists who received grants, as well as exchanges of guest lectures and field excursions. The Swedish–Polish collaboration also gave rise to the IGCP Project 158 "Paleohydrological changes in the temperate zone in the last 15 000 years" led by Professor Starkel and Professor Berglund.

Owing to these contacts and a current need for the exchange of new ideas, it was not difficult to encourage and invite Swedish colleagues to hold a joint seminar in Poland. Environmental changes in Poland and Sweden after the maximum of the last glaciation, i.e. phenomena that occurred after 20 000 BP, were accepted as the main topic of the seminar. The decision was made after discussions with Professors Rapp, Berglund and Hoppe whilst I visited Sweden in 1984. Sponsorship of the seminar was received from the Polish Academy of Sciences within the framework of an agreement with the Royal Swedish Academy of Sciences on collaboration. The organizer of the seminar was the Committee of Geographical Sciences with which the Committee for Quaternary Research of the Polish Academy of Sciences collaborated. The Poznań Branch of the Polish Academy of Sciences where paper sessions were held was also involved in the organization.

The seminar programme included paper sessions and excursions with demonstration of sites in case study areas. The sites represented examples of research conducted in NW Poland and were related to the topics of some Polish papers. The seminar was begun on 9 October 1986 with the demonstration of sites in Wolin Island.

## PAPER SESSIONS

Papers presented at the seminar were chiefly concerned with the area of the last Scandinavian glaciation. Merely one paper (by L. Starkel) dealt with paleogeography of the periglacial domain in Poland during the maximum of the last glaciation. Successive papers discussed a sequence of major events and problems associated with the last ice-sheet waning in Poland and Sweden, as well as Late Vistulian and Holocene environmental changes.

During three paper sessions held on 10, 11 and 13 October, 1986, twenty papers were altogether presented. They tackled the following problems:

– Biostratigraphy, geomorphology, sedimentology and radiometric measurements of Vistulian deposits in the zone of maximum extent of the last Scandinavian glaciation in the vicinity of Konin (K. Tobolski, A. Stankowska and W. Stankowski);

– Dynamics and chronology of deglaciation in NW Poland and S Scania (S. Kozarski, B. E. Berglund, S. Bjorck and G. Digerfeldt);

– Proglacial drainage, ice melting and phases of development of subglacial channels (J. Szupryczyński, W. Niewiarowski);

– Climate and Late Vistulian and Holocene landscape evolution in SW Scania and Bornholm Island (B. E. Berglund and A. Rapp, E. Lindstrom);

– Late Vistulian and Holocene stratigraphy on the Polish coast (Vistula delta plain), and in SW Scania, Hanö Bay and Bohuslan (J. E. Mojski, B. Dennegard, S. Bjorck, U. Miller and P. Sandgren);

– Holocene slope processes in high mountains, the High Tatra Mts and North Scandinavia (A. Kotarba, L. Stromqvist and Ch. Jonasson);

– Human impact on the Holocene vegetation in North Poland and Baltic islands (M. Ralska–Jasiewiczowa, L. K. Konigsson);

– Present-day development of the Wolin Island cliff (A. Kostrzewski and Z. Zwoliński).

Most of the papers are included in this volume.

## EXCURSIONS AND DEMONSTRATION OF SITES

Excursions and demonstration of test sites located in case study areas in various portions of NW Poland remained complementary to some of the Polish papers and to the seminar topic:

– Wolin Island, 9 October: the development of cliff at Międzyzdroje and Holocene stratigraphy of inland dunes at Troszyn, with special reference to human impact, were demonstrated (Leaders: A. Kostrzewski and Z. Zwoliński, B. Nowaczyk);

– The vicinities of Konin, 12 October: this excursion combined a presentation and discussion of bio- and lithostratigraphy and radiometric measurement of Vistulian deposits and of environmental changes produced by brown-coal open-cast mining (Leaders: K. Tobolski, A. Stankowska and W. Stankowski);

– The environs of Poznań, 14 October: this excursion demonstrated results of new study of the marginal zone belonging to the Poznań Phase of the last glaciation at Ceradz Kościelny (Leaders: L. Kasprzak, S. Kozarski), of braided river paleochannel scars and fills at Żabinko (Leaders: B. Antczak, S. Kozarski, B. Nowaczyk, K. Tobolski) and paleomeanders (Leaders: P. Gonera, I. Okuniewska–Nowaczyk);

– The vicinities of Ujście at the margin of the Toruń-Eberswalde Pradolina: litho- and kinetostratigraphic evidence of the last ice-sheet readvance during the Chodzież Subphase was demonstrated in detail (Leadres: S. Kozarski, L. Kasprzak).

Throughout the demonstration of sites and case study areas, special attention was given to research procedures and careful interpretation of sedimentological and paleobotanical records and radiometric measurements, including a comparison between $^{14}C$ and TL dating.

## SOCIAL EVENTS

In order to offer the guests from Sweden the opportunity for learning about Poland's cultural life and all the participants that for relaxations after effort-consuming day-long paper sessions, we had a varied social programme including an evening out at the Opera House in Poznań (10 October: performance of the national opera Straszny Dwór [= The haunted mansion] by Stanisław Moniuszko, a Polish composer), sight-seeing in the medieval part of Poznań (11 October) and a visit to the Library of the Polish Academy of Sciences in the Kórnik Castle where Professor J. Wisłocki, its director, showed its collection to the participants. We also had two evening parties, one at the Quaternary Research Institute of Adam Mickiewicz University on 11 October and the other in the Reception Hall of the Polonez Hotel.

RESOLUTION

At the end of the seminar its participants admitted that:

(1) the seminar was a significant scientific event which first of all promoted discussion on some main palaeoenvironmental events related to the last ice-sheet waning and to environmental changes in Poland and Sweden during the Late Vistulian and Holocene;

(2) field excursions and exact demonstration of sites in case study areas were a very important part of the seminar, they enabled the participants to learn about a record of the events under discussion and about research methods;

(3) papers presented at the seminar would be published as the proceedings of the Polish-Swedish Seminar in a special issue of Geographia Polonica;

(4) joint seminars on environmental changes should be continued, as coordinator, Professor A. Rapp announced on behalf of the Swedish group that the next seminar would be held in Sweden in 1989;

(5) tribute should be paid to the Polish Academy of Sciences that allowed the seminar to be held by sponsoring it and by providing help throughout the organization.

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# GEOMORPHOLOGY, CLIMATE AND VEGETATION IN NORTH-WEST SCANIA, SWEDEN, DURING THE LATE WEICHSELIAN

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ABSTRACT. A reconstruction of the conditions of the cold climate environment during the Late Weichselian is made by means of methods of geomorphology and Quaternary stratigraphy. The geomorphological part deals particularly with landforms and material related to nivation and local glaciation, making use of comparisons with contemporary snow-fields and mini-glaciers in northern Lappland. The two main field localities investigated in NW Scania are the horst ridges Söderåsen and Kullaberg, where geographical evidence and TL-datings are combined with sedimentological and biostratigraphical studies of lake sediments. The results are used for a tentative reconstruction of palaeoclimate and palaeoenvironment 13 500–10 000 BP based on past vegetation and insect fauna, occurrence of stagnant ice, wind and water erosion, nivation and local glaciation.

## INTRODUCTION

This paper is a joint effort to reconstruct the cold climate environment during part of the Late Weichselian period in South Sweden by means of methods in geomorphology and Quaternary geology. Fossil ice-wedges and frost-crack polygons, wind polished rocks and landforms related to nivation and local cirque glaciers have been described and discussed from the province of Scania as tools for reconstruction of former phases of cold climates (Mattsson 1957; Johnsson 1982; Rapp 1982, 1984; Rapp et. al. 1986; Svensson 1972, 1980, 1983). Lake sediments, micro- and macrofossils are important objects of study, used by Quaternary geologists for reconstructions of the Late Weichselian climates of Scania (Berglund 1971; Berglund & Lagerlund 1981; Berglund et al. 1984).

This paper presents on-going research from the above-mentioned two fields of study. The geomorphological part deals particularly with forms related to nivation and local glaciation, and utilizes comparisons with contemporary snow-fields and mini-glaciers in northern Lappland (Figs. 1, 2). Then follow presentations of

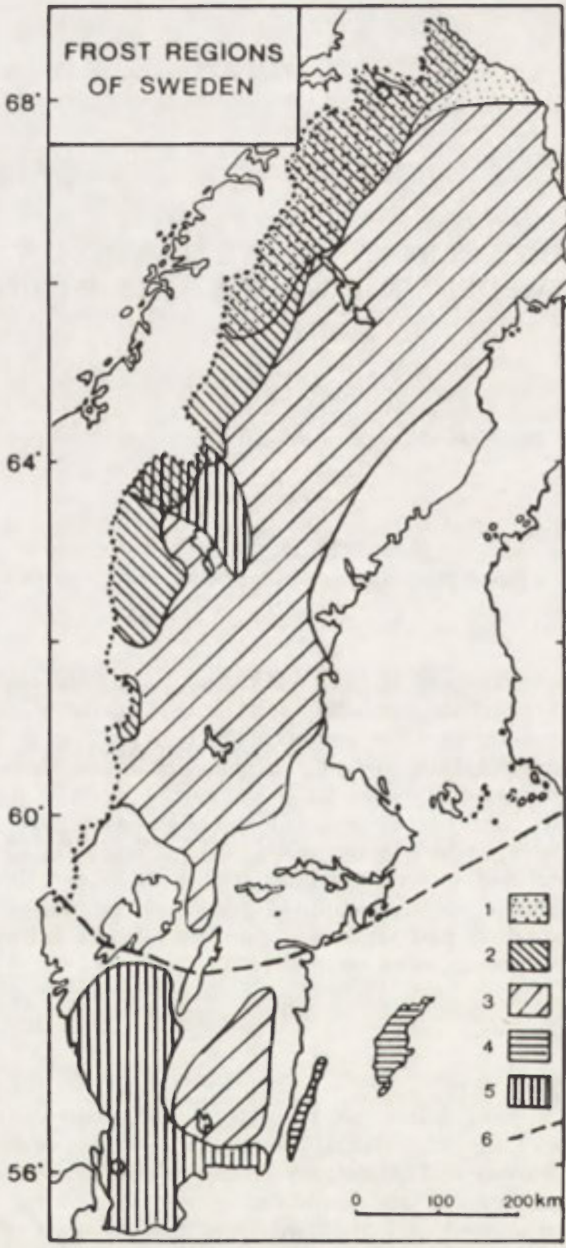


Fig. 1. Regional division of Sweden according to the types of frost phenomena in the ground

1 – region with discontinuous permafrost on low levels, mainly palsa mires, 2 – Caledonide mountains, 3 – region of boulder depressions, 4 – island of Oland and Gotland, 5 – regions with fossil ice-wedges, 6 – Younger Dryas icefront advance until 10 500 BP cf. Fig. 5.

Map modified after Lundqvist 1962 and Johnsson 1981, 1986. Small circle at 56° N shows location of Söderåsen horst ridge in Scania. Same symbol at 68° N shows location of Abisko mountains, N. Sweden results from two localities in NW Scania, namely the horst ridges Söderåsen and Kullaberg, where geomorphological evidence is combined with sedimentological and biostratigraphical studies of lake sediments. The results are then used for a tentative reconstruction of palaeoclimate and palaeoenvironment 13 500 to 10 000 BP.

## COMPARISON OF ACTUAL NIVATION-RELATED ENVIRONMENT IN LAPPLAND AND LATE WEICHSELIAN TUNDRA IN OTHER AREAS

In the mountains surrounding Abisko in northern Sweden (Fig. 2) the main deglaciation is thought to have ended about 9000 BP. Above the present tree line at about 600 m a.s.l. strong snow drifting and nivation may have occurred during the whole Holocene, in favourable places. Late-lying snow patches are characteristic of the north Swedish mountains, but the geomorphic effects seem to have been too weak to create nivation hollows in bedrock during the available time of about 9000 years. Rudberg (1974) made a study of the actual nivation processes in south Lappland and concluded that gelifluction and other processes connected with late-melting snow patches were active and measurable, but of a slow rate (cf. Thorn 1976).

Figure 2 is a map of the mountain area of Abisko. It shows the study sites selected for our studies of nivation and "mini-glacier" action. The map also shows the sites of sub-recent frost-crack polygons, palsa mires and thermokarst ponds, indicating discontinuous permafrost. The majority of the cirque glaciers in the area are facing E–NE direction and thus show the strong predominance of snow drifting from westerly directions.


Fig. 2. Map of mountain area of Abisko, N. Sweden. Study sites of nivation marked by rectangular boxes: 1 – lake and streams, 2 – palsa mires, 3 – frost crack polygons, 4 – thermokarst ponds, 5 – mountain top with altitude in m, 6 – study site for nivation, 7 – road, 8 – glacier

Frost-crack polygons (Fig. 2:3) occur mainly in wind-swept valley bottoms and plateaux areas at 800–1000 m altitude, well above the tree line. Table 1 gives data on precipitation and temperature at three sites in the area (Eriksson 1982; Åkerman and Malmström 1986).

TABLE 1. Mean annual precipitation and air temperature at Abisko and nearby sites (1951 – 1980). From Åkerman and Malmstrom (1986) and Eriksson (1982)

| Place | Altitude (m) | Annual precipitation (mm) | Air temperature (°C) |
|---|---|---|---|
| Abisko | 388 | 322 | −0.9 |
| Laktatjakko | 1220 | 1750 (estimate) | −5.8 (estimate) |
| Katterjokk | 515 | 807 | −1.7 |

Figure 3 is a photo of a nivation study site at the canyon of Nissunjokk, ca 6 km south of Abisko. The canyon is cut into quartzitic schists and is 50–80 m deep. The tree line is at 650 m on the west-facing slope. Above this zone large lee-side snow drifts are formed by the predominant westerly storms. Note the colluvial cone below the largest hollow and snow drift. We call it a nivation cone. It is deposited below a steep nivation hollow and was accumulated by combined processes of small debris flows, running water and gelifluction, due to summer meltwater from the shrinking snow patch.

Fig. 3. Nivation hollows in bedrock at Nissunjokk canyon, Abisko. Note nivation cone of till removed from pre-Weichselian hollow by gelifluction, debris flows, wash and small snow avalanches during 9000 years of post-glacial tundra conditions. Slope is East-facing and 70 m high. Photo A. R., 12 June 1983

Fig. 4A

Fig. 4B

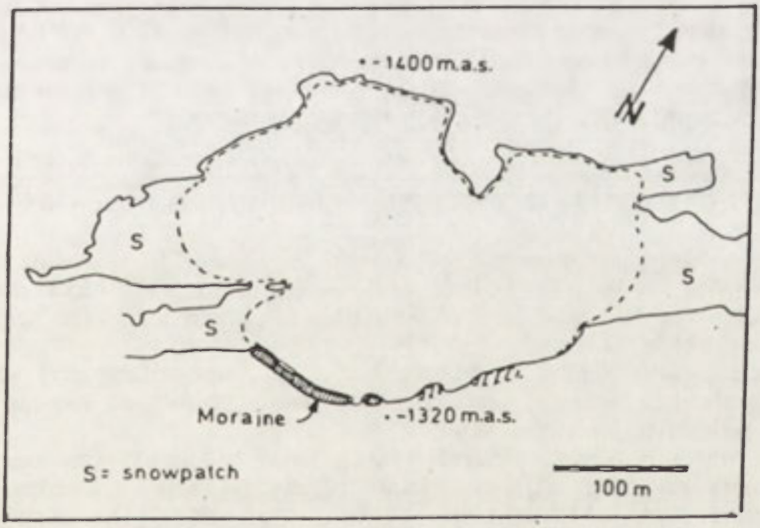


Fig. 4A and B. Aerial photo and map of a small glacier/snowfield facing south on Latnja mountain, SW of Abisko.
Note many small debris flows over icy surfaces. Flows stop at contact with remaining snow patch to the left. A 100 m long end-moraine appears at lower end and some meters from the main glacier/snowfield. Note also talus slope with large blocks at the base, to the far left.
Air photo, N.Å. Andersson 14.9.1985. Map by L. Lindh, autumn 1984

Figure 4 shows the Latnja mountain glacier/snowfield. It has a distinct moraine-like ridge in front of a 60–80 m high ice body of 30° gradient, covering a cliffy bench. The morainic ridge is about 2 m high and 100 m long, situated at 1320 m altitude. The glacier/snowfield is accumulated by snow-drifting in the lee of a plateau, and is facing south-to-southeast. The ice surface has many small mudflows. The site seems to offer very interesting opportunities to test the concepts of development of end-moraines contra protalus ramparts at small and steep glaciers and snow-fields. The interpretations by Ono and Watanabe (1986) refer to a protalus rampart from an area in the Japanese Alps of steeper ice surface, probably of more available debris, higher intensity of rainfall and larger debris flows than at the Latnja morainic ridge. Other small glaciers with moraine-like ridges in the Låkta area occur down to an altitude of only 1000 m, e.g. the NE-facing Kåppa glacier (Lindh 1984). This is about 500 m lower than the glaciation limit of the area (Östrem et al. 1973), and shows the importance of snow-drifting to create small glaciers and moraines at very low levels in favourable localities. The Kåppa glacier site occupies a steep U-valley side which traps drifting snow from a plateau with a long fetch area.

Comparisons with other areas are also of great interest, and we would particularly like to mention the work now in progress to reconstruct the effects of local glaciers and snow fields during the Younger Dryas period in Western Norway where glaciers occurred down to sea level (Longva et al. 1983; Larsen et al. 1984), in Northwest England down to altitudes of 300–400 m a.s.l. (Sissons 1979, 1980; Vincent and Lee 1982; Pennington 1977; Boardman 1985), and in the Vosges, Black Forest and other mountains of Middle Europe (Poser 1953; Fezer 1953; Tricart 1981; Vogt 1984) down to 600 m.

Judging from the experiences expressed by Rudberg (1974) and Strömquist (1983), it is a reasonable working hypothesis to assume, that nivation processes such as slope wash and gelifluction in Lappland are likely to be slow at low elevations near the tree line, increasing upwards towards an optimum zone of late-summer-melting snow patches, and decreasing again in intensity at still higher elevations with a very short snow-free time in summer.

## PRESENT-DAY AND LATE WEICHSELIAN CONDITIONS IN SCANIA

The southernmost province of Sweden is Scania. It is today characterized by cultivated flat or rolling plains and by contrasting, horst-block plateaux-areas, mainly covered by forests of beech or planted spruce trees. The climate is mild temperate like in Denmark.

But the winter winds can be strong and the snow-drifting very severe in this open landscape, blocking roads with thick snow drifts and causing damage to open fields through wind erosion.

The whole province of Scania was covered by the Weichselian and earlier Quaternary ice sheets. The deglaciation pattern in southern Sweden is discussed in another paper of this volume (Björck et al. 1988). The main ice-marginal zones are presented in relation to the Late Weichselian time-scale in Fig. 5. A large number of ice-wedge casts (Johnsson 1956 and later; Svensson 1962 and later) found in gravel pits on the sandy plains of Scania and south Halland are proofs of severe permafrost conditions probably during Younger Dryas as well as other phases of tundra climates in Scania.

Other signs of severe tundra climate in phases after the major deglaciation are occurrences of wind-polished bedrock and boulder pavements. The wind flutes and facets of wind polishing are from two main directions: the east and the

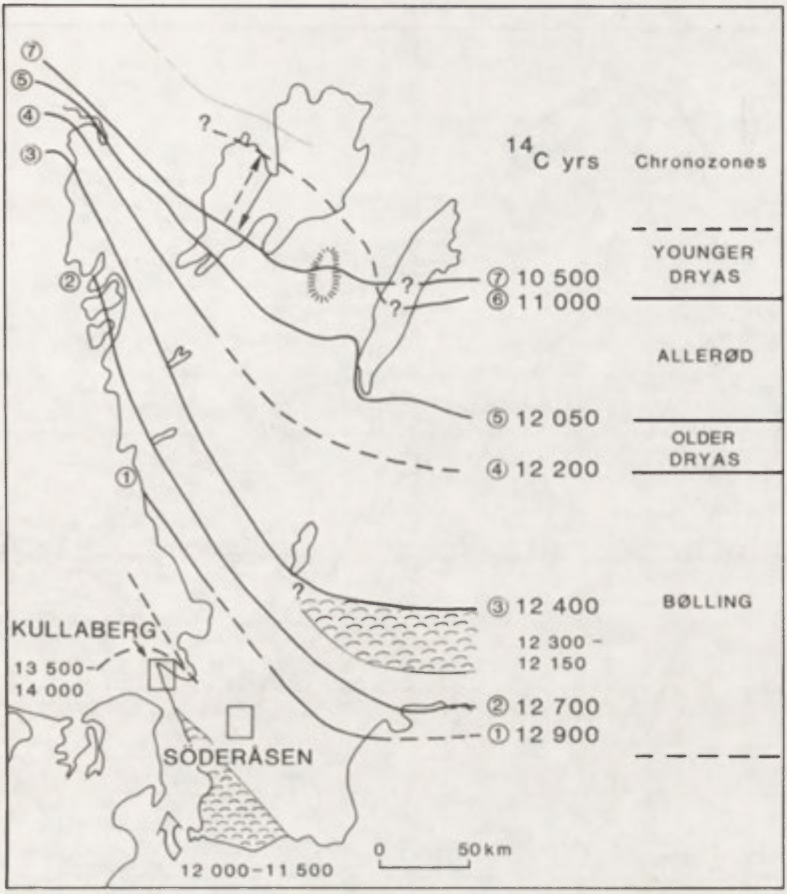


Fig. 5. Late Weichselian ice-marginal zones in south-west Sweden according to the revised deglaciation chronology by Bjorck et al. (this volume). Numbers refer to the system described by Berglund (1979)

west quadrants (Mattsson 1957; Johnsson 1982 and earlier; Svensson 1972, 1980, 1983).

The horst ridges like Söderåsen and Kullaberg are built of Precambrian gneiss and surrounded by Mesozoic sedimentary bedrock. The fault lines give the landscape a marked relief – the plateau of Söderåsen being 100–150 m higher than surrounding plains and the narrow ridge of Kullaberg rising more than 150 m above surrounding sea and coastal plain (Fig. 6). Both ridges are dissected by valleys, in Söderåsen formed as deep canyons. In our working hypothesis (Rapp et al. 1986) the sides of the canyon valleys have had long-lasting snow fields and even some small, local glaciers, e.g. one which created the over-deepened cirque basin of Lake Odensjön (Fig. 7). Most of this sculpture of nivation and local glaciation was created before the Weichselian ice covered the area from about 20 000 BP. But the processes of nivation and local glaciation were repeated as late as during the Younger Dryas cold period, as we will discuss below.

Figure 7 is a map of the central part of the gneiss plateau of Söderåsen. It shows the location of the three canyon valleys of Odensjön, Uggleröd and Skäralid. The sandy plains north of the fault-line slope of Söderåsen are below

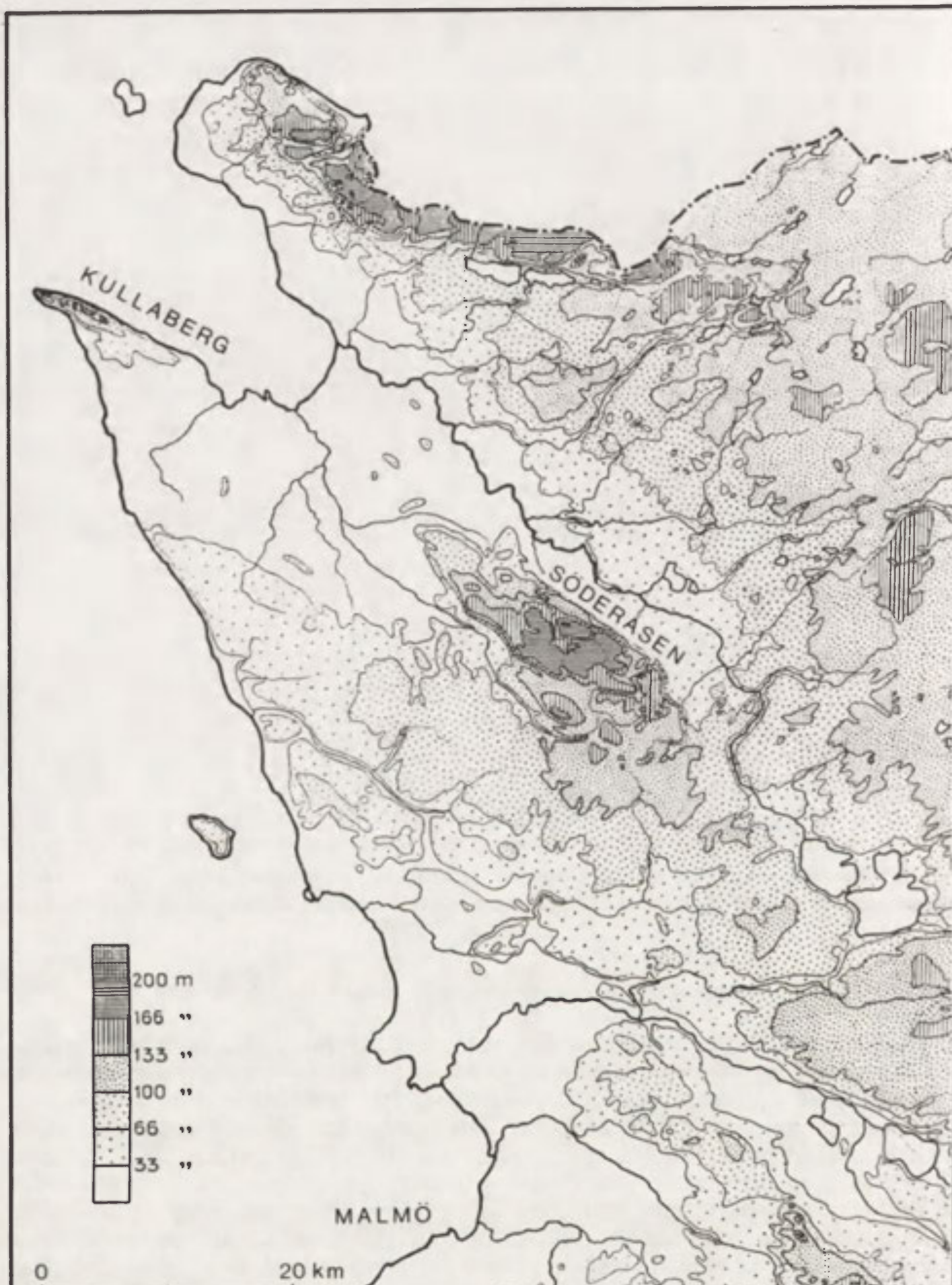


Fig. 6. Topography of western Scania with the horst ridges Soderasen and Kullaberg indicated

the highest coast-line (HK) at 55 m a.s.l. of the Weichselian deglaciation phase (Ringberg 1984; Sandgren 1983). The forms interpreted as nivation hollows are best developed at the rims of canyon valleys (Mattsson 1962, Rapp 1982).

The most widespread interpretation of some of the cirques and hollows of Söderåsen has been the plunge pool hypothesis of Lennart von Post (1938), based on the concept of glaciofluvial scour by meltwater at the Weichselian deglaciation. The hollows of the valley sides are not interpreted as carved by meltwater in our concept, but as nivation hollows. Several of them show typical characteristics of nivation: a hollow in bedrock which has semicircular back rim, gently sloping floor without incision by a stream trench, and with angular rock debris over bedrock on the floor and sides. Many of the hollows have a hanging position to the main valley, that is, a longitudinal profile with a break in inclination between a gently sloping floor of the hollow and a steeper lower part. There are also many nivation hollows which do not have a hanging floor, but reach the main valley bottom and show a simple concave long profile of lower gradient than pure gravity talus slopes.

Two well developed nivation hollows of 50–60 m width, ca 60 m length and semicircular upper rim occur on the west side of the narrow mouth of the Odensjön valley. These two hollows, like many similar ones in this valley and Skaralid valley, have a smooth, semicircular upper rim, undissected by stream erosion. On their gently sloping upper floor are a few rounded gneiss boulders of 0.5–2 m length. The long axis in most cases points downslope in the direction

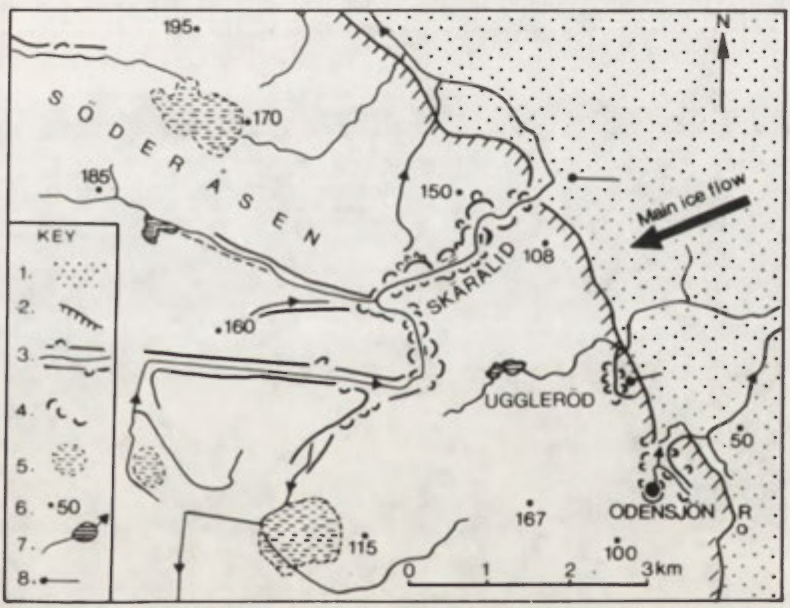


Fig. 7. Eastern part of Söderåsen, Scania, a gneiss plateau with canyon valleys and periglacial nivation hollows and probable glacial cirques

1 – sandy plains with many ice-wedge casts and wind-polished rock surfaces (H. Svensson 1983; G. Johnsson 1982) from periglacial tundra periods, 2 – main fault scarp slopes, 3 – canyon valley with small stream and nivation hollows, 4 – nivation hollows and glacial cirques (larger arcs) in bedrock of canyon sides, 5 – swamps, 6 – elevation in m above sea level, 7 – lake and stream, 8 – locality with wind polished bedrock or blocks

This black arrow marks direction of main Weichselian ice from ENE. Maps of the article are published with the permission of Lantmateriverket (Swedish Geographical Survey) of creep and ealier gelifluction. They are interpreted by us as erratic boulders, dropped by the Weichselian ice together with till, and later moved somewhat by gelifluction.

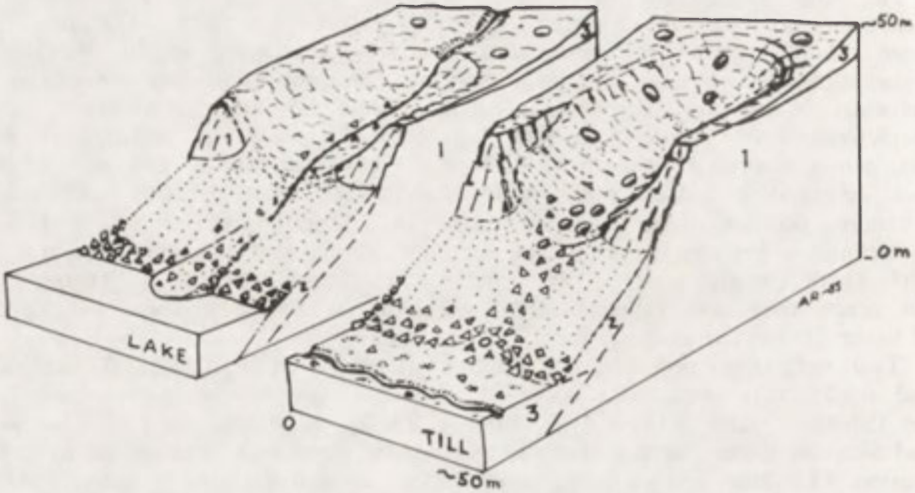


Fig. 8. Block diagrams of nivation hollows and debris slopes, Soderåsen 1 – gneiss bedrock, densely jointed, 2 – talus, 3 – till, on the upper plateau from NE, with rounded, erratic boulders brought by the Weichselian ice

Fig. 9. Wind-polished block with distinct facet edges and wind flutes. Erratic block of quartzitic sandstone deposited by Weichselian ice from NE at Ugglerod, Soderåsen. On the side of the block are two types of pointed wind flutes, here called straight flutes and spiral flutes. The latter type is a particularly reliable indicator of wind direction. Photo A.R., October 1984

If this interpretation is correct, it means that till from the Weichselian ice is blanketing several of the largest and less steep nivation hollows. Thus these hollows were formed mainly before the Weichselian glaciation and survived its glacial scour. This condition is quite evident at the locality of Uggleröd (Fig. 7, 8 and Rapp, Nyberg and Lindh 1986).

## GEOMORPHOLOGICAL EVIDENCE OF WIND EROSION, NIVATION AND LOCAL GLACIATION IN SCANIA

### DEBRIS DEPOSITS CONNECTED WITH NIVATION HOLLOWS AND GLACIAL CIRQUES

Our interpretation of fossil glacial cirques and fossil periglacial nivation hollows in the valleys of Söderåsen is supported by the forms, locations, material and orientations of the erosional amphitheatres. The interpretation is also supported by the debris deposits which exist in connection with the cirques or hollows, either on some hanging hollow floors, or in the valley bottoms. The closer analyses of these accumulations including SEM-studies of the material will be reported on in a forthcoming paper (Nyberg, in prep.).

The most end-moraine-like ridge in the area is in a small cirque, near the place Liagården, Skaralid (Fig. 10). It is about 100 m long, is clearly connected with the talus slope at one end and forms a slight bow in front of the talus. Excavations to 1.5 m depth show a material of angular blocks in a sandy matrix which speaks against the alternative interpretations of formation by rock-glacier creep or blocks gliding on snow as in protalus ramparts. Those two explanations may be more relevant for the second ridge closer to the base of the same talus slope. Measurements of stone orientation in the first-mentioned ridge show

Fig. 10. Photo of the end moraine ridge and the "inner ridge" at the base of talus slope of Fig. 4 viewed from above. Photo A.R., November 1982

preferred orientation transverse to the ridge axis and hence agree with the end-moraine alternative of genesis (Nyberg).

The supposed glacier cirques, the nivation hollows, the debris accumulations of all types discussed here, except debris-flow lobes, all point to an environment of treeless tundra with strong snow-drifting, large snow-drifts, nivation, active small cirque glaciers and snow avalanches also in post-Weichselian time, e.g. Younger Dryas.

The nivation hollows of the small valleys at Ugglerod and Odensjon are about 30–60 m wide and thus rather small in size. In other parts of Soderasen, with larger canyon valleys of 50–60 m depth, the nivation hollows are up to 150 m in width. Valley-side or valley-end cirques of larger size, 220 to 400 m width occur. Some of these are connected with overdeepened basins, or have debris accumulations similar to end-moriane ridges. Other kinds of deposits here called nivation cones are marked in Fig. 8 and. shown in actual form in Fig. 3.

## TL-DATING OF SILTY-SANDY MATERIAL IN NIVATION HOLLOWS AND MORAINE RIDGES OF SODERÅSEN

An assembly of forms and deposits related to former nivation and local glaciation in Soderåsen is shown in the diagram sketches of Fig. 11. It also shows the results of several datings of sandy-silty material in this environment, The datings were made by Dr Vagn Mejdahl, Riso Laboratories, Denmark, the most experienced expert in TL-dating in Scandinavia. The basis of the TL method is that the minerals quartz and feldspar are deposited with a reduced TL level as a result of exposure to light during transport. "The level cannot be reduced to zero, even by prolonged bleaching, but the residual TL level can be quite small especially for eolian sediments. The bleaching effect of light is different for different minerals: potassium feldspar (K-feldspar) is easily bleached by all wave-lengths, whereas quartz is more resistant... After deposition, the TL level in the minerals grows again as a result of exposure to the background radiation, and the increase above the residual level is a measure of the radiation dose received by the minerals and thereby of the age" (Kolstrup and Mejdahl 1986).

The four datings listed in our Fig. 11 are all from the time period 11 000 to 10 400 $\pm$ 700 years BP. This strongly indicates that they were deposited during the Younger Dryas cold period.*

Eolian activity is indicated by the occurrence of wind-polished blocks (Fig. 9.) on the plateau surfaces of Soderasen, e. g. Ugglerod (uppermost profile of Fig. 11). The west-facing nivation hollow at Ugglerod has a sheet of silty-sandy material, about 50 cm thick, mixed with a few stones, some of them wind-polished. The sand is not stratified and rests on till-like material. Sand at 40 cm depth gave TL-ages of 11 000 $\pm$ 700 BP in the upper part of the floor, and 10 500 $\pm$ 700 BP in the downslope part of the floor. We interpret this as niveo-eolian sand trapped by snow-patches in the Younger Dryas, and incorporated into a solifluction deposit when the snow melted every summer.

The two moraine-like ridges at the base of talus slopes in the canyon valley Skäralid are also drawn as schematic slope profiles in Fig. 11. One is at Liagården, facing southwest, at the base of a 50 m high talus and bedrock slope. The morainic ridge contains a matrix of fine sand, carrying angular blocks of gneiss,

* Two newer datings at Liagården gave 10 800 $\pm$ 1000 years BP.

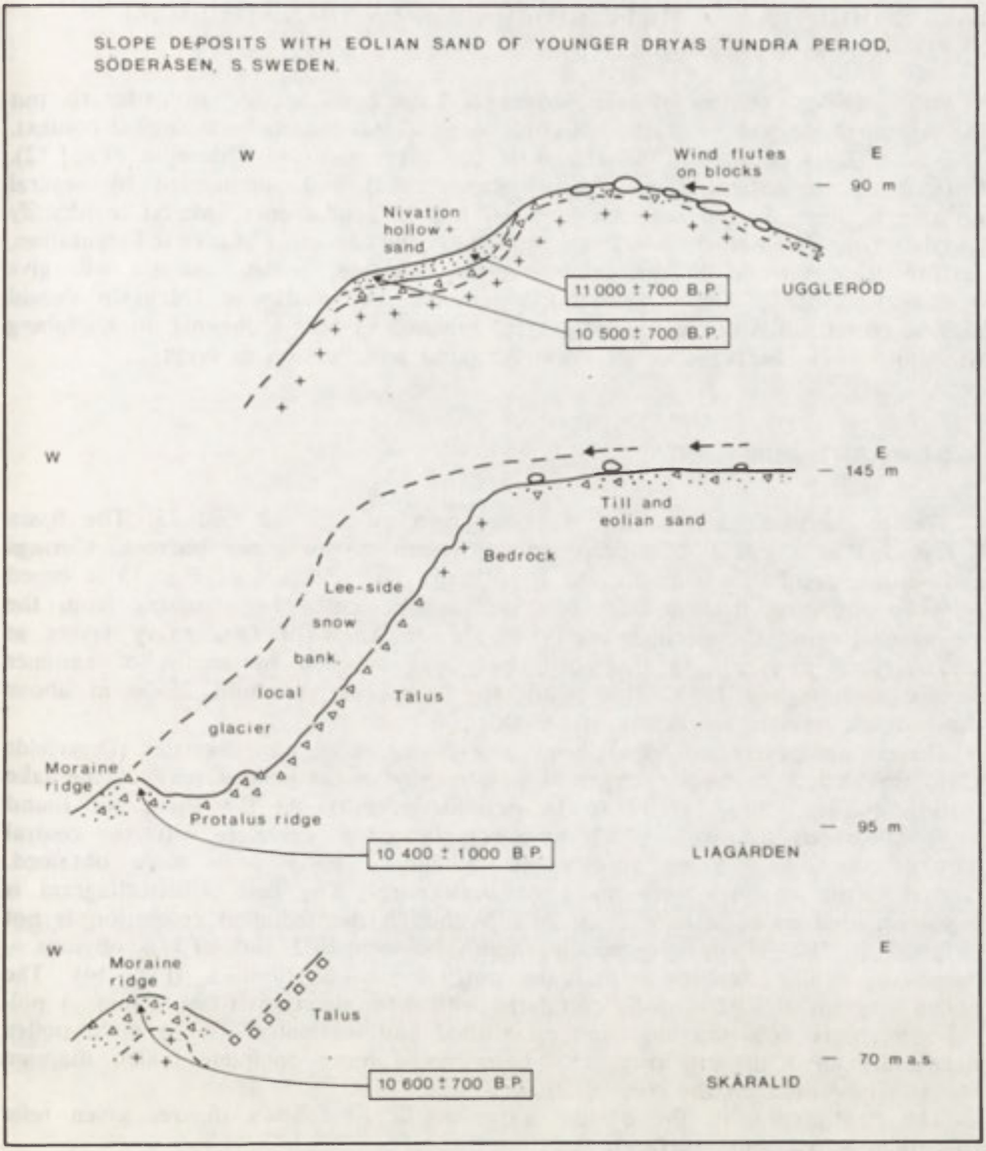


Fig. 11. Diagram of slope deposits with eolian sand of Younger Dryas age. TL-ages are given in boxes. Datings by V. Mejdahl

up to 0.5 m in length. The TL-age of the sand at 1.5 m depth in the ridge is 10 400 ± 1000 BP. A smaller, blocky ridge at the base of the talus contains no sand and is interpreted by us as a protalus rampart. We think the sand content in the outer ridge was brought by wind drifting across the plateau, deposited in lee-side snow drifts, which grew into a small "mini-glacier", similar to the Latnja glacier of Fig. 4 above. The moraine ridge in Skäralid at Forshall is higher (5–10 m), partly filled by a bedrock core. It is covered by a mix of sand and angular blocks and stones at the base of a west-facing talus and rock slope of 80 m height. The TL-age of the sandy material at 1 m depth is 10 600 ± 700 years BP.

## LAKE SEDIMENTS AND THEIR INFORMATION ON THE LATEGLACIAL ENVIRONMENT IN SCANIA

Stratigraphical studies of lake sediments have been applied in order to put the geomorphological events into a chronological and palaeoclimatological context. The most suitable site on Söderåsen is the lake basin of Odensjön (Fig. 12), situated in an ancient cirque basin (Rapp 1984) and surrounded by several nivation hollows. In our research program it is of fundamental interest to identify and date Late Weichselian sediments, particularly the beginning of lake sedimentation, and to trace erosion in the catchment. In addition, pollen analysis will give a general picture of the surrounding vegetation. The studies at Odensjön should also be correlated with earlier and current research of lake sediments on Kullaberg (Berglund 1971; Berglund et al. 1984; Berglund and Persson in prep:).

### LAKE ODENSJÖN ON SODERÅSEN

The geomorphological setting is best shown in Figs. 12 and 13. The basin is regarded as a glacial over-deepened depression in the gneiss bedrock. Corings and seismic profilation support this hypothesis. The selection of Fig. 13 is based on echo sounding from a boat in 1982, complemented by sounding from the ice parallel with the corings in 1984. The thickness of Quaternary layers at the overflow threshold in the north has been checked by means of hammer seismic sounding in 1983. This points to a bedrock threshold 25–30 m above the bedrock level in the centre of the Odensjo basin.

The corings performed by a piston core device of 60 mm diameter (Digerfeldt 1978) revealed a "normal" lateglacial stratigraphy in the central part of the lake (coring points 1 to 3 at 19 to 16 m water depth). At the shore we found more calcareous sediments which were complicated to correlate with the central part of the basin. From core points 2 and 3 good cores were obtained. Granulometric analyses were made on both cores. The first pollen diagram is based on analysis of a core from BP2. Although the sediment resolution is not so good at BP2, a stratigraphic correlation between BP2 and BP3 is obvious – based on ocular similarities and the particle size distribution (Fig. 14). The pollen diagram of BP2 is easily correlated with other diagrams from Scania. A pollen assemblage zonation has been established and correlated with a main pollen diagram from Kullaberg (Fig. 15). Later on a more complete pollen diagram will be elaborated on the core at BP3.

The stratigraphy of the bottom sediments is as follows (figures given refer to depths below water surface).

| | | |
|---|---|---|
| Layer 6 | 18.70–23.53 | Fine detritus gyttja, dark green. Organic matter >70% |
| Layer 5 | 23.53–23.70 | Slightly clayey fine detritus gyttja, dark grey-green. Organic content 20–60%, increasing upwards |
| Layer 4 | 23.70–23.84 | Gyttja clay, grey. Organic content < 8%. Very sharp boundaries upwards and downwards |
| Layer 3 | 23.84–24.00 | Moss clay gyttja, dark brown. Organic content 10–30%. Upper 10 cm a moss filt layer. In BP3 the moss content is lower and layer 3b is a transition layer between 3a and 4 |
| Layer 2 | 24.00–24.05 | Glay gyttja, brown-grey. With scattered mosses. Organic content 5–20% |

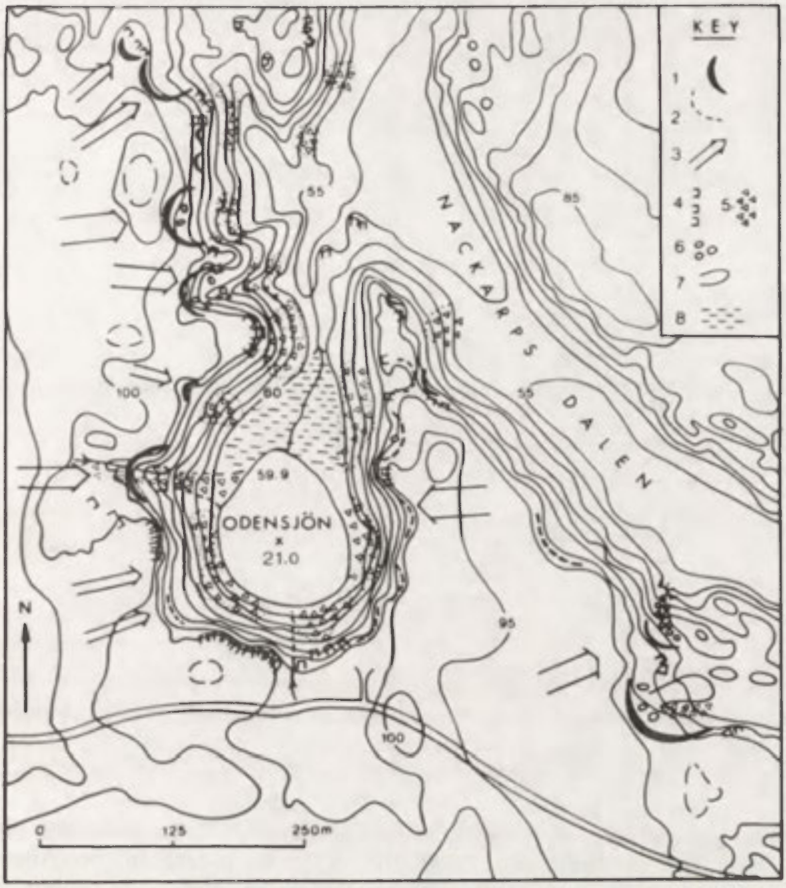


Fig. 12. Geomorphological sketch map of Lake Odensjon and Nackarpsdalen valley in the SE part Soderåsen, Scania

1 — clear or incipient nivation cirque, 2 — weakly developed nivation hollow, 3 — direction of snow drifting into nivation hollow, 4 — bedrock exposure, 5 — angular rock debris, e.g. talus, 6 — rounded boulders in block streams or single "creeping boulders" in former solifluction material, 7 — debris lobe, possibly from periglacial solifluction, 8 — swamp

| | | |
|---|---|---|
| Layer 1 | 24.05–24.10 | Slightly muddy, sandy silt, light brown-grey. Organic content < 5%. Some material lost at BP2, but at BP3 the coring was more successful and layer 1 probably about 20 cm thick above boulder-rich till |

The chronology has to be based on pollen-analytical cross-correlations with earlier dated reference profiles in South Sweden (cf. Berglund 1971, 1979; Björck et al. 1988). The chronozone sequence applied here follows Mangerud et al. (1974) with the exception of the Older Dryas Chronozone which is given the age 12 050–12 200 (Björck 1984), see Figs. 14, 15. Lacustrine brown mosses at 23.84–23.90 m have been radiocarbon dated to 14 800 ± 180 BP (Håkansson 1986), but this age is much higher than expected, probably caused by the reservoir age (cf. Håkansson 1979).

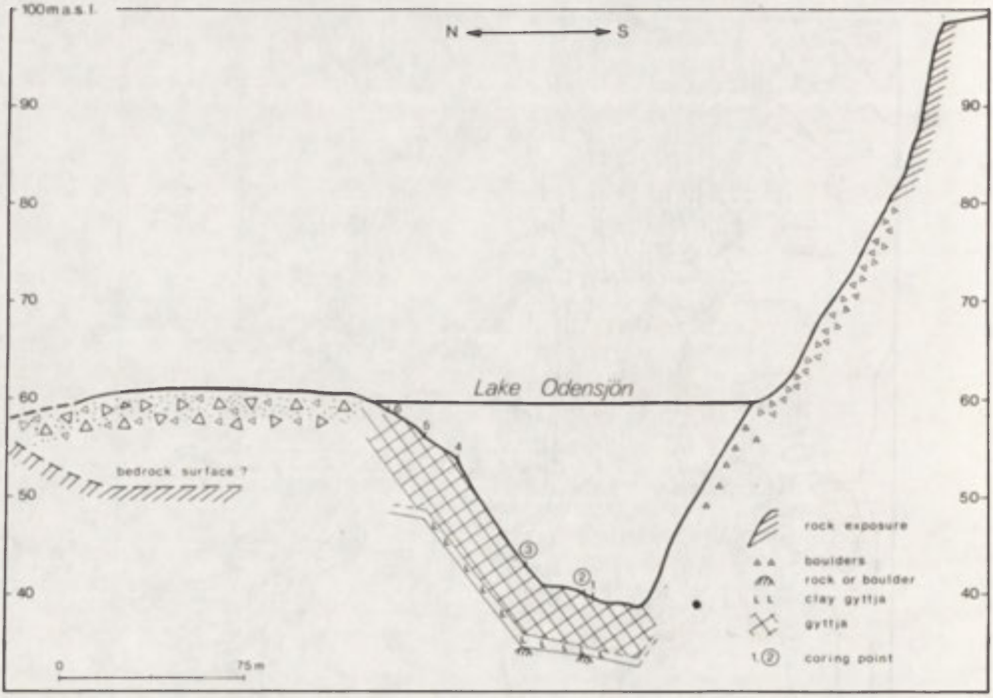


Fig. 13. Cross-section of Lake Odensjon basin with stratigraphy of the lake sediments based on corings 1 to 6

The oldest lacustrine sediments — bottom of layer 1 — are definitely older than Alleröd, probably of the age 12 500 BP or even older. In pre-Alleröd time a coarse sediment with sand and silt was deposited, probably caused by water erosion of the exposed basin slopes. Possibly, nivation processes, as described above, favoured this kind of sedimentation. We are unable to demonstrate the impact of eolian activity. Anyhow, the sediments change towards a clayey silt with some organic content at the end of the pre-Alleröd period (the Older Dryas event is not yet identified sedimentologically). This is caused by decreased erosion parallel with the revegetation of the ground. This development continues into the Alleröd period during which the vegetation cover protects the mineral soil as efficiently as during the Preboreal period. The sediment at the end of the Alleröd period about 11 200 BP is a clay gyttja with silty clay. At the transition to Younger Dryas 11 000 BP the sediment is suddenly becoming more minerogenic, with increasing silt content. We assume an increased slope erosion parallel with a deteriorated vegetation cover. Slightly later, possibly about 10 800 BP, there was a change to a sedimentation of a very minerogenic silty clay. Most probably the vegetation was very poor then, the soils exposed and easily eroded due to harsh climatic conditions. Nivation processes favoured such sedimentation. This situation continued until about 10 400 BP when the silt-clay deposition suddenly changed towards sedimentation of a highly organic gyttja with clay. The vegetation became denser and the ground was protected. During a short period around 9500 BP there was again increased minerogenic sedimentation (mainly clay). Possibly, this was a result of a water-level lowering (cf. Digerfeldt 1976).

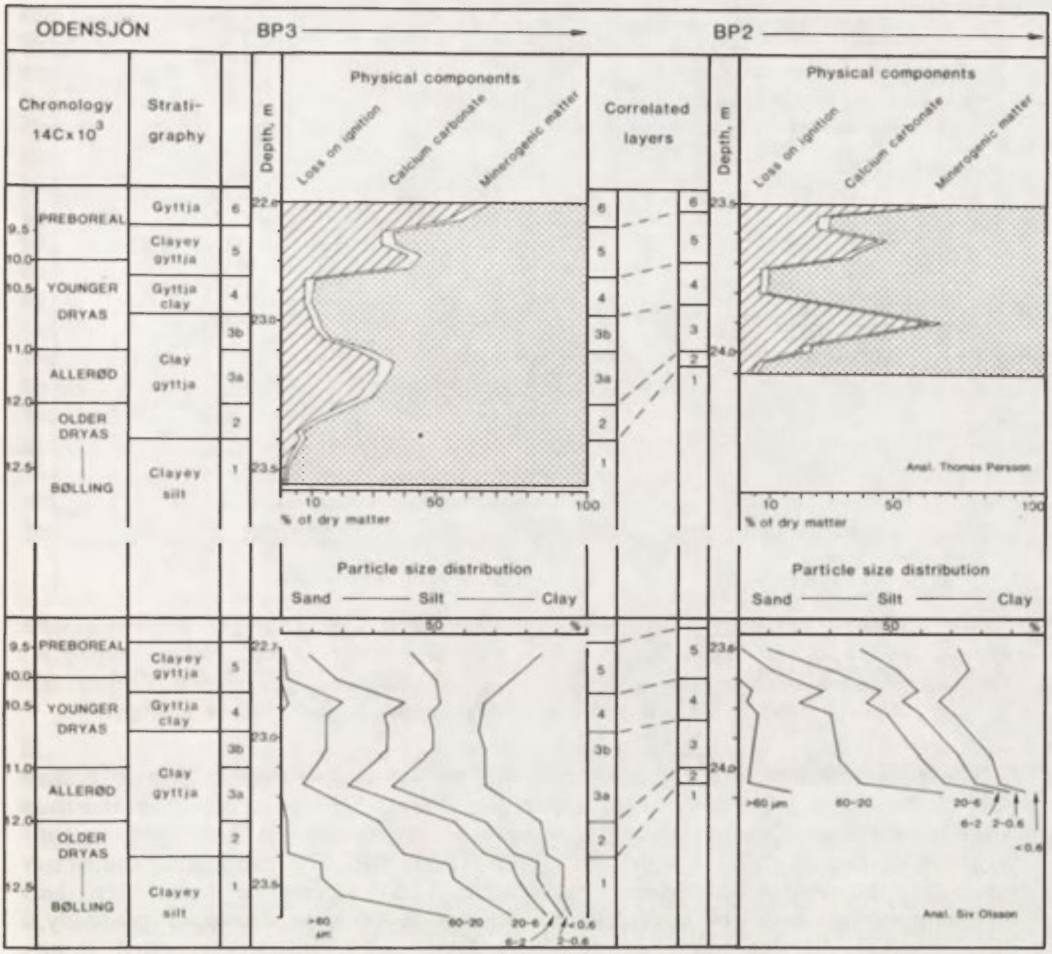


Fig. 14. Physical component and particle-size stratigraphy of the sediment cores BP2 and 3. Ocular sediment correlation between the cores. Chronology based on pollen diagram in Fig. 15

ANCIENT LAKES ON KULLABERG

Several ancient lakes have been studied on the horst ridge Kullaberg (Berglund 1971; Berglund et al. 1984) and current research of the lateglacial sediments is to be published in the near future (Lemdahl, Liedberg-Jönsson). The work has been concentrated to two basins: Hakulls Mosse about 125 m a.s.l. with a lake size of 70 × 200 m, and Björkeröds Mosse about 75 m a.s.l. with a lake size of 50 × 400 m. The analyses are more complete here than at Lake Odensjön and comprise studies of sediments, pollen, plant macrofossils and insect remains. The sedimentation starts directly after deglaciation 14 000 to 13 500 BP with varved clay at the lower basin and about 12 700 with non-laminated silt at the upper basin. This means that we have a good resolution of the sediment record after deglaciation.

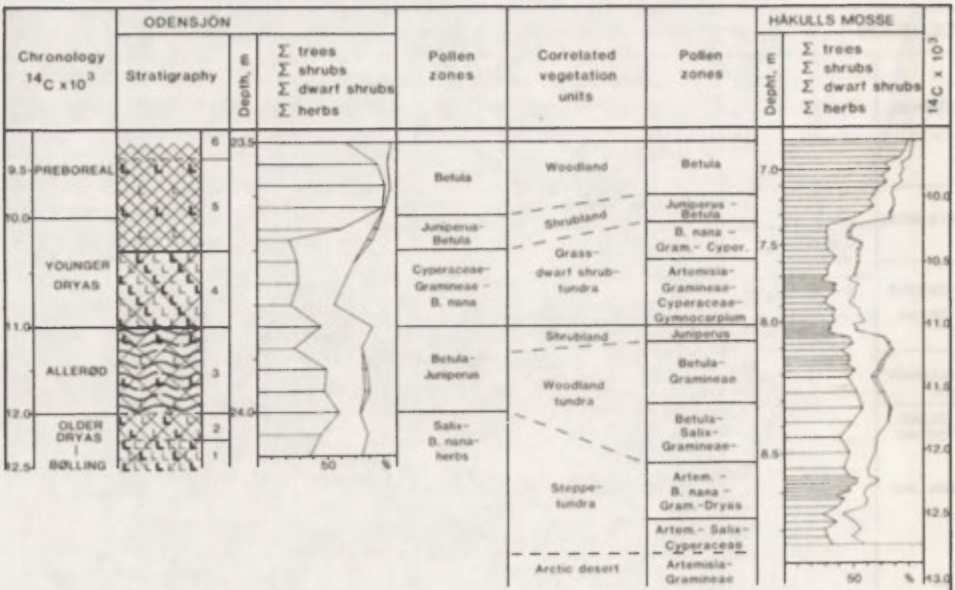

Fig. 15. Sediment stratigraphy and total pollen diagram of Lake Odensjon BP2 correlated with the Håkulls Mosse pollen diagram from Kullaberg (Berglund and Persson in prep.). The first diagram plotted against depth, the second diagram against time. Correlation of pollen assemblage zones means that general vegetation units can be identified

The palaeoecological events based on the studies at the Håkulls Mosse basin were discussed in a paper by Berglund et al. (1984). The most important results concerning lake sedimentation and soil erosion are mentioned here. Melting of stagnant ice caused instable soils until about 12 500 BP. The continuous sedimentation in the Bjorkerods Mosse basin after 13 500 seems to have been an exception in this area. In both lake catchments the erosion decreased gradually, but increased silt deposition during Older Dryas time indicates renewed slope erosion at that time. During Allerod time the minerogenic deposition decreased distinctly and silt was to a great extent replaced by clay (clay gyttja and gyttja). At the boundary Allerod/Younger Dryas there is a sudden change towards a more minerogenic sediment with increased silt content (silty clay gyttja). Strong erosion seems to have been prevailing 11 000 to 10 500 BP, particularly before 10 700 BP. Soil conditions became more stable with the gradual revegetation at 10 500 and 10 200 BP, and the sediment changed towards a clayey gyttja with decreased silt content.

## REVEGETATION AND VEGETATION COVER DURING THE LATE WEICHSELIAN

The pollen stratigraphy at BP2 of Lake Odensjon is correlated with the corresponding pollen sequence at Hakulls Mose in Fig. 15. The resolution and the dating control is by far better at the Kullaberg site, and the pollen assemblages are easy to compare. An interpretation of the corresponding vegetation units helps to describe the common vegetational changes as follows.

During the earliest deglaciation period after about 13 500 BP an arctic desert with very sparse vegetation of mainly graminids and *Artemisia* seems to have been prevailing (this is documented at Björkeröds Mosse on Kullaberg). Certainly, the ground was exposed to a great extent. At some time around 12 700 the vegetation became denser – a steppe tundra developed with grasses, sedges, dwarf-willows, *Artemisia*, *Chenopodiaceae*, many herbs, in the later part particularly *Dryas* and *Leguminosae*. Between 12 500 and 12 000 juniper, buckthorn and dwarf-birch also contributed to the wind protection. About 12 000 BP a woodland tundra became characteristic of the landscape, with more birch trees in protected areas and more shrubs and dwarf-shrubs in exposed places. At Odensjön the Alleröd sediments show unusually high values of *Juniperus*. However, in most areas a humus cover was developed during the Alleröd millenium. At 11 000 BP the transition to the Younger Dryas period meant a transition to a tree-less grass-dwarf-shrub tundra dominated by grasses, sedges, dwarf-birch, *Artemisia*, some herbs and in the Kullaberg area also ferns (*Gymnocarpium*). Probably the ground lost its earlier humus cover and some areas got exposed mineral soils. Around 10 500 there was a change towards a slightly denser vegetation with less herbs and more grasses and dwarf-shrubs. Possibly the first birch trees colonized the area again. But the real change to shrubland and woodland occurred about 10 200 BP with a *Juniperus-Empetrum* stage during a transition period of less than two hundred years (cf. Iversen 1954).

## CONCLUSIONS CONCERNING LATE WEICHSELIAN PALAEOENVIRONMENT AND CLIMATE

Our conclusions concerning the lateglacial environment and climate in NW Scania are mainly based on earlier and present studies on Söderåsen and Kullaberg, but we have also considered evidence related to regional studies of the deglaciation and the periglacial features in South Sweden (Björck et al. 1988; Lagerlund 1987: Johnsson 1981 nad later: Svensson 1980 and later. Details related to vegetation and insect fauna are under elaboration – Berglund and Persson (pollen), Liedberg-Jönsson (plant macrofossils), Lemdahl (insects). Our tentative synthesis is illustrated by the scheme in Fig. 16. We distinguish the following phases:

I. About 13 500–12 700 BP. Deglaciation 14 000 to 13 500 BP. Climate high-arctic with low temperatures, rather dry conditions. Strong wind activity, stagnant ice and permafrost prevailing. Large areas of mineral ground exposed. Arctic fauna and vegetation.

II. 12 700–12 000 BP. Climate subarctic to boreal of continental character, probably with high summer temperature, dry conditions particularly around 12 200–12 050 (Older Dryas) which is interpreted as a short event with cold winters, dry summers and high wind activity. Still stagnant ice in low-lying areas with thick till layers. Slope wash at least at the end of the period. Steppe tundra with nitrogen-fixating plants and with boreal fauna. Distinct time-lag for the revegetation (cf. Berglund et al. 1984; Pennington 1986; Prentice 1986).

III. 12 000–11 000 BP. Climate subarctic and humid with moderate temperature, alternating dry and humid periods. Melting of stagnant ice at the beginning. Progressive vegetation succession leading to woodlands and heaths with a protecting humus cover. Boreal fauna changing to subarctic at the end of this period.

IV. 11 000–10 500 BP. Climate arctic with low temperatures and humid conditions. Strong wind activity and slope wash. Nivation processes leading to formation of large snow-fields and small cirque glaciers. Arctic to subarctic tundra, in the coastal NW-part with suboceanic elements. Arctic fauna.

V. 10 500–10 200 BP. Climate subarctic with increasing temperature at the

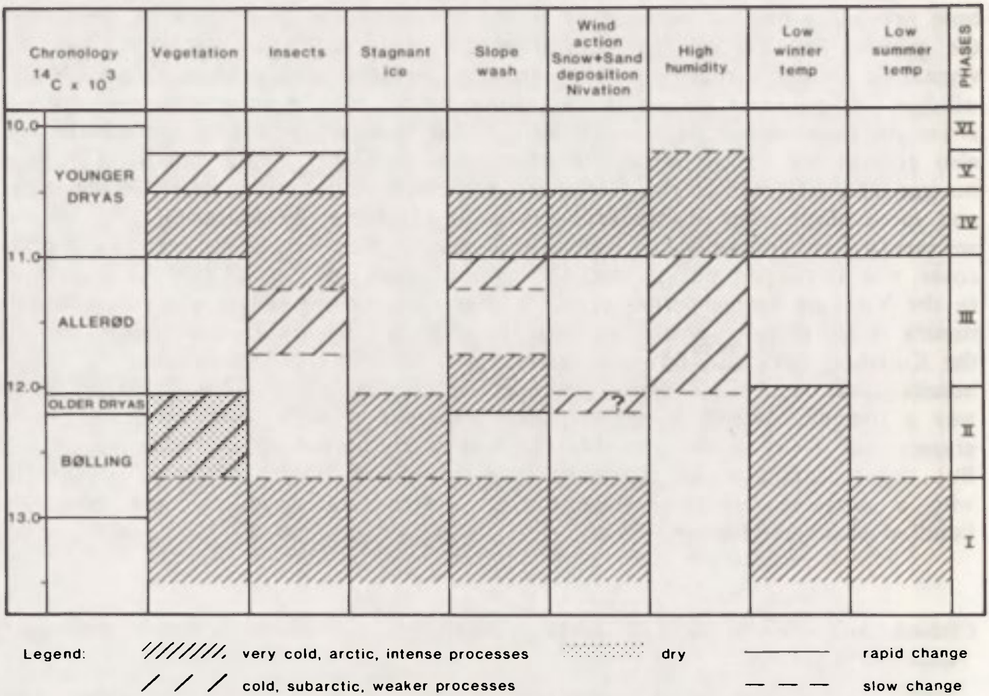


Fig. 16. Survey scheme showing a correlation of environmental changes in NW Scania during the Late Weichselian

Interpretation based on the following information: Vegetation (Berglund, Liedberg-Jonsson, Persson), insects (Lemdahl), stagnant ice (Berglund, Björck, Lagerlund), slope wash (Berglund, Rapp), wind action etc. (Rapp, Lagerlund). Palaeoclimatic interpretation leads to distinguishing of six phases, discussed in the text

beginning. Decreasing wind and water erosion. Subarctic heath and shrub vegetation together with subarctic fauna. A protecting humus cover formed.

VI. 10 200–9500 BP. Climate warm-temperate with temperatures higher than today. Lower humidity particularly around 10 000 BP. South-boreal fauna but north-boreal forest vegetation indicating time-lag for the revegetation (cf. Berglund et al. 1984).

The palaeoclimatic interpretation may be seen as an alternative to the climatic reconstruction by van Geel and Kolstrup (1978). There is also an agreement with the recent reconstruction by Atkinson, Briffa and Coope (1987) for Britain although the situation of Scania close to a large inland-ice cap be a reason for a deviating climatic development before 12 000 BP. Further multidisciplinary research is needed in the Peribaltic region for a better understanding of the climate and its impact on geomorphological and biotic processes during the Late Weichselian.

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# NEW ASPECTS ON THE DEGLACIATION CHRONOLOGY OF SOUTH SWEDEN

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ABSTRACT. The absolute dating of the deglaciation in South Sweden has been based on a number of different methods. Among those are $^{14}C$-dates on a) marine molluscs, algae, mammals, and sediments, b) limnic sediments and macrofossils, and c) terrestric macrofossils and mammals. Other methods have been to use varved clay connected with the present or local varve-chronologies added to the $^{14}C$-chronology. Correlations between the differently dated regions have therefore appeared to be difficult. A correlation method based only on glacial deposits and striations indicated a time-discrepancy between the differently dated deglaciation ages on the west and east coasts. Recent research on the difference between $^{14}C$-years and varve-years during the Late Weichselian suggests that much of the found differences between differently dated regions can be explained by steadily increasing $^{14}C$ production during the deglaciation of South Sweden. The hitherto available data indicate that the time-scales "meet" sometime between 12 700 and 12 800 BP. During the following millennia the gradually higher $^{14}C$-activity led to gradually younger radiocarbon dates (relative to varve dates). A partly new, preliminary deglaciation chronology, taking these new data into account, is presented with correlations between the west and east coasts. Differences between regions, regarding deglaciation pattern, are discussed as well as possible glaciodynamic and climatic reasons for these anomalies.

## INTRODUCTION

Different deglaciation chronologies have been presented for southern Sweden during the last decades (see Berglund 1979). The original varve-chronologically dated deglaciation maps (e. g. Lundqvist 1961; Nilsson 1968) were later complemented and/or revised by $^{14}C$ based deglaciation chronologies (e.g. Mörner 1969, 1970; Berglund 1976, 1979; Berglund and Mörner 1984). A major problem with all these deglaciation chronologies was, however, to successfully correlate between supposedly synchronous ice-marginal zones on the east and west coasts.

Lagerlund et al. (1983) presented an activity-limit recession map where correlations were based solely on interpretations of glacial deposits and striations. These glaciodynamic correlations indicated that the $^{14}C$-chronology on the Swedish west coast, mainly based on dates on marine molluscs (e.g. Wedel 1969; Mörner 1969; Fredén 1975; Håkansson 1975, 1977; Hillefors 1975; Lagerlund 1980a), is 400–500 years older than the chronology on the south-east coast (Björck 1979, 1981, 1984). The latter was based on a combination of $^{14}C$-dates on lake sediments and aquatic mosses and the local varve-chronology. Different explanations for these discrepancies were put forward but without satisfying answers.

There is, however, a striking difference in the way the deglaciation chronology is built-up on the west and east coasts (Fig. 1). In the west the ice-sheet receded in a marine environment. This meant that deglaciation of an area was quite rapidly followed by an immigration of marine organisms such as molluscs. If the molluscs' stratigraphic position, in relation to the deglaciation, is clear, a rather good comprehension of the deglaciation age can be obtained by $^{14}C$-dates on these molluscs. Some reservation on the accuracy of the "sea correction" must, however, always be considered on this type of dated material (cf. Berglund 1979).

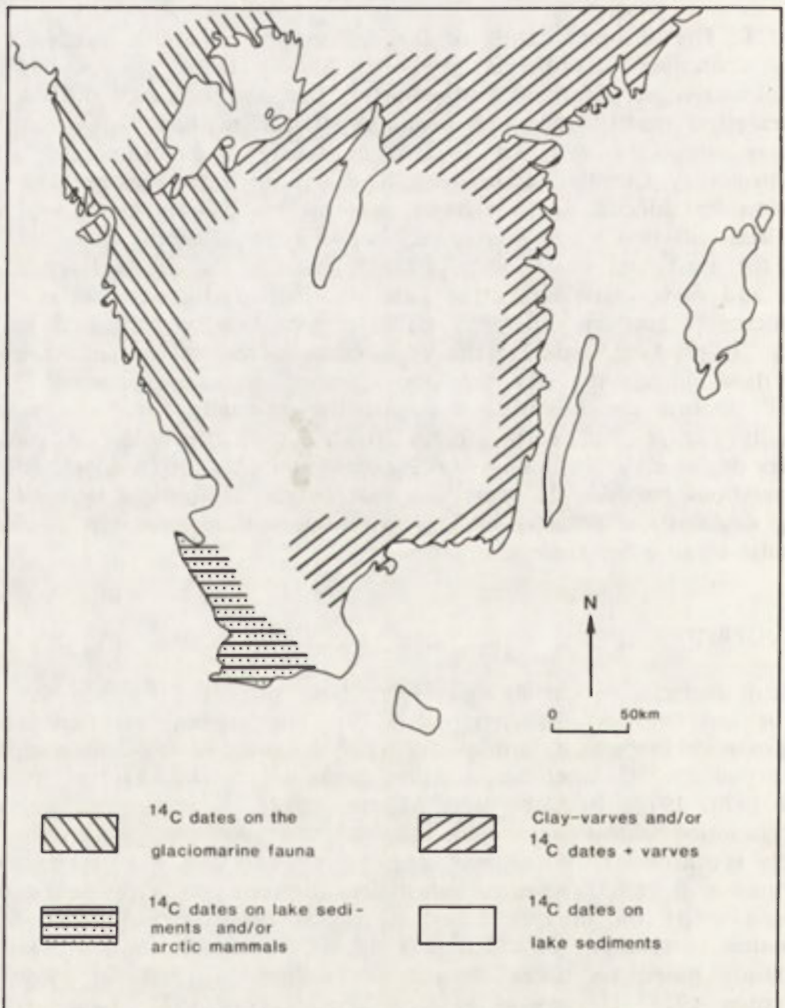


Fig. 1. Map of South Sweden showing how the deglaciation in different regions has been dated

On the east coast the deglaciation has been relatively dated by local varve-chronologies (e.g. Ringberg 1971, 1979; Rudmark 1975; Ringberg and Rudmark 1985; Kristiansson 1986) and absolutely dated by adding local varve-years to the regional $^{14}C$ based pollen stratigraphy (Björck 1979, 1981, 1984).

In the large areas between the highest coast-lines on the west and east coasts the deglaciation has been dated by correlations between the coastal areas or by $^{14}C$-, and pollen-analysis of the oldest datable lake sediments. This has often only given a minimum value of the deglaciation age.

The complexity of dating methods and the uncertainties of the dating methods themselves would of course seem to be the natural explanation for the above mentioned discrepancies and correlation problems. But it will be shown that the possibly largest source of error lies in the relationship between varve-years (calendar years) and radiocarbon years.

## VARVE-CHRONOLOGY VERSUS RADIOCARBON-CHRONOLOGY

When Cato (1985) presented his connection of the Swedish geochronological time-scale (based on varved clay) with the present, an important piece of the deglaciation chronology was still not possible to connect with any certainty, namely the varves from the Middle Swedish end-moraine zone and southwards (Strömberg 1985). Unfortunately neither Kristiansson (1986) was able to connect his varve-series in eastern Småland and Östergötland to varve-series north of the Middle Swedish end-moraine zone. He found, however, that he could approximately correlate a sequence of varves, around the local varve year 1760, to the beginning of the Younger Dryas cooling which is radiocarbon dated to c. 11 000 $^{14}C$ years BP.

By magnetostratigraphic studies (Sandgren et al. in press) on one of Kristiansson's (1986) sites and another site further northeast, which is preliminary connected with the Swedish geochronological time-scale (Brunnberg 1986), it was, with statistical correlations and geological reasoning, possible to connect the varve based time-scale with varve-series south of the Middle Swedish end-moraine zone (Björck et al. 1987). These studies show that around 10 300 $^{14}C$ years BP the varve-chronology exceeds the radiocarbon-chronology with 500 varves, while at 11 000 $^{14}C$ years BP, thus corresponding to Kristiansson's local varve year 1760, the difference is 300–400 varves. Within the 1050 varves preceding varve year 1760, Kristiansson's (1986) varve-diagrams show a very marked, uniform, and unusal increase (varves usually thin out with time) in varve thickness (Fig. 2) between the local varve years 2570 and 2550. This probably corresponds to the climatic amelioration at the beginning of the Allerød interstadial (Björck

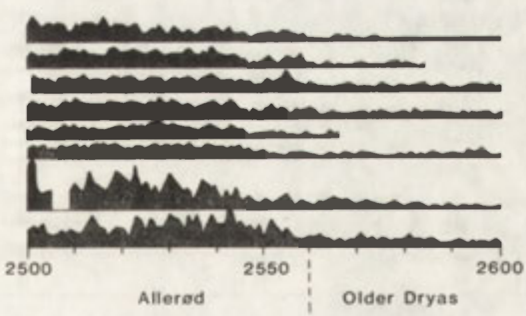


Fig. 2. Eight varve-diagrams from eastern Småland redrawn from Kristiansson (1986). The varves cover the local varves 2500 to 2600 (see Fig. 3). The thickness scale is 1:8

and Moller 1987) with an increased influx of melt-water transported material. From varve-chronologic viewpoint these interstadial conditions would have lasted c. 800 varve-years (2560–1760). However, according to Björck (1984) and Björck and Moller (1987) Allerød lasted 1000–1100 $^{14}$C-years. This difference is also in accordance with what Björck et al. (1987) found regarding the relationship between varve-years and $^{14}$C-years during the later part of the Late Weichselian, i.e. that one varve-year corresponds to 1.2–1.3 $^{14}$C-year. If this ratio for the relationship varve-years/$^{14}$C-years is valid for the whole period during which S Sweden was deglaciated it certainly has meant much for the difficulties in establishing a uniform deglaciation chronology since this is based on different dating methods (Fig. 1).

## A FURTHER EXTENSION OF THE VARVE-CHRONOLOGY

What we have shown above is that the varve-chronology, with the combination of magnetostratigraphic correlations and Kristiansson's (1986) long varve-series, can be extended to pre-Allerød time. South of Kristiansson's study area two local varve-chronologies (Ringberg 1979: Rudmark 1975) exist and have also been connected with each other (Ringberg and Rudmark 1985). Kristiansson (1986) never tried to connect his time-scale with their time-scale of SE Småland-Blekinge as he possibly never thought the time-scales would overlap each other. However, if the younger parts of Ringberg's (1971, 1979 and the unpublished diagram from Karlshamn) varve-series are compared with the oldest parts of Kristiansson's

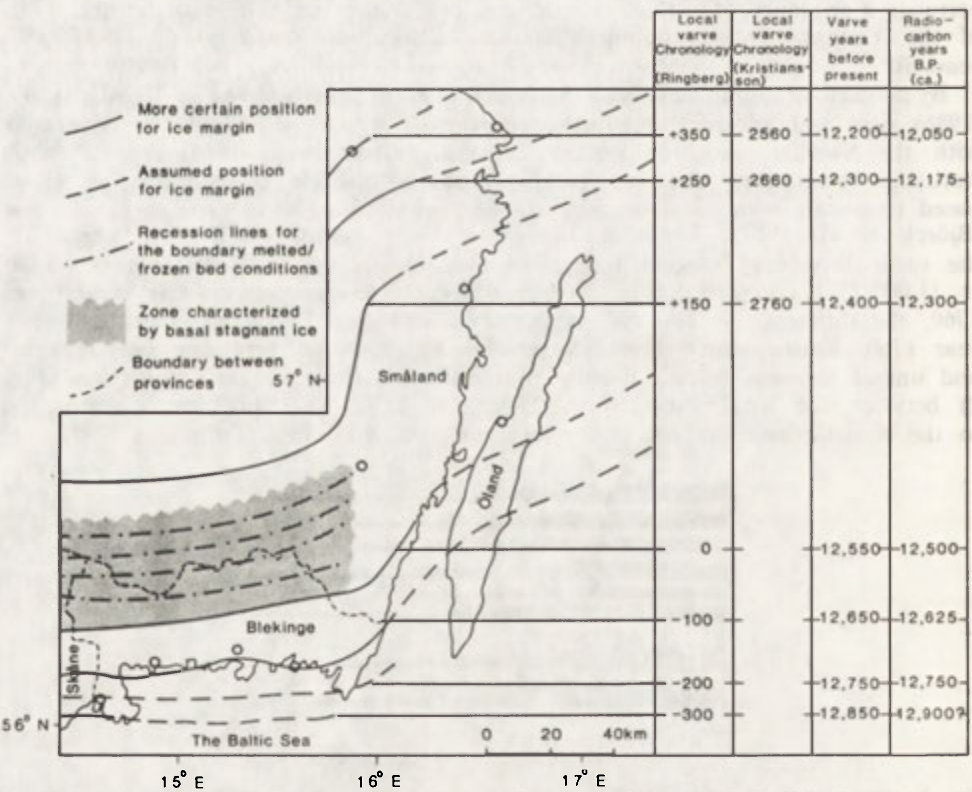


Fig. 3. Map showing the deglaciation chronology of Blekinge, S Småland, and areas below the highest shoreline along the east coast of Småland as concluded by Björck and Moller (1987)

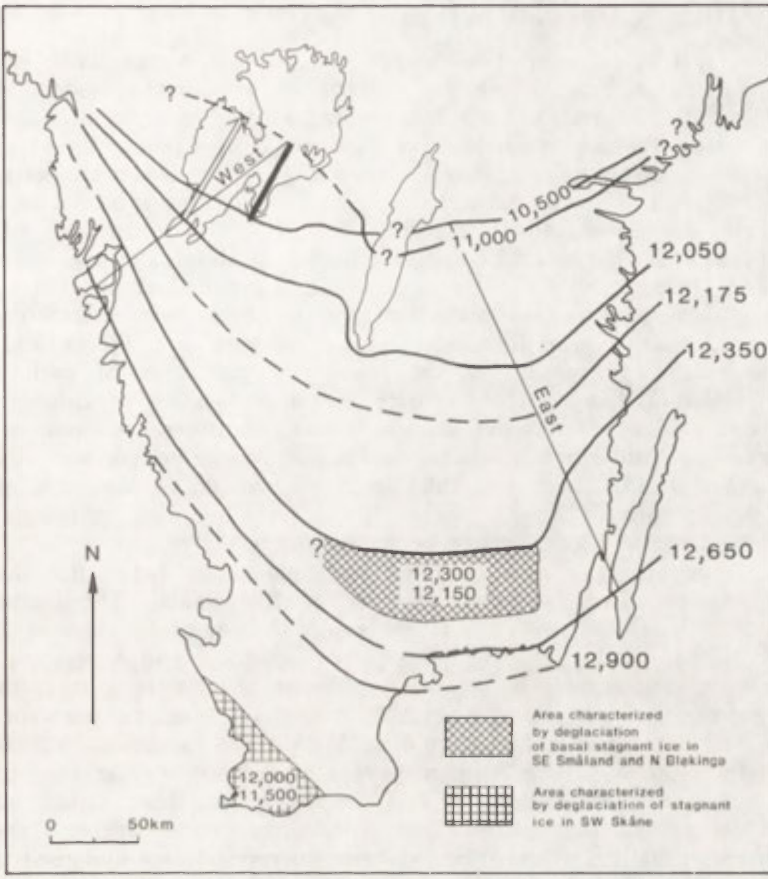


Fig. 4. The deglaciation chronology of South Sweden based on $^{14}$C-years BP (on the east coast transformed from the local varve-chronologies). The drawn ice-recession lines over the South Swedish Uplands are based on the concepts of Lagerlund et al. (1983). Note that the chronology in Skåne is not presented here (cf. Lagerlund 1987). The two transects (east and west) of Fig. 5 are indicated by thin lines between the ice-recession lines 12 650 and 10 500 BP

(1986) varve-series there is a striking agreement between the shaping of the varve-series from Småland and Blekinge if Ringberg's (1979) local year + 100 in Blekinge is approximately connected with the local year 2800 in Småland (Kristiansson 1986). This preliminary connection (Björck and Möller 1987; B. Ringberg pers. comm.) shows that the deglaciation of the coast between southernmost Blekinge and central Småland, a distance of c. 100 km, only took 200 varve years (Fig. 3). If the ratio between varve-years and $^{14}$C-years was the same during this period as later on (see above) the time-scales would "cross" each other at c. 12 750 varve- or $^{14}$C-years BP (Figs. 3–4). This "calibration" of the $^{14}$C-chronology reduces the difference between the deglaciation ages of Lagerlund et al. (1983) and Björck (1979, 1984) deglaciation ages from 400 to 200–300 $^{14}$C-years.

In northeastern Skåne Ringberg (1979) has been able to prolong his time-scale, in relation to southernmost Blekinge, 200 varves further back in time. Ringberg's southernmost equicess (– 300, see Fig. 3) is the restricting ice-recession line in the south for this partly new deglaciation chronology.

## IRREGULARITIES IN THE DEGLACIATION PATTERN

This paper will be an attempt to describe the position of the active ice-margin at various points of time during the deglaciation of South Sweden. As long as the ice receded in not too shallow water, as for example along the east and west coasts, the ice-recession lines constituted a calving ice-margin with ice-free conditions distal to the active ice-margin. However, when the ice-recession took place in shallow water or under subaerial conditions special climatic and glaciodynamic conditions could possibly lead to the formation of zones of stagnant ice outside the active ice-margin. This was postulated by e.g. Bjelm (1976) and Berglund (1976, 1979) for parts of Småland. Lagerlund et al. (1983) showed that certain areas, during the retreat of the ice-sheet, were characterized by formation of deposits typical for deglaciation of stagnant ice. Björck and Möller (1987) described the formation of the possibly largest area of such deposits in South Sweden and were able to date the down-wasting of this area (Fig. 3). It became evident that when this hummocky/transverse moraine area was more or less free from stagnant glacier ice the active ice-margin was situated c. 100 km further north. There was thus, during a couple of hundreds of years, a gradully wider zone of ice-free terrain between the receding ice-margin in the north and the large area of stagnant ice in the south.

Another large area that was occupied by stagnant ice long after the active ice-margin left the area is the main part of SW Skåne. The landscape is dominated by a hummocky or rolling topography dominated by clays or clay-tills. Ever since Nilsson (1935) described the pollen zonation and vegetation history in Skåne since deglaciation it has been obvious that it is guite difficult to find lake sediments from pre-Allerød time in the southwestern parts of Skåne, both below and above the highest shore-line. With a few exceptions (e.g. Berglund and Digerfeldt 1970, and later Sandgren 1986) lake deposits older than c. 12 000 $^{14}$C-years have been impossible to find. In many of these small, nowadays overgrown, basins the bottommost layer constitutes coarse organic (and often also minerogenic) matter, which often has been interpreted as a collapsed terrestric soil originally formed on top of the stagnant ice (cf. the "trash-layers" in the Midwest of N America, see e.g. Florin and Wright 1969). A possible reason for the extreme delay of melting of stagnant ice even below the highest shore-line can possibly be explained by a combination of high debris-load in the stagnant ice and the following transgression (Lagerlund 1980a), which covered the stagnant ice with Lomma Clay/Lund Till (Lagerlund 1980a, b; Berglund and Lagerlund 1981). This first prevented the ice from floating away, and when covered by clay, from rapid melting (cf. Lagerlund 1987). After the regression soils were formed on these clays/tills on top of very slowly melting areas of buried stagnant ice. Around 11 500 $^{14}$C-years BP most of the stagnant ice in SW Skåne had possibly melted, but by that time the active ice-margin of the retreating ice-sheet was situated c. 300 km further north. This is also discussed by Berglund and Rapp (1988).

Apart from these rather large zones there are also many smaller areas in South Sweden characterized by the same type of deposits indicating that stagnant ice was left behind when the ice-sheet retreated. However, since these have not been studied in any detail regarding time of deglaciation they will not be included in the deglaciation chronology map.

## CORRELATIONS BETWEEN THE WEST AND EAST COASTS

It shall be noted that in the following text the deglaciation chronology will be based on $^{14}$C-years BP if not otherwise stated.

Based on both the new chronology and palaeoclimate Björck and Möller

(1987) concluded that the complex Göteborg Moraine (Hillefors 1975) on the west coast corresponds approximately to the time when the ice-margin was situated just south of Blekinge, i.e. approximately between Ringberg's (1979) equicesses – 100 to –300 (Fig. 3). The following 500 years (Bølling) were characterized by a rather mild climate (see e.g. Björck and Möller 1987) favouring a rapid deglaciation. The ice-recession along the Swedish south-east coast was very rapid (Fig. 3) and the icesheet seems to have more or less collapsed in the Baltic basin (Lagerlund et al. 1983; Björck and Möller 1987; Lagerlund 1987). Around 12 400 BP the active ice-margin was situated at the boundary between the hummocky/transverse moraine in the south and the streamlined/drumlinized terrain in the north (Björck and Möller 1987). On the west coast the ice-margin at that time seems to have been located at the Berghem Moraine (Berglund 1979; Hilldén 1979; Lagerlund et al. 1983). The rapid deglaciation in the eastern part of South Sweden seems to have continued, both on land and in the Baltic basin. The previously rather mild climate during the Bølling interstadial was replaced by more arid and cool conditions (Older Dryas) for 100–150 years. According to Kristiansson's (1986) varve-chronology the ice-recession on the east coast was rather slow during this period. The ice-margin position in the west is, however, more uncertain, but it seems to have been situated between the Trollhättan and Levene Moraines. This is based on the fact that the ice-margin in the east was situated at the Vimmerby line, correlated to the Levene Moraine (cf. Berglund 1979), at the end of this period. Björck and Digerfeldt (1982a, b) have also shown that the ice-margin possibly retreated from the Levene Moraine at or slightly before 12 000 BP, and Dennegård et al. (1987) have shown that this ice-marginal line south of Lake Vänern is characterized by a readvance, possibly correlated to the Older Dryas stadial, before the ice receded north of the Levene Moraine.

During the following 1000 years (Allerød) the ice-recession rate was moderate both in the west and in the east. In the east this can be concluded from Kristiansson's (1986) varve-chronology (the local varve-years 2560–1760, corresponding to c. 1000 $^{14}$C-years), while in the west this conclusion can be drawn on Björck and Digerfeldt's (1984, 1986), Dennegård's (1984) and Johansson's (1982) stratigraphic studies in the Mt. Billingen area, southern and western areas of Lake Vänern, respectively. According to their studies the ice-sheet receded rather far north of the Middle Swedish end-moraine zone during late Allerød time, and the Baltic Ice Lake was drained for the first time (Björck and Digerfeldt 1984). Exactly how far is difficult to say, but possibly at least 20 km north of Mt. Billingen (Björck and Digerfeldt 1984, 1986). It seems as if Lake Vänern acted as an efficient calving-bay with water-depths ranging between 120–180 m during the later part of Allerød.

At the end of Allerød (c. 11 000 BP) the ice-margin in the east was situated at the southernmost parts of the Middle Swedish end-moraine zone (Kristiansson 1986), while its exact position in the west is much more uncertain. During the following 500 years, corresponding to the Younger Dryas cooling, there was a significant readvance in the west (Björck and Digerfeldt 1984) and a very slow ice-recession in the east to the northern parts of the Middle Swedish end-moraine zone (Kristiansson 1986). The readvance in the west reached the southernmost parts of the end-moraine zone (the Skövde Moraine), and the Baltic Ice Lake was once again dammed (Björck and Digerfeldt 1984).

Around 10 500 BP a general, rather rapid, ice-recession began both in the west (Björck and Digerfeldt 1984) and in the east (Kristiansson 1986) resulting in a dramatic drainage of the Baltic Ice Lake and a c. 1000 year long contact (Björck 1987) between the sea in the west and the Baltic in the east through the Närke Straits, i.e. the so-called Yoldia Sea.

The correlations of the ice-margin position at 11 000 and 10 500 BP thus show that most of the ice-marginal formations of the Middle Swedish end-moraine zone in the east were formed between 11 000 and 10 500 BP during a generally very slow ice-retreat with occasional ice-oscillations (Kristiansson 1986). In the west, however, the ridges of the end-moraine zone (the Skövde and Billingen moraines) were formed between c. 10 500–10 400 BP i.e. at the beginning of the final retreat from the lowlands between the Vanern and Baltic basins. This also implies that previous correlations (summarized by Berglund 1979) within the end-moraine zone were erroneous and that the most distal parts of the zone in the west shall be correlated to the more proximal parts in the east. The missing link between these eastern and western parts of the end-moraine zone is possibly situated in the Lake Vattern area. Waldemarsson (1986) reported two glacial readvances in the Vattern basin, but as long as they neither can be directly correlated to events in the surrounding region nor be absolutely dated, any correlations to the Younger Dryas readvance in the west are merely more than vague assumptions.

The ice-margin possitions in the Baltic basin during the above discussed time-span is difficult to establish before any varve-chronological studies have been performed on the Late Weichselian sediments of the Baltic Sea. This is, however, under way (B. Ringberg, pers. comm.), but until that work is completed conclusions have to be drawn from other data. From Björck and Möller's (1987) studies (Fig. 3) it is obvious that the ice-recession along the east-coast was extremely rapid and that the ice-recession lines point northeastwards indicating an early deglaciation in the Baltic. The same is suggested by the pollen diagram of Pahlsson and Bergh Alm (1985) from the oldest found sediments NW of the island of Gotland, which suggests that this part of the Baltic was deglaciated before Allerød. In the Gavle area, c. 150 km NNW of Stockholm, G. Lundqvist (1963) found lumps of organic material in till deposits. The $^{14}C$ dates of this material (Lundqvist 1963) indicate that it was formed during late Allerød before the readvancing glacier incorporated it in its till. This suggests that the ice-margin receded into the Gulf of Bothnia during late Allerød and that the deglaciation during Bølling-Allerød in the Baltic basin was extremely rapid, indicating that this part of the ice-sheet experienced a downdraw (hinted at by Lagerlund et al. 1983). In large areas north of the 12 650–12 700 line in Fig. 3 the water was 2–300 m deep (and occasionally even deeper) favouring intense calving. It is also worth mentioning that, since the Baltic Ice Lake was dammed (Björck 1979), the regression was slow and thus a substantial water-depth was maintained during the greater parts of the deglaciation. The above mentioned possible downdraw of the ice-sheet in the Baltic basin was, during Younger Dryas, most likely followed by a significant readvance.

The correlations discussed above are summarized in Fig. 4 showing the probable positions of the active ice-margin during certain points of time.

## SOME GENERAL CONCLUSIONS ON THE DEGLACIATION PATTERN

As already indicated above there seems to be some general differences between western and eastern Sweden regarding the deglaciation pattern. This is also obvious from the map in Fig. 4, especially with respect to the ice-recession rate. In Fig. 4 two sections on the west and east coast, respectively, are drawn from the 12 650–12 700 BP line to the 10 500 BP line. These are then converted to a time-distance diagram for the retreating ice-sheet (Fig. 5). From this diagram it is clear that the ice-recession was not only much more rapid in the east, but also did the ice-sheet behave quite differently on the west and east coast, respectively, during the Younger Dryas cooling. Even in a smaller scale there are significant differences, regarding deglaciation pattern, between various areas in South Sweden. These areas are approximately shown in Fig. 6 (excl. Skåne) and their rough deglaciation characteristics are outlined below:

Area a) This area, which consists of the southeastern and eastern coast-land of South Sweden, is in rough terms situated below the highest coast-line of the Baltic Ice Lake. It was characterized by rapid deglaciation with a calving ice-front. No significant ice-margin deposits or ice readvances have been reported from this area.

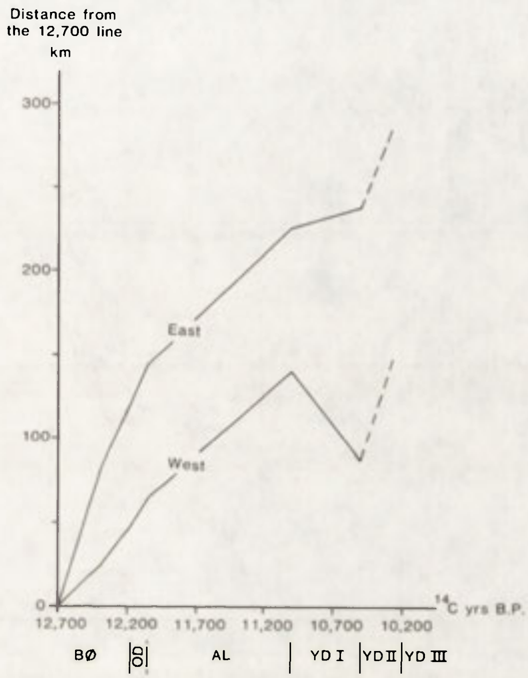


Fig. 5. The ice-margin position between c. 12 700 and 10 200 along two transects (see Fig. 4), related to the 12 650–12 700 BP ice-recession line

Area b) This is the southwest and west coast areas of South Sweden. It is approximately situated below the marine limit of the Late Weichselian Kattegatt Sea. The deglaciation of this area was rather slow with a calving ice-front. Significant ice-margin deposits with ice-oscillation sequences are common.

Area c) This is the area above the highest coast-line in eastern South Sweden. The uncertain limit towards area d in the west is indicated by a dotted line in Fig. 6. The deglaciation was characterized by both stagnant glacier ice (mainly in the south, see Fig. 4) and an ice-sheet margin coinciding with the activity limit. A few ice-margin deposits occur but no ice readvances during the Late Weichselian have been reported. As a whole the deglaciation was rapid.

Area d) This is mainly the area above the marine limit in the western parts of South Sweden. The deglaciation rate was moderate. Some significant ice-margin deposits occur and ice-oscillation sequences have been reported.

Area e) This is mainly the eastern part of the Middle Swedish end-moraine zone. The main part is situated below the highest coast-line. Major ice-margin deposits occur. The ice-recession was first very slow (see Figs 4–5), with minor ice-oscillations, followed by rapid deglaciation.

Area f) This area has been regarded as the western part of the Middle Swedish end-moraine zone. It is situated both above and below the marine limit (or highest coast-line east of Mt. Billingen). Ice-margin deposits occur rather frequently. A fairly rapid deglaciation was followed by a significant readvance, which in turn was succeeded by rapid deglaciation (see Figs. 4–5).

Area g) This area corresponds to Lake Vättern and its immediate surroundings.

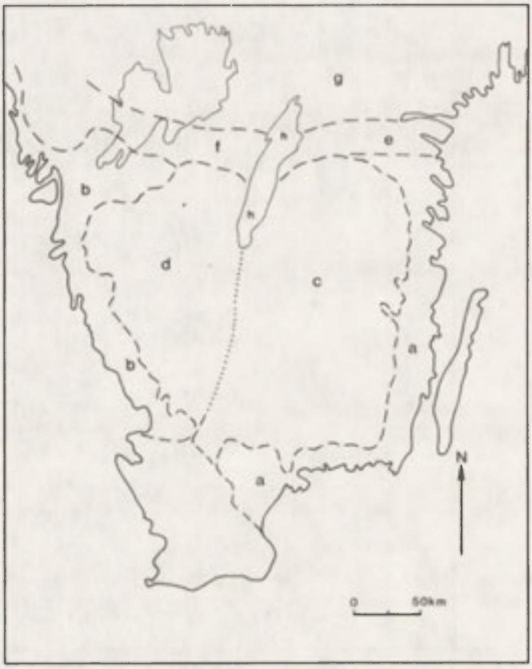


Fig. 6. South Sweden divided into 8 different regions (a-h), excl. Skåne, characterized by different deglaciation features. The characteristics for the regions are described in the text. The boundary between regions c and d is very preliminary

The lake itself is deep (max. 119 m) and in the southern part of the basin 375 m of unconsolidated sediments have been reported (Axberg and Wadstein 1980). In the south the surrounding topography reaches more than 200 m above the lake surface. This shows that the bedrock topography is almost of an alpine character. Extensive and thick deposits of glacial origin occur both west (Norrman 1971) and south (Waldemarsson 1986) of the lake. Waldemarsson (1986) found strong evidences for two, possibly late-glacial readvances in the Lake Vättern basin.

From these facts we can draw some main conclusions about the deglaciation pattern and dynamics of the ice-sheet during the deglaciation of South Sweden:

1. The deglaciation was more rapid in the east than in the west. Along the east coast the water-depth was greater than in the west (due to the damming of the Baltic Ice Lake), in the Baltic basin the ice-recession rate was possibly accelerated by a downdraw, and the precipitation (= snow accumulation) was probably generally lower in the east owing to a greater continentality. Cool

and dry high-pressure systems, created by the ice-sheet and the enormous amounts of icebergs in the Baltic (produced during the downdraw), possibly blocked the western maritime air-masses.

2. The ice-sheet seems to have been more active in the west than in the east, indicated by e.g. the abundance of (minor or major) glacier oscillations and ice-marginal deposits on the western side of South Sweden. This greater activity could be explained by more maritime conditions (Sugden and John 1976) along the west coast during the deglaciation.

3. In all the three main depressions (the Baltic Sea, Lake Vättern, and Lake Vänern) the fluctuations of the ice-margin position were much more extensive and rapid than in the surrounding regions. These rapid fluctuations could be explained by downdraws and surges. The latter has been suggested by Lagerlund (1980a) for the Baltic and by Waldemarsson (1986) for Lake Vättern.

4. In areas with debris-rich or sediment-covered stagnant ice the local/regional deglaciation was delayed for a shorter or longer time owing to climatic and/or glaciodynamic reasons.

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# PRELIMINARY STRATIGRAPHIC STUDIES ON THE LATE WEICHSELIAN AND HOLOCENE DEVELOPMENT OF THE HANÖ BAY, SOUTHEASTERN SWEDEN

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ABSTRACT. Marine hydroacoustic and stratigraphic studies were carried out on Late Quaternary deposits in the Hanö Bay, combined with sampling and $^{14}C$ datings of pine stumps and peat deposits, pollen analyses of peat and soil horizons and mapping of the sea-bottom by divers. It shows that major water-level changes of the Baltic have been the most important factor for understanding the occurrence and absence of deposits in the Hanö Bay. The most spectacular phases in the history of the area is a "pine-phase" between c. 9700–9300 BP when forest spread to areas which today are situated c. 40 m below sea level and a "peat-phase" between c. 8800–7000 BP when peat accumulated rapidly inside the shore bank of that time in areas which today are situated between −7 and −14 m. These quite dramatic water-level changes caused extensive erosion and resedimentation which have resulted in a stratigraphy characterized by many and long hiatuses. No evidences for recent sedimentation of fine-grained sediments have been found in waters of less than c. 60 m water depth.

## INTRODUCTION

While quite detailed studies have been carried out recently on the Quaternary stratigraphy on the Swedish mainland west (e.g. Ringberg 1979; Åmark 1984) and north (Ringberg 1971; Björck 1979, 1981; Lagerlund and Björck 1979; Björck and Möller 1987) of the Hanö Bay (Fig. 1) very little is known about the stratigraphy in, and the history of, the Bay itself.

From shore-line studies further north along the Swedish east coast it has been assumed that the earliest Holocene shore-lines cross today's sea level north of the Hanö Bay region. The existence of these submarine shore-lines have been confirmed from studies along and outside the southeastern (e.g. Gudelis 1979; Veinbergs 1979) and southern (Duphorn 1979; Rudowski 1979; Kolp 1986) coasts of the Baltic. It should thus not be surprising to find evidences for low water-levels in the Hanö Bay region. However, when supposedly rooted pine stumps were found on the bottom of the Bay down to perhaps as much as − 50 m and dated (Håkansson 1972, 1974, 1976, 1982) much scepticism was raised against the palaeogeographic consequences these findings could indicate.

Previous investigations around the island of Bornholm (Kögler and Larsen 1979; Duphorn et al. 1979) have presented quite detailed marine stratigraphies at considerably larger water depths than in the Hanö Bay. But it was quite clear that neither these nor the mainland stratigraphies could explain the Late Weichselian and Holocene development of the Hanö Bay without any knowledge about Quaternary deposits in the Bay itself.

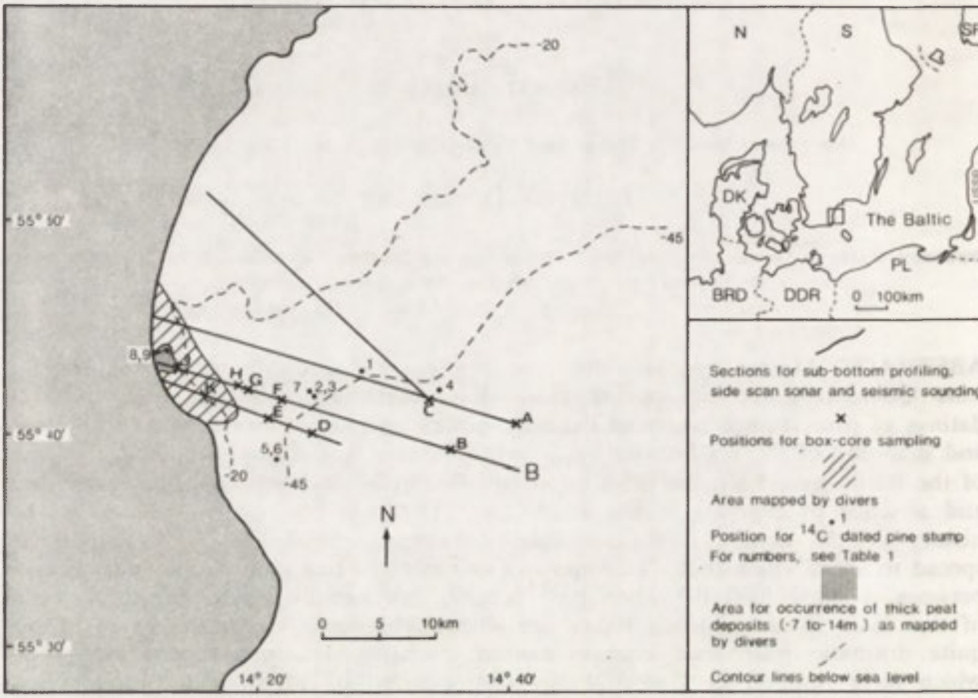


Fig. 1. Map of the study area showing marine stratigraphic sections, points for box-core sampling, dated pine stumps, area for found thick peat deposits and area mapped by divers. On the inserted map in the upper right corner the study area is related to the southern Baltic region

Owing to both the potentially good stratigraphic control around the Hanö Bay and the many puzzling finds (e.g. the pine stumps and thick peat deposits) an inter-departmental project, between a Sport Diving Club in Malmö led by Lars Hansen, the Departments of Marine Geology and Geology in Gothenburg, and the Department of Quaternary Geology in Lund, was started. The aim of the project was to investigate, both horizontally and vertically, the Quaternary deposits found in the area and relate them to neighbouring stratigraphies and development. Such correlations would then place Hanö Bay's Late Weichselian-Holocene history in a regional context.

This article is an attempt to give a broad picture of this history based on the hitherto available data from the Hanö Bay.

## METHODS

The marine hydro-acoustic measurements were conducted with a UNIBOOM system recording the frequency interval 400–1200 Hz. Complementary hydro-acoustic surveys were done using a Klein Hydroscan side scan sonar and sub-bottom profiling system. This research was carried out from a 40 feet research vessel of trawler type.

Within acoustically investigated sections surface sea-bed layers (0–50 cm) were sampled with a box-corer. This part of the study was carried out from the Swedish research vessel R/V Argos. In shallower waters (down to c. −25 m) divers made bottom surveys, detailed mapping of smaller part-areas, sampled stratigraphic sections (e.g. peat) and pine stumps and photographed interesting details.

During the offshore surveys positioning was done using a Navstar system which provides signals to a track plot record. In nearshore areas, during the diving operations, the conventional Decca system, radar and optical determinations were used for positioning.

$^{14}C$, pollen and diatom analyses were carried out according to standard procedures at the laboratories in Lund.

## RESULTS

### MARINE STRATIGRAPHY

Within the four selected and hydro-acoustically investigated sections (Fig. 1) details of the uppermost seabed layers were detected by side scan sonar and sub-bottom profilers. Deeper penetration of the Late Weichselian and Holocene sediments were achieved by the Boomer system.

The section presented in Fig. 2 gives a rather rough and general picture of the stratigraphy in the study area. It is a compilation of the hydro-acoustic studies complemented with the box-core samplings. This section runs from the shallow areas with peat-covered sand and till to areas with almost 70 m of water, where the Late Weichselian clay seems to be covered by a very thin

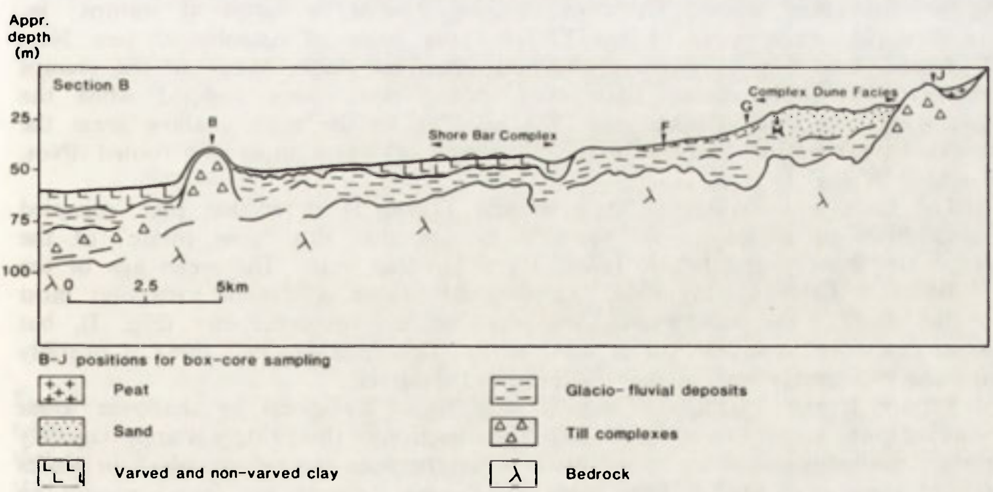


Fig. 2. Generalized stratigraphy of section B in Fig. 1, based on acoustic stratigraphy, box-core sampling and mapping by divers

(< 10 cm) layer of sand and gravel. Below follows a summary of the sampling with the box-corer (points A-J, Fig. 1):

A – 35 cm reddish-brown homogeneous clay covered by 5 cm sand and gravel,
B – 25 cm reddish-brownish grey varved clay with thick varves covered by 5 cm sandy gravel,
C – 10 cm sand,
D – 10 cm reddish-brownish grey varved clay covered by 10 cm gravelly sand,
E – 15 cm sand,
F – 15 cm fine sand covered by a cm thick layer of gravel,
G – 10 cm fine sand,
H – 10 cm coarse sand,
J – Nothing (till?),
K – 5 cm reddish-brown clay (varved?).

The ridge at point B is at least partly covered by varved clay. This ridge, which most likely is built up by till, runs from point B in a NNE direction. This ridge will hopefully be mapped in more detail during the planned continuation of the study. West of the ridge in section B (Fig. 2) thin clay beds seem to be underlain by coarser deposits, but further west the clay thickens and is partly covered by sand and gravel complexes. In the middle of this section a "channel-like" feature appears which has been identified in the other sections as well. Its western slope is always rather steep and is situated at –45 to –50 m. Just west of this channel many of the pine stump finds appear to have been found.

Further west 10–20 m thick sand complexes were found, which on the surface appear to have dune-like structures. Further inland a well-pronounced till ridge, sampled by the divers, was found in three of the sections. Inside this ridge the bottom usually consists of sand and gravel, but in section B also peat is very common either resting on till or on sand.

### TERRESTRIAL DEPOSITS IN THE HANÖ BAY

The abundance of pine stumps on the bottom of the Hanö Bay is not only documented by the many $^{14}C$ dated stumps (Table 1), but also from observations by fishermen and divers. The former have reported "a forest of stumps" in, for example, an area c. 10 km ENE of the coast of Stenshuvud (see Nos. 2, 3, and 7 in Fig. 1) where fishing-nets often get stuck. Many of the stumps seem to be firmly rooted since even fishing-vessels were stopped when the fishing-nets got stuck (Håkansson 1972, p. 386). In the more shallow areas the divers have reported frequent finds of stumps of which most are rooted (Nos. 8 and 9 in Fig. 1).

The finds and datings of these stumps (Table 1) show that pine occupied areas down to at least –40 to –50 m and that this "pine phase" of the Hanö Bay history might have lasted a few hundred years. The mean age of the 11 dates in Table 1 is c. 9540 $^{14}C$ years BP. From a statistic viewpoint most of the dated pines could have lived more or less simultaneously (Fig. 3), but some dates do hardly, or not at all, overlap. This indicates that there is possibly at least two generations of pine in the dated data set.

Stumps 8 and 9 (Figs. 1 and 3) were found by divers in shallower areas where also thick peat ridges were found. The direction of these ridges is approximately parallel to today's coastline. Sand often occurs between the ridges, which in places are at least 4 m thick. These often thick peat deposits are found resting on minerogenic deposits. The water depth for the contact between peat and sand/till varies between 7 and 14 m. At greater depths, down to at least –24 m, the divers

TABLE 1. Ages for radiocarbon dated pine stumps found on the bottom of the Hano Bay within or close to the investigation area. The dates were published in Håkansson (1968, 1972, 1974, 1976, 1982, 1986). The dated stumps within the investigation area are numbered below and positioned in Fig. 1.

| Obtained $^{14}C$ age years BP | Laboratory no. | Approximate depth of finding | No. in Fig. 1 |
|---|---|---|---|
| 9420 ± 100 | Lu – 16 | – 80 m | |
| 9750 ± 95 | Lu – 807 | – 57 m | |
| 9420 ± 95 | Lu – 890 | – 52 m | 1 |
| 9520 ± 95 | Lu – 891 | – 49 m | 2 |
| 9620 ± 95 | Lu – 892 | – 49 m | 3 |
| 9480 ± 95 | Lu – 702 | – 48 m | 4 |
| 9680 ± 95 | Lu – 1901 | – 40 m | 5 |
| 9660 ± 90 | Lu – 1900 | – 40 m | 6 |
| 9330 ± 95 | Lu – 551 | – 35 to – 40 m | 7 |
| 9590 + 90 | Lu – 2342 | – 13 to – 14 m | 8 |
| 9450 ± 90 | Lu – 2341 | – 13 to – 14 m | 9 |

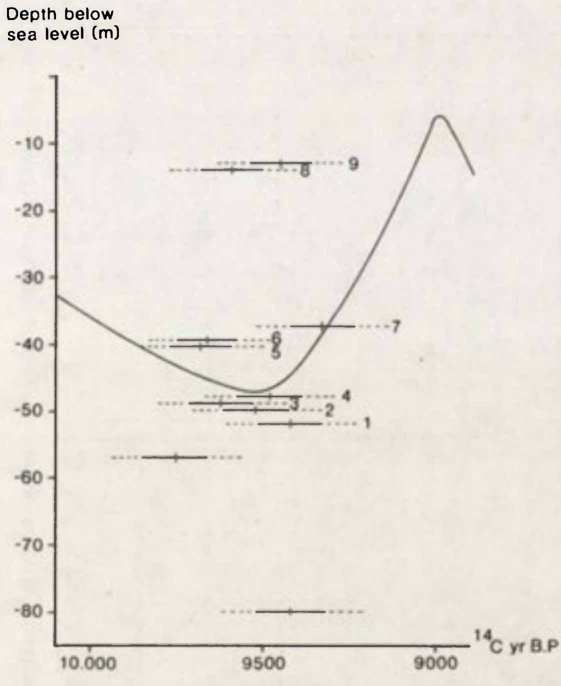


Fig. 3. Radiocarbon dated pine stumps within or close to the study area (Table 1) related to find depths and the supposed shore displacement between 9000–10 000 BP. Note that the find depths are according to the fishermen's rough estimations, which means that Nos. 1–4 could belong to a shallower find-group (see contour lines in Fig. 1). The $^{14}C$ dates were published by Håkansson (1968, 1972, 1974, 1976, 1982, 1986). Note that both single and double (dashed line) standard errors are shown

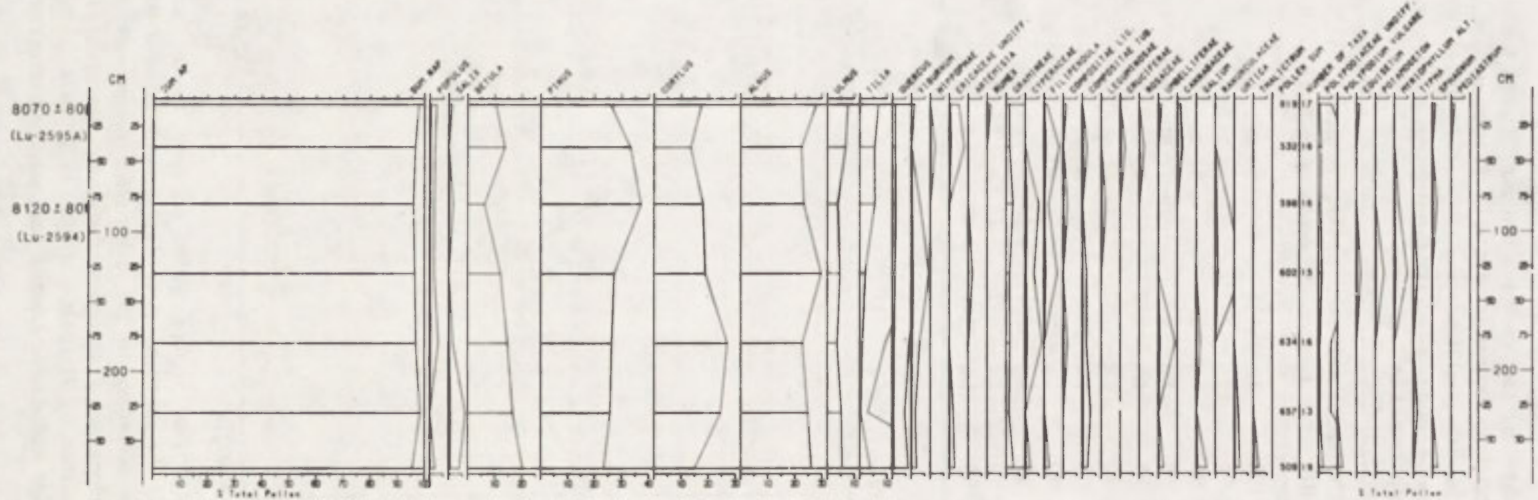


Fig. 4. A pollen diagram from a sequence of peat situated between 8.1 and 10.8 m below sea level and sampled by divers. Note the $^{14}C$ dates to the left

have found 5–10 cm thick organic sand and silt in which pine stumps are rooted. Two levels in this assumed soil were pollen analysed. No *Corylus* pollen grains were found in the lowest level, while c. 4% was found in the level above. This clearly suggests that the assumed soil was formed slightly before 9500 BP.

In order to date the thick peat deposits a 2.7. m thick sequence was sampled by the divers. Both pollen analyses and radiocarbon datings were performed (Fig. 4) and the two independent dating methods support each other. The rational limit of *Tilia* in Skane is dated to c. 7800–8000 BP (Digerfeldt 1977), which also the $^{14}C$ dates in Fig. 4 show. The dates and the pollen stratigraphy also suggest that the accumulation of peat was rapid since the rational limits of neither *Alnus* (c. 8500–8800 BP according to Digerfeldt 1977) nor *Ulmus* (slightly later than *Alnus*) can be seen in the diagram. This means that those rational limits might be found in the lowermost parts of the peat sequence, i.e. the parts that could not be sampled by the divers.

## CHRONOLOGY OF THE HANÖ BAY DEPOSITS

Since only the terrestial deposits have been absolutely dated, the chronology of the other deposits in the study area can only be relatively dated by means of stratigraphic correlations to the surrounding regions.

The two till ridges, B and J (Fig. 1), do possibly correspond to different deglaciation stages of the area. Since the ridge at sample point B is covered by a thin layer of varved clay the till itself could not be sampled. The ridge at sample point I was sampled by the divers and from petrographic analyses it appears that it is a clayey till dominated by Precambrian bedrock (74%). It also contains sandstones of different types (9%), Palaeozoic limestones (7%), shales (4%), Cretaceous rock types (3%) and others. The bedrock content of the till suggests that it is correlatable to the older till in the area (Åmark 1984; Lagerlund personal communication). On the other hand the direction of this ridge (NNW) suggests that, if the ridge was formed at the ice-margin, it could correspond approximately to the position of Lagerlund's (1980, p. 108) ice-margin at c. 13 300 BP.

On top of the formations which have been interpreted as till, coarser, possibly glacio-fluvial sediments, occur abundantly. These are in turn covered by varved and non-varved clays. The varved clay was deposited some time between 13 000 and 12 000 BP (Björck and Möller 1987) after which deposition of non-varved clay followed. Diatom analyses were performed on both types of clay and only a few fragments were found. This absence of diatoms suggests that also the non-varved clay was formed in the Baltic Ice Lake (cf. Björck 1979, 1981) and not during a later stage. After these clays were deposited in the investigated parts of the Hanö Bay there are no clear evidences for any clay or mud sedimentation in the study area in spite of the fact that the sections reached water-depths of c. 60 m.

The Hanö Bay's "pine stage" should, according to Table 1 and Fig. 3, have occurred between c. 9300 and 9700 BP, which is also supported by the pollen dating of the soil in which the pines grew.

According to the rapid accumulation rate of the thick peat deposits, indicated by the pollen stratigraphy and $^{14}C$ dates (Fig. 4), these deposits are possibly not older than 9000 BP and not younger than 7000–7500 BP. This means that these terrestrial deposits are the only certain Holocene deposits in the investigated parts of the Hanö Bay, apart from the sand (Fig. 2) and some gravels and pebbles whose genesis will be discussed below.

AN INTERPOLATED SHORE DISPLACEMENT CURVE FOR THE HANÖ BAY

In order to understand the presence and absence of different type of deposits outside Skåne's present east coast (Fig. 1) it is obviously of importance to have an idea of the Late Weichselian-Holocene shore displacement in the area. This is not only obvious from the presence of stumps and peat deposits, but also from the long hiatuses in he Hanö Bay stratigraphy.

Since no shore-line studies have been carried out in eastern Skåne a hypothetic shore displacement curve has to be constructed for the area. Such a construction has to be based on interpolations between dated shore-lines in Sweden (Berglund 1964; Björck 1979, 1981; Björck and Digerfeldt 1982a 1982b, 1984, 1986; Björck and Moller 1987; Konigsson 1968; Liljegren 1982; Magnusson 1970; Mikaelsson 1978; Morner 1969; Påsse 1983; G. Persson 1973; C. Persson 1978, 1979; Svedhage 1985; Svensson 1985) and the southern-southeastern Baltic region (Duphorn 1979; Gudelis 1979; Kessel and Raukas 1979; Kolp 1986; Krog 1979; Rudowski 1979; Veinbergs 1979). For understanding the Ancylus stage in a more regional context, Björck's (1987) hypothesis, about this partly puzzling event in the Baltic's history, has been used.

When the surrounding regions' shore-lines have been transferred to the study area a tilting direction perpendicular to synchronous shore-lines at different points of time (Björck 1987) has been used. With this method the transformation of shore-lines from one area to another has to be done step-wise since the tilting directions often varies from place to place. These transformations of course make

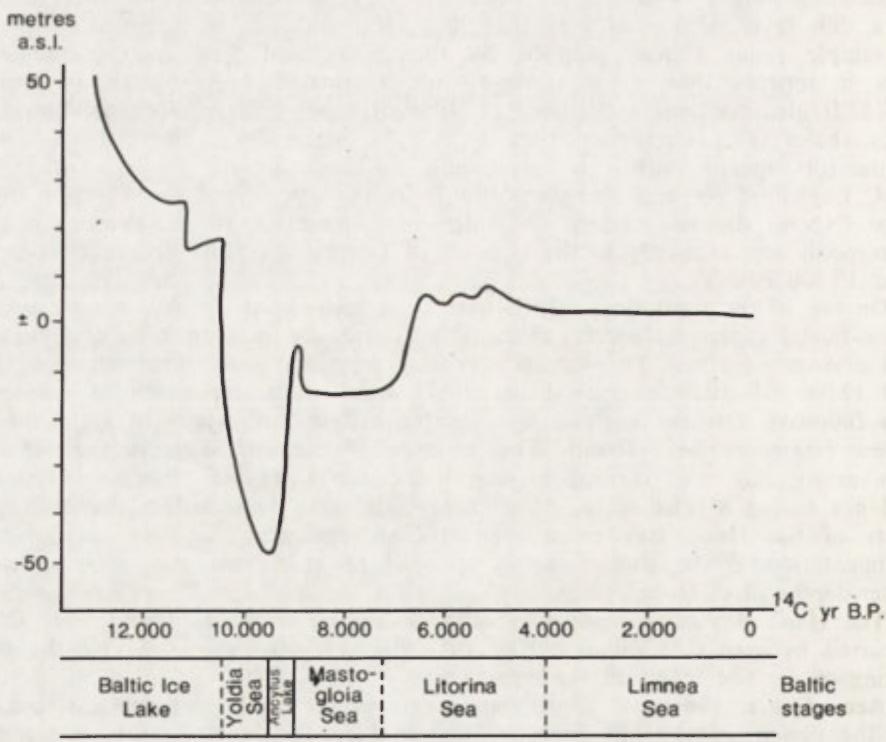


Fig. 5. An interpolated shore displacement curve for the SE Hanö Bay, based on studies from both Sweden and areas in the southern Baltic region. The bases for the construction of the curve are referred to in the text

the resulting curve much less uncertain than if a local shore displacement study had been carried out. Nevertheless an attempt has been made to construct such a shore displacement curve for the SE Hanö Bay i.e. an area c. 15 km east of Stenshuvud (Fig. 5). The main features of this curve are possibly correct, but with regard to fixed levels at certain time points undoubtedly certain errors have to be included. The most dramatic time in the area's sea shore-line history occurs between c. 10 400 and 9000 BP, i.e. between the drainage of the Baltic Ice Lake, with a more or less momentary 25–30 m drop in water-level, and the Ancylus transgression maximum, with a 35–40 m rise in water-level in c. 500 years. In Fig. 3 the dated pine stumps are related to the estimated shore displacement between 9–10 000 BP. It shall, however, be pointed out that the level of the regression maximum at c. 9500 BP is probably the least certain of all the levels in Figs. 3 and 5 since all observations of this level in the southern Baltic are from below sea level and as such more difficult to map. A rough estimation of the error of this level shows that it should be situated somewhere between −40 and −50 m.

## SOME TENTATIVE CONCLUSIONS OF HANO BAY'S LATE WEICHSELIAN-HOLOCENE HISTORY

During deglaciation, at or slightly before 13 000 BP, the clayey till was covered by glacio-fluvial and glacio-lacustrine (varved clay) sediments. The deglaciation shore--line was situated at c. 50 m a.s.l. and the Hanö Bay sediments from that time were deposited in rather deep water. Some hundreds of years of varved clay deposition (Duphorn et al. 1979; Ringberg 1979) were followed by non-varved clay deposition. Apart from the sedimentation of clay in deeper waters, the sedimentation during the remaining part of the Baltic Ice Lake was probably dominated by fluvial deposits from rivers entering the bay of the Kristianstad plain as well as beach and shallow water deposits formed during the rather rapid shore regression. These more coarse-grained sediments were gradually deposited further out on the Kristianstad plain and into the Hanö Bay as the shore-line dropped. When the Baltic Ice Lake was drained at Mt. Billingen (Björck and Digerfeldt 1984, 1986) the shore-line in the Hanö Bay more or less momentarily fell to c. −10 m. During this quite dramatic event erosion and resedimentation processes must have been considerable. This continued during the following Yoldia regression (Figs. 3 and 5). Erosion of the former, possibly rather thick clay deposits was extensive down to at least −40 m. Above this level clay is actually quite rare. In areas below that level the clay deposits were thinned out. For example the till ridge at point B (Fig. 2) was a small island outside the coast when the regression maximum was reached and most of the clay, which most likely previously draped the ridge, was eroded.

Pine and possibly also birch (and later also hazel) trees invaded the quickly exposed new land-surfaces which possibly were dominated by well-drained soils. When the lowest level was reached at c. 9500 BP sub-aqeous bar complexes were built up by wave and stream action on the regression deposits. Inside this complex a shore cliff was formed at c. −45 m (Fig. 2). According to the abundance of pine stumps a pine forest grew west of this cliff. Birch trees were most likely also common, but since pine is preserved much better this can explain why only pine stumps have been found. Tree ring studies on these pines are planned in order to, for example, map the spreading and, later on, flooding of this forest.

At c. 9500 BP the regression turned into a transgression possibly caused by the isostatic uplift of the threshold in Lake Vänern which at that time was a bay of the Baltic. In order to compensate for the gradually shallower threshold for Lake Vänern (and the Baltic) the Baltic's water-level had to rise and the Ancylus transgression had begun (Björck 1987). During the following 500 year period the areas between c. −45 and −10 m in the Hanö Bay were flooded. This did not only annihilate the forest but also, once again, cause considerable erosion and resedimentation. The transgression maximum was reached c. 9000 BP followed by a rapid regression down to −15 to −20 m when the out-flowing Ancylus water eroded and lowered the Baltic's thresholds in the south between Denmark and Sweden (Björck 1987).

For the next 1500–2000 years the shore was situated at more or less the same level (Fig. 5). The inner till ridge (Fig. 2) functioned as a shore bank. Owing to the clayey till and the vicinity to the sea the conditions behind this bank were moist and thus peat began to accumulate. This rather long period with approximately the same shore-level explains the thick sand deposits (Fig. 2) at and slightly outside the shore of that time.

The peat accumulation continued until sea began to rise at c. 7000 BP. The peat was flooded and, during the Litorina transgressions (Fig. 5), possibly covered by sand. This sand protected and compacted the peat for some thousands of years until the shore dropped to a level where wave and stream action could erode and redistribute the sand. The former protection of the peat was thus to a large extent gone and instead erosion of it began, both by water action and by the migrating sand. This is what nowadays still can be seen outside today's coast.

As previously mentioned no clear evidences of clay or mud deposition during the Holocene has been found in the study area. This must be explained by a specific hydrographic situation with high energy levels even in rather deep waters during the greater part of the Holocene. Indications of very thin (< 0.5 m) Holocene clay do, however, occur sporadically in greater water depths than 50 m.

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# A STRATIGRAPHIC SECTION OF THE KATTEGAT BETWEEN LÄSÖ, DENMARK AND BILLDAL, SOUTH-WEST SWEDEN

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ABSTRACT. A stratigraphic section of the Kattegat, from Läsö, Denmark to Billdal, SW Sweden, has been investigated, using hydroacoustic techniques complemented with bottomsampling. A transition occurs, from a predominating Precambrian bedrock surface in the east to Mesozoic strata in the west. The Djupa Rännan channel separates these two bedrock provinces. Glacigenic deposition, also the late Weichselian and Holocene sediments, has been influenced by the various bedrock configurations. Gas is widespread in the Holocene sediments, rich in organic content.

## INTRODUCTION

The sea bed's topographical characteristics and sub-bottom stratigraphy has been investigated in a section, stretching from Billdal, SW Sweden to the Danish island of Läsö (Fig. 1).

In this transect the Djupa Rännan trench, with water depths of up to approximately 100 m, is found oriented in a generally north-south direction. This trench markedly differs from the general water depth of the Kattegat with average values of 25–30 m. New bathymetric measurements, presented by Ulrich (1983), of an area just east of Läsö, reveal extensions of Djupa Rännan with depths up to 130 m.

The hydrography in the Kattegat is at present influenced by the water exchange between the Baltic Sea and the North Sea. The hydrographic situation in relation to wind conditions has been presented by e.g. Dietrich (1951) and a synthesis of the oceanography by Svansson (1975).

Discussions concerning the stratigraphy of unconsolidated sediments in the Kattegat area were recently presented by Fält (1982). Flodén (1973) and Fält (1982) have described a large scale accumulation of Holocene muds east of Skagen, Flaket, (Fig. 1), presumed to be eastwards prograding. These sediments are postulated to be derived from the southern North Sea area. Bahnson et al. (1986) have described the geology of Läsö.

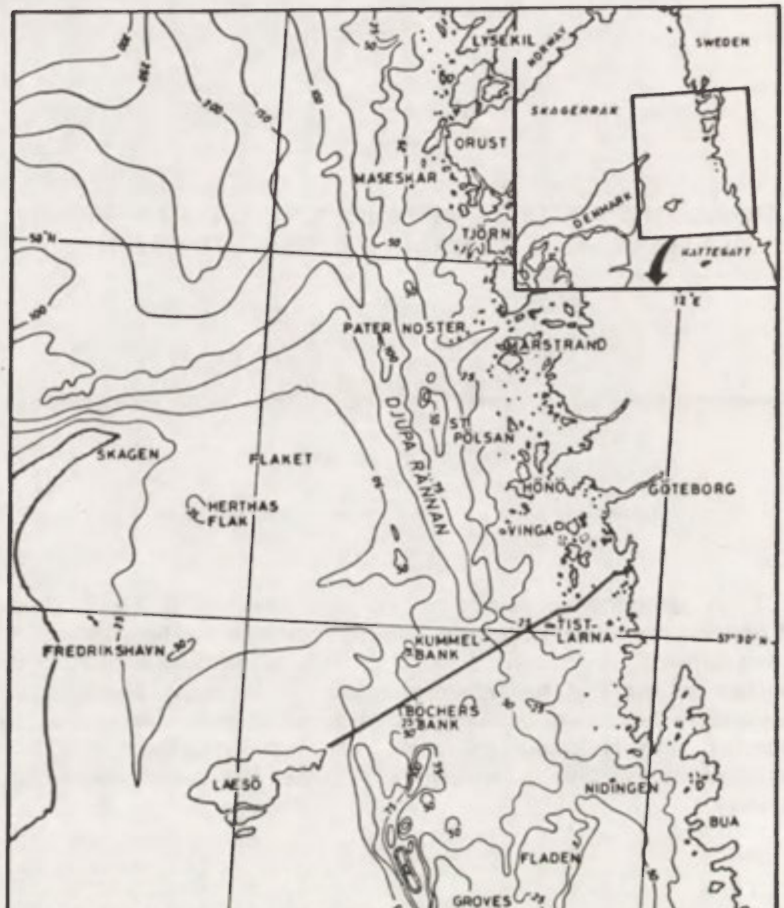


Fig. 1. Review of the northern Kattegat area and investigated transect. The map is reproduced after Falt (1982)

The primary objective of this report is to give an illustration of a section showing the various differences within an area where transition occurs from a bedrock configuration, almost exclusively dominated by Precambrian strata on the west coast of Sweden, to Mesozoic sedimentary strata in Danish territorial waters more to the west.

## METHODS

The presented research results were achieved during a voyage with R/V Arne Tiselius, a 146 tonnes steel trawler equipped for research and surveys at sea. A Del Norte microwave radio positioning system with positioning computer, videosteering indicator and track plotter has provided a navigation precision of +/− 1 m and resolution of 0.1 m.

A Klein Hydroscan Side scan sonar and sub-bottom profiler system and Simrad Skipper 417 echosounding equipment has provided the basis for the hydroacoustic study. All hydroacoustic data has been redrawn after digitalizing and computer processing. On occasion, an underwater video system has been available on board for bottom studies. Sub-bottom strata sampled with a gravity corer have been investigated by visual examinations and in addition the water content, grain-size distributions and loss on ignition has been determined.

## RESULTS AND INTERPRETATIONS

In Fig. 2 the complete investigated transect between SW Sweden and Läso, Denmark, is presented. In the eastern part of the section the Precambrian bedrock reveals a partly irregular configuration. In the Tistlarna area the bedrock's slightly higher elevation is an example of the bedrock topography in the archipelago off the west coast of Sweden. Generally, a westerly dipping bedrock surface is indicated. The bedrock is a remnant of a peneplane, partly broken and eroded in the area, in Lind (1982) a review of the evolution of the Precambrian bedrock of southwestern Sweden is presented. Samuelsson (1975, 1978, 1980) presents interpretations of major features and tectonising of the Precambrian bedrock in the Gothenburg area, SW Sweden.

The reflector indicative of the Precambrian bedrock (unit A) is coming increasingly ill-defined approaching the Böchers Bank moraine complex. Previous studies has shown that in the Böchers Bank area Mesozoic bedrock strata are present (e.g. Bergström et al. 1973). The boundary between the Precambrian bedrock predominant in the easterly area and the westerly area with its huge moraine deposits and Mesozoic strata is found in the transition area between Tistlarna och Böchers Bank. The Djupa Rännan trench marks this boundary.

The Böchers Bank, however, is a vast accumulation of glacigenic deposits (unit B). These deposits probably have their main origin in deposition during active glacial phases and, further, superstructure in the deglaciation stage. On the Böchers Bank the moraine formations outcrop and are only covered by thin (less than 1.5 m) layers of in the Holocene reworked strata. Boulders and stones can be frequent, locally, in the surface layers. Due to the dominance of coarse material in the surroundings, the sediments on Böchers Bank have proportions of sand-silt not found in the eastern part. The uppermost sediments on Böchers Bank vary from silty gyttja-clays, sandy silty gyttja-clays, to gyttja silty sand in the superficial layers.

It is concluded that larger occurrences of Mesozoic strata are absent east of the Djupa Rännan trench and onwards to the Swedish west coast (see also Flodén 1973, Fält 1982). It is also shown that no significant accumulations of till or coarser sediments are visible in the eastern part of the section. Whenever such deposits are present here, they occur as a thin cover on the bedrock surface.

The glacially influenced fine-grained sediments – mainly Late Weichselian clays (unit C) – cover a greater part of the bedrock and moraine deposits in the area east of Böchers Bank. The Precambrian bedrock has only partially outcrops. This occurs where higher elevated parts of the bedrock surface are found. A marked feature of the section is found by the Tistlarna islands where a coherent exposed bedrock surface predominates.

The glacio-marine clays have their greatest thickness in depressions, partly

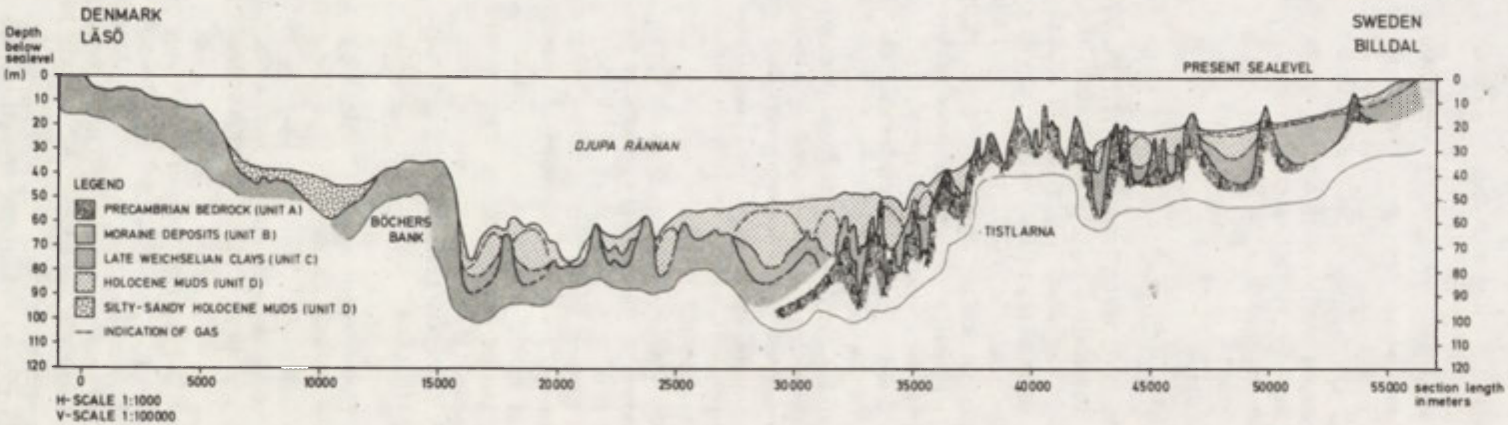


Fig. 2. A stratigraphic section between Läsö, Denmark and Billdal, Sweden

filling these depressions and thus smoothing the topography. The uppermost stratigraphic units (D) investigated are the Holocene – sub-recent fine-grained sediments. These are present as organic rich silty muds in the area east of Böchers Bank.

By definition, the Holocene strata between Böchers Bank and the Swedish west coast can be subdivided into gyttja-clays and clay-gyttjas. In the part of the section with greater water depths, the organic content of the sub-recent sediments is considerable. Analysis of these sediments show loss on ignition values persisting between 6 and 8%.

Except for the Tistlarna area and an occasional outcrop of the Precambrian bedrock elsewhere the Holocene sediments smooth out the bottom topography to a great extent. This smooth bottom topography is a typical feature of the investigated section.

In Fig. 2 the presence of gas in Holocene sediments is illustrated. The presence of gas is traceable from Böchers Bank and onwards to the very nearshore area close to the Swedish mainland. In this figure it is also clearly illustrated how the accumulations are related to bedrock configuration (depressions). The hydroacoustic reflector indicative of the gas is a signal response interpreted as coming from gas in pores in the more or less low permeable sediments, and not from coherent gas accumulations. In parts of the Kattegat, however, layers with trapped gas are indicated. In these areas the hydroacoustic signal is completely extinct. This effect also reduces the possibility of detecting lower stratigraphic units in the areas concerned.

## CONCLUSIONS AND SUMMARY

It can be postulated that the Djupa Rännan trench separates two different sedimentary environments also with marked differences in bedrock lithology. Bedrock of Precambrian origin predominates east of the Djupa Rännan trench, Mesozoic strata west of the trench. The transect shows a section of the Scandinavian Border Zone, subdividing the crystalline Fennoscandian shield in the northeast from the sedimentary Danish-Polish Trough in the southwest.

Also in Quaternary active glaciation and deglaciation stages, erosion and accumulation of deposits is related to the differently developed bedrock and its configuration. It is obvious that more extensive moraine deposits occur west of the Djupa Rännan trench. Deposition of fine-grained sediments in the late Weichselian and Holocene is also related to the different topographic characteristics. These deposits are responsible for smoothing out, to a large extent, the bottom topography in the area.

The Holocene muds of Kattegat are generally rich in organic matter, with contents reaching nearly 10% in areas of greater water depths. The occurrence of gas is, for the most part, confined to these sediments rich in organic matter.

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# MORPHODYNAMICS OF THE CLIFFED COAST, WOLIN ISLAND

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ABSTRACT. The basic research problem of this study involves presenting morphodynamic functions of the Wolin cliffed coast along sandy and till sections. Documentary evidence provided by a ten-year cycle of observations (1977–1986) has been utilized. Analysis of the present morphological development of the cliffed coast in Wolin Island has been based on repeated morphological mapping of cliffs and on the compiled morphological and morphodynamic maps. The present approach is a new methodological proposal for the study of the coastal zone.

The action of the storm surge is the main process that operates in the coastal zone. Simultaneously, the cliff becomes shaped under the influence of gravity force, water, wind, snow and frost. The cliffs retreat continuously.

The extent of the morphodynamic zones recognized, including those of degradation, transport, transport-degradation and accumulation-degradation as well as zone of dynamic equilibrum, varies throughout a year or over a multi-year period and it is dependent on seasonal variations of weather. The denudative system of the cliffed coast in Wolin Island is characterized throughout the year by four morphogenetic seasons, i.e. autumn-winter, early spring, summer and summer-autumn. The development of the cliffed coast tends to follow the trend of abrasion equilibrium profile.

## INTRODUCTION

The present-day denudation system of Wolin Island comprises various relief features which may be regarded as separate systems having specific structure and their own energy-mass circulation. The modern shore zone of the Baltic Sea displays exceptionally distinctive characteristics in the Wolin landscape. It is an internally varying zone which consists of diversified morainic and dune hinterland and wide foreland making up a portion of the submerged land.

Distinctive characteristics of the denudation system of the shore zone are determined by the fact that it is a zone where the lithosphere, hydrosphere, atmosphere and biosphere interact. Note should also be made of the fact that it is affected by multidirectional human activity. The functioning of the shore zone may be discussed if particular relief details present in this zone are considered subsystems. It is an approach concerning component parts. The comprehensive approach to the functioning of the shore zone involves combined study of lithosphere, hydrosphere, atmosphere and biosphere subsystems and establishment of relationship between them (Kostrzewski 1987a). When conducting research in the shore zone of Wolin Island, the present authors were concerned with the lithosphere subsystem, especially cliffed coasts.

## RESEARCH PROBLEM

The ancient and modern development of cliffed coast has been relatively poorly recognized (Kostrzewski 1984, 1987b; Subotowicz 1982). The lithological study of deposits of which the shore zone of Wolin Island is built up has supplied a lot of new data in this respect (Borówka et al. 1986a, b; Borówka et al. 1982; Kostrzewski 1983, 1985). Systematic field studies which have been carried out on the cliffed coast of Wolin Island since 1977 deal with the following problems (Kostrzewski and Zwoliński, in press):

1) morphological variability of the cliffed coast throughout a year and over a multi-year period,

2) the influence of lithology, slope angle and a cover of the cliff on the characteristics of newly produced forms,

3) qualitative and quantitative assessment of the effects of morphogenetic processes on the cliff,

4) morphodynamic functions of the cliffed coast of Wolin Island.

The present authors are chiefly concerned in this study with morphodynamic functions of the cliffed coast of Wolin Island along sections composed of sand and till. Interpretation of the main research problem is based on detailed studies mentioned under 1 and 3. The determination of morphodynamic functions of the cliff is of key significance in studying the present-day development of the cliffed coast in Wolin Island, as well as is of major importance for actual practical purposes.

## STUDY AREA, METHODS

The cliffed coast of Wolin Island may be divided into two portions, i.e. the eastern portion (Międzywodzie – Grodno) and the western one (Grodno – Międzyzdroje), which differ in morphologic characteristics. The western portion cliff

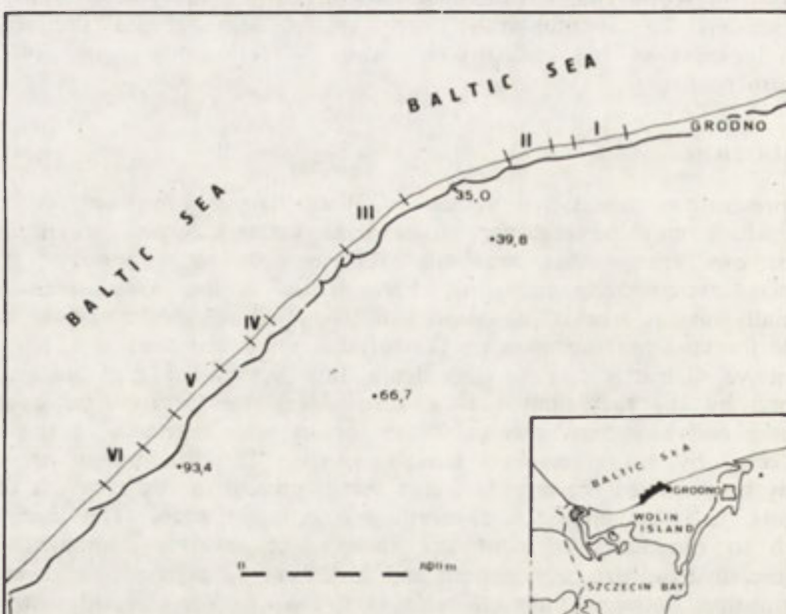


Fig. 1. Location of test section I–VI on the cliffed coast of Wolin Island (Dots indicate height points on the Wolin end moraine)

is higher and varies strongly in respect of morphology. It is just the cliff of the western portion that is subjected to study here.

Throughout a ten-year cycle of observations, 1977 to 1986, studies were carried out 4 to 6 times a year along six test sections of the cliffed coast of Wolin Island (Fig. 1) which differ in morphology and lithology (Photos 1–6).

The determination of morphodynamic functions of the coast, based on the compiled morphodynamic maps, is preceded by taking repeated stereoscopic photos of the cliff sections under investigation (Kijowski et al., 1984), by carrying out detailed lithological mapping, by measuring cliff recession rates and by compiling morphological maps (Kostrzewski and Zwoliński 1985; in press). Thus, the determination of morphodynamic functions of particular test sections of the Wolin cliff constitutes the final stage in carrying the research programme through.

## CONDITIONS OF PRESENT-DAY CLIFF MORPHODYNAMICS

The present-day morphodynamics of the Wolin cliff depends on local lithologic characteristics, weather conditions, a terrain cover and human activity. All these local characteristics need to be considered against regional characteristics, the extent of which is larger.

Throughout preliminary field studies account was taken of the above characteristics (Kostrzewski and Zwoliński 1985). Test sections of the cliffed coast chosen for study are composed of sand or till, or consist of till-sandy or sandy-till material. Till cliff sections contain two till series, i.e. Vistulian (Baltic) brown and Middle-Polish grey tills (Kostrzewski 1985). Grey till is mostly included in sandy tills, whereas brown till falls under slightly sandy silt-sized tills. The series of tills are overlain by fluvioglacial and aeolian sands and the intervening fossil soil horizons (Borówka et al. 1986a, 1986b; Borówka et al. 1982). The sandy-gravelly series is up to 40 m thick in the sandy sections. Aeolian sand covers make up the top of the sandy-gravelly fluvioglacial series.

An important lithologic group within the cliffed coast test sections under investigation is represented by slope and subslope deposits which are largely due to mass movements and slope-wash. In order that susceptibility of deposits of which the Wolin cliff is built up to destructive and constructive processes might be studied, investigations of shear resistance were carried out. Interesting results have allowed definite sets of forms to be related to corresponding deposits (Kostrzewski and Zwoliński 1985; in press).

The rhythm of functioning of the present-day denudative system of the Wolin cliff is closely dependent on the seasonal structure of climate experienced in Wolin Island. Four seasons may be distinguished (Woś 1977), namely:

a) autumn-winter = 12 November till 5 April,
b) spring = 6 April till 7 June,
c) summer = 8 June till 12 September,
d) autumn = 13 September till 11 November.

The autumn-winter season is characterized by cool weather conditions with average diurnal temperature of below 5°C. In December through February days with temperatures of below 0°C may be recorded. Precipitation is moderate. There exist identical weather conditions in the spring and autumnal seasons. Mean diurnal temperatures most frequently range between 5 and 15°C. Days with precipitation occur often. Days with diurnal temperatures of above 15°C prevail in summer. Warm weather types without precipitation gain an advantage. Rarely occurring precipitation is rather heavy. The annual precipitation totals fluctuate around 600 mm. Throughout the year northerly and westerly winds prevail.

The test sections of the Wolin cliffs constitute largely exposed surfaces and are, to a slight degree merely, covered with trees and grass. Single trees or clusters of beech and pine trees and grass are frequently encountered on slope slumps.

There are no protective devices there in order to prevent cliff destruction. The Wolin Island cliffed coast under investigation has been incorporated into the Wolin National Park, which is undoubtedly convenient for research purposes. Adverse action of man on the cliff can be chiefly seen there in disorderly walking on sandy walls, which has a relatively slight effect on the morphology.

All the above elements of the denudation system of the Wolin cliff have been presented in cartographic form and statistically. This represents the initial stage of structural study of morphological variability of the cliff under investigation.

## MORPHOLOGICAL VARIABILITY OF THE WOLIN ISLAND CLIFF THROUGHOUT THE ANNUAL WEATHER CYCLE

The rhythm of operation of shore processes is determined by the seasonal structure of climate experienced on the Baltic coast and exactly, in Wolin Island. The proportions and intensity of morphogenetic factors and corresponding processes are thus closely related to the seasonal weather pattern and local conditions, including lithology, height and slope angle of the cliff, terrain cover and human activity. Over the ten-year cycle of observations, the qualitative assessment of shore process effects was mainly applied and merely to the slightest degree was the quantitative assessment performed.

The basic method that was applied to the study of morphological variability of the cliff involved taking repeated stereoscopic photos of the test sections under investigation (Kijowski et al. 1984). For the purpose of identifying precisely forms, the photos were viewed under a stereoscope or an interpretoscope. Stage-made morphological maps were compiled on the basis of a specially listed set of symbols (Kostrzewski and Zwoliński 1986; in press). The proposed divisions represent a systematic attempt to group forms of the present-day denudative system of the Baltic coast cliff. The classification consists of six groupings, namely morphology, lithology, hydrography, vegetation, land management and others. The grouping under morphology is most comprehensive as it contains 87 symbols (Fig. 2). It takes account of all the morphogenetic factors that model cliff, of variants of their action and of corresponding forms (Kostrzewski and Zwoliński 1987a, b). Large-scale morphological maps presenting the spatial distribution of forms on the cliff (Fig. 3), in combination with field observations and measurements provide the basis of qualitative and quantitative assessment of the intensity and effects of processes modelling the shore zone. Note should be made of the fact that each interpretation of a morphological map of the cliff face is merely correct for the date of observation, i.e. the date of stereoscopic picture recording.

Wave action and abrasion associated with storms are the basic factors governing the morphological characteristics of the shore zone. As a consequence of mechanical abrasion in the main, abrasion shelters, undercuts and niches form. The topographic expression of the Wolin cliff is largely affected by mass movements and slope-wash. Earth falls from the till and sandy faces often occur, resulting in steep slopes and accumulative sub-slope forms, frequently irregular in shape. Rock-fall occurs in the sandy sections on a large scale. Fall furrows and loose talus cones result. The effects of rock-fall from the till faces comprise mantles and small cones with cloddish structure. Landslide should be regarded as one of the most

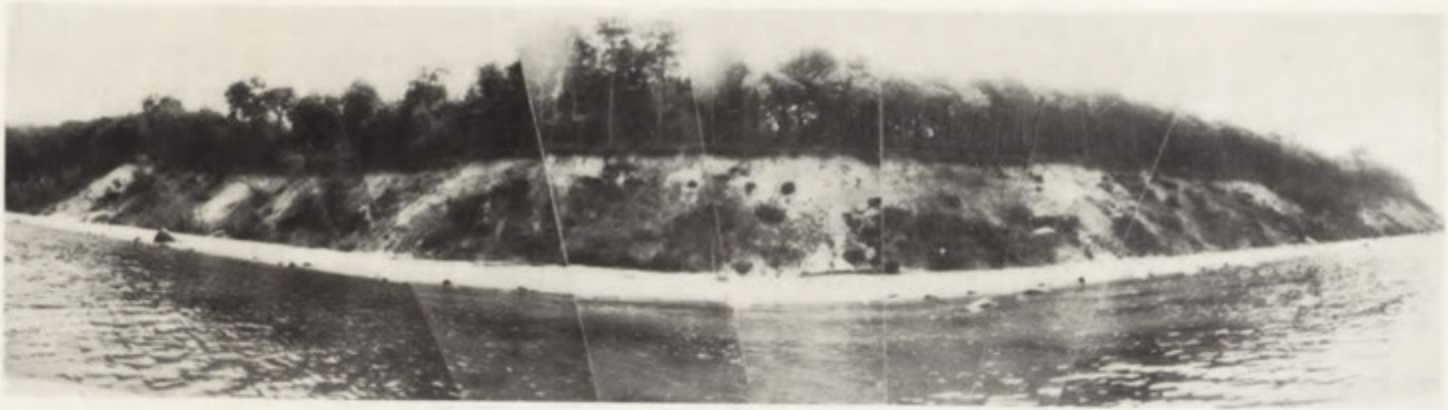

Photo 1.

Photo 2.

Photo 3.

Photo 4.

Photo 5.

Photo 6.

Photo 1. General view over test section I on the cliffed coast of Wolin Island. Length of 215 m, height of ca 20 m. Till and sandy section

Photo 2. General view over test section II on the cliffed coast of Wolin Island. Length of 200 m, height of ca 25 m. Till section

Photo 3. General view over test section III on the cliffed coast of Wolin Island. Length of 315 m, height of ca 55 m. Sandy and till section

Photo 4. General view over test section IV on the cliffed coast of Wolin Island. Length of 140 m, height of ca 40 m. Sandy section

Photo 5. General view over test section V on the cliffed coast of Wolin Island. Length of 275 m, height of ca 30 m. Till section

Photo 6. General view over test section VI on the cliffed coast of Wolin Island. Length of 225 m. Sandy section important processes operating on sandy cliffs. The downslope movement of sandy material takes place along slip planes. An increase in the cliff steepness due to abrasion and that in overburden resulting from strong saturation of sand with water accelerates the landslide. As a rule, the landslide affects the entire surface of the cliff but at varying intensity. Well-developed slides are mainly present in the sandy sections of cliffs.

Slumping with which slumps or systems of slumps on both sandy and till faces are associated is an important morphogenetic process modelling the cliff surface. Soil creep results in small wrinkles which are visible in grass-covered parts of the cliff. Mudflows and earth flows occur when the ground thaws or heavy rains fall. They are particularly well perceptible on till faces and at the foot of the cliff on the beach. The effects of slope-wash are well marked on till faces in the form of a network of rills and grooves. Exposed sandy surfaces are convenient places for deflation which produces small niches at the cliff top, ledges, benches, shelters, pavements and sandy upper cliff caps at the cliff top, sand shadows, streaks and aeolian veneers. The cliff microrelief is shaped by frost and a snow cover.

## THE CONCEPT OF A MORPHODYNAMIC MAP : METHODOLOGICAL REMARKS

The compiled morphological maps allow arriving at stage-made, annual and multi-year morphodynamic maps. A morphodynamic map distinguishes between separate cliff zones differing in morphodynamic function. Dominant and secondary processes define morphodynamic functions of particular cliff zones (Kostrzewski and Zwoliński, in press). Detailed field observations allow recognition of the following zones that are presented on the morphodynamic maps, namely degradation zone, transport zone, accumulation zone and their combinations as well as that of dynamic of equilibrium. These zones vary internally, which allows again subzones to be recognized on the basis of spatial distribution of processes and indirectly, on that of the existing forms.

The degradation zone comprises vertical till and sandy faces in the upper and middle parts of cliff slope as well as the base of the cliff. Its total balance is negative since material, as a rule, is transported permanently downcliff.

The transport zone is marked mainly in middle and lower parts of sandy and till sections. Particularly, it is sharply marked in the sandy cliff. It is a zone where material is constantly transferred downslope. The volume of lost material is greater than that of the material delivered.

The accumulation zone is identified with the cliff base and some local places on the cliff slope. It displays great variability throughout the year. This zone varies internally depending on deposit type.

The zone of dynamic equilibrium occupies these cliffed coast areas that are entirely or predominantly covered with trees, shrubs or grass. Slow action, for example of soil creep and slumping, is observable. It is marked by the tilting trees, undulating grass-covered surfaces, minor surface fissures, etc.

The concept of the morphodynamic map is known in the Polish literature (Dorywalski 1958). However, it is applied for the first time to the determination of variability of cliffed coasts. A characteristic of the cliff geosystem is its great differentiation. The extent of the zones and subzones recognized varies throughout the year and over a multi-year period. The above method facilitates qualitative assessment of successive stages of cliffed coast development in different timescales.

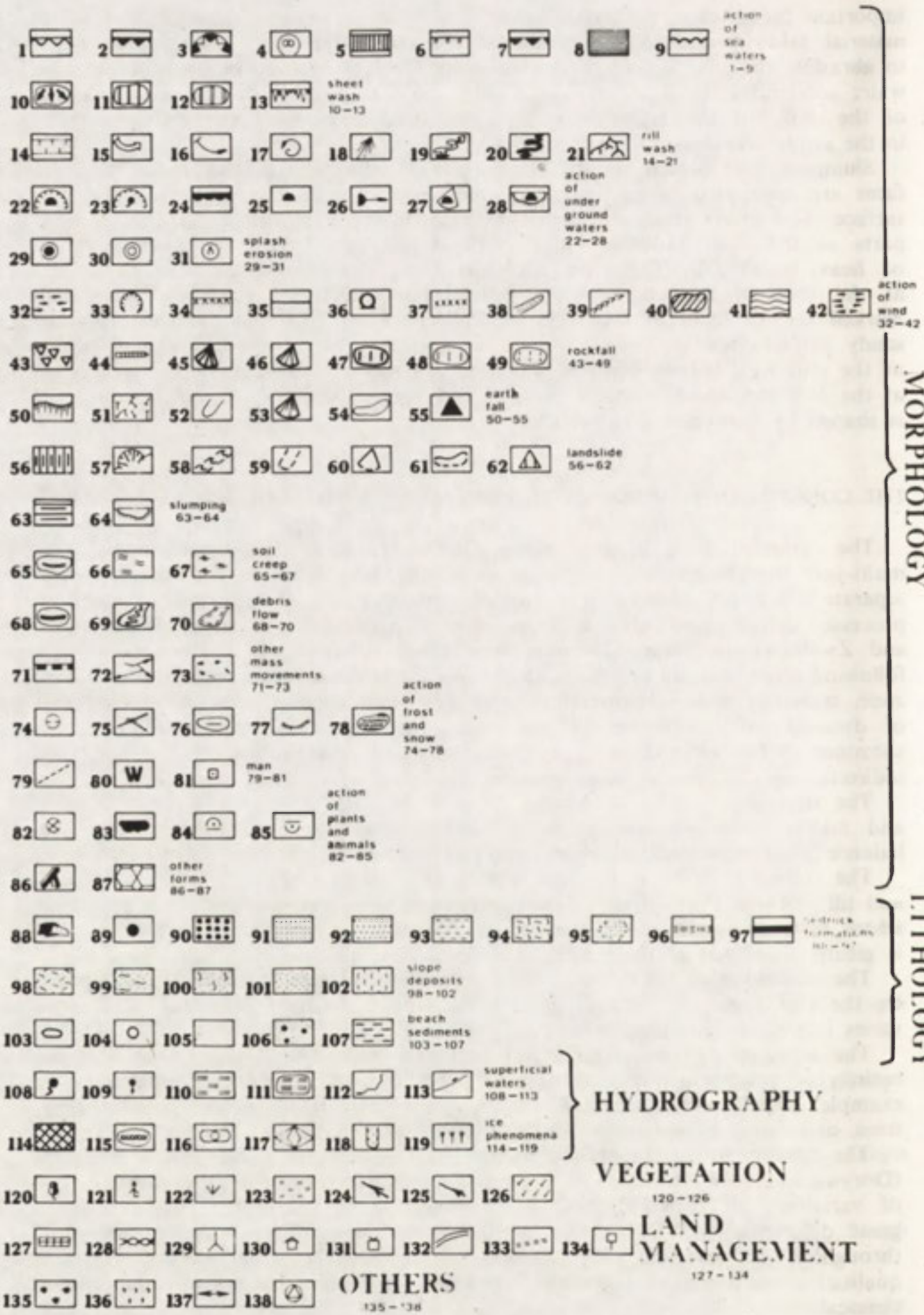
action of sea waters 1–9
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Fig. 2. A set of cliffed coast morphological mapping symbols

## MORPHOLOGY

Forms created by the action of sea waters: 1 — abrasion undercuts, 2 — abrasion shelters, 3 — abrasion niches, 4 — cavitation kettles, 5 — abrasion benches, 6 — microcliffs, 7 — abrasion marks, 8 — beaches, 9 — storm ridges; Forms created by sheet-wash: 10 — sheet-wash pavements, 11 — sheet-wash covers, 12 — sheet-wash veneers, 13 — clay varnish; Forms created by rill-wash: 14 — erosional cuts, 15 — grooves, 16 — rills, 17 — erosional kettles, 18 — alluvial fans, 19 — mud trickles, 20 — earth trickles, 21 — dendritic clay pattern; Forms created by the action of underground waters: 22 — suffosion niches, 23 — spring niches, 24 — suffosion thresholds, 25 — suffosion tunnels, 26 — suffosion rills, 27 — suffosion fans, 28 — suffosion slumps; Forms created by splash erosion: 29 — raindrop craters, 30 — hail craters, 31 — earth pyramids: Forms created by the action of wind: 32 — deflation pavements, 33 — deflation niches, 34 — deflation benches, 35 — deflation ledges, 36 — deflation shelters, 37 — upper cliff caps (cliff naspas), 38 — sand shadows, 39 — aeolian streaks, 40 — aeolian veneers, 41 — ripple marks, 42 — aeolian surfaces; Forms created by rockfall: 43 — fall marks, 44 — fall furrows, 45 — consolidated talus cones, 46 — loose talus cones, 47 — consolidated talus covers, 48 — loose talus covers, 49 — loose talus veneers. Forms created by earth fall: 50 — bluffs, 51 — dumps, 52 — earth tongues, 53 — earth cones, 54 — earth ridges, 55 — earth blocks: Forms created by landslide: 56 — slip plane, 57 — arcuate heads, 58 — slide tracks, 59 — slide tongues, 60 — slide cones, 61 — slide ridges, 62 — turf failures; Forms created by sluping: 63 — rupture surfaces, 64 — slumps: Forms created by soil creep: 65 — creep hollows, 66 — earth wrinkles, 67 — turf wrinkles; Forms created by debris flow: 68 — flow troughs, 69 — mudflows, 70 — earth flows; Forms created by mass movements: 71 — thresholds, 72 — fissures, 73 — cracks; Forms created by the action of frost and snow: 74 — cryogenic pans, 75 — frost cracks, 76 — nivation hollows, 77 — cryogenic swellings, 78 — post-nivation mantles; Forms created by man: 79 — paths, 80 — holes, 81 — imprints; Forms created by the action of plants and animals: 82 — uprooted tree hollows, 83 — turf overhangs, 84 — burrows, 85 — conduits; Other forms: 86 — spurs, 87 — desiccation cracks

## LITHOLOGY

Bedrock formations: 88 — boulders, 89 — stones, 90 — fluvioglacial gravels, 91 — fluvioglacial sands, 92 — aeolian sands, 93 — muds, 94 — tills, 95 — Cretaceous rafts, 96 — soils, 97 — fossil soils; Slope deposits: 98 — till deposits, 99 — till-sandy deposits, 100 — sandy-till deposits, 101 — sandy deposits, 102 — mud deposits; Beach deposits: 103 — boulders, 104 — stones, 105 — gravels, 106 — sands, 107 — muds

## HYDROGRAPHY

Superficial waters: 108 — springs, 109 — ephemeral springs, 110 — swamps, 111 — puddles, 112 — trickles, 113 — cracks; Ice phenomena: 114 — beach ice, 115 — ice covers, 116 — hummocks, 117 — floes, 118 — icefalls, 119 — needle ice

## VEGETATION

120 — deciduous trees, 121 — conifers, 122 — shrubs, 123 — grasses, 124 — fallen trees, 125 — fallen shrubs, 126 — marine grasses

## LAND MANAGEMENT

127 — concrete sea-walls, 128 — fascine sea-walls, 129 — breakwater, 130 — buildings, 131 — fortifications, 132 — roads, 133 — routes, 134 — signs, boards

## OTHERS

135 — snow cover, 136 — shells, 137 — drift-wood, 138 — pollutants

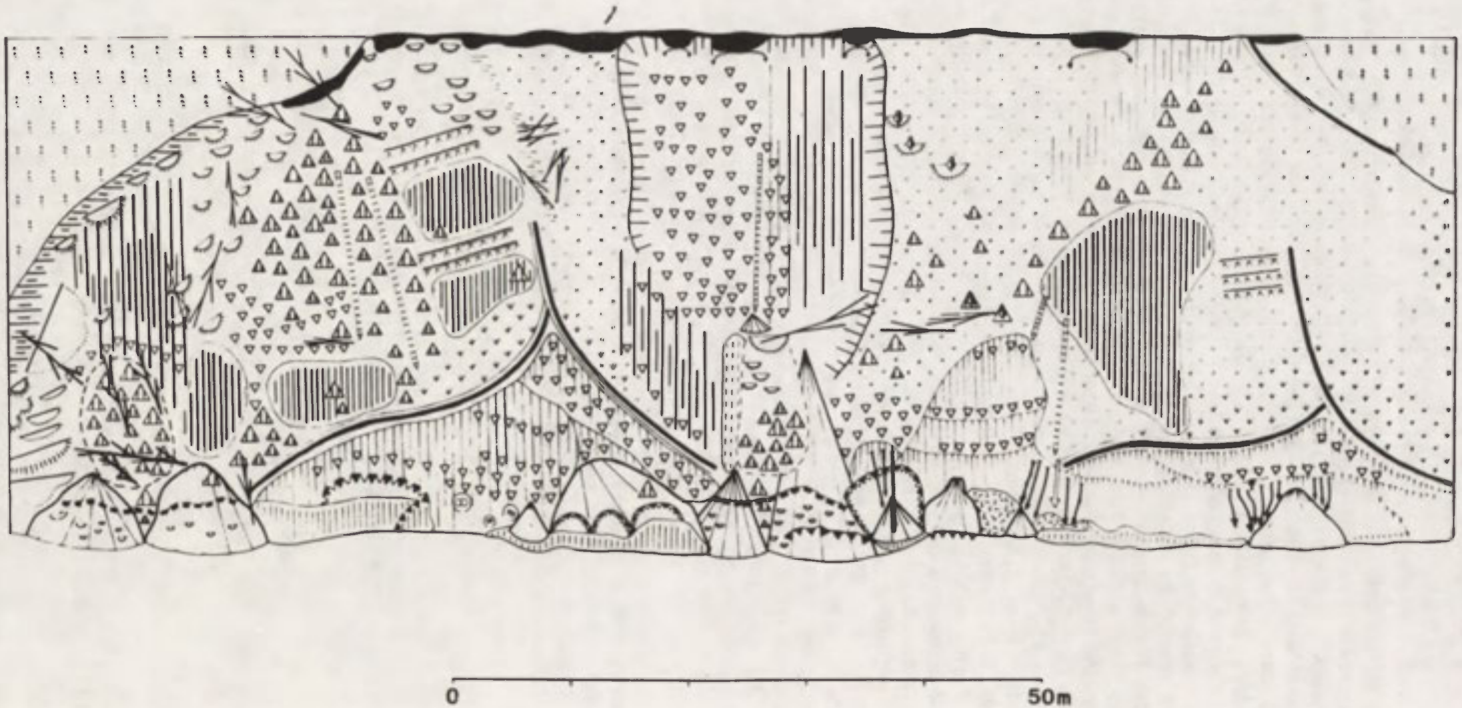


Fig. 3. An example of a morphological map of the cliffed coast, compiled for part of sandy section IV in Wolin Island, 18 November 1983. (For key see Fig. 2)

## MORPHODYNAMICS OF THE WOLIN ISLAND CLIFFED COAST

The morphological mapping has allowed the spatial distribution of forms to be traced in the test sections of the cliffed coast in Wolin Island. Repeated mapping has allowed a better understanding of temporal variations of forms. Definite variations of morphogenetic processes correspond with them. The morphological map represents a detailed analytical approach. This concept of the morphodynamic map is a proposal for inductive definition of morphodynamic functions of the cliffed coast in Wolin Island. The enclosed examples of morphodynamic maps (Figs 4 and 5) reveal differences in morphodynamic functions of cliffed coasts composed of till and/or sand for definite weather seasons.

As can be seen in the till and sandy sections, the degradation zone occupies the top cliff portions and partially, the middle ones. The transport zone is located in the middle cliff portions, while the accumulation zone lies at the cliff base. The transport-degradation and accumulation-degradation zones have varying morphodynamic functions according to the weather seasons. If comparison is made between the distribution of zones in the till and sandy sections, a more diversified distribution pattern is observed in the till sections. The zones of the till sections are also more diversified internally by subzones. Morphodynamic zones lying in the sandy sections are larger in extent and are less internally diversified by subzones. There is also a zone of dynamic equilibrium with different extent within particular test sections. It occupies grass-covered areas of the cliff face. Small surface areas occupied by the zone of dynamic equilibrium are indicative of high activity of the test sections on the cliffed coast of Wolin Island.

The above characteristics allow a description of morphogenetic seasons recognized over the annual weather cycle on the basis of geosystem functioning in the Wolin Island coastal zone. The data from the ten-year cycle of observations suggest that the denudative system of the cliffed coast in Wolin Island is characterized by four morphogenetic seasons, i.e. autumn-winter, early spring, summer and summer-autumn.

During the autumn-winter season there are generally high sea stages, frequent temperature fluctuations at the ground surface and generally low temperatures. Abrasion, rockfall, slope-wash and deflation are in operation. The last process operates frequently on the frozen ground. In winter when there is no beach ice cover, abrasion is particularly effective. A substantial amount of material is carried away from the cliff base. Landslide and sluping become active and numerous slides and slumps result, especially in the sandy sections. The till cliff faces become smoothed and rejuvenated.

In the early spring season the denudation system of till and sandy cliffs functions at an accelerated rate. It is a period of high intensity of slope-wash, debris flow, landslide, slumping and earth fall as well as rockfall and deflation during dry spells. Numerous slides, earth blocks, markedly displaced slumps and turf failures, mud and earth flows become produced.

The summer season exhibits relatively high .morphogenetic stability. The most important processes include rockfall and deflation as well as slope-wash during wet spells. Many surfaces are covered with aeolian veneers, sometimes – aeolian streaks, ripple marks and sand shadows which are characteristic of the sandy surfaces. Small loose and consolidated talus cones are frequently encountered in different places of the cliff slopes. They occur at the cliff foot in maximum extent. Covers and veneers due to rockfall are also of frequent occurrence at the foot of till faces. A system of fall furrows is marked on both till and sandy faces. They are deepened often. Sheet-wash and rill-wash effects are sharply marked during wet summer. Major changes in the cliff morphology may be

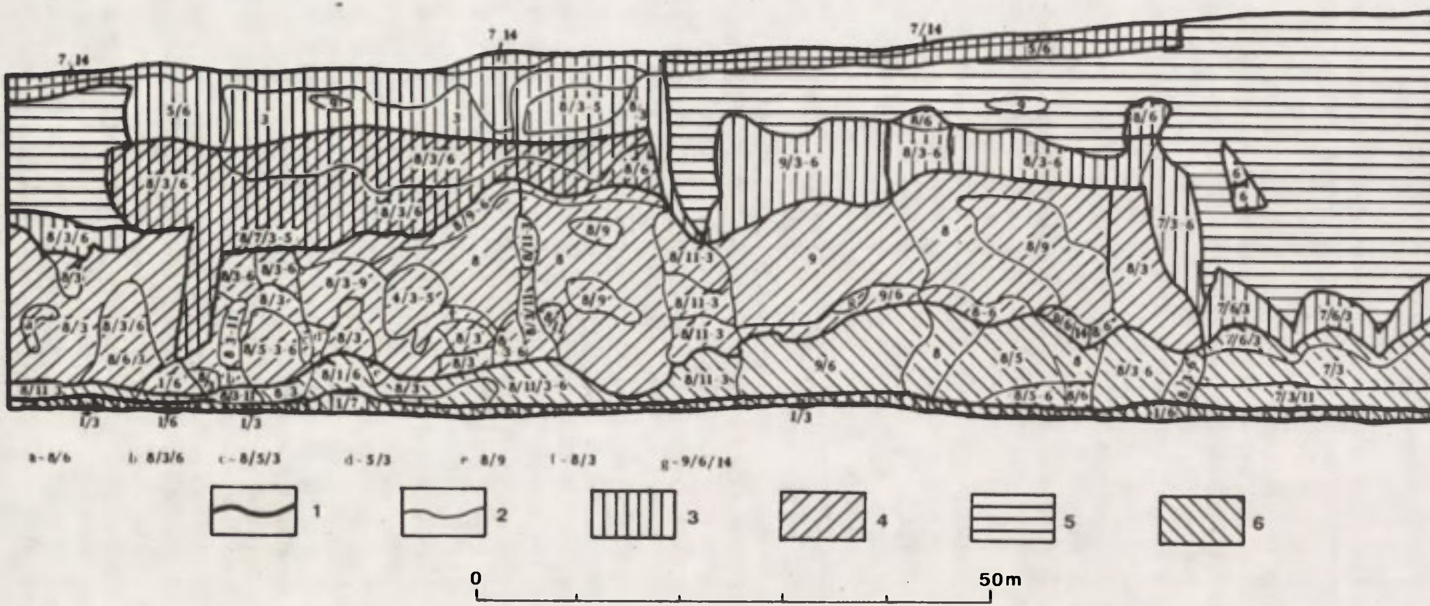

Fig. 4. Morphodynamic map of the cliffed coast for part of sandy section IV in Wolin Island, March 1983 – August 1984

Key to symbols

1 – morphodynamic zone limit, 2 – morphodynamic subzone limit, 3 – degradation zone, 4 – transport zone, 5 – accumulation zone, 6 – zone of dynamic equilibrium

Key to subzone numbers

1 – abrasion subzone, 2 – marine accumulation subzone, 3 – slope-wash subzone, 4 – suffosion subzone, 5 – aeolian subzone, 6 – rockfall subzone, 7 – earth fall subzone, 8 – landslide subzone, 9 – slumping subzone, 10 – creep subzone, 11 – debris flow subzone, 12 – gelation-nivation subzone, 13 – anthropopressure subzone, 14 – bioaction subzone

Note: – operation of similar – intensity morphogenetic processes / operation of decreasing-intensity morphogenetic processes

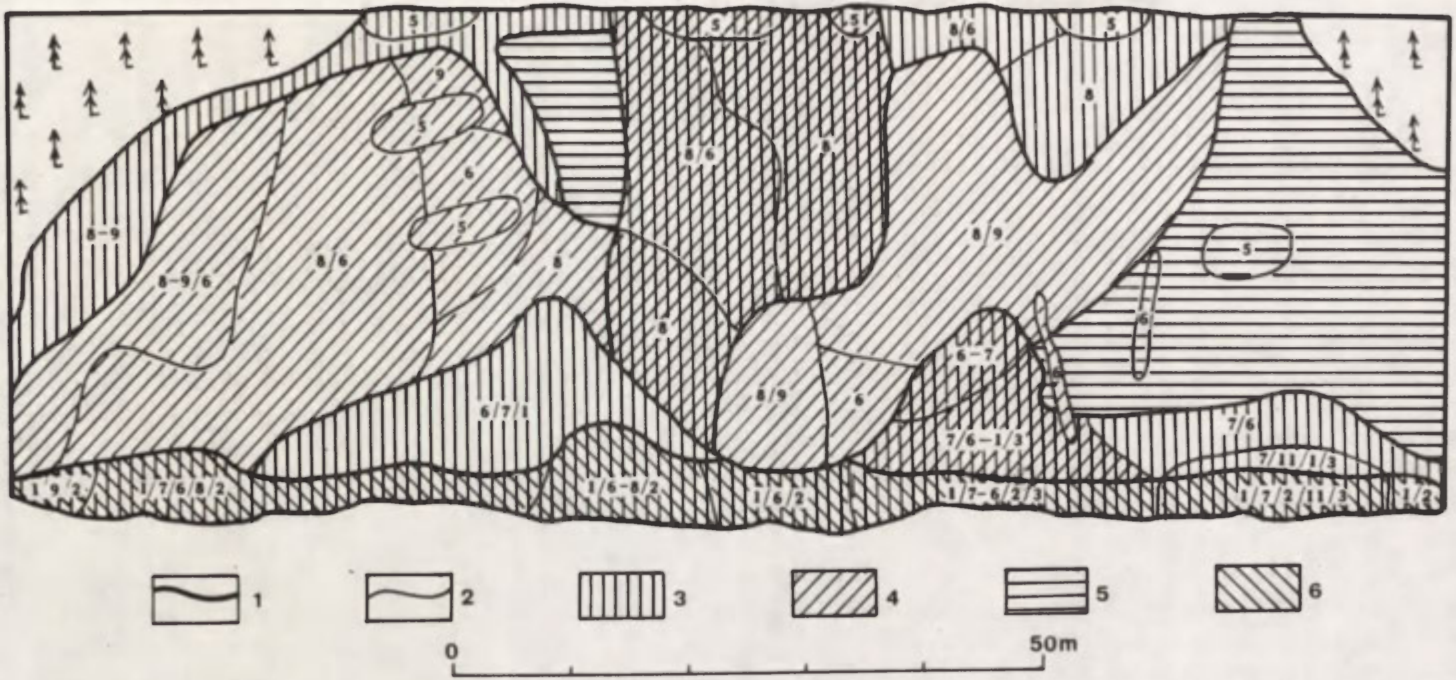


Fig. 5. Morphodynamic map of the cliffed coast for part of till section V in Wolin Island, March 1983 – August 1984 (For key see Fig. 4)

produced by heavy rains. During the summer months ephemeral springs, the morphologic action of which is distinct, may be traced rather clearly.

The summer-autumnal season is characterized by increased intensity of morphogenetic processes. Attention should be given to great qualitative and quantitative variations in the influence of morphogenetic processes on the cliff slope. It is a period of intense slope-wash, soil creep, debris flow, landslide, rockfall, earth fall and relatively intense abrasion. Mud and earth flows form, especially at the base of till cliffs. Numerous craters result from raindrop attack on sandy surfaces. There are cracks and fissures on the sandy and till cliff faces, which give rise to the formation of slides and slumps. Abrasion associated with a storm surge produces different abrasion forms at the cliff base.

Mention should be made of the fact that characteristics of particular morphogenetic seasons may be reappraised in respect of operating morphogenetic processes, which is largely generated by seasonal variations of the weather.

## CONCLUSIONS

The analysis of morphological development of the cliffed coast in Wolin Island, based on the proposed repeated morphological mapping of cliffs and the compiled morphologic and morphodynamic maps, is a new methodological proposal for study of the coastal zone. The main process operating there involves the action of a storm surge. The cliff is simultaneously shaped by gravity force, water, wind, snow and frost as well as, to a minor extent, by bioaction and human impact. The cliff that undergoes permanent destruction retreats gradually. However, not long, a few-year periods of dynamic equilibrium can be recognized throughout the present-day development of the Wolin Island cliff.

The extent of morphodynamic zones and their functions vary throughout the year and over a multi-year period and are indicative of great variability of the coastal geosystem under investigation. The denudation system of the cliffed coast in Wolin Island is determined by morphogenetic seasons throughout the year. The combination of morphogenetic processes during particular morphogenetic seasons is largely dependent on seasonal variations of weather conditions. The development of the cliffed coast in Wolin Island proceeds towards the profile of abrasion equilibrium.

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# TENDENCIES OF DEBRIS SLOPE EVOLUTION IN THE HIGH TATRA MOUNTAINS

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ABSTRACT. An alpine cliff and related to it talus slope, are the most typical features of mountainous areas transformed by glaciers. The development of a young cliff starts at the moment when the inclined rock surface becomes free of ice. Very often such surfaces are very steep, vertical or even overhanging and devoid of weathered material. Weathering processes on rock surfaces produce debris which is transported downslope due to gravitational processes such as falling, sliding and toppling. Principal debris slope units formed below cliffs are differing from each other in terms of grain size composition and sorting. Active and non-active or partially active slope surfaces occur adjacent to each other. They are the diagnostic features of past and contemporary periglacial/cryonival environment. Therefore, talus slope features are good indicators of postglacial evolution. Seven types of cliff-debris slope systems have been distinguished in this paper.

## INTRODUCTION

Rock slope evolution is determined by the properties of its rocks (lithology, tectonics) as well as by the processes acting to modify it. The rate of slope formation is also closely related to climatic conditions. Steep rock slopes termed as alpine cliffs towering above glacial valleys, have hard rock masses fractured and jointed. Natural rock is thus not continuous material. Water pressure fluctuations and severe physical weathering within bare rock surfaces lead to slope weakening and couloir and chute development within them. Therefore, cliff faces are retreating through non-uniform weathering and non-uniform thick layer of rock is removed from the cliff. Theoretically constructed model of bare rock slope development cannot be useful in the studies of alpine cliffs. In the High Tatra Mountains taluses below cliffs are usually of the type of well developed talus cones and compound talus formed by coalescing cones. Only sporadically taluses form a band of accumulation parallel to alpine cliff face.

## VIEWS ON TIME AND RATE OF SLOPE DEVELOPMENT

In geomorphological investigations carried out in the High Tatra Mountains an attempt was made to distinguish different Holocene tendencies in debris slope evolution in relation to rock slope configuration, absolute and relative height and climatic conditions.

Most of alpine cliffs within glacial troughs in the High Tatra Mountains were ice-free at least since 13 000–12 000 years BP. Radiocarbon dates (10 100 ± 140 BP and 9900 ± 120 BP) of lacustrine deposits give the limit between the Late Glacial and the Holocene (Wicik 1979; Krupiński 1983) in the middle part of the mountains, i.e. on an altitude of 1500 m. The uppermost, youngest glacial cirques above 1700 m were exposed to subaerial modelling from about 10 000 years BP. The data of Koperowa (1962) and Krupiński (1983) indicate that the last small glaciers in this part of the mountains were preserved till the Younger Dryas (Klimaszewski 1967). After that time Holocene warming of climate generated subaerial denudational system above the upper timberline. Due to vertical shifting of timberline, the hipsometric belt of 1500–1950 m was subject to the largest morphogenetic changes during the Holocene. Based on the volume of debris covers accumulated at the base of rock slopes, Lukniš (1968) estimates average lowering of ridges during postglacial period of 5 m. According to Klimaszewski (1967) this rate of Holocene remodelling of Pleistocene relief is relatively small.

Contemporaneous sediment transfer within the whole slopes occur as a result of a variety of processes. Particular processes operate with various intensities depending on the slope unit present, its position in a given geoecological belt (Kotarba 1976, 1984) and within a particular slope sequence, and on the stage of development reached by that slope unit (Kotarba and Strömquist 1984). Various rock resistance of the crystalline core-granite rocks in the High Tatra Mountains is responsible for differences in the height and fragmentation of rock slopes.

Six basal debris slope forms had developed below alpine cliffs (Kotarba 1985). They are to be observed at present at different heights above sea level. Their present-day activity is related to the geoecological belts. Both climatic and floristic conditions control the rate of principal morphogenetic processes within belts (Kotarba 1984). The general scheme is shown on Tables 1 and 2.

TABLE 1. Mean annual temperatures, mean monthly air temperatures for January (coldest month) and July (warmest month) in altitudinal climatic and vegetation belts (climatic data from Hess 1965)

| Climatic belt | Vegetation belt | Mean annual temperature | Mean monthly temperature | |
|---|---|---|---|---|
| | | | January | July |
| Cold<br>2200–2663 m | Alpine summit zone | −4.0°C | −12.0°C | 4.0°C |
| Temperate cold<br>1850–2200 m | Alpine meadow zone | −2.0°C | −10.0°C | 6.0°C |
| Very cool<br>1550–1850 m | dwarf pine zone | 0.0°C | −8.5°C | 8.2°C |
| Cool<br>1100–1550 m | upper forest belt | 2.0°C | −6.0°C | 10.5°C |
| Temperate cool<br>700–1100 m | lower forest belt | 4.0°C | −5.5°C | 13.0°C |

TABLE 2. Dominant morphogenetic processes on basal debris slope forms in the Polish Tatra Mountains

| Principal factor | Principal transfer process | Principal slope form | Present-day activity in vertical climatic zones | | | | |
|---|---|---|---|---|---|---|---|
| | | | temperate cool 900–1100 m | cool 1100–1550 m | very cool 1550–1850 m | temperate cold 1850–2200 m | cold 2200–2663 m |
| Gravity | falling, rolling, bouncing, sliding | rockfall talus | + | + + | + + + | + + + | ? |
| Snow | snow avalanching | avalanche talus | – | + | + + | + + + | + + |
| | slow-moving snow | | no information | | | | |
| | sliding on snow surface | | – | – | + + | + + | + + |
| Snow and water | slush avalanching | alluvial talus | – | – | + + | + | – |
| Snow meltwater | ephemeral stream flow | | + | + | + + | + + + | + |
| Rain storm water | rainwash, debris flow | | + | + | + + + | + | + |
| Interstitial ice | internally induced mass movement of debris creep | rock glaciers | | | – | – | |
| Freeze-thaw changes | creep | block slope | – | – | + + | + | + |
| | frost creep, sliding | debris-mantled slope | + | + + | + + + | + + + | + + |
| | | Activity index: | 4 | 7 | 19 | 17 | 9 |

Process: – inactive (0)
+ weak (1)
+ + strong (2)
+ + + very strong (3)

## TYPES OF CLIFF-DEBRIS SLOPE SYSTEMS

There are seven types of process-response alpine cliff-debris slope systems which are typical of the High Tatra Mountains. Their existence was identified on the base of multiannual field experiments (see Kotarba et al. 1983).

### *Type I*

The youngest cliffs situated above 1700 m comprise elements of the glacier cirque relief system and reach relative relief of the order of 450–500 m. They were the last areas to became free of ice. When facing to N and E they are vertical or overhanging. At their bases there are typical talus slopes formed from loose debris which is very mobile and easily subject to movement (scree slope). The ratio of the height of the talus slope ($H_s$) to that of the rock cliff ($H_c$) is always less than 0.1. As a rule they are very well sorted (coefficient of sorting $r = 0.98$). Talus slopes as a whole are subject to movement triggered by debris showers (rockfalls). The material supplied from the cliffs slides over the surface and is only gravitationally sorted. The rate of debris accumulation is uniform along the whole talus slope. If the glacial cirques are filled with lakes, transport of debris material continues under subaqueous conditions which by infilling leads to a decrease in the lake basin's capacity. Talus slopes are developed in a form of regular talus cones or sheets (Fig. 1–I).

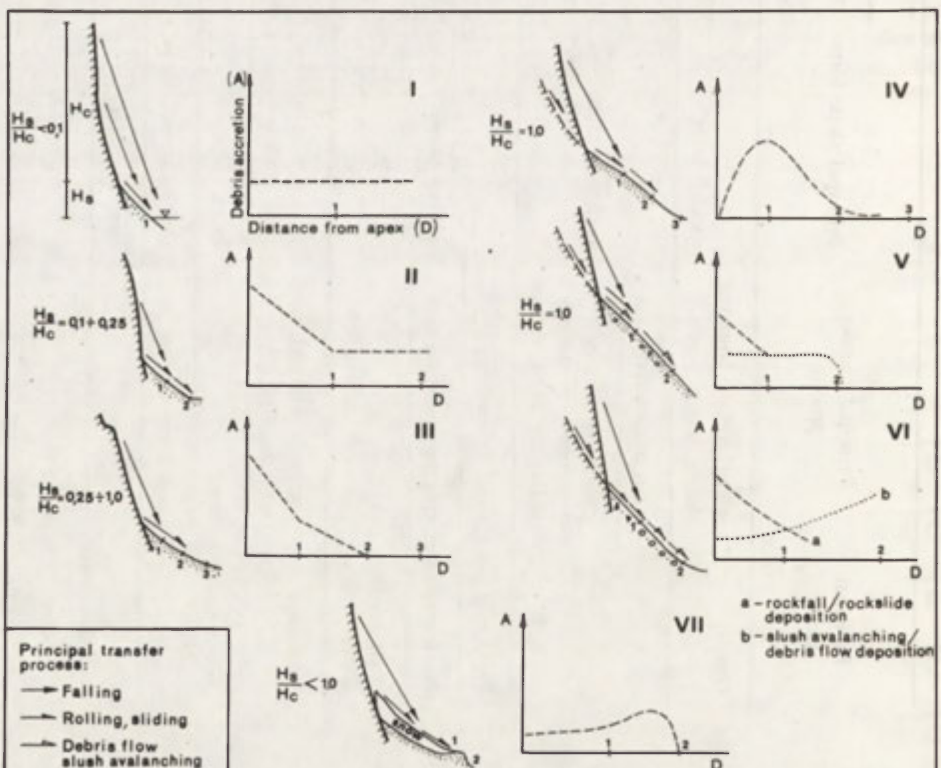


Fig. 1. General dynamics of debris slope surfaces related to cliff morphometry in the High Tatra Mts

*Type II*

Alpine cliffs forming glacial troughs in the main valleys of the Tatra reach relative heights of 350 m. In contrast to type I they possess a profile consisting of at least two elements: the upper one with the smaller gradient is situated above the vertical limits of the valley glaciers, and the lower one has a vertical or nearly vertical profile. The ratio of the height of the talus slope to the height of the rock cliffs is determined by an index within the range. 0.1–0.25. Cliffs are usually undissected and the correlation coefficients measuring sorting on the sheet talus are of the order of $r = 0.90$. The weathered material on the cliffs which falls onto a talus cone has a limited size range. Thus, two zones are distinguished: 0–1 debris falling zone also called the upper scree subsystem, and 1–2 debris sliding zone (lower scree subsystem). The material transported on the surface of the sheet talus extends down to the base of the slope and buries the valley floor which is infilled with morainic deposits. The rate of debris accumulation varies with distance from the apex as is shown in Fig. 1-II.

*Type III*

In areas of complex glacial cirques which developed more fully cliffs have smaller gradients (lower than 60°) and extensive debris slopes which were formed at their bases. $\frac{H_t}{H_c} = 0.25-1.0$ with the most frequent heights of cliffs and debris slopes being of the order of 200 m. At the foot of the undissected cliffs sheet taluses are formed. The latter consist of three elements: 0–1 upper scree subsystem developed under the control of debris shower and surficial displacement by rolling, bouncing and sliding, 1–2 middle subsystem developed by debris supplied from the upper subsystem and transported on the surface, and 2–3 lower subsystem either at present relict, (i.e. not subject to aggradation resulting from debris supplied from the above subsystem) or only sporadically subject to change. Sorting coefficient $r = 0.90$ indicates that the main factor affecting the morphology of the slope surface is gravity. There is an inverse relationship between debris accumulation and distance from the cliff (Fig. 1-III)

*Type IV*

Alpine cliffs within glacial cirques or cliffs of glacial troughs subjected to differential erosion related to systems of cleavages and dislocation zones, display a modified relief. These cliffs are densely channelled by chutes which are initiated in the broad rock niches beneath the ridges and supply many weathered products. They form large talus cones at the base of the cliffs and are 200 m or more high. $\frac{H_t}{H_c} = 1.0$. As in the type III three elements can be distinguished within the debris slopes. The coefficient of sorting is only $r = 0.80$ because the formation of the surface occurs in response to gravity and dirty snow avalanches. Measurements of fresh debris over a few years have shown (Kotarba 1981) that there is a different relationship between the rate of talus cone accumulation and distance from the cliff. This relationship can be expressed by a second order polynominal $y = a+bx+cx^2$ where $y$ – magnitude of accretion and $x$ – distance from cliff. This demonstrates that a relatively small amount of debris accumulates at the cone apex. The rate of accumulation reaches a maximum in the middle part of the cone and then frequently decreases to zero at the cone base. This type of relation is well supported by the fact that the main period of delivery of material from the cliff to the debris slope occurs in spring (May–June). During that period the snow cover starts to disappear. The middle parts of the slope are subject to the most frequent thawing (Photo

1). The rockfall (debris shower) related to spring thawing on the cliff falls on the debris slope close to the apex zone. The debris than slides on the firn surface towards the middle part of the slope where it stops on making contact with the rough debris slope surface which has no snow cover. The largest rate of debris accumulation is observed here. Some of the debris rolls further down and is deposited in small volumes at the base of the cone (Fig. 1–IV).

*Type V*

Alpine cliffs within glacial cirques or glacial troughs are highly dissected by chutes. These are formed as a result of selective erosion corresponding to narrow zone of cleavages. Sheet taluses have been formed at their base as a result of the interlocking and jointing of relatively small adjacent talus cones. At the front of the deep chutes dissecting the cliffs, the sheet taluses are dissected by debris flow gullies. Such slopes develop both in response to gravitational and debris flow processes. Section 0–1 represents debris accretion caused by deposition of debris flow products. Sorting coefficients are low $r = 0.20$–$0.50$. These systems are very common in the Tatra, especially on the south- and west-facing slopes (Fig 1–V).

*Type VI*

Long cliffs (500–600 m) and rocky slopes deeply channelled by chutes corresponding to the resistant zones (gradients of the order of 60°) and located in two climatic zones are subject to a complex pattern of development. Rockfalls and rockslides form section 0–1 within the debris slope, while debris flows and slush avalanches create section 1–2. Debris flows are produced by the selective transport of weathered material of various size across the slope. Therefore, no sorting of material is observed ($r < 0.20$) and the surface of the debris slope is characterised by a complex of secondary forms such as: rockfall debris streams, rockslide tongues, gullies, furrows and levees (Fig. 1–VI).

*Type VII*

The north-facing and east-facing cliffs located higher than 2000 m in the uppermost part of the Tatra are subject to a seminival climate. Steep perennial snow patches (up to 30°) occur directly at the base of the high cliffs and are subject to impact by weathered products supplied to them. Debris sliding on the snow surface accumulates at front of the patch to form protalus rampart. This material is poorly sorted ($r = 0.20$). The area of the largest firn fields reaches 5 ha (Luknis 1973). Protalus ramparts are also formed in basins associated with the highest lakes which are shaded from direct insolation (Fig. 1–VII).

## DEBRIS SLOPE EVOLUTION

As a result of the above described processes acting on debris slopes and closely related to cliff features, different tendencies of their evolution can be distinguished:

I – parallel accretion of the whole debris slope profile affected by debris shower ($G$), sliding and rolling ($s$);

II – strong accretion on debris slope element located close to the apex (debris shower element) and parallel accretion in the lower element affected by debris sliding and rolling on the surface ($S$);

III – strong accretion on debris slope element located close to the apex and moderate accretion on the middle element. The lowest element is inactive;

Photo 1

Photo 1. Alpine cliff and talus cone below Skrajna Turnia at Hala Gąsienicowa. An example of the fourth type (IV) of debris accretion pattern. Photo: 29 May 1979

Photo 2. Rockfall talus slope (type III) and alluvial talus slope (type V) near Morskie Oko Lake
Photo: Polish Society of Earth Sciences, Kraków Section

Photo 3. Rockfall talus slope entirely modified by parallel accretion due to debris shower (type I). To the left type VI, slope modified by slush avalanching and debris flow deposition.
Czarny Staw Lake near Morskie Oko Lake.
Photo: Polish Society of Earth Sciences, Kraków Section


Photo 2


Photo 3

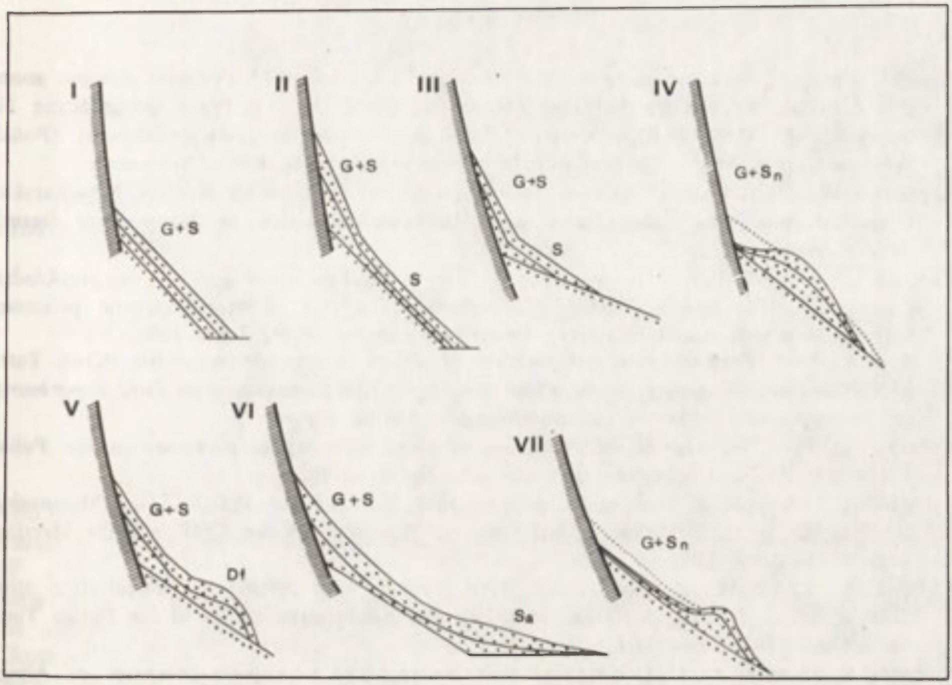


Fig. 2. Patterns of accretion of loose material on debris slopes in the High Tatra Mts. Supply of material by: falling ($G$), sliding and rolling on debris slope surface ($S$), sliding and rolling on snow surface ($S_n$), by debris flows ($D_f$) and slush avalanching ($S_a$)

IV – strong accretion on the middle element. Debris shower is accompanied by debris sliding on the snow surface during spring;

V – strong accretion close to the apex and at the base of slope. Debris flow processes ($D_f$) are responsible for debris redistribution along the slope profile;

VI – strong accretion on the apex element ($G+S$) and slush avalanching/debris flow deposition at the base ($S_a$);

VII – strong accretion at the base of debris slope. Most of the material supplied to debris slope ($G$) slides over the long lasting snow pathes ($S_n$) and is deposited in form of protalus ramparts.

Figure 2 summarizes all the above developmental tendencies.

## CONCLUSIONS

The evaluation of the alpine debris slopes in the High Tatra Mountains shows that they have developed through the operation of mass wasting at least since 10 000 years. As magnitude of principal slope processes in altitudinal belts is controlled mainly by climate, thus relative heights and fragmentation of alpine cliffs (related to structural controls) seem to be decisive factors of principal debris slope formation during the Holocene. This conclusion is summarized in seven simple models, which relate debris slope forms to the processes responsible for them.

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# TIME AND DYNAMICS OF THE LAST SCANDINAVIAN ICE-SHEET RETREAT FROM NORTHWESTERN POLAND

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ABSTRACT. A time-scale calibrated in radiocarbon years has allowed the last glacial episode in northern Poland to be ascribed to the interval of ca 22 000–13 200 yr BP and its maximum of approximately 20 000 yr BP. An attempt has also been made to present a map of deglaciation isochrones and to approximate the age of major ice-sheet positions; i.e. the Poznań Phase: 18 400 yr BP, the Chodzież Subphase: 17 200 yr BP and the Pomeranian Phase: 15 200 yr BP. Whilst occupying the main positions, the dynamics of the ice sheet front had varied to a higher degree than it was assumed in the former deglaciation models with special preference for the zonal ice sheet waning model. Evidence has been provided by new studies based on the complex facies analysis of marginal phenomena. It allows construction of more realistic depositional models in order to establish a dynamic tendency ($D_t$) of the ice-sheet front on the basis of depositional efficiency ($D_e$) estimation. The procedure by means of which models can be constructed is a new methodological proposal in studying the ice-sheet marginal phenomena produced in low energy conditions.

## INTRODUCTION

Another study (Kozarski 1981) has shown that till the midseventies the last glacial episode in northwestern Poland was erroneously identified (Krygowski 1975) with nearly the whole Vistulian. It was a view emerging from a very traditional concern with climatic fluctuations occurring during the Pleistocene, especially the Vistulian. The latest of three general models (Table 1) of genetic interpretation of marginal phenomena, including deglaciation, i.e. the model of zonal waning of the last ice sheet resulting in stagnant and dead ice forms (Bartkowski 1967, 1969), corresponded with it.

The methodological background of this model and the resulting misinterpretations have initially been reviewed critically (Kasprzak and Kozarski 1984; Kozarski and Kasprzak 1987) because of research results provided by a new research project concerning the marginal zones of the last ice sheet in northwestern Poland. The problem of a time interval assigned to the last glacial episode and the dating of the maximum last ice-sheet advance in northwestern Poland have been discussed (Kozarski 1980, 1981) on the basis of a geologic profile found at Konin-Maliniec and described by Stankowska and Stankowski (1979).

TABLE 1. Various models of genetic interpretation of ice-marginal features in NW Poland (after Kozarski and Kasprzak 1987)

| Deposits and structures | Processes | Preferred model |
|---|---|---|
| Till rich in block material | dumping and tectonic stacking | "Alpine" end moraine forms |
| Clays, tills, sands and gravels Glaciotectonic structures | shearing, thrusting, folding | thrust ridges and push moraines |
| Sands, silt and gravels Distinct bedding and gravitational deformations (mainly faults) | deposition in meltwater on and in glacier ice | stagnant and dead ice forms |

Organic sediments present at this site were radiocarbon dated (Pazur and Walanus 1979; Pazur et al 1981) and palaeobotanically studied (Tobolski 1979, Pazdur et al 1981).

The dating of organic sediments occurring in the zone of maximum ice-sheet extent and in the coastal zone, as well as detailed studies of the main positions of the last ice sheet in test areas have provided an entirely new basis of the consideration of time and dynamics of the last ice-sheet retreat from northwestern Poland.

## DATING OF THE LAST ICE-SHEET MAXIMUM ADVANCE IN POLAND AND THE NEIGHBOURING COUNTRIES

There is no doubt that tremendous progress in tentative determination of time intervals at which the Vistulian climatic fluctuations took place, including the most dramatic event, i.e. the last glacial episode, has been stimulated by the radiocarbon dating of organic sediments. For instance, this is the case with the last ice-sheet advance in northwestern Poland since radiocarbon measurements allow estimation of:

(1) the duration of the ice-sheet advance and retreat which lasted for scarcely seven thousand years, and

(2) the age of maximum advance (Kozarski 1980, 1981; Pazdur et al 1981) as ca 20 000 yr BP.

This is due to the discovery (Stankowska and Stankowski 1979) and investigation (Pazdur et al. 1981) of organic sediments covered by the last in the zone of maximum extent of the Vistulian ice sheet at Konin-Maliniec. As can be inferred from radiocarbon measurements of the uppermost organic layer (Pazdur and Walanus 1979; Pazdur et al 1981) having an age of 22 050 ± 450 yr BP (Gd 645) and 22 230 ± 480 yr BP (Gd 646), the maximum advance recorded by the till was younger than the organic layer. After correlations with radiocarbon-dated sequences in the GDR (Cepek 1965) and in the Soviet Union (Krasnov 1978; Vasnyachuk 1971; Gerasimov and Velitchko 1982), its age (Kozarski 1980, 1981, 1986; Pazdur et al 1981) can be estimated as ca 20 000 yr BP (Table 2). It is important to note that this estimate is not applicable to the zone of maximum extent, lying east of Konin towards the Vistula valley and along the southern fringe of the Mazurian Lakeland since it is not the Leszno Phase but the Poznań Phase that produced it (Mojski 1984, Fig. 94).

TABLE 2. Correlation of major ice-sheet positions in N Poland and the neighbouring countries, as well as the radiocarbon age (Morner 1981) of relevant horizons in the Grand Pile profile (after Kozarski 1986)

| | $^{14}C$ age $\cdot 10^3$ BP | BALTIC (Morner et al. 1977) GRAND PILE (Mörner 1981) | GDR (Cepek 1965) | POLAND (Kozarski 1986) | USSR (Gerasimov, Velitchko 1982) |
|---|---|---|---|---|---|
| | 13 | Low Baltic Stadial 13 100 | Velgaster Staffel | Gardno Phase ~13 200 | Luga |
| N | 14 | | | | |
| A | 15 | East Jylland Stadial 15 300 | Pommersches Stadium | Pomeranian Phase ~15 200 | Vepsovo |
| I | 16 | | | | |
| L | 17 | | | Chodzież Subphase ~17 200 | maximum ~17 500 |
| U | 18 | | | | DRETSHALUKI 17 770 ± 170 18 370 ± 180 |
| T | 19 | Frankfurter Stadial 18 600 – 19 700 | Frankfurter Staffel | Poznań Phase ~18 400 | |
| S | 20 | | | Leszno Phase ~20 000 | |
| I | 21 | Brandenburger Stadial 20 500 – 22 500 | Brandenburger Stadium SKADO (W4) 20 270 ± 1000, 20 475 ± 600 | | |
| V | 22 | | KL KOSCHEN (W6) 21 160 ± 800 | KONIN MALINIEC II 22 050 ± 450 22 230 ± 480 | PUCHKA 21 410 ± 150 21 880 ± 110 |
| | 23 | | | | SHAPUROVO 22 410 ± 210 GOSHA 22 950 ± 440 |

If the above situation is concerned farther eastward, in the vicinity of Grodno in the Soviet Union territory, it may probably be accounted for in terms of stratigraphy by the Upper Pleni Vistulian sequences that have been studied by Vasnyachuk (1971). He has established with reference to the Dretshaluki site that the till due to the maximum advance covers organic sediments which have a radiocarbon age of 28 370 ± 180 yr BP (cf. Table 2). Their telecorrelation (Kozarski 1986) indicates that they correspond, to a larger extent, to the Poznań Phase (= the Frankfurter Stadial) rather than to the Leszno Phase (= the Brandenburger Stadial). This implies that the Leszno Phase became overridden by the Poznań Phase in the Grodno region as well, likewise in the major portion of Poland's territory (Woldstedt 1932).

## DEGLACIATION ISOCHRONES

As yet the period between ca 22 000 yr BP and ca 13 000 yr BP has not been divided for the North Polish Plain into smaller units by means of biostratigraphic facts if the constantly insolvable problem of the so called Mazurian Interstadial is left out of account. It must be stated that during the recession of the last ice sheet from northern Poland, there existed too severe climatic and edaphic conditions for a rapid plant succession in areas from which the ice sheet had just retreated. From a scant palaeobotanical record (Borówko-Dłużakowa 1969; Kopczyńska-Lamparska et al 1984), it follows that recolonization by plants began as late as during the second half of the Oldest Dryas, i.e. when the ice-sheet margin was situated on the southern fringe of the present-day Baltic coast. Therefore, the study area should not be supposed to contain organic deposits belonging to the ca 22 000 – ca 13 000 yr BP period, to which radiocarbon dating may be applied, as is the case in North America, for example (Clayton and Moran 1982).

Thus, there is a need to use an indirect procedure for the determination of age of the major events associated with the last ice-sheet waning. This is mostly related to the main ice-margin positions after the maximum advance

(= Leszno Phase), which are marked by marginal form trains of known morphostratigraphic value and importance. In nortwestern Poland they are called the Poznań Phase, the Chodzież Subphase (= Kujavian Phase), the Pomeranian Phase and the Gardno Phase (Fig. 1). All the main ice-sheet positions are of supraregional importance and can be correlated with similar lines in the neighbouring countries.

Recently a simple procedure .has been proposed for the estimation of age of the main ice-sheet positions during deglaciation (Kozarski 1986). It is based on age estimates for extreme ice-margin positions in northwestern Poland during the maximum advance (= Leszno Phase) and Gardno Phase in the Baltic coastal zone (Fig. 1, Table 2), i.e. approx. 20 000 and approx. 13 200 yr BP, respectively, as well as on measurements of distance between them along the 16°40′ E meridian. The Gardno Phase in the Polish coastal zone has been identified with the Oldest Dryas (Rosa 1968; Roszko 1968). Its radiocarbon age is obtained from Gross's (1958) first age estimates based on the radiocarbon chronology. Recently it has been proved to a certain extent by a radiocarbon date available for Niechorze (Kopczyńska-Lamparska 1976; Kopczyńska-Lamparska et al 1984) where lacustrine deposits of Oldest Dryas and Bølling age occur in front of the Gardno Phase line. The date of approximately 13 200 yr BP or that of 13 100 yr BP recurs in new publications (Mörner 1979, 1981; Mörner et al 1977; Berglund 1979; Duphorn et al 1979, 1981). Its validity appears to receive support also from radiocarbon measurements of organic deposits giving a date od 23 390 ± 500 yr BP, which lie beneath the till considered (Čebotareva and Faustova 1975) a result of the ice-sheet readvance during the Oldest Dryas (= Luga) and from those of organic deposits over this till, which have an age of 12 650 ± 500 yr BP in the Soviet Union. In Rügen Island in the GDR, the Frankenthal site with organic deposits of the age 12 365 ± 180 yr BP at the bottom is known for its similar stratigraphic setting (Brose and Kliewe 1975). The organic deposits lie there over the Velgaster Staffel till outside the GI end moraine which is correlated by Kliewe (1975) with the Oldest Dryas (= Low Baltic Advance) ice-sheet readvance.

The above age estimates for the Leszno and Gardno phases and the known distance between them permit calculation of the average annual rate of ice-front retreat as 44 m year$^{-1}$. By means of this value and the measured distances between the main ice-front positions during deglaciation has a map of isochrones been produced (Kozarski 1986), with the approximate age of the Poznań Phase, Chodzież Subphase and Pomeranian Phase on the order of ca 18 400, 17 200 and 15 200 yr BP, respectively (Fig. 1). The approximate age is obviously a tentative estimate since the dates are affected by inaccurate age estimation concerning the extreme ice-front positions. Nevertheless, there is a strikingly great similarity (Table 2) between the available approximate date for the Poznań and Pomeranian phases and the radiocarbon-determined age of cold phases at Grand Pile, which can be correlated (Mörner 1981) with the Frankfurter (18 600–19 700 yr BP) and East Jylland (15 300 yr BP) stadials. It seems not to be accidental since other circumstances provide interesting arguments for the discussion of climatic conditions during deglaciation.

These circumstances include the oceanic record of climatic fluctuations in the North Atlantic (Ruddiman and Mc Intyre 1981) which reveals low temperatures and increased volume of ice sheets in the Northern Hemisphere over the 20 000–16 000 yr BP period. This time interval can be correlated with the Leszno Phase as long as the Pomeranian Phase in the presented (Kozarski 1986) sequence of glacial events in Poland. Moreover, rapid deglaciation of 16 000 to 13 000 yr BP, as postulated by Ruddiman and Mc Intyre (1981), took place in northern Poland's coastal zone where it was recorded by ephemeral ice-dammed lakes with a smaller quantity of varves than in basins occurring to the south of the ice-sheet position during the Pomeranian Phase (Kozarski 1986).

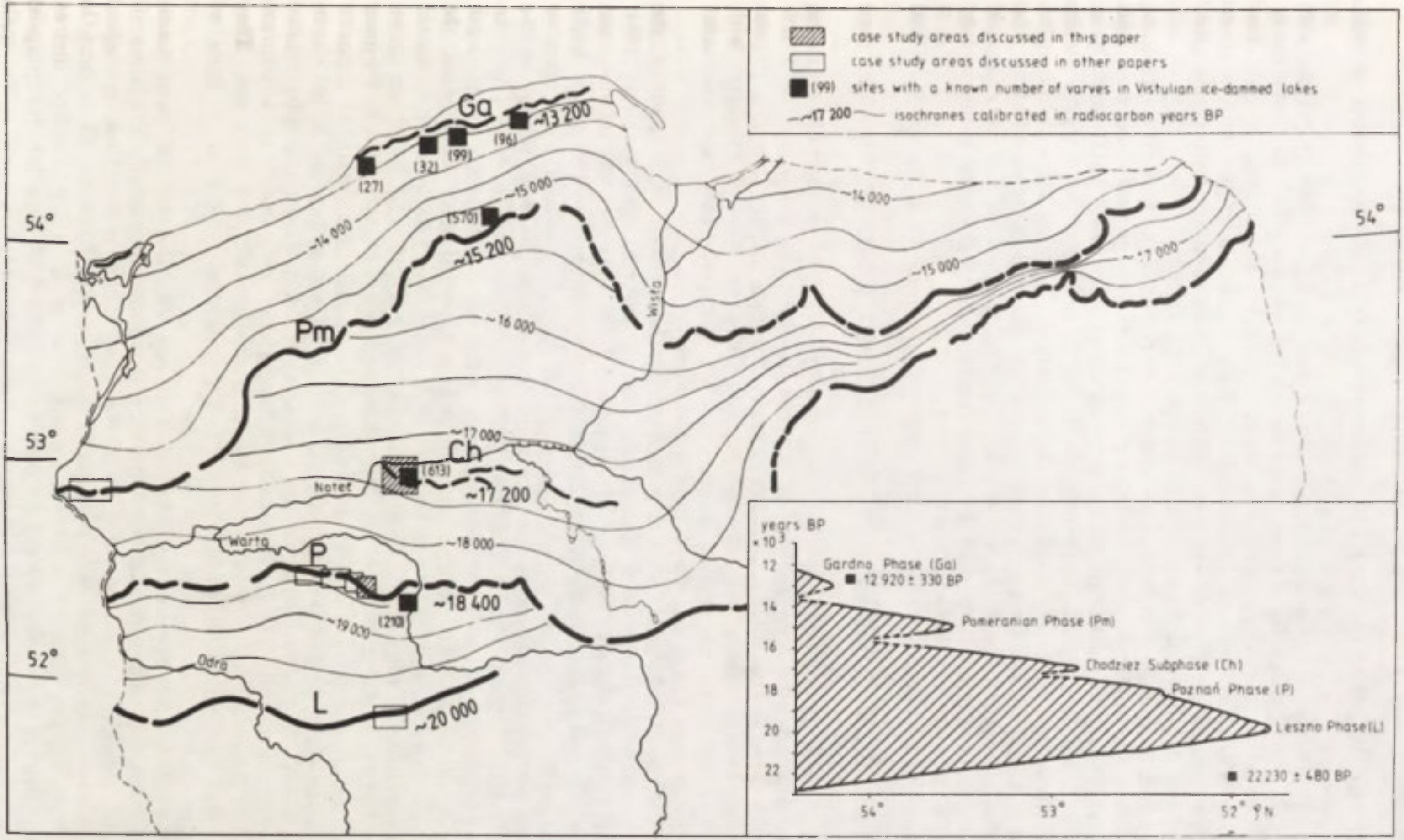


Fig. 1. Deglaciation isochrones and estimated age of major ice-sheet positions in N Poland
Inset: Time-distance diagram of the last ice sheet in Poland (after Kozarski 1986)

## DEPOSITIONAL MODELS AND ICE-SHEET BEHAVIOUR IN MARGINAL ZONES

Recently it has been established (Kasprzak and Kozarski 1984; Kozarski and Kasprzak 1987) that the former general models of the last ice-sheet decay (Table 1) and behaviour in marginal zones in nortwestern Poland are too far-reaching generalizations and simplifications. Such evaluation of the models arises from too general research procedures which have been used in studying marginal phenomena, i.e. their geological and geomorphological record.

New impulses to the interpretation of marginal phenomena and their palaeoenvironmental implications have stemmed from a new research approach (Kasprzak and Kozarski 1984; Kozarski in press; Kozarski and Kasprzak 1987) and have brought about new views concerning the ice-sheet behaviour in marginal zones or have provided better arguments for the discussion of the earlier well-recognized dynamic stages of the ice-sheet front. The new approach is represented by the so called complex lithofacies analysis of deposits that serves as an analytical tool. It is an expanded version of the facies analysis known from the sedimentology and is concerned with the establishment of

(1) all diagnostic properties of each sediment, resulting from its structure and texture;

(2) spatial distribution and thickness of sediments, including, if possible, their three-dimensional analysis, and

(3) geologic interrelationships between deposits and their geomorphic positions.

The complex lithofacies analysis of deposits has been applied so far to a few test sites lying within the main last ice-sheet positions in northwestern Poland (Fig. 1). Two of them will be discussed in brief by way of example, with special reference to the main elements of the analysis. They are associated with the Poznań Phase and the Chodzież Subphase.

*Poznań Phase ice-marginal features.* Analysis has been made of a part of the marginal zone developed as a depositional scarp (Kasprzak and Kozarski 1984, Kozarski and Kasprzak 1987) 5 m high, which is built up of ablation and fluvioglacial sediments. There is an extensive depression behind the scarp, while an outwash plain occurs in its foreland. The scarp is replaced by a number of longitudinal closed depressions which are surrounded by low ridges, in the northwestern portion of the test site 24 km$^2$ in the total surface area. An extensive gravel and sand pit, 56 excavated small pits 2.5 to 3.5 m in depth and 22 boreholes down to the depth of 4.5 m provided basic data about the thickness, distribution, structure and texture of deposits. The data were supplemented by material derived from about 40 additional shallow pits and boreholes down to the depth of 1.0 to 1.5 m. An old geological map (Assmann and Dammer 1916) has proved very helpful in determining the spatial distribution of deposits.

Diagnostic properties have served as a basis of recognition of 8 lithofacies units which are orderly arranged as a scheme of local lithostratigraphy related to geomorphic units. Thus, postulates (Clayton and Moran 1974) which are necessary to construct a simple process-form glacial model have been met. They have also proved useful in discussing the depositional model as a basis of reconstruction of the ice-sheet front behaviour.

*Chodzież Subphase ice-marginal features.* This test site lies in the westernmost part of the marginal zone on the fringes of the Toruń-Eberswalde Pradolina. It is nearly identical in surface area as it occupies 25 km$^2$. The basic geological and sedimentological data were collected from large pits down to 30 m deep in the glassworks at Ujście. The section lengths were up to a few hundred metres. A few tens of boreholes were also used as a source of data for the purpose of determining the horizontal distribution of sandy sedimentary units in the main and their thickness. Geomorphological maps have been employed throughout the analysis of the surface distribution of deposits and relief features (Kozarski 1962).

As is the case for the exmaple described previously, six lithofacies units have also been distinguished there on the basis of diagnostic properties. Owing to the deformations found, kinetostratigraphic units (Berthelsen 1978) have also been studied in detail. They have proved very useful in interpreting the dynamic character of the marginal zone under consideration and its morpho- and lithostratigraphic positions, and next, in building up a depositional model.

*Depositional models.* It is rather easy to distinguish icemarginal zones formed in high energy conditions because, on an average, they are well marked in the relief as thrust ridges and push moraines, and in the substratum as macroglaciotectonic deformations. However, many difficulties arise from dynamic interpretations concerning marginal zones produced in low energy conditions, i.e. in a steady-state position (= dynamic equilibrium), and during disintegration of the ice-sheet margin. The "dead ice model" is usually proposed (Bartkowski 1967, 1969) for such conditions. However, it has been shown recently (Kozarski in press; Kozarski and Kasprzak 1987) that this model cannot be considered universal since there are sedimentological procedures applicable to clear identification of deposits left by an active ice margin in a steady-state position or by a non-active disintegrating one.

Therefore, a new methodological approach has been presented (Kozarski in press). It is based on:

(1) the complex lithofacies analysis (Kasprzak and Kozarski 1984; Kozarski and Kasprzak 1987);

(2) the aggregation of sedimentary properties (structure and texture) of diagnostic value for the ice-front dynamics interpretation (Kozarski in press);

(3) the construction of local summary sequences (LSS) with a definite dynamic tendency ($D_t$) based on depositional efficiency ($D_e$).

The clustering of deposits and structures of diagnostic value for various dynamic states of the ice-sheet margin is considered a necessary first step involved in the depositional model construction. The second step comprises volume computation or mean thickness estimation concerning the lithofacies units. It is possible to carry out the volume computation in cases when geometric parameters (three--dimensional analysis) of lithofacies units have been measured. As rather rarely are such cases encountered, the mean thickness can be only used as a basis of the estimation of depositional efficiency ($D_e$) if limited data are available. Under favourable conditions, i.e. when a threedimensional analysis of lithofacies units has been made, the depositional efficiency can be calculated as

$$D_e = L_1 + L_2 + L_3 + \ldots L_n$$

where $L$ is the volume of a lithofacies unit (= lithosome), $D_e$ can be largely estimated merely as the mean thickness of a lithofacies unit. However, it is also useful in the drawing of a local summary sequence (LSS) and in comparisons between lithofacies units formed under known depositional conditions associated with an active and/or non-active ice-sheet margin. Thus, as can be seen on the scale, the mean thickness of various lithofacies units also reveals the depositional efficiency ($D_e$) which is of significance for the dynamic tendency ($D_t$) reconstruction of the former ice-sheet margin.

In spite of differences in data base for both cases, the dynamic tendency ($D_t$) reconstructed for the ice-marginal zone under investigation can be expressed as a ratio between depositional efficiency, for instance, in a steady-state position ($D_{ess}$) and depositional efficiency during ice-sheet recession and disintegration ($D_{ere}$), i.e.

$$D_t = \frac{D_{ess}}{D_{ere}}$$

When a slow advance occurred before the steady-state position of the ice margin has been reached, the equation has the form of

$$D_t = \frac{D_{ea} + D_{ess}}{D_{ere}}$$

In conditions of full data supply by the three-dimensional analysis of lithofacies units, the result of $D_t$ reconstruction is numerical. In the case of limited data base the dynamic tendency can be only expressed as an advantage of one depositional efficiency over the other. For instance,

$$D_{ess} > D_{ere} \quad \text{or} \quad D_{ea} + D_{ess} > D_{ere} \quad \text{and}$$

$$D_{ess} < D_{ere} \quad \text{or} \quad D_{ea} + D_{ess} < D_{ere}.$$

The above new methodological approach has been adopted recently and used (Kozarski in press) for the reconstruction of dynamic tendencies of the ice-sheet margin in northwestern Poland. Local summary sequences (LSS), together with diagnostic depostis and primary sedimentary environment description and dynamic tendency ($D_t$) determination are regarded at the present stage of research as depositional models for case study areas associated with the main ice-margin positions. Two of them are shown in this paper (Fig. 2) as examples.

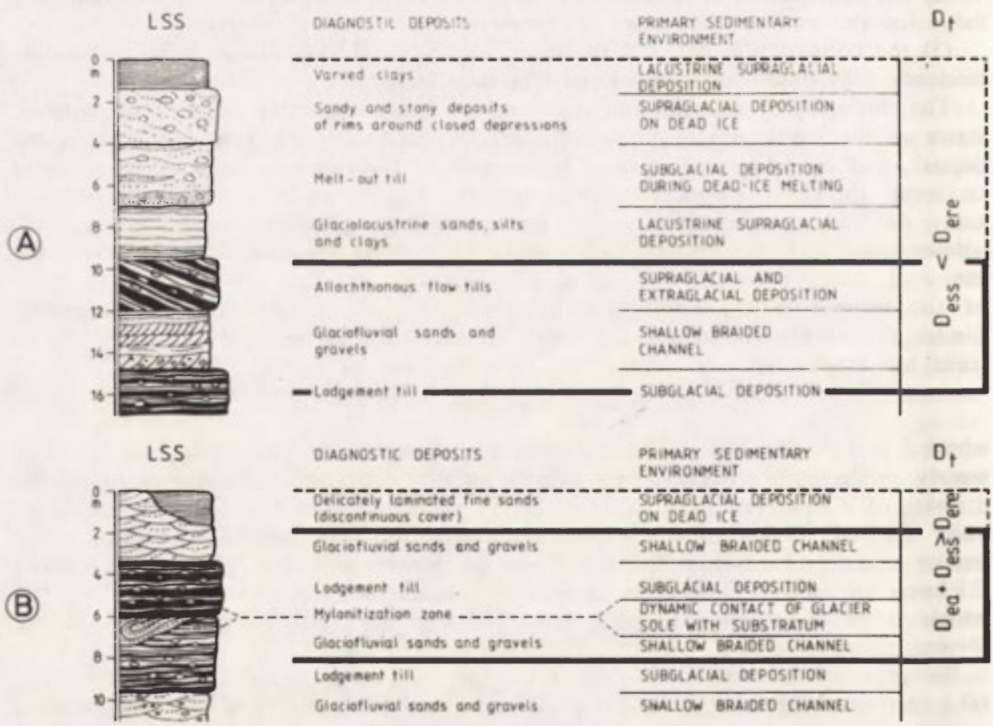


Fig. 2. Depositional models for case study areas within marginal zones of the Poznań Phase (A) and Chodzież Subphase (B); after Kozarski (1987)

## CONCLUSIONS

The above interpretation and discussion promotes a view that progressing detection of radiocarbon dated organic sediments in northwestern Poland has given rise to new possibilities of the determination of a time interval at which the last ice sheet advanced so far as the maximum extent line and then, it retreated as far back at the Baltic coastal zone. This interval is assigned to the ca 22 000–13 200 yr BP period with reference to not only sites in Poland and the neighbouring countries, but also, to those in Sweden (Berglund et al 1976) and in Denmark, as has been found out recently (Petersen 1984).

The determination of a time interval at which the glacial episode took place has also allowed a time-scale calibrated in radiocarbon years to be used for the estimation of age of the main last ice-sheet positions during recession. It approaches radiocarbon dates available for organic deposits included in horizons in the known profile Grand Pile (Mörner 1981). As concerns the Pomeranian Phase, there is also a certain similarity between the age estimate given in this article and that of ca 16 000 yr BP (?) which is provided by Boulton et al (1985).

It also appears that sedimentary interpretations, research procedures, methodological approaches and research results concerning the last ice-sheet marginal zones in northwestern Poland, as presented here in brief, are of importance for a few reasons.

Firstly, they reveal that traditional, routine interpretations of sediments, structures and forms in the marginal zones of the last ice sheet call for reinterpretation in order to produce more realistic models of the ice-sheet front bahaviour in the main positions during deglaciation.

Secondly, they show how to combine sedimentological tools with geomorphic research and interpretations in the complex facies analysis which is the most efficient research procedure at the present stage of study.

Thirdly, they offer possibilities of advanced palaeogeographic reconstructions which can refer at least to (1) ice-flow direction and character, (2) general thermal conditions and tendencies at the ice-sheet margin, and (3) ice-front behaviour.

Fourthly, they allow objective description of morphogenetic conditions in the marginal ice-sheet zones by means of simple depositional models, as well as application of such models to the testing of theoretical ice-sheet models.

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# LATE WEICHSELIAN AND HOLOCENE ENVIRONMENTAL CHANGES IN BOHUSLÄN, SOUTHWESTERN SWEDEN

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ABSTRACT. Biostratigraphical studies on palaeoenvironmental changes have been carried out on sediments from 12 basins in the central part of Bohuslän, southwestern Sweden.

Diatom, pollen and radiocarbon analysis have been applied to reconstruct a shore displacement curve for the region.

The palaeoenvironmental changes contemporary with the Pleistocene/Holocene boundary and the Holocene transgression maximum have been of special interest.

Biostratigraphical studies on palaeoenvironmental changes have been carried out on sediments from 12 basins situated between 135 and 12 meters above the present sea level in central Bohuslän on the west coast of Sweden (Fig. 1). Earlier 8 of the sites in the same area have been investigated very thoroughly pollenanalytically by Fries who studied the regional development of the lanscape and vegetation (Fries 1951). He constructed a preliminary shore displacement curve, which was mainly based on the occurrence of marine dinoflagellate cysts (*Hystrix*) and colonies of freshwater greenalga *Pediastrum* in the sediments (Fig. 2). The material was not radiocarbon dated.

The primary intention of the present investigation was to study and date the main sea level changes during the Late Weichselian and the Holocene by diatom and radiocarbon analysis and use the results to reconstruct a shore displacement curve for the region. At the same time a more complete picture of the development of the sedimentary environment and its relation to sea level in the central part of Bohuslän would be achieved.

By means of diatom analysis the isolation of the basins from the North Sea (the Skagerrak) was determined and the sediments radiocarbon dated. Pollen analysis was used for preliminary dating of the sediments and for comparison with Fries' material.

Some results were reported or published in connection with the investigations attempting to find a marine stratotype site for the Pleistocene/Holocene boundary on the Swedish west coast (Miller and Robertsson unpublished report 1974; Miller 1982; Robertsson 1982).

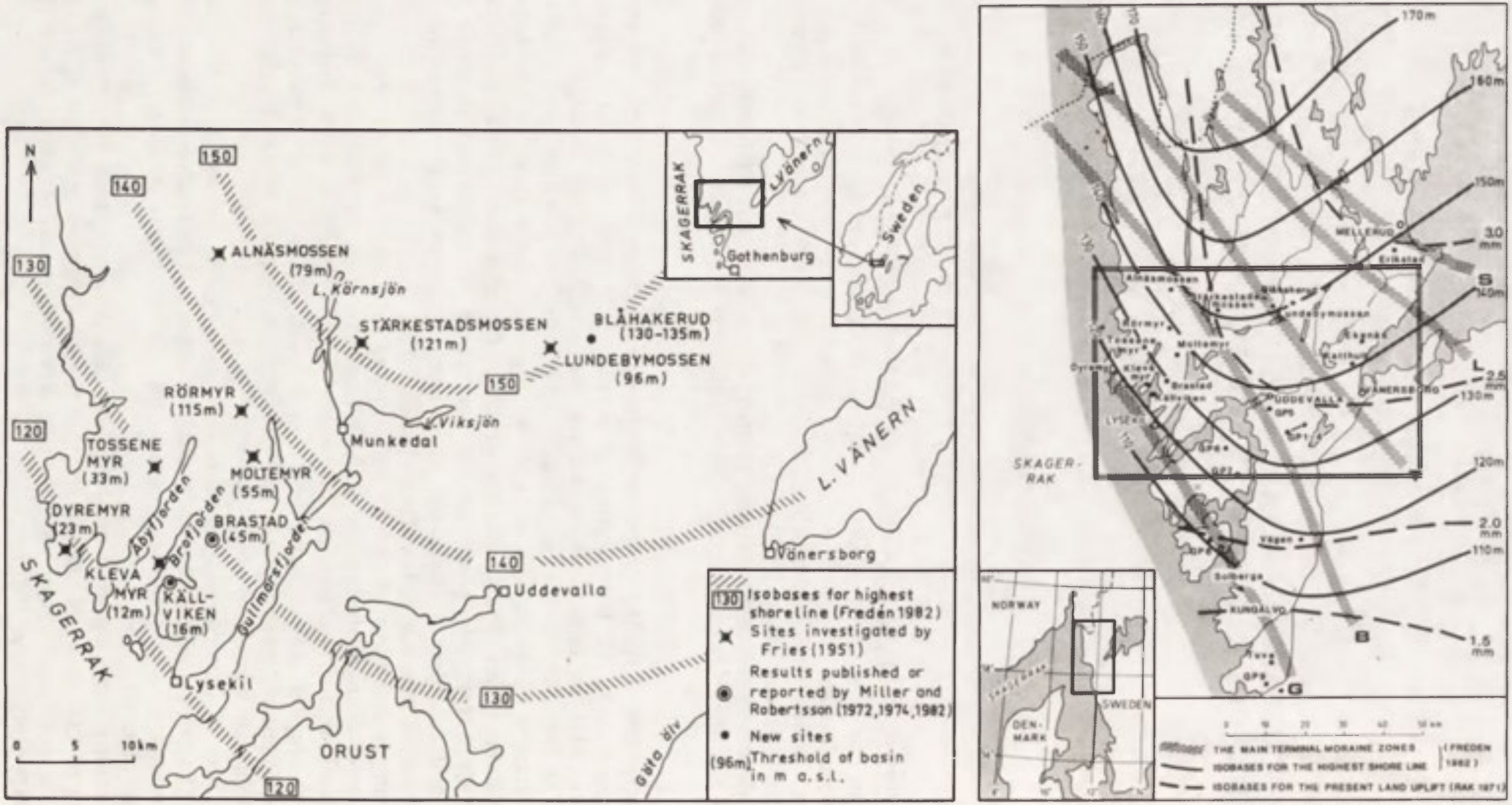


Fig. 1. Location of the investigation area in Bohuslän southwestern Sweden

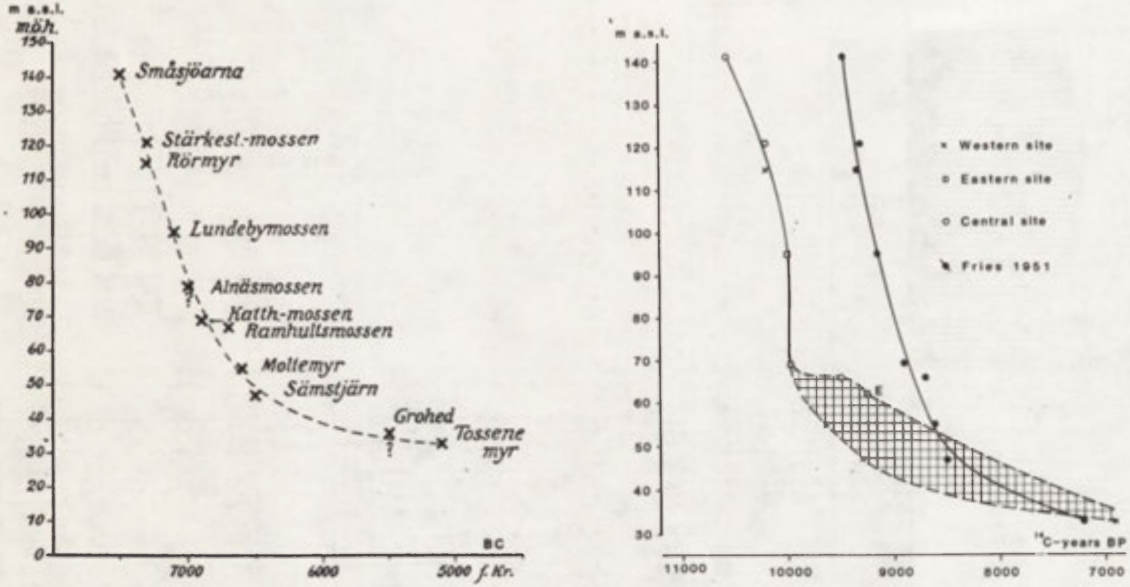


Fig. 2. Shore displacement curve according to Fries (1951, Fig. 43) and Svedhage (1985, Fig. 34)

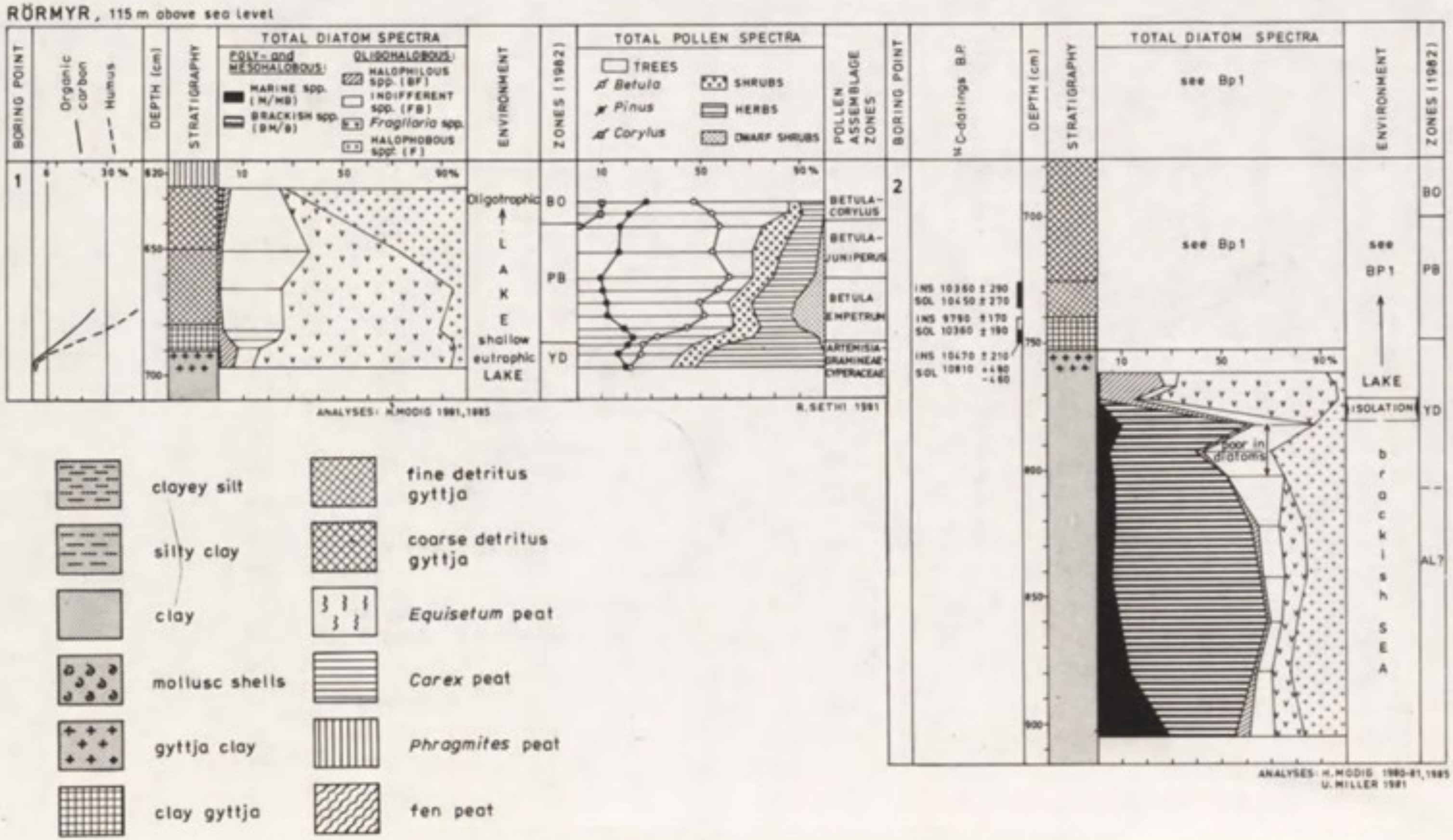


Fig. 3. Diatom and pollen diagram of the core from Rormyr, altitude 115 m above present sea level. Key of the stratigraphical symbols used in Figs 3–6

Changes of the palaeoenvironmental conditions prevailing around and after the Pleistocene/Holocene boundary are registered in the fossil diatom and pollen floras. Diatoms are representative of the aquatic environment and pollen of the terrestrial and to some extent the aquatic. The main part of the sediments in the basins studied represent marine as well as brackish and freshwater conditions bacause of the isostatic land uplift combined with effects of the eustatic sea-level changes in the area. The vegetational changes from an open landscape dominated by herbs and shrubs to birch woods are registered very clearly in the pollen diagram (Fig. 3).

Sediments from one of the basins studied, Moltemyr, have been analysed also in foraminifera and molluscs (Knudsen 1982; Feyling-Hansen 1982; Klingberg 1984). The interpretation of the Moltemyr stratigraphy in connection with the Pleistocene/Holocene boundary project was based on analyses of the strata down to 6.5 m (Fig. 4). Two alternatives were given where to place the Pleistocene/Holocene boundary: 350–345 cm or around 500 cm (Miller 1982, pp. 192–193, 208; Robertsson 1982, pp. 234–237; Knudsen 1982, pp. 154–159; Feyling-Hansen, 1982 pp. 126–132). New corings have penetrated the sediments down to 16.3 m. The thickness of the sedimentary strata in the Moltemyr basin is more than 27 m (Cato 1982).

The preliminary diatom stratigraphy in the sediments between 16.3 and 4.2 m indicates repeated changes in the environmental development of the basin. These

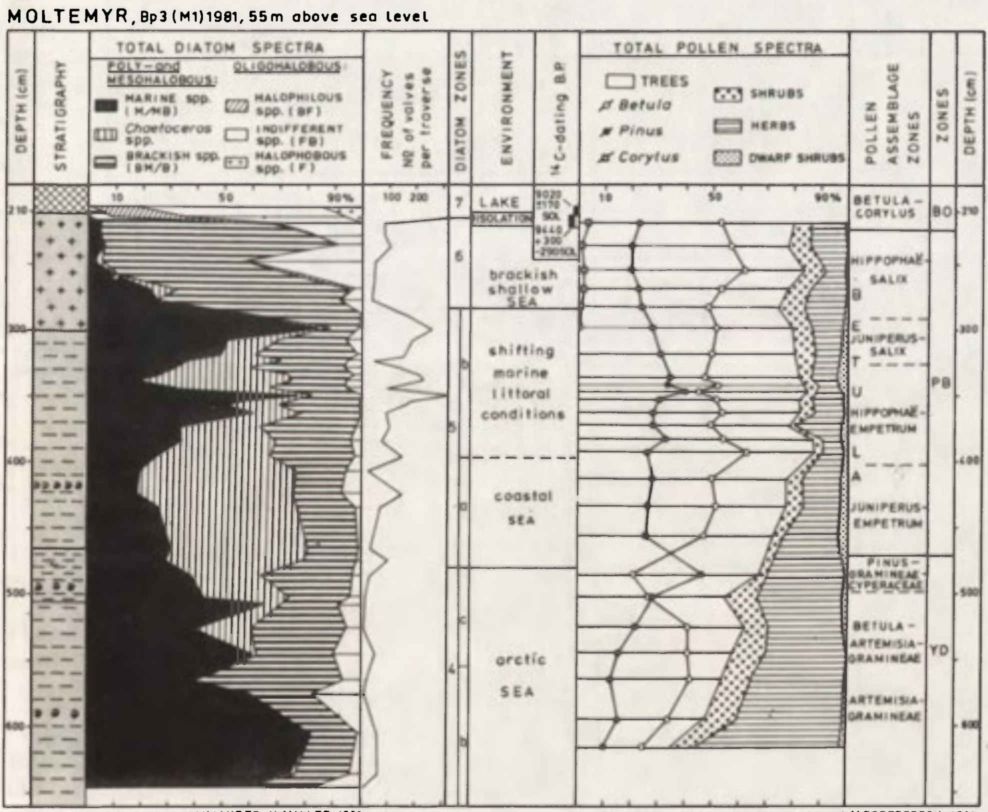


Fig. 4. Diatom and pollen diagram of the core M1 from Moltemyr, altitude 55 m above present sea level. For stratigraphical symbols see Fig. 3

changes were caused by varying sedimentary and climatic conditions influenced by distance to the continental ice margin, presence of sea-ice, altitude of the sea-level, quantity of meltwater outflow and reworking of older sediments.

Two artic phases were identified according to diatom stratigraphy: around 16–13 m and 8–5 m (Fig. 5). They may correspond to Older Dryas (DR 2) and Younger Dryas (DR 3). In this case the intervals with higher diatom productivity (13–8 m and 5–2 m) represent Allerod (AL) and Preboreal (PB). This interpretation is in good accordance with the foraminiferal stratigraphy alternative 2 (Klingberg 1984) and the pollen stratigraphy (Robertsson 1982). If alternative 1 of the Pleistocene/Holocene boundary stratigraphy is applied to the sediments the artic phases must correspond to Oldest Dryas (DR 1) and Older Dryas (DR 2). The warmer phases (5–4 m) consequently should be Bölling (BO) and Allerod (AL). The question is when were the sediments between 27 and 16.3 m deposited? They may represent the Middle-Weichselian Gota alv Interstadial (Brotzen 1961; Miller 1964, 1982).

The isolation of the Moltemyr basin from the sea was completed at the transition clay gyttja/gyttja (2.15 m) just after the rational *Corylus* limit at the transition Preboreal/Boreal.

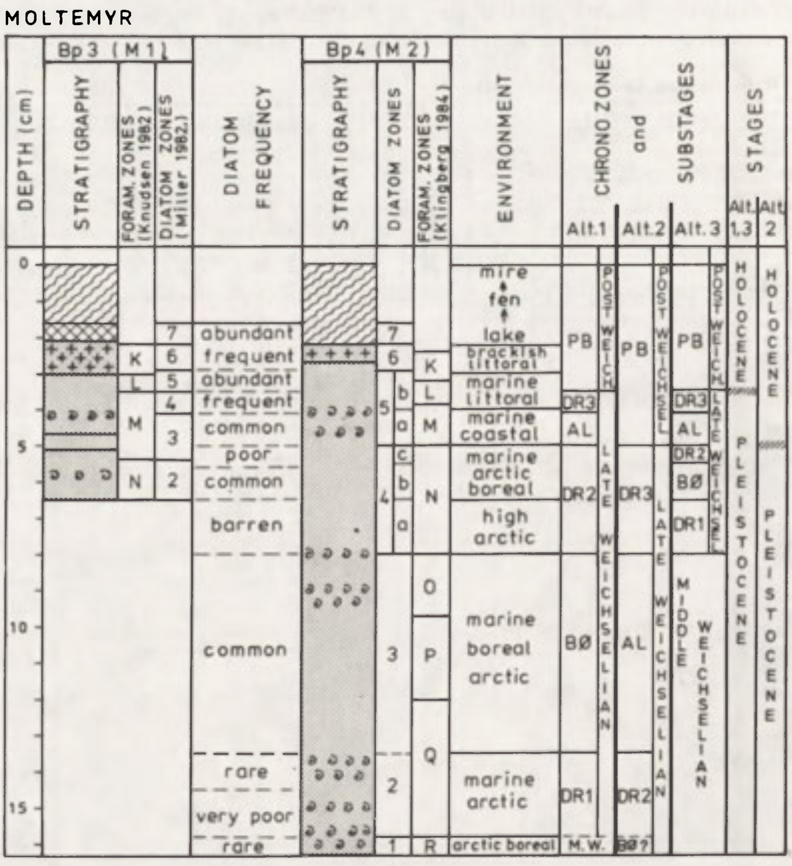


Fig. 5. Correlation of foraminiferal and diatom zones in the cores from Moltemyr with alternative stratigraphical interpretations (Knudsen 1982; Miller 1982; Klingberg 1984). For stratigraphical symbols see Fig. 3

If the Pleistocene/Holocene boundary at Moltemyr is placed at 5.0–4.8 m the diatom flora in the sediments of the proposed Solberga type site indicates a similar change at 22–21 m (Miller 1982, Fig. 16:5). In this case the radiocarbon dating 9170 ± 400 BP (T–2982) of shells between 19 and 18 m in the Solberga core can be considered more reliable than the datings of the shells in the Moltemyr core $9160\pm^{1460}_{1210}$ BP, U–4408, Olsson 1982), but still too young.

In general the radiocarbon datings of the gyttja sediments deposited around the isolation phases of the basins studied have yielded ambiguous ages (Olsson 1982). This is obvious particularly when dating the sediments around the Pleistocene/ /Holocene boundary. A marked change in the stable carbon isotope ratio $^{13}C/^{12}C$ at the Pleistocene/Holocene boundary in southern Sweden has newly been reported (Hakansson 1986).

The Late Weichselian and Holocene shore displacement and stratigraphy have been studied recently in southwestern Sweden (Björck and Digerfeldt 1982; Påsse 1983; Dennegård 1984; Svedhage 1985; Fredén 1986). To those studies can be added three Norwegian investigations dealing with Late Weichselian and Holocene shore displacement in southern and central Norway (Stabell 1982; Krzywinski and Stabell 1984; Kjemperud 1984).

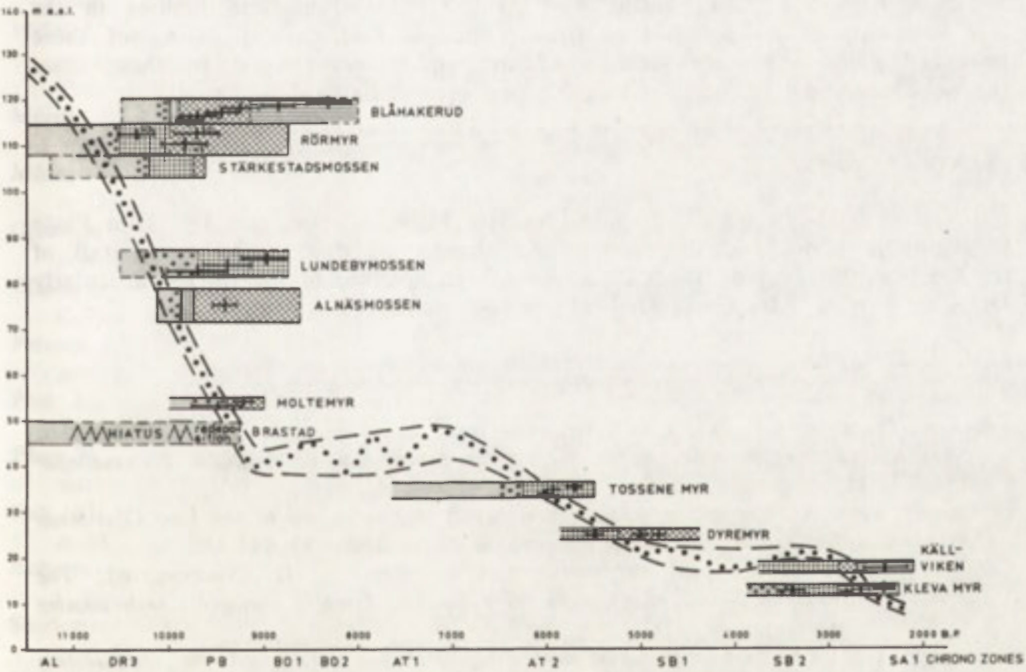


Fig. 6. Shore displacement curve for Central Bohuslän based on diatom and pollen analytical record combined with radiocarbon dating. For stratigraphical symbols see Fig. 3

In the central part of Bohuslän the isolines for the present land uplift (shore displacement) vary between 2 and 3 mm/year (Fig. 1). The isolines for the highest coastline vary between 120 and 150 m. The shore displacement curve for central Bohuslän is calculated in relation to 130 m highest coast line isoline and 2.5 mm/year land uplift isoline (Fig. 6). The curve which represents the time period 11 000–2 000 BP can be divided into 5 parts:

1. The Late Weichselian and Early Holocene (YD–PB) steeply falling regressive part (11 000–9000 BP)
2. The Early-Middle Holocene (BO–AT1) undulating part with several transgressions and regressions (9000–6000 BP)
3. The Middle Holocene (AT2) regressive part (6000–5000 BP)
4. The Middle-Late (Upper) Holocene (SB) undulating part (5000–3000 BP)
5. The Late (Upper) Holocene (SB2 – SA1) regressive part (3000 – 2000 BP)

According to the biostratigraphical studies the Pleistocene/Holocene boundary in the central part of Bohuslan is indicated by a series of repeated palaeoenvironmental changes, which took place during the approximative time period 10 500–9500 BP. The combination of eustatic lowering of sea level and increasing isostatic land uplift resulted in an extremely quick regression of shoreline. Around 9000 BP the regressive trend was changed to transgressive. The big postglacial Holocene transgression maximum (Sandegren and Johansson 1931; Thomasson 1934; von Post 1947; Morner 1969) seems to have been caused by several transgressive phases during 9000 to 6000 BP (Persson 1973; Mörner, ed. 1976; Pásse 1983). This is also confirmed by the results of the present study. In the central part of Bohuslan the sea-level changes which took place during the transgressive period, 9000 to 6500 BP, have strongly affected basins situated between altitudes of 40 and 50 m above present sea-level. Repeated transgressions have caused redeposition and hiauses in the few sedimentary basins found at these altitudes. Radiocarbon dating of these reworked gyttja sediments yields confusing and reversed ages. In these cases the pollenanalytical record is a valuable and more reliable dating tool.

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# LEVELS IN SUBGLACIAL CHANNELS AND THEIR SIGNIFICANCE IN DETERMINING THE CHANNEL ORIGIN AND EVOLUTION

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ABSTRACT. The present author points out that only in some large subglacial channels of the Dobrzyń Lakeland there are glacial till horizons indicative of ice-sheet participation in their formation. In subglacial channels cut out subglacially by glacial water, which prevail in the study area, the horizons are likely to occur as fluvioglacial erosional levels. There are fluvioglacial accumulative horizons along subglacial channels where outwash water flowed away, independently of their origin, while kame terraces occur in some of large channels filled up with dead ice. There exist lake terraces in subglacial channels in which there were or still are lakes, whereas fluvial terraces are present in considerably transformed channels. Analysis of these levels reveals the origin of channels and allows establishment of channel evolution stages during the Lateglacial and Holocene.

## INTRODUCTION

Subglacial channels, also called, particularly in Denmark, tunnel valleys, constitute a very important element of the relief of areas glaciated during the last Scandinavian Glaciation (Vistulian, Weichselian). According to Majdanowski (1950), the average density of channels in the south Peribalticum area is 4.59 km/100 km$^2$. In the lake districts of northern Poland it amounts to 8.12–10.93 km/100 km$^2$ and ranges from 0.1 to 36.3 km/100 km$^2$. In his calculations Majdanowski included both the subglacial and ice-walled channels.

Subglacial channels were first identified at the end of the 19th century and discussion concerning their origin has been continued ever since. Recent summaries of the discussion are found in Galon's (1965, 1983) and Kozarski's (1966–67) articles in the Polish literature. These articles and others on the subject under consideration have underlined the complexity of the problem. Without going into detail, it can be stated that there are several main views concerning the origin of subglacial channels occurring in the south Peribalticum area.

At least since Ussing (1903) carried out his investigations in the Jylland Peninsula, the opinion has become prevalent that subglacial channels formed as a result of subglacial erosion by glacial water. Another idea put forward by Woldstedt (1954) and Gripp (1964), for example, assumes that the chief process leading to the formation of most subglacial channels was glacial erosion and in particular, erosion by comparatively narrow ice lobes adavancing out of an ice sheet. Landforms, the formation of which was due to this kind of erosion are called Glaziellen by Gripp (1964). It is also assumed that in some cases the ice lobes entered the existing wide channels to cause their transformation. Another view is (e.g. Kozarski 1966/67, Liedtke 1975) that a part may have been played in the formation of some subglacial channels by both glacial and glacial water erosion.

As concerns the area of Poland covered by the Scandinavian ice sheet during the last glaciation, the prevailing opinion is that nearly all subglacial channels occurring in this area were subglacially eroded by glacial water (Majdanowski 1950; Galon 1965, 1983; Kozarski 1966/67 and many others). Nevertheless, several examples of transformation of the existing channels by ice lobe erosion have been reported from this area (Kozarski 1966/67; Pasierbski 1979). There are also a few authors who think that the formation of some subglacial channels was due to glacial erosion.


Fig. 1. The study area
x – distribution of geomorphological maps included in the paper

The problem of further evolution of subglacial channels is not less important than their formation, independently of the way in which they originated. This problem is much less frequently the subject of investigations and, as demonstrated by Niewiarowski's (1986) studies, for example, it is not less complex. In order to throw some light on the evolution of subglacial channels, it is necessary to study in detail the morphology and geological structure of subglacial channels and their close surroundings. The studies of subglacial channels situated within the Vistulian Glaciation extent lines in the Kuyavian, Chełmno, Dobrzyń and Brodnica lake regions (Fig. 1) have revealed that some of them contain clearly marked levels differing in origin, which can be one of the indicators of the formation and evolution of subglacial channels. Among these levels, there are erosional and accumulative ones, those of glacial and meltwater origin, lake and fluvial levels. Analysis of these levels by way of actual example is the subject of this paper.

## GLACIAL AND GLACIO-AQUEOUS LEVELS IN SUBGLACIAL CHANNELS AND THEIR CLOSE SURROUNDINGS

An example of a subglacial channel of complex origin is the Brynica river channel believed so far to have been cut out by meltwater. It is situated within the extent lines of the Kuyavian subphase of the Vistulian Glaciation, not far from its maximum extent line. It lies in a wide zone of a flat and undulating moraine plateau in the hinterland of the end moraines of the Kuyavian subphase and perpendicular to them. It adjoins an outwash plain in its distal part. Although morainic hills are marked in the Radoszki region in Fig. 2, their origin has as yet not been studied in any detail.

North of the Brynica channel there are several subglacial channels more or less parallel to it but those exhibit typical characteristics of a channel eroded subglacially by meltwater (they run in a winding course, are comparatively deep and narrow, have uneven bottoms with swells and pools). Unlike those channels, the Brynica channel is short (up to 10 km) and wide (up to 4 km). Within it, especially in an extension on the north side, there is a flat level up to 1 km wide, lying 10–15 m below the moraine plateau surface at 99–100 m a.s.l. and separated from it by a pronounced scarp. Narrow fragments of this level also occur in the southern portion of the channel. The level is built up of till reaching in places the thickness of 5–6 m but in other places a thin layer of till is underlain by fluvioglacial or limnoglacial deposits. Both the orientation of stones in the till with their long axes parallel to ice movement and the till structure indicate that this is lodgement till. As there are no traces of subglacial erosion by meltwater within the till horizon, it cannot have been formed so. The occurrence of a lodgement till horizon indicates that it is due to erosive action of ice (the formation of a scarp between the till horizon and the moraine plateau) and to the accumulative action (the till horizon). This horizon developed under the active ice layer while it was in slow movement. During ice stagnation other landforms developed in the Brynica channel, such as an esker and numerous sand hillocks differing in origin. On the south side of the Brynica channel there is also a sandygravelly horizon where structure and texture of deposits prove that it is a fluvial level. This would indicate that while dead ice was still ocurring in the Brynica channel, the earliest water flow of the rivers Pissa and Górznianka, derived largely from the melting dead ice on the moraine plateau, flowed at its margin towards the Drwęca ice marginal streamway.

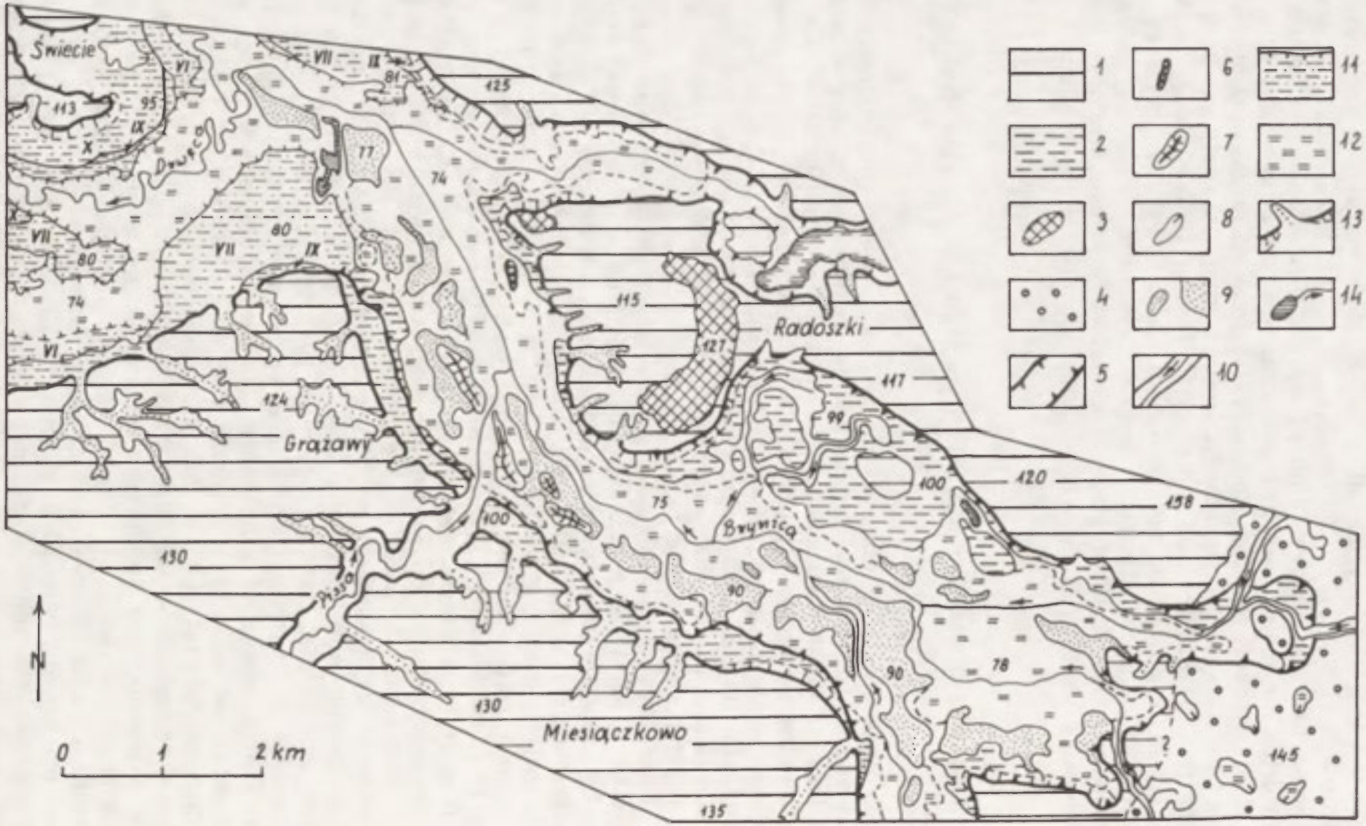


Fig. 2. Geomorphological map of the Brynica subglacial channel and its surroundings: 1 – moraine plateau, 2 – till horizon in subglacial channel. 3 – morainic ridges. 4 – outwash. 5 – subglacial channels. 6 – limnoglacial kame. 7 – esker. 8 – kettles. 9 – sandy elevations of different origin. 10 – valleys. 11 – river terrace. 12 – peat plains. 13 – Lateglacial and Holocene denudative and erosional small valleys. 14 – lakes and rivers

It seems very likely that the deepest peat-filled part of the channel having an uneven bottom was due to subglacial erosion by glacial water flowing in the final stage of active ice or in the initial stage of ice stagnation. It is thus an example of a channel, in the formation of which both ice and glacial water took part and a great variety of landforms in the channel points to the complexity of evolution of this form which is however not analysed in detail in this paper.

Analysis of the relationship of this landform to the earlier fossil relief shows that there existed there a wide and deep valley of Mazovian Interglacial age along its course. Therefore, it can be assumed that before the last ice-sheet advance there was a tendency towards the formation of an ice stream within the ice sheet. The stream moved faster than that on the neighbouring moraine plateau, which deposited the till horizon referred to earlier. The formation of the till horizon has been presented in Fig. 8, 1A and 1B.

The second example of a subglacial channel of composite origin is the Ruziec channel in the hinterland of end moraines belonging to the Kuyavian subphase, in the zone of a flat and undulating moraine plateau (Fig. 3). Its course is also perpendicular to the end moraines and its pecularity is that it cuts through the Zbójno drumlin field. Investigations carried out so far have shown that it consists of several sections, the southern one showing characteristics typical of a channel cut out by glacial water and the northern one being of more complex origin. The latter contains a till horizon up to 0.5 km wide, situated 5–15 m below the moraine plateau surface and separated from it by a pronounced scarp. This horizon is largely built up of lodgement till. Evidence for this is provided by its compact structure, clay contents and orientation of pebbles conformable to the direction of ice movement, but in places also to ablation (flow) till. There are no traces of erosive action of meltwater in this horizon. It should be stressed that this till horizon is related to the level at which drumlins occur. It is an additional argument for its glacial origin. It was then formed subglacially in active ice which probably moved there somewhat faster than on the moraine plateau.

A deep channel cuts into this horizon, further southward, it has the characteristics of a typical channel cut out by meltwater and there is no till horizon within it. This indicates that either while the ice sheet was still in motion or during its stagnation, a channel was eroded in a depression formed by ice by subglacially flowing water. Then the channel was filled with dead ice. Evidence for the occurrence of dead ice in the channel and in its surroundings are kame hillocks of kame limnoglacial type, built up of horizontally rhythmically stratified fine sands and silts. Besides the kame hillocks, there also exist similarly built kame terraces. In the case of this channel there are glacial and glacio-aqueous levels. Kame terraces have been reported from subglacial channels in other areas of northern Poland (Pasierbski 1975; Andrzejewski 1984 and others). The development of a glacial level in subglacial channels and a channel cut out by subglacial water is presented in Fig. 8, 2A and 2B and that of kames and a kame terrace is illustrated in Fig. 8 4A and 4B and 5A and 5B.

Apart from the above examples, Nechay (1927) made inferences as to the participation of glacial erosion in the formation of a wide subglacial channel in the region of Kikół in the Dobrzyń Lakeland. However, it should be stressed that the lake districts under investigation contain nearly exclusively subglacial channels formed by meltwater and that examples of channels where direct involvement of ice action can be proved, as is the case for till horizons, are scarce and have not been found so far outside the Dobrzyń Lakeland.

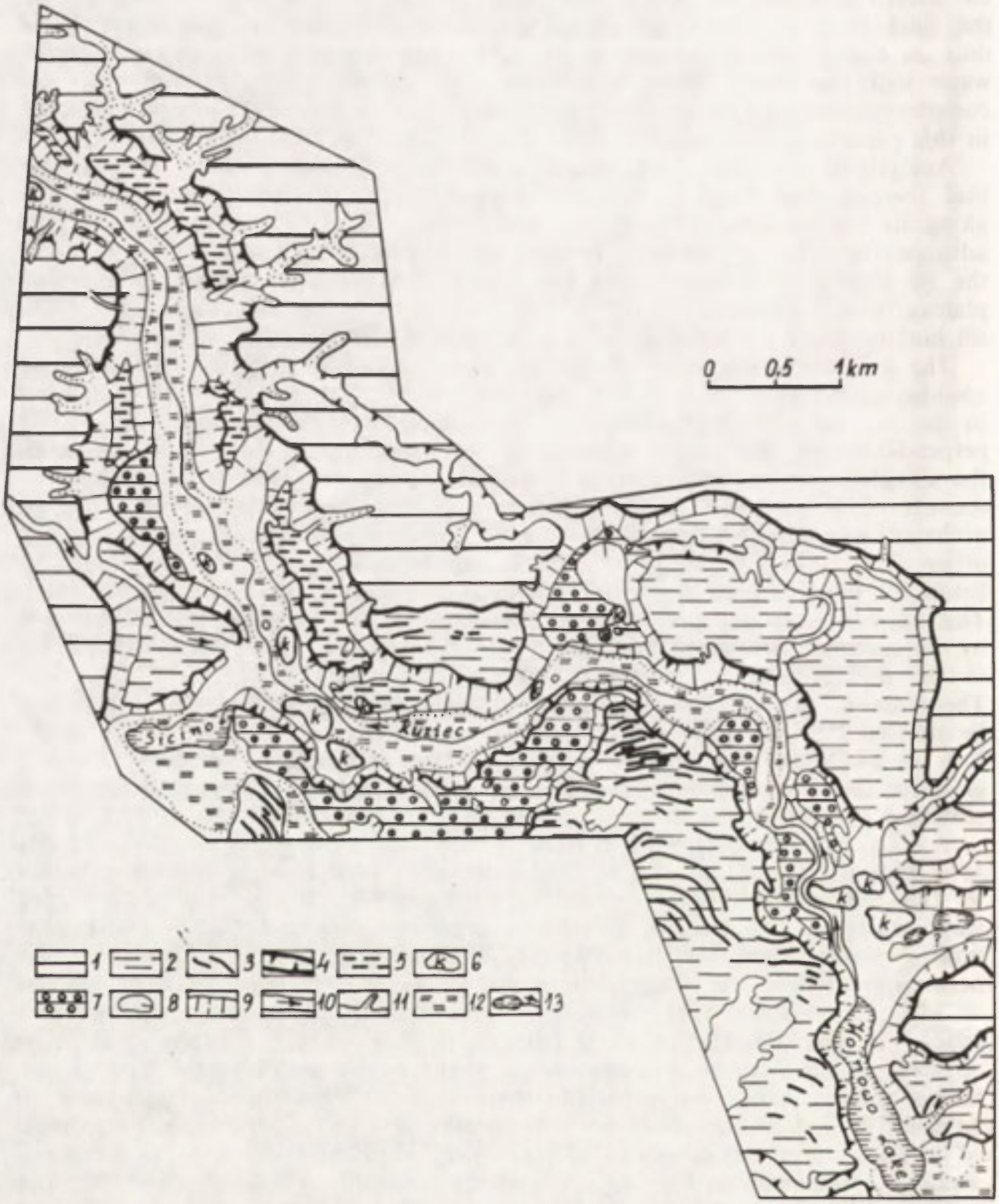


Fig. 3. Geomorphological map of the Ruziec subglacial channel (northern section) and its surroundings: 1 — moraine plateau, 2 — drumlin field level, 3 — drumlins, 4 — subglacial channels, 5 — till horizon in subglacial channel, 6 — kame hillocks, 7 — kame terrace, 8 — kettles, 9 — slopes, 10 — valleys, 11 — Lateglacial and Holocene denudative and erosional small valleys, 12 — peat plains, 13 — lakes and rivers

While considering the origin of subglacial channels, Galon (1965) reports that in the Byszewo subglacial channel in the Krayna Lakeland there is a level which, according to him, was subglacially cut out by meltwater as a result of flow blockage by ice in the main flow channel and water flow alongside the ice (Fig. 8 3A and 3B). It would be then an erosional level cut out subglacially by meltwater. Evidence for this is supposed to be the fact that there are traces of water flow (a thin cover of fluvioglacial deposits, erosional pavement) on the till at that level. This does, however, not rule out the possibility that water flowed subaerially and not subglacially. The problem will be solved after the completion of new investigations.*

The present research carried out in the above area has revealed only one case which may suggest similar origin of a level in the subglacial channel. This is associated with the Skarlin channel in the Brodnica Lakeland (Fig. 4). This channel is part of a larger marginal channel (30 km in length) developed in the marginal zone of the Krajna–Wąbrzeźno subphase (Niewiarowski 1986). It runs parallel to the end moraines and perpendicularly to the ice movement. This alone, as well as the fact that the higher level of the Brodnica outwash begins near that channel rule out its formation by glacial erosion and support the idea that its origin was due to meltwater erosion.

Part of this channel is occupied by Lake Skarlin 293 ha in surface area and 15.1 m in maximum depth. Along the lake shores there are fragments of a flat level up to 0.8 km wide, raised 15–18 m above the lake surface, but lying several to a dozen or more metres below the moraine plateau surface and separated from it by a pronounced scarp (slope). On the north side the level height corresponds with that of the flat moraine plateau occurring to the west. It is joined from the north by small channels cut out by meltwater, which may have taken part in its formation. There is, however, an essential difficulty in accounting for the origin of this level as a result of subglacial


Fig. 4. Geomorphological map of the Lake Skarlin channel and its surroundings: 1 – moraine plateau, 2 – end moraines, 3 – outwash, 4 – subglacial channels, 5 – level in subglacial channel (disputable origin), 6 – kettles, 7 – valleys, 8 – Lateglacial and Holocene denudative and erosional small valleys, 9 – slopes, 10 – lake terraces, 11 – peat plains, 12 – lakes

* Investigations carried out in 1987 have shown that the above mentioned level was formed by subaerial meltwaters.

erosion by meltwater for it is built up of till, including ablation till more than 2 m in thickness and in some places even lodgement till with no traces of water erosion. On the other hand, such traces have been found in the slope separating this level from the moraine plateau. Therefore, it cannot be ruled out that the level was cut out subglacially and the ablation till is a result of the melting out of the collapsed ceiling of the ice tunnel. Cases of ice collapsing in ice tunnels have also been reported from the present-day glaciers (Embleton and King 1969). However, it must be pointed out that though theoretically, the development of levels in subglacial channels as a result of subglacial erosion by meltwater cannot be ruled out, its evidence has been insufficient so far. In the case of the Skarlin channel, it cannot be excluded that the level found there is one of the moraine plateau levels occurring in this area (Niewiarowski and Wysota 1986).

It is, however, indisputable that there may exist levels in subglacial channels, which developed under subaerial conditions, namely outwash levels and kame terraces. The existence of the latter ones has been documented, e.g. in the Ruziec channel (Fig. 3), whereas good examples of outwash shelves in channels or their surroundings are provided by channels in the Brodnica Lakeland. In this lake district like in others, there exist subglacial channels, the origin of which is related to particular stages of glaciation and deglaciation; therefore, they are of differing age.

A peculiarity of the subglacial channels of the south Peribalticum and in particular, of areas where deposits due to the last glaciation reached considerable thickness is the fact that, irrespective of their origin, they were preserved for a long time by buried ice. The term "buried ice" is applied here without reference to its origin. It may then have been glacial ice or the so called winter ice derived from frozen meltwater. It was buried as a result of covering with ablation till or till flows from the neighbouring higher areas or with glacio-aqueous sediments, mostly fluvioglacial ones. That cover, cold climate and the presence of permafrost caused buried ice to be left in the subglacial channels for a long time.

At the time of the standstill position of the ice-sheet margin along the line of the end-moraines Konojady-Wawrowice, subglacial channels occurring in the foreland of the ice sheet in the Brodnica Lakeland had already been preserved by buried ice. This is testified by the fact that they were not filled with outwash deposits. The outwash water spread widely near the ice-sheet margin and flowed southward. However, the moraine plateau surface becomes raised in that direction. Thus, it flowed in trickles along large radial subglacial channels, e.g. along the Brodniczanka channel (Fig. 5) or the channel of the lakes Ciche-Zbiczno-Bachotek (Fig. 6).

Evidence for the outwash flow in the Brodniczanka channel (Fig. 5) is provided by outwash shelves 100–500 m wide, built up largely of fine sands up to 5 m thick and present within it or in its neighbourhood. The outwash surface slopes down 0.4–0.6% southward and the sand fraction becomes finer in that direction. The fact that the channel was filled up with buried ice at that time is testified by highly diversified topography of its bottom with differences of up to 20 m in relative height, moving from the lake surface and those of up to 40 m, moving from the bottom, and by the occurrence of till hummocks below the outwash shelf surface without a cover of outwash deposits. Similar fluvioglacial levels occurring on both sides of channels are found in other channels of the Brodnica Lakeland and have been reported from many other areas of northern Poland. Their relationships to the channels are often similar to those of kame terraces (Fig. 8, 4A and 4B).

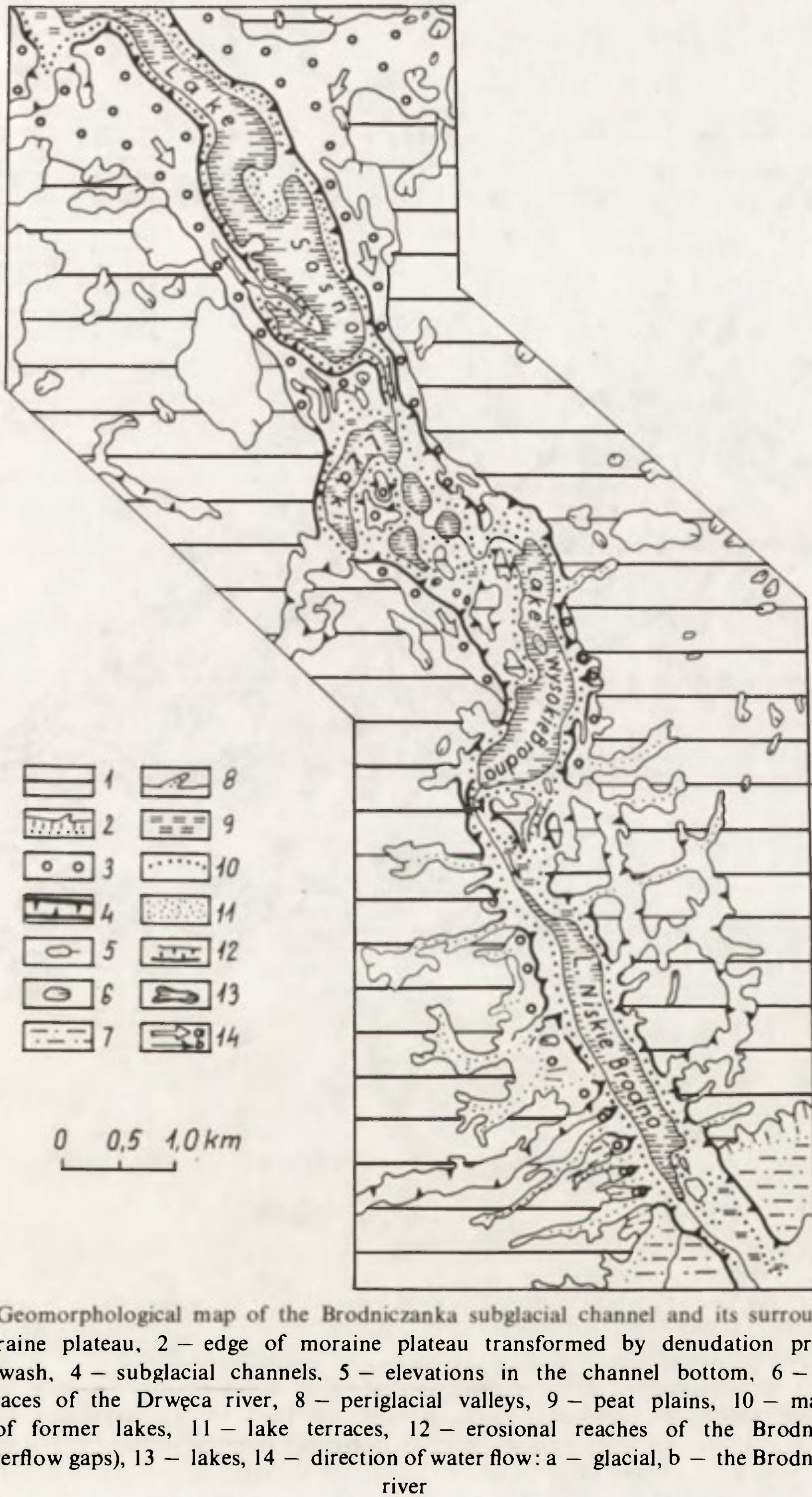


Fig. 5. Geomorphological map of the Brodniczanka subglacial channel and its surroundings: 1 – moraine plateau, 2 – edge of moraine plateau transformed by denudation processes, 3 – outwash, 4 – subglacial channels, 5 – elevations in the channel bottom, 6 – kettles, 7 – terraces of the Drwęca river, 8 – periglacial valleys, 9 – peat plains, 10 – maximum extent of former lakes, 11 – lake terraces, 12 – erosional reaches of the Brodniczanka river (overflow gaps), 13 – lakes, 14 – direction of water flow: a – glacial, b – the Brodniczanka river

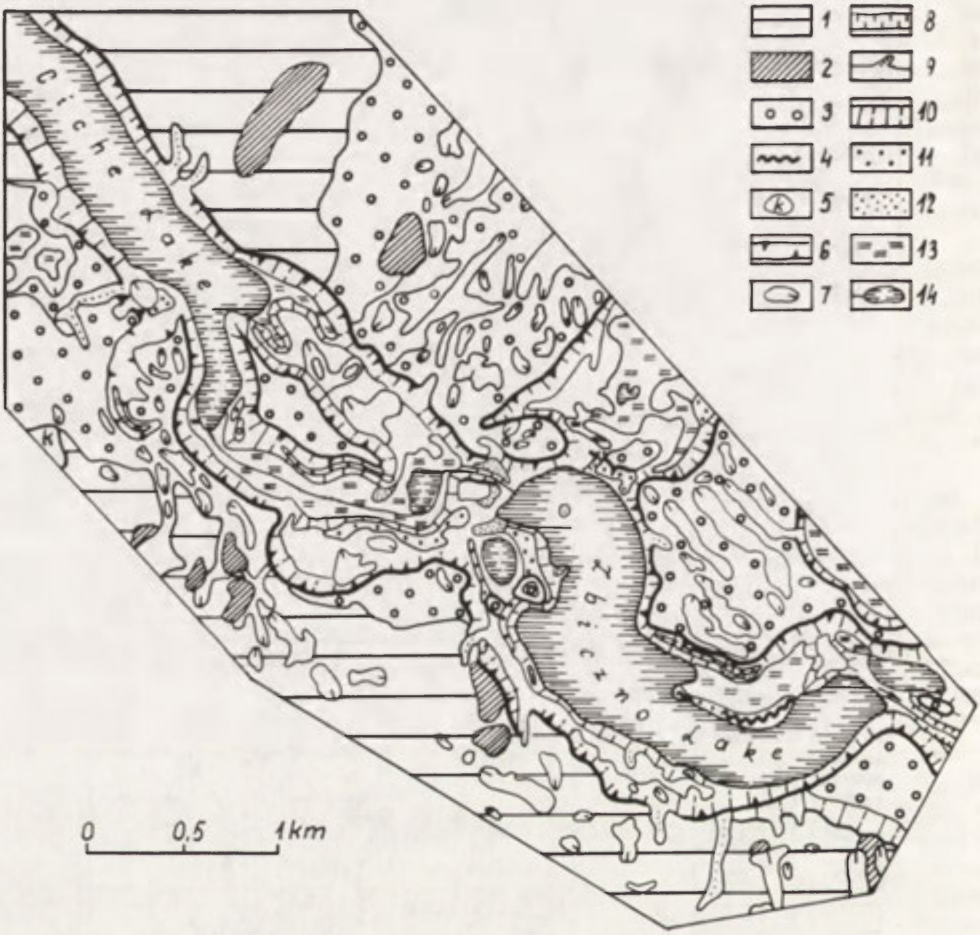


Fig. 6. Geomorphological map of part of the lakes Ciche-Zbiczno-Bachotek channel: 1 — moraine plateau, 2 — dead ice moraines, 3 — outwash, 4 — esker, 5 — kame, 6 — subglacial channels, 7 — kettles, 8 — overflow gaps, 9 — Lateglacial and Holocene denudative and erosional small valleys, 10 — slopes, 11 — Lateglacial (pre-Allerod) lake terrace, 12 — Sub-Atlanctic lake terraces, 13 — peat plains, 14 — lakes and rivers

## LAKE AND FLUVIAL LEVELS OR TERRACES IN SUBGLACIAL CHANNELS

As has already been mentioned, a common phenomenon in Poland's territory affected by the Last Scandinavian Glaciation and in the surrounding area was the occurrence of buried ice in subglacial channels. The melting out of the ice took long and was not simultaneous, depending on the thickness of the ice and the blanketing mantle and on whether the channel was used by a river or not. The river cut through the sedimentary cover and produced faster ice melting, also as a result of the thermal influence of water. In cases of considerable ice thickness, such as in the Brynica channel (Fig. 2), the river initially ran along the channel margin. It was already at the end of the Pleistocene when the ice sheet was still covering the northernmost area of Poland. Its age can be established by analysing the correspondence of the fluvial level with terraces along the Drwęca ice marginal streamway. In other subglacial channels of the area under investigation the development of river flow took place in different stages of the Lateglacial, most frequently during the Oldest Dryas and in some cases even during the Holocene.

Subglacial channels provided favourable conditions for the development of lakes because of the melting out of buried ice. Many workers claim that lakes of northern Poland appeared as late as during the Alleröd which had been preceded by a-few-thousand-year-long lakeless period. The present author's investigations carried out in the Brodnica Lakeland (Niewiarowski 1986, 1987) revealed the presence of lacustrine deposits belonging to a generation earlier than the Alleröd in some subglacial channels. This is exemplified by the channel of the lakes Ciche-Zbiczno-Bachotek (Fig. 6) in which there are fragments of a lake terrace lying 10–20 m below the outwash level and 7–8 m above the surface of the present-day lakes. The lacustrine character of the terrace is testified by deposits that it is built up of, i.e. fine sands, the percentage of which reaches up to 82%, lacustrine marl and chalk with up to 83% $CaCO_3$ contents. The calcareous deposits are up to 1.1. m thick. The lake in which they had been deposited was relatively shallow and developed above the buried ice that still filled the channel. This is testified by the following facts: (a) kettle occurrences in the lake terrace, (b) correspondence of the lake terrace with the Drwęca river terrace which developed most probably during the Older Dryas, (c) detailed investigations of bottom sediments in Lake Strażym located in the same channel have demonstrated that the melting out of buried ice ceased at the end of the Alleröd (Niewiarowski 1987) and lakes have been present in the channel ever since.

Similar age of dead-ice decay in channels was recorded in Greater Poland (Kozarski 1963) and in the Kuyavian Lake District (Niewiarowski 1978). Obviously, it cannot be excluded that this process may have ceased to operate in the deepest channels later on. Irrespective of the time of dead-ice decay in channels, cases of shallow lakes occurring prior to the Alleröd and developed above the buried ice have also been found in other channels outside the Brodnica Lakeland.

The statement that subglacial channels contain lake terraces belonging to an older generation, i.e. of pre-Alleröd age, as well as those of present-day lakes, testifying oscillations in their water levels is most significant for the problem considered in this paper. Two lake terraces belonging to the younger generation, i.e. of Subatlantic age, have been found in the Brodnica and Kuyavian lake districts. The lake terraces of differing age present in subglacial channels and numerous lakes still existing there validate the distinction in the evolution of channels in buried-ice conservation and lacustrine phases.

Subglacial channels which were or are used by rivers have been, to a greater or lesser extent, transformed by them. The extent of transformation depends

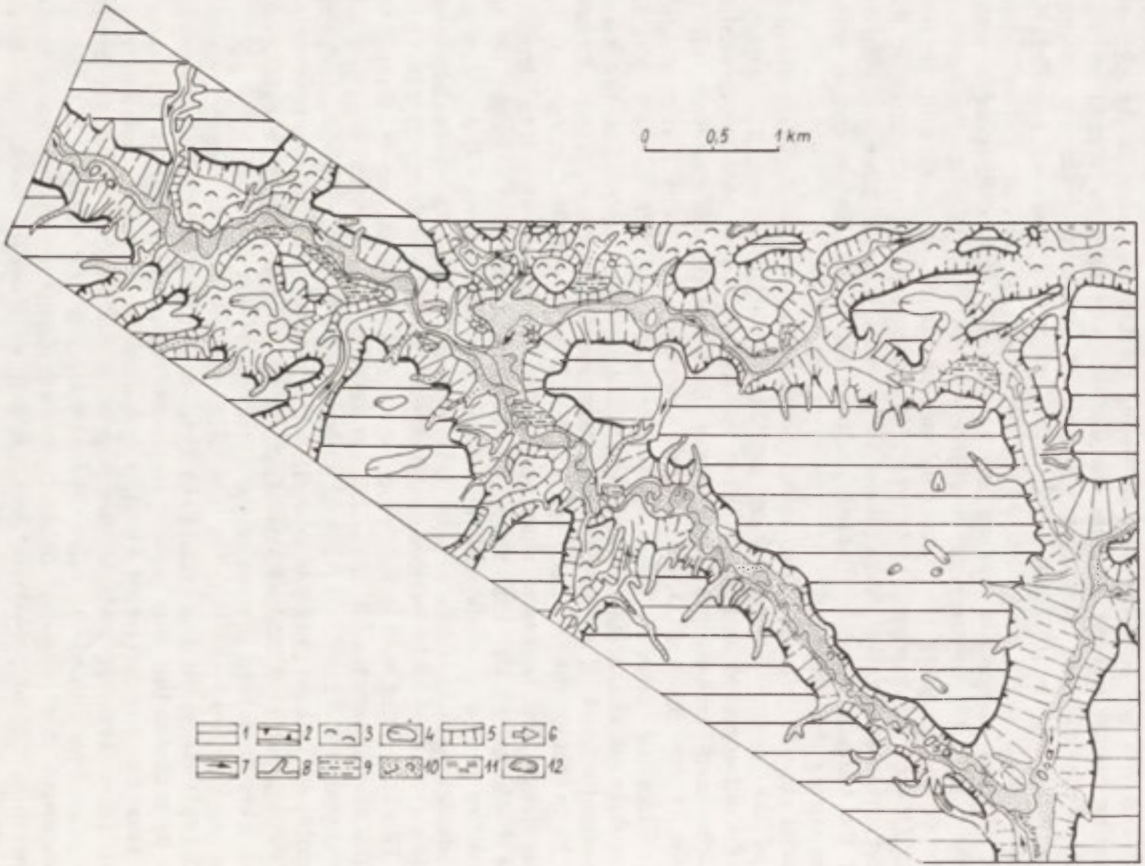


Fig. 7. Geomorphological map of the Osa valley and its surroundings: 1 – moraine plateau, 2 – subglacial channels, 3 – bottoms of large subglacial channels, 4 – kettles, 5 – slopes, 6 – valleys abandoned by the Osa and Lutryna rivers (dry), 7 – other valleys, 8 – Lateglacial and Holocene denudative and erosional small valleys, 10 – floodplain with meanders, 11 – peat plains, 12 – lakes

on the size of a river and on the time it started in the channel. Hence, there are channels only very slightly transformed, e.g. the Brodniczanka channel (Fig. 5), as well as channels changed considerably. A good example of considerable transformation of several channels by a river is a reach of the Osa valley (Fig. 7). From the fragments of terraces preserved in this reach, it can be inferred that the oldest of them corresponds with terrace VI or V in the lower Vistula valley. The age of this terrace can be approximately identified with the Oldest Dryas (*sensu lato*), which determines the time when the Osa became a tributary of the Vistula.

The Osa river first developed along the line of several shallow marginal channels and then, also used a fragment of a deeper radial channel. It follows from Fig. 7 that the earlier flow of the Osa took place down the northern channel, while the southern channel was used by the Lutryna river. After the buried ice in the deeper channel of the Lutryna had melted away, which most probably took place during the Allerod, the whole outflow of the Osa directed itself down the southern channel and a long section of the former valley became dry and dead. Flowing down the southern channel, the Osa has transformed and deepened it considerably (the present-day Osa valley is 40–55 m deep). The abandoned fragments of the subglacial channel are hanging above the present valley bottom. The deepening of the valley proceeded in stages reflected in the form of five fluvial terraces, mainly erosional, which are cut in the channel sides (higher) or in the newly cut valley. Such instances of changes in the direction of river flow resulting from buried ice decay in subglacial channels, as well as different degrees of transformation and development of fluvial terraces in them have also been reported from other areas of northern Poland. This phase of evolution of subglacial channels is referred to here as the fluvial phase.

## CONCLUDING REMARKS

Selected examples from the study area testify the existence of glacial levels in some subglacial channels, which is indicative of the participation of ice-sheet erosion in their formation. In the area under discussion it concerns only some of channels of the Dobrzyń Lakeland. Channels cut out subglacially by meltwater are thus the prevailing type. Erosional levels cut out subglacially by meltwater are likely to occur in this type of channels. The existence of outwash levels and kame terraces in subglacial channels, irrespective of their origin, has been proved. Their development is associated with the phase of channel conservation by dead and buried ice. The older generation of lake terraces which are traces of shallow lakes that existed under periglacial tundra conditions is also linked to the existence of this ice in the channels. Lake terraces developed after the dead-ice decay are associated with the phase of present-day lakes. The development of these terraces is connected with Holocene oscillations in the lake water level. Higher lake levels than the present ones occurred during the Sub-Atlantic period.

Among lake terraces there have been fragments of abrasion terraces but accumulative terraces built up of lacustrine sediments prevail. Among these terraces there are some which are linked with the anthropogenic phase in the evolution of subglacial channels, i.e. with artificial lowering of the lake level, or as a consequence of artificial ponding of water level in lakes, new lake levels develop.

The fluvial phase must be distinguished in subglacial channels used by rivers. First steps and elevations in the bottoms of the subglacial channels are cut through; then, the water table of lakes is lowered until they often disappear

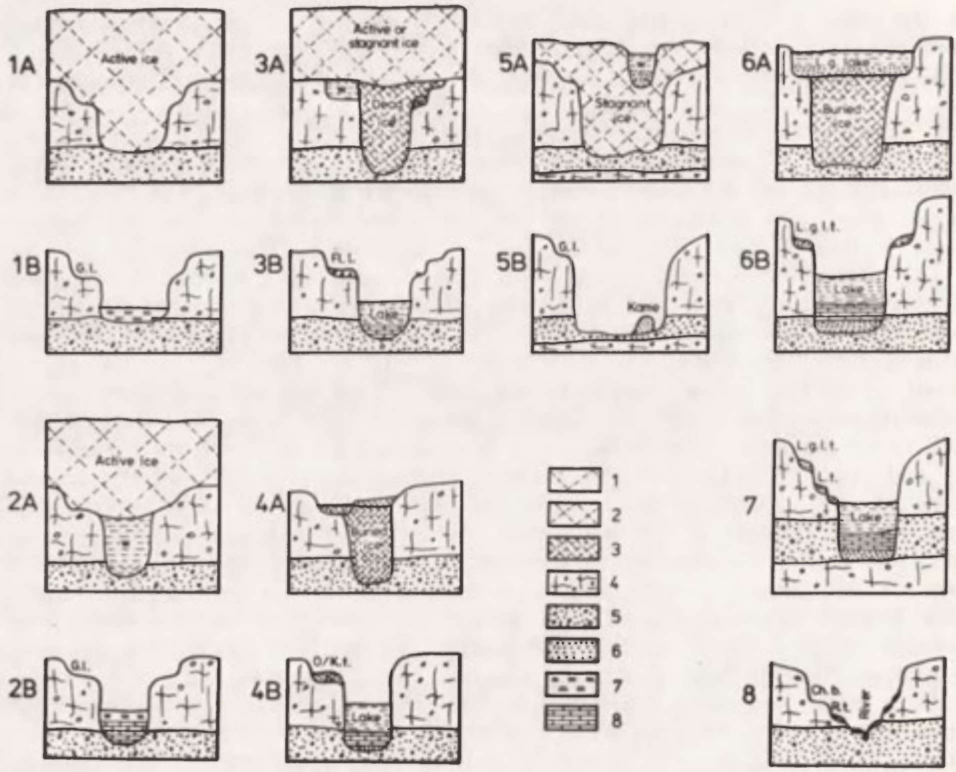


Fig. 8. Diagrams illustrating the origin of levels in subglacial channels: 1 — active ice, 2 — stagnant ice, 3 — dead and buried ice, 4 — till, 5 — glacio-aqueous deposits, 6 — sands of different origin, 7 — peat, 8 — calcareous deposits

Key to symbols: G.l. — glacial level, Fl.l. — fluvioglacial level, O/K.t. — outwash level or kame terrace, L.g.l.t. — Lateglacial lake terrace, L.t. — Holocene lake terraces, Ch.b. — subglacial channel bottom, R.t. — river terraces

completely; after that the channels are deepened and widened, and fluvial terraces are formed. Fig. 8 gives a diagrammatic picture of the way in which genetically different levels develop in subglacial channels. The knowledge of the origin of these levels provides one of the criteria which prove helpful in the study of the origin of subglacial channels and their subsequent evolution during the Lateglacial and Holocene.

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# LATE WEICHSELIAN AND HOLOCENE PALAEOMAGNETICS STUDIES IN SOUTH SWEDEN

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ABSTRACT. This paper summarizes the results of palaeosecular variation (PSV) studies carried out since 1983 in south Sweden. South Sweden is for a number of reasons considered to be a suitable area for these studies. Different types of sediment, covering different periods almost back to the deglaciation, have been analysed. From the Late Weichselian a repeatable pattern of PSV curves have been established. Sediment, dated both with radiocarbon chronology and varve chronology (absolute years BP), indicate a difference between the two time scales. Further investigations will be carried out to confirm these results. Long core palaeomagnetic analyses of a ca 20 m long Holocene sediment succession have resulted in a preliminary PSV curve from this period. With respect to the documented pattern and age of identified turning points, features from both the east European and west European master curves, as presented by Thompson (1983), are found. The preliminary result from this single profile are promising and further investigations from this site will be carried out in the near future.

## INTRODUCTION

Under favorable conditions lake sediment carry more or less detailed records of palaeosecular variation (PSV). This was first reported by Mackereth (1971). He showed that long period declination oscillations existed in the Lake Windermere post-glacial sediment. Since the pioneer work of Mackereth a number of PSV studies have been carried out. Based on records from Europe and other parts of the world Thompson (1983) presented dated Holocene PSV declination and inclination master curves for seven regions of the world, dated in tree ring calibrated years BP. No regular period or amplitude exists neither in the declination nor in the inclination, and the curves vary between regions. Master curves can serve as a dating tool of new sediment successions by matching the palaeomagnetic features. The west European master curves (WE) are based on records from Loch Lomond and Lake Windermere in Britain, published by Thompson and Turner (1979) and the east European master curves (EE) are based on records from Lake Paajärvi and Lake Lovojärvi in Finland, published by Huttonen and Stober (1980) and Tolonen, Siirianinen and Thompson (1975). PSV studies in USA (King et al. 1982) indicate that secular variaton patterns are reproducable on a regional scale from 1000 to 3000 km and Thompson (1983) concludes that a number of observations suggest that the regions will be at least 1000 km across. With respect to the British and Finnish sites, at a distance in both directions of about 1000 km, south Sweden takes up a strategic position in north Europe in the continuous study of the PSV.

From the Late Weichselian in Europe no master curves have as yet been published and the knowledge of the PSV from this time period are more limited, especially for north European sites. This might be due to difficulties of adequately dating pre-Holocene sediment as well as the absence of long Late Weichselian sequences. Data for the Late Glacial of northern Europe are also somewhat confusing as discussed by Creer (1985). The question is whether reports in the literature (e.g. Mörner et al. 1971; Mörner 1977; Noël 1975; Abrahamsen and Readman 1980) of large amplitude departures (so called excursions) of measured palaeomagnetic directions from the present axial dipole field direction is of geomagnetic origin or not. The aberrant palaeomagnetic directions have been explained as caused by mechanical sedimentation processes (Thompson 1976; Thompson and Berglund 1976) or as the result of other processes (Creer 1985). From central and eastern Europe records from the time period 9–21 000 BP have been published (e.g. from Lac du Bouchet (Bonifay et al. 1984) and from south Europe extending back to 25 000 BP (Creer 1974).

## STUDY AREA

For several reasons south Sweden is considered a suitable area for PSV studies of both Late Weichselian and Holocene sediment.

– The basement in south Sweden is dominated by igneous bedrocks which contribute to a comparatively high amount of ferrimagnetic minerals in the sediments. These are the main carriers of the remanent magnetization.

– As south Sweden is situated between the areas in northern Europe for which Holocene master curves already exist, well dated curves from this area would increase the knowledge and understanding of the behaviour of the geomagnetic field and the nature of the secular variations.

– Good possibilities to find long, suitable Late Weischselian sequences in south Sweden with a well defined chronostratigraphy.

– Possibilities to study PSV in different types of sediment dated by independent methods:

a) in lake sediments dated by pollenstratigraphy and/or radiocarbon analysis,

b) in varved clays dated in varve (calendar) years BP according to he Swedish Geochronological Time Scale.

By comparing the two sets of curves it should be possible to estimate the relation between radiocarbon years and varve (calendar) years back to the deglaciation. This is especially important for the Late Weichselian, where calibrations between the radiocarbon chronology and the calendar year time scales based on dendrochronology still does not exist.

– Possibilities of using the magnetic features as a chronological tool for sites where radiocarbon analysis cannot be used, as for example in the southernmost parts of Sweden.

## LATE WEICHSELIAN PALAEOSECULAR VARIATION

From the Late Weichselian (sensu Mangerud et al. 1974) PSV have been studied in sediments where

a) the age is obtained in radiocarbon years BP, based on $^{14}C$ analysis and/or pollenstratigraphy,

b) the age is obtained in varve years BP, based on the Swedish Geochronological Time Scale, wich recently was connected with the present (Cato 1987).
The former type consits of more or less organic sediments while the latter consists of varved clays.

PSV studies of varved clays are associated with special problems as these sediments generally are deposited in high energy environments resulting in coarse grain sizes, disturbances due to slumping and erosion, deflection of non-spherical grains due to currents etc. According to our experiences so far, varved clays can be used for PSV studies if the sediments:

1) are fine grained, which means deposited in a low energy environment far away from the ice margin,
2) have a homogenous lithology,
3) are horizontally deposited and without disturbances,
4) are still situated below ground water level. If not, the sediments become compacted, which tend to give too low inclination values,
5) cover a sufficient number of varves (= years) so that prominent magnetic features can be seen.

Both with respect to total sediment thickness and time, the longest and most detailed PSV records are obtained from three profiles in the Torreberga basin (Sandgren 1986), situated in the southernmost part of Sweden, 10 km south of Lund (Fig. 1). The 6 m long sequence of Late Weichselian sediment in

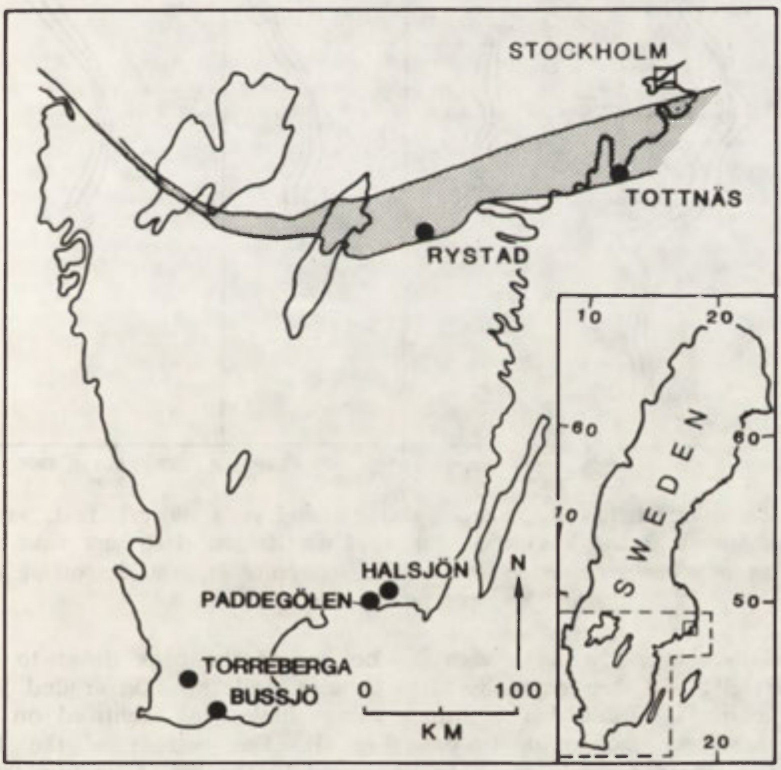


Fig. 1. Map of south Sweden with sites discussed in this paper. The light dotted area corresponds to the middle Swedish end moraine zone

Torreberga are dominated by clay with a very low organic content (3–5% loss on ignition) and span the time interval 12 800–10 000 $^{14}C$ years BP. In Sandgren (1986) dates were only available from a few levels but a complete pollen analyses on one of the profiles (profile 4) have now been performed. According to Jonas Ising (pers. comm.) there are only minor differences compared to the preliminary dates. The sediment in the bottom of the profile is possibly 200 years younger. Since the results could be repeated in the different cores the variations in the declination and inclination records are considered to reflect true variations in the magnetic field throughout this time period. As can be seen in Fig. 2 the declination in the bottom of the profile is close to north

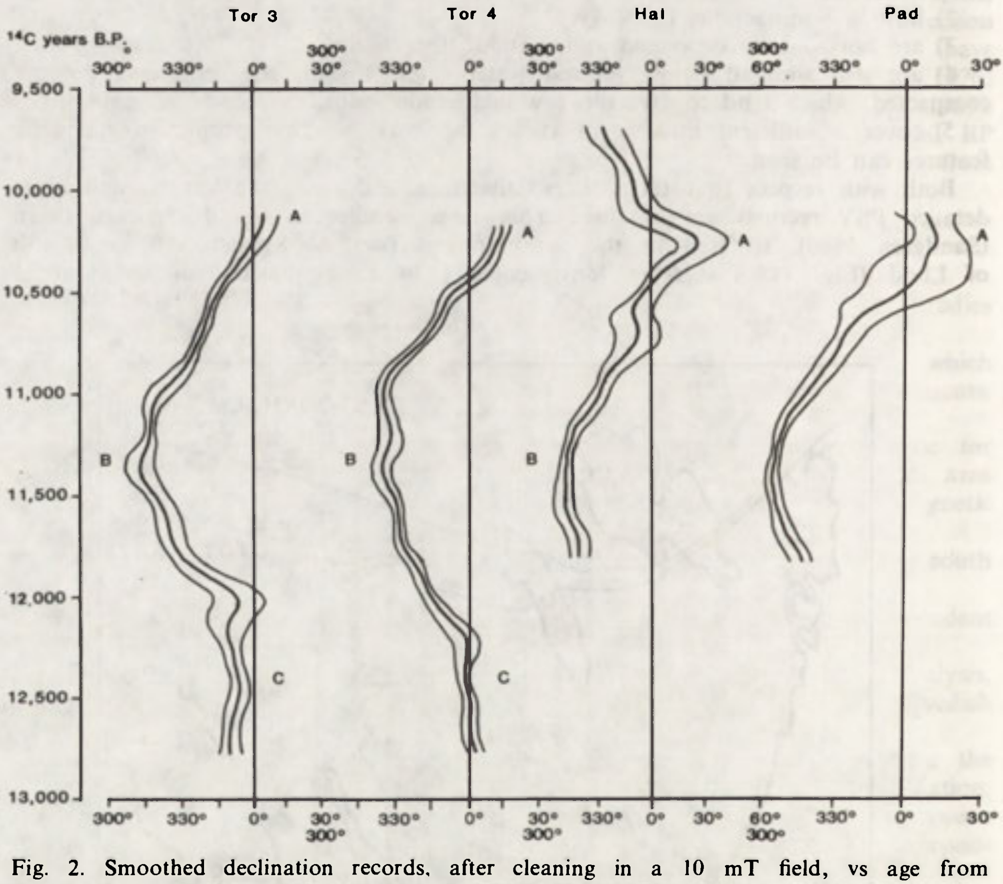


Fig. 2. Smoothed declination records, after cleaning in a 10 mT field, vs age from Torreberga profile 3 and 4 (Tor 3, Tor 4), Lake Halsjon (Hal) and Lake Paddegolen (Pad). The outer lines correspond to a 95% confidence interval. Smoothing degree 200 years. Magnetic turning points labelled A-C

followed by a westerly swing with the bottom of the peak dated to ca 11 300 $^{14}C$ years BP, and then gradually turns to an easterly position around 10 200 BP. In the inclination a number of minor swings have been identified on a general trend of steeper inclination upcore (Fig. 3). The bottom of the Torreberga inclination record possibly could be correlated to the low inclination feature ε in the core from Lac de Bouchet in France (Bonifay et al. 1984), and ε in the Lac du Joux in Switzerland, cores 3 and 4 (Creer et al. 1980), while the top could correspond to ε in the same cores. Both with respect to the absolute inclination value and age this correlation seems reasonable. Comparisons with the declination records from the same are less clear. Profiles from Lake Halsjon and Lake Paddegölen (Bjorck and Sandgren 1986) in Blekinge (Fig. 2) in southeastern Sweden (Fig. 1), reflect the same pattern as seen in the Torreberga cores. The Late Weichselian sediment of these profiles are only about a meter thick, but compared to the Torreberga cores the chronology is very detailed, based on a number of radiocarbon dated levels.

In the Middle Swedish end moraine zone two varved clay sites, Tottnäs and Rystad (Fig. 1), have been sampled and analysed (Sandgren et al., in press) The chronology of the Tottnäs site is in varve years BP, (BP = 1984) according to the Swedish Geochronological Time Scale (Brunnberg 1986). The chronology of the Rystad site is based on a floating varve chronology (Kristiansson 1986), which indirectly can be correlated to the radiocarbon chronology. The varved clay at both sites is very fine grained with varve thicknesses in the order of a couple of mm. The varves are horizontally deposited without disturbances. Two parallel profiles were sampled from Tottnäs spanning 357 and 381 varves

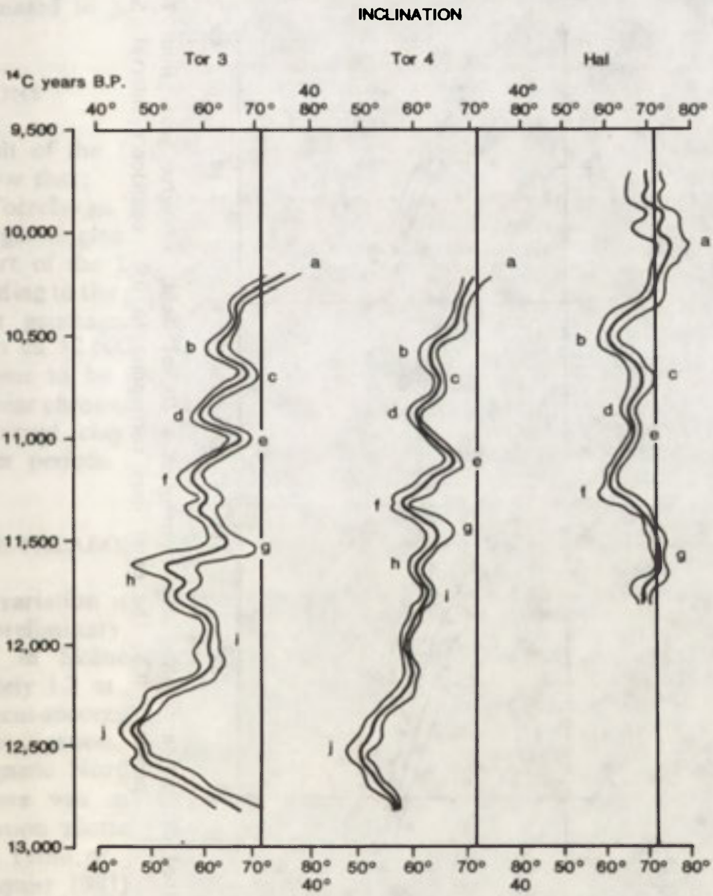


Fig. 3. Smoothed inclination records, after cleaning in a 10 mT field, vs age from Torreberga profile 3 and 4 (Tor 3, Tor 4) and Lake Halsjon (Hal). The outer lines correspond to 95% confidence interval. Smoothing degree 150 years. Magnetic turning points labelled a-j

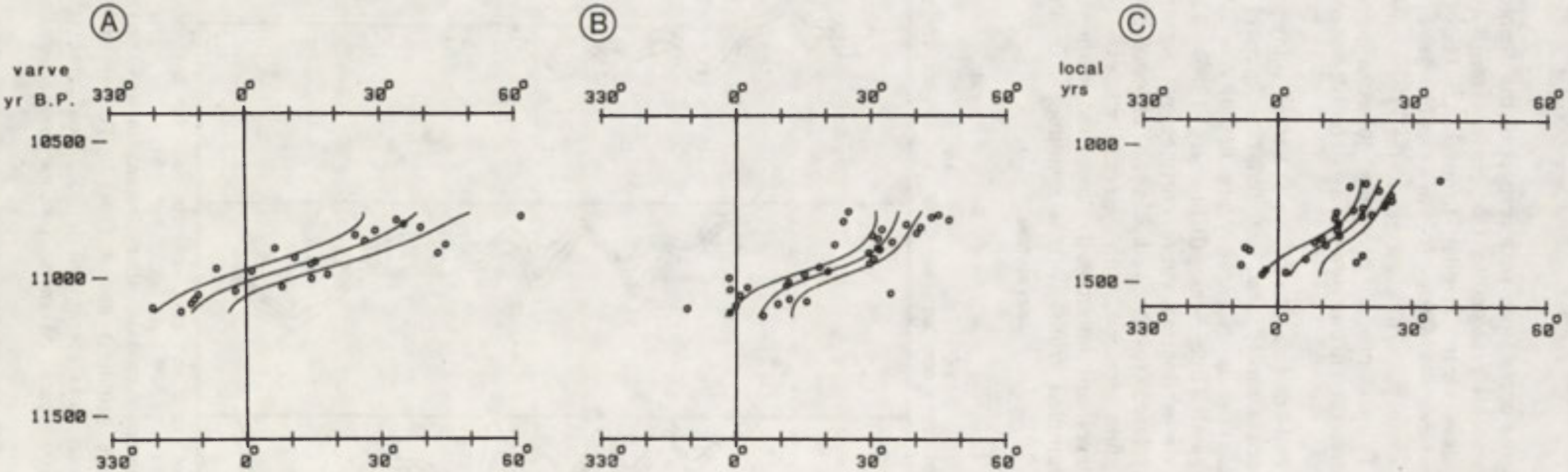


Fig. 4. Smoothed declination records, after cleaning in a 10 mT field, vs varve years from Tottnas profile 1 and 2 (A and B) and Rystad profile 1 (C). The outer lines correspond to 95% confidence interval. Smoothing degree 200 years

(= years) respectively, and from Rystad two partly parallel profiles were recovered, of which the longest spans 347 varves (= c years). At both sites both the declination and the inclination records could be repeated in the replicate core. At both sites there is a gradual easterly declination upcore (Fig. 4). The inclination of the Rystad profiles appeared to be lower (15–20°) than expected on the basis of a general axial dipole field and with respect to the site latitude. This can be explained by the fact that the sediment is situated above the ground water level. At the Tottnas site, where the sediment were recovered in a bay of the Baltic Sea, the inclination values seemed to be correct.

A statistical comparison of the declination records (Figs 2 and 4) from Lake Halsjon, Tottnas and Rystad are presented in Björck et al. (1987) and the difference between the radiocarbon chronology and the absolute chronology between 11–10 000 $^{14}$C years BP is discussed. The conclusion is that radiocarbon dating, in relation to the Swedish varve chronology, give too young ages in the order of 550–650 year at 10 300 $^{14}$C years BP but at 11 000 $^{14}$C yeras BP this difference has decreased to 400–500 years. These relationships do not take into account the differance between the varve chronology and calendar years, which can be estimated to possibly as much as 200 years (Cato 1987).

## CONCLUSIONS

The result of the Late Weichselian magnetostratigraphic investigations in south Sweden show that:

1) the Torreberga profiles, supported by the profiles from Lake Halsjön and Lake Paddegöen, give reliable and detailed records of the PSV variations throughout a major part of the Late Weichselian;

2) according to the palaeomagnetic results from Torreberga, there are no indications of aberrant geomagnetic field directions during the Late Weichselian, at least not back to ca 12 600 $^{14}$C years BP.

3) it seems to be possible to correlate the radiocarbon, chronology and varve (calendar) year chronology based on magnetostratigraphy, but additional investigations on long varved clay sequences are necessary to confirm these results and to cover longer periods of the Late Weichselian.

## HOLOCENE PALAEOSECULAR VARIATION

Secular variation studies of Holocene sediment have recently started and hitherto only the preliminary results from one locality exist. The locality called Bussjö is situated in Skåne, the southernmost province of Sweden (Fig. 1). Sixteen approximately 1.2 m long Livinstone piston cores were recovered from the 19.5 m long sediment-succession, in the late winter 1986, for a preliminary magnetostratigraphic investigation. The azimuth of each individual core was determined relative to the magnetic North Pole in the field. In the laboratory the long core magnetism of each core was measured on a Digico long core fluxgate magnetometer and the declination plotted against depth. The sediment at the Bussjö site is studied within the Ystad project with respect to a number of other paramaters (Berglund and Stjernquist 1981). Awaiting a complete pollen analytical investigation, (which is under way), four pollen analyses have been carried out in the lower part of the sequence at 1 metre interval between 15.7 and 18.7 metres below the water surface The three lower-most samples reflect typical Atlantic pollen-assemblages (J. Regnell, pers. comm.). The uppermost of these three falls within the Late

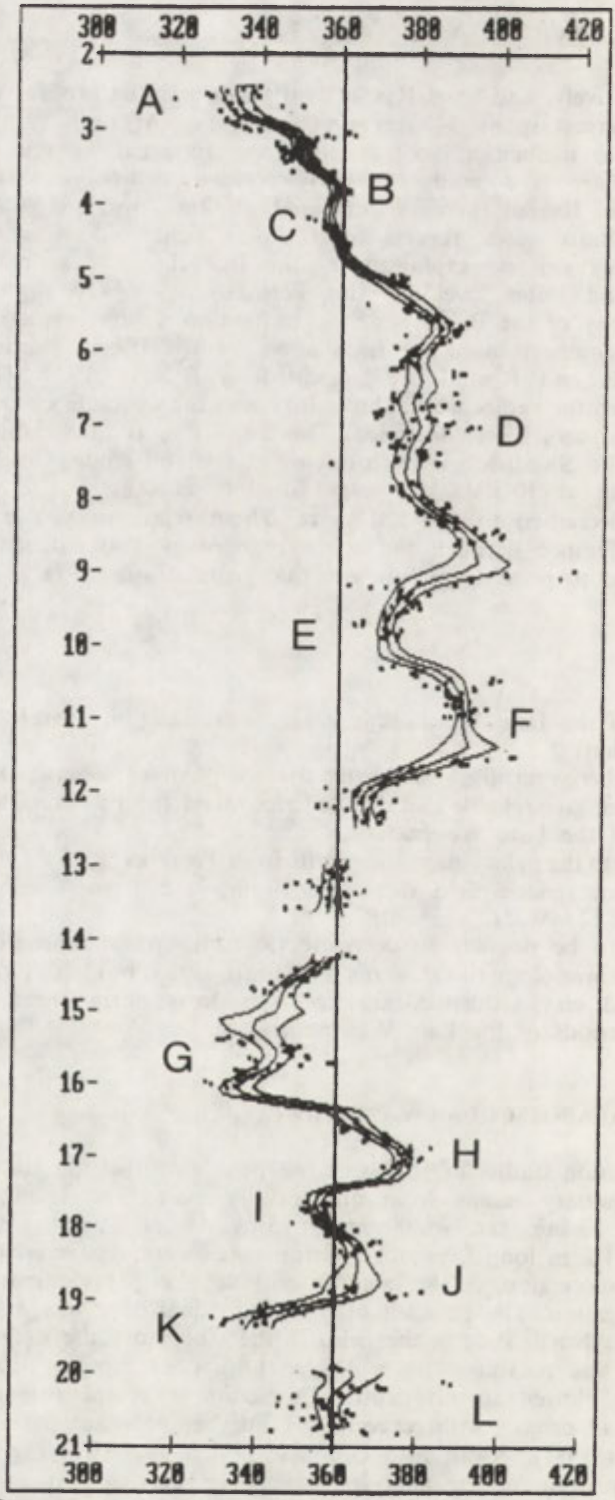


Fig. 5. Composite declination log for the Bussjo cores, with a smoothed average (central line) and 95% confidence limits (outer lines) vs depth. Smoothing degree 0.5 m. Prominent magnetic features labelled A-L

Atlantic pollen zone and the other two in the Early Atlantic pollen zone (Nilsson 1961, 1964). The uppermost of the four samples falls within the Early Subboreal pollenzone (Nilsson 1961, 1964). The Holocene/Late Weichselian transition, represented by a sedimentological change from gyttja to non-varved clay takes place below 18.70 metre but above 20.9 metre below the water level. A layer below the non-varved clay, very close to the bottom, composed of coarse detritus gyttja resting on sand and gravel is, until further analyses have been performed, considered to have been deposited during the Allerød Chronozone (Mangerud et al. 1974). The information from the pollen analysis combined with the litostratigraphic information thus gives some idea of the time span of the sediment succession. It probably represents the entire Holocene and a major part of the Late Weichselian. The results of the long core measurements are shown against depth in Fig. 5 as a composite log of the individual cores. As can be seen, the declination swings around zero with an approximately greatest amplitude of $\pm 30°$. Declination turning points are labelled A–L, with A being the youngest. When the results are compared to to the East and West European (EE and WE) magnetic master curves (Thompson 1983) the general trend of the Bussjo declination log is in agreement with these curves. In closer detail there are, however, also differences and the Bussjo log seems to invoke features from both master curves, although the EE curve appears to be the closest. The double nature of the easterly peak D seen in the EE curve is reflected in the Bussjö sediment while the following swing E is more westerly in the master curves. The gradual change from the easterly peak F to G in the Bussjö core again, is very similar to the EE curve, whereas the shape of the easterly peak H is more like the WE curve. Based on the declination turning points a correlation with either of the two mastercurves seem to be reasonable.

In Fig. 6 the depth of the Bussjo declination turning points are plotted against age, where ages are obtained from the two master curves according to Thompsom (1983). The ages of the magnetic features are tree ring calibrated years BP (according to Thompson 1983), and $^{14}C$ years BP are according to Björck et al. (in prep.) The pollen zones (in $^{14}C$ years BP) back to 10 000 BP are according to the definition of Nilsson (1961, 1964). Figure 6 also shows the levels of the pollen-analysis and the pollen zones to which they have been determined. The age/depth points are connected with lines, which gives the sediment accumulation rate based on the two master curve time scales. If the results of the pollen analyses are projected onto the depth/age curve based on the west European chronology, two of the four pollen spectra fall within the wrong pollen zone. If, on the other hand, the east European chronology is used all four samples fall within the expected pollen zone. This suggests that the EE master curve could be the most relevant to use in south Sweden as an indirect chronological tool in dating sediment cores.

## CONCLUSIONS

The results of these preliminary Holocene magnetostratigraphic investigations show:

1) that the Bussjö sediment seem to be suitable for further magnetostratigraphic investigations,

2) the possibility of using geomagnetic master curves as a chronological tool in dating sediment successions that otherwise cannot be absolutely dated or where the chronological information, as in this case, is very limited,

3) the problems of using master curves over greater distances and the importance of establishing master curves from different regions.

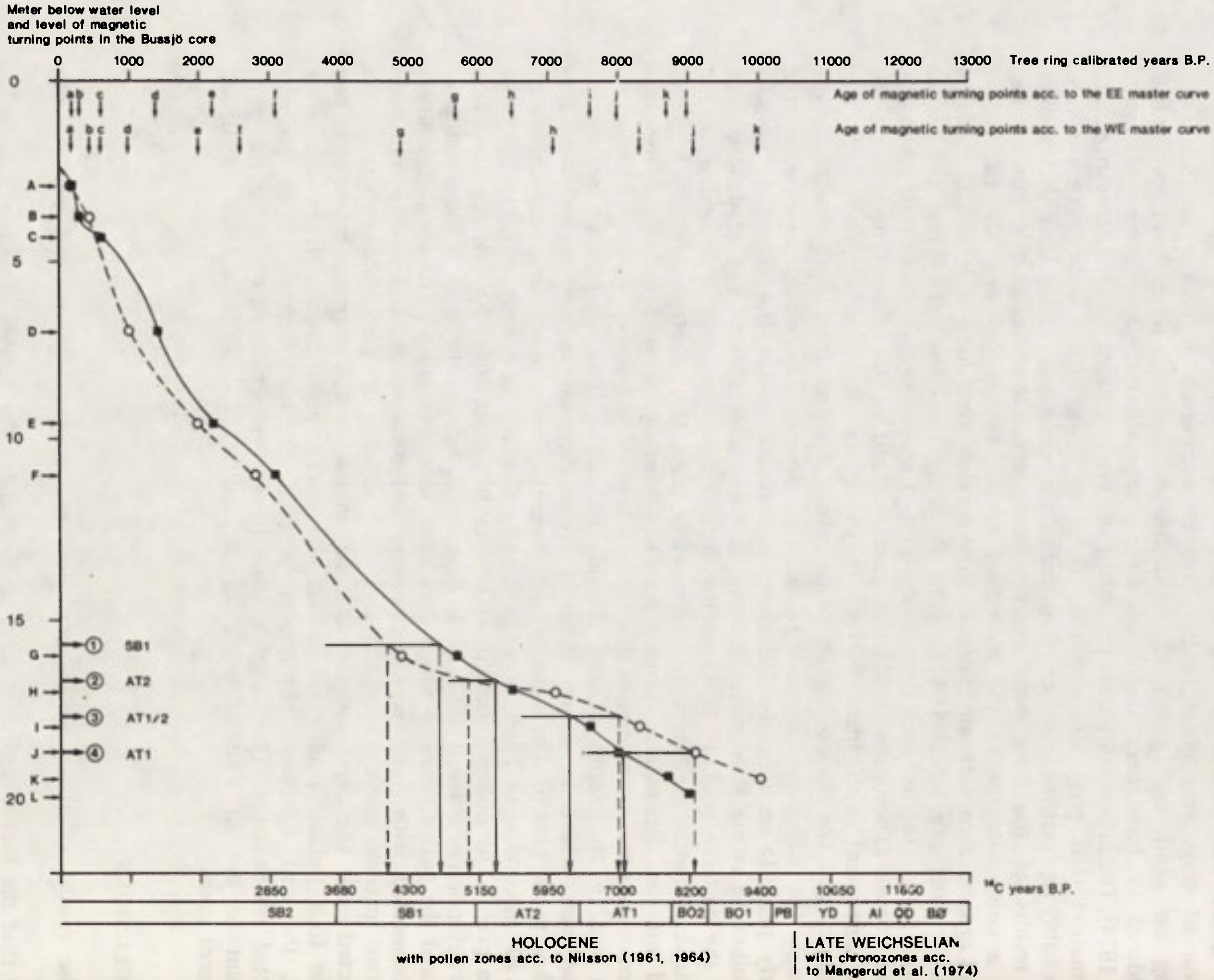


Fig. 6. Diagram showing the sedimentation rate of the Bussjo sediment based on correlation of the magnetic turning points with the WE declination master curve (hatched line) and the EE declination master curve (solid line). Futher comments see text

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GEOGRAPHIA POLONICA 55, 1988

# MAXIMUM EXTENT OF THE VISTULIAN ICE SHEET IN THE VICINITY OF KONIN, POLAND: A GEOMORPHOLOGICAL, SEDIMENTOLOGICAL AND RADIOMETRIC EVIDENCE

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ABSTRACT. The results of geomorphological mapping of the vicinities of Konin have been verified by sedimentological, stratigraphical, radiometric and geochemical studies. The age of the maximum advance of the Vistulian ice sheet has been estimated as about 20 000 yr BP. The $^{14}C$ and TL dating is considered not only in terms of palaeogeography but also as a methodological approach. Geochemical analyses of tills differing in age, which occur in the vicinity of Konin, serve as a tool, by means of which the deposits can be described and allow an attempt to be made to use the resulting data for the purpose of relative dating of deposits accumulated directly by the ice sheet.

## INTRODUCTION

The vicinities of Konin are contiguous to the Lateglacial relief produced by the Vistulian Glaciation (Würm) and to the periglacial relief due to the Middle-Polish Glaciation (Riss) (Fig. 1). The dividing line runs in a northeast-southwest direction. A parallel-oriented extensive valley, i.e. the Konin section of the Warsaw-Berlin Pradolina, is a conspicuous feature of the geomorphologic setting. The east side of the pradolina lies within relief features produced by the Middle-Polish Glaciation, whereas its west side adjoins areas of the Vistulian accumulation of glacial and fluvioglacial origin to the north. There is a zone of marked indistinctness of the Lateglacial relief to the south. Merely fragments of thin till patches due to the Vistulian Glaciation are encountered in vast areas of terraces in the so called Pyzdry Basin. A general outline of geomorphology of the study area has been documented by the studies of Krygowski (1952, 1961), Rotnicki (1963), Konieczny (1965), Bartkowski (1967), Rutkowski (1967) and Kłysz (1981). The earlier studies, as well as investigations which have been caried out by the present authors for many years have not brought about fully satisfactory determination of the maximum extent of the Vistulian Glaciation. Neither a geomorphological solution nor a sedimentological one has been provided to the problem of interrelationships between the Leszno Phase (maximum) and the somewhat later Poznań Phase.

A zone of classical marginal forms due to the maximum advance is absent to the west of Konin. On the other hand, continuous end-moraine ridges belonging to the Poznań Phase remain conspicuous (Rotnicki 1963; cf. Fig. 1).

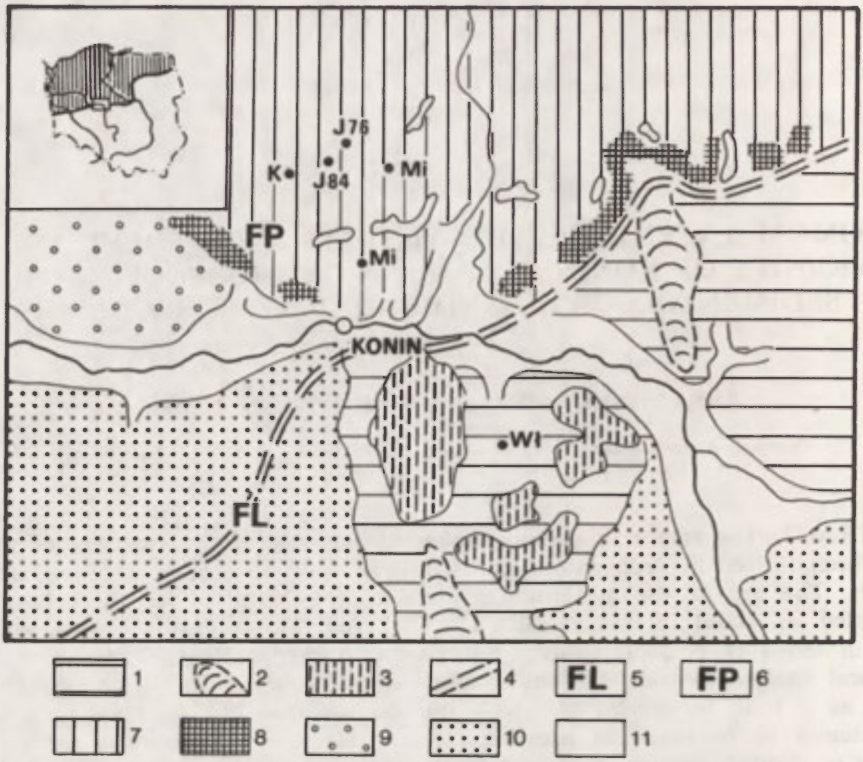



Fig. 1. Simplified geomorphologic setting: 1 – morainic plateaus due to the Middle-Polish Glaciation, 2 – old Pleistocene residual, 3 – table-land and kame hills, 4 – maximum extent of the Vistulian Glaciation, 5 – Leszno Phase, 6 – Poznań Phase, 7 – morainic plateaus due to the Vistulian Glaciation, 8 – morainic hillocks, 9 – multi-level outwash plains, including outwash water erosional surfaces, 10 – river terraces, 11 – floors of valleys, pradolinas and lake basins

A wide marginal zone varying in morphology extends to the east of Konin. It is due to process which controlled the modification of the ice-sheet margin during two phases of the Vistulian Glaciation i.e. the Leszno and Poznań phases (cf. Fig. 1).

## CHARACTERISTICS OF NEOPLEISTOCENE SERIES

The palaeosurface of deposits belonging to the penultimate glaciation (the Middle-Polish Glaciation = Riss) is of fundamental importance for the sediments under investigation. Deposits dating back to the last Pleistocene warm unit, i.e. the Eemian, were preserved in depressions over that surface. The present authors' studies begun in 1975 led to the detection of a number of sites containing Eemian deposits, including those containing fossil organic deposits laid down over that period. The series detected provide confirmation of inferences based on the geomorphological criteria as to the extent of the Vistulian ice sheet in the vicinity of Konin. One of the newly discovered sites that is Maliniec provides information on the age of the maximum advance of this ice sheet.

The Eemian organic series identified successively are derived from the following

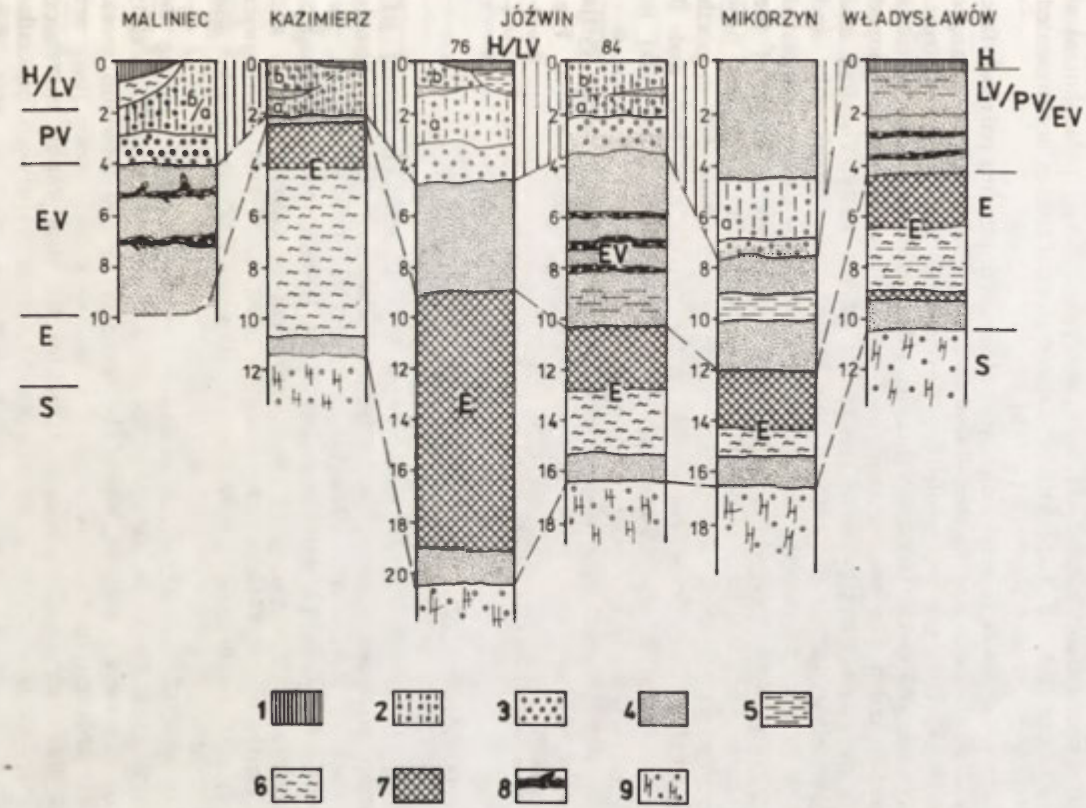



Fig. 2. Selected Neopleistocene profiles: 1 – peats and surface organic series, 2 – tills due to the Vistulian Glaciation, 3 – fluvioglacial deposits, 4 – fine and medium-grained sands, 5 – silts, 6 – gyttjas, 7 – moss earth peats, 8 – thin fossil organic series, 9 – tills due to the Middle-Polish Glaciation. H – Holocene, LV – Late Vistulian, PV – Pleni-Vistulian, EV – Early Vistulian, E – Eemian, S – Middle-Polish Glaciation

sites: Jóźwin 76 (Stankowska and Stankowski 1976), Kazimierz (Stankowski and Tobolski 1982), Władysławów (Kłysz and Stankowski 1986), Jóźwin 84 (Stankowski in press), Mikorzyn (Stankowska and Stankowski in preparation). Recently, another profile was produced for Maliniec (cf. Fig. 3 – Maliniec 86A and B).

Palaeobotanical and palaeoecological documentary evidence provided by Tobolski is available for all the sites (cf. Figs 1 and 2). In some cases it consists of complete pollen spectra and macrofossil ones; otherwise, merely palaeobotanical expertises constitute the evidence. The results of palaeobotanical studies are presented in a separate article (cf. Tobolski pp. 181–186).

The Eemian organic deposits, mineral sediments and the corresponding palaeosurface are overlain by deposits belonging to the Würm cold unit. Their thickness ranges from 2–3 to 8–10 m. They are represented by fluvial sands and silts, lacustrine, ice-dammed-lake and pool sands, fluvioglacial sands, tills distributed in two horizons separated indistincly and by, in general, thin fossil organic intercalations developed to a varying degree. The latter preserve the record of the Early Vistulian and Pleni-Vistulian warming up. In the case of the Maliniec site (Stankowska and Stankowski 1979, Pazdur, Stankowski and Tobolski 1981), they can be associated with weak positive climatic fluctuations prior to the maximum advance of the Vistulian ice sheet. The Maliniec site that was discovered in 1978 has now received confirmation in the form of a similar profile located about 500 m north of the starting profile. The newly discovered profile of Maliniec 86A and B is the subject matter of a new compilation (Stankowska and Stankowski in preparation).

The Holocene deposits occurring in this area are represented by peat, gyttja and silt series of sporadic occurrence and a continuous soil cover (cf. Fig. 3 – Maliniec 1986 B).

## RESULTS OF RADIOMETRIC DATING

Three of the sites discovered were subjected to $^{14}C$ and TL dating (cf. Fig. 3). The dates given were arrived at in different Polish laboratories, namely (1) the Laboratory of Radiometric Chronology of the Institute of Physics, the Silesian Polytechnic ($^{14}C$ and TL dates provided by M. F. Pazdur and A. Bluszcz, respectively), (2) the TL Laboratory of the Department of Geomorphology and Geology, the University of Gdansk (the dating carried out by S. Fedorowicz and J. Olszak), (3) radiometric laboratories in Warsaw (M. Prószyński and H. Prószyńska-Bordas).

There are many $^{14}C$ and TL dates available for the Maliniec, Jóźwin 84 and Władysławów sites. From the viewpoint of the Late Quaternary stratigraphy all the dates are generally consistent within the range of 80–100 thousand years. However, their evaluation in terms of physics is not so explicit since there are rather essential distinctions and inversions. As far as the stratigraphic expression of the dates available is concerned, the existing discrepancies between particular laboratories and inversions of dates within a given dating series are acceptable.

The dating of samples originating from tills to older ages than those of the underlying fluvioglacial deposits may be accounted for by the incomplete removal of thermoluminescence defects during glacial transport. Rather insignificant inversions of dates in the deposits of the aquatic sedimentary environment or even apparently wide discrepancies between dates available for complementary samples investigated in different laboratories are, as a rule, included within the 10–20% error range.

If account is taken of all the dates available for the Maliniec and Jóźwin

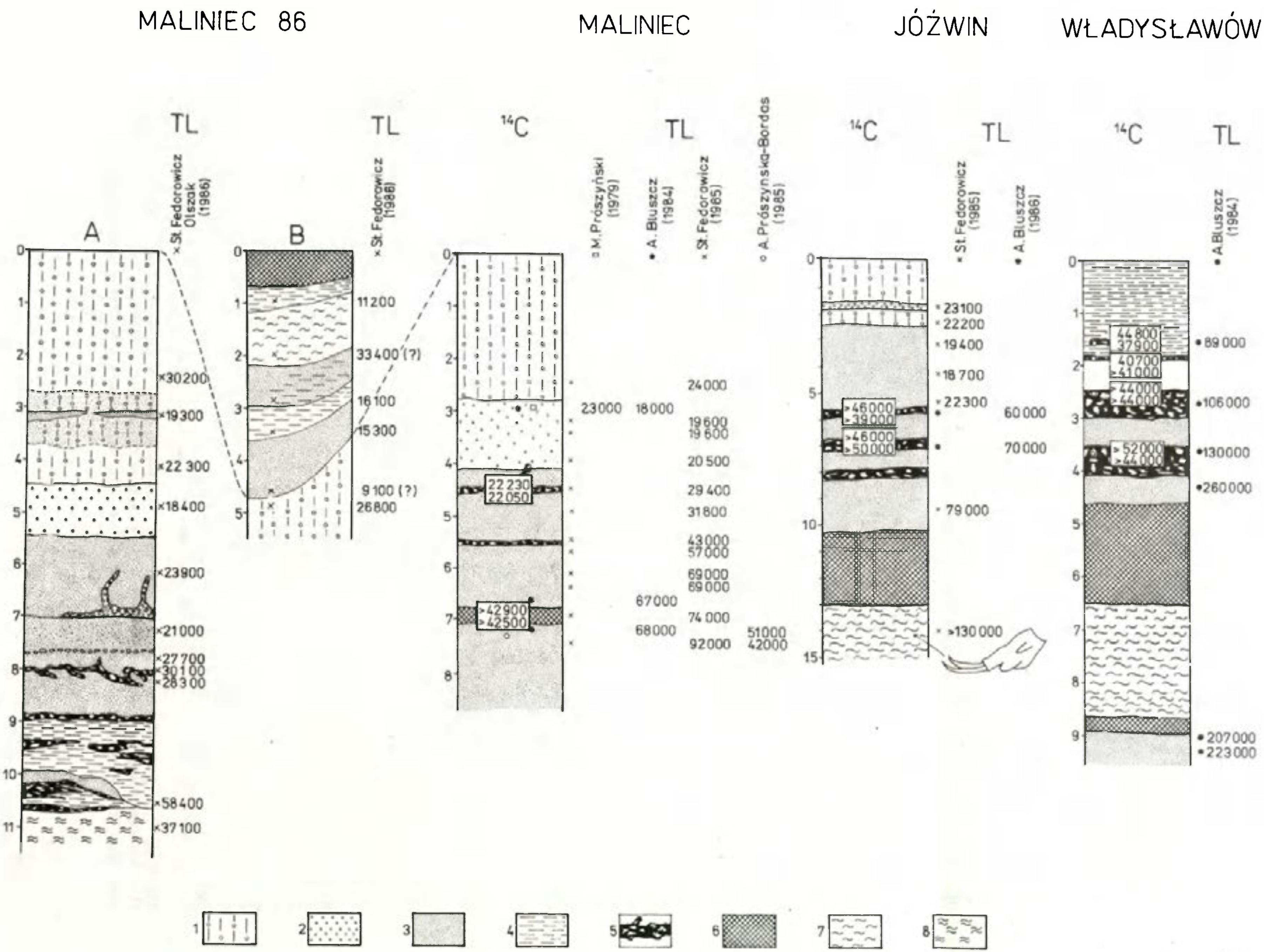



Fig. 3. Radiometric dates: 1 – tills due to the Vistulian Glaciation, 2 – fluvioglacial deposits, 3 – medium-grained and fine sands, 4 – silts, 5 – fossil organic series, 6 – moss and earth peats, 7 – gyttjas, 8 – clays

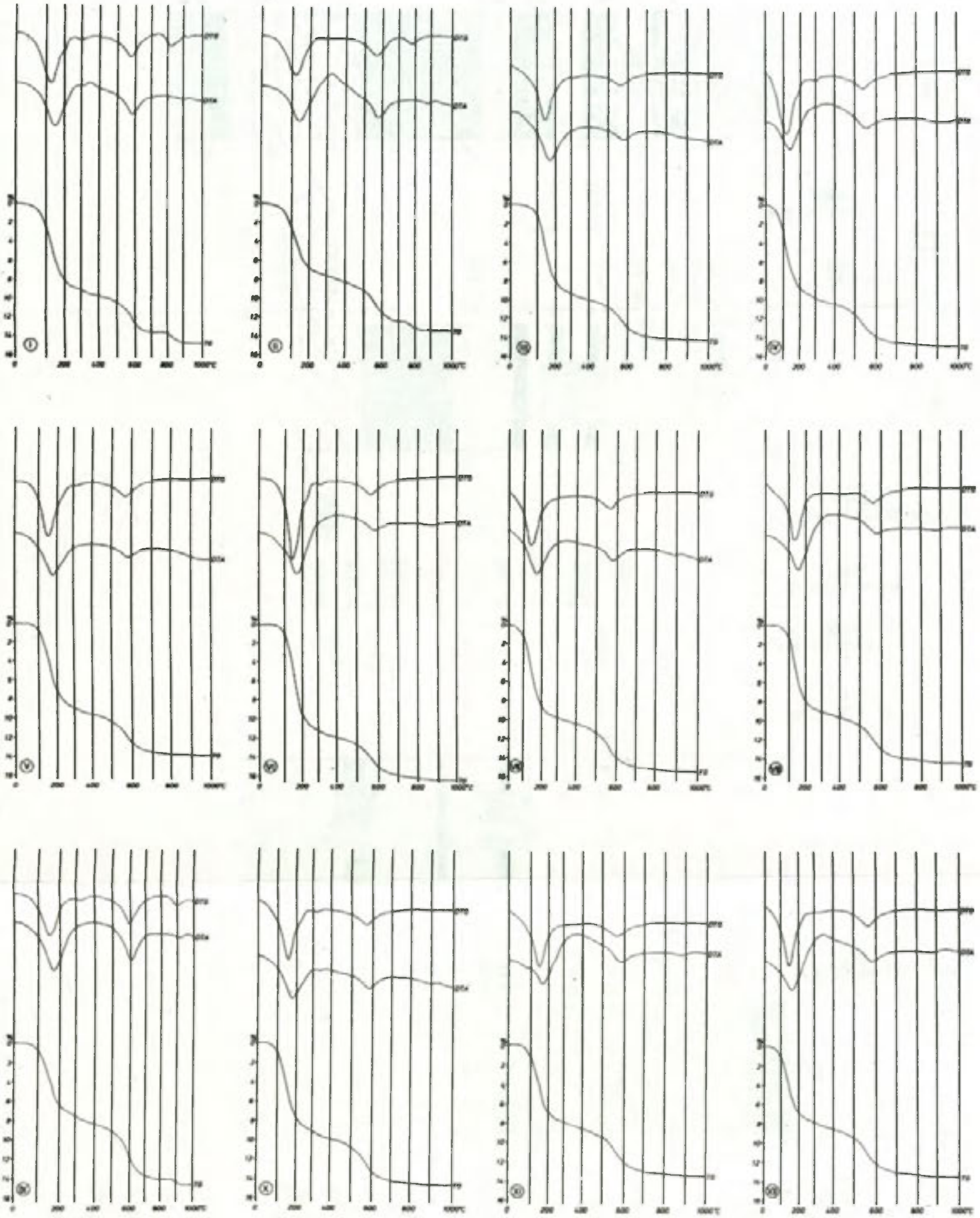


Fig. 4. Derivatograms of tills, fraction ø< 2 m: I – Jóźwin–Würm, II – Jóźwin–Riss, III – Kleczew-Würm, IV – Maliniec–Würm, V – Maliniec–Würm, VI – Kazimierz–Würm, VII – Kazimierz-Würm, VIII – Kazimierz–Riss, IX – Konin S–Riss, X – Władysławów–Riss, XI – Władysławów-Riss, XII – Władysławów–Mindel

84 sites, it should be assumed that the maximum advance of the Vistulian ice sheet took place in the vicinity of Konin about 21 000–20 000 yr BP at the earliest. Moreover, the lithofacies analysis of profiles for Maliniec, Jóźwin 84 and Władysławów, as well as of numerous sites in the vicinity of Konin, which do not contain fossil organic series and have as yet not been dated radiometrically, rules out the presence or even close neighbourhood of any older ice sheet during the whole last cold period (the Würm).

Mention should also be made of the fact that some the TL dates for sands and silts which cover and lie beneath the Eemian organic series are excessively early. Serious inversions of dates are also recorded there. It may be thought that the resulting data from measurements are due to the objects under investigation, i.e. deposits, the TL age of which is beyond the linear extent of variations of this parameter.

In general, the radiometric dating provides support of the results of palaeogeographical study concerning the extent of the maximum advance of the Vistulian ice sheet. Simultaneously, the dating defines the age of this advance, it is of Pleni-Vistulian ice advance.

## MINERALOGICAL AND CHEMICAL PROPERTIES OF TILLS OF DIFFERENT AGE

Laboratory investigations of tills from the vicinity of Konin were not only concerned with the above sites of fossil organic series but also with many other sites. Among the characteristics under investigation, carbonates, sorption complex, amorphous substance and clay minerals are considered in this article. *Carbonates* were gasometrically identified in the fraction smaller than 1.02 mm by means of the Scheibler method. The tills occurring in the vicinity of Konin possess relatively low carbonate contents which average several percent. The extreme values for particular samples are 0.4 to 15.6%. There are certain variations of carbonate contents between the Vistulian tills and the older ones. Samples from the Vistulian tills generally contain less than 9% of carbonates, with the extreme values being 0.4–11.6%. On the other hand, earlier deposits usually contain over 9% of carbonates, with the extreme values ranging from 8.1 to 15.6%. *The sorption complex* was identified by removing exchangeable cations with ammonium chloride (Szczepańska, Walna 1983, Stankowska, Walna in press). Calcium ion is the main ion included in the sorption complex of the tills under investigation. It occurs as 19.0 to 21.0 mval/100 g, which accounts for 88.8 to 93.1% of the whole sorption complex. The percentage of the remaining cations, i.e. magnesium, sodium and potassium, is markedly lower. Magnesium sodium and potassium ions assume values of 0.5–1.3 mval/100 g (2.1–5.5% of the whole complex), 0.5–0.8 mval/100 g (2.5–4.4%) and 0.4–0.9 mval/100 g (1.9–3.9%), respectively. *The amorphous substances* were driven off by dissolving them in 5% sodium carbonate through heating. Silica was identified in the extract by the use of the weighing method by vaporizing it twice with hydrochloric acid. After the isolation of silica in the filter, the $R_2O_3$ oxide sum was also identified by the weighing method through precipitating hydroxides with ammonium hydroxide.

Silica assumes values of 9.1 to 28.7% whereas the $R_2O_3$ oxide sum ranges from 1.8 to 10.1%. The parameters identified vary within one age series and between series belonging to the same cool unit, as well as between tills originating from different units. However, no marked trend in these variations is observable. *Clay minerals* were identified by a few analyses, i.e. sorption capacity determination in relation to cations, thermal analysis and X-ray analysis. The study was made

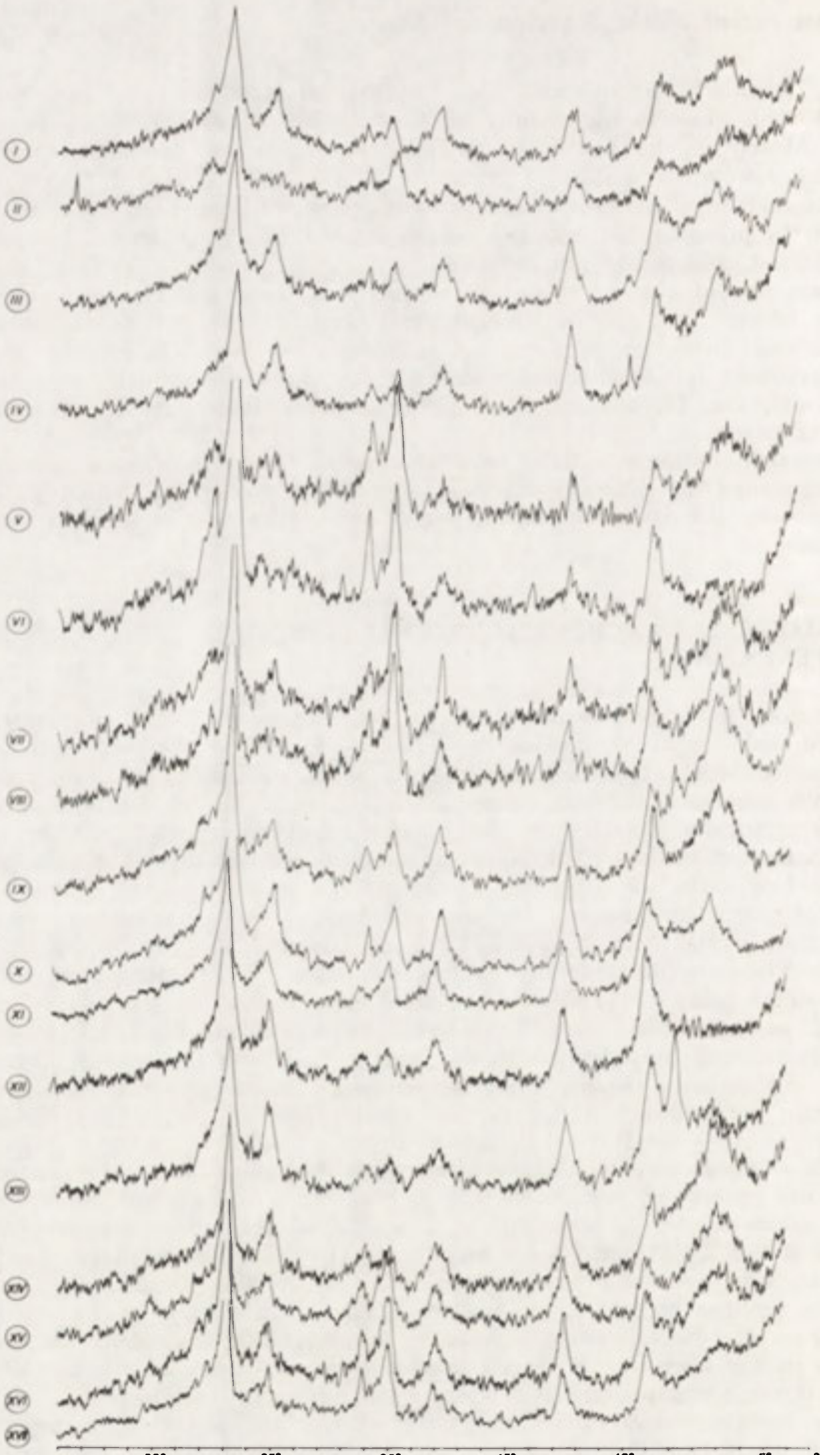



Fig. 5. Diffractograms of tills, fraction ø < 2 m: I – Jóźwin–Würm, II – Jóźwin–Würm, III – Jóźwin–Riss, IV – Jóźwin–Riss, V – Kleczew–Würm, VI – Kleczew–Würm, VII – Maliniec–Würm, VIII – Maliniec–Würm, IX – Kazimierz–Würm, X – Kazimierz–Würm, XI – Kazimierz–Riss, XII – Kazimierz–Riss, XIII – Konin S–Riss, XIV – Kuny–Riss, XV – Władysławów–Riss, XVI – Władysławów–Riss, XVII – Władysławów–Mindel

of the < 2 μm fraction that comprises clay minerals. The X-ray analysis was carried out in crude powder samples, glycerine-saturated ones, those heated previously at the temperature of 550°C and those subjected to hydrochloric acid treatment. The ratio of integral intensities of the 001 basic lines was subjected to analysis. Account was also taken of the intensities of the basic lines of pure minerals.

The complex mineral composition of the samples analysed did not allow determination of the quantitative composition of clay minerals. On the other hand, it was possible to determine the qualitative composition and to establish which minerals remained dominant in a given deposit. The order of occurrence of the remaining clay minerals was established on the basis of their quantity. Figures 4 and 5 show selected examples of derivatograms and diffractograms.

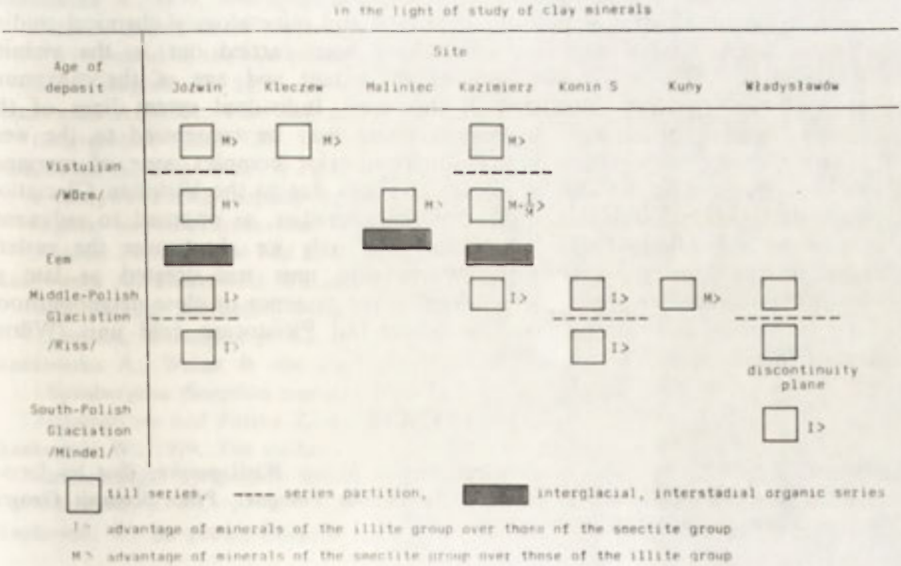
There are no differences on the qualitative composition of clay minerals between tills occurring in the vicinity of Konin and accompanying deposits, i.e. sand sediments of aquatic origin, which were studied for comparative purposes. Yet, differences were found in relative quantities of the mineral components. It mostly holds for minerals of the smectite and illite groups.

Clay minerals which occur in the deposits under investigation comprise chiefly minerals of the illite and smectite groups present in similar amounts. They are accompanied by minerals of the mixed illite/smectite type. In addition to these minerals, much smaller amounts of minerals of the chlorite and kaolinite groups are found.

The presence of minerals of the mixed illite/smectite type points to a genetic relationship between illite and smectite. A wide basic line of smectites is indicative of poor arrangement of their structure. An increased amount of minerals of the mixed illite/smectite type correlates with a diminishing crystallinity index of the basic line of illite. Thus, it appears that the presence of mixed illite/smectite type structures is due to weathering processes which convert minerals of the illite group into those of the smectite group.

A matter of interest is that minerals representing mixed structures exhibit differing characteristics on diffractograms. A conspicuous maximum of about 10 Å is shown by some samples. X-ray-gram characteristics for other samples point to a closer affinity of the mixed-structure minerals for those of the smectite group than for

Table 1. Stratigraphic variability of tills occurring in the vicinity of Konin in the light of study of clay minerals



those of the illite one. X-ray-grams of other samples indicate a closer affinity for minerals of the illite group. It may be presumed that this diversity is the result of intensity and duration of weathering processes affecting the input material.

The results of identification of clay minerals allow certain inferences to be made as to the stratigraphical variability of tills occurring in the vicinity of Konin (Table 1). It has been established that the main clay minerals included in the tills under investigation comprise minerals of the smectite and illite groups. Smectites prevail in all the Vistulian tills under investigation, irrespective of lithological and stratigraphical distinctions (phase separation). An exception is provided by a sediment lense composed of poorly modified Mesozoic material from the profile for Kleczew (Stankowska and Stankowski 1979). Certain variations in the results of sorption capacity, thermal and X-ray analyses are observable within a particular series, as well as between the Vistulian till series. However, they are not so great as to become reflected in the composition of clay minerals.

Minerals of the illite group gain an advantage over those of the smectite group in tills produced by the Middle-Polish Glaciation. The only exception is represented by tills occurring at the Kuny site lying somewhat north of the Władysławów site, where smectites dominate over illites.

Note should be made of the fact that the study of clay minerals occurring in tills of the Konin vicinities which are earlier than the Middle-Polish Glaciation (the Władysławów site) reveals that illites assume dominance over smectites.

The results of study of tills occurring in the vicinity of Konin reveal their distinctiveness in respect of the general stratigraphic trend in the differentiation of the clay mineral components present in tills of different ages in Poland (Stankowska 1970, 1979, 1981). In general, the amount of smectites increases with the age of tills. The reverse trend, i.e. an increase in the amount of smectites in younger tills, compared with the earlier ones, is observable in the vicinities of Konin. This distinctiveness may be conditioned by regional characteristics but nevertheless it allows mineralogical distinctions between till series of different ages to be documented.

## CONCLUSIONS

Geomorphological, stratigraphical, radiometric and mineralogical-chemical studies of Neopleistocene glacial deposits, which have been carried out in the vicinity of Konin since 1975, provide evidence of the extent and age of the maximum advance of the Vistulian Glaciation in this area. Individual extent lines of the maximum (Leszno) phase and the Poznań phase may be recognized to the west of Konin. During both phases a well-developed wide complex zone of marginal forms was produced to the east of Konin. Deposits due to the Vistulian Glaciation display lithological-mineralogical distinctive characteristics, as opposed to sediments produced by the Middle-Polish Glaciation. The only ice sheet over the eastern Greater Poland area throughout the Würm cold unit was created as late as about 20 000 yr. BP. No geological evidence on the presence or close neighbourhood of any earlier ice sheet throughout the whole last Pleistocene cold unit (Würm) has been found.

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# PALEOGEOGRAPHY OF THE PERIGLACIAL ZONE IN POLAND DURING THE MAXIMUM ADVANCE OF THE VISTULIAN ICE SHEET

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ABSTRACT. The paleogeographic reconstruction of the periglacial zone during the maximum extent of the last ice sheet is based on various studies of cryogenic phenomena, loess, slope and fluvial forms and deposits as well as paleobotanical and paleozoological records. These data show a very high climatic gradient in N–S transect as well the increasing continentality towards east and great diversity of geoecosystems.

## INTRODUCTION

The concept of the periglacial zone on the Polish territory during the last cold stage has a long history starting with Łoziński's (1909) classic paper on the block fields in the Holy Cross Mts. The definitions of the processes characteristic of various parts of that zone from the arctic desert to the subartic tundra and forest-tundra with distinct regions of the dominance of loess deposition or solifluction processes were presented in the fundamental papers by Dylik (1953, 1956) and Jahn (1956), as well as Klimaszewski (1948) and Malicki (1950). A detailed recognition of the deposits and forms is found in the papers published by Dylik (1967, 1969), Goździk (1973), Jersak (1973), Maruszczak (1968, 1980), Mojski (1968), Rotnicki (1966), and Starkel (1969). In the last 15 years the number of chronologically fixed data is increasing due to the radiocarbon datings of organic layers and TL datings of loess and fluvial deposits (Fig. 1).

In this paper special attention will be payed to the phase of the maximum extent of the last ice sheet, which in the Polish sector occurred 20–19ka BP because the glacial till near Konin covers the fluvial sands with a dryas flora dated $22150 \pm 330$ BP (Pazdur et al. 1979). In other sectors the maximum extent followed not later than 17.5ka BP (dating near Witebsk 17 700 BP – cf. Chebotarieva ane Makaricheva 1974). This phase will be discussed on the phone of the whole upper pleniglacial including the period 29–13ka BP, after the Denekamp warming, which is slightly older than the Komorniki loess soil in Southern Poland (Jersak 1973) or the East-European warmings Bryansk phase (Velichko et al. 1984) or the Dunayevo phase (Chebotarieva and Makaricheva 1982) dated 29–25ka BP, and before the Bølling phase. This time interval causes up till now many questions, among them the increase of climatic continentality, the possibility of survival of various plant communities north of the Carpathians, and the character of the reflection of secondary climatic fluctuations.


Fig. 1. Localitiew of the upper Vistulian deposits mentioned in the text: 1 — Konin (Pazdur et al. 1979), 2 — Ostrzeszów Hills (Rotnicki 1966), 3 — Bełchatów (Baraniecka 1980), 4 — Łódź region (Dylik 1969, Goździk 1973 and others), 5 — Kamion (Manikowska 1985), 6 — Radzymin terrace (Nowak 1974), 7 — Łomżyczka (Straszewska 'and Goździk 1978), 8 — Wąchock (Lindner 1985), 9 — Holy Cross Mts. (Kłatka 1962), 10 — Raj Cave (Kowalski 1974, Madeyska 1977), 11 — Łążek (Bielecka 1960, Mamakowa 1968), 12 — Nieledew (Maruszczak 1980, 1985), 13 — Odonów (Jersak 1973, 1985), 14 — Opatowiec (Jersak 1976), 15 — Kraków–Zwierzyniec-two sites (Konecka-Betley and Madeyska 1985, Kozłowski et al. 1974), 16 — Nowa Huta (Mamakowa and Środoń 1977), 17 — Kaniów (Gilot et al. 1982), 18 — Wadowice (Sobolewska et al. 1964), 19 — Dobra (Klimaszewski 1971), 20 — Lipowe (Starkel 1969), 21 — na Grelu (Klımaszewski 1961), 33 — Krościenko (Klimaszewski et al. 1950), 23 — Brzeźnica (Mamakowa and Starkel 1974), 24 — Podgrodzie (Mamakowa and Starkel 1977), 25 — Pilzno (Pazdur 1985), 26 — Błażowa (Krygowski 1960), 27 — Kępa and other hollows in the Jasło–Sanok Depression (Gerlach et al. 1972), 28 — Dziurdziów (Starkel 1965), 29 — Solına (Starkel 1965), 30 — Smerek (Ralska-Jasiewiczowa 1980)

## PERMAFROST, WEATHERING AND MASS MOVEMENTS

In the territory of Poland south of the maximum extent of the last Scandinavian ice sheet, as well as in the zone of older recessional phases, there are many traces of permafrost and frost weathering. Among them there are the pseudomorphoses of poligonal structures, the layers of frost desintegration, involutions and solifluction blankets as well as residual tors, cryoplanation terraces, cryopediments and block fields (Fig. 2). In the zone reaching about 100–150 km from the maximum extent of the ice sheet over the interpleniglacial layer there occurs the horizon with involutions and next ice wedge poligons with a diameter up to 20 meters, wind polished stony pavement, and higher up the next horizon of poligonal structures filled with sand and indicating the existence of an arctic desert (Dylik 1969, Goździk 1973). In Łomżynka (Straszewska and Goździk 1978) the lacustrine sediments with tundra pollen are dissected by contraction fissures and covered with stony pavement indicating a turn to aridity. In that zone near Łódź J. Dylik (1967) distinguished earlier on the slopes from below: the member

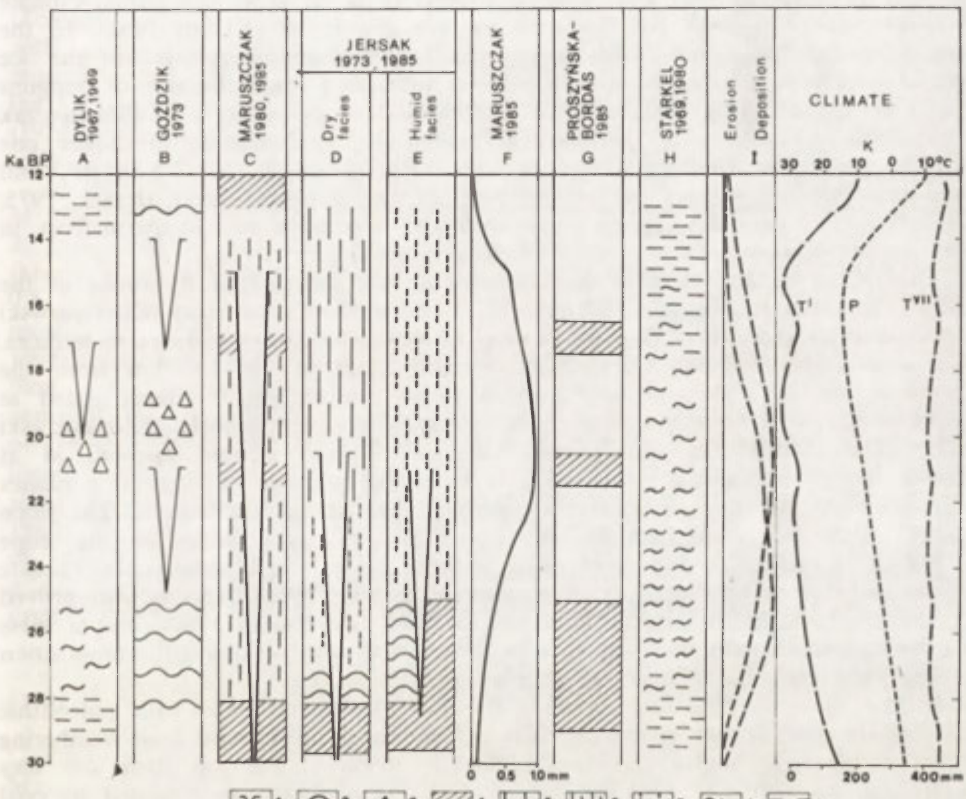


Fig. 2. Sequence of events in the periglacial zone during the upper Pleniglacial 30–13 ka BP (after different authors).

A – frost and slope processes in Łódź Plateau (Dylik 1967, 1969), B – permafrost phenomena (Goździk 1973), C – loess deposits (Maruszczak 1980, 1985), D – loess deposits in dry facies (Jersak 1973, 1985), E – loess deposits in "humid" facies (Jersak 1973, 1985), F – rate of loess deposition (Maruszczak 1985), G – dating of arctic soils in the loess profiles (Prószyńska-Bordas 1985), H – sequence of slope processes in the Carpathians (Starkel 1969, 1980), I – tendency to erosion or aggradation in the upper-Vistula basin (two variants), K – course of summar temperature ($t^{VII}$), winter temperature ($t^{I}$) and precipitation (P): at 50°N latitude

1 – frost wedges, 2 – cryoturbations, 3 – eolian stone pavement, 4 – arctic solis, 5 – loess deposits, 6 – „dry facies" of loess, 7 – "humid facies" of loess, 8 – solifluction processes, 9 – deluvial processes

of rhythmically stratified deposits, the solifluctional member, and the upper rhytmically stratified deluvial horizon. Kuydowicz-Turkowska (1975) described the fossil earth flows. According to Manikowska (1985) the congelifluction member is older and represents the period about 25ka BP. The slope processes on the permafrost formed in that arctic desert residual hills with extensive cryopediments (Dylik 1953). In the Ostrzeszów Hills, which belong to the marginal zone of the Warta glaciation on the sides of the interpleniglacial age valleys there was formed a lower level of cryopediments (Rotnicki 1966). The coexistence of wash, creep and eolian processes has been proved by the findings of mixed niveoeolian-niveofluvial deposits on the slopes of periglacial valleys (Klatkowa 1985).

In the Holy Cross Mts. at a distance of ca 150–200 km from tne ice heet the layer of weathered debris and solifluction formed in the relatively humid climate is overlain with a loess cover related to arid conditions (Klatka 1962). In the whole belt of the South-Polish loess uplands the pseudomorphoses of the ice wedges developed above the interpleniglacial soils, and reach the size of poligons 20–25 m, the depth of wedges up to 4–6 m, and width up to 1 m (Maruszczak 1980, 1985, Jersak 1973). There are also here in two horizons: the lower one directly above the Komorniki soil or loess layer dated at 24±2.5 ka BP. and the upper smaller wedges in the top part of the youngest loess (Jersak 1975, Lindner 1985), probably synchronous with the expansion of the permafrost in the deglaciated area (Maruszczak 1960, Kozarski 1974).

The relicts of an intensive frost weathering are represented by rocks in the Holy Cross Mts (Klatka 1962, Lindner 1971) and in the Carpathians (Klimaszewski 1948, Baumgart-Kotarba 1974), where there also exist narrow cryoplanation terraces. The slopes of the flysch Carpathians at a distance of 300–400 km from the ice sheet are covered with solifluction deposits. In Dobra, the layer dated at 32 550+450 BP is overlain with 9 m thick loams with debris (Klimaszewski 1971). The existence of 3–4 clayey horizons in that section, as well as at Krościenko (Klimaszewski et al. 1950) indicates the existence of alternating phases with thick and shallow active layers controlled by climatic oscillations. The slope hooks of the upper solifluction bed rooted in the clayey shales on the slope in Solina in the upper San river valley indicate an extremely continental climate at the end of the existence of the permafrost (Starkel 1965). This is also proved by the rhythmically stratified deluvia in Błażowa (Krygowski 1960), and a loess blanket on the solifluction sequence at Wadowice with a different composition of heavy minerals (Sobolewska et al. 1964).

In the caves of limestone uplands the interpleniglacial beds with paleolithic cultures are covered with coarse debris, which indicates an intensive frost weathering under a relatively humid climate (Madeyska 1979). At the top there are silty layers without debris and with a steppe fauna with *Dicrostonyx* related to cold and dry climate (Kowalski 1974, Madeyska 1977).

## EOLIAN ACTIVITY AND LOESS DEPOSITION

In all paleogeographic reconstructions the direct foreland of the ice sheet is described as the zone of strong falling winds. The deposition of loess followed at a distance of 200–300 km (cf. Maruszczak 1968, Velichko 1984). The lateglacial dunes are superimposed either on older fluvial deposits (the oldest in Kamion dated at 14 590±270 BP – cf. Manikowska 1985), or more frequently on ventifact pavement related to the coldest phase (Dylikowa 1967). This thin stony blanket exists also below the younger slope deposits and indicates the existence of a polar desert (Dylik 1969). How far southwards did that zone extend? It depended on the character of the substratum and on the vegetation cover. The N–S elongated deflational troughs in the zone of the wide transversal depression of the Carpathians filled with lateglacial lacustrine sediments (Gerlach et al. 1972) and surrounded by hills with ventifacts (Gerlach et al. 1983) indicate that before the lateglacial warming the arctic desert under the favourable aerodynamic conditions was present even 300–400 km to the south.

The deposition of the younger upper loess in the zone of the Southern Polish Uplands and at the margin of the Carpathians and Sudetes Mts. followed at the elevation between 180 and 400 m a.s.l. mainly at the windward side (Maruszczak 1968). This opinion is corroborated by the thickness pattern of the loess blanket, as well as by the unrecognized extensive granulometric studies of the Podolian loess showing the diminishing of the grain size towards SW (Tokarski 1936). In the opinion of Malicki (1950) most loess deposits are connected with short distance transport which is proved by the composition of heavy minerals (Chlebowski and Lindner 1975, Racinowski 1976). To the valley floors the loess layer was redeposited from the slopes (Maruszczak 1985). In Jersak's (1976) opinion the accumulation was restricted to fine-grain substratum with warm soils covered with a dense grass-shrub tundra vegetation.

In the light of many TL and several $^{14}C$ and collagen datings the deposition of the younger loess started in the more humid phase 28–24ka BP; next it followed in the colder and drier climate, and finished ca 15–12ka BP (Maruszczak 1985). The sedimentation rate between 24 and 15ka BP reached 0.5–1 mm $yr^{-1}$ and the total thickness of younger loess reached 8–10 m. The most recent dating at Zwierzyniec give $14 \pm 2$ka BP (Konecka-Betley and Madeyska 1985). The distinct horizon of the arctic soil dated at 29–25ka BP at the base of the younger upper loess may be correlated with the Komorniki soil complex (cf. Jersak 1973) and in the Podolian Plateau it has the character of chernozem soils (Ivanova 1969). Two younger soil horizons inside the younger loess have the character of poorly developed brown artic soils with gleying. In some profiles they were dated at $21 \pm 3$ and $17 \pm 3$ka BP (Proszczyńska-Bordas 1985). In the W–E transect there are differences expressing the increasing continentality of the climate (Jersak 1975, 1985). In western Poland the younger loess has more distinct gley horizons and very rare pseudomorphoses of ice wedges. An opposite situation occurs in the Lublin Upland (Fig. 2).

## FLUVIAL ACTIVITY IN THE PERIGLACIAL ENVIRONMENT

There exists a general opinion, that the rivers of the periglacial zone have braided channels and that they aggradate. But we know that the valleys of the great rivers: Vistula, Oder and Warta were at first blocked by the ice sheet and later on dissected during the ice retreat. Simultaneously, small catchments responded with great delay to those variations of the base level. At the ice foreland in the Warsaw Basin there was formed a big dammed lake at the Radzymin level (Nowak 1974) and during its dissection a sequence of cut and fill terraces was formed (cf. Baraniecka and Konecka-Betley – 1987). In the basin of the upper Vistula there exists evidence of downcutting during the Interpleniglacial (Mycielska-Dowgiałło 1977, Starkel 1977a) as well as during the upper Pleniglacial. The latter fact is proved by the deposition of loess and deluvia on the floodplain sediments dated at 28–27ka BP in the Vistula valley at Opatowiec (Jersak 1976), Nowa Huta (Mamakowa and Środoń 1977) and Kaniów (Gilot et al. 1982). The bedded sands overlying the horizon dated $28\,500 \pm 430$ BP at Brzeźnica on the Wisłoka river (Mamakowa and Starkel 1974) and the peat layer at Łążek on the Sanna river dated at 25 580 ± 3270 BP (Bielecka 1960) indicate that the dissection started a little later. Similar alluvial fills from the first phase of the upper Pleniglacial were described at the Łódź Upland; at Bełchatów the peat leyer at the top of such a fill was dated at 21 970 + 810 BP (Baraniecka 1980). There is no doubt that a general downcutting started with the retreat of ice and with the increasing continentality (Fig. 2). Even in the Nowy Targ Basin the lower erosional step is covered with Bølling deposits (Klimaszewski 1961) and in the middle course of the San river valley the pre-Alleröd erosion reached the level of the present-day channel (Mamakowa 1962). But the valley floors of the small Carpathian creeks were continuously filled up with solifluction deposits, which is indicated by the interfingering of slope and fluvial deposits at Dziurdziów (Starkel 1965) and Lipowe (Starkel 1969).

## THE VEGETATION COVER

There are many records that during the Interpleniglacial there prevailed in Southern Poland open woodlands reaching in the Carpathians an elevation of ca 650 m a.s.l. and the northern forest limit crossed the centre of the Małopolska Upland and followed the southern margin of the Roztocze Upland (Mamakowa and Środoń 1977). In the localities on the wide valley floors of the Sandomierz Basin there were found the layers with a typical tundra spectrum dated in Brzeźnica at 28 500 ± 430 BP (Mamakowa and Starkel 1974) and in Nowa Huta at 27 745 ± 300 BP (Mamakowa and Środoń 1977). But independently of those datings, at the margin of the Carpathian Foothills at Podgrodzie in the solifluction intercalation the wood of *Picea* (vel *Larix*), was found dated at 26 980 ± 345 BP, and directly above it, in the alluvial muds dated at 22 450 ± 340 BP, the pollen spectrum dominated by trees (50–60% AP) shows the presence of *Pinus silvestris*, *Pinus cembra*, *Larix*, *Betula* and *Populus* (Mamakowa and Starkel 1977). These are clear indications of the survival of tree patches on the Carpathian margin during the coolest phase (Fig. 3). The forest-tundra spectrum from Łążek on the northern margin of the Sandomierz Basin, dated at 25 580 ± 3270 BP, includes abundant pollen of *Pinus*, *Betula*, *Larix* and *Alnus* (Mamakowa 1968). At the site of the mammoth hunters at Zwierzyniec in Cracow dated ca 23–20ka BP there were found charcoals of unidentified trees (Kozłowski et al. 1974). Moreover,

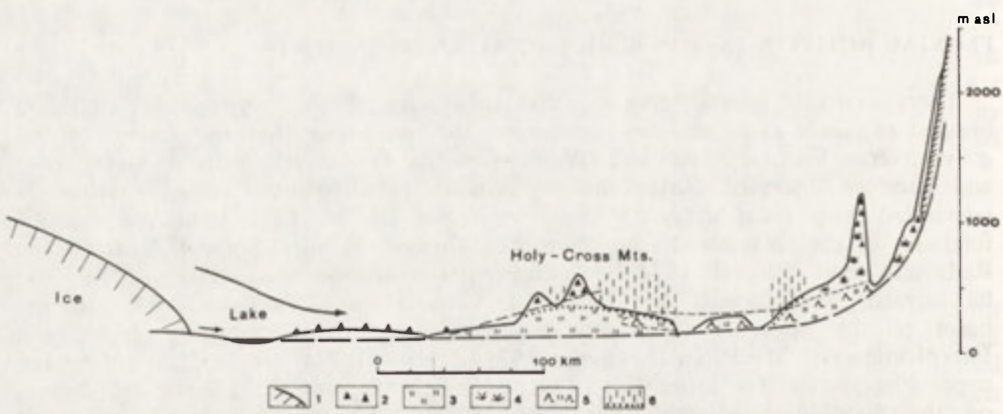


Fig. 3. Paleographic profile across Poland during the maximum advance of the Vistulian ice sheet: 1 – glaciers (ice sheet and in Tatra mountains), 2 – arctic desert, 3 – tundra-steppe with some trees, 4 – alpine tundra, 5 – tundra-forest-steppe, 6 – zone of loess deposition

in the cave deposits of the Cracow Upland a mixed tundra-steppe-forest fauna indicates that in that highland landscape there existed various ecosystems side by side, as we observe them in the carstic canyons at present (Madeyska 1979). Especially interesting is the presence of bird species typical of boreal forests, such as *Lyrurus tetrix, Falco columbaria* and *Strix uralensis* (Bocheński 1974).

All these facts support the opinion on the survival of at least small tree groups under favourable mesoclimatic conditions similar as those which exist in central Mongolian highlands, where the upper and the lower tree line bring them so close together (cf. Kowalkowski and Starkel 1984). The existence of local tree refugia in the Carpathians is evidenced by the locality at Smerek in the Polish Eastern Carpathians at the elevation of 600 m a.s.l. (Ralska-Jasiewiczowa 1980). In the peat layer dated at 16 925±325 BP the percent of AP decreases to the top from 80 to 60% and among those there are *Pinus silvestris, Pinus cembra, Larix, Picea* and *Betula* (as well as pine wood fragments). This spectrum is characteristic of the upper timber line and may be correlated with the Lascaux or Ula interstadial (Chebotarieva and Makaricheva 1974) or with the Masurian interstadial previous to the Pomeranian phase (Halicki 1960).

The data presented above may be compared with the general pictures of vegetation patterns during the maximum glaciation reconstructed by Frenzel and Grichuk who had not at their disposal most of these data. Frenzel (1967) shows the narrow zone of the shrub tundra, to the south – the loess tundra belt, and along the northern margin of the Carpathians a narrow belt of the foreststeppe – foresttundra. Grichuk (1982) shows in the territory of Poland in the immediate ice foreland the arctic tundra with single trees (*Larix, Betula*) changing to the east in tundra-steppe communities with single trees. The south-eastern part was occupied by a periglacial steppe with patches of forests (*Pinus, Betula, Larix*), and the higher Carpathians by an alpine-steppe vegetation. This reconstruction stands in contrast with the indications of the arctic desert. The areal pattern of habitats and vegetation were unquestionably very differentiated.

## CLIMATE AND GEOECOSYSTEMS DURING THE MAXIMUM EXTENT OF THE LAST ICE SHEET

Trying to reconstruct the habitats, geomorphic processes as well as controlling them climatic and hydrologic conditions in the periglacial zone we should remember first that at present we do not find full equivalents of the pleniglacial hyperzonal pattern of geoecosystems (Velichko 1984). Therefore all reconstructions are burdened with many errors. Secondly, under the conditions of limited warmth and water, needed by various plant communities, the leading role in the areal distribution was played by the relief and lithology, deciding on the mesoclimatic variation and thickness of the active layer (cf. Madeyska 1979, Starkel 1977b). And, thirdly, that the period of the upper pleniglacial had no uniform climate, which is indicated by the soil horizons in the loess, clay layers in the solifluction debris, and oscillations of the ice sheet front as well as of the small Tatra glaciers (Maruszczak 1985, Klimaszewski 1967). In the circumstances of not very precise datings this can cause quite incorrect correlations of events.

Taking the period between 22 and 18 ka BP as basic for our climatic reconstructions we conclude that it was characterised by very low annual temperatures ca minus 5–8°C, which enabled the formation of large poligonal structures (Jersak 1975, Starkel 1977b, Maruszczak 1985). The winter temperature fell to −30°C and below (Maruszczak 1980, Starkel 1977b), which is in agreement with the reconstructions made for the Dnestr and Dnepr river basins (Velichko 1984).

But summer temperatures are the most controversial. In the opinion of Mamakowa and Srodoń (1977) and Jersak (1985) the July temperature was below +10°C. But the localities not far from the eastern border of Poland show the temperature +14°C and more (Grichuk et al. 1984, Velichko 1984), as well as a steep gradient from the ice sheet margin.

The preservation of the tree species at the margin of the Carpathians indicates that the July temperature was above +10°C. If 17 ka BP at 600 m a.s.l. there existed a vegetation of the forest-tundra ecotone (cf. Ralska-Jasiewiczowa 1980) then at the elevation of 200 m at the margin of the mountains the temperature might have reached at that time in July +13–14°C. Short and warm continental summers made possible the existence of trees which reproduced also in a vegetative way. Their western limit probably ran across western Poland.

The reconstruction of the annual precipitation is more complicated. Maruszczak (1985) estimates it at 250 mm, and Jersak (1985) at 300 mm. We should take into account that the active layer of permafrost fed the forest communities with water, and that the annual precipitation might have even lower; (now in the forest zone of Central Yakutia precipitation below 200 mm are also recorded). In the mountains precipitation was doubtless higher, which is indicated by the prevailing solifluction deposits; later on, during the ice retreat and increasing aridity, they decreased in lower elevations and were replaced by deflation or deposition of loess. Grichuk (1982) gives very interesting information for Eastern Europe on the progressing aridity and invasion of steppe to the north from 52°30′N latitude at 25 ka BP to 60°N at 21 ka BP. This indicates a change of the annual precipitation with time, and an increase of aridity. Trying to find analogs in the zonal differentiation of the present-day Asia we may say that the periglacial climate of Central Europe during the maximum cooling turned from the Siberian to the Mongolian type (cf. Starkel 1977b).

Undoubtedly, the characteristic feature of the periglacial climate in the Polish territory during the maximum extent of the last ice sheet consisted in the increasing continentality towards east (Jersak 1973, 1985), as well as a steep latitudinal gradient (Fig. 3) expressed in the fast transition from an arctic desert to subarctic tundra, forest-tundra and tundra-forest-steppe (cf. Grichuk et al. 1984). As a result, in the Polish territory south of the narrow arctic desert belt there extended two periglacial provinces (Fig. 4): the criosemihumid one covering the Oder river basin and the Carpathians, and the criosemiarid one extending over the S-Polish uplands and the Subcarpathian basin.

On this general pattern there was superimposed a mosaic of geoecosystems: inversional floors of valleys and basins, wet loess plain and dry sandy plains, warm slopes and cool slopes, windy ridges and intramontane gates. This variety conditioned the existence, side by side, of patches of cold deserts and enclaves of dense vegetation with trees, of areas of deflation and loess deposition.

There is no doubt that the climate of the phase had a very variable weather. The NE winds falling from the ice-sheet (Velichko 1984) were not blowing throughout the whole year. During the slow ice retreat there probably occurred occasionally even very hot summers, which supported the preservation of more demanding trees. There also seldom happened the advection of humid Atlantic air masses with heavy snow or rain. Future investigations should solve the questions: which of the recognised periglacial processes of the maximum ice extent phase were of a more secular character (every year) or if they were related to extreme events (with long recurrence interval) and which reconstructed picture is more realistic: either that of a synchronity of various processes areally differentiated, or the image of many short-term climatic variations during the general cooling of the upper Pleniglacial (cf. Wysoczański-Minkowicz 1982).

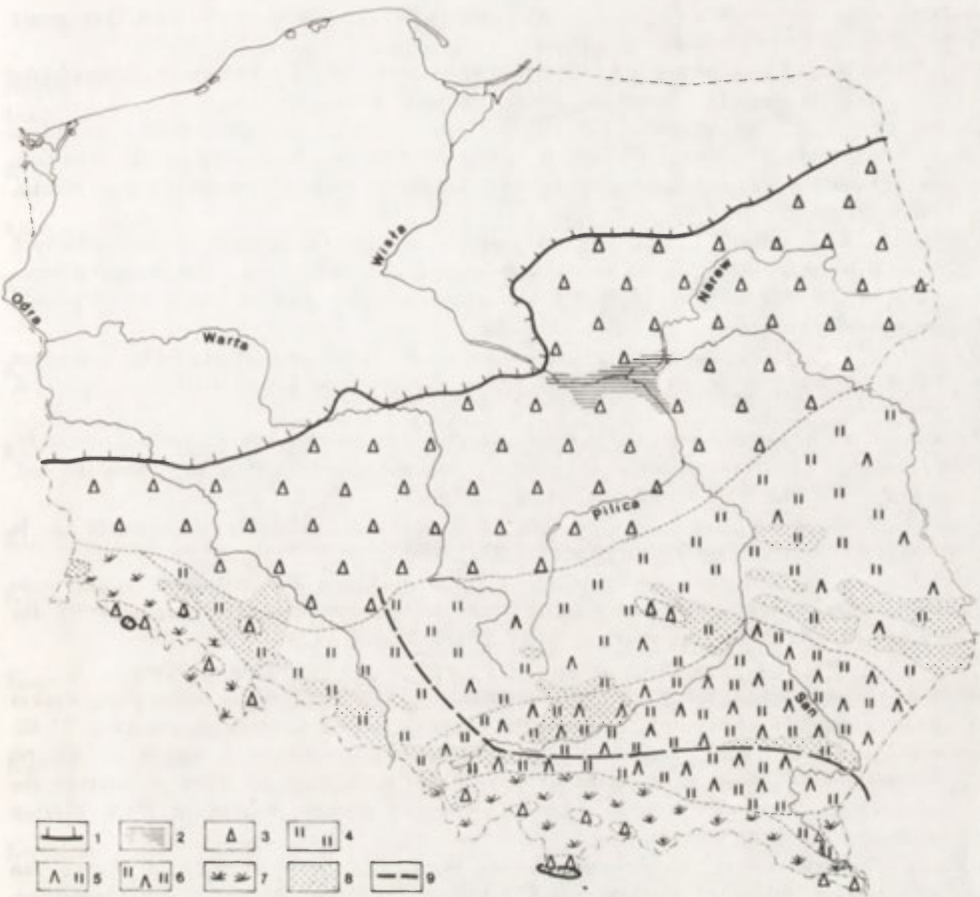


Fig. 4. Paleogeography of the periglacial zone during the maximum advance of the Vistulian ice sheet: 1 – ice margin, 2 – glacial lake, 3 – zone of arctic desert, 4 – subarctic tundra, 5 – subarctic tundra-steppe with some trees, 6 – tundra-forest-steppe, 7 – alpine tundra, 8 – main regions of loess deposition, 9 – facultative boundary between the criosemihumid province and the criosemiarid one

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# SLOPE PROCESSES IN NORTH SCANDINAVIA AND LATE HOLOCENE CLIMATIC IMPACT

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ABSTRACT. A review of previous studies on slope erosion in Scandinavia introduces a case study of past and recent slope process intensity on Andöya Island in Northern Norway. The pilot study is based on geomorphological mapping from aerial photographs, field controls and quantifications of sediment storage in slope deposits and investigations of lake sediments. The frequency of past catastrophic events (i.e. debris flow activity) was estimated by dendrochronology and studies of lake sedimentation.

## INTRODUCTION

Mass movements and other processes have in the past been regarded too much as accidental and rare events without much geomorphological significance, due to their erratic occurrence. In earlier papers by A. Rapp (1960, 1961) he concluded that mass movements in many environments were of major geomorphological importance. An improved analysis of type, frequency and magnitude of mass movements will strengthen the basis and the applications of geomorphology. Later observations of mass movements and their relations to climate (Kotarba and Strömquist 1984) and studies of Holocene climatic variations (Karlén 1973) have opened up new avenues to study the process activity in relation to climatic variations and to use the process activity as an indicator of past climatic changes.

One of the first projects on mass movements and their relations to the present climate was Rapp's (1961) project in Kärkevagge during the period 1952–1960. These studies dealt mainly with mass wasting processes of two groups: a) slow and continuous processes (gelifluction and creep), and b) rapid and sporadic processes (rockfall, avalanche, debris slides or flows). After the publication of the report on Kärkevagge several follow up studies have been performed in nearby areas of Norway and Sweden.

Reviews and bibliographies of literature on mass wasting in the Scandinavia have previously been published by Rapp (1961, 1964, 1965), Rapp and Rudberg (1964), Rudberg (1962, 1970) and Rapp and Strömquist (1976). A recent work on slope processes by Nyberg (1985) gives new information on the process activity as well as reviews of previous works.

### SLOW MASS MOVEMENTS

Periglacial environment and mass wasting of mountain slopes was mapped and described by Rudberg (1960, 1962, 1964, 1970, 1974). He published detailed geomorphological maps of process type and activity. The rate of downslope soil movements by gelifluction and creep has been estimated by several scientific workers, working in different Northern Scandinavian environments – cf. Rudberg 1958 and later, Rapp 1961, Williams (1957, 1960) Strömquist (1973, 1975, 1977 and 1983) and Harris (1973 and 1982). The surface movement rates are generally low (mm to cm/yr). Geomorphological mapping of the forms has aimed at establishing a zonation of forms and processes in relation to altitudinal zonation of present day climate (cf. Harris 1982). Findings from for instance Kärkevagge Valley (Strömquist 1983), however, indicate that the distribution and frequency of gelifluction lobes is limited by the topography and supply of material in the upper part of slopes and by an increased fluvial activity at the bases. The same study also showed that the gelifluction lobes were rather short lived features due to fluvial slope erosion of the forms. Therefore the forms cannot be used for dating the total Holocene activity as made by Benedict (1970) in the Colorade Rockies.

#### RAPID MASS MOVEMENTS

Rapp's (1960, 1961) works on talus development concentrated on estimating the recent process types and activities as well as the total post glacial deposition and relocation of material. In Kärkevagge, for instance, Rapp (1961) concentrated on observations of recent deposition of materials and relocation by talus creep and other processes. Later works on talus development in other periglacial environments (i.e. Bones 1973; Church et al. 1979; White 1981) has classified the talus slopes into different types according to the dominant recent land forming processes.

A study in the Polish Tatra Mountains by Kotarba and Strömquist (1984) has illustrated the possibilities to use different types of recent talus transformation processes (i.e. increased fluvial activity, less deposition of materials etc.) to discuss the impact of climatic fluctuations. A similar approach is used in this project.

Judging from the morphology and slope deposits of several Scandinavian mountain areas the rapid and sporadic mass movements (i.e. slides and debris flows) have a dominant impact on the postglacial evolution of slopes. Previous works are summarized by Rapp and Strömquist (1979), Rapp and Nyberg (1981) and Nyberg (1985).

## A PILOT STUDY ON WESTERN ANDOYA ISLAND

The study area, the Åknes catchment (Fig. 1) is situated on the Andoya Island in northern Norway. The bedrock consists of different gneiss-granites. The catchment is exposed towards north-west. The hillslopes have pronounced glaciated profiles and are covered by till or slope deposits. Interpretation of aerial photographs from 1978 have used to map the spatial distribution of sediment sources and sediment storage areas. The fieldwork concerning the late Holocene process activity has been concentrated to studies of debris flow activity.

#### METHODS

The study is based on conventional field methods, air-photo interpretation and studies of lake sedimentation and analyses of sediment cores according to methods developed by Axelsson (1983). The sediment budget approach (Gilbert 1917) is tested for the inventories of sediment sources, sediment routing and sediment storage.

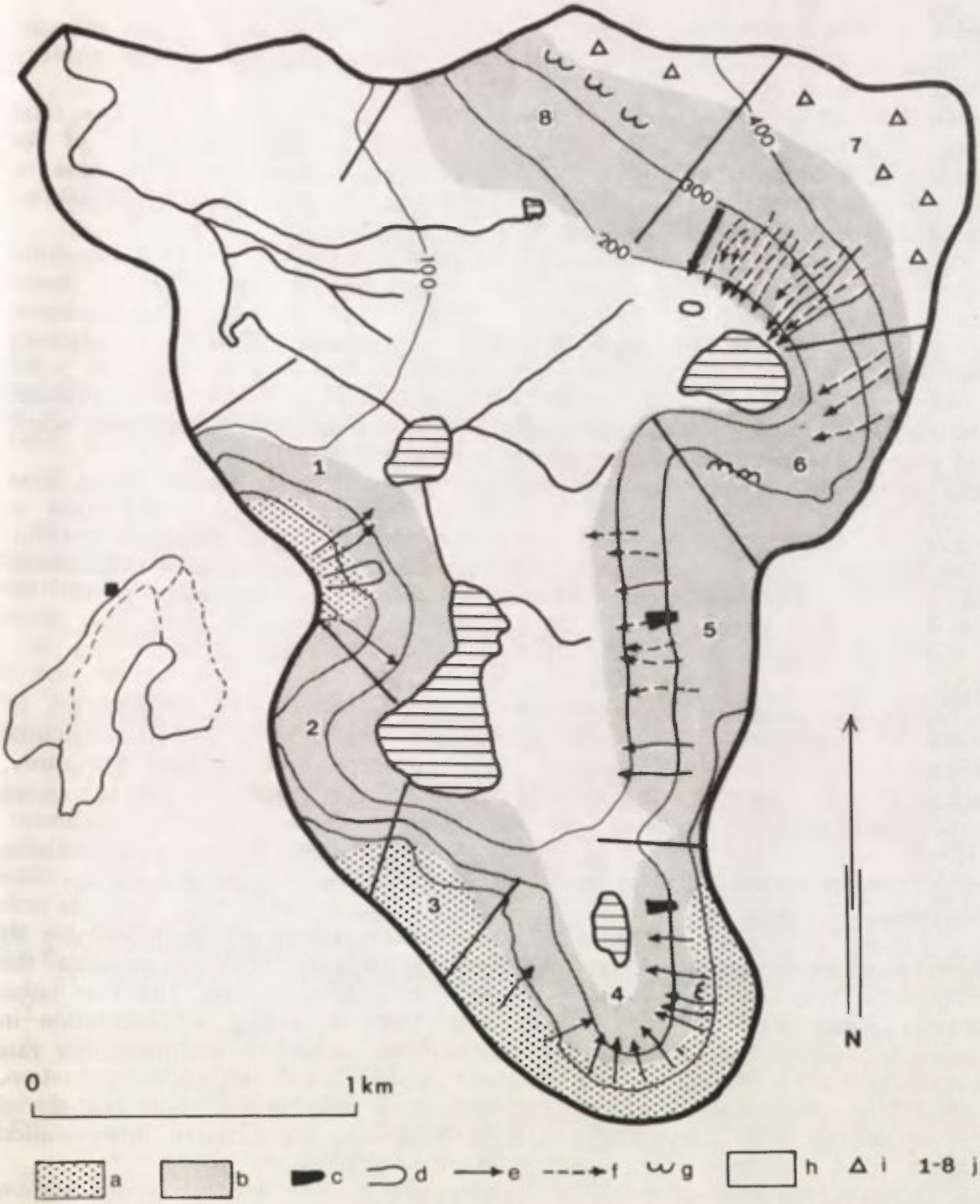


Fig. 1. Geomorphological map. Åknes catchment: a – free face, b – vegetated talus, c – landslide, d – rockslide, e – fresh debris flow scars, f – vegetated debris flow scars, g – gelifluction lobe, h – till deposits, i – block field, j – slope sections (1–8)

SLOPE PROCESSES

The hillslopes are dominanted by talus sheets (55% of the hillslopes) but the rate of recent talus formation seems to be rather low. Fresh debris accumulation occurs only at one place in the study area (Section 1, Fig. 1). Most of the scree slopes are, except from debris flow scars, totally covered by vegetation. The relation between surfaces covered by unvegetated and vegetated talus slopes is 1 to 21, indicating a low recent rate of scree formation. A rough estimate of the volume based on area, height, pore volume and bedrock morphology indicates a total storage of about $150 \times 10^6$ tonnes.

Gelifluction occurs in three of the sections: 4,6 and 8 (Fig. 1). The total area dominated by gelifluction is 3% of the total catchment area (4% of the hillslopes). The probable amount of material being relocated by gelifluction is, based on observations of soil depth and surface area, approximately 96 000 $m^3$ or 144 000–201 000 tonnes.

The last known event of intense debris flow activity was the 1959 rainstorm caused by a front passage from the south-west (cf. Strömquist 1976), hence the most recent debris flow scars are more frequent on the west-facing hillslopes. Quite a few of the scars are completely unvegetated indicating a relatively recent process activity. Other scars are totally covered by vegetation, in some cases by birches with a stem diameter of more than 10 cm. In order to evaluate the sediment transfer made by the debris flows and the age the different scars, the eastern part of section 7 was examined in detail (Fig. 1).

Dendrochronological measurements indicate a return period for the debris flows of 25–60 years. The corresponds with observations of extreme events made at Bjornskinn (some 5 km from the study area) and at Åknes (Strömquist 1976). The most recent observable debris transport within the hillslopes in the catchment, as estimated from the fresh debris slide scars, is approximately 52 000 $m^3$ (78 000–109 200 tonnes).

## LATE HOLOCENE PROCESS ACTIVITY

As it is possible to date the past debris flow activity by studying lake sedimentation we have concentrated our efforts to this process type. Previously, sediment studies have been made by Ostrem and Olsen (1975) to study sub-recent process activity. Sediment cores were taken in three of the lakes in the catchment. The sediment cores had been analysed by X-ray analyses according to methods developed by Axelsson (1983). Two of the cores were found suitable for our purpose.

In order to identify the different periods with extreme events (noticeable as light tone layers on the X-ray photographs) we have tried to correlate the uppermost layers to periods of increased debris flow activity. The two latest known events occurred in 1959 and about 1900. According to observation in parts of Swedish Lappland (Karlen, oral communication) the sedimentation rate vary between 0.1 and 0.2 mm/year. Lacking accurate local data on sedimentation, which ought to be calculated from sediment cores covering the whole post-glacial period in the area, Karlen's figures have been used for a rough interpretation of our results.

The first sediment core of 25 cm reached, due to the low sedimentation rate in that lake, back to about 3000 years BP. The second 25 cm core, from a smaller lake with surroundings more favourable for formation of debris flows, covers only about 600 years. The sedimentation rate is important for an accurate identification of periods with catastrophic events as the thin layers from the first core are very difficult to identify and correlate.

To test the possibilities of this method we have chosen to use the 600 year sediment core as it is easier to correlate with Karlen's (1984) dendrochronological curve. Twenty periods/events of increased slope activity could be identified. In general the present frequency has been predominant since about the year 1800. A previous period from 1350 to 1550 was characterized by a higher intensity in general (Fig. 2), a period with warmer summer temperatures.

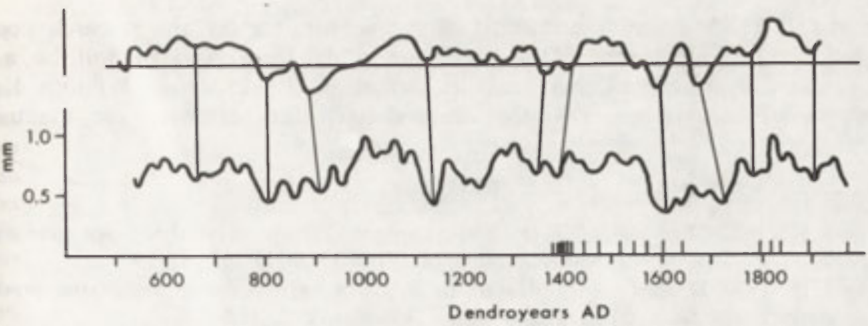


Fig. 2. Curves showing tree ring width variations (lower curve) in North Scandinavia and fluctuations in the organic content (upper curve) of pro-glacial lacustrine sediment in a core from Vuolep Allkasjaure, a lake situated in Northern Sweden (from Karlen 1984). The markings on the X-axis show periods or events of an increased debris flow activity in the Åknes catchment area on Andöya, Northern Norway

From about 1550 years AD the periods of increased activity seem to be correlated with periods with falling summer temperatures. According to licheno-metrical measurements made by Nyberg (1985) near Abisko, many debris flow events appear to correspond to cold periods.

A question for future research in this field is therefore if it is possible to correlate the activity also with more long lasting changes of climate. Does the period with an increased activity from 1350 years AD correspond with periods with more convective rains?

## FUTURE RESEARCH DEVELOPMENT

It is notable to see that the sedimentation-curve from the proglacial lake in Karlen (1984), at least in the upper part, is more or less inversed to the core taken at Andöya. Periods woth lower summer temperatures decrease the sedimentation rate in the proglacial lake but increase the amount of triggered debris flows. The future work will be concentrated on the following questions:

1. The differences in lake sedimentation rate (covering post-glacial time).
2. Differences in interfluvial material relocation (covering late Holocene).
3. Differences in sediment-budget (recent process activity).

## CONCLUSIONS

The preliminary studies on western Andöya have clearly illustrated the possibility of a catchment area based approach on studies of past and recent slope process activity. Hence it is possible to modify the sediment budget approach, first suggested by Gilbert (1917), to this type of studies identifying the main sediment source areas (i.e. debris flow scars etc.), sediment routing and sediment storage areas (i.e. slope deposits, lake sediments). It is furthermore possible to use dendrochronology and lake sedimentation figures to date past slope activity related to other climatic situations.

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# MORPHOLOGY AND ICE MELTING IN A POMERANIAN OUTWASH PLAIN, WDA VALLEY

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ABSTRACT. In the proximal zone of the Wda outwash the occurrence of eight outwash levels associated with the outflow of glacial meltwater from the Pomeranian phase was confirmed. Within the outwash a number of subglacial channels and kettle-holes occurs. Part of these landforms is filled with organic sediments – peat and gyttja up to 15 m in thickness. The oldest sediments from the base of these sediments are of Bölling age – dating $^{14}C$ is 12 170 ± 250 yr BP.

## INTRODUCTION

In his monograph on the morphology of the Brda valley and outwash plain Galon (1953) put forward a hypothesis that the most important problems associated with the Polish Lowland involved setting up an end-moraine system and determining the development of pradolinas. Sandur tracks serve as a link between end moraine forms and pradolinas. Outwash plains lying between end moraines belonging to the Pomeranian phase or stage of the Vistulian glaciation and the Toruń–Eberswalde pradolina have become particulary extensive. A genetic relationship of the moraines belonging to the Pomeranian phase of the last glaciation and the outwash sediments with the Toruń – Eberswalde (Noteć – Warta) pradolina was first established by Ost (1933) as a result of the study of the Gwda and Drawa outwash morphology. It was confirmed twenty years later on by Galon (1953) on the basis of detailed investigations carried out in the Brda outwash and valley. Further research conducted in the Noteć – Warta pradolina (Kozarski and Szupryczyński 1958; Galon 1961b; Kozarski 1965) and in the outwash plains associated with the marginal zone belonging to the Pomeranian phase supports the suggestion that meltwater which issued from the marginal zone of end moraines belonging to the Pomeranian phase flowed southward towards the pradolina and was next carried westward through a pradolina track (Galon 1956; Kozarski 1965; Niewiarowski 1968).

## STUDY AREA

When carrying out investigations in the Wda (Czarna · Woda) river outwash, the present author's main aim was to reconstruct particular stages of outwash and valley relief development, as well as to establish relationships of the outwash with the Vistula valley and with the Noteć–Warta pradolina through the lower Vistula valley reaches. Over the period 1947–1950 Okołowicz (1948, 1949, 1952) with a group of colleagues and students conducted research in the Wda outwash area. The resulting data were published as a morphogenetic sketch of an area between the Słupia and Vistula river catchments, which showed the main sandur tracks in this area, sandur tracks associated with the end moraines belonging to the Pomeranian phase and those developed in the hinterland due to the recession of an ice sheet northward. A few years later, Churska and Rosa gave valuable contributions concerning the proximal side of the Wda river outwash in the first part of the 6th INQUA Congress Guidebook which was devoted to areas where the last glaciation had taken place. Churska (1961) distinguished between four outwash levels absolute heights of 220–210, 195, 185 and 175, m a.s.l. The earliest level which remains highest is associated with a complex of the Wieżyca end moraines, whereas the second level joins end moraines crossing the Raduńskie Lake basin and the so-called moraines of the western lobe. According to Churska, the third level 185 m a.s.l. high is connected with end moraines lying north of Kartuzy, whereas the lowest level marks the standstill position of an ice sheet along the Szemud-Koleczkowo line and in the vicinity of Bytów. In a brief report Rosa (1961) provides a description of the geologic structure of an outwash level in the vicinity of Korne village to find out that it is erosional-accumulative in some places. I published a brief note in Belgium, concerning the preliminary data on the development of erosional valleys in the Wda outwash area in the vicinity of the Korne village (Szupryczyński 1967). A map of outwash levels in the vicinity of the Korne village lying 8 km west of Kościerzyna was enclosed with the note.

## THE WDA OUTWASH LEVELS

The outwash areas of the marginal zone belonging to the Pomeranian phase have also been studied by Sylwestrzak (1973) who distinguishes between eight outwash levels to the west of Kościerzyna, the absolute heights (m a.s.l.) of which are as follows: I – 220–210, II – 205–200, III – 190–185, IV – 180–175, V – approximately 170, VI – 165–160, VII – 155–150, VIII – approximately 145. When compiling the map of distribution of outwash levels, Sylwestrzak utilized the results of studies carried out by Okołowicz (1956), Churska (1961) and Szupryczyński (1967). The present author is not against the number of the levels distinguished but in some cases he does not agree with their boundaries or the heights given. Note should be made, however, of the fact that Sylwestrzak carried out geomorphological mapping in the field, using 1:25 000-scale topographic base-maps, whereas the present author's field studies were based on new topographic base-maps at the scale of 1:10 000.

The highest levels occur as outwash fans formed immediately in the foreland of the moraines. In consequence of the action of meltwater issuing from the glacier during the late recession phases, the levels became seriously destroyed. They have been preserved in the shade of morainic islands and along some subglacial channels (Fig. 1).

There are small fragments of the third level with the absolute height of over 182 m, frequently not more than 50 m wide, left along the glacial channel of Wielkie Długie Lake. Forest-covered fragments of this level lie over 15 m higher, compared with level V adjacent to it in the east, and over 35 m higher, compared with the lake level in the glacial channel. Fragments of level IV

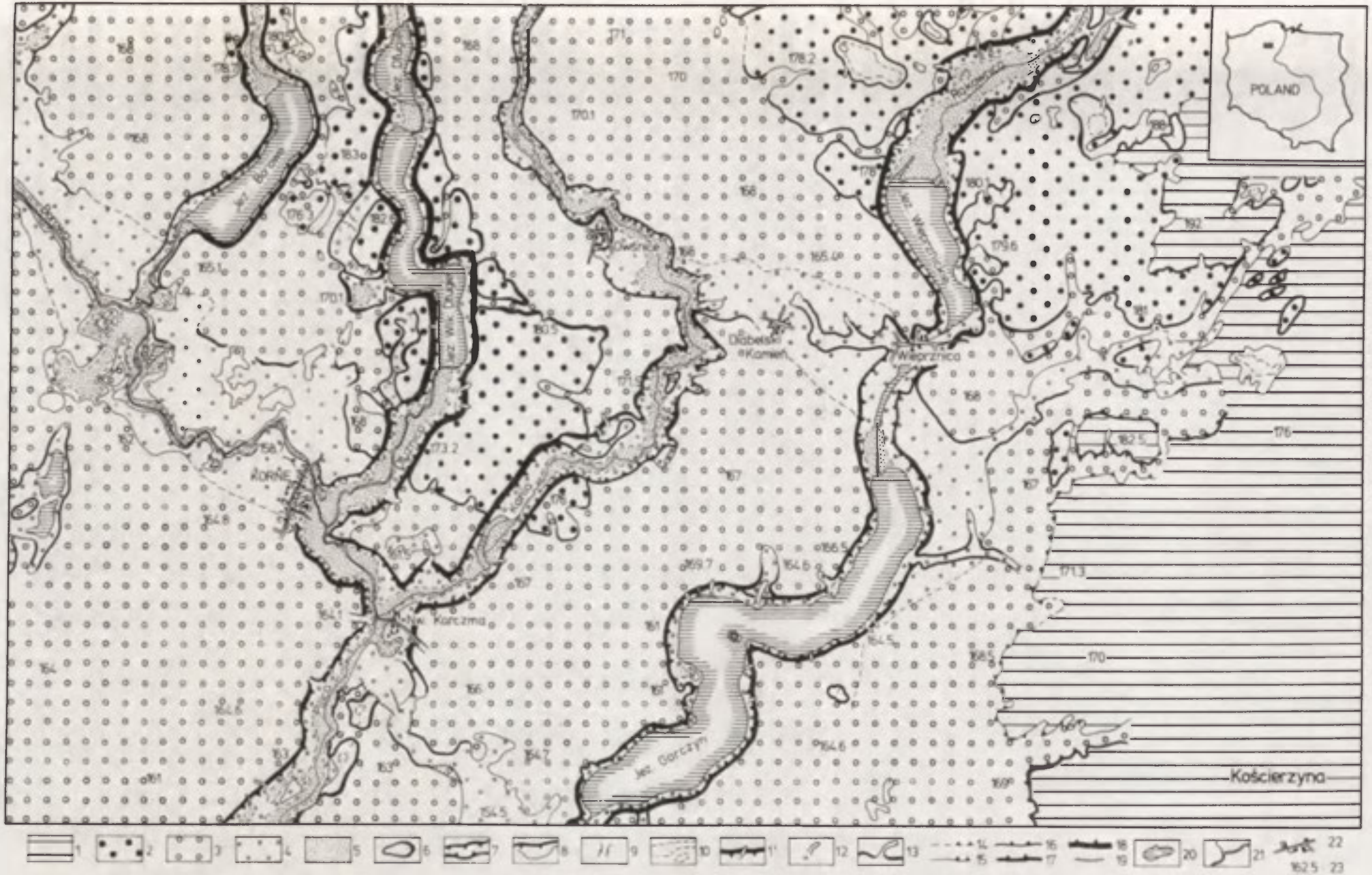


Fig. 1. 1 – moraine plateau, 2 – III-outwash level, 3 – IV-outwash level, 4 – V-outwash level, 5 – VI-outwash level, 6 – melt-out froms, 7 – subglacial channels, 8 – edge with aggradation zone, 9 – sill in the bottom of subglacial channels, 10 – peat areas, 11 – edge with small gully, 12 – gully with alluvial fan, 13 – large gully, 14 – indistinct-edge, 15 – edge up to 5 m, 16 – edge up to 10 m, 17 – edge up to 20 m, 18 – edge more than 20 m, 19 – erosional undercuttes, 20 – lakes, 21 – streams, 22 – village, 23 – hypsometric points

have also survived, their settings being similar to those of the higher levels. Outwash levels V and VI are most expanded in terms of space. Meltwater flowed southward over them through the Wdzydze Lake route into the Wda valley. Their fragments are found in the Wda valley as far as its entry to the Vistula river near Świecie (Drozdowski 1967).

## GEOLOGIC STRUCTURE

The study area of the Wda outwash and valley has contained so far 173 deep boreholes and 191 shallow boreholes drilled throughout field studies and about 150 shallow boreholes drilled by the "Geoprojekt" of Bydgoszcz. Seventy three exposures have been studied and 59 holes have been made to depths of 2.5 to 4.5 m. None of the deep boreholes on the proximal side of the outwash run through the Quaternary deposits. The greatest thickness of the Quaternary deposits has been found in the following localities: Nowa Kiszewa 84,0 m. Fingurowa Huta 75 m, Sylczno 77,0 m and Stężyca 76,0 m. Their thickness is likely to exceed 150 m in the adjacent morainic plateau (Nowy Klincz: 150 m, Quaternary deposits non-drilled, Nowa Karczna: 122 m, Quaternary deposits non-drilled).

Geological documentary evidence available for the middle Wda reaches is very scant. Several deep borings have been only taken but they do not exceed 35 m in depth. Some of them run through the Quaternary deposits in the lower reaches of the Wda valley. The greatest thickness of the Quaternary deposits, i.e. 157 m, has been found at the locality Wierzchy. The Quaternary deposits lie there over the Tetiary sediments. In the lower reaches of the Wda valley Tertiary deposits approach nearly the surface. Merely 1.2-m-thick and up-to-9.0-m-thick deposits occurring in the localities Gródek and Przechowo, respectively, provide a record of Quaternary time.

A question to be now asked is what thickness outwash sediments reach. It is difficult to answer since no reliable method of separation between younger outwash sediments and older fluvioglacial series is available. The present author is certain that it can be inferred from the documentary evidence collected that the depth of accumulation of outwash exceeds 20 m within the higher outwash levels and that it may attain greater thickness. Outwash sediments occurring in the foreland of small glaciers in Spitsbergen are 5 to 8 m thick (Szupryczyński 1968; Sendobry 1983).

The boreholes analysed from the surface downward reveal the greatest thickness of sandy-gravelly sediments in the localities Dąbrowa, i.e. 51 m along a narrow sandur track, and Łubiana within extensive outwash level VI, i.e. 15–31.5 m. A 41.0-m-thick series of sandy-gravelly deposits has been found within lower level VII in the locality Rybaki. Level VI in the locality Korne contains very thin, i.e. 3.2 to 6.0-m-thick, sandy-gravelly deposits or even glacial till at the surface in some places. This receives confirmation from Rosa's earlier conclusion (1961) that fragments of a meltwater erosional surface are present at this level. Thin outwash sediments, i.e. those 5.9–8.4 m thick, have been detected in boreholes in the localities Wdzydze and Olpuch.

A few tens of samples have been taken from the exposures and boreholes for particle-size and mineralogical analyses. No major differences have been established in the mineral contents between particular levels. The analysis has revealed that quartz which accounts for 87 to 94% prevails among minerals. A percentage of feldspars on the order of 4.3 to 9% is also of importance for the samples

Photo. 1. Wielkie Długie Lake in a subglacial channel

Photo 2. Subglacial channel near Korne


Photo 3. General view of VI outwash level



Photo 4. Small gully of subglacial edge at Wielkie Dlugie Lake

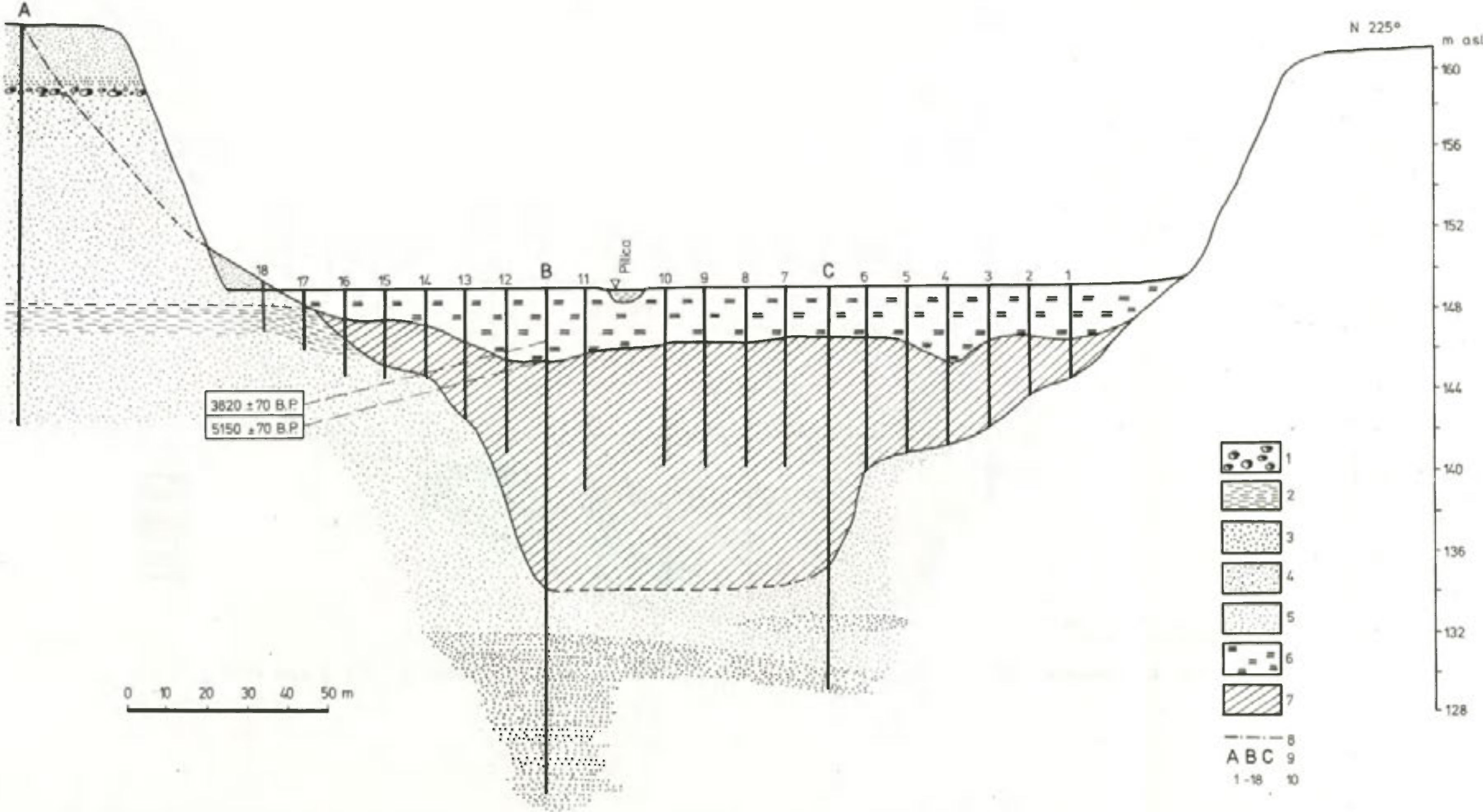


Fig. 3. Cross-section through subglacial channel at Korne: 1 – boulder pavement, 2 – water, 3 – coarse-grained sand, 4 – medium-grained sand, 5 – fine-grained sand, 6 – peat, 7 – gyttja, 8 – indistinct base of organic fill, 9 – deep drill-holes, 10 – shalow drill-holes

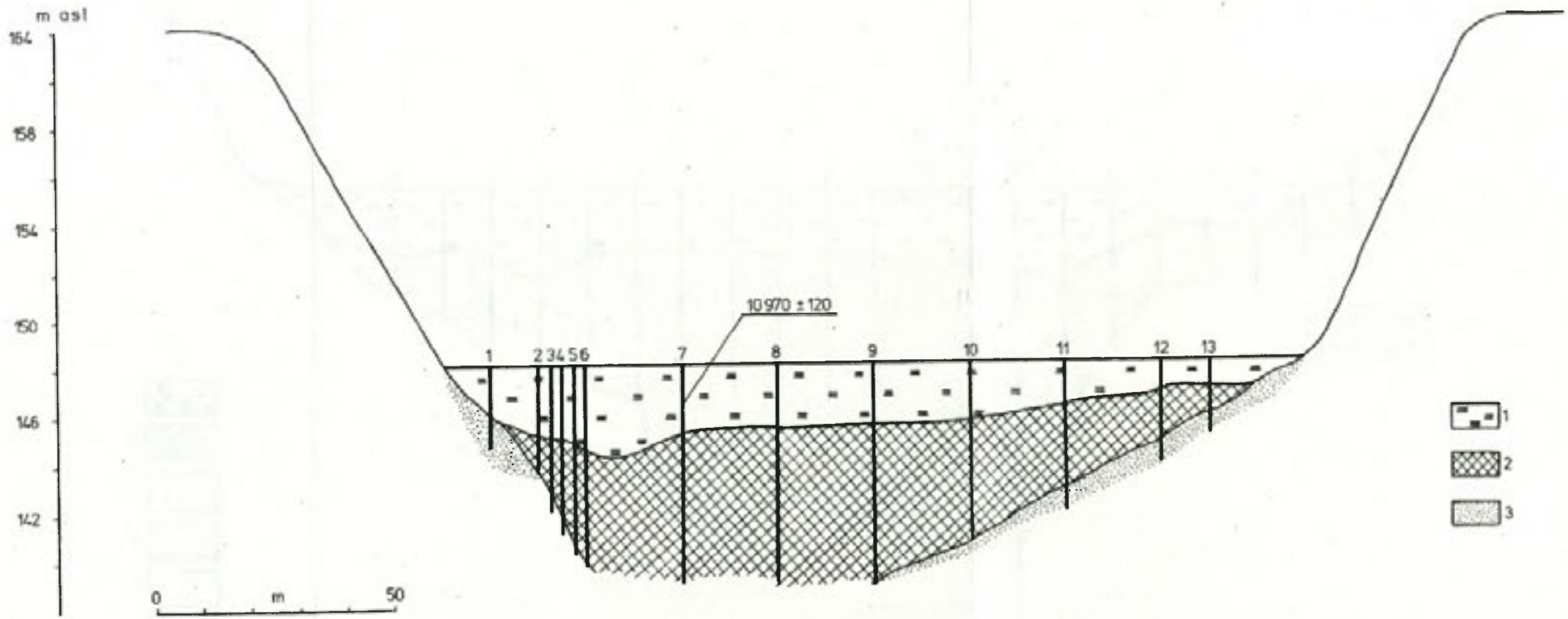


Fig. 4. Cross-section through subglacial channel of Kania stream: 1 – peat, 2 – gyttja, 3 – fine-grained sand

under investigation. Basides, other minerals are also present in minor amounts, namely garnet 0.2–0.6%, zircon, goethite, ilmenite, amphibole, biotite and pyroxene. As a rule, they represent up to 0.2% of the weight of a sample.

## HOLLOWS AND ICE MELTING

Particular outwash levels are separated from each other by conspicuous steps, relative heights of which approach 20 m. Glacial channels and melt-out forms which were produced after the melting out of buried winter ice are also characteristic of the landscape on the proximal side of the Wda outwash. Some of these forms are occupied by lakes or filled in by organic deposits (Figs 2–4). Some melt-out forms with depths of over 20 m are devoid of any organic fills. Organic sediments were sampled from the glacial channels and melt-out forms for palynological analysis and $^{14}C$ dating. The thickness of peat and gyttja organic fills of the channels approaches 15 m (Fig. 3). In 1983 five samples from these deposits were dated. The date of 12 230 ± 250 BP is indicative of the presence of the Bolling sediments in a narrow trough lying north of the Korne village, at the depth of scarcely 3.5 m. These sediments are overlain by the Allerod (11 120 + 120) and Younger Dryas (10 170 ± 90) deposits. The above dates indicate that the melting out of dead-ice blocks which filled the channels and melt-out forms began relatively abruptly and took place as early as during the Bolling. The melt-out forms probably vary in age. The youngest of them must have formed at the close of the Lateglacial or at the beginning of the Holocene. They are generally dry, i.e. remain unoccupied by lakes and nonfilled with organic deposits. The ice melted away after the permafrost had disappeared. Meltwater produced by the melting out of blocks infiltrated downward through sandy and gravelly deposits to feed deep-lying warter-bearing horizons.

The melt-out forms are present at all the outwash levels and on terraces along the Wda valley. A dead-ice depression has been found even on very

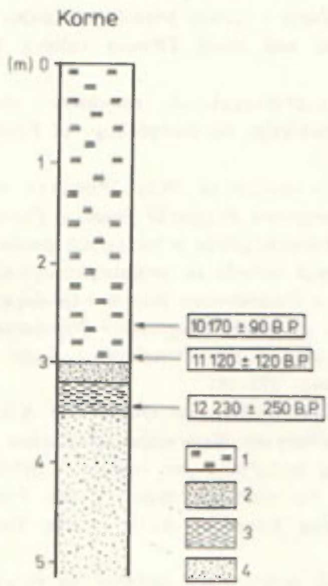


Fig. 2. Sediment sequence in bottom of subglacial channel at Korne: 1 – peat, 2 – medium-grained sand, 3 – silty-gyttja, 4 – fine-grained sand

low terraces of the Wda river, about 5 m in relative height above river level in the vicinity of the Mały Dolsk village (Okolowicz 1949). Galon (1953) detected dead-ice depressions on low terraces III or even II in the Brda outwash, to the south of the railway station Pruszcz-Bagiennica.

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# BIOSTRATIGRAPHICAL RECORD OF VISTULIAN DEPOSITS AT THE MAXIMUM EXTENT OF THE LAST ICE SHEET IN THE KONIN REGION, POLAND

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ABSTRACT. In this paper new sites with Eemian and Vistulian floras at the maximum extent of the last ice sheet near Konin are discussed. They show several climatic changes which can be correlated with some of the stadials and interstadials known from NW Europe, NW and Central Poland, i.e. Amersfoort, Brørup, Odderade, Stare Kurowo, Rudunki, Keller and Oerel.

## INTRODUCTION

In the recent years (until 1986) a few new sites with Upper Pleistocene floras have been found in northwestern Poland. Most of them lie in the vicinity of Konin (Fig. 1) where biogenic and lacustrine Quaternary sediments have been exposed many times in brown-coal open-cast mines (Stankowski and Tobolski 1981; Kłysz and Stankowski 1986).

The profile from Władysławów has provided the most complete record of Pleistocene events so far. It is from a discussed part of the brown-coal mine Władysławów situated halfway between Konin and Turek. The profile contains a 7 m thick continuous undisturbed sedimentary series rich in sporomorphs and macroscopic remains which has been identified with the close of the Saalian, the Eemian Interglacial and a considerable part of the Vistulian (Tobolski 1986).

The Eemian Interglacial is represented at Władysławów by an abundant macroflora of local aquatic and telmatic vegetation. *Brasenia purpurea*, *Cyperus glomeratus*, *Aldrovanda vesiculosa*, seeds and vegetative parts of *Stratiotes aloides* seeds of *Lemna sp.*, etc., are present. Forest vegetation belonging to this part of the interglacial is represented selectively in the macroscopic flora by single specimens. Numerous macrofossils of trees and shrubs occur in the transitional parts of this interglacial. Fruits and fruit scales of *Betula pubescens* and *B. pendula* and *Pinus sylvestris* needles have been found. A simplified pollen diagram from the profile for Władysławów is shown in Fig. 2. An equally diverse macroflora belonging to the Eemian Interglacial has been detected in deposits from Józwin, profile 76 (Tobolski 1986 manuscript).

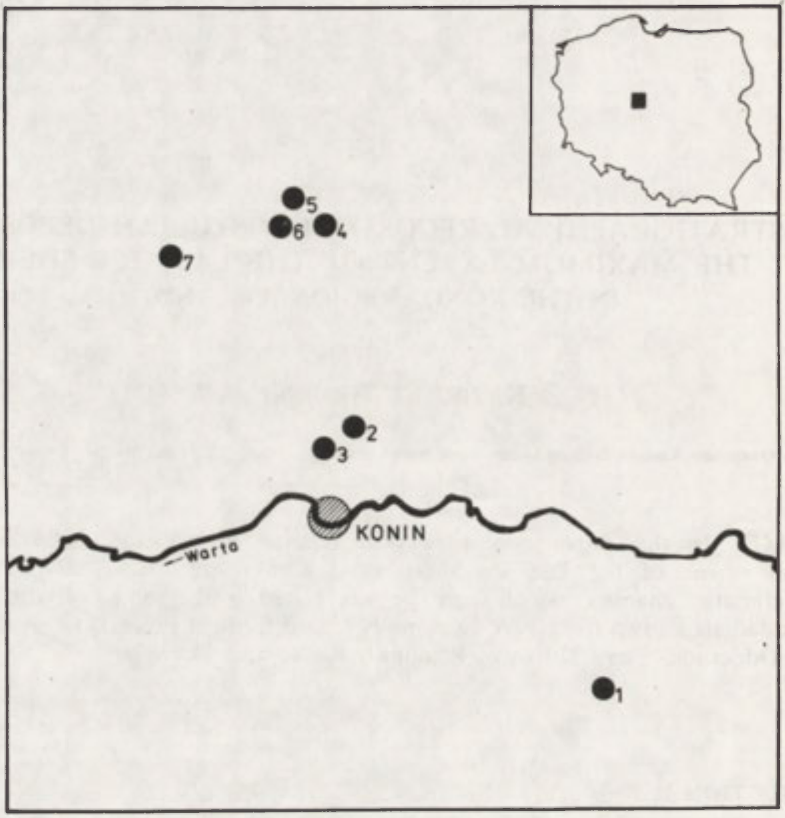


Fig. 1. Distribution of sites with Upper Pleistocene floras: 1 — Władysławów, 2 — Maliniec, 3 — Konin—Gosławice, 4 — Konin—Jóźwin (postglacial), 5 — Konin—Jóźwin—84, 6 — Konin—Jóźwin—76, 7 — Kazimierz

VISTULIAN FLORAS

Palaeobotanical studies of the profile from Władysławów reveal two Vistulian interstadials, both with forest cover and a consecutive interstadial with shrub vegetation. The earliest interstadial is separated from the Eemian Interglacial by a deposit whose palaeobotanical contents are indicative of a treeless landscape and continental climate. The AP values drop below 50% (*Cyperaceae* is absent from the NAP sum). Among the herbaceous plants pollen grains of heliophytes occur, i.e. *Artemisia* in large amounts and *Ephedra* t. *distachya*. Indicator plants include *Potamogeton praelongus* and *Rumex maritimus*. Their presence appears to indicate a continental climate characteristic of this stadial cooling but excluding Arctic tundra. *Potamogeton praelongus* is a species with an Euro—Siberian and North—American distribution. It grows in Asia within a belt at 50–65° latitude. It has been found many times in northwestern Poland as a constituent of floras displaying Sub—Arctic climatic characteristics (Tobolski 1986). The Euro—Asian range of *Rumex maritimus* is also large.

The earliest interstadial is conspicuously bipartite. In the profile for Władysławów it represents a succession culminating in a Boreal pine forest with some spruce

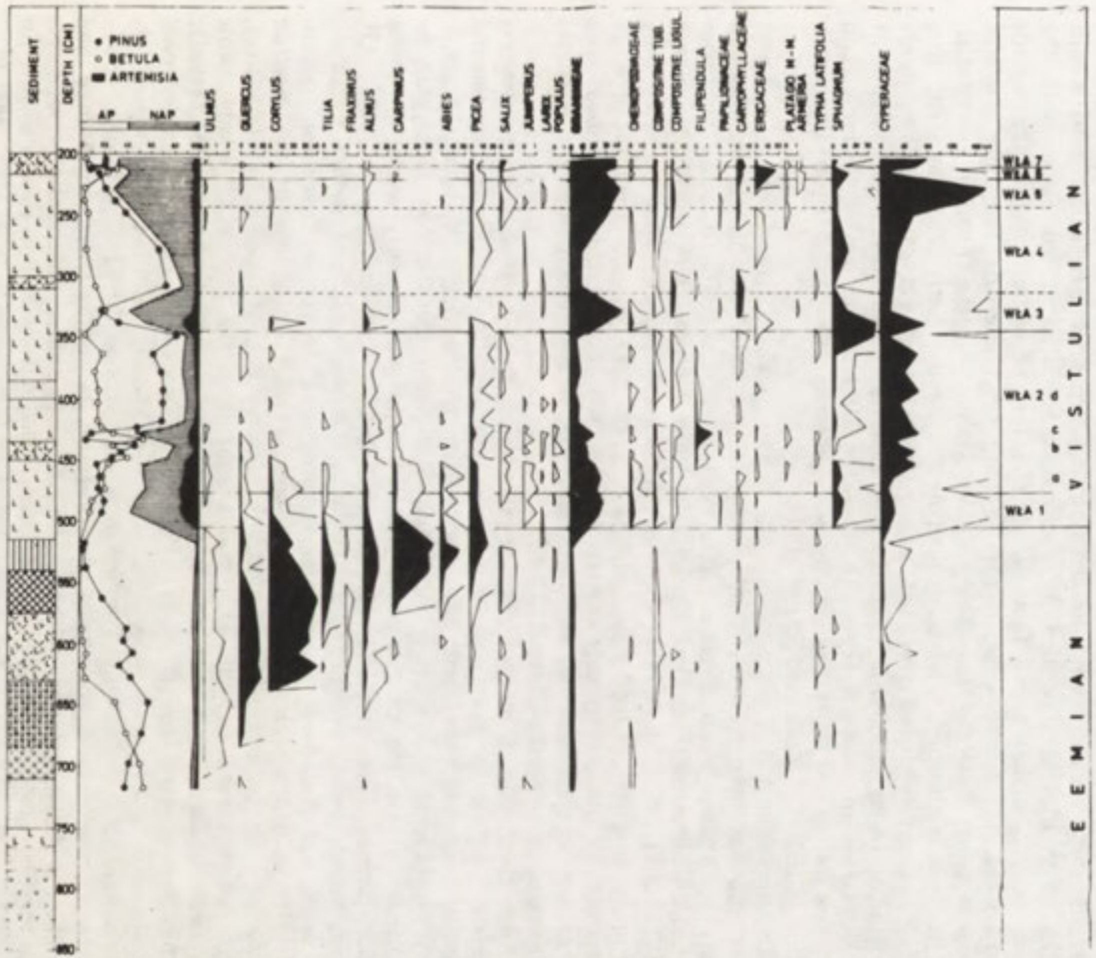


Fig. 2. Simplified percentage pollen diagram from Władysławów

and alder. This floristically diverse interstadial is correlated here with the Brørup Interstadial, as suggested by Menke and Behre (1973), and Menke and Tynni (1984). Emphasis should be put on the continental characteristics of this interstadial. During this period, a marked zonation of forest distribution occurred in middle and western Europe. Menke and Tynni (1984, p. 81) as well as Behre and Lade (1986) give attention to this. If this important palaeophytogeographical fact is ignored, estimates of age of particular fossil floras may be erroneous.

The Early Vistulian flora from Rudunki, i.e. from an area adjacent to Konin in the Łódź Plateau described by Jastrzębska-Mamełka (1985) agrees with the above concept of the Brørup Interstadial. The flora of interest at this site records in detail events that took place over a long time interval. The base contains deposits dating back to the close of the Saalian, to the Eemian Interglacial, and to two interstadials and three stadials during the Early Vistulian. The first Early Vistulian Interstadial is referred by Jastrzębska-Mamełka (1985) as the Amersfoort Interstadial. In my opinion it is bipartite, because traces of cold climatic oscillations are detectable in samples 42 to 45 from the Rudunki II profile. I am certain that this oscillation correlates strictly with horizon WF IIa 2 from Rederstall, as recognized by Menke and Tynni (1984).

The next interstadial younger than the Brørup is also characterized by a forest cover and has been recognized in the profile for Władysławów. It is separated from the Brørup by a treeless phase of stadial rank. Up to now the upper organogenic horizon from Stare Kurowo, i.e. Stare Kurowo II (Kozarski, Nowaczyk and Tobolski 1982) comparable with the Odderade Interstadial, has been the only site in northwestern Poland that contains a forest flora younger than the Brørup. The Rudunki site corresponds closely with the second interstadial described from the profile for Władysławów. This forest interstiadial has been termed the Rudunki Interstadial (Jastrzębska-Mamełka 1985).

After the predominance of pine forests during the interstadial period treeless vegatation appeared again. In that phase climate did not display continental characteristics, compared with the two earlier treeless Early Vistulian stages.

A vegetational oscillation possibly due to different climatic conditions, can be recognized at the top of the profile for Władysławów, as early as in a treeless stadial phase. This oscillation is demonstrated by an increase to 40% AP, among which *Betula* pollen prevails, including *Betula nana* and *Betula t. "alba"*. *Ericaceae* and *Sphagnum* expand, while there is a decline in *Gramineae* pollen. This change in the sporomorph composition provides indications of the presence of *Betula-nana*-containing shurb tundra with some contribution from birch trees, maybe birch shrubs as well. This pollen zone from the profile at Władysławów has been labelled WŁA 6 (Tobolski 1986). The optimum phase from Maliniec I may be its correlate (Tobolski 1984). This oscillation which displays cool interstadial characteristics is equated with the Middle Vistulian. The mean July temperature for the Maliniec I optimum is estimated as 10–11°C (Tobolski 1984). The Keller Interstadial could be the corresponding interstadial in Menke and Tynni's (1984) stratigraphic scheme. The Middle Vistulian Interstadial also agrees with the Oerel Interstadial (Behre and Lade 1986). A radiocarbon date available for a sample taken 50 cm above the interstadial at Władysławów is 44 800 + 5000 to − 3000 yr BP. An organic fraction soluble in NaOH has been dated. The control date applicable to the whole organic material is much later. This implies that the organic matter contains an admixture of younger substances and thus the fraction soluble in NaOH may be of older age.

It appears that the flora Maliniec II which has been found beneath a bed of the only Vistulian boulder clay is of stratigraphic significance. The radiocarbon dates available for it are 22 050 ± 450 and 22 230 ± 480 yr BP. Its characteristics are a dominance of herbaceous plants, especially *Gramineae* and *Cyperaceae*, and small amounts of *Artemisia* and other heliophytes. Among the macrofossils, sedges as well as *Drepanocladus revolvens*, *Calliergon turgescens*, and *C. trifarium* occur in large amounts. The oldest radiocarbon dates given by sediments that overlie a bed of boulder clay near Konin are 12 980 ± 130 and 11 880 ± 180 yr BP (Borówko-Dłużakowa 1969).

## CORRELATION TABLE

Comparisions have been made between the most important data on Vistulian floras near Konin and similar biostratigraphic information available from nortwestern Poland and selected sites in lowland Central Europe.

Figure 3 shows the correlations of the Władysławów, Maliniec and Stare Kurowo sites in northwestern Poland with the Rudunki site in middle Poland. These sites are compared with the Behre and Lade division system (1986).

| N-W POLAND | | | CENTRAL POLAND | CORRELATION Behre & Lade 1986 |
|---|---|---|---|---|
| WŁADYSŁAWÓW | MALINIEC I | STARE KUROWO | RUDUNKI | |
| WŁA-7 | | | | Ebersdorf-Stadial |
| WŁA-6 | D C B | | | Oerel-Interstadial |
| WŁA-5 | A | | VS 3 | Schalkholz-Stadial |
| WŁA-4 | | STARE KUROWO II | Rudunki | Odderade-Interstadial |
| WŁA-3 | | E | VS 2 | Rederstall-Stadial |
| WŁA-2 d c b a | | D C B | Amersfoort b a | Brörup-Interstadial |
| WŁA-1 | | | VS 1 | Herning-Stadial |
| Eemian | | | Eemian | Eem-Warmzeit |

Fig. 3. Correlation table

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