ISSN 0067-3064
ISBN 9986-615-05-4

LIETUVOS MOKSLŲ AKADEMIJA
GEOLOGIJOS INSTITUTAS

Baltica

An international yearbook
of Baltic Sea geology,
geomorphology and paleogeography

Volume 9

Vilnius ACADEMIA 1996

BALTICA An International Yearbook for Geology, Geomorphology and Palaeogeography of the Baltic Sea

ISSUED ANNUALY IN COOPERATION WITH THE GEOLOGISTS OF THE CIRCUM-BALTIC STATES

SPONSORED by The Science Council of Lithuania.

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INDEXING * BALTICA is at a present time included in the international abstracting and indexing services and databases: Bibliography and Index of Geology and GeoRef, Current Geographical Publications, Petroleum Abstracts, and Ulrich's International Periodicals Directory.

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SUBSCRIPTION PRICE * The 1996 subscription rate of volume 9 (airmail postage included) : USD 8.00; Baltic countries only: LTL 4.00, EEK 12.00, LAT 0.5. Communications regarding subscription should be sent to the Editor's office.

This Yearbook can be ordered at the Lithuanian Institute of Geology, T. Ševčenkos 13, 2600 Vilnius, Lithuania.

ISSN 0067-3064

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Editor's Address to BALTICA Readers

EDITORIAL

"Navigare necesse est" – with such wishes Academician Vytautas Gudelis saw the BALTICA Yearbook off already twice: in 1963, when this publication was born in the scientific world, and in 1994, when he passed to me the honour to renew BALTICA's publication from its volume 8.

BALTICA's way to the family of research publications is not easy. I think, firstly, that Baltic Sea investigators are insufficiently informed about the rebirth of this Yearbook. Secondly, actively organized conferences and various international projects publish their own special editions "collecting" quite a few of research results. Thirdly, distribution of the BALTICA yearbook and its presentation via international information systems is not well enough organized. But, these are not very large problems, they can be and will be gradually solved.

Thus, BALTICA came back, BALTICA again took its road. The nearest future will show how this road looks. BALTICA would like to join the family of such NORDIC journals as BOREAS, LETHAIA, GFF etc. Volume 8 of BALTICA has already shown its direction – serious scientific papers and information typical of a yearbook at the same time; such is the credo of BALTICA. Studies of the Baltic Sea and its shores make up a field where successful development of BALTICA is quite real.

The success of BALTICA fully depends on how much new papers and information is being received. Therefore, Editor expects favour of authors, grace of readers and attention of publishers.

Traditions of BALTICA created by Vytautas Gudelis can be continued. Please, write and send articles. BALTICA's papercase is open all the time.

On this occasion I would like to express my gratitude to the authors of BALTICA volumes 8 and 9 and, with a great pleasure, to the manuscript referees whose assistance was really of great value. I expect that BALTICA's sponsor – The Science Council of Lithuania – and supporters – Lithuanian Academy of Sciences and Institute of Geology will help us further as before, that this publication could reach readers each year.

Algimantas Grigelis

Chemical Components and Elements in the Suspended Matter and Sediments of the Western Baltic

Emelyan M. Emelyanov

Emelyanov, E.M. 1996: Chemical components and elements in the suspended matter and sediments of the Western Baltic. *Baltica, Vol. 9*, pp. 5–15. Vilnius. ISBN 9986–615–05–4 / ISSN 0067–3064

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Chemical components ($CaCO_3$, SiO_{2am}) and elements (C_{org}, Ca, Mg, P, Fe, Mn, Ti, K, Na, Rb, Li, Zr, Ba, Zn, Cu, Cr, Ni, Co, V, Mo, Sn) have been investigated in the surface and Late Quaternary sediments of the Bornholm Basin, Fe, Mn, Zn and Cu – in the suspended matter. Some sediment samples were also investigated for Sr, B, Sc, Cs, Hf, Ta, Th, La, Ce, Nd, Sm, Eu, Tb, Yb, Lu, U, Au. The behaviour of the elements in the water column and bottom sediments is discussed.

Keywords: Baltic Sea, Arkona, Bornholm, sediments, sedimentation, geochemistry, sapropels, manganese, rare elements, suspended matter.

INTRODUCTION

The Baltic Sea is an extremely interesting area from a geochemical point of view. Periodical stagnation of the near-bottom waters in the deep basins causes sedimentation in these deeps of sapropel-like muds, which are enriched not only in organic matter (C_{org}), but also in Mn, P, Fe, Mo, and in some cases, Cu, Zn, Ni, Ba. V.T.Gorshkova (1960), A.Blazhchishin (1976), and E. Emelyanov (1976) compiled maps for the Baltic proper showing areas with highest concentrations of $CaCO_3$, C_{org}, Fe, Mn, P, Cr, Ni, V, Ba, Zr, Mo, Sn in surface sediments. Later some of these maps were improved (Emelyanov 1988,1992, 1995). The distribution of chemical components and elements in the sediments of the Western Baltic was shown fragmentarily in the published maps. During the last decade hundreds of new samples of bottom sediments were collected and analysed (our unpublished data, data of L.Brügmann & D.Lange 1990, and M.Pertilla & L.Brügmann 1992).

The main purpose of this paper is, on the one hand, to show (1) the influence of the land (terrigenous input) on the chemical composition of the sea water, suspended matter and bottom sediments of the Bornholm and the Arkona Basins, and (2) the influence of stagnant environment in the Bornholm Basin to the chemical composition of the suspended matter and bottom sediments. On the other hand, author should like to compare the average contents of elements (1) in Late Glacial and recent sediments, and (2) in the Western Baltic and other parts of this sea.

SAMPLING AND ANALYTICAL PROCEDURES

Sediment samples were collected mainly during Soviet expeditions (1970–1991) on R/V "Professor Dobrynin", "Shelf", "Akademik Kurchatov", "Akademik Sergey Vavilov", and "Professor Shtokman". Surface sediment samples were collected by grab "Okean-25", cores were taken by simple light gravity corer (diameter 72 mm), heavy gravity corer (diameter 127 mm) and Niemistö corer. Sediment samples were collected from 106 stations (48 cores) in the Bornholm Basin and 43 stations (15 cores) in the Arkona Basin (Emelyanov 1995). The longest core in the Arkona Basin exceeds 4.70 m, in the Bornholm Basin it is 7.47 m long. The cores consist of Boreal sediments in the Arkona Basin and of tills in the Bornholm Basin (station AK–1414). In wet samples the grain size distribution was analysed by sieving and pipette methods, and the sediments were classified according to the classification of the Institute of Oceanology, USSR (Bezrukov & Lisitzin 1960). For chemical analyses samples were dried (50–70°C), pounded and analyzed. Carbon (total and organic) was analysed by a volumetric method on a Leco analyser AN 7529, phosphorus and titanium – by colorimetric method, amorphous silica – in sodium carbonate extraction, Fe, Mn, K, Na, Li, Rb, Ca, Mg, Zn, Cu, Cr, Ni, Co – by atomic absorption spectroscopy (Khandros & Shaidurov 1980), Ba, V, Zr, Mo, Be, Ge, Sn – by emission spectroscopy (Meishtas 1970). Soviet (Russian) state geological standards SDO-1, SDO-2 and SDO-3 were used for control. All analyses were perfomed in the Atlantic Geological Laboratory of the P.P.Shirshov Institute of Oceanology (Atlantic Branch) of the Russian Academy of Sciences, Kaliningrad. Some rare microelements and the rare earth elements (REE) were analysed by neutron activation. These results were given to the author by G.Baturin.

The water sampling was carried out by using 5 l plastic samplers at drift stations and with Niskin samplers attached to a hydrochemical sonde "Rozett". Water from the samplers was filled into polyethylene bottles and filtered immediately using pressure filtration through precleaned polycarbonate 0.5 mm nuclepore filters. The filters were soaked for 3 days in cold 6 mol/l HCL, rinsed again soaked for 3 days in 1% HCl and twice distilled at 50°C. Filters with suspension were successive by treated with acid (ultrapure HF, HNO_3 and HCl) in teflon dishes, heated and poured into special 10 ml test tubes for final concentration in solution less 2% HCl. The content of elements in the suspended matter was analyzed by atomic absorption spectroscopy.

HYDROGRAPHIC CONDITIONS

The Arkona and Bornholm Basins are located in the Western Baltic near the Danish straits (Fig. 1). The maximum depth is 51 m in the first basin and 105 m in the second one. The North Sea saline waters reach first the Arkona Basin and enter the Bornholm Basin through the Bornholm gate. There is a well-expressed halocline. Its level is located between 20-25 m in the Arkona Basin and 55–65 m in the Bornholm Basin. A hydrogene sulphide contamination of near-bottom waters was observed in the Bornholm Basin only. During our expeditions the O_2–H_2S interface (strongest redox-barrier) was found at a depth of 72–75 m in many cases. During our expeditions the deepest part of Bornholm Basin was under stagnant conditions.

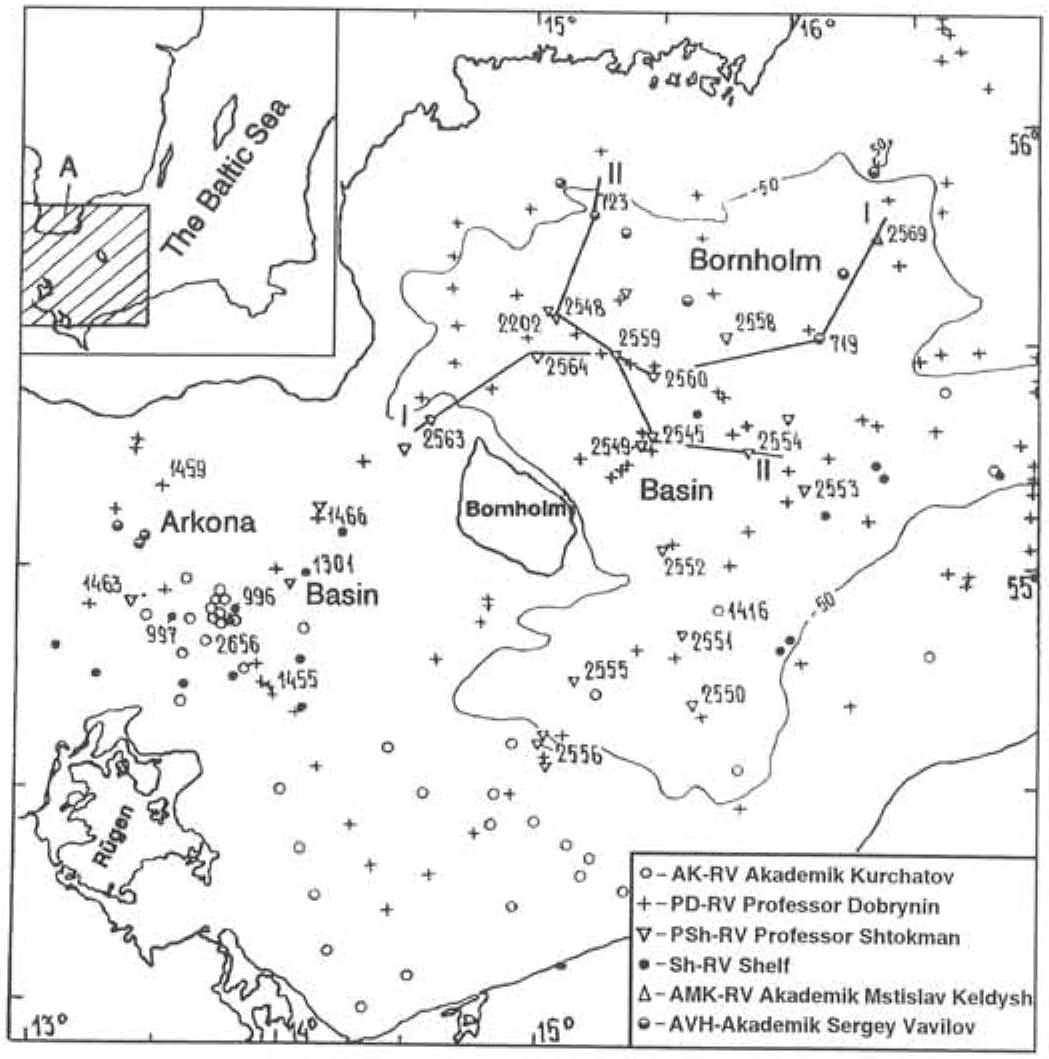

Fig. 1. Map of geological stations perfomed on board of Soviet vessels. A – studied area. Isobath is drawn in m, I–I and II–II – lines of geochemical transects (see Figs. 3 and 4).

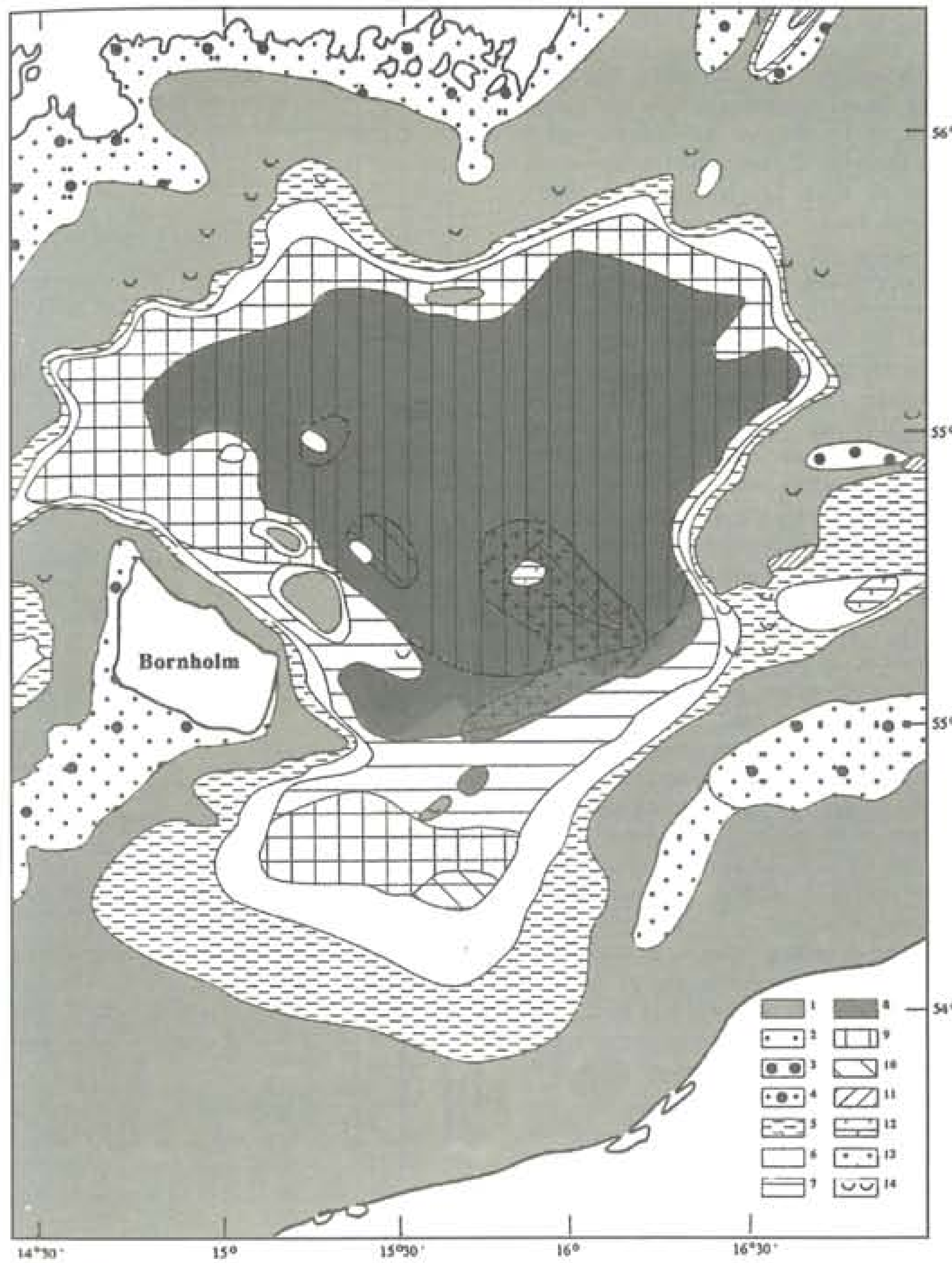

Fig. 2. Map of bottom sediments (0–3 cm layer), Bornholm Basin. The distribution of sediments is shown according to the Institute of Oceanology classification (Bezrukov & Lisitzin, 1960). 1 – fine-grained sand, quartzose; 2 – medium and coarse-grained sand; 3 – gravelly sediments; 4 – various-grained sand with gravel, pebble and boulders; 5 – coarse aleurite; 6 – fine-aleuritic mud; 7 – aleuritic-pelitic mud; 8 – pelitic mud; 9 – sapropelic mud (or sapropel-like mud, 3–5% C_{org}); 10 – sapropel mud, >5% of C_{org}; 11 – subcrops of till and clay; 12 – low calcareous sediments (10–20% $CaCO_3$); 13 – low manganese mud (>0.2% Mn, up to 0.40% Mn); 14 – presence of Fe–Mn concretions and crusts.

SEDIMENT TYPES AND THEIR COMPOSITION

In all cases sediments were classified by lithostratigraphy and faunal contents (benthic and planktonic foraminifera), by absolute age determination (^{14}C, and some of them by diatom and pollen analyses (Kuptsov et al. 1984; Emelyanov et al. 1989).

In the Arkona Basin, sands are widespread in the nearshore areas. At a depth of 30–40 m they are replaced by silt and fine aleuritic (fine silty) mud and aleuritic-pelitic mud (with a content of 50–70% of the fraction < 0.01 mm). Only some spots in the central part of the Arkona Basin have a pelitic (clayey) mud with a content of 70–75% of the fraction < 0.01 mm (Fig. 2). The Holocene marine muds of the Bornholm Deep are accumulated beneath the halocline, where the environment is calm, sands occur above the halocline (Emelyanov, 1995), coarse and fine aleurites (silts) are at the level of the halocline or just beneath it. In the Arkona Basin with depths lower than of the Bornholm Deep, the sediments are not so well-sorted: the inluence of the halocline in this basin on the sediment distribution is much less than in the Bornholm Deep. The mud thickness in the Bornholm Deep is 280–300 cm according to the coring results. However, according to CSP data, it may exceed this value. The thickness of mud presumably reaches 8–10 m in some places. Here, accumulation rate exceeds 0.3 mm per year, and at some stations (station PSh-2564) it reaches even 1.0 mm per year. It seems to be a characteristic feature that, in the deep-sea areas of the Bornholm Basin, muds are either completely absent (stations PSh-2551, PD-2202) or considerably less thick. One may assume that strong near-bottom currents preventing accumulation of large masses of muddy material occur in these regions.

The sediments of the Bornholm and the Arkona Basins are terrigenous: biogenic and chemogenic components do not exceed 10–15% of the total sample. Sands and silts consist mainly of quartz. Content of SiO_2 (total) exceeds 93.70% (Blazhchishin 1985). Muds are slightly reduced – the Eh value is usually negative and reaches – 135 mV. The muds contain 46.92–53.63% SiO_2 (total) and 12.51–16.67% Al_2O_3. They are enriched in C_{org} (up to 7.31% in Bornholm Basin, up to 5.13% in Arkona Basin), P_2O_5, and sometimes in MnO_2. The mean contents (Tables 1 and 2) of $CaCO_3$ and Ca are decreasing, C_{org}, Zn and Cu are increasing from till to sulphidic clay (Emelyanov 1995). There are more Fe, Zn, Cu, K and Na in sulphidic clay than in homogeneous (grey or brown) clays accumulated beneath the sulphidic clay .

Unlike tills and Late Glacial clays, the Holocene sediments contain distinctly increased amounts of

Table 1. **Chemical composition of the sapropel-like muds of the Arkona Basin (with content of C_{org} more than 4 %)**

Station	Depth, m	Horizon, cm	Sediment type	Fraction <0.01 mm, %	%											10^{-4} %						
					$CaCO_3$	SiO_{2am}	C_{org}	P	Fe	Mn	Ti	K	Na	Ca	Mg	Zn	Cu	Cr	Ni	Co	Rb	Li
PD-1455	35	0–20	C.a.	10.3	1.84	2.40	7.75	0.09	4.33	0.02	0.38	–	–	–	–	–	–	44	10	–	–	–
PD-1457	42	0–25	A.p.m.	62.2	0.84	2.42	4.76	0.07	3.58	0.05	0.39	–	–	–	–	–	–	93	32	–	–	–
PD-1463	44	0–25	A.p.m.	61.8	0.75	1.81	4.08	0.06	4.52	0.03	0.40	–	–	–	–	–	–	96	50	–	–	–
PD-1466	47	0–25	A.p.m.	56.2	0.66	3.20	4.30	0.06	3.22	0.05	0.32	–	–	–	–	–	–	91	27	–	–	–
AK-2656	47	0–5	F.a.m.	–	0.00	–	4.44	–	3.60	0.03	–	–	–	–	–	–	36	52	40	–	–	–
		120–125	F.a.m.	–	1.75	–	5.13	–	3.20	0.03	0.47	–	–	–	–	68	26	42	47	–	–	–
		160–165	F.a.m.	–	2.75	–	4.30	–	3.79	0.05	0.41	1.90	1.34	–	–	78	29	70	57	–	89	–
Sh-996	46	25–30	A.p.m.	53.3	0.33	–	4.08	–	9.98	0.04	0.45	1.90	1.67	1.00	1.13	148	38	88	52	26	88	42
		40–45	A.p.m.	5.54	2.20	–	4.82	–	3.65	0.05	0.43	2.08	1.47	0.89	1.28	52	41	74	56	20	62	38
Sh-997	47	0–5	P.m.	–	1.25	–	4.34	0.08	3.86	0.04	0.39	2.48	2.06	0.73	1.22	60	40	74	68	34	66	36
		40–45	A.p.m.	58.8	15.91	–	4.70	–	4.28	0.03	0.42	2.06	1.91	0.96	1.20	136	38	84	78	34	102	42
		80–85	A.p.m.	–	15.50	–	4.90	–	5.16	0.05	0.43	2.06	1.62	0.61	1.34	59	38	62	78	26	72	38
Sh-1301	47	0–3	A.p.m.	62.3	14.55	–	4.75	–	3.50	0.03	0.41	2.24	2.08	–(?)	1.51	129	37	82	86	–	119	40
Average					5.03	2.46	4.57	0.07	3.86	0.04	0.41	2.10	1.74	0.68	1.28	91	32	67	52	28	85	39

Sediment type: C.a. – coarse aleurite, F.a.m. – fine aleuritic mud, A.p.m. – aleuritic-pelitic mud, P.m. – pelitic mud

Table 2. **Maximum (for Hl2–3) and average content of chemical elements in the sapropelic-like (3–5% of C_{org}) and sapropelic (>5% of C_{org}) mud of the Bornholm Basin.** (Note: $CaCO_3$–Mg in %, otherwise in 10^{-4}%)

Component, element	Maximum	Average (0–3 cm)	Shales (clay)[2]
$CaCO_3$	19.90	3.70	–
SiO_{2am}	3.53	1.77	–
C_{org}	7.75	4.29	–
P	0.10	0.08	0.08
Al	7.83	7.29	–
Fe	7.31	4.54	3.33
Mn	0.64	0.14	0.07
Ti	0.53	0.42	0.45
K	3.16	2.41	2.28
Na	2.62	1.90	0.66
Ca	2.09	0.66	–
Mg	1.96	0.29	–
Rb	239	141	200
Li	78	47	50
Ba	2300	846	800
Sr	170	150[1]	450
Cu	343	44	57
Zn	268	148	95
Cr	134	77	100
Ni	88	57	95
Co	64	32	20
V	220	134	130
Mo	28	9	2
Pb	101	86	20
Zr	520	214	200
Sn	101	42	10
B	110	168	100
Sc	51	171	10
As	28	24	10
Sb	26.70	4.10	1.50
Cs	12.40	8.10	12.00
Hf	5.10	4.40	6.00
Ta	1.70	1.20	–
Th	22	14	11
La	58	54	–
Ce	127	90	–
Nd	65	47	–
Sm	8.90	7.00	–
Eu	1.60	1.50	–
Tb	1.50	1.30	–
Yb	6.00	4.90	–
Lu	0.60	0.50	–
U	4.90	4.30	–
Au	0.01	–	–

[1] The microelements Sr, Sc, As, Sb, Cs, Hf, Ta, Th, La, Ce, Nd, Sm, Eu, Tb, Yb, Lu, U are given after the data of G.Baturin and E.Emelyanov (9 samples) (Emelyanov 1995).

[2] Shales (clay of the land area) are given as an average, according to A.P.Vinogradov (1962).

C_{org}, sometimes Mn, obviously increased amounts of P, Na, Co, and decreased amounts of K, Rb, $CaCO_3$ in the marine Holocene (Hl2–3) mud (Tables 1 and 2). The rest of the elements is observed in the same amounts as in the Late Glacial clays. The maximum amounts of elements and components are usually found in the sedimentary strata. In the upper layer (0–3 cm) there are maxima for SiO_{2am}, Zn, Ba, Rb only. The maximum content exceed the minimum ones in many times for some elements and 1.5–3 times over the average content.

The elements and components in the upper sediment layer (0–3 cm) are distributed in different ways throughout the main granulometric sediment types (sand – coarse aleurite – fine aleuritic – aleuritic/pelitic – pelitic muds). According to the average content (Emelyanov 1995), they can be grouped as follows: (1) SiO_2 total and Sn; (2) $CaCO_3$, Zr, W; (3) Cr, Zn, Cu, Mo, SiO_{2am}, C_{org}; (4) Al, P, K, Na, Fe, Ti, Rb, Ba, V, B, (Co, Li); (5) Mn and Ni. The elements of the first group are represented by minerals of the sandy (1–0.1 mm) and silty (0.1–0.01 mm) fractions: quartz and feldspar prevail among clastic minerals in the fraction > 0.01 mm. According to granulometric sediment types the distribution of SiO_{2total} and Sn is similar to the quartz distribution in the bulk samples: the maximum in the sands (or in shallow water areas of the basins at the map), the minimum – in pelitic muds (hence, in the deep areas of the basins) (Emelyanov 1986).

Calcium carbonate, Zr and W are also associated with clastic minerals (Emelyanov 1976, 1986; Pilipchuk & Emelyanov 1979) in the aleuritic fraction (0.1–0.01 mm) but not in the sandy one (1–0.1 mm). So, their maximum average contents are confined to aleurites related to the slopes of the basin and bottom depressions in the Bornholm Basin. Zirconium as a typical element-hydrolysate is present in sediments mainly with ilmenite, titanomagnetite, biotite, and amphiboles. Tungsten in sediments has a clear positive correlation with clastic iron (that part of Fe which is in the crystalline lattice of minerals) (Pilipchuk & Emelyanov 1979). Positively charged colloids of ferric hydrate well adsorb anions, including $(WO)^{2-}$. In the Bornholm Basin this process should occur in the sediments above the O_2 – H_2S boundary, i.e. within the oxid zone where aleuritic-pelitic muds are replaced by fine-aleuritic ones. This explains why W is confined to the periphery of the basin and to aleuritic sediments.

The elements of the third group are present in aleuritic/pelitic muds. The sediment distribution map shows this group corresponds either to the periphery of the Bornholm Deep floor (see Cr and Zn) or to its central part (see Cu). Chromium, as one of low-mobile elements, is represented by fine-dispersed (0.01–0.001 mm) clastic minerals (including clay minerals) in the marine sediments.

The largest amount of the element is found in aleuritic-pelitic muds. The elements Zn, Cu and Mo, on the one hand, are found in the same minerals as Cr, and, on the other hand, they are found in authigenic minerals (hydrous ferric oxides, sulphides) and organic detritus. The latter ones are also concentrated mainly in the fraction of 0.05–0.001 mm. Amorphous silica and C_{org} are being grouped together with Cr, Zn, Cu and Mo, but they are not genetically related to them. These two components are represented in sediments by light biogenic components – diatom skeletons (SiO_{2am}), spore and pollen grains, fragments of algae tissue. All these particles are found usually with grain size fraction of 0.05–0.001 mm. They are accumulating mainly in the deep part of the basin. Besides, C_{org} is sorbed on clayey particles.

The fourth group of elements is closely connected with the pelite fraction <0.01 mm (with clayey and subcolloidal particles, mainly). These elements are distributed in accordance with the rule of the pelitic (<0.01 mm) fraction (Emelyanov 1986, 1982, 1995): the higher the content of <0.01 mm fraction the higher is the element content. The map shows maxima of element concentrations located in the central areas of the basin. Marine Holocene muds are sometimes enriched in barium, vanadium, molybdenum and lead. The map shows barium maximum located in the peripheral area of the Bornholm Deep, i.e. in the transitional environment (between stagnant and oxic conditions) (Emelyanov 1976). This element occurs in the mud usually in a form of $BaSO_4$ (barite) (Blazhchishin 1976). Vanadium and molybdenum (and probably lead) are attracted towards the central area of the basin taking them closer to manganese. The concentrations of boron in the Baltic Sea sediments (and in the Bornholm Deep sediments as well) are close to the Clarke values (Sukhorukov & Emelyanov 1969), i.e. they are typical for sedimentary rocks: a low boron content (20–35 ppm) is observed in sands and aleurites and a high content – in pelitic muds (150–180 ppm). Boron is connected mainly with illite in the Baltic Sea sediments.

Mn and Ni stand apart from the elements of the fourth group, though in other regions of the Baltic Sea (Emelyanov 1986, 1987) they strictly follow the rule of the pelitic fraction. This could be explained by the fact that the major part of Mn in the sediments is contained not in clastic but in fine authigenic minerals (Manheim 1961; Hartmann 1964; Emelyanov et al. 1982). These are either hydroxides of Fe and Mn (flakes and hydroxide particles of irregular form, micronodules) or fine dispersed carbonates of Mn which predetermine increased Mn content in the sediments (Emelyanov 1981). Manganese carbonates are found in higher amounts in the pelitic muds of Gotland, Fårö, and Landsort Deeps (Emelyanov et al. 1982). Nickel is associated with manganese. Copper is sorbed by manganese hydroxides and iron hydroxide, therefore, according to granulometric sediment types, its distribution is similar to that of Mn and Fe.

Marine Holocene (Hl2–3) sediments in the cores contain approximately the same amounts of chemical components and elements as the 0–3 cm layer of corresponding sediment types. The maxima for C_{org}, Mn, K, Li, Cu, Cr, Co and Mo are displaced from the top of the cores down to the sedimentary strata . In the Holocene muds Ni, Cu, Zn, P, Rb, Fe, K coincide with Al; whereas Rb, P, Mn, Ti and Ni coincide with Fe and Al; whereas Ti, Na and Li – with K. As mentioned above, sapropelic (5–10% C_{org}) and sapropel-like (3–5% C_{org}) muds are characteristic for the Bornholm Deep, as well as for many other Baltic Sea basins.

DISCUSSION

Thus, on the basis of data from the Bornholm, Arkona and other deeps (Gdańsk and Gotland basins) of the Baltic Sea one may conclude that the mechanical separation of the sedimentary material during transportation in the sea is the main factor controlling the differentiation of the chemical elements in the Baltic Sea and the formation of different geochemical facies: the shallow (or silicious) facies and the deep (or alumosilicate-sapropelic) facies.

The halocline is the main depth boundary between these two facies. Another conclusion concerns the chemical separation of the elements. This separation begins in the water column, especially in the basins with hydrogene sulphide (Emelyanov 1981, 1982, 1986; Kremling 1983; Kravtzov 1994). In the Bornholm basin the minimum amount of iron in the suspended matter is found in the upper water layer (Figs. 3 and 4). At the depths of 20–50 m its content increases and within the halocline (or below it) decreases again. The maximum Fe content was observed within the transitional layer O_2–H_2S (between oxic and anoxic conditions). In the layer with an oxygen content of 1 ml/l, iron concentration reaches its maximum (10.0%). In stagnant water, the Fe content in the suspension decreases. If we consider individual layers in the upper hydrodynamically active water layer (0–1 m), the Fe content ranges within 0.48–4.85%, in the O_2–H_2S layer – within 1.7–10.0% and in the near-bottom layer – within 1.67–10.0%.

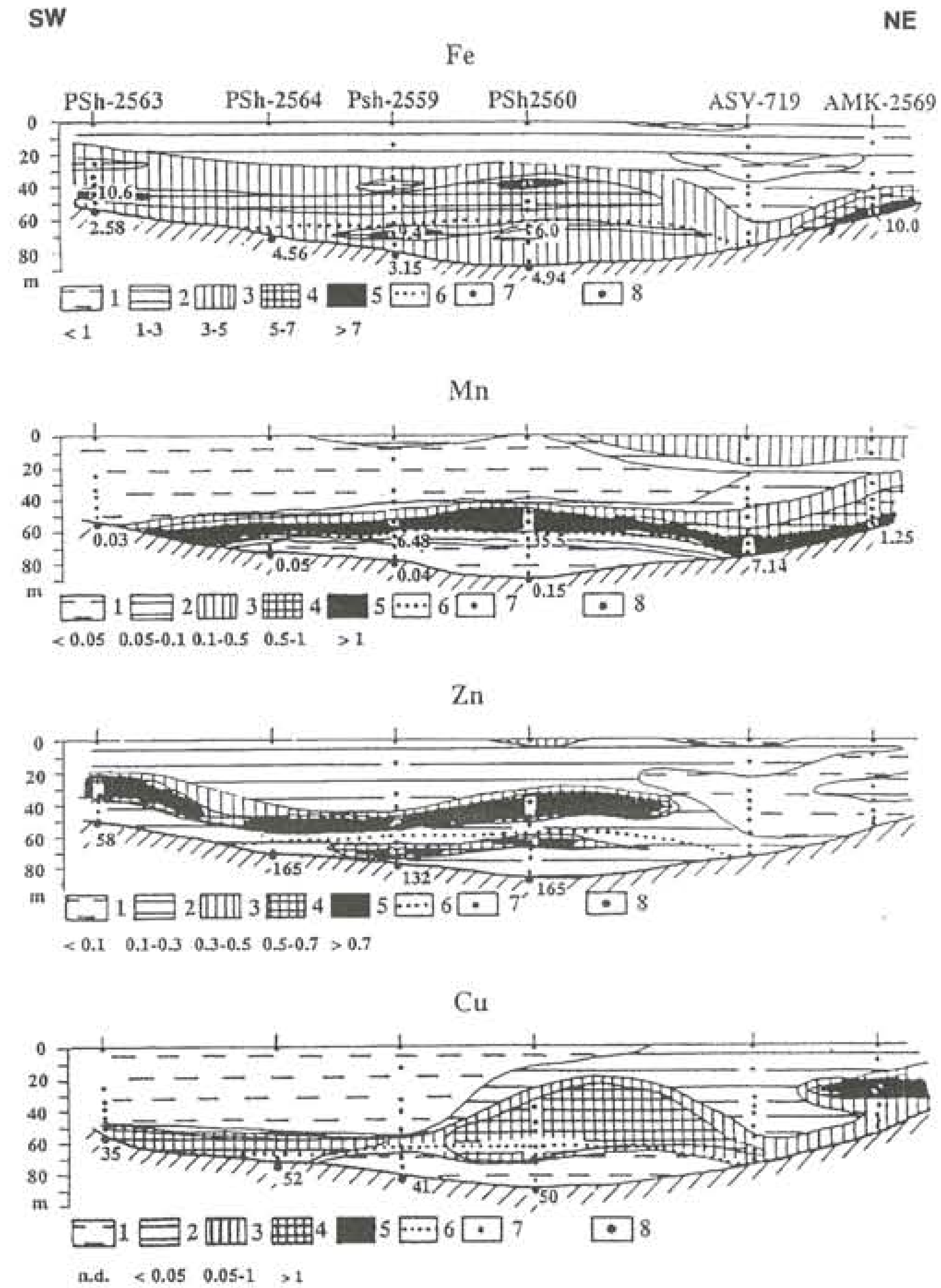

Fig. 3. Distribution of Fe, Mn, Zn and Cu in particulate matter of water from the Bornholm Basin based on several stations along transect I–I (see Fig. 1). Isolines are also drawn manually. All data are given in per cents. 1 to 5 are contents (n.d. not detected); 6 – isoxygen 0.7 ml/l; 7 – sample sites (maximum content is also indicated); 8 – surface sediment contents (Note: Zn and Cu are given in ppm).

A more complicated picture is observed for Mn distribution in the suspension (Emelyanov & Stryuk 1981). In the upper water layer (0–1 m), two areas of suspension enrichment in Mn can be outlined (up to 0.17 and 0.40% respectively). At the depths of about 60–70 m, there is a layer (lens) with a maximum Mn content in suspension for all stations studied (Figs. 3 and 4).

Within this layer Mn content reaches 35.5% generally. A layer of the sharply increased (>1%) and maximum (20–35.5%) Mn content is confined strictly to the isoxygene of 1 ml/l. Usually, a layer of maximum Mn content in the suspended matter is located several metres above the analogous layer for Fe. The reasons of suspension enrichment in metals are related to the chemical properties of

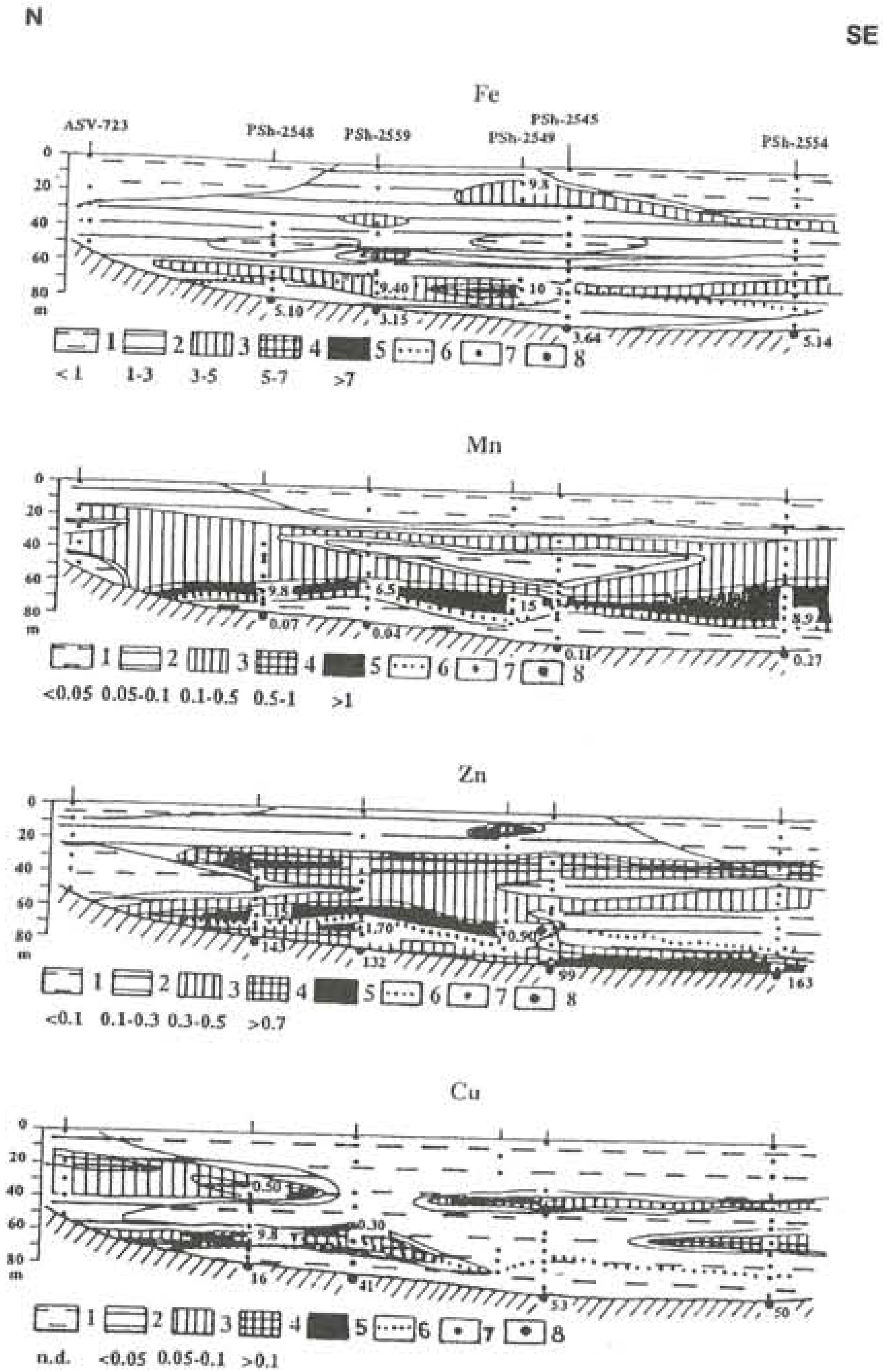

Fig. 4. Distribution of Fe, Mn, Zn and Cu in particulate matter of waters from the Bornholm Basin based on several stations along the transect II–II (see Fig. 1). The legend is given in Fig. 3.

the elements, discussed in previous publications (Emelyanov 1981, 1982, 1986, 1995; Kremling 1983; Kravtzov 1994).

In the near-bottom layer, the areas of high Mn content in the suspension form a ring-shaped structure. This structure encircles the Bornholm Deep at a depth of 60–70 m, i.e. it is located in the place where the 1 ml/l isoxygene touches the bottom. This occurs at the foot of the slope. The gentle slope places contain the band widening, and if the slope is steep (near Christiansö Island), the band becomes significantly narrower.

Below the isoxygene of 1 ml/l, the Mn content in suspension sharply drops down to <0.05%: under stagnant conditions, manganese is mainly concentrated not in a particulate (Mn oxide), but in a dissolved form (Mn^{2+}). This explains low Mn contents in the suspended matter of the near-bottom layer in the central part of the Bornholm Deep (Figs. 3 and 4).

A complex, mottled distribution is observed for Zn and Cu in the suspension (Figs. 3 and 4). However, nearly all the profiles have an intermediate layer distinguished where the suspended matter contains increased amounts of Zn and Cu. This layer is usually confined to the halocline. Another layer of high content of both Zn an Cu is confined to the transitional layer (O_2–H_2S). But, in contrast to Mn, the increased content of Zn and Cu in suspension of the transitional layer between oxic and anoxic conditions is not constantly observed.

A comparison between Fe, Mn, Zn, and Cu contents in suspension of the near-bottom layer and bottom sediments shows, that the surface sediments contain 1.5–3 times more of Fe, 10 times more of Mn and 10 times less of Zn and Cu. The main reason of these phenomena is suggested to be the high solubility of manganese under the stagnant conditions and its (Mn^{2+}) diffusion from pore water to the near-bottom water. The solubility of iron is less than the solubility of manganese (Kremling 1983; Kravtsov 1994). The behaviour of Zn and Cu depends, in the most cases, on the behaviour of Fe and Mn. This behaviour was discussed in previous publications (Kremling 1983; Kravtzov 1994).

The mobile iron (Fe^{2+}) in the sediments is mainly in a form of iron sulphides dispersed (horizontally and vertically) in the reduced sediments of the Arkona and Bornholm basins.

In comparison with Fe, the maximum for Mn in the sediments shifted towards the centre of the Bornholm Basin. This is caused by the fact that stagnant conditions are frequently prevailing in the centre of the deep. That leads to high (maximum) accumulation of Mn that, under strongly reduced conditions, accumulates in the Bornholm Basin diagenetically in a form of manganese carbonates of a complex composition (Manheim 1961; Emelyanov 1981). According to M. Hartmann (1964), at the station E13 (depth of 68 m), 1.05% Mn (the maximum Mn content for the Bornholm Deep muds) and 4.05 % Fe were found in the layer between 0.5 and 1 cm.

The sediments (especially muds) of the Bornholm Deep contain much more $CaCO_3$, C_{org}, K, Na and Mn than the sediments of Gdańsk Basin (maximum depth 110 m) (Blazhchishin & Emelyanov 1977; Emelyanov 1986, 1987). This is explained by (1) a wide occurence of carbonate rocks in the sediment source areas adjacent to the Bornholm Basin (Rügen Island, Swedish shore) and a high content of clastic calcite in the sediments (1–14% of heavy 0.1–0.05 subfraction); (2) a higher salinity of the sea and pore waters (for Na and, partially, K); (3) longer periods of stagnation of the near-bottom water (for Mn). It is significant that the P concentration in aleurites (silts) and muds of the Bornholm Deep is appreciably less than in analogous sediment types of the Gdańsk Deep (Emelyanov 1986). The Bornholm area is apparently less affected by agricultural activity on mainland (the Bornholm Deep is farther from the mainland than the Gdańsk Deep). Moreover, the large Vistula river brings high amounts of phosphates into the Gdańsk Basin (Emelyanov 1986).

Thus, the maximum content of C_{org} as well as of SiO_{2am} is located in the central part of the Bornholm Basin, i.e. beneath the halocline (Emelyanov 1991).

Sapropelic and sapropel-like (or sapropelitic) muds in the Bornholm Deep as well as in some places of the Arkona Basin and in all the deeps of the Baltic Sea were formed during the Middle and Late Holocene (Atlantic phase, Hl2–3) (Blazhchishin 1976; Emelyanov 1981, 1986, 1988, 1992). These are terrigenous microlaminated muds of greenish- and dark-grey (to black) color, they contain up to 7.75% C_{org} and a considerable amount of carbonates (up to 19.90 % $CaCO_3$). Sapropel-like and sapropelic muds in the Bornholm Deep as in the Gdańsk and the Arkona Basins are not laminated because hydrogen-sulphide contamination in Hl2 and Hl3 was not so strong as in the Gotland, Fårö and other deeps (Emelyanov 1981). This explains, why the muds of the Bornholm Basin and, especially, in the Gdańsk (Emelyanov 1986) and Arkona Basins contain such low contents of manganese, phosphorus and microelements. Carbonaceous matter in the central parts of the Fårö, Gotland and Landsort Deeps is almost entirely represented by manganese carbonates. At the same time Mn content here is rather high (up to 12.86 %) (Emelyanov 1995). Sapropelic laminated layer regularly reaches 30–50 cm thickness and sometimes up to 1 m. They cover the bottom of the deeps in an irregular way (Emelyanov 1988). If the thickness of sapropelic muds increases, the degree of organic matter enrichment decreases. Maximum C_{org} content (6.0–9.5%) are found in a 50–80 cm thick layer of the Gotland Deep. A degree of organic matter enrichment is variable both in horizontal and vertical directions. The sapropelic and sapropel-like muds are enriched not only in Mn but also in a number of minor elements (Tables 1 and 2). The maximum and average concentrations of C_{org}, Mn, P, Mo, Se and As are increasing with the increase of the depth and the degree of stagnation, i.e. from the Arkona – Gdańsk – Bornholm Basins to the Fårö – Gotland Deeps (Emelyanov 1981). Compared with the shales and clays from the land (Table 2), sapropel-like and sapropelic muds of the Bornholm Basin are enriched in Fe, Mn, Na, Mo, Sn, B, Sc, As, Sb and depleted in Sr, Cr, Ni, Ta. Other elements are found in the same concentrations as in the shales of the land.

The formation of the sapropel-like mud took place under conditions of a warm humid climate and water transgression. Geochemical indicators (Hallberg 1974; Emelyanov 1978) testify the sapropelic mud accumulation to have occured under conditions of a hydrosulphur contamination of the near-bottom water. The sapropelic microlaminated layer is usually represented by a aleuritic-pelitic mud with 50–70% content of the fraction <0.01 mm. Organic matter enrichment is obviously the prevailing process and is not directly connected with terrigenous organics. The planktonic nature of the major part of organic matter is manifested in the character of its areal extent as well. Due to powerful inflow of ocean saline water, a stratified water column was formed in the basins, abyssal layers were enriched in nutrients. Near-bottom water enriched in nutrients ascended periodically into the photic layer in the centres of cyclonic rotation that resulted a rapid phytoplankton bloom and further fossilization of organic matter in the sediments. This process was very intensive because of small (60–70 m) depths. The deeper northern part of the Gotland Deep and the Fårö, North Baltic, and Landsort Deeps, where winter heat loss was stronger than in the Bornholm Basin, is typical for a seasonal vertical water mixing accompanied by carbonate precipitation. However, the accumulation of sediments rich in organic matter does not take place in a large scale in the Baltic Sea. They are usually restricted to small-sized basins and individual layers in the Holocene strata column. These small basins serve as effective traps not only for organic matter but also for many chemical elements (Emelyanov 1986, 1988).

CONCLUSION

One may use the Baltic Sea deeps as a recent natural model for the explanation of the origin of recent sapropelic and sapropel-like muds, ancient black clay and ancient sedimentary manganese ores formed during Oligocene in the southerm part of the Russian Plate (Emelyanov 1986; Emelyanov & Lisitzin 1986). The Bornholm Basin with slightly increased contents of manganese in the mud is considered to be the first one in the chain of those basins (deeps) leading to the ancient basins where manganese ores have developed.

ACKNOWLEDGEMENTS

This paper was supported by the Russian Academy of Sciences (RAS), the Russian Fund of Fundamental Researches (RFFR grant 93–05–08685) and Institut für Ostseeforschung Warnemünde (IOW, Germany). The staff of the Marine Geology Section of IOW and especially its head Prof. Dr. J. Harff helped me. My English language and scientific value were improved by J. Harff, L. Brügmann, C. Christiansen, W. Lemke and S. Clasen.

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Accelerated Coastal Erosion – Implications for Latvia

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Eberhards, G. & Saltupe, B. 1996: Accelerated coastal erosion – implications for Latvia. *Baltica, Vol. 9*, pp. 16–28. Vilnius. ISBN 9986-615-05-4 / ISSN 0067-3064

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The article deals with the changes of the Latvian sea coast erosion rate during the last 50 years and recent decades with the relationship to annual and extreme sea level rise, as well as with the wind direction and frequency changes. Accelerated sea coast erosion around the Latvia coastline during the last decades has increased 2–5 times in comparison with past 50 years period, and mainly it is controlled by constantly increased wind velosity and energy during storms. Beach and bluff erosion rapid growth in Latvia is related mainly to the global and regional climate change realized by changes in wind direction, velocity and frequency, as well as an accelerated rise in sea level (extreme sea level during storms), closely connected with human impact such as construction of seaports and deepening of sea-entering channels.

Keywords: coastal erosion, erosion rate, sea level, wind, climate change, sea coast of Latvia.

INTRODUCTION

Throughout the history of mankind, coastal areas – the boundary between the land and the sea – have played an important role in society and have primarily been the sites for human settlements.

Coastal zones are particularly vulnerable to the impact of nature and man and are physically very unstable. Erosion and associated loss of land is the most evident sign of this instability. Negative shoreline trends cause secondary effects that affect society through threats to human settlements, harbors, coastal recreation areas, wetlands, marches etc. These impacts are expected to increase as a result of climate change leading to a rise in sea level.

Thus, in a short- as well as in a long-term perspective, coastal erosion must be regarded as an environmental problem that deserves much attention.

Loss of land has direct and indirect consequences for coastal communities, as well as it creates a conflict between coastal users and interest groups.

A more or less widespread opinion among scientists involved in studies of coastal processes suggests that there may be an increased global erosion pattern (Bird 1987, 1990; Kaplin & Lukjanova 1992; Kaplin, Porotov & Selivanov 1992; Hanson & Lindh 1993).

This hypothesis stems from the results of the study performed by the International Geographical Union from 1972 to 1976.

A survey of coastal areas supports the notion of increased global erosion. In the U.S., currently, 68% of coastal beaches are moderately eroded and 53% are severely affected by erosion (Williams, Dodd & Gohn 1991). An increased erosion has been observed in European coastline. The French, Dutch (Lousse & Kuik 1990), Danish and Swedish coasts (Hanson & Lindh 1993) are also subject to accelerated erosion. An increased erosion has been observed also in the Black Sea. About 80% of Crimea Peninsula and 71% of Georgia coastal zone are moderately eroded, 58% of coastal areas in the Azov Sea (the Ukraine coast) are also subjected to accelerated erosion (Kaplin, Porotov & Selivanov 1992).

Global warming and the associated rise in sea level are processes of vital significance to Latvia which has the sea coast length exceeding 496 km. In 1993, a total of 150 km of the sea coast was being severely affected by erosion. In the Latvian coastal areas threatened by erosion (in a zone 20 to 50 metres wide) there are 11 towns and small coastal settlement areas with 38 residential zones, 2 cemeteries, 3 lighthouses, local roads, the sewage treatment plants in Liepāja and Ventspils (Eberhards & Saltupe 1993).

This paper describes the initial results of the investigations:

– historical topographical maps and plans, hydrological and meteorological data (sea level, wind, storms) applied to determine historical changes and erosional rates attributed to sea level rise, increased changes in storm intensity or frequency, the kinetic energy of sustained storm winds during the last 50 years;

– examples of characteristic cliff erosion rates and accelerated erosional trends in last decade, as well as the results of recent bluff erosion studies;

– short analysis of factors that influence the accelerated coastal erosion;

– an attempt made to quantify rates of erosion and to identify processes responsible for increased erosion at the site and to map the spatial distributions of high-risk coasts.

COASTAL EROSION RATE IN THE OPEN BALTIC COAST DURING THE LAST 50 YEARS

The most recent data include topographical maps and plans at various scales (1:25000, 1:10000, 1:5000 and 1:2500) and data back to 1900–1930. The important data include plans and maps from 1932/1935, 1947 and 1981/1985. The erosional rates were calculated according to the changes in location of erosional cliff top line.

Examination of charts and maps compiled in 1935–1938 and 1981–1985, as well as 1981–1985 and 1993 was done. The examination of the coastal zone development and bluff face erosion during the last decade is based on field mapping and measurements.

Total maximum coastal zone erosion and accretion during the last 50 years is shown in Fig. 1. From the 1935/1938 to 1981/1985 the shoreline generally was eroded only in some localized erosional areas, while during the last decade the interval of about 150 km on the shoreline area was critical or significantly eroded (Fig. 2).

We present here three separated coastal areas in the open Baltic Sea without direct influence of seaports, each coastline is about 10–25 km in length. The first study area is situated in SW Latvia between the Pape and Nida (Fig. 1).

During the last 48–50 years (before 1981/1985) the width of wash-out zone in the Pape-Nida area has reached 20–50 m, in the Strante-Jūrkalne and

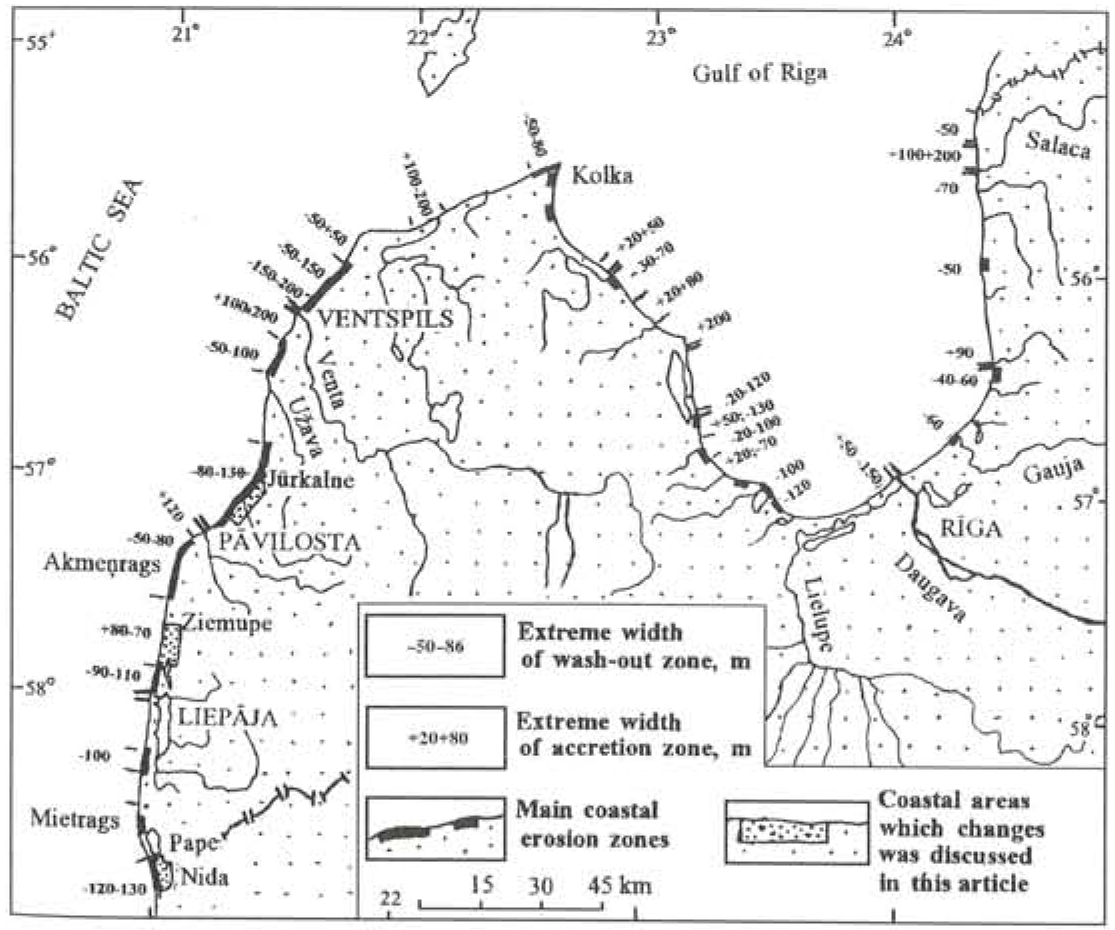


Fig. 1. A map showing total coastal zone changes during the last 60 years.

Fig. 2. A map showing increasing coastal erosion along the Latvian coast. A: 1960–1965; B: 1992–1993.

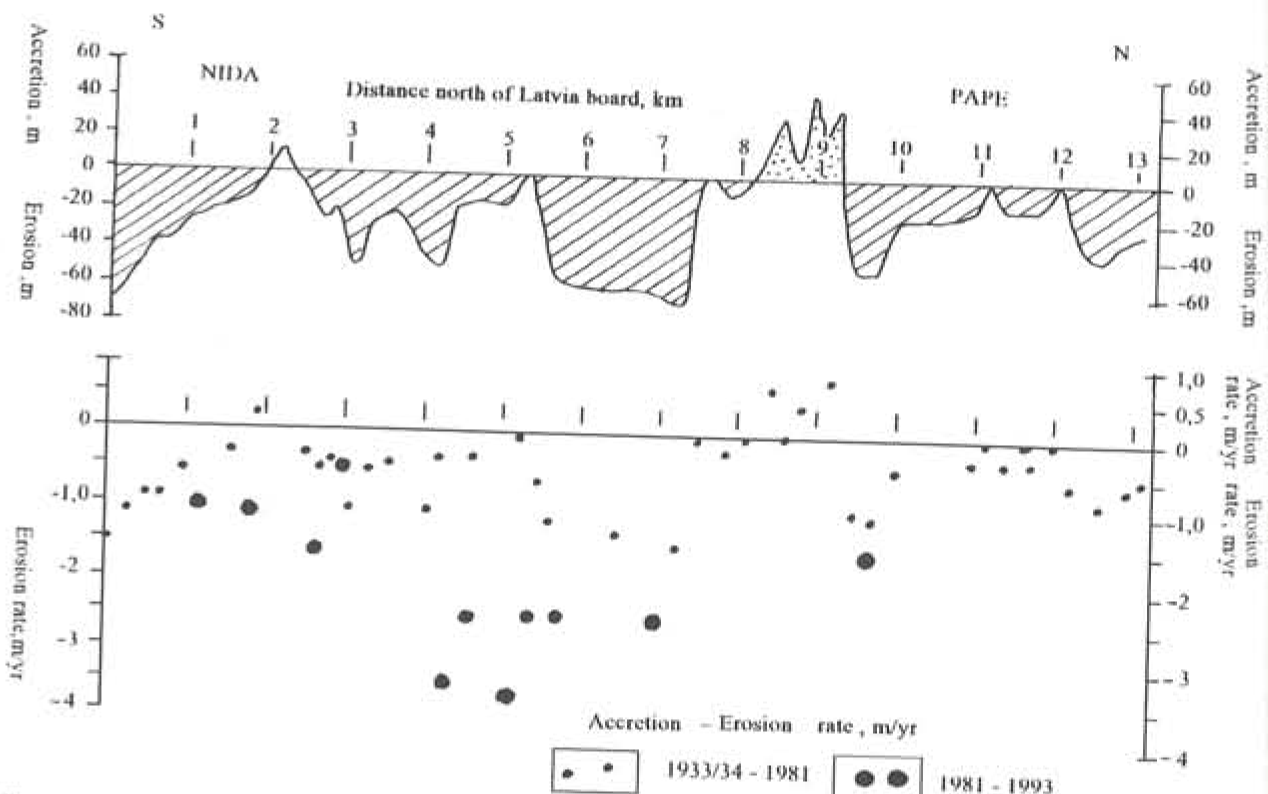


Fig. 3. Erosion and accretion along the Nida-Pape coast during the last decades.

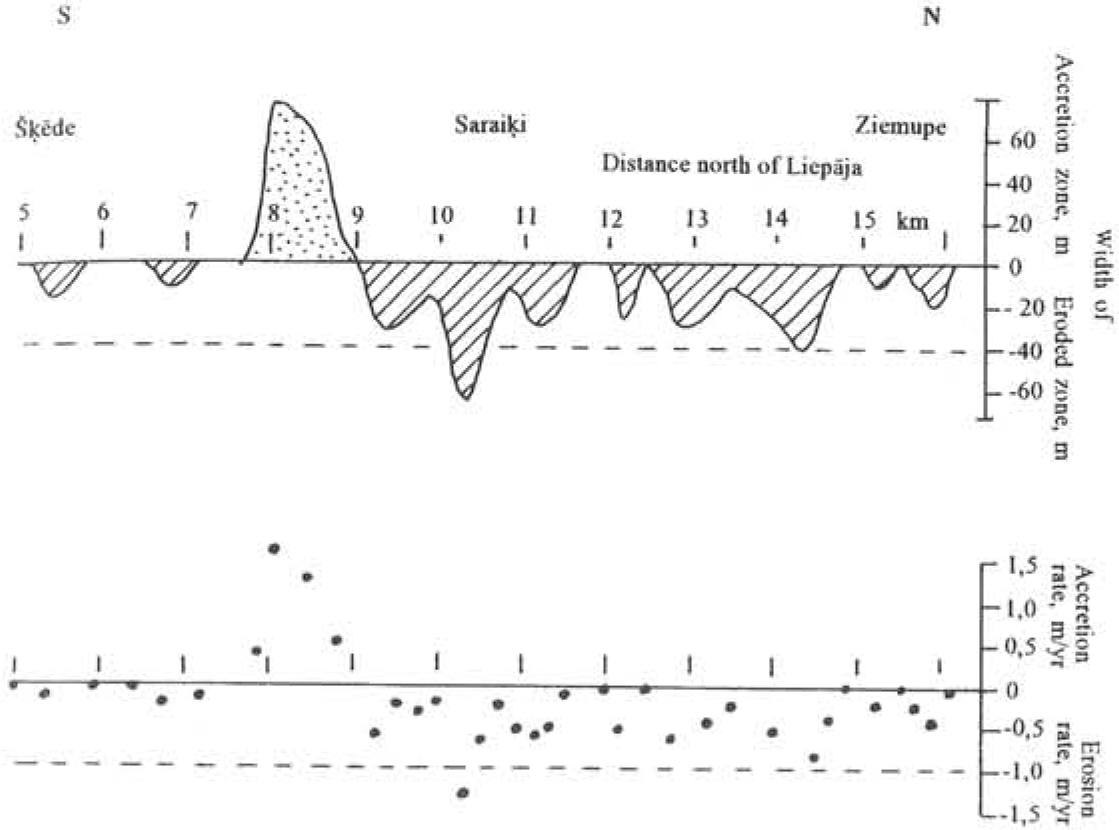


Fig. 4. Erosion and accretion along the Šķēde and Ziemupe coast (1936–1985).

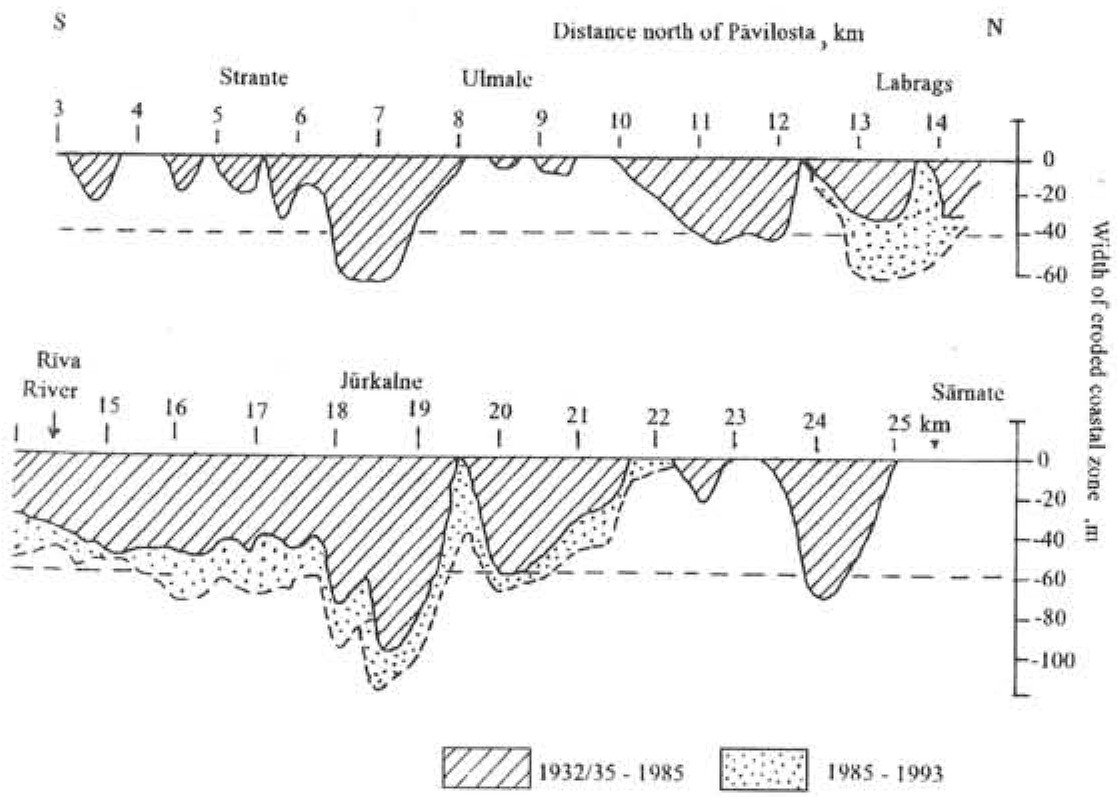


Fig. 5. Cliff erosion along the Ulmale and Jūrkalne coast.

Table 1. **Coastal erosion and accretion rate between Nida and Cape Oviši during the last 50 years (1931/35 – 1981/85)**

Coastal zone area	Coastal erosion				Accretion			
	maximum		prevalent		maximum		prevalent	
	width of washout zone, m	erosion rate, m/yr	width of washout zone, m	erosion rate, m/yr	width of accretion zone, m	rate of coastal zone increase, m/yr	width of accretion zone, m	rate of coastal zone increase, m/yr
Jaunupe-Staldzene	50	1.0	10–20	0.2–0.5	10	–	–	–
Staldzene-Ventspils	130	2.6	30–60	0.6–1.2	–	–	–	–
Sārnate-Jūrkalne	90	1.2	20–40	0.4–0.8	–	–	–	
Jūrkalne-Ulmale	55	1.1	20–40	0.4–0.8	–	–	–	–
Ulmale-Strante	65	1.3	5–20	0.1–0.5	–	–	–	
Pāvilosta-Akmeņrags	50	1.0	–	–	50	1.0	20–30	0.4–0.5
Akmeņrags-Ziemupe	20	0.4	–	–	30	0.3	10–20	0.1–0.4
Ziemupe-Saraiķi	65	1.3	10–25	0.2–0.5	–	–	–	–
Saraiķi-Šķēde	30	0,6	–	–	80	1.6	20–40	0.4–0.8
Pape	50	1.0	15–30	0.3–0.6	40	0.8	–	–
Pape-Nida	125	2.7	15–50	0.3–1.1	35	0.8	–	–
	20–130 (50–65)	0.6–2.6 (1.0–1.3)	10–60 (25–35)	0.1–1.2 (0.5–0.6)	0–80 (20–50)	0.3–1.6	20–30	–

Šķēde area it ranges in 20–100 and 20–65 m, correspondingly (Figs. 3, 4 and 5). The calculation has shown that mean erosion rate during this period ranged in 0.5–1.0 m/yr, mainly, and only in some local areas it reached 1.5 m/yr. Similar rates were figured out for other coastal areas (Table 1).

Coastal Erosion Rate During the Last Decade

Different situation was observed during the last decade. The width of wash-out zones in all three areas separated and determined before reach 10–40 m, but mean erosional rate of soft cliffs increased considerably to 1.5–5.5 m/yr (Figs. 3 and 6).

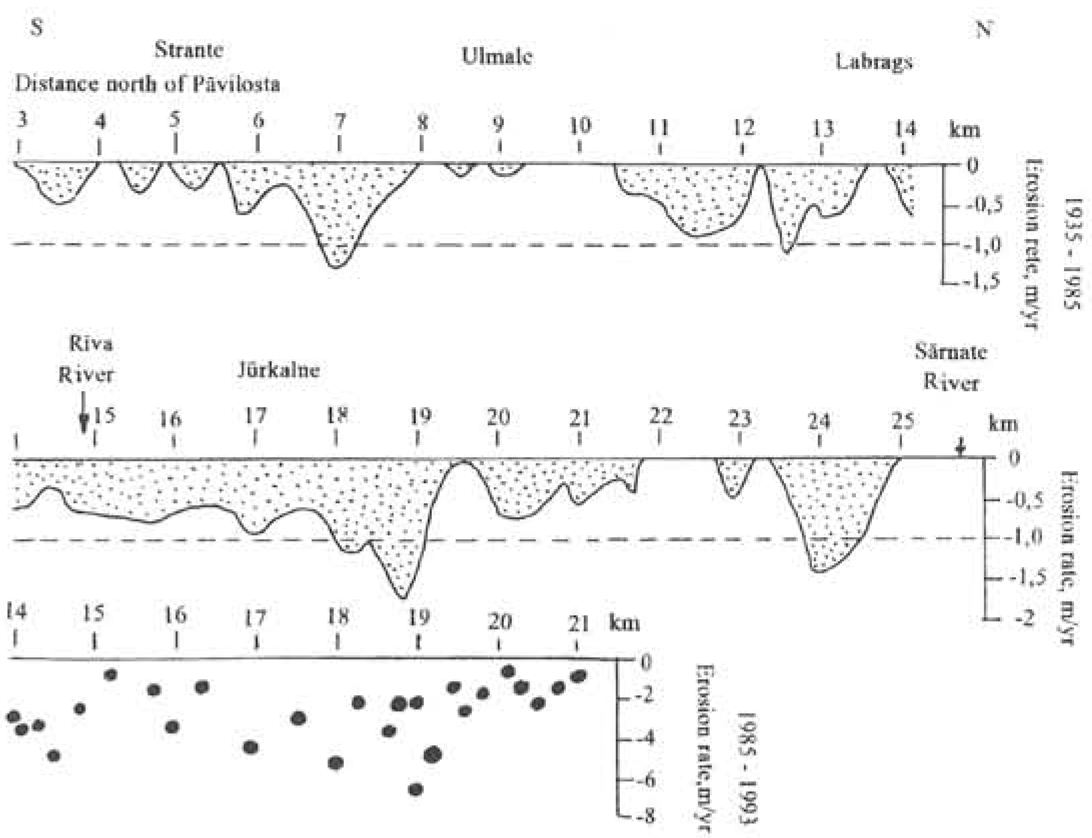


Fig. 6. Erosion rate of cliffs at the Ulmale and Jūrkalne coast.

Table 2. **Coastal erosion rate on the open Baltic coastline (Kolka—Nida) during the last 6–12 year period (1981/85 – 1993)**

Coastal zone area	Height of bluffs, m	Coastal erosion	
		Width of wash-out zone, m	Erosion rate, m
Kolkasrags	2–9	16–20	1.7–2.2
Jaunupe-Liepene	5–15	70–80	12–15
Liepene	5–12	40	2.5
Liepene-Staldzene	5–17	20–50	2.2–5.5
Staldzene-Ventspils	8–17	25–30	3.0–4.0
Vārve	2–17	30–70	2.5–5.8
Jūrkalne-Strante	10–15	20–40	1.5–5.0
Cape Bernātu	2–8	50–60	5.0–6.0
Cape Mietrags	2–8	20–30	2.0–3.0
Pape-Nida	1.5–4	10–40	1.0–3.0
		10–80 (20–50)	1.0–15.0 (1.5–4.0)

Table 2 gives the erosional rates along the open Baltic coastal zone (250 km) with a mean erosional rate in some separated areas ranging from 1.0 to 15 m/yr and prevailing in 1.5–4.0 m/yr. It gives evidence that coastal erosion rate during last decade has experienced a 2–5-fold increase if compared to that for the past 50 years.

COASTAL EROSION ALONG GULF OF RĪGA

Coastline of Gulf of Rīga during last 50 years has changed as well (Fig. 1). More important wash-out took place only on a low-lying accumulative western (Kurzeme) coast particularly from Roja to Jūrmala. Significant coastline changes could not be fixed in the accumulative southern part of the Gulf with the equilibrium of coastal processes or with accumulation near the mouths of the Gauja and the Lielupe rivers. Insignificant mainly low erosion (2–6 m) of soft cliffs or those with sandstone erosion are located in the zone from the Age to the Vitrupe rivers.

A total width of the eroded zone in the Kurzeme coast has reached 40–100 m (Fig. 1, Tab. 3) and mean erosional rate is 0.1–0.5 m/yr.

The recent soft low bluffs erosion rates were calculated according to the measurements in

Table 3. **Coastal erosion and accretion along the western side (Kurzeme coast) of the Gulf of Rīga during last 50 years (1932/37–1981/83)**

Coastal zone area	Erosion				Accretion	
	Maximum		Prevalent		Maximum, m	Prevalent, m
	Width of wash-out zone, m	Erosion rate, m/yr	Width of wash-out zone, m	Erosion rate, m/yr		
Roja	30–32	0.6	5–15	0.1–0.3	–	–
Rojnieki	50	1.0	0–15	0–0.3	30	0–10
Kaltene I	45	0.9	10	<0.1	–	–
Kaltene II	40	0.8	5–20	0–0.4	14	–
Kaltene-Valgalciems	27	0.5	5–20	0–0.4	43	15–25
Engure	115	2.5	20–50	0.3–1.0	120	50–60
Ķesterciems	80	1.7	15–30	0.3–0.6	–	–
Plieņciems	65	1.4	–	–	20	0–15
Apšuciems	70	1.5	0–10	0–0.1	10	–
Klapkalnciems	70	1.5	10–40	0.2–0.8	25	–
Ragaciems-Bigauņciems	100	2.0	20–50	0.2–1.0	–	–
	30–115 (50–100)	0.5– 2.5 (0.8–1.5)	5–50 (10–30)	0.1–1.0 (0.1–0.5)	10–120 (0–40)	10–60 (0–20)

15 stationary areas fixed up specially with more than 290 measuring lines. During the recent strong storm in winter 1992 low coastal sandy bluffs erosion rate prevailed from 1.0 to 2.0 m per storm, rarely 3–5 m (Table 4).

The recent rates of high (8–15 m) soft cliff erosion on the Baltic coastline during the latest strong storm period (January 1993) was more considerable than along the Gulf of Rīga. Maximum rate in some measured profiles has reached

Table 4. **Soft cliff erosion rate during the January 1993 storms along the Kurzeme coast of the Gulf of Riga (Eberhards & Saltupe 1993)**

Coastal erosion research area	Coastline lenght, m	Height of erosional bluff, m	Measurement profiles	Coastal erosion	
				Total, m	Prevalent, m
Aizklāni I	160	5–12	8	0–2.0	–
Roja	300	1–3	44	0–2.0	–
Sillidumi	200	2–7	13	0–4.0	–
Valgalciems	330	1–1.5	10	0–0.7	–
Upesgrīva I	450	1–1.7	17	0–1.7	–
Upesgrīva II	560	1–1.5	19	0.2–3.5	1.0–2.0
Abragciems I	480	1–1.	24	0–4.5	2.0–3.0
Abragciems II	200	1–1.7	13	0–2.0	1.0–1.3
Engure	900	1–4	15	1.0–4.0	1.0–2.0
Ķesterciems I	590	0.5–1.2	33	0–2.4	1.0–1.8
Ķesterciems II	490	0.5–2.2	21	0–3.7	1.5–2.0
Plieņciems-Apšuciems	180	0.7–0.5	10	0.2–2.1	1.0–1.5
Ragaciems (cape)	1280	1–2.5	33	1.5–4.6	1.0–2.0
Ragaciems	250	1–2.5	16	1.0–4.5	0.5–1.0
Bigauņciems	320	1–1.5	19	1.5–3.6	2.0
				0–4.6	1.0–2.0

Table 5. **Recent rates of soft cliff erosion on the open Baltic (storm series January 1993). Stationary measurement data**

Coastal erosion research area	Height of cliff, m	Lenght of coastline, m	Number of profiles	Cliff erosion rate	
				Maximum, m per storm series	Mean, m per storm series
Kolka I	2–8	940	36	8.8	4.0
Kolka II	1.5–10	530	19	1.0	–
Jaunupe	2–5	150	9	7.5	4.8
Liepene	5–13	330	20	2.0	–
Bušnieki III	5–13	390	21	5.7	–
Bušnieki II	4–17	330	16	3.1	–
Bušnieki I	4–6	130	9	4.3	2.9
Staldzene	5–8	460	20	2.0	–
Ventspils IV	7–16	630	29	5.9	3.2
Ventspils III	8–10	420	21	1.0	–
Ventspils II	8–16	120	9	3.8	1.2
Ventspils I	8–12	350	18	5.6	3.2
Vārve	4–15	820	38	6.7	3.2
Jūrkalne IV	15–17	360	18	6.0	3.2
Jūrkalne III	14–15	200	12	3.3	2.0
Jūrkalne I	10–15	360	18	7.2	3.0
Jūrkalne II	10–15	170	8	4.3	3.3
Muiža	11–15	300	32	10.9	6.0
Riva	11–15	480	22	5.0	1.5
Ulmale II	15	270	7	6.6	4.8
Ulmale I	10–12	420	30	7.0	2.4
Strante	10–12	530	34	6.5	2.7
Bernāti III	1.–8	750	22	12.6	7.0
Bernāti I	1.5–5	1400	29	18.0	14.5
				1.0–18 (4.0–18.0)	1.2–14.5 (2.6–6.0)

10–18 m, mean rate ranged from 1.2 to 14.5 m, or prevailed in 2.5–6.0 m per storm (Table 5).

EROSION AND ACCRETION ZONES ALONG THE OPEN BALTIC SEA COASTLINE

First special geomorphological mapping of the coastal zone was done in 1989 and repeated in 1992/1993, it enabled us to indicate recent permanent erosion zones in succession with accumulative (accretion) or equilibrium zones along the Baltic Sea coastline from Nida to cape Ovīši (Fig. 7). The length of separated erosion areas mainly is 7–10 km, but areas with accelerated sediment accumulation and foredunes growth or zones with coastal processes equilibrium reached 3–8 km in length. Only to the north from Pāvilosta, there stretches a 10–15 m high soft cliff zone for about 25 km, it is constantly ancient because of permanent wash-out which was in progress during the last centuries (Daņilāns & Veinbergs 1992). Other much vaster and continuous soft bluff zone stretches to the northeast from the seaport of Ventspils. In the 60s, active eroded sandy bluffs stretched only for 3 km, usually with only 8–17 m high bluffs, later the zone increased and became 14 km long in January 1993 (Fig. 7).

Particularly visible increase of coast erosion feature has appeared on those accumulative coastal areas where the accretion dominated during the last century. Such most typical examples, as Cape Bernāti and Mietrags situated southwest from Liepāja, are characteristic. The episodic erosion near Bernāti lasted all this century and width of the eroded cliff zone was only 1 km. During recent decade, the accelerated wash-out area spread and reached about 7 km in 1993. The erosion during winter and autumn storms destroyed backshore zone with 3–4 foredunes. In recent times, the wash-out of coastal dune ridges has started (Fig. 8). A similar accumulative coast wash-out is widespread also at Cape Mietrags located 15 km to the south. V.Venska measurement data northwards from the seaports of Liepāja and Ventspils show that high annual erosional rate in some sandy cliff areas reach 5–10 m .

The zones of significant accumulation during last century existed southwards from Liepāja and Ventspils.

The general direction of beach and nearshore sediment transport is northward or northeastward. This littoral flow is known as the South Baltic longshore current (Fig. 9; Knaps 1966). There is an evident deposit transport southeastwards along the western coast and southwards—northwards along the eastern coast of the Gulf of Rīga.

The calculated erosion rates show values differing from place to place in terms of the shoreline direction, and high cliff erosion being caused by the combination of marine and subaeral factors and controlled by local geology.

Examination of hydrographic charts compiled in 1947 and 1978 or 1984 show an increase in nearshore water depths at the open Baltic and Gulf of Rīga areas with the deterioration of shoals, wave energy almost certainly has increased and caused the increased potential for erosion.

The wave refraction modelled by L.Weishar, W.Tiffney and De Kimpe Jr.N. (1991) clarified that focusing of wave energy on the beach is controlled strongly by the configuration and location of shoals. Episodicity in longshore sediment transport is controled by wind azimuth, velocity and wave climate (Seymour & Castel 1985). Main values of longshore sediment transport must be completed during the only one strong stormy day or some decades of hours.

Lasting permanent wind azimuth in storm has caused successional erosional and accretion zones exchange in longshore sediment transport stream. The patterns of erosion and accretion along the coast appear to be related to the position of large accumulation of sand referred to here as the "lobe"

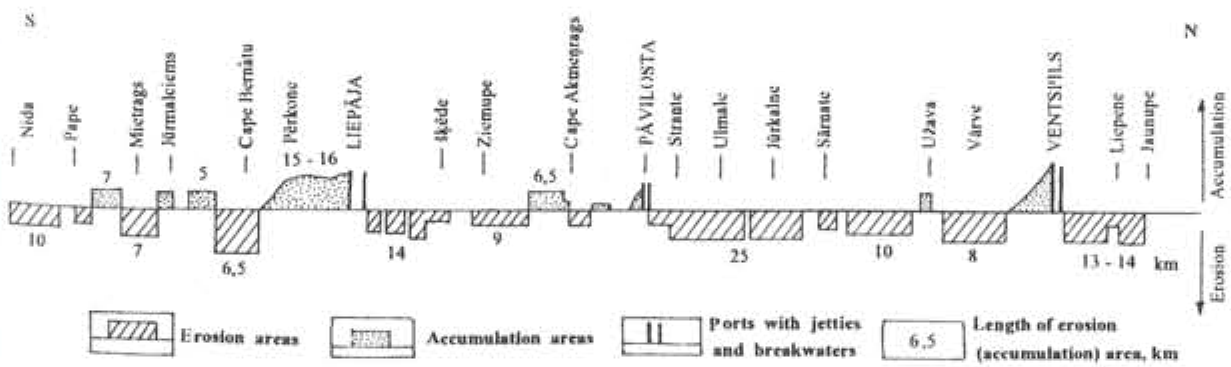


Fig. 7. Distribution of recent erosion and accretion zones along the open Baltic coast from Nida to Ovīši.

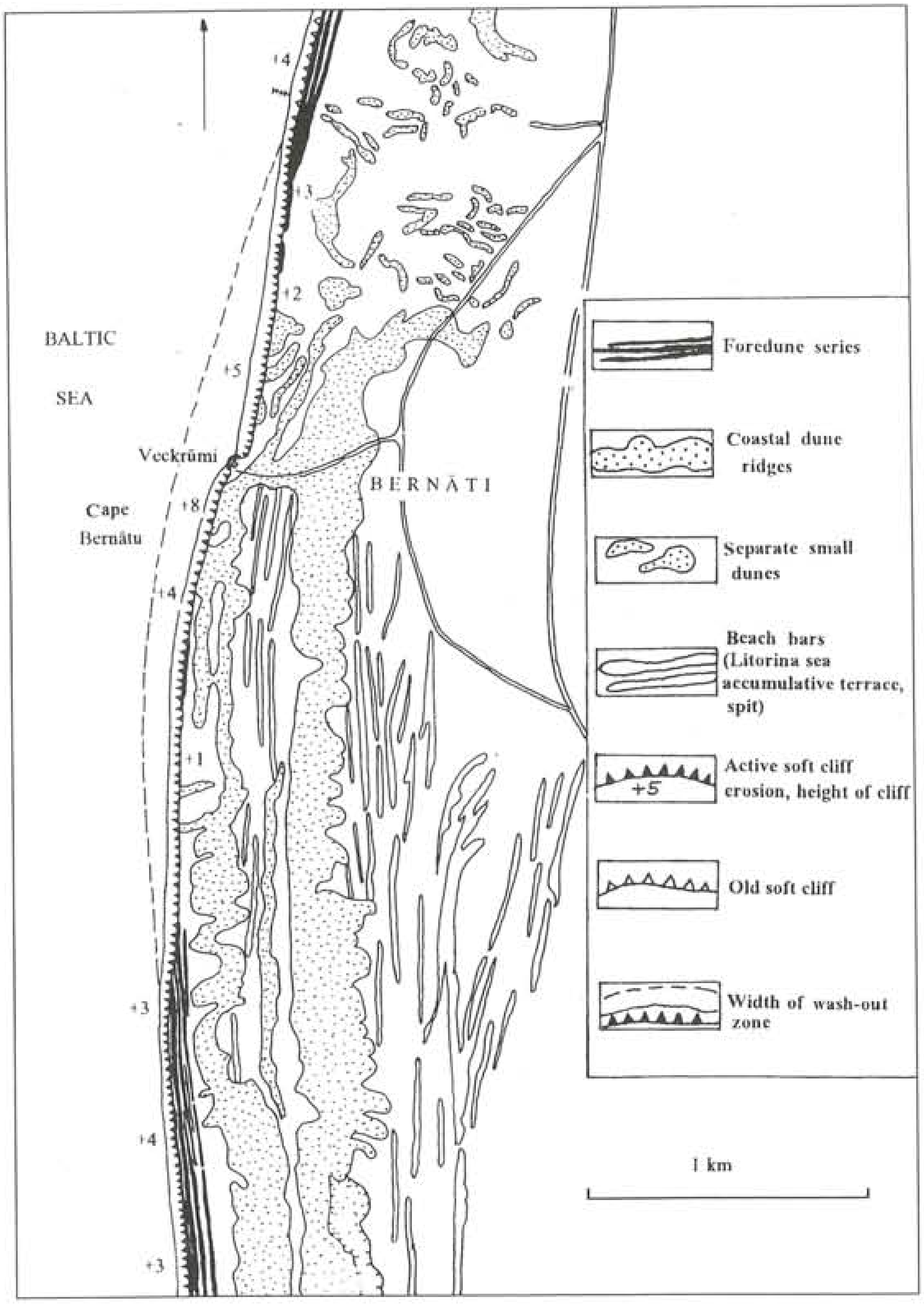


Fig. 8. Geomorphological map showing the accelerated accumulative coastal zone erosion along the Cape Bernati.

(Weishar, Tiffney & De Kimpe 1991), or "sand-waves" (Verhagen 1991) with about 0.3–0.6 million m³ of deposits and length as far as 1.5 km or more (Safjanov 1987; Kozuhov 1968).

We have discovered similar sandy bluff erosion zones and zones without cliff erosion successional exchange northeastwards from Ventspils after the 1993 strong winter storm series mainly with south-

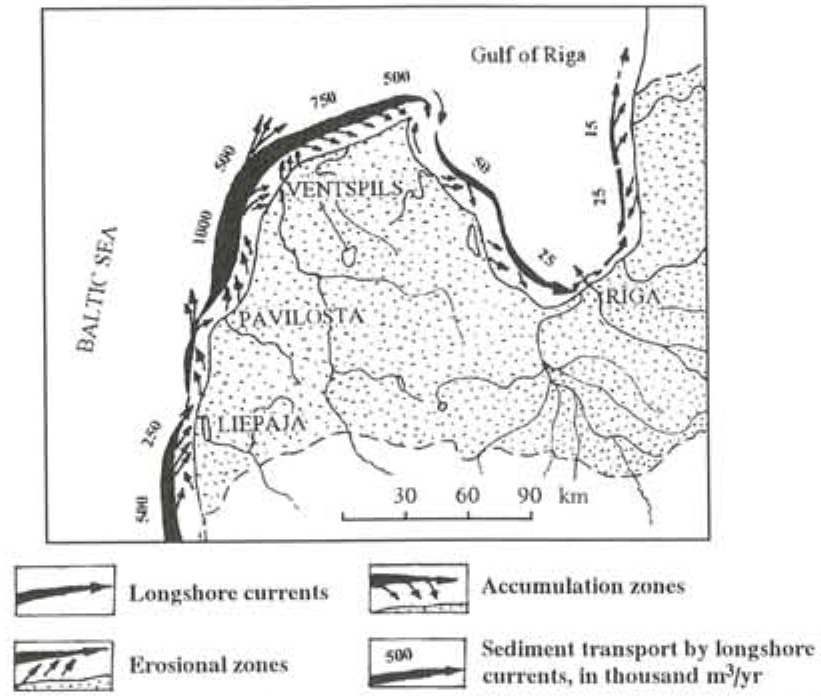


Fig. 9. Longshore currents, erosion and accumulation zones along the Latvian coastline (Knaps 1965).

west and west azimuth. The length of these zones ranges from 0.8 to 1.0 km.

SEA LEVEL CHANGES

Accelerated worldwide beach and cliff erosion which took place during the last decades is estimated by the scientists as the result of an increase in global mean temperature, a rise in mean sea level and an increase in storm frequency (Bruun 1962; Armah 1991; Wiles 1989). During the century, sea level rise has been approximately 10 cm (Hanson & Lindh 1993).

Observations of sea level rise at 10 coastal zone stations of Latvia cover 50–100 years of measurements. The analysis of annual mean sea level time scales could not confirm the general increase trends. The accelerated annual sea-level rise is observed in three stations (Skulte, Daugavgrīva and Grīva) in the southern part of the Gulf of Rīga (Fig. 10A; Eberhards & Saltupe 1993), while in other coastline

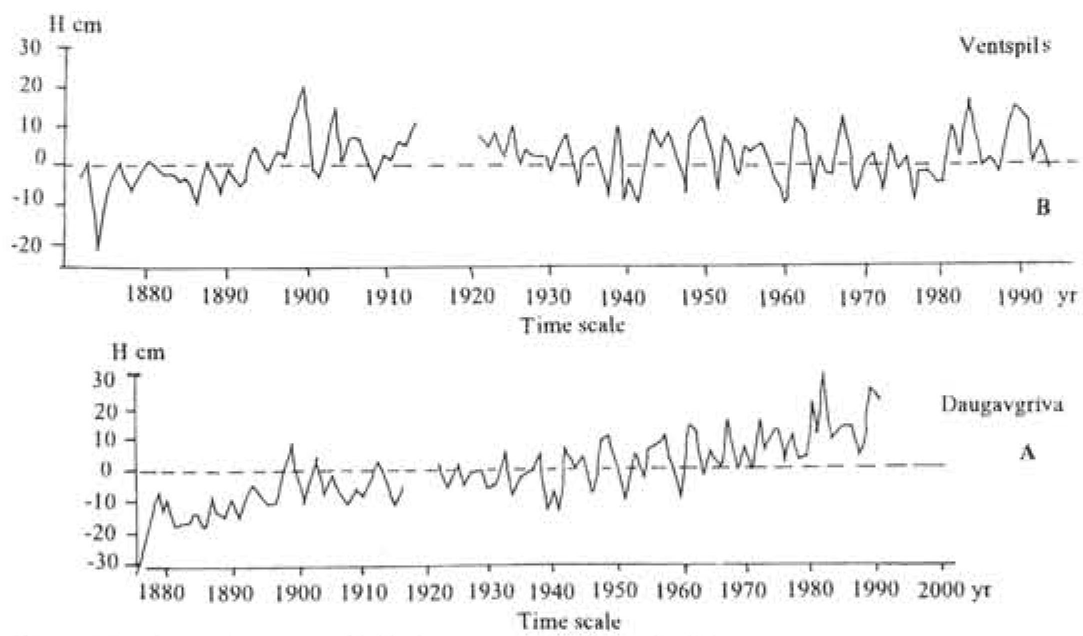


Fig. 10. Time series of annual mean sea-level: Daugavgriva (A), Ventspils (B).

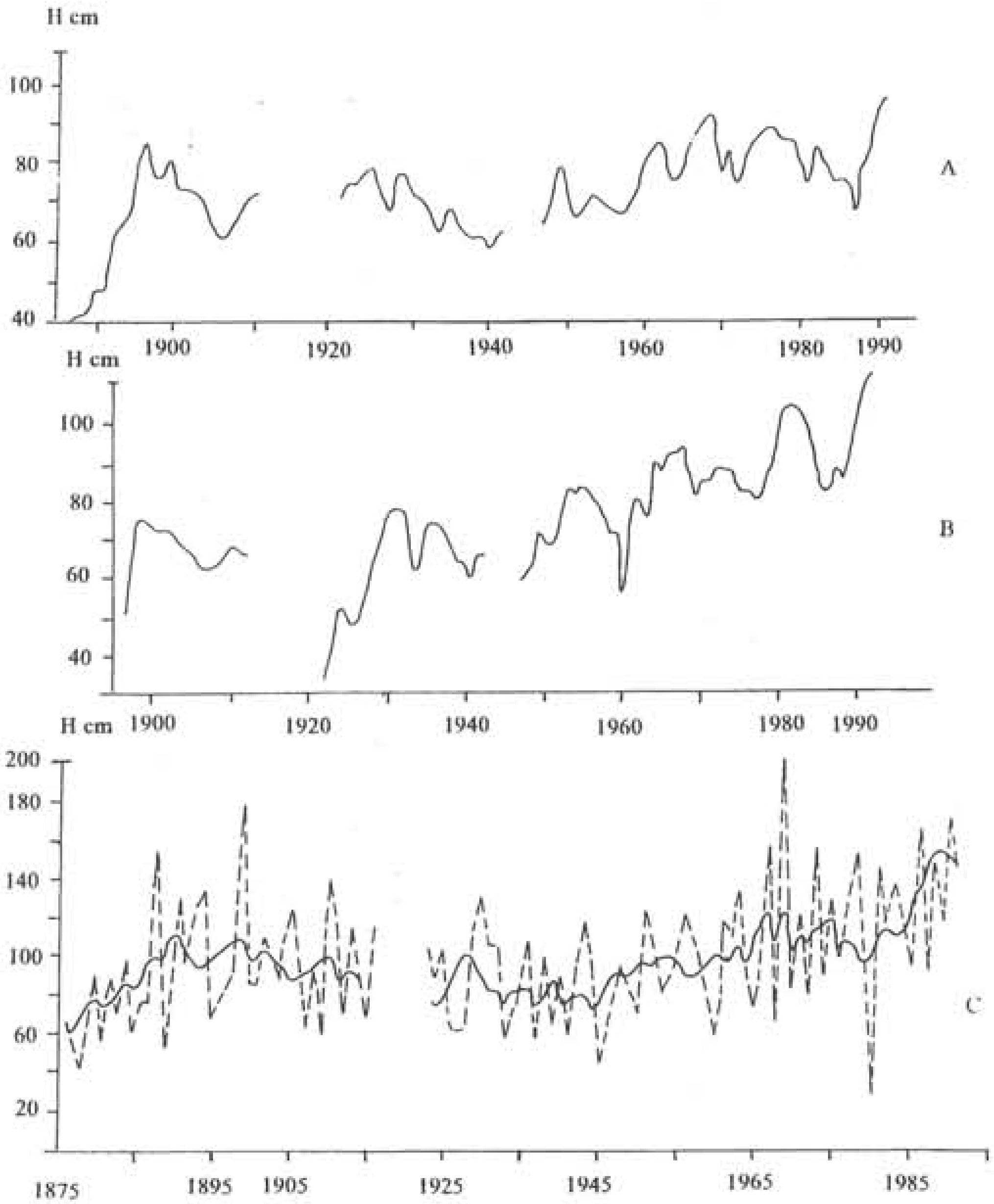


Fig. 11. Time series of annual high sea-level: Ventspils (A), Liepāja (B), Daugavgrīva (C).

areas, including the open Baltic Sea coast, similar tendency could not be observed (Fig. 10 B).

The effect of global sea level rise on the shoreline will be spatially non-uniform because of the presence of local vertical crustal movements, differences in resistance to erosion, varying wave climate and longshore currents (Gornitz, White & Custman 1991).

Moreover, along all the Latvian coast, the sea level observation data show the increase in annual extreme (storm) water levels during the last century, especially since sixties-seventies (Fig. 11). This annual extreme sea level rise trend has a good correspondence with an increased accelerated coastal erosion rate during the last decade, as it is discussed above in this article.

It is able to tolerate that extreme sea level rise reflected by the number of severe storms in winters and stormy autumn seasons.

STORM FREQUENCY

Still the analysis of frequency of storms (the number of observed storms exceeding 14 m/s) could not confirm a similar increase (Fig. 12), but quite the opposite – seawards the number of strong winds decreased during the last decades, although the frequency of south, southwest and west winds has increased during winter (Table 6).

Therefore, it seems that accelerated sea coast erosion around the Latvian coastline during the

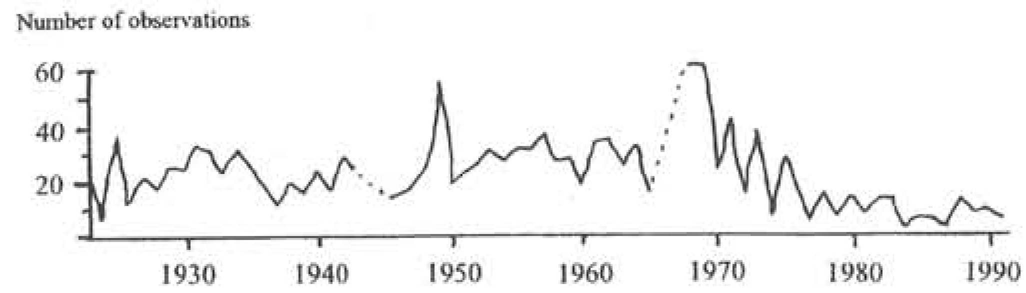
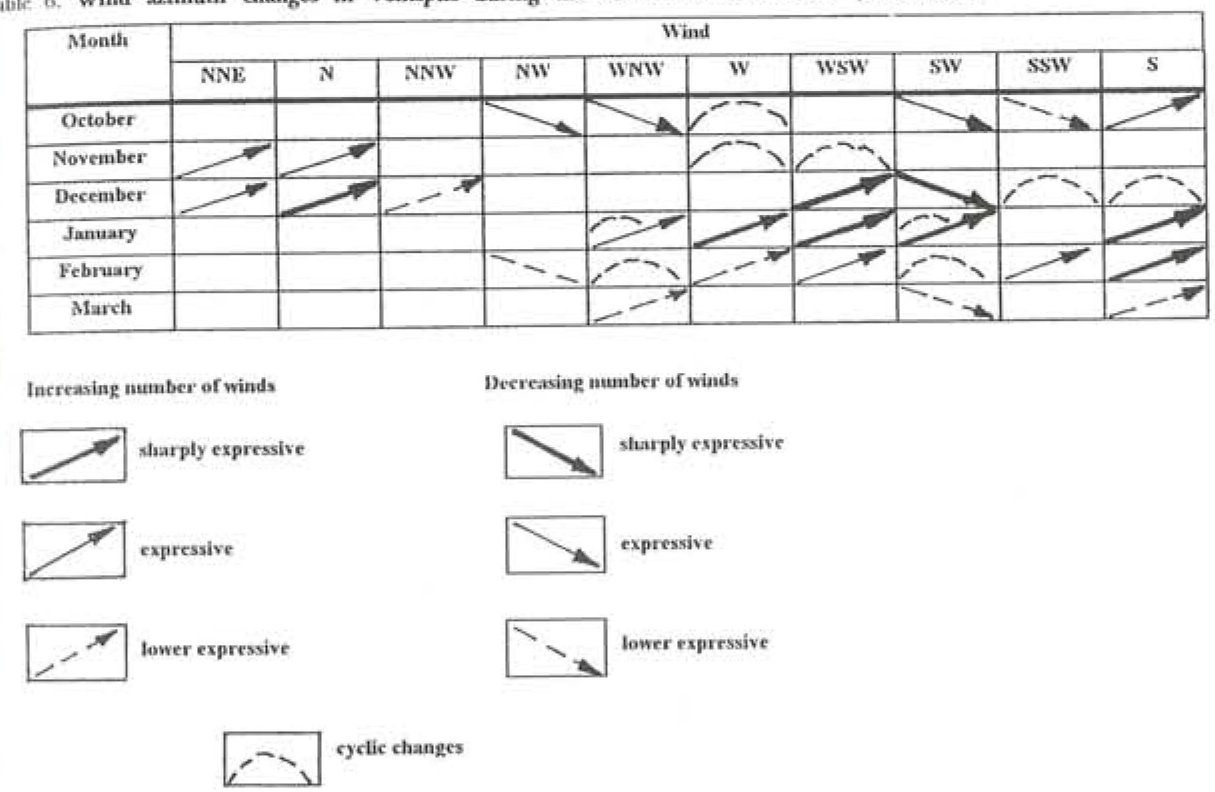


Fig. 12. Time series of annual strong winds (>14 m/s) during the autumn–winter seasons (I+II+III+X+XI+XII) in Ventspils.

Table 6. **Wind azimuth changes in Ventspils during the autumn–winter seasons (1966–1991)**



last one-two decades is mainly controlled by constantly increasing wind velocity and energy.

The opposite point of view is expressed by I.Veinbergs and I.Daņilāns (1992), whereas G.A.Safjanov (1987) explained the increased erosional rate by sand deposits which dry up in the littoral zone, as well as where there are a large number of seaports (with breakwaters, jetties) and regular deepening of sea-entering channels, as well as where dragged material is being dumped in the offshore zone and the longshore sediment flow is interrupted.

CONCLUSIONS

It seems that beach and cliff erosion rapid growth in Latvian coastline during last decades is connected mainly with global and regional climate change expressed in changes of wind azimuth, velocity and frequency and an accelerated rise (mainly extreme) in sea level under a human impact: construction of seaports, deepening of sea-entering channels, as well as sand and gravel extraction from the littoral zone, beaches and large river channels.

1. Bluff erosion rates calculated from the topographical charts during the last decade mainly produce a good coincidence with the recent coastal bluff erosion rate observed on coastal bluff erosion research stationary areas.

2. The coastal environment represents an extremely complex interaction between climatic factors, waves, currents and bottom sediments causing continuous physical change in beaches and cliffs at a scale ranging from local (e.g. erosional of particular beach or cliff) to global (e.g. the rise in mean sea level.). The resulting conflicts between the eroding coast and the human activities, that

we see along coastlines around the world, may be traced back to misconceptions about the natural behavior of sandy beaches and to actions based on these misconceptions. The differences between our perceptions and the reality of coastal behaviour may often lead to unforeseen degradation of coasts.

3. The only way to avoid this is by developing methods which allow quantitatively to determine the behaviour of beaches and cliffs.

ACKNOWLEDGEMENTS

The authors are indepted to Prof. Dr. G.Lampe (Greifswald University, Germany) and Dr.hab. I.Veinbergs (Latvia) for the critical reading of the manuscript. Our thanks are due to Dr. S.Kraže for helping with the English language.

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ESR Dating of Subfossil Mollusc Shells of the Limnea Sea

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Molodkov, A. & Raukas, A. 1996: ESR dating of subfossil mollusc shells of Limnea Sea. *Baltica Vol. 9*, pp. 29–35. Vilnius. ISBN 9986–615–05–4 / ISSN 0067–3064

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The dating of the Baltic Sea deposits and shorelines needs new precise physical methods. In the paper the reliability of the results on young subfossil Holocene shells from the Limnea Sea deposits (Hiiumaa Island, Estonia) by the ESR dating method is discussed. Most of the dates fit well with predicted geological ages.

Keywords: chronostratigraphy, *Cerastoderma glaucum*, *Macoma baltica*, electron-spin-resonance (ESR) dating, Limnea Sea, subfossil mollusc shells, Baltic Sea, Hiiumaa Island, Estonia.

INTRODUCTION

An accurate and high-resolution chronology is extremely important in the Holocene stratigraphy but to realize it will be rather difficult, because all available physical dating methods have a lot of limitations and errors. Especially complicated is the dating of sediments which do not contain organic material. The stratigraphy of the deposits of the Baltic Sea in the coastal area of Estonia is based mainly on the dating of buried organic layers under the transgressional formations of the sea and lagoonal sediments, which allow to establish approximately also the age of neighbouring coastal relief forms. But in the history of the last stage of the Baltic – Limnea Sea in Estonia transgressional series are absent. It means that all the sediments of the Baltic Sea which were formed in Estonia during the Subboreal and Subatlantic chronozones since 4000 years ago until present, were formed in regressional conditions what hampers the correlation of deposits and shorelines.

The Limnea Sea was recognized in 1886 by G. Lindström on the basis of disappearance of the genus *Littorina* and the introduction to the Baltic of the fresh-water mollusc *Lymnaea ovata* (Drap.) f. *baltica* Nilss. According to Kessel (1958), the immigration of *L. ovata* to the coastal waters of Estonia took place about 4000 years ago. Therefore, and in agreement with the image common in Sweden (Fredén 1979), it was proposed to define the Litorina/Limnea boundary at that chronostratigraphical level (Hyvärinen et al. 1992), however the boundary itself is gradual and the palaeoenvironmental conditions at the beginning of the Limnea Sea were rather similar to those at the end of the Litorina Sea. The salinity of the coastal waters in the western part of the Gulf of Finland in the first half of the Limnea Sea was approximately 10‰. The salinity of the water in the coastal areas of the Litorina Sea varied between 8 and 15‰ (Kessel & Raukas 1979). Already about 2500 years ago the salinity of the water was only 2–3‰ higher than in the contemporary sea (Raukas, Kessel & Hyvärinen 1992).

The Limnea Sea in Estonia is represented by five regressive phases (Lim_{I-V}). The maximum level of the Limnea Sea (Lim_I) is at an absolute height of 13–3 m, and the level of the latest phase Lim_V at a height of only 2.5–1.5 m above the contemporary water level (Kessel & Raukas 1979). In several places (Niibi, Õlbiku a. o.) it was possible to date lagoonal deposits of Limnea age using ^{14}C method, but these did not give a precise answer to the question concerning concrete shorelines and therefore direct dating methods are highly needed.

Königsson and Olsson (1981) attempted radiocarbon dating of subfossil molluscs. However, there is a potential danger that small shells can be contaminated with nonrepresentative carbon from the environment. In several cases the thermoluminescence (TL) method has also been applied to dating of marine sands, but as to the Upper Holocene, the method has been limited by uncertainties associated with the residual TL signal found in recent sediments. More recently, a dating method based on new luminescence readout technique using a laser (Huntley et al. 1985) or infra-red (IR) diodes with λ=880 nm (Godfrey-Smith et al. 1987; Godfrey-Smith et al. 1988) to stimulate electrons from light-sensitive traps in quartz and feldspars has been developed. Although almost all other steps in the optically stimulated luminescence (OSL) dating procedure are similar to those used in the conventional TL dating of unheated sediments, the new readout technique can utilize light-sensitive traps for dating, because lose of the electrons when exposed to direct sunlight, can occur for a very short time, and hence the residual signals for modern sediments can be easily zeroed. This, in principle, will enable to date in much reliable way the younger sediments. However, subaqueous sediments seem to be not the best objects for the OSL dating because the use of wavelengths outside the range ca. 500 to 625 nm would be inappropriate for zeroing the residual signals due to strong attenuation of the shorter and longer wave length by water (see e.g. Berger 1988). The aforesaid accounts for the interest of geologists in the ESR dating of Late Holocene shells which have given already promising results (see e.g. Molodkov 1986, 1988, 1989a; Molodkov & Raukas 1988; Molodkov et al. 1992).

In order to check the reliability of the results, obtained on young subfossil Holocene shells by the ESR dating method, we undertook a detailed research into the Limnea sediments on Hiiumaa Island, where due to the intensive crustal uplift (at present 2.5 to 3 mm per year) the Limnea shorelines are better preserved, and Limnea sediments of different age lie at some distances from each other. Four localities – Vanajõe, Muda, Kaderna and Viiri were subject to detailed geological, malacological and ESR studies (Fig. 1).

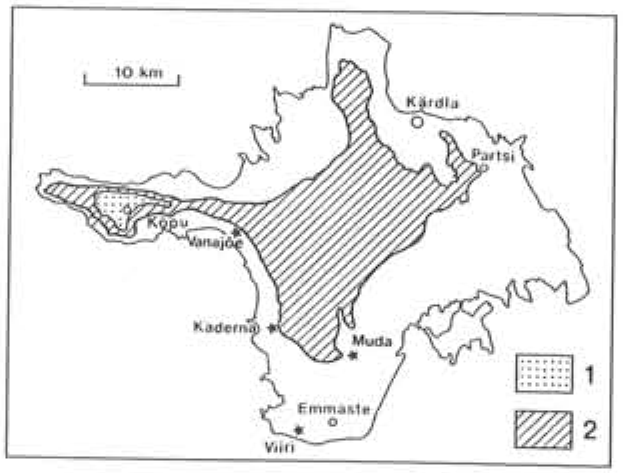

Fig. 1. Location of investigated sites and the evolution of the Hiiumaa Island in postglacial time after Kessel and Raukas (1967), with complements. 1 – small island (about 1 sq km) at the time of the Ancylus transgression about 9000 (^{14}C) years ago; 2 – Hiiumaa at the end of the Litorina Sea. Asteriks represent the locations where subfossil mollusc shells for the ESR analysis were collected.

We had no task in this study to compare our results with those obtained by ^{14}C method because shells (and bones) are acknowledged as the nonpromising material due to the possible contamination from the environment or during sampling, handling, preparation or storing. According to some authors (see e.g. Goslar et al. 1986), thinwalled freshwater mollusc shells are entirely unsuitable for 14C dating. Our main control here was geological. We believed that the neotectonic uplift in the Late Holocene during last 4000 years was more or less linear.

ESR–DATING

ESR–dating is based on the direct measurement of the amount of radiation-induced paramagnetic centers (radiation damage) in shell substance that are created by the natural radiation resulting from radioactivity in the shell itself and from the environment (embedding matrix and cosmic). In ESR spectrometry the presence of paramagnetic carbonate centers in mollusc shell substance can be detected because they show characteristic absorption patterns at certain locations within the recorded spectrum. The position of the peaks within the spectrum are identified by their g-factor values (Fig. 2a). The concentration of the centers can be determined from ESR signal intensity. The signals are absent in modern shells, but increases in intensity (amplitude or peak height) as a function of the total radiation dose absorbed by the shell over the time of burial.

The ESR age, T, of the shell samples studied in the present work was determined by the equation (Molodkov 1986):

$$T = \frac{P_s}{\dot{D}_c + W_\gamma \dot{D}_{ext\gamma} + W_b k_1 \dot{D}_{ext\beta} + k_3 k_\alpha \dot{D}^U_{int\alpha} + k_2 \dot{D}^U_{int\beta}},$$

where P_s is the accumulated palaeodose determined by ESR; $\dot{D}_c$ is the cosmic dose rate; W_γ and W_β are the correction factors for water; k_1–k_3 are the correction factors for α- and β-radiation; k_a is the α-efficiency; $\dot{D}_{ext\,\gamma,\beta}$ is the external dose rate; $\dot{D}^U_{int\,\alpha,\beta}$ is the internal dose rate due to ^{238}U in the shell.

To determine the unknown value of accumulated palaeodose, the cleaned, etched, powdered and sieved to 75–400 μm fraction of shell samples were divided into 6–10 aliquots, 200–300 mg each, and then irradiated from 15 to 300 Gy using a ^{60}Co source delivering 1.5×10^{-2} Gy s^{-1}. After irradiation, all shell aliquots were annealed for 2.5 hr at 100°C to reduce the effect of short–lived signals. For reconstruction of the grow curve outside the additive dose section, logarithmic transformation was made of dose-vs-ESR intensity.

Typical spectrum of Holocene shells from Hiiumaa Island recorded with microwave power of 2 mW and a modulation amplitude of 0.25×10^{-4} T is shown in Fig. 2a.

In our previous works (see e.g. Molodkov 1986, 1988, 1989b, 1993) it was shown, that paramagnetic centers with the g-factor at 2.0012 and the relevant signal (line width ≈ 2.25×10^{-4} T) are most reliable for quantitative analysis and dating of Quaternary mollusc shells including those of marine and brackish-water origin. The signal at

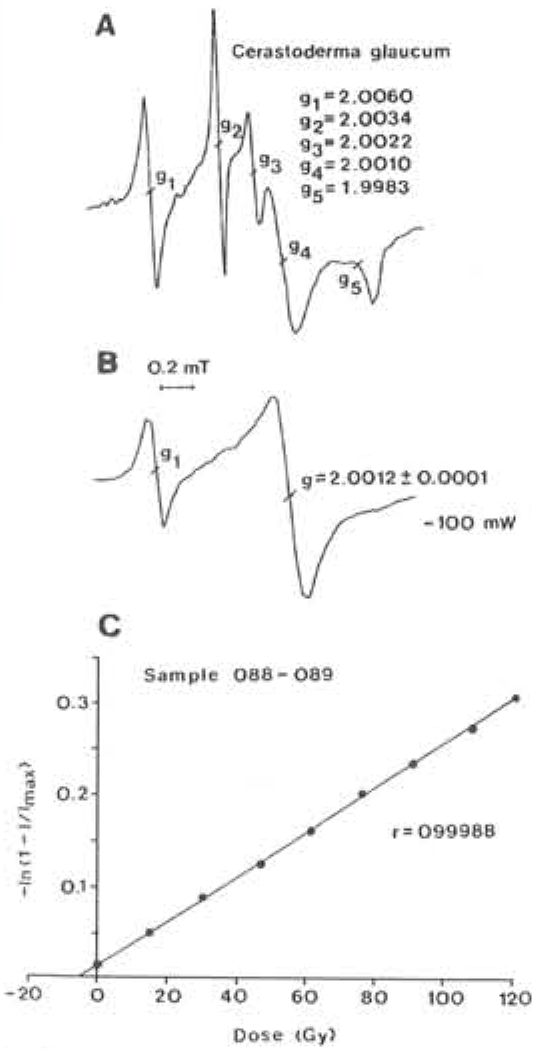

Fig. 2. Typical ESR-specra of Holocene brackish-water mollusc shells from Limnea Sea deposits (A); analytical signal at g=2.0012 (B) is separated from multi-component ESR-spectrum of shell sample by high microwave power (HMP) technique; (C) dose-response of the carbonate center at g=2.0012 for the shell samples from Limnea Sea deposits and mathematical evaluation of the accumulated palaeodose by the logarithmic fitting of the data points; I is the ESR intensity; I_{max} is the ESR intensity at saturation dose, r is the correlation coefficient.

g=2.0012 is not usually detected in multi-component differential spectra of the shells due to its interference with the signals of other centers, including those of non-radiation origin. In the present study the signal was separated by high microwave power (HMP) technique described in Molodkov (1988). Dosimetric reading was performed with microwave power of 100 mW and a modulation amplitude of 0.25×10^{-4} T (Fig. 2b).

U concentration in the shells was measured by neutron activation analysis. The external dose rate was deduced from the gamma-spectrometric analysis of the embedding matrix (Molodkov 1992) using conversion factors of Nambi and Aitken (1986) and taking into account the shell geometry and water content in the sediments.

The cosmic dose rate was calculated following Prescott and Stephan (1982) and the half-depth of burial.

The accumulated palaeodoses were determined from the enhancement of the intensity of the g=2.0012 signal by exposure of the shell samples to increasing gamma-doses. For determination of palaeodose outside the additive dose section, the ESR response to irradiation must be known and mathematically expressed. In contrast to other five lines, normally observed in the ESR spectra of the aragonite shell substance, the dose dependence of the g=2.0012 signal is well described by exponential function in the range of the doses applied (Molodkov 1988). It allows to convert the data point into a straight line to determine the unknown value of accumulated palaeodose by extrapolation of the dose dependence to zero-ESR-intensity (Fig. 2c).

A SHORT DESCRIPTION OF LOCALITIES AND DISCUSSION OF RESULTS

Fig. 1 shows the Holocene shorelines of Hiiumaa Island and the location of investigated sites. The evolution of Hiiumaa Island in postglacial time has been described by Kents (1939), Kessel and Raukas (1967) and in most detail by Urve Sepp-Ratas (Sepp 1974). The Kõpu Peninsula, the westernmost part of Hiiumaa, which is up to 68 m above the present sea level, was the first piece of land to rise above sea level at the end of the Baltic Ice Lake regression about 10,500–10,200 conventional non-corrected ^{14}C yr BP. But due to the rather flat topography, most of Hiiumaa Island emerged from the sea during last several thousand years only. The shoreline of the regressive phase of the Litorina Sea L_{III} lies at Hiiumaa at an absolute height of 17–19 m and regressive phase L_{IV} 14–16 m above the sea level (Fig. 1), when in the central part of the contemporary Hiiumaa (989 sq. km) the island about 120 km² in area is formed (Sepp 1974). According to Sepp-Ratas (Sepp 1974) the shorelines of Lim_I are on Hiiumaa Island at a height of 11.8–13.6 m, Lim_{II} 9.5–10.6 m, Lim_{III} 6.5–8.5 m, Lim_{IV} 4.0–5.2 m and Lim_V 2.0–3.0 m.

The Vanajõe section

The Vanajõe section is situated in the south-western part of Hiiumaa Island (Fig. 1) at an absolute height of 12–15 m, where the Vanajõe River flows over the coastal and bottom deposits of the Litorina and Limnea seas (Fig. 3). The river banks, about 400 metres to the west of the Emmaste-Luidja highway expose medium- and fine-grained cross-bedded sands. The upper 4–5 m of the section are barren of molluscs, while several lower layers (after every 0.6–1.2 m) abound in subfossil brackish-water mollusc shells (mainly *Cerastoderma glaucum* – 59.6% and *Macoma baltica* – 19.1%). H. Kessel found in the section also the shells of *Littorina littorea* (14.2%), *Lymnaea peregra* (Müller) f. *baltica* Nilss. (4.8%), *Theodoxus fluviatilis* (L.). f. *littoralis* L. (2.3%), and a lot of broken fragments of *Mytilus edulis* L. The shell samples 085–089 and 086–089 were taken from the left bank of the river: the

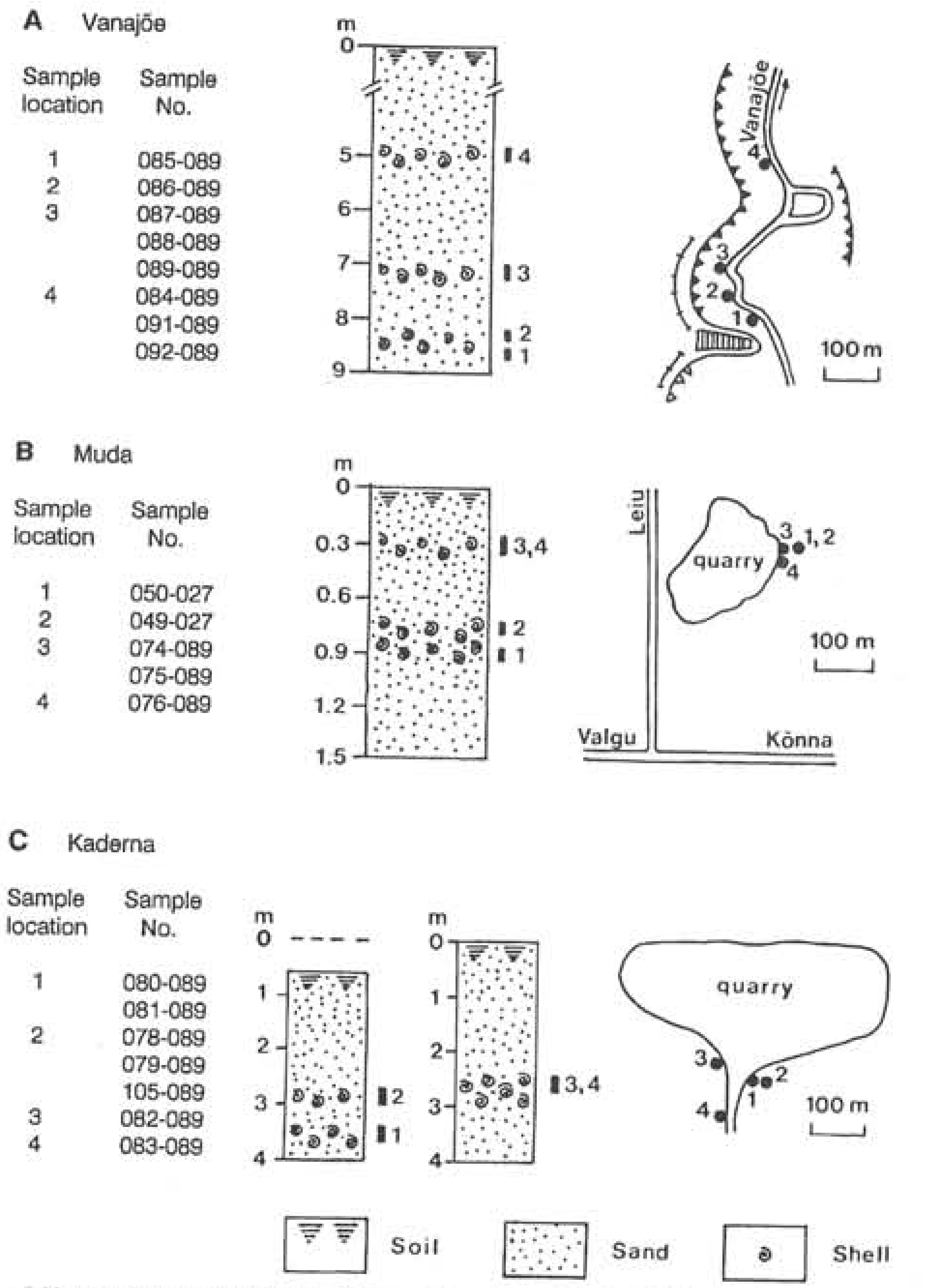

Fig. 3. Positions of Vanajõe (A), Muda (B) and Kaderna (C) sections with the indication of sample locations.

Table 1. **ESR-dating results on subfossil mollusc shells from Limnea Sea deposits**

Sample No.	Site	Depth (m)	Molluscs	ESR age (yr BP)
083–089	Kaderna	2.50	*Cerastoderma glaucum*	2940
082–089	Kaderna	2.50	*Cerastoderma glaucum*	2990
079–089	Kaderna	3.00	*Macoma baltica*	1960
105–089	Kaderna	3.00	*Cerastoderma glaucum*	2500
078–089	Kaderna	3.00	*Cerastoderma glaucum*	2590
080–089	Kaderna	3.60	*Macoma baltica*	2790
081–089	Kaderna	3.60	*Cerastoderma glaucum*	3270
076–089	Muda	0.30	*Macoma baltica*	2330
074–089	Muda	0.30	*Cerastoderma glaucum*	2700
075–089	Muda	0.30	*Macoma baltica*	2880
049–027	Muda	0.70	*Cerastoderma glaucum*	3880
050–027	Muda	0.90	*Cerastoderma glaucum*	4060
092–089	Vanajõe	5.00	*Macoma baltica*	1920
084–089	Vanajõe	5.00	*Macoma baltica*	2030
091–089	Vanajõe	5.00	*Cerastoderma glaucum*	3050
047–027	Vanajõe	7.00	*Cerastoderma glaucum*	3200
046–027	Vanajõe	7.00	*Cerastoderma glaucum*	3730
007–124	Vanajõe	7.00	*Cerastoderma glaucum*	3760
088–089	Vanajõe	7.30	*Macoma baltica*	3580
089–089	Vanajõe	7.30	*Cerastoderma glaucum*	3820
087–089	Vanajõe	7.30	*Cerastoderma glaucum*	4120
086–089	Vanajõe	8.30	*Cerastoderma glaucum*	4100
085–089	Vanajõe	8.50	*Macoma baltica*	4170
051–027	Viiri	0.40	*Cerastoderma glaucum*	980
052–027	Viiri	0.45	*Cerastoderma glaucum*	1020

Note: The precision of the ESR dates is typically arround 10 %.

Table 2. **Clusters of ESR-ages obtained on the Limnea Sea shell-bearing deposits from Hiiumaa Island**

No.	Site	Average ESR-age, yr BP
1	Kaderna	2350 ± 280
2	Kaderna	2970 ± 25
3	Kaderna	3030 ± 240
4	Muda	2640 ± 230
5	Muda	3970 ± 90
6	Vanajõe	1980 ± 270
7	Vanajõe	3700 ± 280
8	Vanajõe	4135 ± 35
9	Viiri	1000 ± 30

Note: The quoted errors represent the dispersion between the ages obtained from the several shell samples.

former at the water level and the latter ca. 20 cm higher. The dates are mutually consistent, giving an average age of 4135 ± 50 years BP. The next series of samples (087–089, 088–089 and 089–089) was taken 1.2 m above the water level, several metres downstream. Samples 007–124, 046–027 and 047–027 originate in the same layer, but these were collected from the other outcrops of the Vanajõe River some metres downstream (location 5, not shown in the Fig. 3). The average age of the samples is 3700 ± 300 years BP. Samples 084–089, 091–089 and 092–089 (2030, 3050 and 1920 years BP, respectively) were taken 3.5 m above the water level. If to omit an "outlier" with the age of 3050 years BP (sample 091–089) the average age of this series of the samples makes up 1980 ± 80 years BP.

The obtained ESR dates (see Tables 1, 2) are in a rather good agreement with the predicted geological age and show a clear dependence on the mode of occurrence. Thus, they enable one to establish not only the rates of sediment accumulation, but also the gradual fall of the sea level and retreat of the shoreline.

The Muda section

In the central part of southern Hiiumaa near Muda the Emmaste-Männamaa esker ridge of the Palivere stage (Raukas 1992) ends, and is bordered by descending beach ridges of limestone gravel. The esker was overwashed by the waves of the Litorina Sea, and the beach ridges formed during the ensuing regression of the Litorina and Limnea seas as waves reworked the sandy-gravel esker deposits. The Muda quarry lies at a height of 10 m above the sealevel and exposes sands and shingle up to 4 m thick. Sands are rich in subfossil shells *Macoma baltica*, *Cerastoderma glaucum* and other species typical of the Limnea Sea. The supposable non-corrected ^{14}C age is about 4000 years, because the deposits related to such absolute heights in southern Hiiumaa belong to the end of the Litorina Sea and the beginning of the Limnea Sea (Kessel & Raukas 1979; Sepp 1974).

Series of dates were obtained for two layers. One of those, about 0.2 m thick occuring between depths 0.93-0.65 m, consists of sand and shingles with an abundance of *Cerastoderma glaucum* shells. The sand in the lower part of the layer is coarse-grained, in the upper part it is fine-grained. Shells for ESR-dating were taken from both parts of the layer to check the resolution of the method. ESR-analysis yielded internally consistent ages: 4060 ± 370 and 3880 ± 360 years BP for the lower and upper parts, respectively. The average age of the layer is 3970 ± 130 years BP. Series of dates for the samples 074–089, 075–089 and 076–089 taken from the upper layer, occured at depths of 0.23–0.35 m, yielded an average age of 2640 ± 280 years BP (see Tables 1 and 2).

The Kaderna section

The Kaderna section is situated in the western part of Hiiumaa, north of the settlement of Nurste, at an absolute height of 6–8 m, showing the deposition of Limnea Sea sediments. The sand quarry, about 4 m deep, contains a lot of subfossil mollusc shells *Cerastoderma glaucum* and *Macoma baltica* typical of the Limnea Sea. From the bottom of the quarry two samples (080–089 and 081–089) yielded an average age of 3030 ± 340 years BP, samples 078–089, 079–089 and 105–089, situated 0.6 m higher – 2350 ± 340 years BP, and close to the same level from the other side of the quarry (Fig. 3) 2970 ± 40 years BP (samples 082–089 and 083–089). On the whole, the dates obtained are consistent with the predicted geological ages of the deposits.

The Viiri site

The youngest shells, analyzed in the present study, were taken in southern part of Hiiumaa Island near the Metsalauka–Sõru highway (Fig. 1). Deposits abundant of *Cerastoderma glaucum* shells were found here near Viiri farm some hundred metres from the modern seashore at an absolute height of 1–2 m. According to Kessel and Raukas (1979), these deposits may be attributed to the Limnea Sea phase Lim_V. The shells for dating were taken at depths of 0.45 m (sample 052–027) and 0.4 m (sample 051–027) from the surface. The sampling points were 2 m apart. The shells analyzed, yielded concordant dates of 1020 ± 100 years BP for sample 051–027 and 980 ± 100 years BP for sample 052–027. The results are consistent with the predicted geological ages of the deposits and stratigraphical position of these samples. Besides, the results of ESR analysis of these shells are encouraging and suggest that even shells as young as 300–400 years BP can be successfully dated by ESR.

CONCLUSIONS

The aim of the present work on ESR dating is to point out that the new dating method can be used in correlating Holocene deposits on the basis of both relative and numerical age. The results of the study confirmed a complex history of regressive phases of Limnea Sea in north-west Estonia, where a large variety of ancient and contemporary sea deposits, rich in subfossil mollusc shells were found. It gave a good possibility to check the reliability of the ESR datings of young Holocene shells. Most of dates on shells studied are in agreement with the stratigraphy and relative ages of the deposits. Revealed clusters of ESR-ages imply some sea-level fluctuations or other palaeoecological conditions around 4000, 3700, 3000, 2600, 2300, 2000 and 1000 years BP. On the other hand, the experiments made with young shell samples give every reason to hope that even very young (300–400 years BP) Holocene shells can also be successfully dated by ESR in addition to those from Pleistocene deposits.

ACKNOWLEDGEMENTS

The authors are grateful to Mrs. Kai Vares who provided us with some of the samples and helpful information, Mr. Rein Vaher for help in figure preparation, and Mrs. Helle Kukk for a thorough revision of the English text. The authors acknowledge financial support from the Estonian Science Foundation (grants 326 and 971).

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Formation and Distribution of Storm Layers in Western Baltic Sea Muds

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Milkert, D. & Werner, F. 1996: Formation and distribution of storm layers in Western Baltic Sea muds. *Baltica, Vol. 9,* pp. 36–50. Vilnius. ISBN 9986-615-05-4 / ISSN 0067-3064
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Storms are an important agent for sedimentary processes in the western Baltic Sea. Mud sediments were examined using box cores to evaluate the deposition and preservation of storm beds. According to the intensity, direction and duration a storm produces different types of storm layers. The storm layer of a "century storm" (due at New Year 1978/79) can still be mapped in Eckernförde Bay 10 years later. It is possible to establish a storm sedimentation model for this bay. The preservation of storm layers not only depends on high sedimentation rates reflecting direct physical influences but similarly on the lack of anthropogenic disturbance and restricted biogenic activity. Considering these relationships, the analysis of storm layers principally may provide a good tool to reconstruct the Holocene climatological history of the Baltic Sea.

Keywords: western Baltic Sea, Holocene, sedimentation, storm layers, climatic history.

INTRODUCTION

Storm-generated sediments as a product of high energy, and short time events have increasingly drawn attention to geologists. They are not only of great relevance for the dynamics of marine sedimentation processes, they also provide a useful tool for event stratigraphy in ancient sedimentary sequences (Ager 1974; Häntzschel & Reineck 1968; Aigner 1982; Duke 1990). The actualistic counterparts were analysed in various environments all over the world (e.g. United States Gulf coast of Mexico: Hayes 1967; Ball et al. 1967; Morton 1981, Snedden et al. 1988; Siringan & Anderson 1994; Beaufort Sea: Hill & Nadeau 1989; Great Barrier Reef: Gagan et al. 1988, 1990; German Bight, North Sea: Reineck et al. 1968; Gadow & Reineck 1969; Aigner & Reineck 1982).

During these studies, an increasing debate raised on the oceanographic processes generating storm layers. Noteworthy, the question about relationships between shore-parallel (e.g. geostrophic) and shore-normal and offshore directed flows stimulated the discussion (Duke 1990). It became obvious that more data from different environments were needed. In particular, little is known on storm layer formation and distribution in shallow adjacent seas as the Baltic. In this context, the relationships in shallow, tideless seas, as the Baltic, can give good contributions.

Another aim for our study was the recent discussion concerning a possible increase of storms and storm intensity due to greenhouse effects. An important factor in this context is to monitor the storm events as they are documented in the youngest, historical relevant sedimentary record. In addition storm layer may give indications on the significance of reworking and re-distribution of pollutants of the sediments, such as trace-metals (Lapp & Balzer 1993) and PAH's (Gerlach 1990).

In the Baltic Sea, only a few publications exist considering storm sedimentation in basins. Werner (1968) attributed laminated and sandy-silty layers to storm situations. Khandriche et al. (1986) were able to show that a century storm due in winter 1978/79 formed a fine-sandy, ripple like structure of 1 to 1.5 cm thickness in Eckernförde Bay. Niedermeyer et al. (1993) interpreted shelly and sandy layers in Greifswalder Bodden (a semi-enclosed bay in the southern Baltic) as storm layers.

In this paper we emphasize the following questions:

– How can storm layers be recognized?

– Which amount of sediment is reworked and redeposited by a single storm event?

– Can we determine a preferred hydrodynamic and meteorological setting for storms to originate storm layers? Does a storm have to last for a certain minimum period of time before it can be documented in the sediment?

– Can we recognize a Holocene climatic record?

Geological and Hydrographical Setting

Kiel Bay as a semi-enclosed, sub-rectangular basin represents the westernmost part of the Baltic, but typically many features and relationships of the larger basins of the Baltic Sea can be studied quasi in model character.

Hydrographically and geologically Kiel Bay is classified as an 'adjacent sea' with humid climate (Seibold et al. 1971), characterized by a sequence

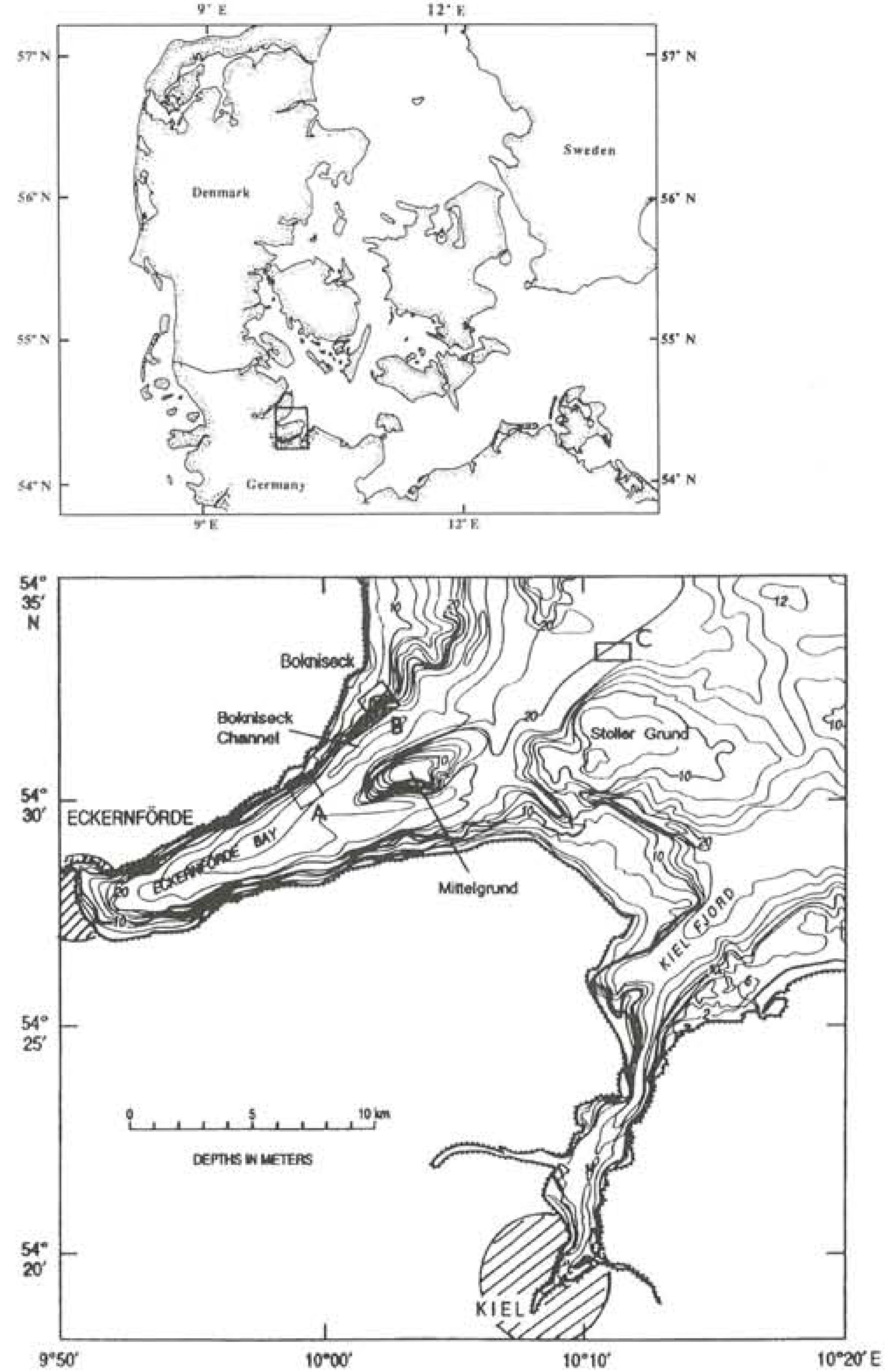

Fig.1. Generalized map of northern Europe with the investigation area in Eckernförde Bay (modified after Orsi et al., in press). **A.** Inner part of Eckernförde Bay, otter trawl fishery prohibited. **B.** Boknis Eck. **C.** NW Stoller Grund.

of basins with decreasing salinity as well as oxygen supply form west to east. Its origin results mostly by the Scandinavian ice sheet, which shaped the depression of the western Baltic (Gripp 1964). Glacial till (boulder clay) deposited during the Pleistocene forms the present coastline and covers the Prequaternary bedrock. In Kiel Bay a system of submarine channels was originated as sub-glacial channels (Seibold et al. 1971; Winn et al. 1982; Atzler 1995).

Currents and waves are the main transporting and eroding agents in Kiel Bay (Walden 1960). Kiel Bay has a minimal tidal range, but strong currents can be generated by wind forcing compensational currents, storm surges and seiches (Dietrich 1951). Wittstock (1982) suggested a dynamic resonance model for water exchange between the Baltic Sea and the Skaggerrak region. Current speeds of 20 cm/sec during inflow of surface water and 9 cm/sec, in 27 m water depth, 3 m above the seafloor, for outflow out of Kiel Bay during 'normal' periods are common (Geyer 1965).

Our investigation focused on sediments in Eckernförde Bay (Fig. 1), which was in the centre of interest for many research activities based at Kiel University during the last decades (e.g. Geyer 1965; Werner 1968; Dold 1980; Rumohr et al. 1987; Gerlach 1990; Krost 1990; Lapp 1991). The bay is divided by the sill "Mittelgrund" in a deeper northern channel with a maximum depth of 29 m, and a shallower southern channel with a maximum of 22 m water depth (Fig. 1).

Generally, a strong depth dependence of the sediments similar to other parts of Kiel (Werner et al. 1987) is observed. In detail the sediment-cover is variable in thickness and grain size. Glacial till, thinly covered with coarse lag sediments and alternating with coarse or medium sand is found from shallower than 10 m down to 16 m water depth, where wave action is ceasing. Below this level grain size is contiguously decreasing with an increasing amount of silt fraction (Werner 1968). Depending on the exposition, the transition to mud dominated sediments occurs between 18 to 22 m water depth. Below 22 m water depth fine mud dominates the central parts of Eckernförde Bay. The thickness of Holocene mud sediment reaches its maximum in the inner part of Eckernförde Bay with more than 10 m, the real thickness remaining unknown because of the strong acoustic turbidity, which masks the deeper strata in the seismic records (Atzler 1995).

Bottom currents in Eckernförde Bay are closely correlated to wind directions. Westerly winds generate surface outflow and compensated bottom water inflow from the open parts. Current speeds with a maximum of 55 cm/sec, 4 m above sea floor were observed west of Mittelgrund by Geyer (1965) (Fig.1) during a gale force 8 Beaufort from SSW as effect of such a compensation current into the bay. Easterly winds produce inflow of surface water and outflow of bottom water (Dietrich 1951; Rumohr et al. 1987) because of the larger fetch from east and therefore higher water levels. The hydrographic conditions show strong seasonal effects by development of a thermocline in summer and salinity layering throughout most parts of the year. Long periods with reduced oxygen content or hydrogen sulphate in the bottom water lead to reduced or even missing benthic biological activity. Intense ground fishery with otter trawls is common in the outer parts of the bay (Krost et al. 1989; Werner et al. 1990).

Weather System and Major Storm Events

The weather system in this area is strongly influenced by the Atlantic circulation system. Main wind direction is west during all parts of the year (Fig. 2). The main direction for winter storms is correlated to the major pathways of the Atlantic depressions (Defant 1974). Storms from west cause sea level low stands in the bay with almost no influence on the coastline because of the lee side. Strong storms from the eastern direction are less common, but create major damages on the coast because of their high water levels and on-land waves.

Regular and continuous daily weather observations started as late as 1870. Former historic storm events were normally only remembered due to the loss of human lives and the destruction, therefore the record of storm events could sometimes be incomplete. Wegemann (1911) tried to summarise the most important storm events between 1044 and 1908 AD. Increasing storm intensity since 1900 was recorded by Petersen & Rohde (1979), who reviewed the storm intensities through the last 100 years. Per definition all easterly storms, which reach a sea-level of more than one meter above zero are major storms. Figure 3A summarizes the water level high stands for major events between 1872 and 1989. Figure 3B shows all events between 1978 and 1990. The investigated storms which were likely to produce a storm layer are marked with an arrow. It can be seen, that there are several storm events with much higher sea levels, and even duration than those observed, and originating a storm layer, which do not effect the sediment.

Table 1 exhibits the observed storms in 1989 with respect to their influence on the sediments. It

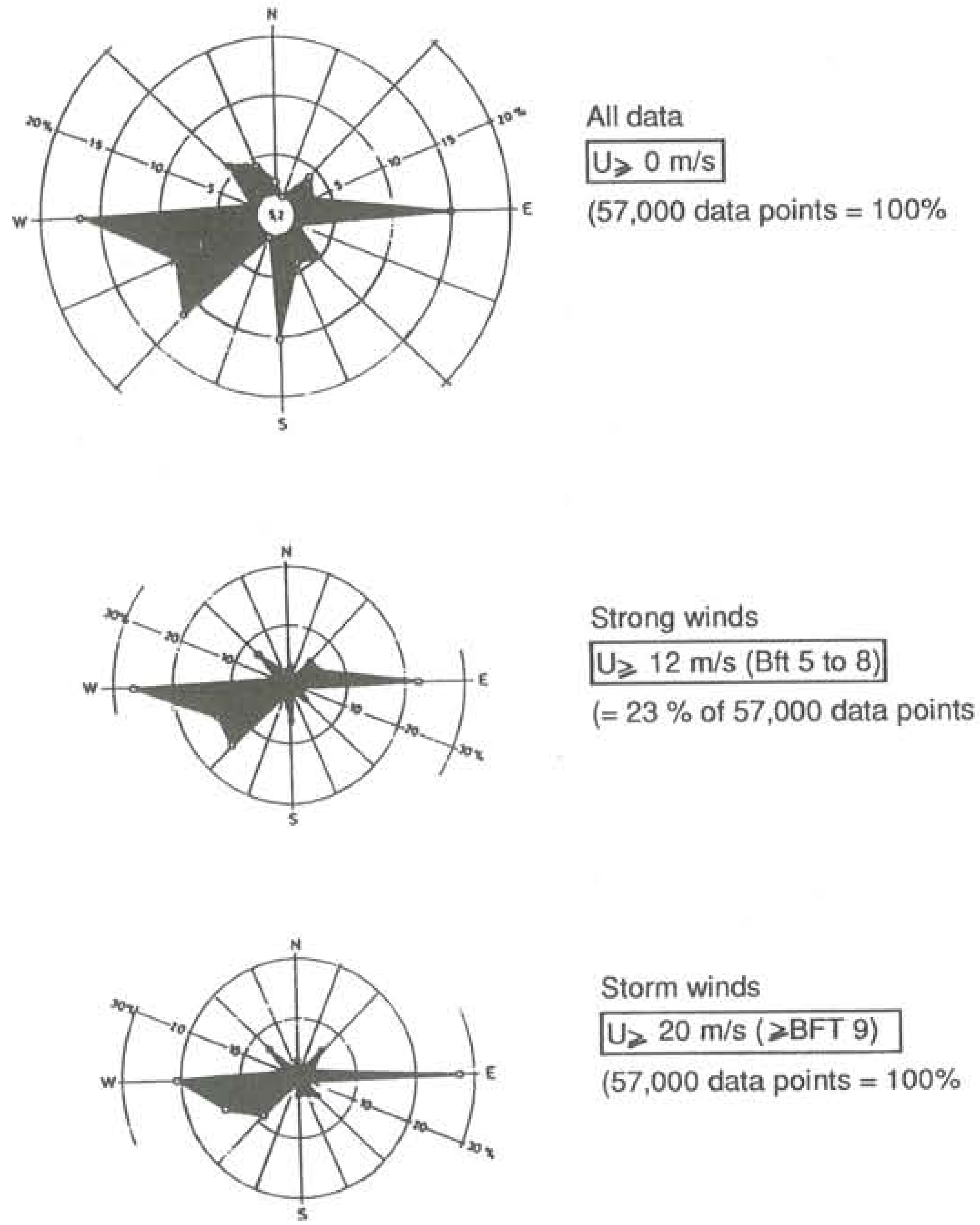

Fig. 2. Distribution of wind directions and force for Kiel Bay in percent after Eiben (1979), based on 57,000 single registrations.

may not be considered as belonging to the first range, when compared to the events in 1872 or New Year 1978/79, but they show a significant degree in intensity, sea level and duration.

During the observation period (1988–1991) three major storms (> 8 Beaufort, duration of high tide more than 10 hours) are documented, in April 1989 mainly from an easterly direction, in August 1989 from N/NE (Table 1), and in January 1990 from SW. The south-westerly storm in January 1990 lead to a sea-level low stand of –160 cm and had a duration of 18 hours (up to 8 Beaufort).

Predicted occurrence (MELF 1984) for sea level high stands caused by major storm events are as follows: > 285 cm (above normal sea level) – 1/150 years, >200 cm – 1/17 years, >175 cm – 1/9 years, > 150 cm – 1/6 years. This estimate, however, does not take into account the duration, which seems to be the most important factor for the formation of distinct storm layers in the sediment. Nevertheless it shows that there is a high of basins with decreasing salinity as well as oxygen supply form west to east. Its origin results mostly by the Scandinavian ice sheet, which shaped the depression of the western Baltic (Gripp 1964). Glacial till (boulder clay) deposited during the Pleistocene forms the present coastline and covers the Prequaternary bedrock. In Kiel Bay a system of submarine channels was originated as sub-glacial channels (Seibold et al. 1971; Winn et al. 1982; Atzler 1995).

Currents and waves are the main transporting and eroding agents in Kiel Bay (Walden 1960). Kiel Bay has a minimal tidal range, but strong currents can be generated by wind forcing compensational currents, storm surges and seiches (Dietrich 1951). Wittstock (1982) suggested a dynamic resonance model for water exchange between the Baltic Sea and the Skaggerrak region. Current speeds of 20 cm/sec during inflow of surface water and 9 cm/sec, in 27 m water depth, 3 m above the seafloor, for outflow out of Kiel Bay during 'normal' periods are common (Geyer 1965).

Our investigation focused on sediments in Eckernförde Bay (Fig. 1), which was in the centre of interest for many research activities based at Kiel University during the last decades (e.g. Geyer 1965; Werner 1968; Dold 1980; Rumohr et al. 1987; Gerlach 1990; Krost 1990; Lapp 1991). The bay is divided by the sill "Mittelgrund" in a deeper northern channel with a maximum depth of 29 m, and a shallower southern channel with a maximum of 22 m water depth (Fig. 1).

Generally, a strong depth dependence of the sediments similar to other parts of Kiel (Werner et al. 1987) is observed. In detail the sediment-cover is variable in thickness and grain size. Glacial till, thinly covered with coarse lag sediments and alternating with coarse or medium sand is found from shallower than 10 m down to 16 m water depth, where wave action is ceasing. Below this level grain size is contiguously decreasing with an increasing amount of silt fraction (Werner 1968). Depending on the exposition, the transition to mud dominated sediments occurs between 18 to 22 m water depth. Below 22 m water depth fine mud dominates the central parts of Eckernförde Bay. The thickness of Holocene mud sediment reaches its maximum in the inner part of Eckernförde Bay with more than 10 m, the real thickness remaining unknown because of the strong acoustic turbidity, which masks the deeper strata in the seismic records (Atzler 1995).

Bottom currents in Eckernförde Bay are closely correlated to wind directions. Westerly winds generate surface outflow and compensated bottom water inflow from the open parts. Current speeds with a maximum of 55 cm/sec, 4 m above sea floor were observed west of Mittelgrund by Geyer (1965) (Fig.1) during a gale force 8 Beaufort from SSW as effect of such a compensation current into the bay. Easterly winds produce inflow of surface water and outflow of bottom water (Dietrich 1951; Rumohr et al. 1987) because of the larger fetch from east and therefore higher water levels. The hydrographic conditions show strong seasonal effects by development of a thermocline in summer and salinity layering throughout most parts of the year. Long periods with reduced oxygen content or hydrogen sulphate in the bottom water lead to reduced or even missing benthic biological activity. Intense ground fishery with otter trawls is common in the outer parts of the bay (Krost et al. 1989; Werner et al. 1990).

Weather System and Major Storm Events

The weather system in this area is strongly influenced by the Atlantic circulation system. Main wind direction is west during all parts of the year (Fig. 2). The main direction for winter storms is correlated to the major pathways of the Atlantic depressions (Defant 1974). Storms from west cause sea level low stands in the bay with almost no influence on the coastline because of the lee side. Strong storms from the eastern direction are less common, but create major damages on the coast because of their high water levels and on-land waves.

Regular and continuous daily weather observations started as late as 1870. Former historic storm events were normally only remembered due to the loss of human lives and the destruction, therefore the record of storm events could sometimes be incomplete. Wegemann (1911) tried to summarise the most important storm events between 1044 and 1908 AD. Increasing storm intensity since 1900 was recorded by Petersen & Rohde (1979), who reviewed the storm intensities through the last 100 years. Per definition all easterly storms, which reach a sea-level of more than one meter above zero are major storms. Figure 3A summarizes the water level high stands for major events between 1872 and 1989. Figure 3B shows all events between 1978 and 1990. The investigated storms which were likely to produce a storm layer are marked with an arrow. It can be seen, that there are several storm events with much higher sea levels, and even duration than those observed, and originating a storm layer, which do not effect the sediment.

Table 1 exhibits the observed storms in 1989 with respect to their influence on the sediments. It

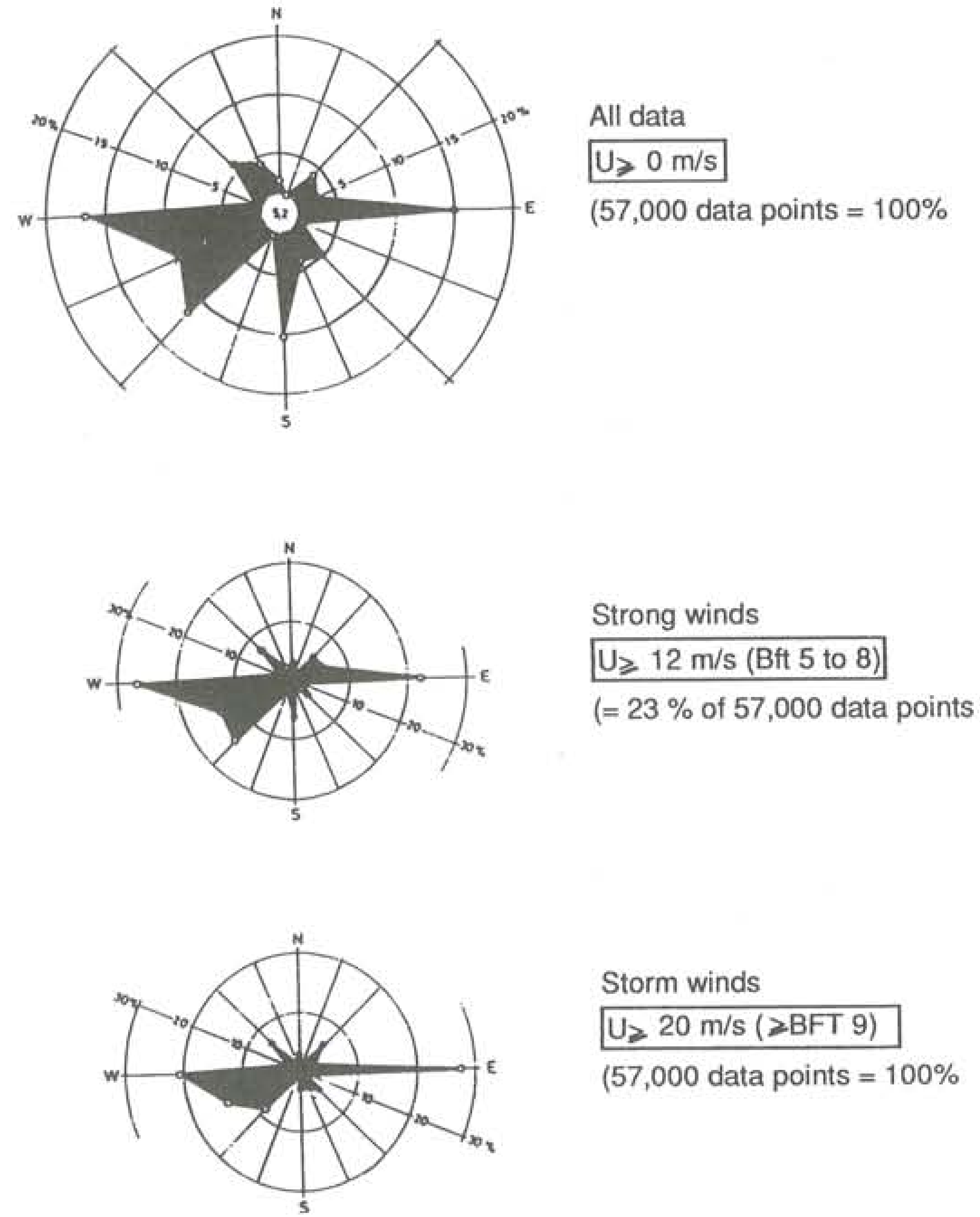

Fig. 2. Distribution of wind directions and force for Kiel Bay in percent after Eiben (1979), based on 57,000 single registrations.

may not be considered as belonging to the first range, when compared to the events in 1872 or New Year 1978/79, but they show a significant degree in intensity, sea level and duration.

During the observation period (1988–1991) three major storms (> 8 Beaufort, duration of high tide more than 10 hours) are documented, in April 1989 mainly from an easterly direction, in August 1989 from N/NE (Table 1), and in January 1990 from SW. The south-westerly storm in January 1990 lead to a sea-level low stand of –160 cm and had a duration of 18 hours (up to 8 Beaufort).

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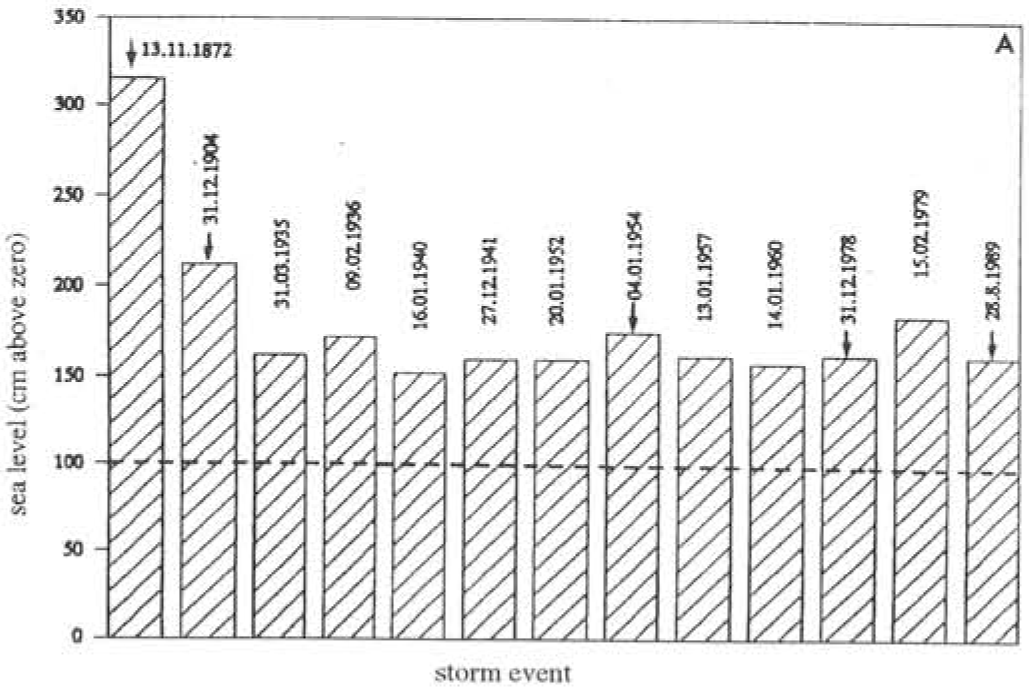

Fig. 3A. Water level gauges for main storm events in Kiel Bay for the period between 1872 and 1989. Arrows mark the investigated storms, and those who originated a storm layer.

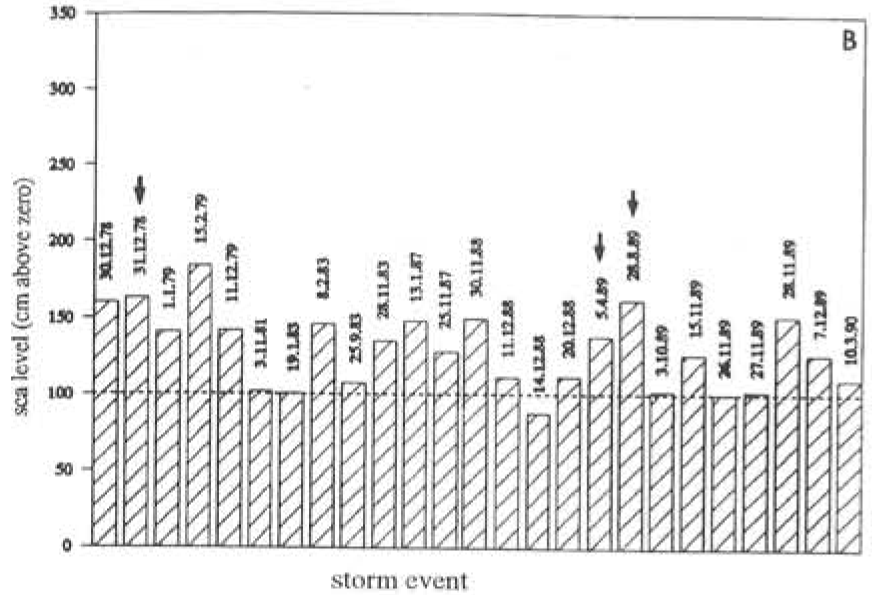

Fig. 3B. Water level gauges for main storm events in Kiel Bay between 1978 and 1989 where sampling of storm layer sequences occurred. Arrows mark the investigated storms.

Table 1. **Major easterly storm events in the last 100 years (after Pralle 1875; Führböter 1979; Klug 1986; and Eiben 1989)**

Period	1872	1904	1954	1978/79	1989 April	1989 August
Direction	NE	E	N/NE	N/NE	E	N/NE
Sea level (above NN)	+350 cm	+220 cm	+183 cm	+175 cm	+138 cm	+162 cm
Duration of high tide (>1 m)	44 hrs	?	24 hrs	72 hrs	12 hrs	12 hrs
Duration (>8 Beaufort)	48 hrs	?	?	48 hrs	38 hrs	36 hrs

possibility for the occurrence of sea level high stands during storm events.

MATERIAL AND METHODS

For our investigation (see also Milkert 1994) around 100 short box cores (maximum length 45 cm long) were analysed. To make sure that the surface sediment was undisturbed by the influence of ground fishery (Werner et al. 1990), the box corer equipped with a video camera. Sediment slabs for X-radiographs and samples for grain size analysis, radiocarbon dating and ^{210}Pb dating were taken.

Continuous bimonthly sampling was carried out in the inner part of Eckernförde Bay (Fig. 1, box A, between 24 and 26 m water depth) at Boknis Eck (Fig. 1, box B, between 23 and 28 m water depth) and Northwest of Stoller Grund (Fig. 1, box C, between 20 and 22 m water depth).

Navigation was done by using GPS and SYLEDIS as high resolution navigational tools with an inaccuracy of less than 30 m respectively 10 m. We use the local names and terms as given and approved by Babenerd & Gerlach (1987).

RESULTS

In particular three major storms (April 1989, August 1989 and January 1990) and their sedimentological impacts could be verified. Referring to the detailed bimonthly sampling we were able to recognize the effect of these storms.

Types of Storm Layers

According to the intensity, direction, and duration, storms produce different types of storm layers.

Observations after the east storm in April 1989 (Fig. 4a) showed erosion of sediment and a very thin (0.4 cm) layer of laminated mud, similar to the type A storm layer displayed in Figure 5A, was detected. Figure 4a shows the variability of the surface sediment after the April 1989 storm. A layer of coarse silt of approximately 0.3 cm was traceable in the deepest parts of the northern channel just at the sediment surface. This layer was soon reworked by benthic activity as well by ground fishery as revealed by subsequent sampling.

The north-north-east storm in August 1989 (Fig. 4b) deposited an approximately 1 cm thick layer of laminated mud, underlain by a 1– to 2–mm thick lamina of fine sand (see also Fig. 5A). This sequence could be mapped in the deeper parts of the bay, during a sampling campaign three weeks after the storm (Fig. 4b, GIK 13674 to GIK 13676). At the shallower positions, northwest of Stoller Grund (Fig. 1, box C; GIK 13677, GIK 13678) in 20 m water depth, the surface sediment was made up by a 1 cm thick fine sand layer. This may give a hint that the duration and/or probably the wave action on the coastline in April 1989 was not sufficient enough to build a storm layer, whereas in August 1989 we could observe the formation at all stations.

As an effect of similar order of magnitude, after the New Year storm 1978/79 Khandriche et al. (1986) were able to show a newly built sandy layer of 1 to 3 cm thickness partially showing lenticular or flaser bedding (Fig. 5B, Fig. 6).

According to these results, three distinct types of storm layers are distinguished. They are shown in Figures 5 a to c.

Figure 5a Type "A" represents laminated mud with silty bases from water depths between 24 to 28 m. The structure consists of a parallel lamination with cross-stratification of clayey silt to sandy silt. It is underlain by a fine-sand layer of ca. 1 mm thickness. This type was observed after the 1978/79 event by Khandriche et al. (1986), the thickness generally varies between 0.5 cm to 1 cm.

A second type (Fig. 5b, Type "B") is made up by very fine sand ripples, occasionally overlain by homogeneous mud (Fig. 5b). The ripples show either internal lamination or homogeneous structure depending on the distance from the shore. With increasing water depth and shore distance they show decreasing ripple height and decreasing sand content. It is possible to map this storm layer throughout most parts of the bay (Milkert 1994). The good preservation of this layer in the inner parts of Eckernförde Bay can be attributed to reduced biological activity and lack of otter trawl net fishery.

The third type of storm layer is a homogeneous coarse silt layer (Fig. 5c). Overlying the 1978/79 storm horizon, usually three of these 3 to 4 mm thick coarse silt layers with virtually no internal structure occur as they are shown in Figure 6.

Box core GIK 13680 (Fig. 6) gives a very good example for the upper sedimentary sequence (0–25 cm depth) in the inner parts of Eckernförde Bay, and includes all described types of storm layers. Using the ^{210}Pb technique for dating these sediments (Milkert 1994), we were able to date sandy layers for the last 100 years.

As a different type of event layer, the occurrence of a unique ripple structure of extremely small ripples ("crypto-ripples") in 21 m water

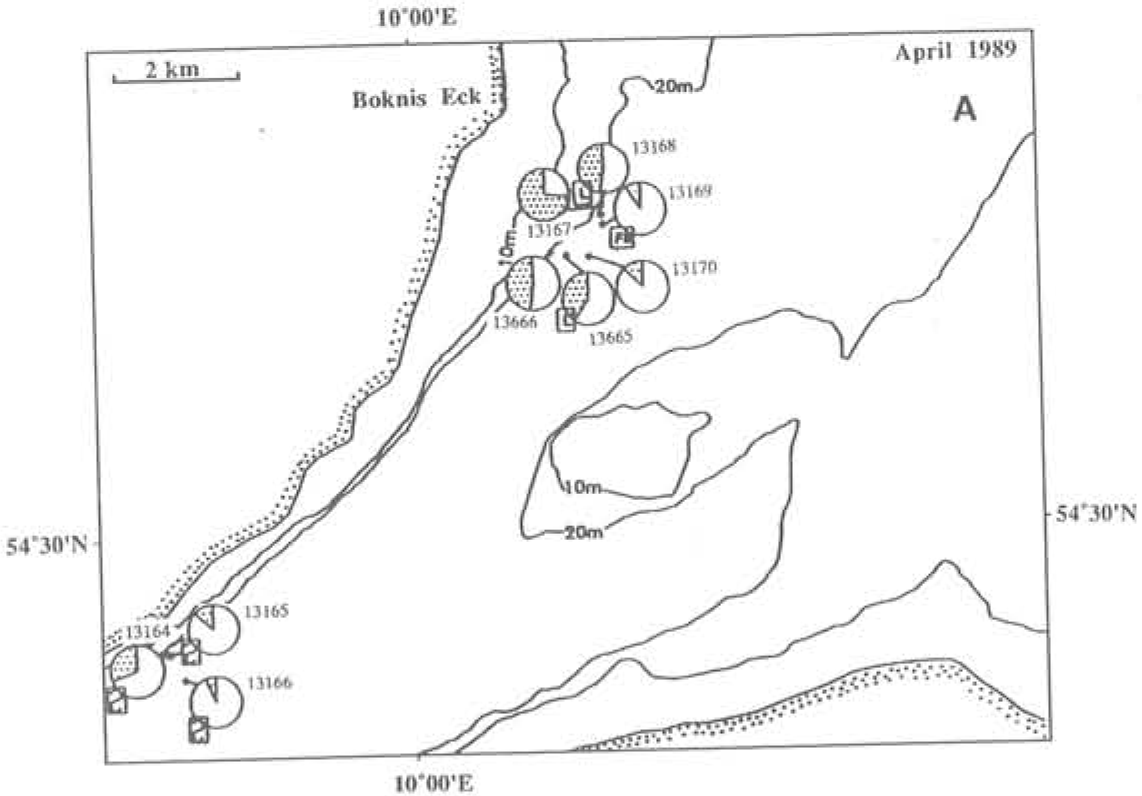

Fig. 4A. Sand content and surface layer lithology after the east-storm in April 1989 and after the north-north-east storm in August 1989. L = lamination, FS = fine sand layer, = bioturbed surface.

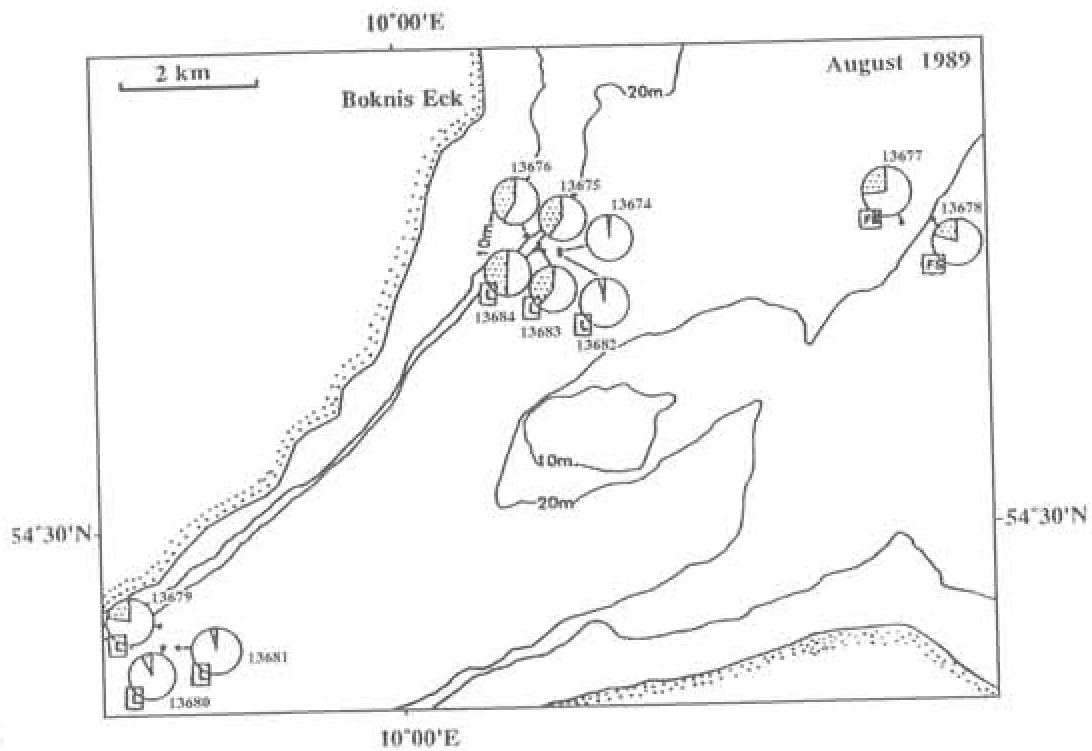

Fig. 4B. Sand content and surface layer lithology after the north-north-east storm in August 1989. L = lamination, FS = fine sand layer, = bioturbed surface.

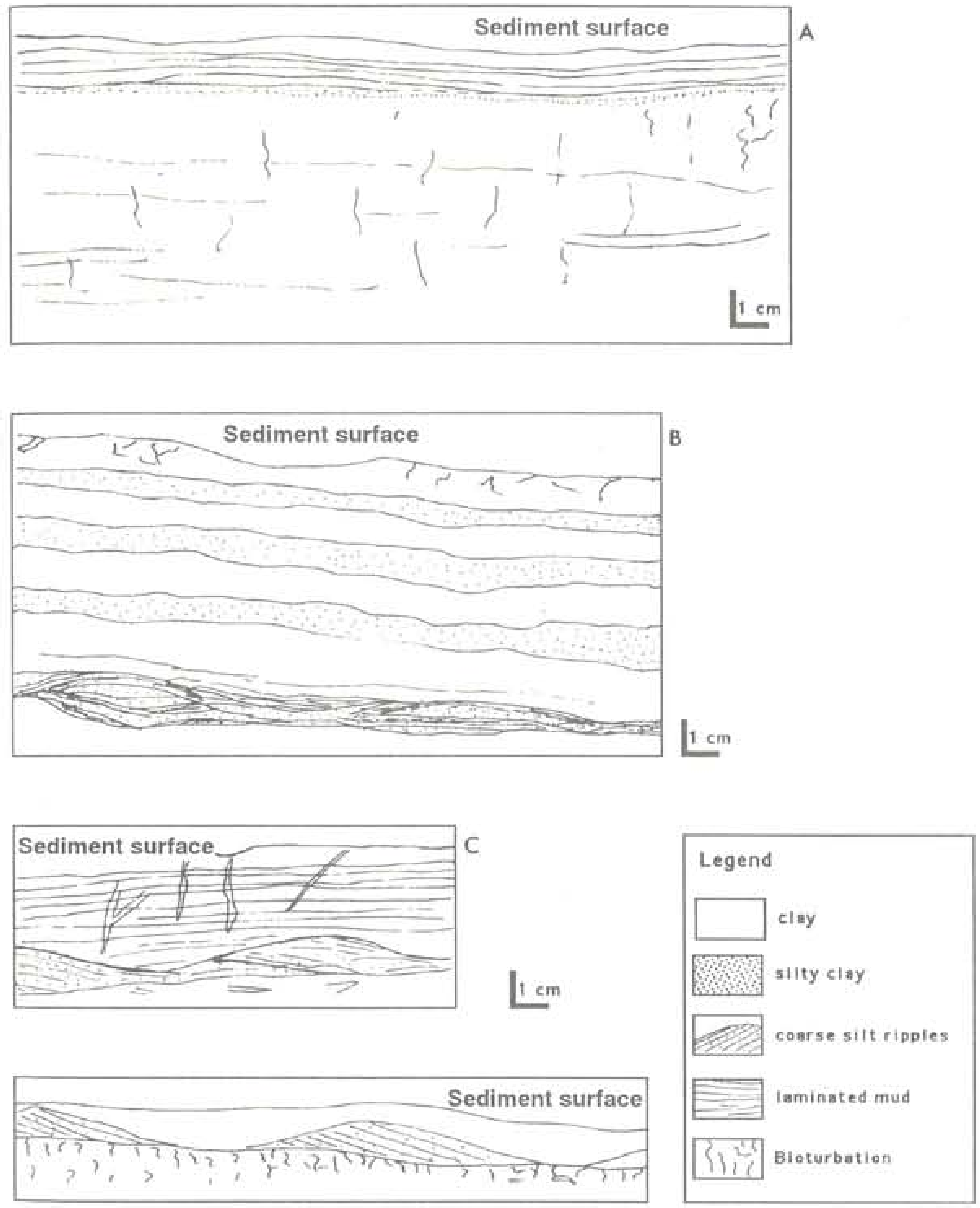

Fig. 5. Different types of storm layers as observed in Eckernförde Bay. A – the storm layer of **Type A** – laminated mud with coarse silt base as observed in Eckernförde Bay. B – the **Type B** storm layer which consists of homogenous coarse silt underlain by fine sand ripples as observed in Eckernförde Bay. C – the **Type C** storm layer, fine sand ripples with overlying laminated mud, as observed in Eckernförde Bay.

depth (for location see Fig. 1, box C) may be mentioned. The whole layer has a thickness of 4 cm (Milkert 1994). It appears to be a kind of slightly wavy lamination with discordant internal laminae (Fig. 7). The thickness of a single ripple layer varies between 0.1 cm at the base to 0.3 cm in the upper part of the sequence. The distance between single ripple crests varies between 0.6 cm and 0.7 cm. It is underlain by an 0.2 to 0.5 cm thick fine sand layer which overlies erosive the

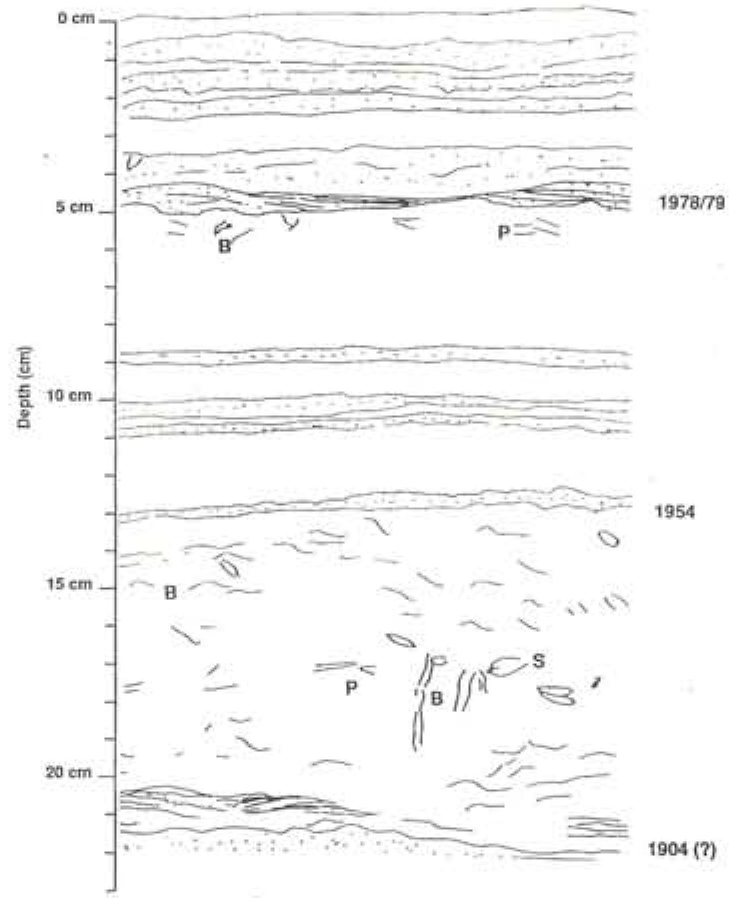

Fig. 6. Sedimentary structure of box core GIK 13680, taken in 26 m water depth with original depth scale. The dates are determined by ^{210}Pb and placed at the base of the observed coarser grained storm layers. B = biogenic structures, P = tubes of *Pectinaria korenii*, S = shells of *Abra alba*.

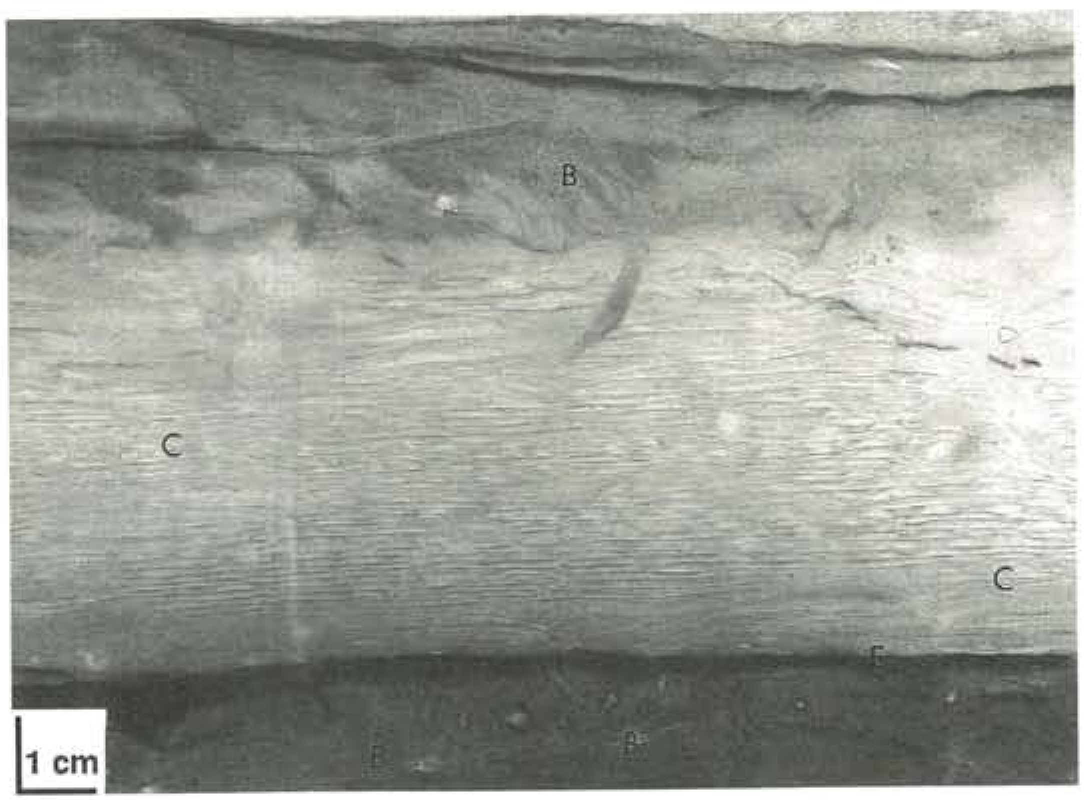

Fig. 7. Example for a complete "rypto-ripple"-sequence (x-radiography, positive, scale 1:1) of box core GIK 16549, 20 m water depth, northwest Stoller Grund (Fig. 1, C). C = crypto ripples, E = erosional base of sequence, B = biogenic structures.

intensely mottled basal mud. A similar type of ripple layer has been described first by Werner (1968) in about 1 m sediment depth from sediment cores taken 4 nautical miles north-east of the present position on the northern channel in 27 m of water depth. Preliminary dating shows that this layer NW of Stoller Grund is older than 1954. We assume that it is the product of a turbulent, unidirectional boundary layer flow on a weakly cohesive bed (due to high water content). Because of the structural constancy we suggest that the layer is the product of a single (storm) event. Further evidence gives the occurrence of an erosional layer at the base. Just a very few escape structures can be observed throughout the layer. Only the top parts of this structure are reworked by benthic fauna, but bioturbation from above never reaches the bottom.

Erosion and Resuspension After Storm Events

Since the amount of erosion is usually an unknown quantity in balance calculation, experiments to estimate the depth of erosion after a major storm were performed in different water depths (23 m, sandy-silty mud; 24 m, silty mud and 25 m, mud) in the outer parts of Eckernförde Bay (Milkert 1990; Milkert 1994). Baryte bearing, soluble tracer bars (similar to those described by Runte 1989) were inserted into the sediment by divers, based on the idea that eroded sediment thicknesses could be measured by the length of the remaining bars observed in x-radiographs.

The storm event in August 1989 eroded up to 3.3 cm of silty clay on the upper parts of the slope, with about 2 cm of silty clay settling at the same position. Other indications for the erosion of mud sediments during storm events were given by Smetacek (1980) by means of peak values of sediments collected in sediment traps. During stormy weather, very high values of resuspended material have been encountered in the water column (Milkert 1994; Smetacek 1980). Sedimentation of material is rapidly following the storm cessation. Material settled in winter shows a higher organic carbon content, and the proportion of humic acids to total organics is lower, as compared to the bottom sediment (Smetacek, 1980). This indicates that particle selection occurs during resuspension by the winnowing effect of turbulence due to bottom currents.

As effects of bottom currents because of the influence of storms, two phenomena on the bottom of Eckernförde Bay, can be mentioned: (1) Plant debris (twigs and leaves) together with beer cans accumulated in fishery trawl tracks as found by ROV observations (Werner et al. 1990); (2) Fields of parallel ripple-like features on muddy bottom with a spacing of ca. 2 m were mapped with high-resolution side scan sonar (Khandriche & Werner, in press), which probably likewise consist of organic debris and may be interpreted as transverse ribs (Allen 1984).

According to these observations we suggest a storm sedimentation model for Eckernförde Bay. It shows that mainly easterly storms cause transport of material into the deepest parts of the channel.

Distribution of Storm Layers

'Century storms', like the New Year storm 1978/79 produce a storm layer which could be recognized and mapped in most parts of Eckernförde Bay muds even after ten years still nearly undisturbed (Milkert 1994).

Figure 8 exhibits the X-radiographs of two box-cores, GIK 14471 (taken in July 1979) and GIK 16595 (taken at the same position in July 1988). In core GIK 14471 the newly deposited storm layer is visible on the sediment surface, whereas in GIK 16595 the storm layer can clearly be identified in 4.5 cm sediment depth. At the base of this layer an assemblage of tubes formed by the polychaete *Pectinaria korenii* occurs. This layer is attributed to a mass extinct during the hot summer of 1976 (Dold 1980). The *Pectinaria* bed can be found in nearly all parts of Eckernförde Bay (Khandriche et al. 1986; Milkert 1994). It seems possible that the conical tubes were eroded from the very soft mud during the New Year storm and swept together without being destroyed, thus, forming an oriented layer.

The recognition of storm layers (tempestite layers) in this case studies is based on lithological differentiation of coarse clastics (fine sand) within a silt/clay sequence. This holds also true for the definition of "tempestite" (Einsele & Seilacher 1982; Aigner 1982). In most cases a differentiation of sandy/silty layers visually discerned as coarser sediments within fine-grained series were the basis for observations. In fact, these observations are therefore related to a transport criterion. The physical properties being originated by bed load respectively suspension load material.

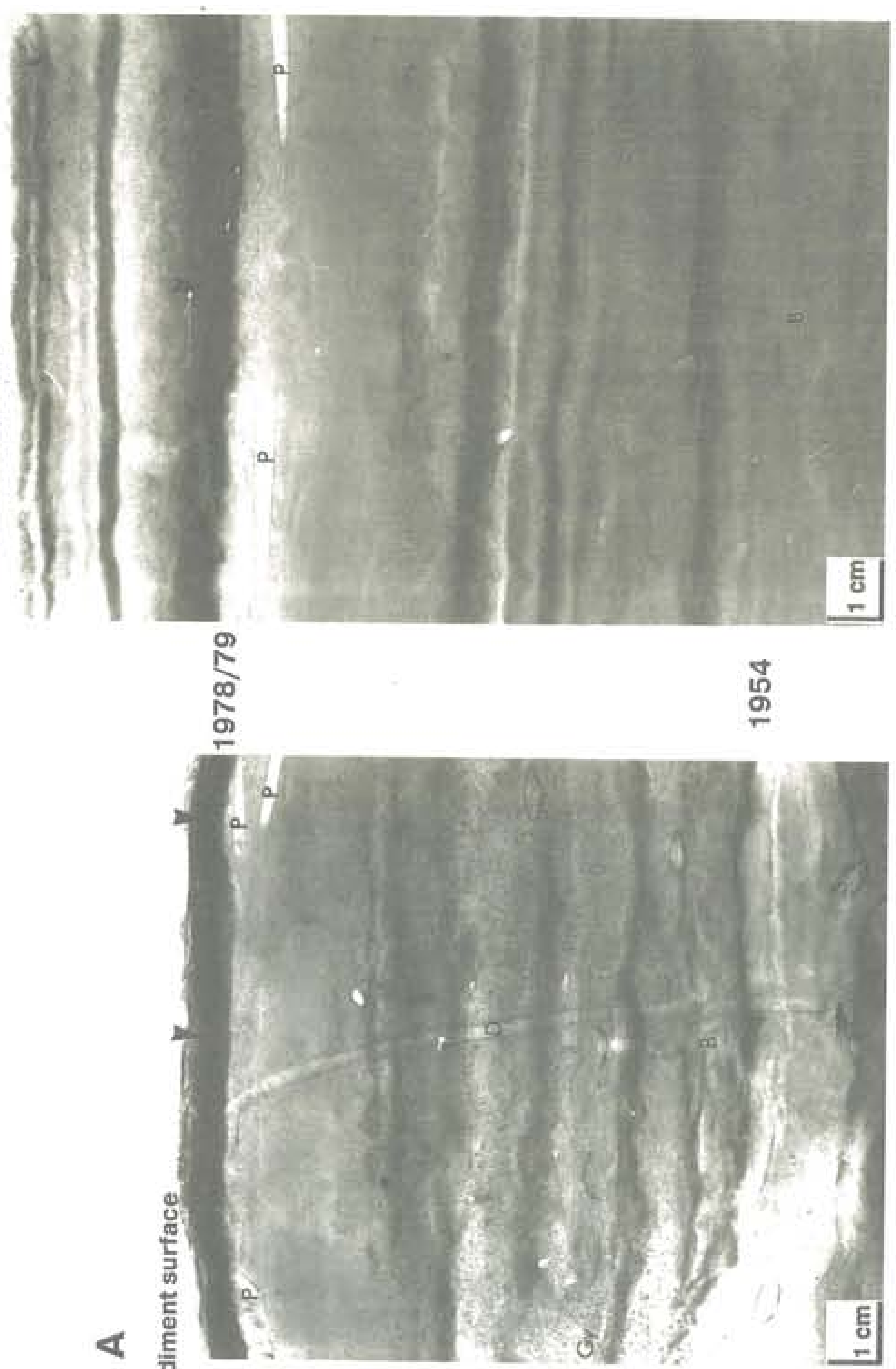

Fig. 8. Comparison between box core GIK 14471 (A), taken in November 1979 in the inner Eckernförde Bay and GIK 16595 (B), taken in August 1989 at the same position. Arrows mark the 1978/79 fine sand layer. Note the good correlation of the registered and dated storm layers. B = biogenic structures, P = tubes of *Pectinaria koreni*, S = shells of *Abra alba*.

A MODEL FOR STORM SEDIMENTATION IN ECKERNFÖRDE BAY AND DISCUSSION

According to the observations described above, a storm sedimentation model for Eckernförde bay is suggested (Fig. 9).

It shows that easterly storms cause water transport into the bay combined with sea level rise and high surface waves (Fig. 9a). This leads to strong wave action on the easterly exposed cliffs and beaches and to a corresponding material input which is transported to deeper water supported by down welling flow. By direct wave action, oscillation ripples are formed in greater depths than under normal conditions, i. e. about 22 m in the exposed parts of the bay. Accompanied by resuspension of material which takes place at the same location, and because of missing availability of non cohesive material, oscillation ripples can not be formed. Simultaneous bottom currents, developed by compensation of the inward transport at the surface, and generated in outward direction, may move the sand and silt sized material, and, thus form coarser, distinct layers, partly as current ripples. With ceasing storm intensity, the finer grain sizes and eventually organic material, e.g. plant debris are deposited, and a complete, graded storm layer sequence is formed.

On the contrary, with strong west storms (Fig. 9b) show outflow of surface water, and a significant drop of the sea level can be observed. In Eckernförde Bay this may expose at least the first, sometimes even the second offshore bar. As a consequence, some material might reach deeper water due to eolian transport (Werner 1968). Oscillation ripples are, of course, not activated because of the small fetch. As inferred from our observations after strong events, the corresponding compensation current at the bottom is commonly (perhaps with extreme situations?) not strong enough to deposit significant and surviving sediment layers, despite the cited measurements of Geyer (1965). This indirectly may give an additional suggestion of the effects of extreme easterly events. Considering our results in the light of the

Model for East Storm Situation

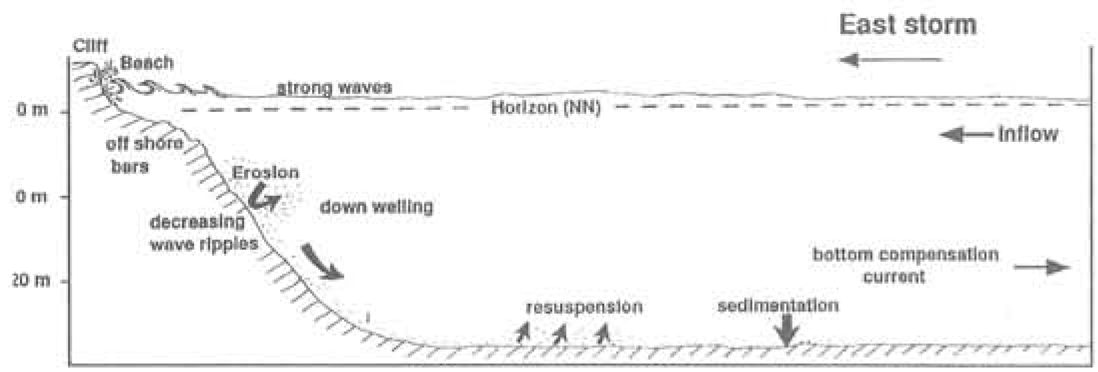

Fig. 9A. Storm sedimentation model for a fjord-like bay in the western Baltic Sea, generalised for Eckernförde Bay, which displays the situation for east storms with sea-level high stand and coastal set-up.

Model for West Storm Situation

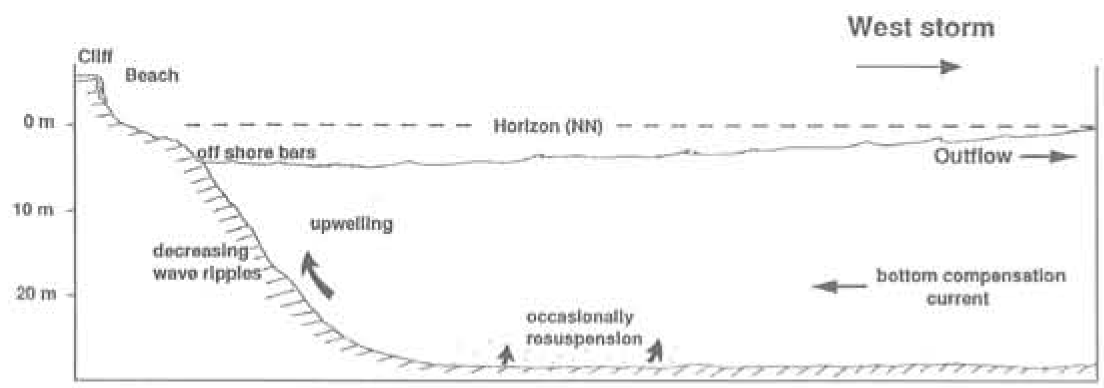

Fig. 9B. Storm sedimentation model for a fjord-like bay in the western Baltic Sea, generalised for Eckernförde Bay, which displays the situation for west storms with sea-level low stand and "wind ebb".

discussions on storm-sedimentation models in the literature (Hayes 1967; Walker 1984; Swift et al. 1985; Duke 1990) we have to emphasize that the geological effects we have observed are essentially connected with a "bay situation" which leads to relative strong bottom currents due to the high water stands and corresponding compensational bottom current.

Consequently, in many parts of the Baltic Sea as well in the deep sea, storm layer cannot be described. In other words, efforts for establishing a "storm stratigraphy" in Baltic Sea sediments are only reasonable for an area of favourable prerequisites of the transport model. This seems to be self-understanding, but it has to be emphasised that only restricted areas will fulfil these conditions.

In contrast to the model presented in this paper, stays the model proposed by Aigner & Reineck (1982) for tidal influenced sedimentary environments, i.e. the North Sea. Sediments are largely derived from the coastal sand and from tidal flats. They are partly transported in channels; by backflowing water, due to wind-induced gradient currents, these sediments are finally swapped onto the open shelf in a "jet-like" way. Storm sand sheets in this environment are most abundant beyond the mouth of the major tidal channels. Aigner & Reineck (1982) postulate that offshore transport does not require density currents. This is the major difference to the formation of storm layers in the (almost) tide-free western Baltic Sea, where most of the storm-induced sedimentation is correlated to density currents, and even to compensational currents flowing downslope (Fig. 9).

CONCLUSIONS

Storms are an important agent for sedimentary processes in the western Baltic Sea. Eckernförde Bay is considered as a good example for the formation of storm layers in similar fjord-like bays for the whole Baltic Sea. Therefore, a storm sedimentation model for Eckernförde Bay is suggested that emphasize the significance of sea level fluctuations and corresponding bottom current intensity correlated to the main wind direction. We can determine that storm layers in this bay are basically originated after east storm events, whereas west storms do not influence the sediment.

It can be summarized that the generation and preservation of storm layer sequences in the Baltic Sea requires: (1) water level rise during storms with near-bottom compensation currents and transport of sediment downslope; (2) mud basins where short coastal distances with steep slopes allow the supply of sand with grain-sizes fine enough to be transported as bed load of weak bottom currents; (3) restricted bottom fauna (as characteristic for many Baltic Sea basins) in order to avoid complete reworking by bioturbation; (4) little anthropogenic influence on the surface sediments.

Major storms from eastern direction could be recognized during the last 100 years, further investigations might lead to a Holocene storm stratigraphy.

ACKNOWLEDGEMENTS

This study and the resulting paper benefited by comments and enlightening discussions with several colleagues. For their valuable comments we would like to thank G. Hoffmann, A. Khandriche, A. Kuijpers, U. Hentschke and B. Lapp. We thank the Scientific Diving Group, Geol.-Pal. Inst., University of Kiel for their support during the erosion experiments. Special thanks are given to the master of RV Littorina, Captain V. Ohl, and his crew for never ending support during numerous hours at sea. This manuscript has benefited from the careful review and helpful comments of Józef E. Mojski and Gerd Hoffmann.

Funding of this work and a grant to D.M. was supported by the German Ministry for Research and Technology (BMFT, now BMBF) under the grants MFG 00686 and MFG 00969 and are greatly appreciated.

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Deglaciation of the Last Ice Sheet and Development of the Late Glacial Basins in the Southeastern Part of the Baltic Sea

Ints Veinbergs, Alexander Savvaitov and Vilnis Stelle

Veinbergs, I., Savvaitov, A. & Stelle, V. 1996: Deglaciation of the Last ice sheet and development of the Late Glacial Basins in the southeastern part of the Baltic Sea. *Baltica, Vol. 9*, pp. 51–56. Vilnius. ISBN 9986-615-05-4 / ISSN 0067-3064

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received 10th April 1995; accepted 1st September, 1995.

Six zones of ice marginal formations of the Last ice sheet are indicated by authors on the bottom of the southeastern part of the Baltic Sea. The ice marginal formations are correlated with Middle Lithuanian, Linkuva /North Lithuanian/, Plieņu and Valdemārpils Stages. Two younger lines of the ice marginal formations are also supposed. The Linkuva and Plieņu Stages are of the pre-Bölling age and the Valdemārpils Stage – of Older Dryas.The retreat of the ice margin from the line of Linkuva formations began about 13,000 years BP. The retreat of the ice sheet from the position of Plieņu formations occurred during Bölling and from the position of Valdemārpils formations – during Alleröd.The begining of the Baltic Ice Lake formation is related to Middle Alleröd.

Keywords: marginal formations, stages of reatreating ice margin, southeastern part of the Baltic Sea.

INTRODUCTION

The southeastern part of the Baltic Sea borders with lands of Russia, Lithuania and southwestern Latvia. The plateaus, the depressions and the other large and small forms of glacial and marine origin are distinguished in the bottom topography of this area of the Baltic Sea. The Liepāja, Klaipėda and Kurish Plateaus, as well as the Gotland and Gdańsk Depressions are the basic macroforms of the topography. The eastern slopes of Gotland and Gdańsk Depressions, the Klaipėda-Liepāja Saddle, the Gdańsk Swell, Palanga, Klaipėda and Kurish Lowerings and Nemunas Lowland are also seen clearly (Fig. 1, Veinberga et al. 1986). Some of enumerated forms of the bottom topography are of a lower class, than macroforms known from the latest regional maps of the Baltic Sea bottom and adjacent land areas (Gelumbauskaitė et al. 1990; Grigelis (ed.) 1991; Bjerkéus et al. 1995).

Locations of mentioned morphological forms correspond to the contours of the bedrock surface (Gudelis 1970, 1978). The Quaternary thickness is increased by dozen metres and more only in the deep incisions and depressions of the bedrock surface. The southeastern area of the Baltic Sea, as the Kurzemes Peninsula, was situated on the slope of the ice sheet with a rather small glacial accumulation on the plateaus (Aseyev 1974). The glacial formations, as a rule, were changed and sometimes even destroyed by wave activity of the Late and the Post Glacial basins and often buried. The marginal forms or their traces are not clear.

DISTRIBUTION OF ICE MARGINAL FORMATIONS

The next lines of the ice marginal formations are observed on the bottom in the southeastern part of the Baltic Sea (Fig. 2). The data given below are rather specific and supplement earlier conceptions about locations and ages of the marginal formations on the bottom in the southeastern part of the Baltic Sea (Serebryanny & Raukas 1966; Mörner et al. 1977; Gelumbauskaitė et al. 1990).

On the coast of the Sambian Peninsula there are the Sambian end moraines (Vaitekūnas 1972), which in Lithuania are related to the complex of the Middle Lithuanian /Vidurio Lietuvos/ marginal formations. The contours of the Sambian end moraines allow to consider that the ice sheet moving at this line was subdivided into Gdańsk and Kurish ice lobes. Large accumulations of the boulders spread in the coastal zone on the bottom of the Baltic Sea northwards from the towns of Yantarnyi and Svetlogorsk can be connected with

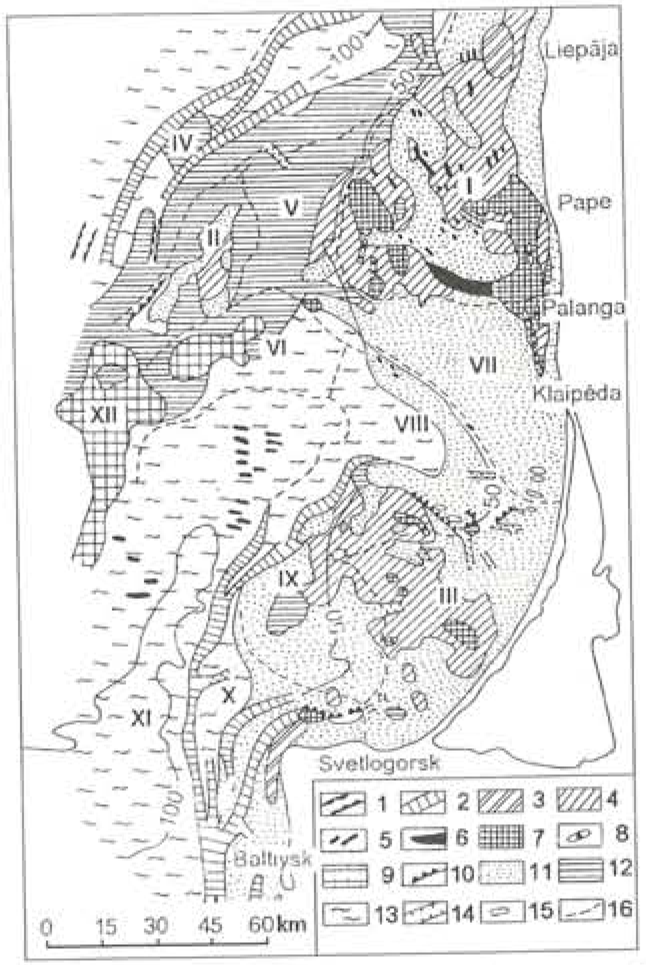

Fig.1. Geomorphological sheme of the southeastern part of the Baltic Sea.
I – Liepāja Plateau, II – Klaipėda Plateau, III – Kurish Plateau, IV – Eastern slope of Gotland Depression, V – Klaipėda-Liepāja Saddle, VI – Klaipėda Lowering, VII – Palanga Lowering, VIII – Nemunas Lowland, IX – Kurish Lowering, X – Eastern slope of Gdańsk Depression, XI – Bottom of Gdańsk Depression, XII – Gdańsk Swell.
The forms of the topography connected with the roughness of the bedrock surface: 1 – sub-Quaternary erosional incisions seen in the topography, 2 – cliffs and steep slopes, 3 – bedrock plains; the traces of the glacial topography forms: 4 – undulating plains of the basal till, 5 – drumlins, 6 – morainic swells formed in the zones of junction between the ice lobes, 7 – hills with the boulders on surface, 8 – esker-like forms, 9 – hills of the glacial topography unchanged by activities of the sea; the formations of the Late and the Post Glacial basins: 10 – cliffs, 11 – sandy plains of coastal-marine accumulation, 12 – glaciolacustrine plains formed by clays and silts, 13 – plains of deep sea accumulation formed by silts and mud, 14 – submerged valleys of the rivers, 15 – swells of uncertain genesis formed by sands and gravels, 16 – boundaries between macroforms of bottom topography.

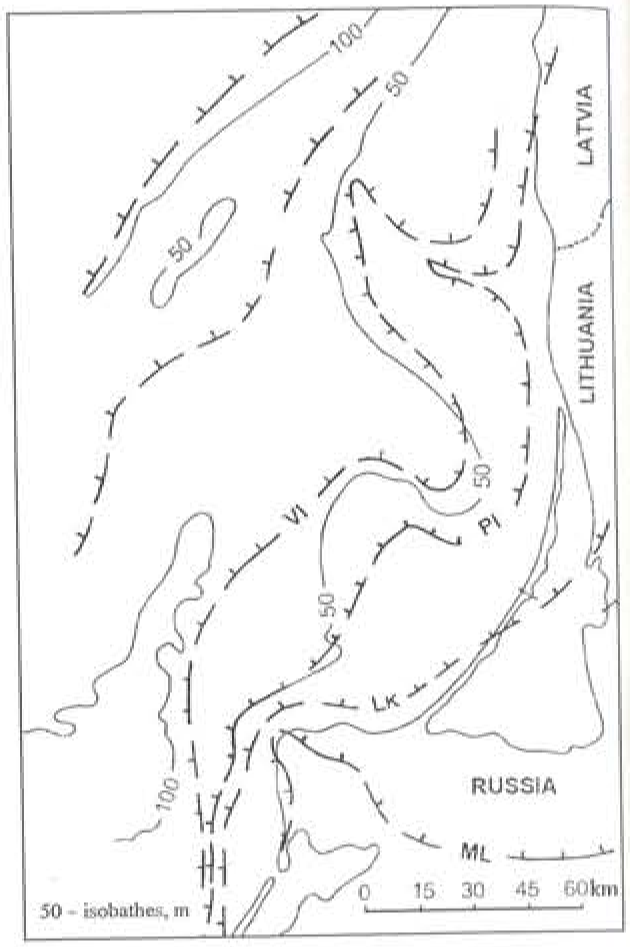

Fig. 2. Scheme showing the location of the ice marginal formations in southeastern part of the Baltic Sea. ML – Middle Lithuanian Stage /Vidurio Lietuvos/, Lk – Linkuva – North Lithuanian Stage /Šiaurės Lietuvos/, Pl – Plieņu Stage, VI – Valdemārpils Stage.

the ice marginal formations of the Middle Lithuanian Stage.

The marginal formations of Linkuva (North Lithuanian /Šiaurės Lietuvos/) Stage are also unclearly marked on the bottom of the Baltic Sea. The line of these marginal formation precipices goes from the land to the sea northwards from delta of the Nemunas River. Probably, the hilly relief on the southern part of the Kurish Plateau changed intensively by wave activity, as well as the precipices – marginal slopes (Veinbergs 1968, 1972) situated higher on the eastern slope of Gdańsk Depression westwards from the towns of Svetlogorsk and Baltiysk – can be considered as the ice marginal formations of the Linkuva Stage.

The remains of marginal formations located in the eastern and the southern parts of Liepāja Plateau on the sea bottom near the Linkuva zone on land (Gudelis 1964; Veinbergs 1968; Āboltiņš et al. 1972; Raukas et al. 1993) are believed to be of the Plieņu Stage. The submerged Papes–Palanga hilly forms and esker-like swells (Veinberga et al. 1986) with accumulations of boulders on their surface mark the position of frontal parts of the Liepāja and Nemunas ice lobes with some distinctions in orientations of ice flow. Probably, the Kunigiškiai morainic swell was formed between these ice lobes. The southern continuation of the Plieņu ice marginal formations is traced by hilly relief sections with boulders on the surface in the Kurish Plateau and the Kurish Lowering, as well as by the cliffs in the middle part of the eastern slope of the Gdańsk Depression.

Hilly formations of glacial topography in the southwestern part of the Liepāja Plateau are attributed to the Valdemārpils marginal zone. These are formed in the junction between the Liepāja and Nemunas ice lobes. Formations of the Valdemārpils ice marginal zone are not marked in the relief within the Nemunas depression, but southwestwards, it is probably represented by the lowest lying precipices of the eastern slope of the Gdańsk Depression.

Besides the characterized ice marginal zones in the southeastern part of the Baltic Sea, there are two more lines of ice marginal formations. The first one is represented by a hilly topography on the Klaipėda Lowering and Gdańsk Swell, with their surfaces covered by glaciolacustrine deposits. The second and younger morphological zone of the edge of the ice sheet is probably fixed at the eastern slope of the Gdańsk Depression.

The distribution and succession of all the zones of marginal formations allow to consider that the edge of the ice sheet, due to deglaciation, is located in the area of the southeastern part of the Baltic Sea and stretching from southeast to northwest.

During the first stages of deglaciation (Middle Lithuanian /Vidurio Lietuvos/ and Linkuva or North Lithuanian /Šiaurės Lietuvos/), the ice sheet was characterized by two ice lobes – Kurish and Gdańsk. The Liepāja and the Nemunas ice lobes were formed at the periods of Plieņu and Valdemārpils Stages. Later the mentioned ice lobes disappeared and the ice sheet edge became relatively straightforward in the plan and was situated along the slope of the Gotland Depression in a direction from southwest to northeast. At that time the ice sheet as a large ice lobe as lying in the Gotland Depression (Gudelis 1964; Serebryanny & Raukas 1966).

LOCAL GLACIAL LAKES AND MELTWATER STREAMWAYS

Wide distribution of glaciolacustrine deposits on the sea bottom is connected with widespread meltwater basins developed in front of the edges of the retreating ice sheet.

According to D. Kvasov (1975), during the Middle Lithuanian Stage, the Lower Nemunas ice-dammed basin was lying in front of the ice edge; the Lake level was at the altitudes about 80 m. The meltwater flowed westwards along the ice edge and reached the valley of the Vistula River and went farther to the ice-marginal valley of Toruń–Ebersvalds. In front of the retreating ice edge, the joint Nemunas–Gdańsk basin (level about 60 m) was formed, with waters running westwards through the valley of Reda–Łeba to the level 40 m. As a result of a subsequent retreat of the ice sheet in Poland, the level of the Nemunas–Gdańsk basin fell to 30 m bellow p.s.l. Later, as the margin of ice sheet retreated from the Slupsk bank, the Nemunas–Gdańsk basin joined to the Oder-Bornholm basin resulting the South-Baltic Ice Lake. According to D. Kvasov, the period when the Nemunas–Gdańsk basin was developed is related to the Middle Lithuanian and North Lithuanian Stages and the formation of the South Baltic Ice Lake – to Bölling.

However, D. Kvasov did not take into account the data about transgressions of the Late Glacial basins, which had taken place due to stage advancements of the ice sheet. These phenomena were established in Latvia and Lithuania (Basalykas 1967; Veinbergs 1968; Āboltiņš et al. 1974). The local ice-dammed basins and the South-Baltic Ice Lake varied in sizes and levels depending on the stages of the ice sheet advancements.

The ancient shore formations of the South-Baltic Ice Lake in the region of Sambian Peninsula occured on the bottom of the Baltic Sea at the depth of 70–75 m and 95–105 m (Blazhchishin 1981; Blazhchishin et al. 1982). Later the level of the South-Baltic Ice Lake rosed to the altitudes above p.s.l. and the general areas of the basin decreased, when the edge of the advancing ice sheet reached the boundary of the Plieņu marginal formations. At that period, meltwaters flowed down to the South-Baltic Ice Lake from the Upper Bārtas basin and the Apriķu basin through the valleys of Durbes-Vārtājas, Ālandes (Veinbergs 1968) and Erla-Salantas-Minija, Šventoji-Kulšė-Tenzė (Fig. 3) (Basalykas 1965, 1967; Gudelis & Klimavičienė 1982).

There was a large and deep regression of the South-Baltic Ice Lake, when the ice sheet edge retreated from the line of the Plieņu Stage. Probably, the South-Baltic Ice Lake was connected at that period with the Upper-Bārtas, Apriķu, Ventas-Usmas basins and with those in the region of the Gulf of Rīga, as well. The indications of this regression are fixed by a presence of incisions cut in the river valleys. Such the incisions were found in the Durbes-Vārtājas, Ālandes and Abavas-Slocenes valleys.

The transgressional stage developed when the ice edge advanced to the position of the Valdemārpils marginal formations, and the former basin connected to the South-Baltic Ice Lake was

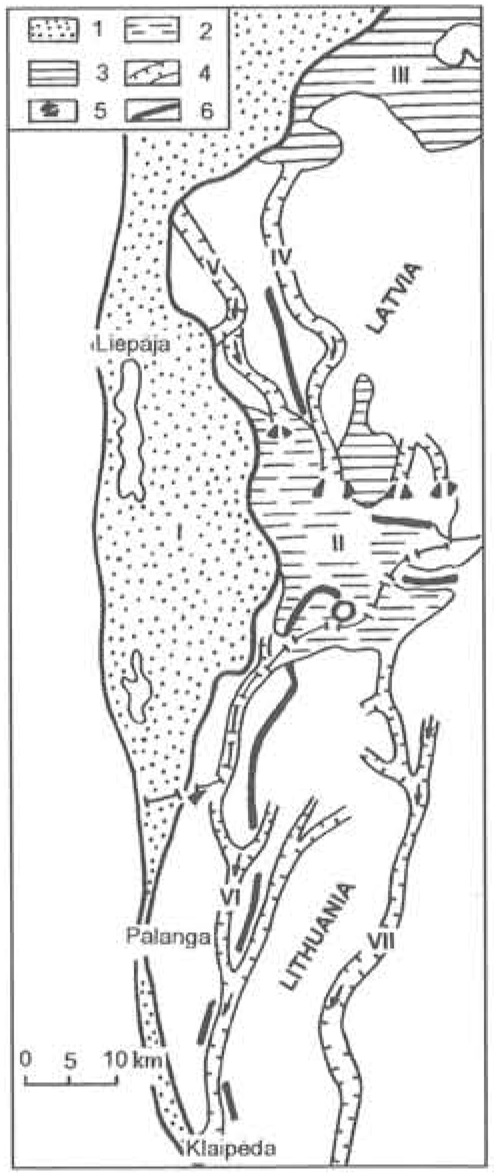

Fig. 3. Scheme showing the location of the glacial streamway valleys and glaciolacustrine basins in the northern part of the Lithuanian seacoast and in the southwestern part of Latvia. I – coast of the Baltic Sea, II – glaciolacustrine plain of the Upper Bārtas Lake, III – glaciolacustrine plain of the Apriķu Lake, IV – Durbes-Vārtājas valley, V – Ālande valley, VI – Šventoji-Kulšė-Tenzė valleys, VII – Erla-Salantas-Minija valleys. 1 – coastal plain, 2 – sandy glaciolacustrine plain, 3 – clayey glaciolacustrine plain, 4 – glacial streamway valleys, 5 – ancient deltas and outwash fans, 6 – ice marginal formations of Linkuva – North Lithuanian /Šiaurės Lietuvos/ Stage.

separated from the northern basins. The incisions in the Durbes-Vārtājas valleys were filled by glaciolacustrine clays to the altitudes of 15 m and those of the Abavas-Slocenes valley to 45 m due to that transgression (Veinbergs 1975). Probably, the final move of the South-Baltic Ice Lake water northeastwards and eastwards and confluence with the relicts of glaciolacustrine basins of Kurzeme and the Gulf of Rīga took place when the ice sheet retreated from the line of the Valdemārpils ice marginal formations.

AGES OF THE LATE GLACIAL BASINS, BALTIC ICE LAKE AND ICE MARGINAL FORMATIONS

The shorelines of the Baltic Ice Lake, well-kown and well-investigated in the Baltic (Grīnbergs 1957; Ulsts 1961; Gudelis & Emelyanov 1976; Gudelis & Klimavičienė 1982) cannot be regarded as the shores of the South-Baltic Ice Lake. The radiocarbon data for sands of the Baltic Ice Lake (Bgl II, Bgl III stages) in the area of the town Jelgava show the Alleröd age and the Younger Dryas age, as well (Veinbergs et al. 1980). The sands of the BgII stage of the Baltic Ice Lake on the coast of the Baltic Sea near Labrags northwards from Pāvilosta are of the age of Older Dryas (Veinbergs et al. 1980). The age of this stage is determined on the base of pollen data. However, not a long time ago, the composition of spores and pollen spectra in the middle part of Alleröd in the Late Glacial sediments form the Gulf of Rīga showed the colder spectra (Al 2), than in the lower (Al 1) and the upper (Al 3) parts of Alleröd (Stelle et al. 1990). The composition of these studied spectra for Middle Alleröd (Al 2) resembles that for spectra of Dryas and reflect the cooling climatic conditions during Middle Alleröd (Al 2), if compared to the conditions of Early (Al 1) and Late Alleröd (Al 3). The Older Dryas spectra near Labrags distinguished before by pollen data should be considered the spectra of Middle Alleröd now, since the begining of the development of the Baltic Ice Lake is determined to be attributed to the Middle Alleröd time.

The sediments of the Late Glacial basins in the southwestern part of the Baltic Sea, for intervals reached by sampling, were formed during Older Dryas, Alleröd and Younger Dryas. The spores and pollen zones of these periods have been established by investigations of V. Stelle, I. Jakubovska, A. Savvaitov, I. Timofeyev, A. Efimov and V. Bergman in the sections situated westwards and northwestwards from Klaipėda, i.e. in the zone of the Palanga Lowering and the Klaipėda-Liepāja Saddle, as well as on the slopes of Gdańsk and Gotland Depressions (Stelle et al. 1972; Stelle et al. 1975). Analogous spores and pollen zones were established by M. Kabailienė, O. Kondratienė, L. Lukoševičius, A. Blazhchishin and A. Gaigalas in the sections of the Gotland and the Gdańsk Depressions (Kabailienė et al. 1978). That zones are known for basin sediments in the Gotland Depression and in the Klaipėda–Liepāja submarine plateau (Kleimionova & Khomutova 1981). The spores and pollen zones of Older Dryas, Alleröd and Younger Dryas have been reported and characterized by G. Kleimionova, V. Gudelis, E. Vishnevskaya, L. Lukoševičius, H. Kessel, J. Zachowicz, K. Zaborowska as the basic palynological subdivisions for Late Glacial (Kleimionova et al. 1985).

The earlier basin sediments of Late Glacial on the coast of Latvia are represented by brown clay and varved clay. These sediments near Labrags northwards from Pāvilosta lie under the sandy deposits of the BgII stage of the Baltic Ice Lake. According to pollen data, the formation of the Late Glacial clay and especially its lower part in the section near Labrags took place at Bölling.

Moreover, the stratigraphical indices enumerated and regarded above in the Latvian land can be useful in determining age of the stage when the ice sheet has retreated. The radiocarbon data from the Raunis section show that the retreat of the the ice sheet edge from the line of the Linkuva ice marginal formations began about 13,000 years BP (Punning et al. 1968). The stratigraphical position of dated deposits in Raunis section are differently regarded. V. Stelle considered that these deposits were formed later than Linkuva Stage, and that the Plieņu ice marginal formations were formed in pre-Bölling.The retreat of the ice sheet from the line of Plieņu marginal formations took place during Bölling. This age does not contradict to the results of spores and pollen analysis done for the Līkaiņi section which show that the deposits have been formed under the conditions of dead ice (Meirons 1972). Probably, the spores and pollen spectra in this section regarded before as Younger Dryas and Alleröd in fact should be attributed to Older Dryas and Bölling. Younger Valdemārpils marginal formations were formed, when the ice sheet advanced during Older Dryas. The retreat of the ice sheet from the line of Valdemārpils formations took place during Alleröd.

In our opinion, during the late phases, the South-Baltic Ice Lake was marked out by a lower level, and probably that basin desintegrated into several local basins, which were connected by meltwater rivers. Later, at the Middle Alleröd time, a large transgression of the basin began, and as a result, the Baltic Ice Lake was formed during Middle Alleröd.

CONCLUSIONS

Marginal ice formations on the bottom of the south-eastern part of the Baltic Sea are mainly related to the Linkuva (North Lithuanian /Šiaurės Lietuvos/), Plieņu and Valdemārpils Stages. The Stages had been characterized by the advancements of the ice sheet edge. The development of the Late Glacial Basins occurred in front of the ice margin under the conditions of advances and retreats of the ice sheet margin.

The South Baltic Ice Lake formed when the ice margin retreated from the Slupsk Bank, later due to a disappearance of the ice sheet, was expanded northwards and northeastwards and even reached the Gulf of Rīga. During the ice margin advancements, the South Baltic Ice Lake desintegrated into several ice dammed basins.

The position of the ice margin at the boundaries of the Linkuva and Plieņu marginal formations is related to pre-Bölling and that on the boundary of the Valdemārpils formations – to Older Dryas.

ACKNOWLEDGEMENTS

The authors are indepted to Prof. Algirdas Gaigalas, Dr. Živilė Gelumbauskaitė (Lithuania), Prof. Anto Raukas (Estonia) for a critical reading of manuscript and important advices. The authors would like to express their thanks to colleagues Irina Jakubovska, Renāta Riņķe and Indra Federe for their help.

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Structure and Lithology of the Beach Ridge of the Baltic Ice Lake in Lithuania

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Gaigalas, A. 1996: Structure and lithology of the beach ridge of the Baltic Ice Lake in Lithuania. *Baltica, Vol. 9*, pp. 57–64. ISBN 9986–615–05–4 / ISSN 0067–3064.

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The beach ridge of the Baltic Ice Lake in Lithuania is about 200–400 m wide and 1.5–2.5 m high. Two layers of sand and gravel sediments have been distinguished in sections of the beach ridge, where horizontal bedding of sediments has been noted.The granulometric and mineralogic-petrographic composition of the sediments as well as the shape and roundness of the pebbles and orientation of the clasts in the beach ridge are discussed. The low sphericity, high flatness, oblongness and anisometricity of the pebbles as well as the roundness of the sand grains are typical for the coastal sediments of the Baltic Ice Lake in Lithuania.

Keywords: Baltic Sea, Baltic Ice Lake, beach ridge, lithology.

INTRODUCTION

The Lithuanian Baltic sea coast is of a low, straightened, accumulative type. The width of the Lithuanian coastal zone (to the upper limit of the extent of the former sea basin) reaches 5–6 km in the northern part (Palanga – Šventoji), a few hundred meters to the north of Klaipėda, and ca. 2–3 km on the northern outskirts of the Kuršių Marios Lagoon (Priekulė – Dreverna) (Gudelis 1979).

The highest shoreline of the Baltic Ice Lake lies at 12–13 m above the present sea level (Palanga – Šventoji). In the northern part of the coast two shore lines and terraces of the Baltic Ice Lake (BG I and BG II) and the Litorina Sea (Lit_3, Lit_4) are established. In the southern part only one accumulative Litorina terrace (Lit_3) could be traced (Gudelis 1955).

LOCATION OF THE BEACH RIDGE

The investigated beach ridge (BG II) extends inside the present shore of the Baltic Sea from Palanga across the Šventoji river to Latvia along the Palanga – Liepaja highway (Figs. 1 and 2, left side). It was formed by a transgression of the Baltic Ice Lake in the Younger Dryas period. This beach ridge complex consists of some minor shore ridges joined together and is built up of coarse-sand and gravel layers interbedded with pebbles (Fig. 2). The beach-bar deposits are distinctly separated from the erosional surface of the underlying strata. No organic layers in the beach-ridge sediments have been discovered. The top of this beach ridge has height of 13.0–12.5 m above sea level in the northern part. In the neighbourhood of Palanga, the height above sea level exceeds 10–11 m.

The beach ridge complex is 200–400 m wide and 1.5–2.5 m high. Due to small height, it is difficult to be distinguished from the surrounding topography. On both sides of it there are terraced plains: a rather wide terrace (BG I) of the Baltic Ice Lake from the east and fragments of the Baltic Ice Lake terrace (BG II) and the Litorina terrace (3–6 m above sea level) from the west (Gudelis 1979).

The slopes of the ridge are asymmetric; the western slope is steeper than the eastern one, being of 8–10° and 3–5°, respectively (Mikalauskas & Gaigalas 1973). The western slope has been formed as a micro-cliff.

The terraces BG I and BG II of the Baltic Ice Lake and the beach ridge are most probably the oldest shore formation of the Baltic basin on the Lithuanian coast. The surface of these terraces lies at 13–14 m above sea level from Šventoji to Laukžemė and 10–11 m to the south of Palanga. The terrace surfaces are slightly inclined to the south and west due to isostatic uplift. BG I and BG II are on the average 3–4 m thick and built up predominantly of sands of various grain size and gravel-pebble material. The Baltic Ice Lake material macroscopically does not differ from

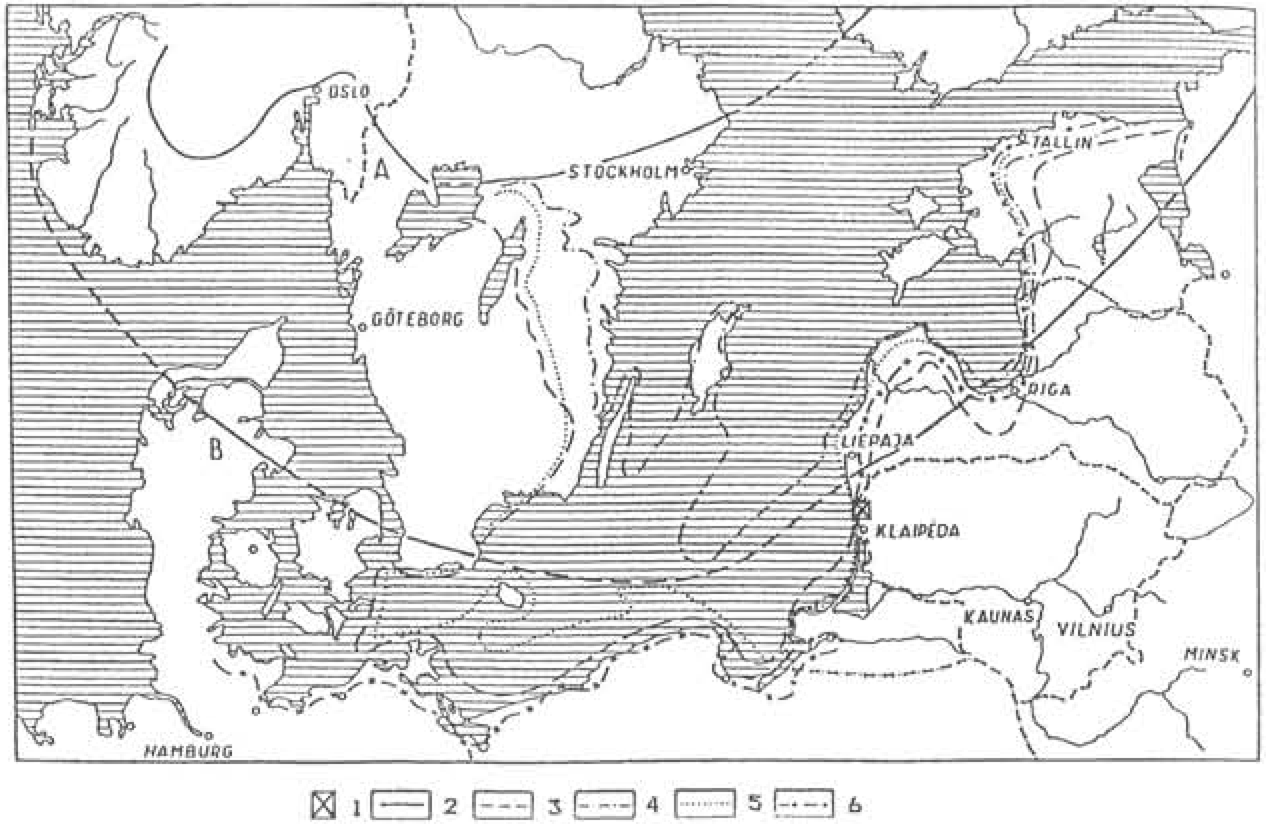

Fig. 1. Location of the studied area and ancient coastal lines in Peribaltic region (after Gudelis, 1961). 1 – studied area, 2 – hinge lines of M. Sauramo: A – Finiglacial, and B – Gotiglacial, 3 – Baltic Ice Lake, 4– Ancylus Lake, 6 – Litorina Sea.

glaciofluvial sediments and lies on the till or melt-water deposits of last glaciation.

STRUCTURE AND TEXTURE OF SEDIMENTS

The beach ridge BG II is made of facially equal gravel-pebble and gravel-sand layers inclined slightly (5°) seawards (Fig. 3 and 4). The horizontal lamination of these layers is accentuated by interlayers of various-grained sand, gravel, pebble, cobble and boulder deposits.

The long axes of pebbles (as well as the layers) are inclined 15° westwards (Fig. 2, right side). Moreover, these axes are perpendicular to the former shore line.

The processes determining the orientation of the pebbles are to be taken into account when reconstructing the conditions of the deposition. The orientation of long axes of pebbles in the beach sand is formed by ongoing and retreating waves above and under the actual mean water level (Gaigalas 1982). The predominant orientation depends on the pebble size and sea conditions (Reineck & Singh 1975; Ruchin 1958). Small pebbles usually acquire the orientation of long axes perpendicular to the shoreline, whereas big pebbles exhibit such a orientation only during stormy seas. The processes of dragging and rolling cause the orientation of pebbles. Strong flow drags the pebbles over sand surface and their orientation becomes perpendicular to the shoreline. Small pebbles can be moved more easily, thus their axes become oriented perpendicularly to the shoreline. On a dry surface of the beach ridge the pebbles are oriented by deflation. The orientation of long axes depends on the direction of the wind carrying away sand from below the pebbles.

A detailed study of the lamination and other structural peculiarities of material provides valuable information about the sedimentation conditions of the Baltic Ice Lake. The coarseness of clasts changes more quickly through a vertical section of the Baltic Ice Lake than in glaciofluvial deposits.

The sediments of Baltic Ice Lake are somewhat better sorted than the glaciofluvial deposits (Jurgaitis & Juozapavičius 1989). This is caused by the more homogenous hydrodynamic environment of sedimentations. The character of distribution of clasts by sand-gravel-pebble fractions in the Baltic

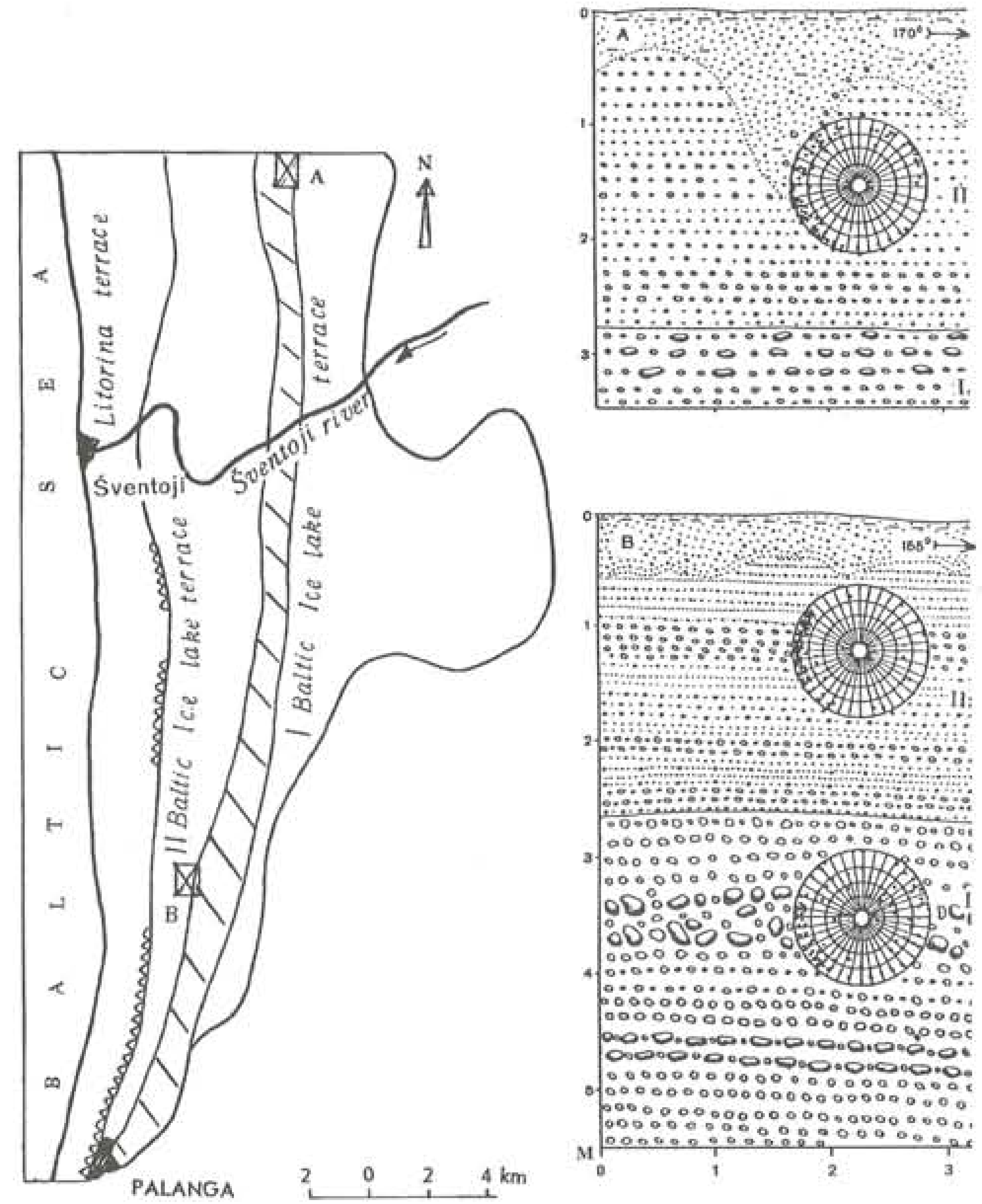

Fig. 2. Location of the study of ancient-shore formations (left side) and the examined sections of beach ridge (BGII) (right side): A – Laukžemė, B – Kunigiškiai. Point – diagrams of the long axe orientation of pebbles in coarse material sediments of two facially equal cobble-pebble-gravel (I) and gravel-sand (II) layers. The arrows indicate the azimuth of sections.

Ice Lake sediments is similar to that of glaciofluvial deposits.

In granulometric exhibits we can see a multimodal distribution of size fractions (Fig. 5). The Baltic Ice Lake sediments were formed by the reworking of glacial and glaciofluvial sediments and do not differ from them by their granulometric composition. The coastal sediments of the periglacial lakes facies consist of gravel (40%), pebbles (33%) and sand (26%), the sand fraction reaching up to 40% in some sections (Jurgaitis & Gaigalas 1993). The medium grain size is about 5 mm, with sorting coefficients ranging from 0.75 to 0.86. The character of the distribution of fragments according to sand-gravel-pebble fractions in the Baltic Ice Lake sediments is similar to that of glaciofluvial deposits. A uniform content of gravel-pebble fractions is a very significant feature (Fig. 5).

In the shore ridge of the Baltic Ice Lake many of the gravel and sand deposits were prospected and exploited at different times for production of building materials. A concentration of heavy minerals is observed in the beach sediments of the Baltic Ice Lake shore formations in Lithuania. According to granulometry, most coarse material is deposited halfways of the length under study, whereas in Palanga and Laukžemė the deposits

Fig. 3. The section at the Kunigiškiai. The cobble-pebble-gravel sediments (BGI) are exposed in the lower part (I), just below the beach ridge (BGII) gravel-sand layer (II). Photo by A. Gaigalas, 1973.

Fig. 4. The cobble-pebble-gravel sediments (lower layer) in the Kunigiškiai section. Photo by A. Gaigalas, 1973.

are finer. Such a distribution suggests that abrasion and accumulation were different in different sections of the Baltic Ice Lake shores. Moreover, the shore was gradually changing into an accumulation, and the northward transport of material prevailed. The ancient beach placers of heavy minerals (ilmenite, magnetite, rutile, leucoxene, zircon, garnets, etc.) were generated as a result of the activity of a whole complex of natural processes, both of denudation (erosion, deflation) and of accumulation character, acting in the littoral zone of the southeastern

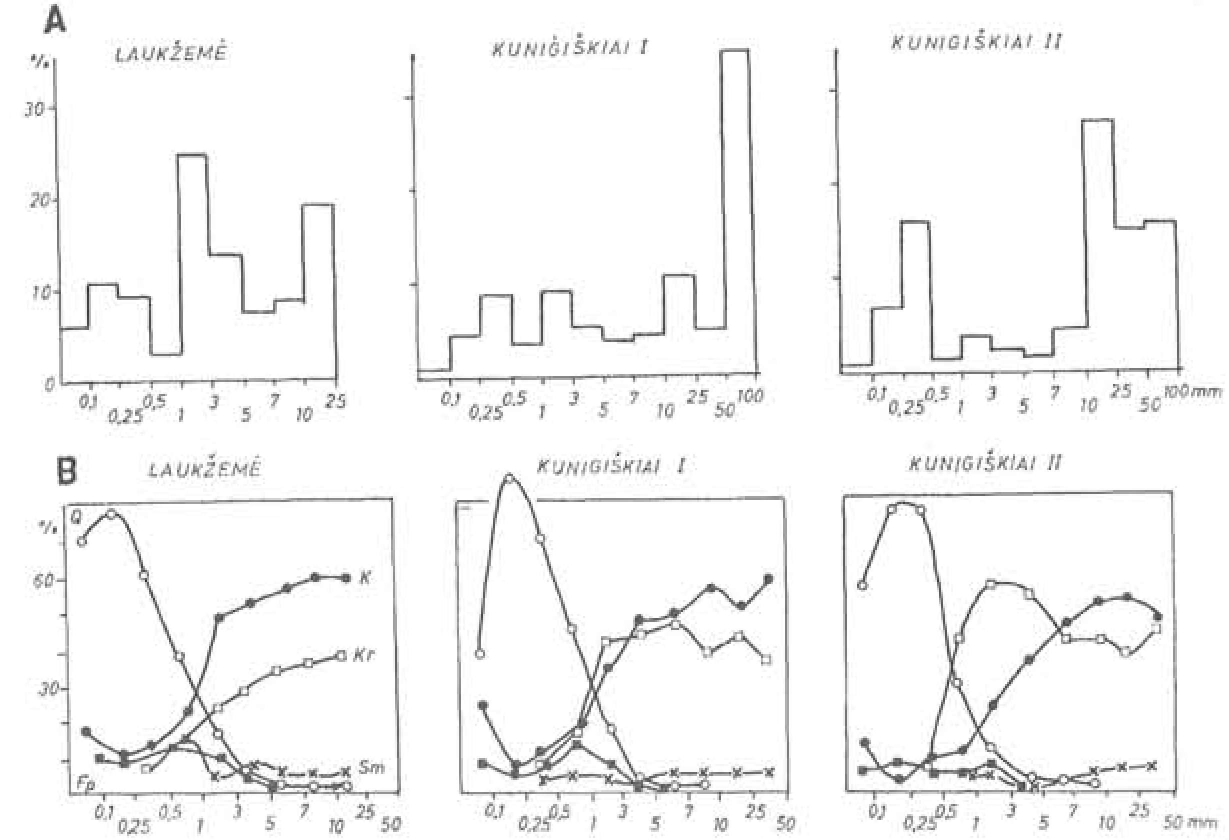

Fig. 5. A – granulometric composition (A) of shore ridge sediments (BG I–II) near Laukžemė and Kunigiškiai. B – distribution of quartz (Q), feldspar (Fp), sandstone (Sm), carbonate (K), crystalline rock (Kr) fragments in different size grades.

part of the Baltic Sea (gravity, waves and currents).

COMPOSITION OF CLASTS

Petrographic analyses of clasts were made along two sections (at Laukžemė and Kunigiškiai) of the beach ridge of the Baltic Ice Lake. The results of clast counts in granulometric spectrum for the upper (II) layer of Laukžemė and two (lower - I and upper – II) layers of Kunigiškiai sections are summarized in Fig. 5 (lower row of diagrams).

Five groups of rock and mineral fragments were counted: crystalline rock, quartz, feldspar, carbonate and sandstone. In the coarse fractions (1 – 50 mm) carbonate rocks (limestone – 19.5–56.1% and dolostone – 1.9–9.8%) and crystalline rocks (22.7–56.2%) prevail, quartz comprising 0.0–16.4%, feldspar 0.0–7.9% and sandstone 0.0–8.2%. Coastal deposits of the Baltic Ice Lake represent the redeposition of clastic material from a melting glacier of the last glaciation. Two categories of provenance-dependent compositional elements are considered: 1) erratic rock fragments including crystalline rocks derived from the Fennoscandian basement, 2) transitional and local rock sources such as carbonates, sandstones and others derived from (pre-Quaternary) Palaeozoic and Mesozoic sedimentary rocks.

The sand fraction (less than 1 mm) of the beach ridge sediments consists mainly dominantly of quartz (30.7–86.5%), feldspar (3.3–13.7%), carbonate – calcite and dolomite (2.5–22.8%), mica (0.5–25.8%) and heavy minerals (0.6–8.0%). The composition of sand fractions mostly depends on hydrodynamic processes of coastal sedimentation.

The clasts of the lower (I) layer in the Kunigiškiai section are less weathered. The most weathered upper (II) layer of the beach ridge of this section in the sand fraction is enriched with micas and heavy minerals. Coarse rock fragments are more weathered in the upper (II) layer than in the lower (I) one. The sediments of the upper layer were formed by the reworking of lower layer clasts. The weathering occurred simultaneously with depositional processes. The clasts were decomposed by the frosts.

As a result of mechanical comminution, at least two groups of clasts in diagrams are observed (Fig. 5). One mode is in the coarse clast size grade, consisting predominantly of rock fragments, and the other is in the sediment matrix, consisting mainly of mineral fragments (Dreimanis & Vagners 1971). Bimodal distribution of rock and mineral fragments is typical for sediments from glacigenic sources (Gaigalas 1974), an example of them being the deposits of the beach ridge of the Baltic Ice Lake. However, a bimodal distribution of grain size is not only typical for sediments from glacigenic sources, but also for beach sediments (Sonu 1972).

ROUNDNESS AND SHAPE OF CLASTS

The roundness and shape of pebbles and sand grains has long been studied to decipher the history of deposits of the Baltic Ice Lake. The shape of clasts may be classified in a number of ways, but the method described by Vassoyevich (1968) was used.

The roundness grades used (as shown in Fig. 6) are as follows: 0 – angular, 1 – subangular, 2 – subrounded, 3 – rounded and 4 – well-rounded. The clasts in the 1–2.5 and 2.5–5 cm size ranges are mostly subrounded and subangular. More subrounded pebbles are found in the 1–2.5 cm size range in the lower (I) layer (40%) and 2.5–5 cm in the upper (II) layer (45%) of the Kunigiškiai section (Mikalauskas & Gaigalas 1973).

The coefficient of oblongness was estimated as

$$K_0 = \frac{A+B}{2C} - 1,$$

and the coefficient of flatness of pebbles as

$$K_0 = \frac{2A}{B+C} - 1,$$

where A, B and C mark long, middle and short axes of pebbles.

The results of these observations are summarized in Fig. 6.

The oblongness and flatness grades used as follows: 0 – near isometric – with class limits from 0.0 to 0.1; 1 – subisometric – 0.1–0.25; 2 – subanisometric (subprolate or subflat) – 0.25–0.5; 3 – anisometric (prolate or flat) – 0.5–1; 4 – distinctly anisometric (distinctly prolate or flat) – 1–5.

The coefficient of anisometricity (Vassoyevich 1958) is $K = K_0 + K_f$ and characterizes the general shape of pebbles (Gaigalas 1965).

The conception of isometricity is not identical to the conception of sphericity of fragments. Emphasizing the difference between conceptions of roundness and of sphericity, H. Wadell (1935) expressed the sphericity coefficient by the following formula:

$$K_0 = \frac{S_0}{S_1},$$

where S_1 – surface area of fragment, S_0 – surface of sphere of equivalent volume.

An analysis of the general shape of pebbles has indicated that the low shpericity, high flatness, oblongness of the sand grains are typical for the coastal sediments of the Baltic Ice Lake in Lithuania.

Variations in petrology also affect the pebble shape (roundness, oblongness, flatness and anisometricity). In and on beach ridge deposit coarse-grained igneous and metamorphic clasts are more weathered than fine-grained clasts of the same composition. The forms of cobbles, pebbles and grains have had a long history before, but the time during which they were exposed to coastal processes was relatively short.

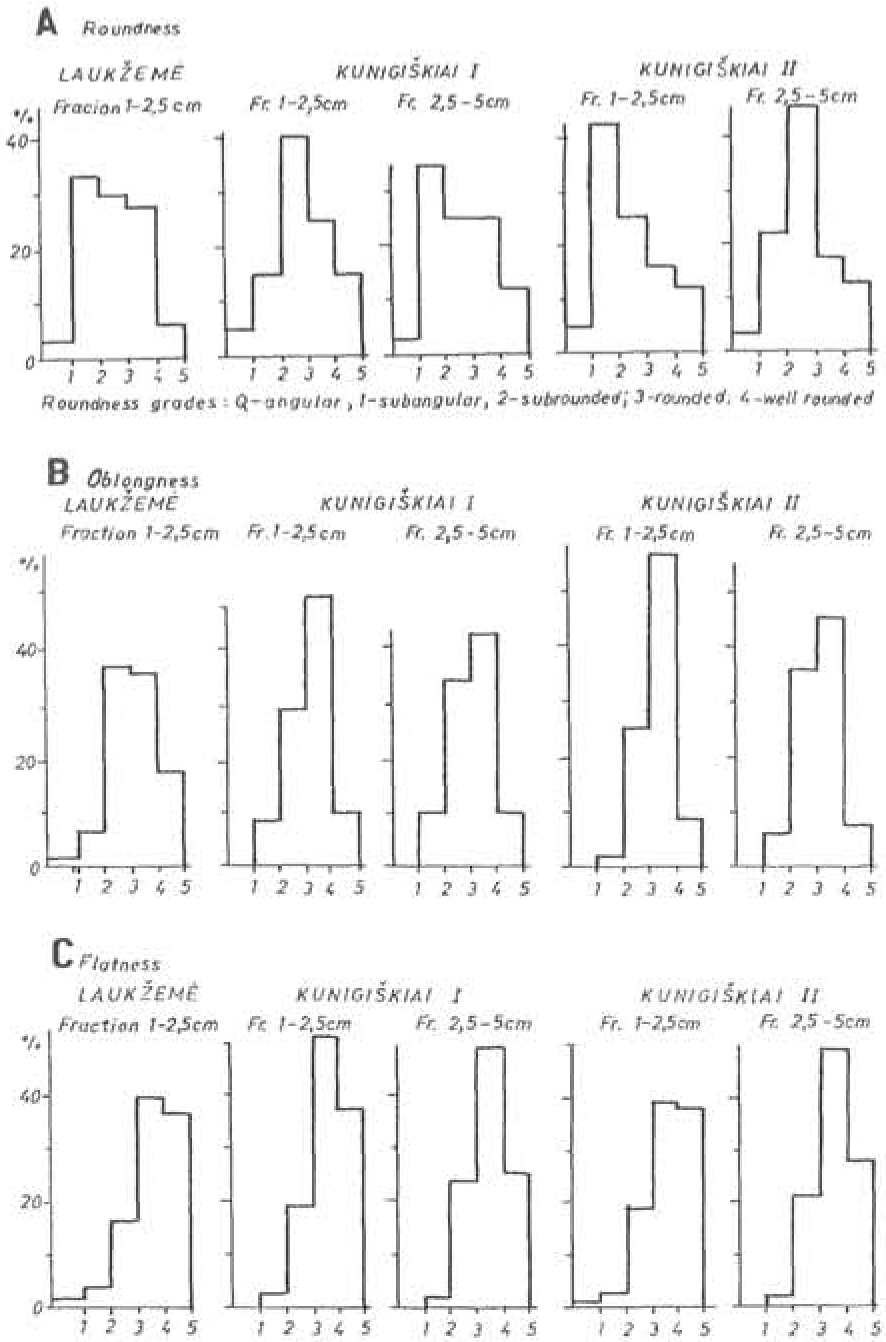

Fig. 6. The roundness (A), oblongness (B) and flatness (C) of clasts (fractions 1–2.5 and 2.5–5 cm) in sediments of shore ridge (BG I–II) near Laukžemė and Kunigiškiai. Oblongness and flatness (anisometricity) grades: 0 – near isometric (0.0–0.1); 1 – subanisometric (0.1–0.25); 2 – subprolate and subflat (0.25–0.5); 3 – prolate and flat (0.5–1); 4 – distinctly prolate and distinctly flat (1–5).

CONCLUSIONS

A detailed field study of the beach ridge deposits (BG II) and laboratory analyses are necessary to differentiate between different hydrodynamic environments. Composition of deposits of the Baltic Ice Lake, low degree of sorting, appearance of boulders and morphometry of fragment material show that the sources of the deposits are local and that their differentiation is low. The Baltic Ice Lake material macroscopically is similar to the glaciofluvial sediments. A detailed study of the lamination and other structural peculiarities of material provides valuable information about the sedimentation conditions. The low sphericity, high flatness, oblongness and anisometricity of the pebbles as well as the better roundness of the sand grains are typical for the coastal sediments of the Baltic Ice Lake in Lithuania. The better sorting of Baltic Ice Lake sediments is caused by the more homogeneous hydrodynamic environment. The processes determining the orientation of pebbles should be taken into account in the reconstruction of the conditions that prevailed during deposition. Detailed measurements on the orientation and dips of the long axes of pebbles allow to reconstruct the direction of waves and of the transportation of material.

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Bottom Relief and Genesis of the Gotland Depression

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Gelumbauskaitė, Ž. 1996: Bottom relief and genesis of the Gotland Depression. *Baltica, Vol. 9,* pp. 65–75. Vilnius. ISBN 9986-615-05-4 / ISSN 0067-3064

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The Gotland Depression, as the entire basin of the Baltic Sea has been formed under the conditions of interaction among tectonic, eustatic, denudational and glacioisostatic processes. Its development is thought to be related with that of the morphostructure/morphosculpture of the Baltic Shield and the Baltic Syneclise, starting from Mesozoic and ending in Holocene. The data for this article have been collected during joint Lithuanian-Swedish geological-geophysical expeditions in 1993–1994. The Pre-Quaternary morphostructure of the sea floor has been analyzed, the thickness and composition of the Quaternary described, and bottom topography analyzed in detail. Four stages have been distinguished in the development of the Gotland Depression: (1) the Mesozoic-Paleogene 1st degree morphostructures, (2) the Neogene 2nd and 3d degree morphostructures, (3) the Pleistocene glacial morphosculptures and (4) the Holocene morphostructures of formation of the present Baltic Sea.

Keywords: Baltic Sea, Gotland Depression, morphostructure, morphosculpture, bottom topography, Baltic Sea development.

INTRODUCTION

Evaluation and prediction of each geosystem, as well as of a present state of an ecosystem, can be done only by means of a comprehensive study of all the parts of the system. In this article we would like to look into the influence of the plasticity and origin of the recent relief in the Gotland Depression on present litho-hydrodynamical processes.

The data for this article have been collected during joint Lithuanian-Swedish expeditions in 1993–1994 (Grigelis & Flodén 1994).

The Baltic Sea is a basin of a mediterranean platformic type. The Gotland Depression, like the entire area of the Baltic Proper, has been formed under conditions of interaction among tectonic, eustatic, denudational and glacioisostatic processes. Its development is thought to be related to that of the morphostructure/morphosculpture of the Baltic Shield and the Baltic Syneclise, starting from the Mesozoic and ending in the Holocene (Flodén 1980, Gudelis 1976, Kumpas 1977, Puura et al. 1991, Sviridov 1981).

RELIEF OF THE PRE-QUATERNARY MORPHOSTRUCTURE

At present, the Gotland Depression lies in the field of the Silurian-Devonian peneplain dissected during the Pre-Quaternary. In the southwestern part the Basin borders upon the zone of tectonic fractures, in the east it is limited by the Klaipėda-Liepaja structural uplift composed of the Upper Devonian Plaviņas-Pamūšis Regional Stage and by the Ventspils-Saaremaa Step of the Middle Devonian Burtnieki-Narva Regional Stage. In the north it is isolated from the Fårö Basin by the Jaani-Jaagarahu glint of the Lower Silurian. Its western slope borders upon the Łeba-Gotland Plateau of the Lower-Upper Silurian (Fig.1A).

The Basin itself is clearly divided into the southern and the northern parts and separated by a double swell of banks. The northern one is known as the Klints Bank, whereas the southern one is named the South Klints Bank and described here for the first time (Fig. 1A). The plasticity of palaeorelief of these morphostructures in the Gotland Depression is directly reflected in the recent relief. This confirmed the fact that glacial accumulation occurred differently and it was controlled by the the Pre-Quaternary surface had been formed already.

Southern part of the Basin separated by the swell makes up Western Gotland Trough and Eastern Gotland Trough. Their floors lie at the depths of 140–180 m with an inclination northwards. Their soutwestern end-parts are dissected by erosional incisions. The floor of the Western Trough consists of the Middle Devonian Narva rocks, whereas the Eastern Trough has the surface formed of Burtnieki rocks with signs of tectonic fractures. Their slopes are step-wise with glints. The morphology and origin of the palaeorelief of this Pre-Quaternary

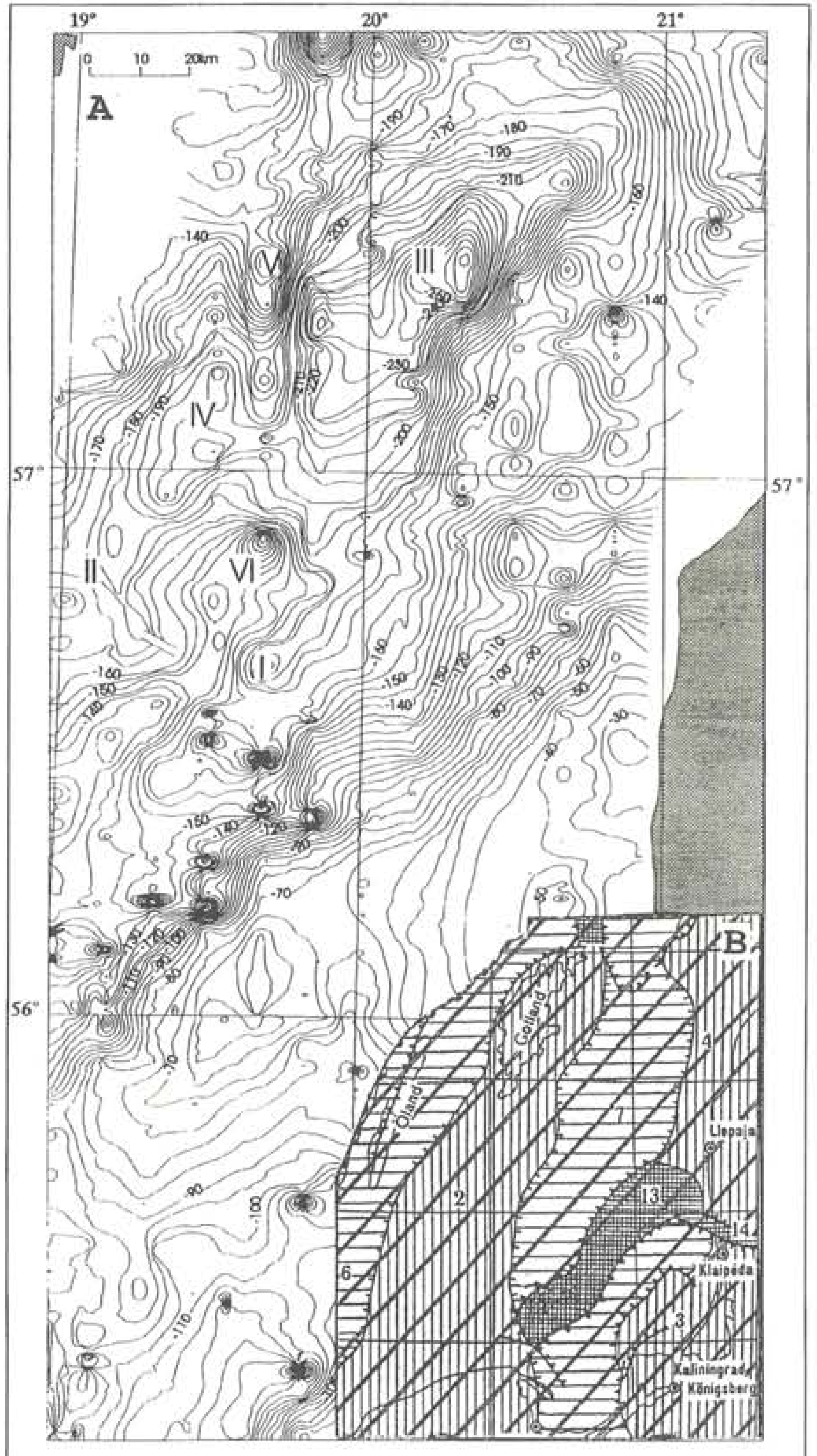

Fig. 1 A. Pre-Quaternary surface of the eastern Central and South Baltic. Structure contours are drawn at 5 m intervals. I – Eastern Gotland Trough; II – Western Gotland Trough; III – Eastern Gotland Basin; IV – Western Gotland Basin; V – Klints Bank Swell; VI – South Klints Bank Swell.

B. Sketch map of morphostructural zones (after Ž. Gelumbauskaitė, 1991). 2 – Łeba-Gotland Plateau; 3 – Sambia-Kurish Step; 4 – Ventspils-Saaremaa Step; 6 – Bornholm-Öland Depression; 7 – Central Baltic Proper; 13 – Klaipėda–Liepaja Uplift; 14 – Telšiai Swell.

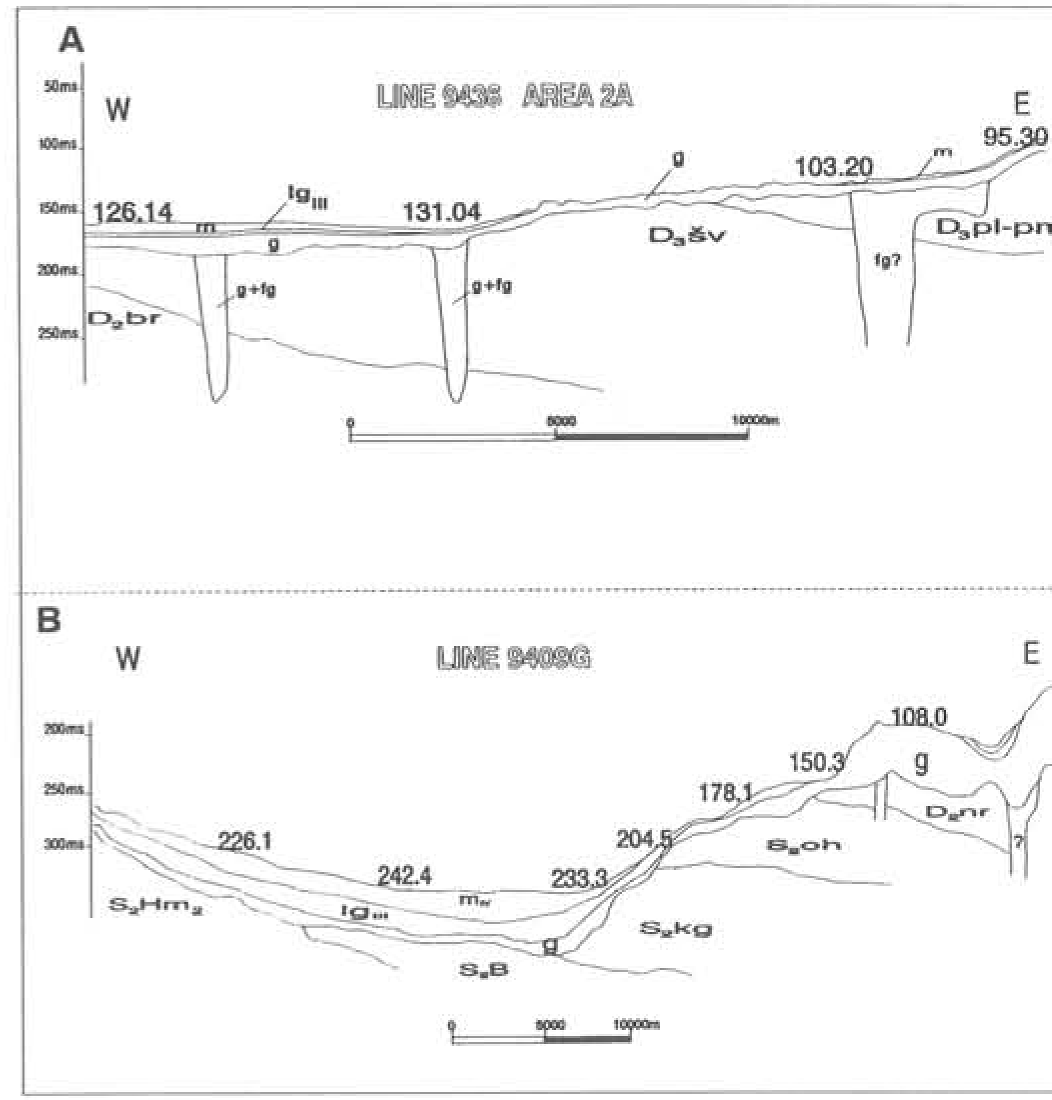

Fig. 2 A. Middle and Upper Devonian seismic recording profile 9436, Area 2A, Eastern Gotland Trough. D_2br – Burtnieki Horizon, $D_3\check{s}v$ – Šventoji Horizon, $D_3pl–pm$ – Pliaviņas-Pamūšis Complex; Pleistocene deposits: g+fg – moraine + fluvioglacial; g – glacial; lgIII – limnoglacial; mIV – Holocene marine deposits. Location see on Fig. 6.

B. Uppermost Silurian and Middle Devonian seismic recording profile 9409G (GOBEX Line), Gotland Deep. S_2Hm_2 – Hemse Group, S_2B – Burgsvik Group; other symbols see on Fig. 3A.

morphostructure in these troughs are reflected by fragments in the profiles 9420 and 9436.

The seismic profile 9436 (Fig. 2A) crossing the Eastern Gotland Trough from the east to the west shows well-expressed stepwise eastern slope hardly covered by the Quaternary deposits. At the top of the slope, the Quaternary layer is about 6 m thick, whereas at the depths of 111–126 m, the Upper Devonian Šventoji Regional Stage crops to the surface. Three V-shaped incisions with depths exceeding 100 m are found in the trough floor. Preliminary data show that they are filled by till, sand and gravel.

The seismic profile 9420 (Fig. 3B), passing from the north to the south, shows in the morphology of the floor in the Eastern Gotland Trough a graben-type depression of about 30 m deep with slopes having flexures of the Middle Devonian Burtnieki Regional Stage and floor having several small incisions. These incisions are filled up with stratified fluvioglacial material.

The Swell consisting of the Klints and the Southern Klints Banks is a structural/denudational remnant of the Middle Devonian Narva and Aruküla Regional Stages lying at the depths of 140–210 m with its northeastern part ending by an impressive cliff at the depth of 105 m (Fig. 3A). It should be noted that such cliffs are detected in other profiles as well. They are found in the northern zone of the faults stretching from NW to SE in the area where Middle Devonian Narva Regional Stage join the Upper Silurian Ohesaare Regional Stage

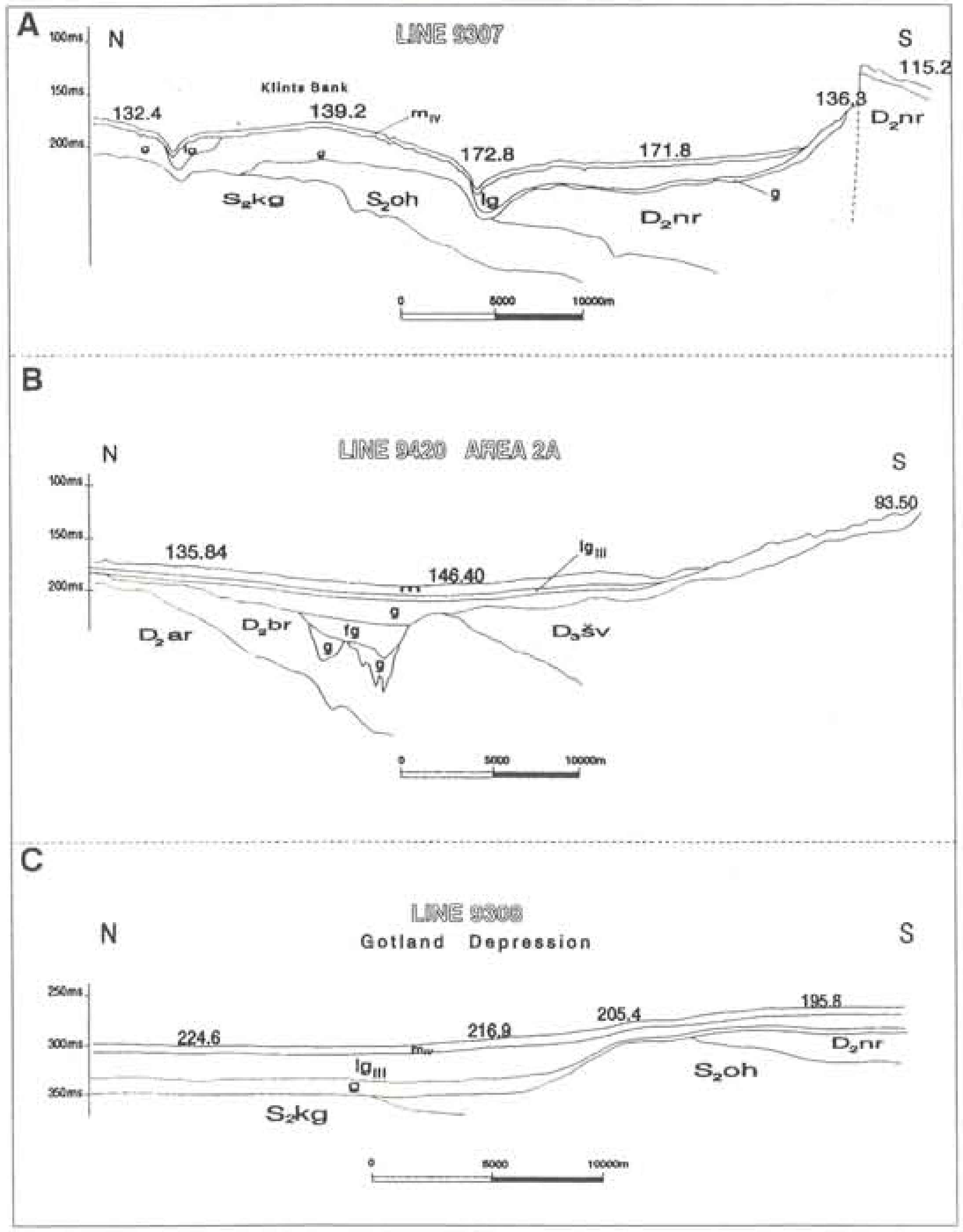

Fig. 3 A. Uppermost Silurian and Middle Devonian seismic recording profile 9307, Klints Bank Area. S_2kg – Kaugatuma Horizon, S_2oh – Ohesaare Horizon, D_2nr – Narva Group; Pleistocene deposits: g – glacial, lg – limnoglacial; fg – fluvioglacial; m – Holocene marine deposits. Location see on Fig. 6.

B. Middle and Upper Devonian seismic recording profile 9420, Area 2A, Eastern Gotland Trough. D_2ar – Arukiila Horizon, D_2br – Burtnieki Horizon, $D_3šv$ – Šventoji Horizon.

C. Uppermost Silurian and Middle Devonian seismic recording profile 9308, Gotland Depression.

(Grigelis et al. 1991, Puura et al. 1991) and separate southern part of the depression from the northern one.

For the first time the Devonian cliffs have been described by M. Kumpas (1977) and thought to be erosional formations. The author of the present paper express their opinion that they are of structural/denudational origin and are related with the NW-SE zone of faults dividing the Gotland Depression into the northern and southern parts.

The data obtained during seismic profiling done in 1993–1994 show that the deepest NE part of the basin is lying lower than the southern one. A fragment of the profile 9308 (Fig. 3C) shows that

the Upper Silurian layers in the northern depression are by 20 m lower than the southern one.

On its turn, the northern part of the Basin is also divided into the Western and the Eastern Depressions. The Eastern one is made of a kettle-like negative morphostructure at the depths of 205–285 m. It overlies the Upper Silurian rocks.

The morphology and morphostructure of the Pre-Quaternary surface are well-reflected in the seismic profile 9409 (GOBEX Project) going across the centre of the depression. A stepwise relief of the Upper Silurian Hemse Group on the western slope is replaced by a reefogenous relief of the Upper Silurian Burgsvik Group and Kaugatuma Regional Stage on the floor, whereas the eastern slope has stepwise relief of the Upper Silurian Ohesaare Regional Stage. The Silurian glints in some areas of the slopes at the depths of 150–180 m are hardly covered with the Quaternary rocks (Fig. 2B).

A surface of the Western Depression lying higher (at the depths of 175–200 m) is divided by flexures of the Upper Silurian Kaugatuma and Ohesaare rocks. There are no incisions in both depressions: the Western and the Eastern.

The Klints Bank Swell separating the depressions at the depths of 170–120 m is a remnant of the Upper Silurian Hemse and Kaugatuma units.

THICKNESS AND COMPOSITION OF THE QUATERNARY

Distribution of Quaternary thicknesses and composition is uneven in the area studied. The Quaternary is made of complicated combination of layers settled during Pleistocene-Holocene time. Its major part consists of glacial, fluvioglacial and limnoglacial matter accumulated. In some parts of the Gotland Depression, the floor is also unevenly filled up with deposits of the Holocene palaeobasins.

The material collected during echosoundings, seismic profiling and sampling of bottom sediments shows the influence of the Pre-Quaternary surface roughness on movement of glacier bodies and on formation of palaeobasins. Analyzing geological section in seismic profiles, signs of activity typical of the Caledonian and the Hercynian disjunctive and plicative structures are detected in the neotectonic period, i.e. vertical tectonic movements have been found to take part in glacial exaration/denudation and accumulation processes.

Interpreting the materials of the 1993–1994 expeditions, the data for evaluation of the thicknesses of the Quaternary deposits were obtained after bathymetric and Pre-Quaternary maps were produced by digitisation seismic profiles, and then the Quaternary map was compiled.

The Quaternary thicknesses in the southern part of the depression range from 5 to 20–25 m. The SW end-parts of the Eastern and Western Troughs at the depths of 140–165 m are covered by a 10–15 m thick layer of the Quaternary deposits; going deeper, this layer becomes thicker (to 20–25 m) at the depths of 165–180 m. The infilling of the incisions is quite complicated. Mainly they are filled up with morainic and fluvioglacial Pleistocene matter (Bjerkéus et al. 1994).

A fragment of seismic profile 9420 (see Fig. 3B) shows that the Quaternary at the floor of the Eastern Gotland Trough is about 20 m thick, with 3–5 m settled during the Holocene. A fragment of seismic profile 9436 (see Fig. 2A) indicates that Quaternary and Holocene thicknesses on the trough floor reach 14 and 5.5 m, correspondingly. The swell separating the troughs is covered with Quaternary deposits 5–10 m thick, or even without them in some places, as for example, on the above-mentioned cliff found at the depth of 105 m.

The northern part of the depression is more complicated. The map of Quaternary thicknesses (see Fig. 4) shows distinctly the Klints Bank and eastern slope of the Gotland Deep. The Upper Silurian structural/denudational remnant of the Klints Bank at the depth of 120 m is covered with the Quaternary deposits, 0–60 m thick, whereas the eastern slope of the Gotland Deep at the depth of 130 m has Quaternary thicknesses exceeding 90 m. These layers are considered to consist of glacial/fluvioglacial deposits (Mörner et al. 1977), and for the first time they are attributed to the Pandivere recession phase marked on the Geomorphological Map of the Baltic Sea and its coasts in 1990 (Gelumbauskaitė et al. 1993).

Total Quaternary thickness in the profile 9409 reaches 40 m. The Holocene deposits at the floor of the Gotland Deep make up 10 m, whereas Late-Glacial clays are 21 m thick, and Pleistocene till occurs in a 10 m thick layer. At the geological site 118 (depth –232.8 m), the pollen analysis (data of O.Kondratienė, Lithuanian Institute of Geology) has shown the boundary between Upper and Middle Subatlantic [SA3 & SA2, correspondingly] to be at the depth of 0.54 cm (Fig. 5). Quaternary thicknesses in the western depression also reach 35–40 m. But their distribution is different: the Holocene and Pleistocene layers are equally thick, 20 m each.

A. Blazhchishin et al. (1985) carried out a revision of the litho- and biostratigraphic data for the Gdańsk and Gotland depressions and divided the Quaternary of the Gotland Depression into 7 lithostratigraphic complexes. According to core

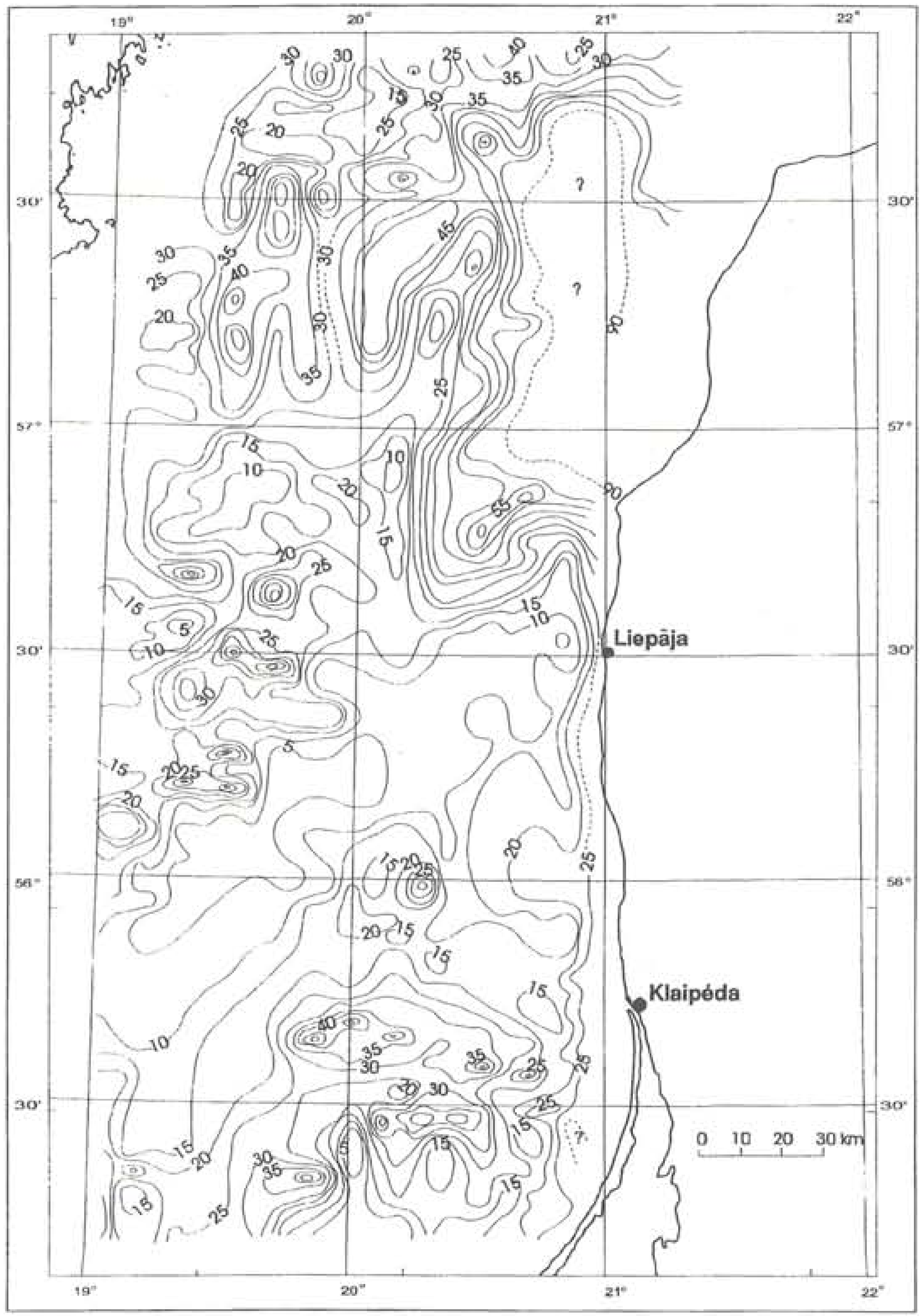

Fig. 4. Quaternary thickness (m) of the Gotland Depression and adjacent SE Baltic area.

analyses and lithostratigraphic studies of the Klints Bank and Southern Depression, the Holocene (i.e. Limnaea+Litorina+Ancylus+Yoldia) makes up 5–7 m, and the Baltic Ice Lake deposits are 3–5 m thick. The deposits of ice lakes, i.e. varved clays and underlying till layer related to the Pandivere recession phase are 6–8 m thick. The sixth lithostratigraphic complex, 20 m thick, seems to be formed of laminated clays during melting of retained glacier lobe. The last seventh

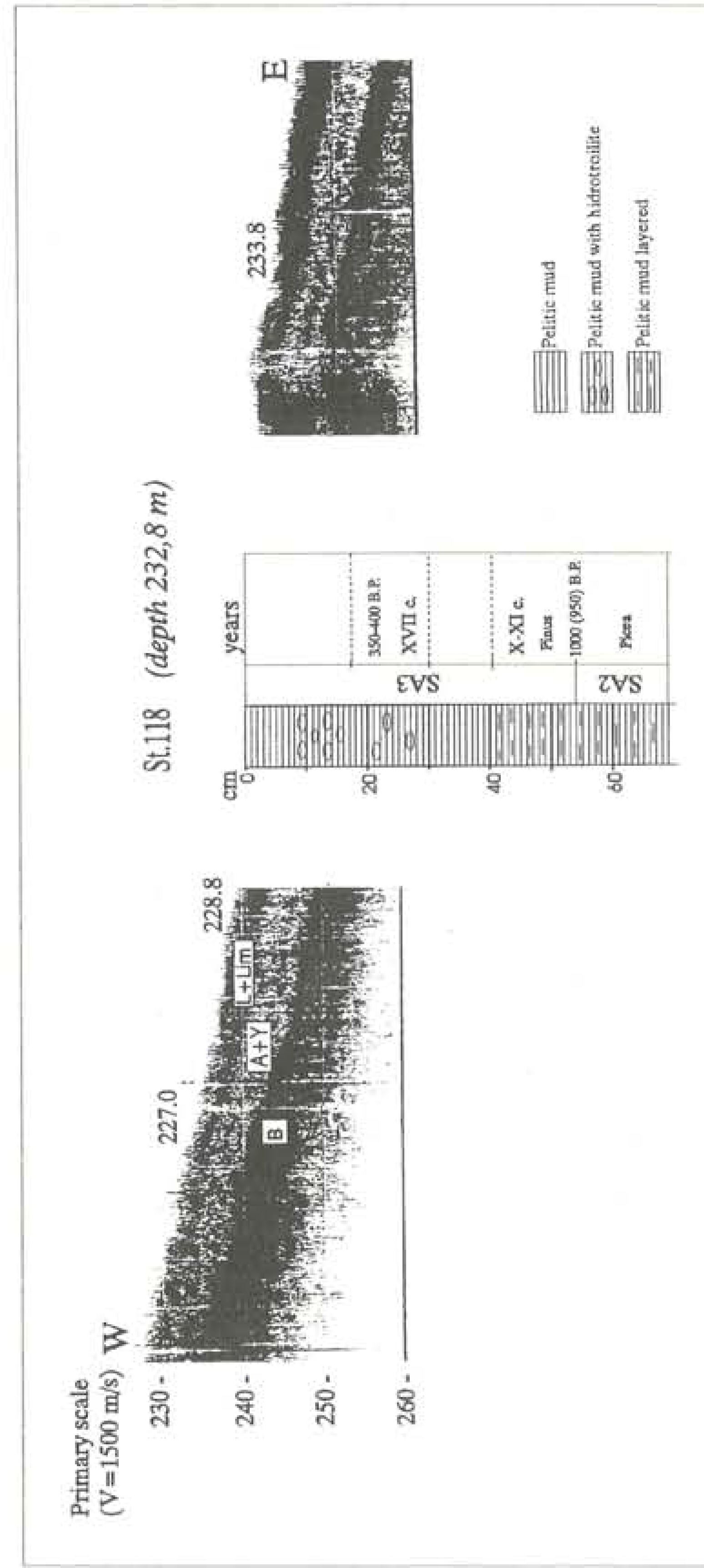

Fig. 5. Holocene sediments echosounding record along profile 9409, and site 118 (1993) (19°59,9' E; 57°18,8' N) from the Gotland Deep. SA3 – Late Subatlantic, SA2 – Middle Subatlantic; L – Litorina, Lim – Limnea, A – Ancylus, Y – Yoldia, B – Baltic Ice Lake.

complex, 3–10 m thick, consists of a double-membered till layer.

BOTTOM RELIEF

The measurements of depth and records of bottom relief plasticity done by echosoundings with SKIPPER 607 and FURUNO 881MK II were digitalized and, applying special software, the bathymetric map of the Baltic Proper was compiled on the scale of 1:200 000 in the UTM coordinate system with a 4 m step of isobathes. This map supplements the other maps compiled in 1988 by Ž. Gelumbauskaitė and V. Litvin on the scale of 1:500 000 in the coordinate system by Gauss-Kryger, and in 1993 by T. Flodén on the scale of 1:200 000 in the UTM coordinate system.

The data obtained are used for compiling new morphological scheme of the Gotland Depression (Fig. 6) with all the macroform details, although known previously but morphologically unstudied and undescribed. The scheme shows that the Gotland Depression is distinctly divided into the southern and the northern parts which both are subdivided by the Klints Bank and the South

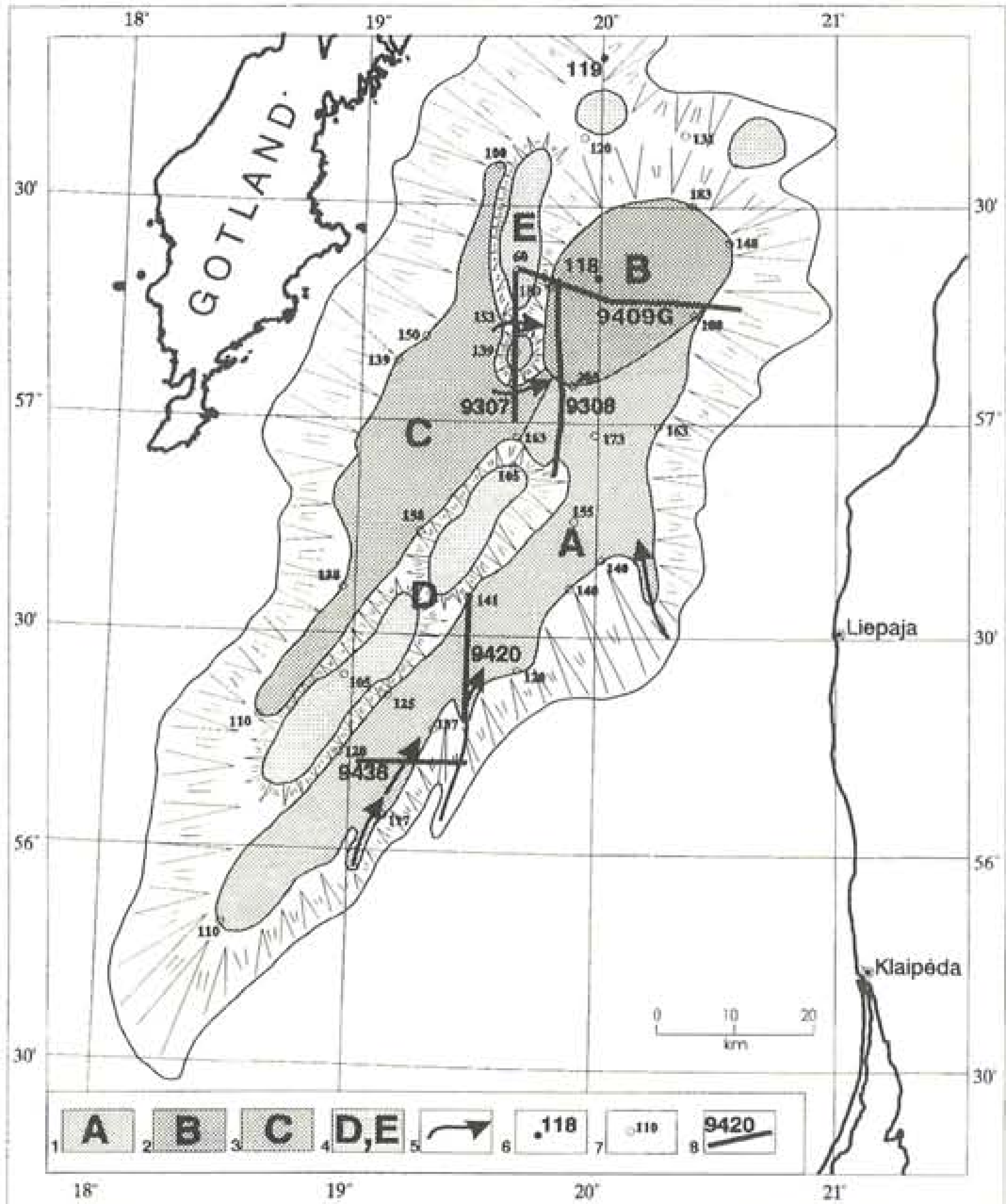

Fig. 6. Morphology of the Gotland Depression. Ž. Gelumbauskaitė, 1995. 1A – Eastern Gotland Depression; 2B – Gotland Deep; 3C – Western Gotland Depression; 4D – South Klints Bank, 4E – Klints Bank; 5 – general current patterns; 6 – sampling site, 7 – depth of the sea floor, m; 8 – seismic recording and echosounding profiles.

Klints Bank into the eastern and the western parts.

The southern part of the Gotland Depression is prolongated, its floor is inclined and stretching from SW to NE from 110 m to 170 m. The Eastern Depression is separated from the Western one by the South Klints Bank. Eastern and western slopes of the southern depression are at the depths from 70 to 120–140 m. The eastern slope is rather steep and stepwise with Upper-Middle Devonian rock glints having no Quaternary cover in some places of the Pre-Quaternary surface; the glints are easily detected in the recent relief. The western slope is less inclined, the Upper-Lower Silurian glints are distinctly seen at its northern part in the recent relief.

Echosounding and seismic profiles reflect close relationship between the Pre-Quaternary palaeorelief of the southern depression and the recent relief. The signs of palaeoincision network fixed in seismic profiles 9420 and 9436 are revealed in echosoundings as the arteries of recent currents directed along the eastern slope of the depression (see Figs. 2A, 3B). Such signs of the arteries fixed in some profile fragments are detected in the Western Depression as well. A large prolongated graben-shaped incision found in the Pre-Quaternary surface is expressed in the recent relief by the zones of active gas seepages.

Plateau of the South Klints Bank is at the depths of 110–105 m with a lowering to 120 m in its central part. The slopes are rather plane and occur at the depths from 110–120 to 140 m.

The northern part of the Gotland Depression differs greatly from the southern one. The northern Gotland Deep with a maximum depth fixed at 249 m is a kettle-like depression inclined from SW to NE at the depths of 205–249–180 m. The fault zone reflected in the Pre-Quaternary surface separates the northern part of the depression from the southern one by distinct cliffs and make up a rise in the relief of about 10 m.

There are four steps on the slopes of the Gotland Deep (profile 9409). They are especially distinct on its eastern slope: Step 1 at 233.3–182.4 m, Step 2 at 182.4–150.2 m, Step 3 at 150.2–108.0 m and Step 4 at 108.0–72.5 m. The western slope is of a similar structure. Steps 1 and 2 correspond to the Silurian glints which are not covered by the Quaternary deposits in some places. Steps 3 and 4 correspond to recessive marginal formations attributable to the litostratigraphical complex No. 5 related to the Pandivere recession phase (Fig. 7).

The core data from the site 118 (see Fig. 5) show that during last millenium 540 mm of pelitic deposits settled down in the Gotland Deep. At the same time, only 25 mm have been settled on the

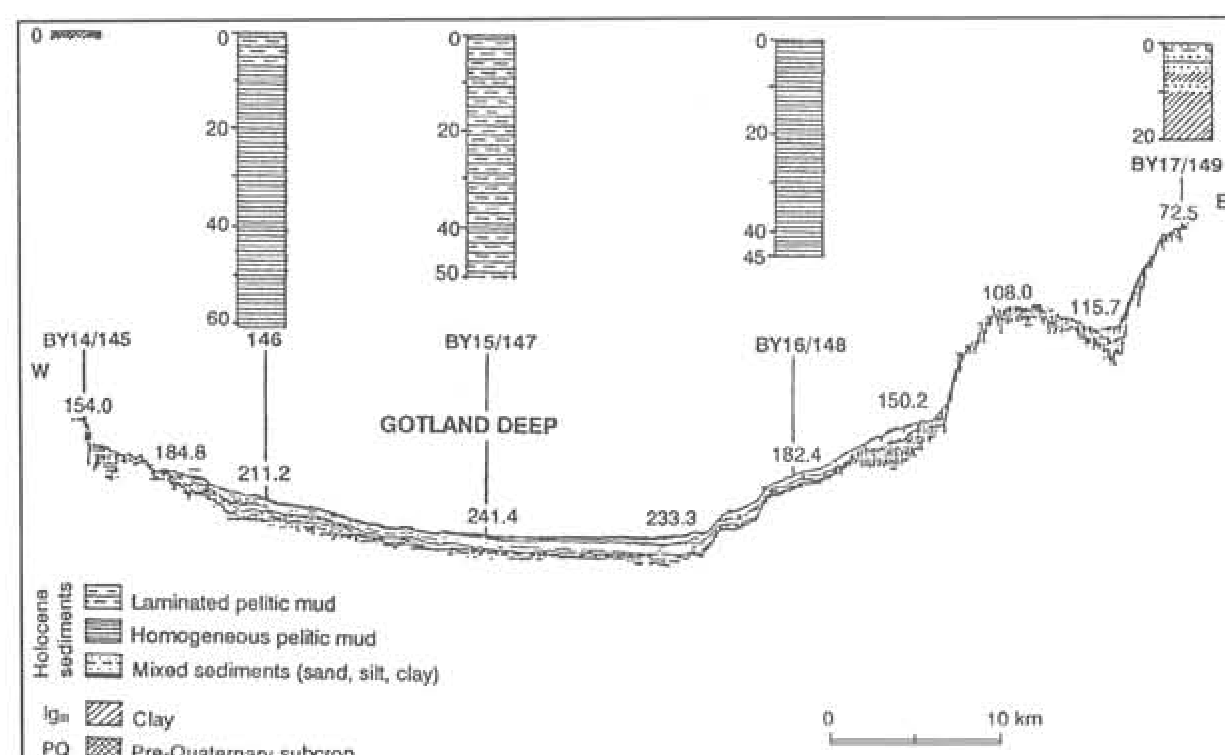

Fig. 7. Echosounding line 9409G (GOBEX Line) and bottom sediment columns (thickness in centimetres), an illustration of lithodynamic processes in the Gotland Deep.

saddle separating the Gotland and the Fårö Depressions in the northern part of the Basin, according to pollen data from the core of geological site 119 (the data presented by L.Kalniņa, Latvian Geological Survey).

The core data from the floor of the Gotland Deep (Fig. 7) show that lithodynamical processes are highly complicated here. Recent sedimentation occurs not everywhere, since it seems to be related to active hydrodynamics of near-bottom water masses.

The morphology of the northern end of the West Gotland Depression separated by the Klints and the South Klints Banks is the same as that in the southern part; only its central part with depths varying in 140–179–171 m is inclined from the west to the east and resembles to its northern neighbour. Its connection with the latter one is obviously shown by a fragment of the seismic profile 9307. Signs of the recent near-bottom currents are detected at the southern end of the Klints Bank at the depths of 172 and 153 m (see Fig. 3A).

DEVELOPMENT OF THE GOTLAND DEPRESSION

Development of the Gotland Depression can be divided into several time stages: (1) Mesozoic-Paleogene 1st degree morphostructures, (2) Neogene 2nd and 3d degree morphostructures, (3) Pleistocene glacial morphostructures, and (4) Holocene Baltic Sea formation (Gelumbauskaitė & Litvin 1990).

During the 1st rather long stage of development, the peneplains of the Baltic Shield base and the monocline relief of the Russian Plate margins expressed by the Palaeozoic, Mesozoic and Palaeogene steps have been formed. During the second stage – Neogene – the peneplained surfaces were being dissected further. During this stage, in the Central Baltic, the peneplains of Łeba-Gotland, Sambija-Curonian and Ventspils-Saaremaa have been formed together with the Central Baltic Depression that, during Neogene-Quaternary, was gradually being divided by the Liepaja–Klaipėda uplift into the Gotland and Gdańsk Depressions. These main morphostructures make up at present the bases of the Baltic Proper macroforms (Gelumbauskaitė 1991).

The next stage of relief development is related to glaciations. The Pleistocene period in the Gotland Depression has inherited heterogenous Pre-Quaternary surface formed under subaerial conditions of Pliocene and dissected by river network in the southern part of the Depression (Grigelis & Flodén 1994).

There were rather scanty studies showing how the mechanism of deglaciation functioned in the Baltic Proper, especially in the Gotland Basin, i.e. how the glacial accumulation interacted with exaration – impact of glacier meltwater, how glaciotectonics was related to vertical tectonic movements.

A. Gaigalas (1974) studied the leading boulders and determined that glacial exaration was most intense during Early and Late Pleistocene, when glacier tongues moved from the north to the south. J. S. Aber (1993) affirmed that glacial accumulation in the Central and South Baltics made 50%, whereas exaration and deformation reached only 25%.

Recently the opinion became clear that glacio-isostatic surging movements were in a close relationship with vertical tectonic ones (Gudelis 1973, Mörner 1988). Interglacials were also very important for the formation of the Baltic Sea. The last Weichselian retreat of glaciers was very important for the Gotland Depression; at that time glacial exaration concluded the formation of the depression of Gotland finally and accumulated the recessive Palivere and Pandivere marginal zones (Gudelis 1976; Mörner et al. 1977).

The structure of the Gotland Depression also plays a role of a key to the history of its development during the Late Glacial and the Holocene. After glacier margin retreated from the marginal Pandivere formations area and paused during the Palivere recessive phase, and after the periglacial lakes joined together, the recent history of the Baltic Sea started (Gelumbauskaitė et al. 1993).

All the phases of the Baltic Sea development – starting from Yoldia and finishing with Limnaea – can be singled out in the Holocene (4th stage) palaeogeography.

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Hydrographic and Climatic Changes Recorded in Holocene Sediments of the Central Baltic Sea

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Huckriede, H., Clasen, S., & Meischner, D. 1996: Hydrographic and climatic changes recorded in Holocene sediments of the Central Baltic Sea. *Baltica, Vol. 9*, pp. 76–91. Vilnius. ISBN 9986-615-05-4 / ISSN 0067-3064

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New insights into the history of the Baltic Sea derive from detailed investigation of piston and gravity cores from the deep basins of the central Baltic Sea. After the drainage of the Baltic Ice Lake 10,400 YBP, water exchange with the North Sea was blocked for approximately 500 years at the beginning of the *Ancylus* stage. With the exception of this short period, brackish conditions prevailed within the bottom waters of the Baltic Sea. Brackish surface waters, as interpreted from diatom assemblages, are restricted to parts of the *Yoldia* and *Mastogloia* stages and to the *Litorina* and *Limnaea* stages. The present hydrographic conditions became established during the *Mastogloia* stage roughly 8,000 YBP. They are characterised by salinity stratification of the water column and drastic changes between anoxic sulphidic and oxic conditions in bottom waters. Oxygenation of bottom waters is controlled by a long-term periodicity of 800 to 900 years which is related to climatic variations such as the Little Ice Age.

Keywords: Baltic Sea – Gotland Basin – Bornholm Basin – Holocene – laminated sediments – stratigraphy – geochemistry – authigenic carbonates – salt water intrusions – palaeo-salinity – climatic cycles.

INTRODUCTION

The modern Baltic Sea is the largest brackish sea in the world, and covers an area of 412.560 km². Water exchange with the North Sea via the Skagerrak is restricted to the shallow and narrow Danish Sounds, and is therefore sensitive to isostatic, eustatic, and climatic changes. Such changes and their impact on the palaeo-environment can be reconstructed from undisturbed and complete sediment profiles.

The main phases of the history of the Baltic Sea have been known since the end of the last century (Hyvärinen & Eronen 1979). Important descriptions and summaries were published by Sauramo (1958), Alhonen (1971), Eronen (1974, 1988), Winterhalter et al. (1981), and Hyvärinen (1988). Modern, comprehensive descriptions of the development of the Baltic Sea are compiled in a volume edited by Gudelis and Königsson (1979). The following assumptions are generally accepted:

After the retreat of the Fennoscandian ice-shield some 12,000 YBP, varved sediments accumulated within a large glacial freshwater lake, the Baltic Ice Lake. Approximately 10,400 YBP, the retreating ice margin opened a new connection to the North Sea via the Skagerrak at Mt. Billingen in Sweden. As a consequence, a sudden drop of the water level occurred and marine water invaded the Baltic Ice Lake. This event marks the beginning of the *Yoldia* stage, which is characterised by brackish conditions particularly in the western parts of the Baltic Sea. Due to the rapid isostatic rise of central Sweden, the connection with the Skagerrak was closed 9,500 YBP, and once again fresh water lake conditions developed, the *Ancylus* Lake. Roughly 8,000 YBP the rising sea level reached the Danish Sounds, and the former fresh water lake was again flooded by marine waters, forming the *Litorina* Sea. The highest salinities occurred during the middle part of the *Litorina* stage. Since then, the salinity gradually dropped, and therefore the *Limnaea* stage is separated from the *Litorina* stage at approximately 4,000 YBP. The sediments of the last few hundred years are sometimes referred to as a separate stage, the *Mya* Sea. However, this stage is no longer used in most recent publications.

A detailed history of the Holocene development is the base for any discussion of the present ecological state and the future of the Baltic Sea. Unfortunately, many uncertainties and unsolved questions remain. With the exception of the drainage of the Baltic Ice Lake, all boundaries between the stages of the Baltic Sea are poorly defined and dated (Hyvärinen 1988). Some transitional stages, especially the *Echeneis* stage (Sauramo 1958) between the *Yoldia* and the *Ancylus* stages, and the *Mastogloia* stage (Hyvärinen et al. 1988), separating the *Ancylus* and the *Litorina* stages, are known from shallow areas but have never been found in deep water (Winterhalter et al. 1981). Furthermore, the existence of a separate *Echeneis* stage is questioned (Alhonen 1971, Eronen 1974, Kessel et al. 1988), also the presence of brackish waters in the central and eastern Baltic Sea during the *Yoldia* stage is not generally accepted (Abelmann 1985, Eronen 1974, Raukas 1994). In the central Baltic Sea, the base of the *Litorina*-sediments is marked by a sharp lithostratigraphic boundary. According to Winterhalter et al. (1981) this points to a catastrophic event in the hydrography of the entire Baltic Sea, but the nature of this event is unexplained. During the *Litorina* and *Limnaea* stages, repeated changes in the oxygenation of bottom waters are represented by deposition of either bioturbated mud or laminated sapropels (Winterhalter 1992, Huckriede 1994). The factors controlling these long-term variations are unknown.

In this paper, we present a reconstruction of the hydrographic and stratigraphic development of the Baltic Sea based on sedimentological, geochemical, and palaeontological analyses of sediment-cores from the Gotland Basin. This basin is up to 250 m deep and is located in the central part of the Baltic Sea. It contains complete, undisturbed sediment profiles which are particularly well suited for a detailed reconstruction of the history of the Baltic Sea. Additionally, cores from the Bornholm Basin are described for comparison.

MATERIAL AND METHODS

Sediments from the Gotland Basin were sampled with piston and gravity corers during cruises No. 184 of R/V GAUSS in 1991 and No. 138/3 of R/V VALDIVIA in 1993. The cores from the Bornholm Basin were taken during cruises of R/V HUMBOLDT in 1993. Generally, the uppermost 5 to 15 cm of the profiles are disturbed or missing. With this exception, the sediment columns in the cores are complete and undisturbed. Figure 1 shows the core stations in the central Baltic Sea; for positions and other technical data see Table 1.

The sediments were studied in detail using thin sections, polished sections, X-ray radiographs, X-ray diffractometry, and AAS-analysis. Carbon and sulphur were analysed with an "ELTRA Metalyt CS 100 / 1000 S" C/S analyser. Organic carbon (TOC) was determined in samples which were pretreated with diluted HCl to dissolve carbonates. For AAS-analysis (determination of Mn, Fe, Ca, Sr, Mg, Cu, Zn), desalted bulk samples were treated with 3% HCl and 20% H_2O_2 at room temperature for 20 hours. Carbonates, sulphides and some reactive silicates were dissolved during this procedure. The sodium content of rhodochrosite was determined from the grain size fraction 10 to 63 μm which was leached with 10% acetic acid for 30 minutes at room temperature. The solutions were filtrated and then analysed with a Philips PU9200 atomic absorption spectrometer according to the details given in Heinrichs & Herrmann (1990).

Samples for grain size analysis and separation of microfossils were treated with 30% H_2O_2 for approximately 14 days. Acids which may form during this procedure were neutralised with NH_4OH. Carbonates and sulphides were dissolved with a mixture of 10% acetic acid and 30% H_2O_2 for grain size analysis of the non-carbonate fraction. The grain size distribution was then determined with a CIS-laser particle size recorder (LOT GmbH).

Diatoms were enriched in the grain size fraction >10 μm by wet sieving. These diatom concentrates are not suitable for quantitative analysis but give valuable information for the reconstruction of palaeo-salinities even if the sediments are very poor in diatoms. Palaeo-salinities were reconstructed from diatom assemblages according to Abelmann (1985).

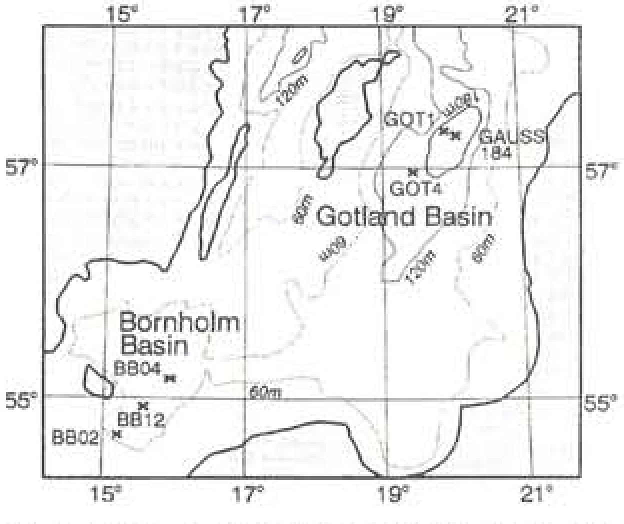

Fig. 1. Bathymetric map of the central Baltic Sea with core locations.

Table 1. **Core locations and stratigraphical range**

Core	Location		Water depth	Core length	Core diametre	Stratigraphy
GOT1KL90	57°17.90' N	19°54.39' E	230 m	4.10 m	85 mm	*Limnaea* – *Yoldia*
GOT1KL120	57°18.24' N	19°54.07' E	230 m	11.45 m	110 mm	*Limnaea* – Baltic Ice Lake
GOT2KL90	58°13.92' N	19°56.91' E	166 m	4.48 m	85 mm	*Limnaea* – *Ancylus*
GOT4KL120	56°58.94' N	19°26.83' E	175 m	10.90 m	110 mm	*Limnaea* – Baltic Ice Lake
GOT5KL120	56°53.50' N	18°56.03' E	141 m	8.82 m	110 mm	*Limnaea* – Baltic Ice Lake
GAUSS184KL	57°17.04' N	20°02.67' E	249 m	11.42 m	110 mm	*Litorina* – Baltic Ice Lake
GAUSS184VL	57°17.04' N	20°02.67' E	249 m	1.82 m	85 mm	*Limnaea* – *Litorina*
BB02	54°39.99' N	15°09.93' E	60 m	5.37 m	110 mm	*Limnaea* – Baltic Ice Lake
BB04	55°09.99' N	16°00.03' E	88 m	5.55 m	110 mm	*Ancylus* – Baltic Ice Lake
BB12	54°55.03' N	15°35.12' E	77 m	4.35 m	110 mm	*Limnaea* – Baltic Ice Lake

The stable isotope ratios $^{13}C/^{12}C$ and $^{18}O/^{16}O$ (corrected for rhodochrosite) of the carbonate minerals were measured at the University of Erlangen by Dr. Michael Joachimski using standard methods. ^{14}C AMS dates of fish bones (five samples) and sea grass (one sample) were determined by Dr. Johannes van der Plicht, Centrum voor Isotopen Onderzoek, Groningen.

LITHOLOGY

According to Winterhalter et al. (1981), the major lithological units in the central Baltic Sea are "postglacial mud", "transition clay", and "glacial clay and silt". These lithological units are well recognisable in our cores. Detailed study allows additionally the distinction of 17 lithological zones as described in Table 2. The main criteria for the distinction of these zones are colour of the fresh sediment (colours are defined according to "Munsell Soil Color Charts"), type of bedding, and distribution of iron-sulphide. These lithological zones present a new approach to stratigraphy and offer the possibility of simple and time saving lithological and stratigraphical correlations between cores from different locations within the central Baltic Sea.

The uppermost 240 to 300 cm consist of alternating layers of laminated and bioturbated sapropelic silty clay ("postglacial mud"). Generally, laminated sequences contain 5 to 9 wt.%, and bioturbated sequences 3 to 5 wt.%, organic carbon. In the basin centre, the presence of authigenic rhodochrosite causes exceptionally high manganese concentrations of up to 10 wt.%. Here, laminated horizons consist of closely alternating sapropel-, diatomite-, and rhodochrosite-layers (Fig. 2). Layers of bioturbated mud and laminated sapropel are between a few millimetres and a few decimetres thick. Rhodochrosite-layers range in thickness between 0.1 and 0.5 mm, and diatomite-layers reach a maximum thickness of 20 mm.

Sediment colours of laminated sapropels are greyish brown, dark brown or black. The thickness of the laminae varies in the range of the calculated sedimentation rate of 0.25 to 0.35 mm per year. Recent sedimentation rates before compaction are around 1.0 to 1.3 mm per year (Winterhalter et al., 1981). Therefore it is very likely that the lamination is annual. Each lamina consists of a dark, carbon-rich basal part and a lighter coloured upper part. Relics or traces of benthic organisms are absent.

Bioturbated mud is grey, olive grey, or dark grey. Sediment structures are mainly restricted to indistinct traces of burrowing organisms. Tests of the benthic, calcareous foraminifer *Elphidium excavatum* (Terquem) occur in most of the samples.

Diatomite-layers are yellowish white or pale brown. Here, the rhodochrosite content is low, and benthic organisms are absent. Rhodochrosite-layers consist of radially fibrous crystal aggregates of calcium-rich rhodochrosite (30 to 40 mole % $CaCO_3$) with diameters of 1 to 15 μm. Often, rhodochrosite-layers contain well preserved tests of *Elphidium excavatum*. All studied cores from the Gotland Basin show the same succession of bioturbated and laminated horizons; thus a subdivision into seven lithological zones (a1 to a7, Table 2) is possible.

The bluish grey "transition clay" sensu Winterhalter et al. (1981) separates the underlying varved glacial clay from the sapropelic silty clay described above. It reaches thicknesses of 2 to 3 m. With increasing sulphide content, the colour changes to dark grey and black. The dark colour is caused by amorphous or weakly crystalline iron-monosulphides. Locally these sulphides form nodules up to 15 mm thick consisting of clay cemented by minor amounts of black sulphides. Sulphur contents of dark clays vary between 0.05 and 0.1 wt.%, only some thin horizons and sulphide nodules show higher values. Obviously, very small concentrations of sulphides are sufficient to cause black staining.

Table 2. **Lithological zones and concentrations of organic carbon (TOC), sulphur and manganese in the sediments of the central Gotland Basin.**

Zone	General lithology in core GOT1KL120	Thickness	Geochemistry in GOT1KL90		
			TOC [wt. %]	S [wt. %]	Mn [wt. %]
a1	Bioturbated mud, upper part sometimes laminated. Olive grey to dark grey.	15 cm	3.6–6.2	1.4–3.2	1.2–4.3
a2	Laminated sapropel, locally slightly bioturbated. Greyish brown to black.	16 cm	6.1–9.1	2.7–6.0	0.2–4.7
a3	Bioturbated mud, some horizons indistinctly laminated. Dark grey to olive grey.	123 cm	3.2–5.2	1.1–4.6	0.6–4.9
a4	Laminated sapropel with some intercalations of bioturbated mud. Brown to olive grey.	38 cm	4.6–8.4	2.3–6.4	0.7–9.9
a5	Bioturbated mud with some intercalations of laminated sapropel. Olive grey.	15 cm	3.2–5.0	1.1–3.2	1.1–3.7
a6	Laminated sapropel with some intercalations of bioturbated mud. Brown to olive grey.	27 cm	4.3–5.7	1.8–4.2	1.6–6.2
a7	Laminated greyish brown sapropel closely alternating with bluish green silty clay.	19 cm	3.2–6.2	1.2–2.8	1.6–8.2
b1	Bluish grey to bluish green clay with irregularly distributed black sulphide-rich patches up to 5 mm thick. Indistinctly bedded in X-ray radiographs.	10 cm	2.1–3.5	1.7–2.8	0.13–0.5
b2	Bluish grey clay with sulphide-rich streaks and pyrite-marcasite filled branched burrows, these up to 20 mm long and 0.2 to 0.5 mm thick. Indistinct layering indicated by varying contents of monosulphides.	36 cm	1.0–1.8	0.6–2.4	0.07–0.13
b3	Dark grey to black clay with sulphide nodules. Mostly unstratified.	35 cm	0.6–0.9	0.2–2.0	0.06–0.13
b4	Dark grey clay covered with light grey, sulphide-free spots up to 15 mm thick. Pronounced lamination, each lamina is roughly 1 mm thick in X-ray radiographs.	35 cm	0.4–0.8	0.3–0.7	0.06–0.08
b5	Bluish grey clay with isolated, conspicuously round sulphide nodules. Lamination very pronounced, laminae between 1 and 2 mm thick.	34 cm	0.4–0.7	0.03–2.2	0.06–0.08
b6	Bluish grey clay with layers of sulphide-nodules. Often nodules grown together forming continuous sulphide layers up to 10 mm thick, alternating with bluish grey clay and markedly layered. Some horizons bioturbated.	46 cm	0.5–1.4	0.02–1.3	0.05–0.06
b7	Brown grey varved clay. Varves 1 to 4 mm thick. Basal part of each varve consisting of grey, silty clay stained by iron sulphides; upper part is sulphide-free. Ice-rafted debris rare. Single varves easily correlated between different cores.	70 cm	0.4–0.5	0.01–0.04	0.05–0.06
b8	Grey, indistinctly laminated clay, stained by iron-sulphides.	5 cm			
c1	Distal varved clay. Varves 0.5 to 2 mm thick and often visible only in X-ray radiographs. Ice-rafted debris and barite nodules frequent. Clays stained by iron sulphides restricted to some small streaks in the uppermost 20 to 30 cm of this zone. Light grey to reddish brown.	185 cm			
c2	Varved clay. Varves are 2 to 80 mm thick; ice-rafted debris and barite nodules frequent. Thickness of varves generally increasing with core depth. Light grey to reddish brown.	435 cm			

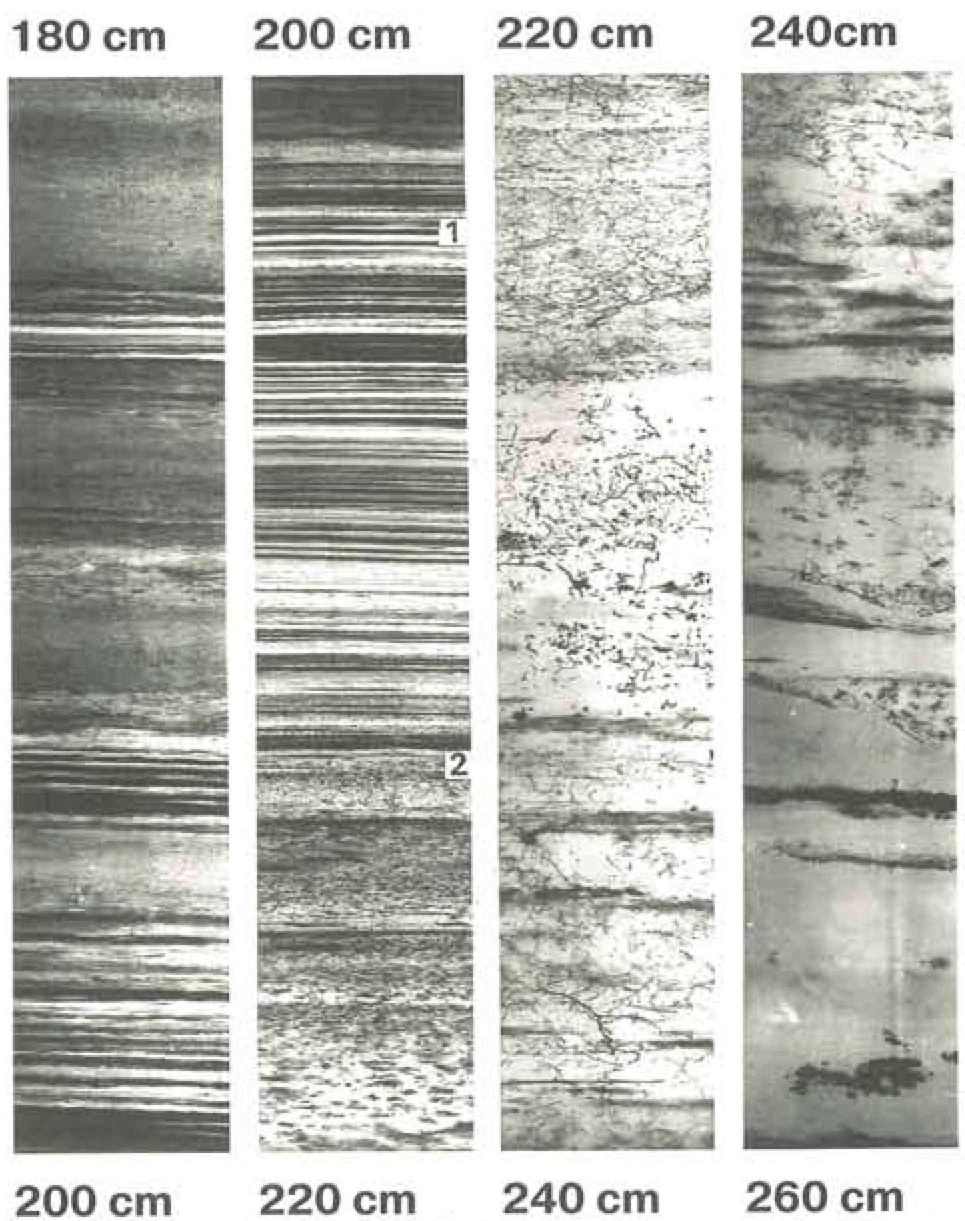

Fig. 2. Sediments transitional between *Ancylus* stage and *Litorina* stage (X-ray radiographs, core GOT1KL90).

180–203 cm: Sediments of the *Litorina* stage (zones a5 and a6). Some horizons are homogenised due to bioturbation; laminated sediments contain numerous rhodochrosite-layers (black). Typical diatom assemblages of the *Litorina* stage first occur at 203 cm core depth (1).

203–220 cm: Sediments of the *Mastogloia* stage (zones a7 and b1). Sediments of zone a7 are characterised by undisturbed lamination and many rhodochrosite-layers. The lithological boundary between "postglacial mud" and "transition clay" is between 211 and 214 cm (2).

220–260 cm: Sediments of the *Ancylus* stage (zones b2 and b3). Accumulations of iron sulphides appear black. Note branched burrows filled with pyrite and marcasite between 220 and 240 cm.

Variations in lithology as described in Table 2 can be used for a subdivision into eight lithological zones (b1 to b8).

The oldest sediments in the investigated cores are varved clays rich in ice-rafted debris ("glacial clay and silt"). Narrow streaks and small patches of clay stained with iron sulphides are restricted to the uppermost 20 to 30 cm of this lithological unit. The varves are 0.5 to 80 mm thick. A typical varve has a sharp lower boundary enriched in dropstones and up to 15 mm thick lumps of ice-rafted, undispersed till (Fig. 3). The lower third of the varve is light grey to white and consists predominantly of calcareous silt and sand. Fragments of Palaeozoic fossils indicate that the carbonates are debris derived from glacial ablation of Palaeozoic carbonates which cover a large area between South Sweden and the Baltic States. The upper part of each varve gradually becomes more fine grained, the colour changes to brown or

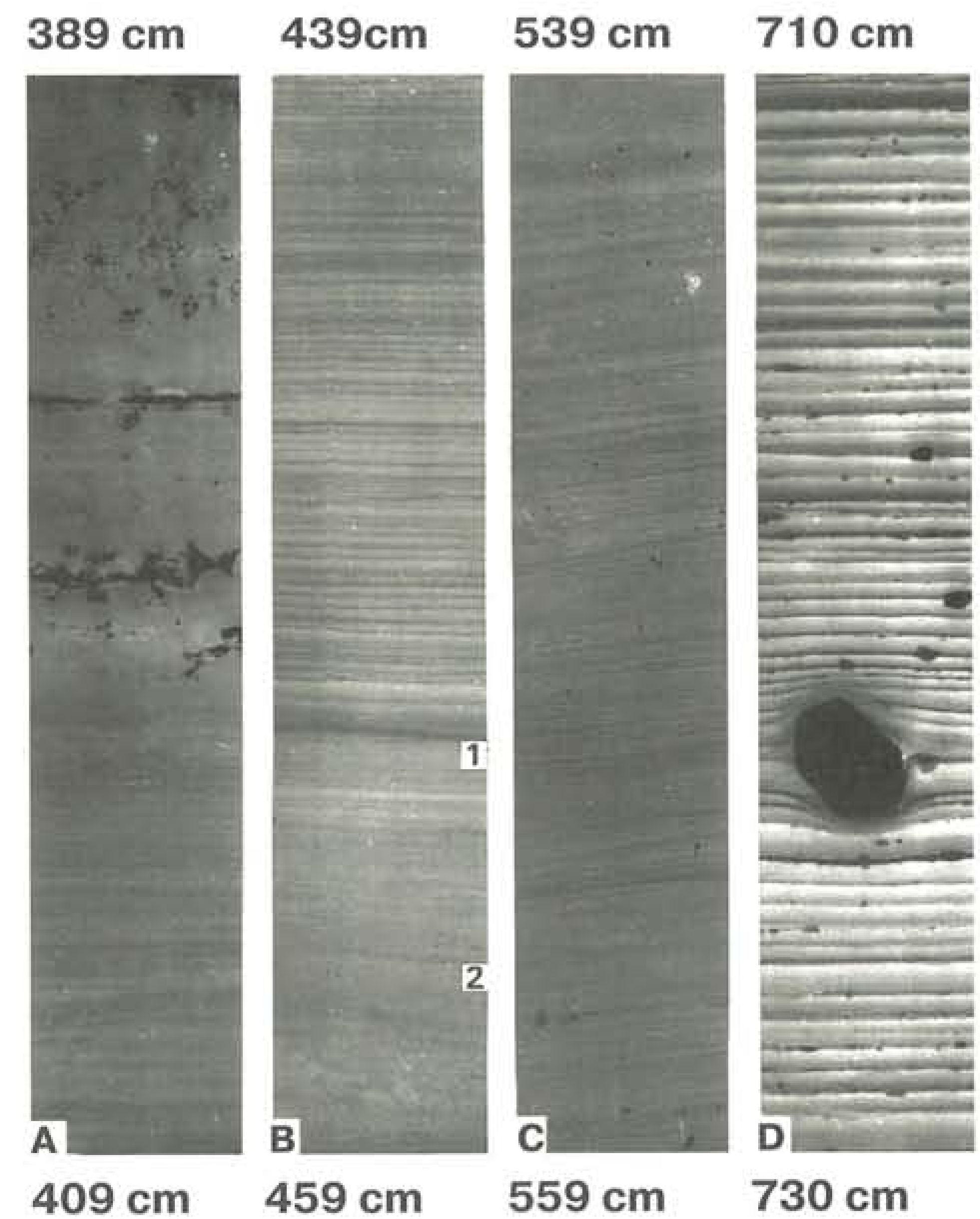

Fig. 3. Sediments of the Baltic Ice Lake and the *Yoldia* stage (X-ray radiographs, core GOT4KL120).

A: Typical, partly bioturbated sediments of the *Yoldia* Sea (zone b6). Iron-monosulphide nodules appear black.

B: Boundary between the sediments of the Baltic Ice Lake and the *Yoldia* Sea (1). The first horizon of dark clay stained by iron sulphides (zone b8) is located between (1) and (2). Note the distinct varves above this horizon (zone b7).

C: Distal varves of the Baltic Ice Lake (zone c1) rich in up to 1 mm thick barite nodules (black points).

D: Varved clays of the Baltic Ice Lake (zone c2) rich in ice-rafted debris.

reddish brown, and often some normally graded silt-layers are intercalated. The upper part of the varved clays (zone c1) consists of distal varves thinner than 2 mm. They are often visible only in X-ray radiographs. Therefore these sediments were described by Winterhalter (1992) as "homogeneous clay". Locally, erosional disconformities occur in this zone. They are enriched in coarse grained material and cut the varves at angles of up to 30°. The lower part of the varved clay (zone c2) consists of clearly visible varves which are between 2 and 80 mm thick.

GEOCHEMISTRY

The concentrations of carbon, sulphur, iron, manganese, and sodium in the Holocene sediments from the central Baltic Sea are sensitive indicators of hydrographic changes. A correlation of profiles in different cores based on geochemistry is often more precise than a correlation based solely on lithology (Fig. 4). The following geochemical description of the Holocene sediments focuses on core GOT1KL90 from the central Gotland Basin (230 m water depth) which contains a complete sequence from the

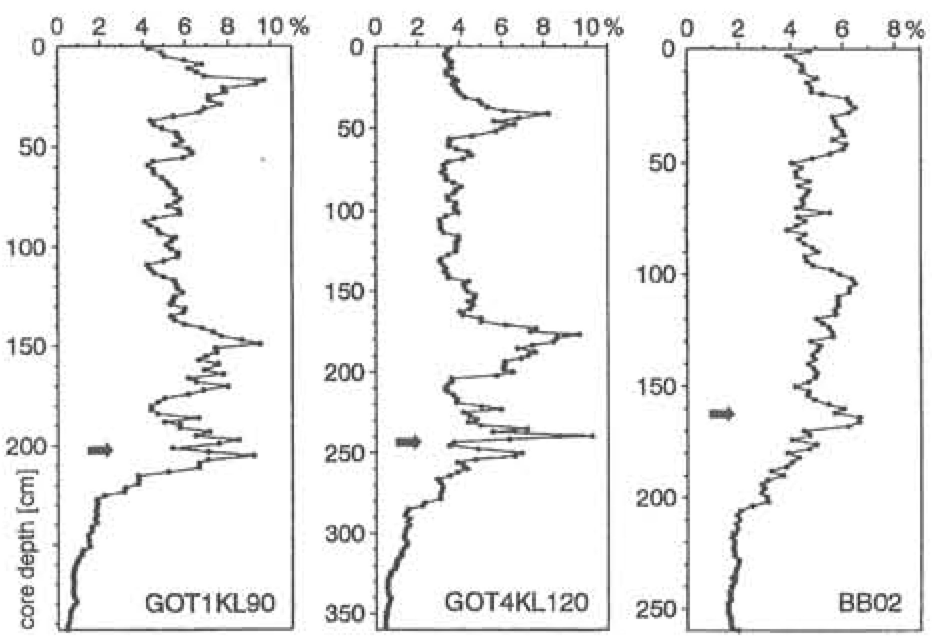

Fig. 4. Total carbon in three cores from the Gotland Basin (GOT1, GOT4) and Bornholm Basin (BB02). The base of the profiles is in the middle part of the *Ancylus*-sediments. Arrows indicate the first occurrence of typical diatom assemblages of the *Litorina* stage. Note the similar pattern of carbon contents in cores from different locations.

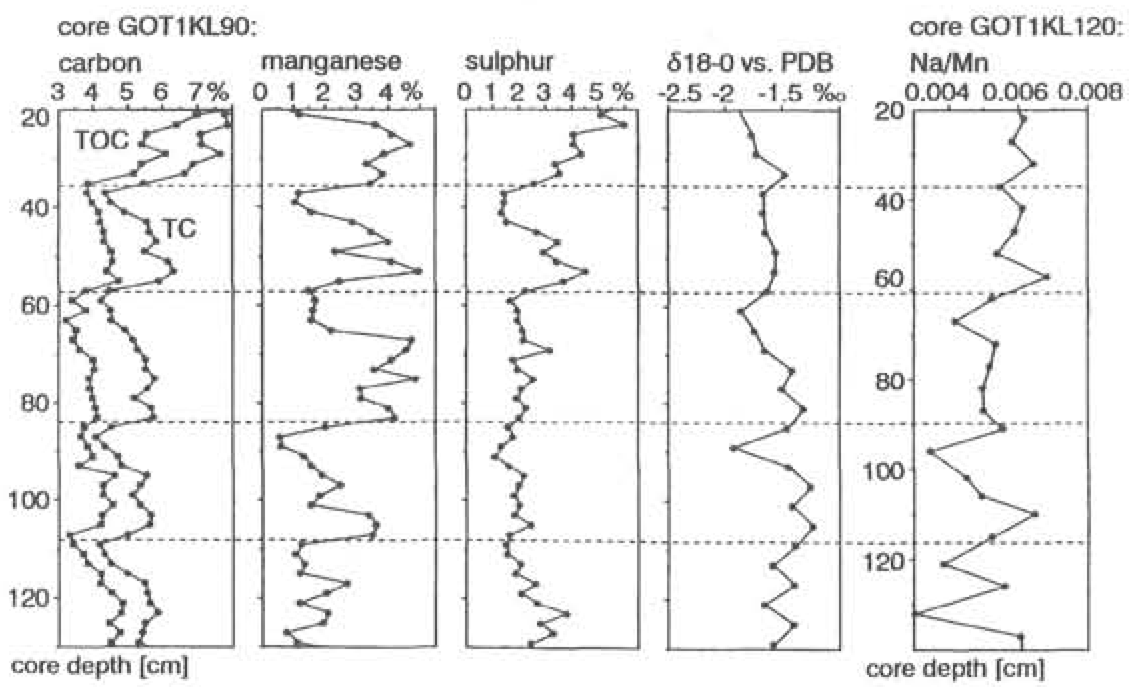

Fig. 5. Geochemical characteristics of the sediments of the *Limnaea* stage. Broken line indicates cyclic increase in carbon contents. High contents of organic and inorganic carbon are often positively correlated with manganese concentrations, $\delta^{18}O$-values, and Na/Mn-ratios.

Limnaea stage to the lower part of the *Yoldia* stage. The lithological zones mentioned in the text refer to Table 2.

The occurrence of inorganic carbon is mainly restricted to lithological zones a1 to a7 (Fig. 5). Inorganic carbon is predominantly fixed in rhodochrosite; highest values are around 2 wt.%. Delta^{13}C-values ranging from −7 to −12‰ vs. PDB indicate that roughly 30 to 50% of inorganic carbon originated from bacterial degradation of organic matter (Suess 1979). Total carbon contents of the laminated sapropel range from 6 to 10 wt.%, whereas bioturbated mud contains only 4 to 6 wt.% carbon. Carbon concentrations of zones a1 to a6 show a marked cyclicity (Fig. 5, Fig. 7). Within the "transition clay", carbon contents decrease from

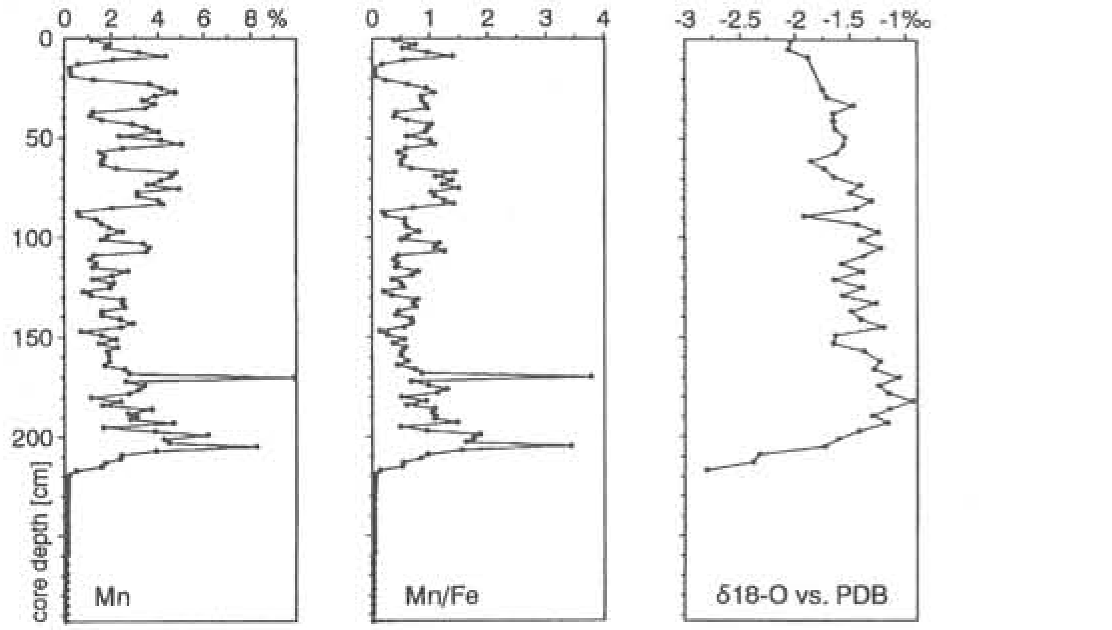

Fig. 6. Manganese contents, Mn/Fe- and $\delta^{18}O$-ratios in core GOT1KL90. The base of the profile is in the middle part of the *Ancylus*-sediments. According to diatoms, the lower boundary of the *Litorina* stage is at 203 cm core depth. The iron to manganese-ratio is exclusively controlled by the changing concentrations of manganese because iron contents are rather constant. $\delta^{18}O$-ratios depend mainly on changing salinities which reach their highest values around 18‰ at 175 cm core depth.

4 wt.% in zone b1 to approximately 0.5 wt.% in zone b4. Zones b5, b6 and b7 have a rather constant carbon content of approximately 0.5 wt.%; higher values of up to 1.6 wt.% are locally observed within zone b6.

The sulphur contents of sapropelic silty clay (zones a1 to a7) vary between 1.1 and 6.4 wt.%. Generally, sulphur and carbon are positively correlated, but the sulphur concentrations have a less pronounced cyclicity. Within the upper part of the "transition clay" (zones b1 to b3), sulphur contents fluctuate more than carbon contents but show the same downwards decreasing trend. Minimum values of around 0.03 wt.% occur between zone b5 and zone b7; considerably higher sulphur-values up to 1.3 wt.% are restricted to some thin horizons within zones b5 and b6. The sulphidisation of detrital iron-minerals due to the reaction with H_2S during diagenesis (Berner 1970) can be expressed by the ratio of sulphur to reactive iron. In our cores, this ratio depends mainly on the concentration of sulphur because reactive iron is remarkably constant with values of 3 to 5 wt.%. High concentrations of sulphur therefore indicate a high degree of sulphidisation which points to the presence of marine water, and also to anoxic sulphidic conditions during diagenesis.

Manganese plays an important role in the geochemistry of the central Baltic Sea because authigenic rhodochrosite occurs in large quantities in the sapropelic silty clay. Manganese concentrations of zones a1 to a7 vary between 0.2 and 10 wt.%, they decrease rapidly in zones b1 and b2 from 2 wt.% to 0.1 wt.%. Zones b3 to b7 have very low and constant manganese concentrations between 0.05 and 0.1 wt.% (Table 2, Fig. 6). The sodium content of rhodochrosite varies between 1500 and 4500 ppm. Generally, the Na/Mn-ratio is higher in laminated sapropels (0.006 to 0.008) than in bioturbated mud (0.004–0.006).

Various models have been proposed to explain the extraordinary enrichment of manganese and the formation of rhodochrosite (Manheim 1961, Hartmann 1964, Suess 1979, Jakobsen & Postma 1989). Emelyanov (1986) first mentioned the significance of major North Sea water inflows for the high concentration of manganese. These well known inflows of oxygenated water occur episodically at intervals of several years (Fonselius 1962). Huckriede (1994) described a model for the formation of rhodochrosite as a result of drastic changes in redox conditions caused by major inflows: Due to pronounced salinity stratification in the water column, bottom waters are anoxic and therefore leach manganese from the sediments. Episodic inflows of denser, oxygenated water from the North Sea into the deeps of the central Baltic Sea cause oxidation of dissolved Mn^{2+} to particulate oxides. Fine-grained manganese oxide particles accumulate preferentially in the deepest parts of the basins. Here anoxic conditions prevail, alkalinity is high due to intense bacterial sulphate-reduction (Berner et al. 1970), and manganese is fixed as rhodochrosite. Multiple repetitions of these drastic changes in redox conditions of the bottom waters cause manganese depletion in most areas below the halocline and simultaneous enrichment of manganese within the deepest basins.

PALAEOHYDROGRAPHY AND STRATIGRAPHY

The correlation of basinal sediments, littoral sediments and the Baltic shore displacement is still a problem (Hyvärinen 1988). One reason for these difficulties is the absence of molluscs and the low diversity and poor preservation of diatoms in basinal sediments. The following correlation between the sediment record in the Gotland Basin and the Baltic stages and the reconstruction of the hydrography are based on a combination of geochemical, lithological and palaeontological evidence. The lithological zones are described in Table 2.

Baltic Ice Lake

Sulphide-free varved clays rich in ice rafted debris (zones c1 and c2) are typical sediments of the Baltic Ice Lake. The upwards decreasing varve thickness corresponds to the increasing distance to the melting ice shield. The brownish red tint of the sediment indicates an oxic environment during sedimentation. Surprisingly, detrital pyrite grains show no signs of oxidation. Apparently, this is a result of rapid deposition and weakly reducing conditions during diagenesis. Frustules of diatoms are completely absent. Most probably, this is due to extensive dissolution of amorphous SiO_2. Correlation of varves in different cores indicates that the transition between lithological zones c1 and c2 is synchronous in different parts of the Gotland Basin.

Barite nodules with diameters up to 2 mm occur frequently in the sediments of the Baltic Ice Lake (Fig. 3). According to Suess (1982), they formed diagenetically due to ion exchange reactions on clay mineral surfaces when freshwater deposits were permeated by marine waters. Therefore these nodules point to a freshwater origin of their host sediments. The upper limit of the occurrence of barite nodules coincides roughly with the lower boundary of zone b8. Within the "transition clay", barite nodules occur as rare exceptions within the lower part of zone b7. More frequently, they are found once again in zone b4 which is clearly of freshwater origin (Fig. 7).

Despite the lack of fossils usable for the reconstruction of the palaeo-environment, a freshwater origin of the sulphide-free varved clay is indicated by the occurrence of barite nodules and the absence of diagenetic iron sulphides. A connection with the North Sea would have caused salinity stratification with resulting oxygen deficient bottom waters favouring the formation of iron-sulphides during diagenesis.

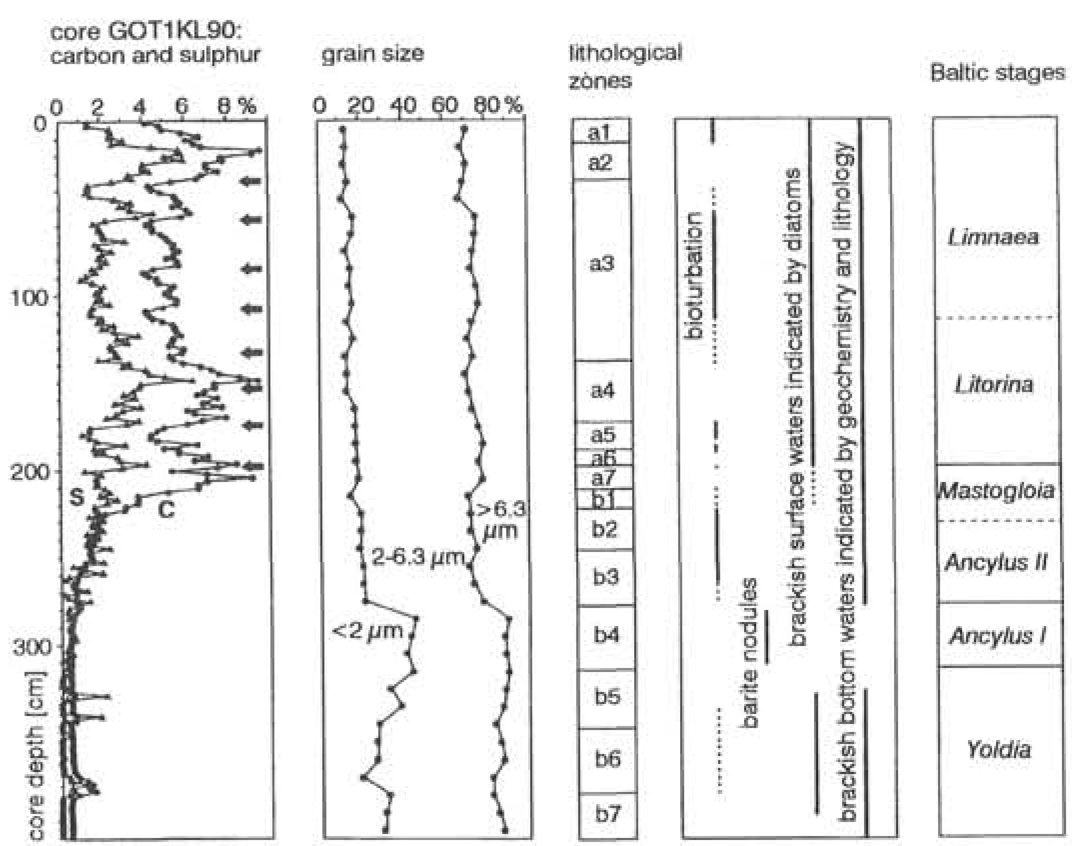

Fig. 7. Lithological zones, carbon and sulphur contents, grain size distribution and stratigraphy in core GOT1KL90 from the central Gotland Basin. Arrows indicate periodic increases in carbon and sulphur contents within the *Litorina* and *Limnaea* stages.

Yoldia Sea

The upper boundary of the sulphide-free varved clays is marked by a layer of dark clay stained with iron sulphides (zone b8). Immediately below this boundary, varves are indistinctly developed with individual thicknesses of 0.5 to 1.0 mm. In contrast, the varves of the overlying clay (zone b7) are up to 4 mm thick, contain little iron sulphides and some layers of silt and sand (Fig. 3). Individual varves of zone b7 can be easily correlated between cores from the Gotland Basin indicating a synchronous deposition of the sediments of zones b8 and b7. Obviously, the transition between "glacial clay and silt" and "transition clay" has been caused by a change in hydrographic conditions that affected large parts of the Baltic basin. The presence of iron sulphides points to marine influence because the formation of sulphide is favoured by sulphate in the water and by anoxic conditions as a consequence of salinity stratification. The increased thickness of the varves can be best explained by a rapid drop of the water level which caused an enhanced sediment supply due to erosion of soft sediments from former subaquatic areas. This event was the drainage of the Baltic Ice Lake at Mt. Billingen in South Central Sweden which lowered the water level by 26 m to 29 m (Winterhalter et al. 1981) and initiated the brackish *Yoldia* stage. Layers of sand and silt within glacial clay in Hanö Bay in south-east Sweden have been previously ascribed to the drainage of the Baltic Ice Lake by Björck et al. (1990). According to the Swedish varve chronology, the drainage event is dated 10,430 YBP (Björck et al. 1992).

The first diatoms occur 5 to 10 centimetres below the boundary between zones b7 and b6. Dominating species is *Thalassiosira baltica* (Grunow) Ostenfeld, which indicates cold and brackish water with maximum salinities up to 10 ‰ (Snoeijs 1993). Other species occurring in rapidly fluctuating quantities are *Stephanodiscus astraea* (Ehrenberg) Grunow, *Melosira islandica* O. Müller, and *Diploneis smithii* (Brebisson) Cleve. The association of *Thalassiosira baltica* with the freshwater indicating diatom *Stephanodiscus astraea* has been previously reported by Abelmann (1985) from sediments of the *Yoldia* stage in the Karlsö Basin. This association clearly indicates salinities lower than in the present Baltic Sea.

The occurrence of *Thalassiosira baltica* is mainly restricted to zone b6 and to the lowest part of zone b5. The fragility of this diatom and a positive correlation between its abundance and the sulphide content of the sediment exclude re-deposition from older marine sediments as proposed by Eronen (1974) for marine diatoms in sediments of the *Yoldia* stage in Finland. In some horizons, high carbon and sulphur contents point to salinity stratification of the water column and to anoxic conditions in bottom waters. In contrast to the modern Baltic Sea, constantly low manganese contents indicate the absence of rapid fluctuations between oxic and anoxic sulphidic conditions in bottom waters. Based on annual lamination the *Yoldia* stage lasted 600 to 700 years.

Ancylus Lake

The diatoms of zones b4 to b1 are typical of the freshwater *Ancylus* Lake. Most abundant are *Melosira islandica*, *Stephanodiscus astraea*, *Gyrosigma attenuatum* (Kützing) Rabenhorst, *Cymatopleura elliptica* (Brebisson) W. Smith, and *Cymatopleura solea* (Brebisson) W. Smith. The occurrence of benthic diatoms (*Gyrosigma attenuatum*, *Cymatopleura elliptica*, *Cymatopleura solea*) indicates redeposition from shallower areas. The uniform distribution of diatom species within *Ancylus* Lake sediments is at contrast with marked changes of carbon and sulphur contents. Sulphur and carbon are low in zone b4, they increase in zone b3, and reach maximum values in the lower part of the "postglacial mud". Boesen & Postma (1988) explained the unexpected high sulphur contents of the upper part of the *Ancylus*-sediments by diagenetic sulphidisation due to downward migrating H_2S. However, the positive correlation of organic carbon and sulphur in this horizon ($r^2 = 0.53$, $n = 40$) and appreciable fluctuations of sulphur contents, which are similarly developed in different cores, disprove this model. In our opinion, the increasing sulphur and carbon contents in the upper part of the *Ancylus*-sediments indicate oxygen-deficient bottom waters due to salinity stratification.

Grain size is another indication of palaeo-salinity (Fig. 7). Brackish conditions favour rapid sedimentation of clay minerals through flocculation. The fine-grained material is preferentially deposited near-shore whilst the basin sediments are relatively poor in clay. In core GOT1KL120, markedly increased contents of fines < 2 µm occur exclusively in the upper part of zone b5 and in zone b4. Between zones b4 and b3 there occurs a drastic decrease in fines whereas the sulphur content rises simultaneously (Fig. 7). This indicates that the connection with the North Sea was blocked for only a short period at the beginning of the *Ancylus* stage. Subsequently, a new connection to the North Sea was established, and brackish bottom waters reached the central Baltic Sea, causing temporary stagnation and flocculation of clay minerals. However, until the beginning of the *Mastogloia* stage (zone b1), salinities in surface waters were too low to induce changes in diatom assemblages from freshwater to brackish waters.

Two phases in the development of the *Ancylus* Lake can be distinguished: The first phase (*Ancylus* I, Fig. 7) shows no marine influence, whereas the second phase is characterised by brackish bottom waters (*Ancylus* II). The first phase lasted 400 to 500 years according to counts of laminae in zone b4. Kolp (1986) presented similar results from mapping and dating of submarine terraces in the western Baltic Sea. He found the level of the *Ancylus* Lake independent from world ocean between 9,300 and 8,800 YBP. Other arguments supporting a bipartite history of the *Ancylus* Lake are discussed by Björck (1987).

Mastogloia Sea

Despite the drastic changes in lithology at the boundary between "transition clay" and "postglacial mud", no associated significant change in the composition of diatom assemblages is recognisable (Fig. 2). Even laminated sapropels of zone a7 with up to 10 wt.% carbon are dominated by freshwater diatoms such as *Melosira islandica* and *Stephanodiscus astraea*. Only the occasional appearance of *Diploneis smithii* in zones b1 and a7 points to temporary slightly brackish conditions in surface waters, indicating the *Mastogloia* stage. The top of zone a7 is the upper boundary of *Mastogloia*-sediments in the Gotland Basin. This boundary shows no significant lithological change (Fig. 2) but is characterised by a rapid increase in sulphur content and by the total disappearance, within a few centimetres, of freshwater diatoms. The boundary to the *Ancylus*-sediments is not exactly defined because brackish diatoms are very rare and salinities of bottom waters rose gradually for approximately 800 to 1,000 years. The increase of carbon and sulphur between zones b2 and b1 probably points to an increased inflow of brackish water which may indicate the beginning of the *Mastogloia* stage.

Litorina Sea and *Limnaea* Sea

Diatoms of zones a6 to a1 indicate brackish conditions. Some typical species are *Actinocyclus ehrenbergi* Ralfs, *Thalassiosira eccentrica* (Ehrenberg) Cleve, *Chaetoceros subsecundus* (Grunow) Hustedt, *Rhizosolenia calcar avis* M. Schultze, *Thalassionema nitzschioides* Grunow, *Rhabdonema arcuatum* (Agardh) Kützing, *Grammatophora oceanica* (Ehrenberg) Grunow, and *Diploneis didyma* (Ehrenberg) Cleve.

A subdivision of the "postglacial mud" based on diatoms is impossible because changes in diatom assemblages mainly result from selective dissolution of opal. Diatoms are best preserved in laminated sapropels from the deepest parts of the Gotland Basin whereas bioturbated mud contains only species less sensitive to dissolution of amorphous SiO_2 like *Actinocyclus ehrenbergi*.

Calcium-rich rhodochrosite is a typical mineral of the "postglacial mud" and originated from early diagenetic reduction of manganese oxides at the sediment-water interface (Huckriede 1994). Its isotopic composition and geochemistry reflect the contemporaneous bottom waters. The oxygen isotope ratios of rhodochrosite (Fig. 6) should be controlled by salinity and temperature of bottom waters. A linear relationship between salinity and oxygen isotope composition in the waters of the Baltic Sea has been documented by Fröhlich et al. (1988). Limited data are available for temperature-dependent fractionation of oxygen isotopes during precipitation of Ca-rhodochrosite (Böttcher 1993); therefore we assume that this fractionation is approximately similar to fractionation in calcite. Oxygen isotope ratios of carbonates at equilibrium with Baltic sea water should be positively correlated to salinity and negatively correlated to temperature. According to Epstein et al. (1953) and data in Fröhlich et al. (1988), the effect of a salinity increase of 1‰ on oxygen isotope ratios in carbonate is roughly compensated by a temperature increase of 1°C. During the *Mastogloia*, *Litorina* and *Limnaea* stages, salinities of bottom waters were most probably more variable than temperatures. Thus changing oxygen isotope ratios can be used for an indication of changing palaeo-salinities. Recent temperatures of bottom waters in the central Gotland Basin are rather constant at 4 to 6°C (Nehring 1990, Nehring et al. 1994). Based on constant temperature during rhodochrosite-formation and on present day salinities of 12‰ during sedimentation of the uppermost sediment, the calculated palaeo-salinities rose from 8‰ in zone a7 to 18‰ in the lower part of zone a5. From the upper part of zone a5 to zone a1, salinities gradually dropped to present day values of 12‰. This general trend is superimposed by smaller variations which show a weak positive correlation with carbon and manganese contents (Fig. 5).

The sodium concentration in rhodochrosite is another indicator of palaeo-salinities. The Na/Mn-ratio is weakly positively correlated with the oxygen isotope composition of rhodochrosite. High oxygen isotope ratios and high Na/Mn-ratios in rhodochrosite occur preferentially in horizons rich in organic and inorganic carbon (Fig. 5). Additionally, these horizons are rich in tests of the benthic foraminifer *Elphidium excavatum* indicating an enhanced water exchange with the North Sea. Oxygen isotopes, Na/Mn-ratios of rhodochrosite, and the distribution of foraminifers together therefore indicate that high salinities of bottom waters cause stable stratification and hence anoxic conditions.

The following model explains the positive correlation between anoxic bottom waters and salinity. Increased inflow from the North Sea results in higher salinity and pushes the halocline to shallower depths. According to data in Matthäus (1986), a salinity increase in the bottom waters of the Gotland Basin from 12.0 to 12.5‰ raises the centre of the halocline from 80 to 62 m water depth. The deep water below the halocline is the main source of nutrients for primary production in the photic zone. As the halocline moves upward, more nutrients can reach the photic zone due to diffusion and mixing and the primary production increases. As a consequence, oxygen consumption by degradation of organic matter is enhanced, and anoxic conditions occur more easily. In contrast, absence of salt water inflows during some years or decades results in decreasing salinities of bottom waters, and eventually cold winter water can penetrate the halocline and transport oxygen into the basins (Nehring 1990). Thus the stability and the depth of the halocline are the controlling factors for the oxygenation of bottom waters during the *Litorina* and *Limnaea* stages. The oxygenation of bottom waters by salt water inflows is of minor importance.

COMPARISON WITH THE SEDIMENTS OF THE BORNHOLM BASIN

The Bornholm Basin has a maximum depth of 105 m and is the deepest region of the Baltic Sea southwest of the Gotland Basin. Due to the shallower water depth and more varying detritus supply, the lithological development of the Bornholm Basin is dominated by local variations compared to the Gotland Basin. Nevertheless, most lithological zones defined in the Gotland Basin can be easily correlated with sediments in the Bornholm Basin. This demonstrates that the sedimentary development in the Gotland Basin is representative for at least a major part of the Baltic Sea.

The "postglacial mud" in the cores from the Bornholm Basin is up to 250 cm thick. A subdivision in lithological zones is difficult because the sediment is intensively bioturbated. Only the lowermost decimetres are indistinctly laminated and contain some rhodochrosite-layers. Diatomite-layers are completely absent. However, carbon and sulphur contents show a similar development as in the cores from the Gotland Basin. Three horizons with high carbon concentrations indicate the positions of lithological zones a2, a4, and a6/a7 (Fig. 4).

Similar to the sedimentary development in the Gotland Basin, the boundary between "transition clay" and "postglacial mud" shows no significant change in diatoms. Only some horizons containing the benthic diatoms *Campylodiscus clypeus* Ehrenberg and *Epithemia turgida* (Ehrenberg) Kützing indicate weakly brackish conditions near this boundary. Typical diatom assemblages of the *Litorina* stage containing *Rhizosolenia calcar-avis*, *Actinocyclus ehrenbergi* and *Rhabdonema arcuatum* first occur 15 to 30 cm above this boundary. Intense leaching of amorphous SiO_2 has destroyed diatoms in most horizons, therefore a reconstruction of palaeo-salinity based on diatoms is difficult.

The thickness of "transition clay" varies between 1.2 and 3 m. Zones b1, b2, b6, b7, and b8 as defined in the Gotland Basin (Table 2) can be correlated to similar horizons in the Bornholm Basin. In cores BB12 and BB02, zones b7 and b8 contain up to 5 cm thick layers of calcareous silt and sand alternating with dark laminated clay. Evidently, the increased supply of detritus caused by the drainage of the Baltic Ice Lake was more intense in the Bornholm Basin than in the Gotland Basin. The sediments of the Bornholm Basin corresponding to zones b3, b4, and b5 show a more indistinct development. They consist of bioturbated to weakly laminated dark clays which are rich in screw-shaped sulphide concretions. In core BB02, these sediments locally contain enriched organic detritus; possibly re-deposited peat. The glacial sediments in the Bornholm Basin are very similar to those in the Gotland Basin but a correlation of varves is not possible.

Kögler & Larsen (1979) introduced the name "AY-clay" for the sediments between clearly layered varved glacial clay and marine mud rich in organic matter, and divided these sediments into three zones. They were not successful in correlating these zones with the Baltic stages. According to our correlation between the sediments of the Gotland Basin and the Bornholm Basin, zones A and B of Kögler & Larsen (1979) correspond to the *Ancylus* and *Yoldia* stages, whereas zone C is composed of indistinctly varved clays of the Baltic Ice Lake. The boundary between the sediments of the *Yoldia* Sea and the *Ancylus* Lake corresponds to the top of the sulphide-rich horizon within the upper third of zone B.

CYCLIC SEDIMENTATION AND CHANGING CLIMATE

Carbon and sulphur contents of zone a1 to a6 show a distinct cyclic periodicity (Fig. 5, Fig. 7). The eight cycles are best developed in the sediments of the central Gotland Deep, but are also recognisable in the western part of the Gotland Basin (core GOT4, Fig. 4). A typical cycle is between 20 and 30 cm thick and starts with a rapid increase in organic and inorganic carbon. Simultaneously, sulphur, manganese, $\delta^{18}O$-values, and Na/Mn-ratios rise. The highest carbon-values are reached approximately 5 cm higher. Often a second carbon peak follows some centimetres further up. Towards the top of the cycle, carbon contents decrease gradually.

As discussed above, the increase of $\delta^{18}O$-values and Na/Mn-ratios at the beginning of each cycle points to a rapid increase in salinity. This is confirmed by simultaneously rising manganese contents, indicating frequent inflows of North Sea water. During such periods of high salinities, anoxic conditions in bottom waters become established for 300 to 400 years. Subsequently, the inflow of salt water from the North Sea diminishes, salinities decrease, and oxic periods in bottom waters occur at increasing frequency until a new cycle starts after a period of another 400 to 600 years.

Assuming constant detritus input and an absolute age of 7,500 YBP (Hyvärinen 1984) for the boundary between *Litorina* and *Mastogloia* stages, the average duration of each cycle is 850 ± 40 years. This is roughly confirmed by ^{14}C AMS dating. The uppermost cycle shows very good correspondence with the climate history of the last 1,000 years: Carbon-rich sediments accumulated during the medieval warm period, which lasted approximately from AD 900 to 1300 (Lamb 1984). Low carbon-values characterise the sediments of the Little Ice Age (AD 1400 to 1900). The climatic changes following the end of the Little Ice Age are again linked with anoxic conditions in bottom waters: Since the beginning of the 20th century, regular hydrographic measurements revealed an increase of salinity until 1975 (Fonselius 1962, Matthäus 1986, Nehring & Matthäus 1990). Simultaneously, oxygen contents of bottom waters decreased rapidly (Fonselius 1969), and laminated sediments spread over large areas (Jonsson & Jonsson 1988).

The unstable hydrography of the central Baltic Sea responds sensitively to even weak climatic changes. The results of these climatic variations are drastic changes between oxic and anoxic conditions in bottom waters which are clearly documented in the sedimentary record. It is very likely that all eight cycles are controlled by climatic events like the Little Ice Age. The sediments of the deep basins of the central Baltic Sea therefore offer the possibility for a detailed reconstruction of climatic changes during the last 7.500 years.

The linkage between climate and oxygenation of bottom waters is possibly controlled by large-scale variations in the atmospheric circulation that cause either changes in the intensity and frequency of storms that induce inflows of North Sea water, or changes in the humidity in the drainage area of the Baltic Sea, which would have drastic effects on the outflow through the Danish Sounds (Fonselius 1969, Kaleis 1976, Börngen et al. 1990). The individual triggering mechanisms for inflow events are still under discussion (Dickson 1971, Welander 1974, Kaleis 1976, Kullenberg 1983, Börngen et al. 1990, Matthäus & Franck 1992).

Up to now, Holocene climatic cycles with periodicities about 1,000 years have rarely been described. Kellogg (1984) discovered periodic changes in the abundance of subpolar foraminifers in Holocene sediments in the Denmark Strait. Maximum abundances occur at intervals of approximately 800 years. Overpeck (1987) investigated twenty pollen records from the midwestern United States and recognised a cyclicity around 1,100 years. Investigations of Mörner (1980), Colquhoun & Brooks (1987) and Ters (1987) point to minor sea-level changes with a similar periodicity.

SUMMARY AND CONCLUSIONS

The sediments of the central Gotland Basin contain a complete and nearly undisturbed record of hydrographic changes during the last 12,000 years. Variations in lithology, geochemistry and diatom assemblages are linked to the history of the Baltic Sea. The lithological development in the central Gotland Basin and in the shallower Bornholm Basin are closely related. Therefore we suggest to use the sediments of the central Gotland Basin as a standard for the lithological and hydrographical development of large parts of the Baltic Sea. Based on marked changes in lithology, the late glacial and Holocene sediments in the Gotland Basin can be subdivided into 17 lithological zones.

The boundary between the sediments of the Baltic Ice Lake and the *Yoldia* stage is identified by the first appearance of dark clays stained by iron sulphides, and by a rapid rise in varve thickness. The varves above this boundary can be correlated between different cores, indicating synchronous deposition. The lithological and geochemical changes are caused by the drainage of the Baltic Ice Lake at Mt. Billingen and by the subsequent inflow of marine waters some 10,400 YBP. The overlying sediments of the *Yoldia* stage contain diatom assemblages characterised by *Thalassiosira baltica*. The diatom assemblages and the geochemistry of *Yoldia*-sediments point to brackish salinities in surface waters which were generally lower than in the modern Baltic Sea. Stable stratification of the water column caused anoxic bottom waters for only short periods.

The sediments of the *Yoldia* stage are immediately overlain by limnic sediments of the *Ancylus* Lake. No evidence has been found of the existence of a separate *Echeneis* stage. The sediments of the *Ancylus* stage are divided into two parts. The lower part shows no marine influence for approximately 500 years. Most probably, it corresponds to the *Ancylus*-transgression known from shore line displacement. The boundary to the upper part of the *Ancylus*-sediments is marked by an increase in sulphur and carbon and by a rapid decrease of the clay-size fraction. This indicates a weak marine influence in bottom waters. Towards the top of this unit, carbon and sulphur concentrations rise gradually due to an increasing stability of the halocline. The boundary to the sediments of the *Mastogloia* stage is a matter of convention because salinities rose gradually for approximately 1,000 years within the upper part of the *Ancylus* stage and during the *Mastogloia* stage. Brackish conditions in surface waters during the *Mastogloia* stage are indicated only by a few brackish diatoms near the boundary between "transition clay" and "postglacial mud". The exceptionally sharp lower boundary of "postglacial mud" is located within the sediments of this stage and does not correspond to any important change in diatom assemblages. Approximately 15 to 20 cm higher, increasing sulphur contents and the disappearance of freshwater diatoms mark the beginning of the *Litorina* stage.

From the beginning of the *Litorina* stage until today, there occurred no significant hydrographic changes. This is proven by a uniform distribution of diatom species and by the occurrence of authigenic rhodochrosite indicating drastic and frequent changes between anoxic sulphidic and oxic conditions in bottom waters. Such changes characterise the modern hydrography of the Baltic Sea. Rhodochrosite originated from early diagenetic reduction of manganese oxides at the sediment-water interface. Its isotopic composition and geochemistry therefore reflect the hydrographical properties of the contemporaneous bottom waters. According to oxygen isotope ratios of rhodochrosite, rapidly rising salinities within the bottom waters of the Gotland Basin reached a maximum of 18‰ during the middle part of the *Litorina* stage. Since this time, salinities decreased gradually to present-day values of 12 to 13‰. Minor periodic changes in salinity are superimposed on this trend. They are documented by oxygen isotopes and sodium contents of rhodochrosite and – most convincing – by periodic carbon and sulphur concentrations. High salinities occur together with anoxic conditions whereas lower salinities favour oxic periods in bottom waters. Controlling factors for the oxygenation of bottom waters are stability and depth of the halocline. The duration of the cycles is between 800 and 900 years.

The good correlation between climate and geochemistry in the uppermost cycle demonstrates that the periodicity is induced by climatic changes: During the medieval warm period, bottom waters were mostly anoxic. Oxic conditions occurred more frequently during the Little Ice Age. The linkage between climate and oxygenation of bottom waters is caused by large scale variations in the atmospheric circulation that caused changes in the water exchange between the North Sea and the Baltic Sea. The unstable hydrography of the central Baltic Sea responds sensitively to even weak climatic variations. The results are drastic changes between oxic and anoxic conditions in bottom waters which are clearly documented in the sedimentary record. Therefore the sediments of the central Baltic Sea offer the possibility for a detailed reconstruction of climatic changes during the last 7.500 years.

ACKNOWLEDGEMENTS

We are indebted to Dr. Klaus Figge (Bundesamt für Seeschiffahrt und Hydrographie, Hamburg) for technical and scientific support. Dr. Michael Joachimski (Erlangen) measured the carbon and oxygen isotopes. Ulrike Meliß (Göttingen) assisted with laboratory work. We wish to thank them all. Further thanks are due to Rolf Kunze, Rencken Paar, the captain and crew of R/V VALDIVIA during cruise 138/3 for skilful assistance. This study

was supported by German Science Foundation (DFG) grant No. Me 267/37–1 to Dieter Meischner. The cores from the Bornholm Basin were investigated within the project No. 03F0101A of the Federal Ministry of Science and Technology (BMBF).

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BALTICA REVIEW

A WEIGHTY CONTRIBUTION INTO THE GEOLOGY OF THE SOUTH BALTIC AREA

ATLAS GEOLOGICZNY POŁUDNIOWEGO BAŁTYKU / GEOLOGICAL ATLAS OF THE SOUTHERN BALTIC. 1:500 000. PAŃSTWOWY INSTYTUT GEOLOGICZNY. SOPOT–WARSZAWA 1995. I-XXXIV PLATES (IN COLOURS)

Of late years the interest in the Baltic Sea has grown tremenduously. This can be explained by a removal of the barrier that divided the Sea into the Soviet and non-Soviet parts. Moreover, marine investigations related to more rational use of its resources and protection from its pollution are very important for millions of people living around the Baltic Sea; finally, the Sea provides excellent opportunities to test different research methods, to realize national and international projects with the Baltic Sea geologists being especially active in these studies.

Among the newest achievements, a wonderful Geological Atlas of the Southern Baltic should be mentioned. Its authors (R.Dadlez, K.Kenig, R.Kramarska et al.) and editorial board (J.E.Mojski, Editor-in-Chief) submitted to the geological society a large generalizing treatise about the geological structure of the Polish part of the Baltic Sea, the Quaternary history, recent sedimentation, Sea bottom topography, bottom deposits, geochemistry and raw materials.

After the World War II, Prof. E.Rühle was interested in the Baltic Sea studies in Poland, the initial investigations were conducted by F.B. Pieczka, K.Wypych and B.Rosa - researchers from Gdańsk Polytechnical Institute. During last three decades, marine geological investigations are being carried on at the Marine Geology Branch of the Polish Geological Institute (J. Zachowicz, J.E. Mojski, Sz. Uścinowicz, R. Kramarska, T. Szczepańska, A. Tomczak et al.).

Geological maps of different types, sections, schemes depicting geological structure of the Sea, its evolution, recent sedimentation and geochemical processes are given in the Atlas. The material of the Atlas can be naturally grouped into several parts: pre-Quaternary structure, Quaternary system and structure, bottom topography and deposits, raw materials; and shore geology.

Some maps cover the entire Southern Baltic Area: Gdańsk Basin and Gdańsk Bay, Bornholm Basin, Słupsk Furrow and surrounding banks, partly Gotland Basin (S part) and Arkona Basin (E part); other maps cover only Polish area of the Baltic Sea: area from Zatoka Pomorska in the west to Zatoka Gdańska in the east. It is difficult to judge, how much the authors are right by making such an option, but a curious reader is astonished, as he sees that the map is partly a white spot. We expect that these white spots will be filled up later.

The pre-Quaternary system is given in the sheets II-XI showing distribution of the crystalline basement, Paleozoic, Mesozoic and Tertiary, as well as thickness, structural surfaces and occurrence conditions presented. Lithology and petrography is shown in the generalized sections for separate tectonic blocks. A smarter reader will be able to calculate the thicknesses of stratigraphic units from the maps himself.

The Quaternary system is depicted in the sheets XII-XX, where a good deal of attention is paid to the evolution of Pleistocene glaciers, distribution and thickness of Quaternary and Pleistocene deposits, as well as structural conditions of the occurrence. The Tychnowsky Sea (Morze Tychnowskiego) of Eemian interglacial is shown as if it was on land (sic!) in the lower reaches of Vistula up to Grudziądz.

Bottom topography (sheets I and XXV) and bottom deposits (XXII-XXX) show in detail the bathimetry of the Sea bottom (by 5 m), morphogenesis, distribution of deposits at the bottom surface and at the depth of 1 m, Holocene thicknesses, Sea evolution during the Late Glacial and Holocene, recent sedimentation and geochemistry of surficial deposits as well.

The raw materials (sheets XXXI-XXXII) on the sea bottom are mainly sand and gravel strata occurring in shallow parts of the Baltic Sea: Southern Middle, Slupsk and Odrzana Banks, and Koszalin Bay.

The shore zone geology (sheets XXXIII-XXXIV) shows the distribution of pre-Quaternary and Quaternary at the joint of land and sea, occurrence conditions along the shore belt over 400 km long from Wolin Island to Vistula Spit. Most detailed picture is given for a phantastic Holocene geological structure of the Hel Peninsula.

Majority of sheets in the Atlas, together with geological generalizations, contain wide analytical information, research data, tables, diagrams etc. All this stimulates the interest of a reader to analyse, consider and evaluate.

The Geological Atlas of the Southern Baltic is expected to live for a long and useful age. I am sure that some new ideas in the marine geological mapping will receive continuation in other Baltic regions; hence, the authors and editors of the present Atlas will be able to accept it with a full satisfaction for their scientific and creative work and their love to the Baltic Sea.

Algimantas Grigelis

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Miidel, A. & Jantunen, T. 1992: Neotectonics and contemporary movements. *In* Raukas, A. & Hyvarinen, H. (Eds.): *Geology of the Gulf of Finland*, 87–98. Estonian Academy of Sciences, Tallinn.

Gudelis, V. 1973: Relief and Quaternary of the East Baltic region. 264 pp. *Mintis*, Vilnius.

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BALTICA is an International Yearbook on Geology, Geomorphology and Palaeogeography of the Baltic Sea.

ISSUED annually in cooperation with the geologists of the circum-Baltic States.

SEVEN volumes of the BALTICA Yearbook have already been published during 1963–1982. The renewed BALTICA Vol. 8 for 1994 is published in March, 1995.

EDITOR Algimantas Grigelis, Vilnius, Lithuania.

ISBN 9986-615-05-4 / ISSN 0067-3064

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