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SIMPLE SEDIMENT MODEL: A TOOL TO ESTIMATE NUTRIENT REGENERATION IN BALTIC SEA SEDIMENTS

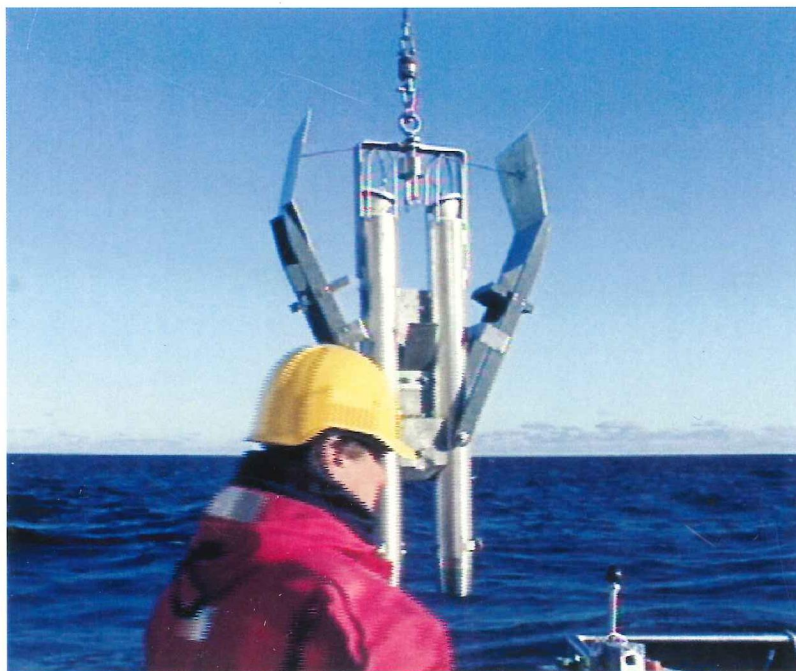
Oleg Andrejev, Oleg Savchuk & Matti Perttilä

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LONG-TERM CHANGES OF ZOOBENTHOS BIOMASS IN THE BARENTS SEA

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No. 45
2002

MERI

Report Series of the Finnish
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Cover photo: “Gemini” sediment corer ready to descend.

Publisher:
Finnish Institute of Marine Research
P.O. Box 33
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Tämän raporttisarjan numeroita voi tilata Merentutkimuslaitoksen kirjastosta.

ISSN 1238-5328

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ABSTRACT

A simple model is presented of coupled nitrogen and phosphorus cycles in Baltic Sea sediments. The state variables of the model are ammonium, nitrate, phosphate and oxygen. Using data on organic matter distribution in sediments in the Baltic Proper, the model simulates both the nutrient vertical profiles in the sediment and nutrient fluxes through the water-sediment interface. The results of numerical experiments to calibrate the model parameters are presented and discussed.

Key words: model, calibration, sediments, nutrients, modelling, Baltic Sea

1. INTRODUCTION

The Baltic Sea is a shallow brackish-water basin, permanently stratified and therefore subject to redox variations. The water exchange with the adjacent North Sea is restricted, and in consequence, sedimentation and sediment processes play an important role in the mass balance of nutrients within the Baltic Sea. Nutrient regeneration in the sediments constitutes an important feedback link in closing the nitrogen and phosphorus biogeochemical cycles. Although the nutrient release from oxic sediments corresponds to only a few percent of the phytoplankton requirements, the contribution may increase up to 7–13% of N demand and over 100% of P demand in anoxic situations or at specific locations (Koop & al. 1990, Conley & al. 1997, Stockenberg 1998). One of the reasons for the transition from a nitrogen-limited system in the Baltic Proper into a phosphorus-limited one in the Bothnian Bay (Graneli & al. 1990) may be the redox-related changes in the sediment biogeochemistry of nitrogen and phosphorus (e.g. Klump & Martens 1983, Caraco & al. 1990, Santschi & al. 1990, Sundby & al. 1992), which modify the N:P ratio of nutrient fluxes across the sediment-water interface. Knowledge of the interactions between pelagic and sediment sub-systems is also essential for nutrient budget calculations and turnover time estimates on basin-wide long-term scales, both in empirical and mechanistic ecosystem models (Wulff & al. 1990, Savchuk & Wulff 1996). A description of the sediments is therefore an inherent part of modern aquatic ecosystem models (Baretta & al. 1995, Ruardij & Van Raaphorst 1995, Tamsalu & Ennet 1995).

Sediment processes are usually quantified by means of two different approaches or their combination. The “direct” approach is based on measurements, during laboratory or *in situ* incubations, of core tubes or benthic chambers, sometimes artificially enriched with algal material (e.g., Balzer 1984, Conley & Johnstone 1995, Hall & al. 1996, Tuominen & al. 1999). The “inverse” approach involves either simple flux calculations based on observed concentration gradients at the sediment-water interface (e.g., Hall & al. 1996, Lehtoranta 1998) or more sophisticated fitting of the sediment models to observed vertical profiles (e.g. Van Raaphorst & al. 1990, Jensen & al. 1994).

In spite of the importance of sediment processes and the possibility to quantifying them, there are surprisingly few studies on sediment processes, particularly on the nutrient water-bottom exchange, in the Baltic Sea (Koop & al. 1990, Conley & al. 1997, Lehtoranta 1998, and references therein). Many uncertainties remain in current estimates, due to methodological and analytical difficulties encountered in the implementation of different schemes of measurements and calculations (Stockenberg 1998). Further, the irregular mosaic distribution of different sediment types and sparse sampling hamper accurate interpolations on a basin-wide scale, even from reliable site estimates. Finally, nutrient

concentrations in the porewater, needed in the “inverse” approach for flux calculations, have been sampled far more seldom than the nutrient content of the sediment solid phase.

The intention of the present study is to develop a simple model of coupled nitrogen and phosphorus sediment processes in order to provide a tool to estimate the main transport and transformation fluxes from given vertical distributions of organic matter.

2. METHOD AND DATA

2.1. Model formulation

The model state variables are ammonium A , nitrate N , phosphate P , and oxygen O . The dynamics of the vertical distribution of these substances is described by a system of general one-dimensional diagenetic equations under the following assumptions: a) an adsorption process is approximated with the Freundlich adsorption equilibrium, b) pore water advection is insignificant compared to the other processes concerned, c) the rates of deposition and compaction are constant, i.e. the porosity profile preserves its shape (Lerman 1977, Berner 1980):

$$(K_i + 1) \frac{\partial C_i}{\partial t} = \frac{1}{\phi} \frac{\partial}{\partial z} \left[\phi D_i \frac{\partial C_i}{\partial z} \right] - R_i + S_i \quad (1)$$

with the boundary conditions:

$$z = 0; \quad C_i = C_{0,i} \quad (2)$$

$$z \rightarrow \infty; \quad \frac{\partial C_i}{\partial z} = 0 \quad (3)$$

In Equations 2–3: $C_i(z)$ (mmol m^{-3}) is the concentration of the i -th substance ($i = 1 \dots 4$, $C_i = A, N, P, O$, respectively) dissolved in the pore water; the vertical distance co-ordinate z (metre) is taken as $z = 0$ at the sediment-water interface, increasing downward; t (day) is time; $\phi(z)$ is the vertical distribution of porosity (volume fraction of sediment occupied by water); D_i ($\text{m}^2 \text{d}^{-1}$) is an effective diffusion coefficient of the substance in pore water; R_i and S_i ($\text{mmol m}^{-3} \text{d}^{-1}$) are removal and supply rates, K_i is a dimensionless distribution coefficient defined as the ratio between the concentrations of absorbed and dissolved phases; $C_{0,i}$ is the given concentration at the sediment-water interface.

All removal processes are treated as first-order chemical reactions, while supply processes are supposed to have zero-order kinetics: fast adsorption is considered to be an instantaneous equilibrium reaction (Lerman 1977, Seitzinger 1990, Ruardij & Van Raaphorst 1995).

Assuming that ammonium A is produced by mineralization of organic nitrogen, its losses being caused by nitrification, which takes place only under oxic conditions, while in anoxic conditions a certain part is adsorbed onto sediment particles, Equation 1 is written as

$$(K_A + 1) \frac{\partial A}{\partial t} = \frac{1}{\phi} \frac{\partial}{\partial z} \left[\phi D_A \frac{\partial A}{\partial z} \right] - k_N A + M_A(z) \quad (4)$$

$$k_N = \frac{1}{1 + \left(\frac{z}{z_{ox}}\right)^{n_1}} \times k_{N0} \quad (5)$$

$$M_A(z) = r_N * N_{org}(z) \quad (6)$$

In Equations 4–6, D_A is the sediment diffusion coefficient for ammonium, k_N is the specific nitrification rate (day^{-1}) and k_{N0} is the value of this coefficient under oxic conditions. The formula to calculate k_N takes into account the fact that nitrification is only takes place under oxic conditions and conserves the form of Equation 4 both for oxic and anoxic conditions. In this formula z_{ox} is the oxygen penetration depth, calculated in accordance to condition $O \leq \varepsilon_1$, where O is oxygen concentration, ε_1 is a very small positive constant, $n_1 = 6$. M_A is the ammonium production rate ($\text{mmol N m}^{-3} \text{ day}^{-1}$) proportional to the organic nitrogen distribution N_{org} , r_N is the specific nitrogen mineralization rate (day^{-1}).

Nitrate is produced by nitrification of ammonium in the oxic layer and lost due to denitrification at low oxygen concentrations:

$$\frac{\partial N}{\partial t} = \frac{1}{\phi} \frac{\partial}{\partial z} \left[\phi D_N \frac{\partial N}{\partial z} \right] + k_N A - d_N N \quad (7)$$

$$d_N = \frac{1}{1 + \left(\frac{z_{den}}{z} \right)^{n_2}} \times d_{N0} \quad (8)$$

where D_N is the sediment diffusion coefficient for nitrate, d_N is the specific denitrification rate (day^{-1}) and d_{N0} is its value in the “denitrification zone”, that is below a depth z_{den} calculated according to the condition $O \leq \varepsilon_2$, where ε_2 is a given oxygen “threshold” concentration, below which denitrification starts, $n_2 = 6$.

Phosphate is produced by mineralization of organic phosphorus. The ratio between the sorbed and dissolved phases is considerably smaller in the oxic layer than in the anoxic layer.

$$(K_P + 1) \frac{\partial P}{\partial t} = \frac{1}{\phi} \frac{\partial}{\partial z} \left[\phi D_P \frac{\partial P}{\partial z} \right] + M_P \quad (9)$$

$$M_P(z) = r_P * P_{org}(z) \quad (10)$$

where D_P is the sediment diffusion coefficient for phosphate, M_P is the phosphate production rate ($\text{mmol N m}^{-3} \text{ day}^{-1}$) proportional to the organic phosphorus content P_{org} , r_P is the specific mineralization rate (day^{-1}).

As seen from Equations 5 and 8, the oxygen concentration is an important governing variable and coupling parameter in the model. Oxygen is consumed during the biogeochemical mineralization of organic nitrogen (ammonification) and nitrification. As mineralization partly takes place with nitrate as an electron acceptor instead of oxygen, the correspondent portion of oxygen consumption is “reimbursed” proportionally to denitrification (Stigebrandt & Wulff 1987, Savchuk & Wulff 1996). Hydrogen sulphide is considered as “negative” oxygen (Fonselius 1969).

$$\frac{\partial O}{\partial t} = \frac{1}{\phi} \frac{\partial}{\partial z} \left[\phi D_O \frac{\partial O}{\partial z} \right] - R - R_n + R_d \quad (11)$$

$$R = a_m M_A(z) \quad (12)$$

$$R_n = a_n k_n A(z, t) \quad (13)$$

$$R_d = a_d d_N N(z, t) \quad (14)$$

Here D_{ox} is the sediment diffusion coefficient for oxygen, a_m , a_n , and a_d are the oxygen consumption equivalents ($\text{mmol O}_2 \text{ mmol}^{-1} \text{ N}$) of mineralization, nitrification, and denitrification, respectively.

To increase the time step an implicit numerical scheme is used to solve the system of Equations (4, 7, 9, 11) with boundary conditions (2, 3). The nutrient diffusive flows are calculated as:

$$J = -\phi D_i \frac{\partial C_i}{\partial z} \quad (15)$$

2.2. Set-up of numerical experiments

The main aim of the simulations presented and discussed in this paper was to calibrate the model parameters against vertical distributions of state variables known from field measurements. For this purpose we used data on both solid phase contents and pore water concentrations for several stations in the Baltic Proper (Fig. 1, Table 1) sampled in January 1990 (Carman & Rahm 1997). All recalculations necessary to convert published values into model units were made with additional data on the “water content” and the “loss on ignition” vertical distributions kindly supplied by R. Carman (pers. com.).

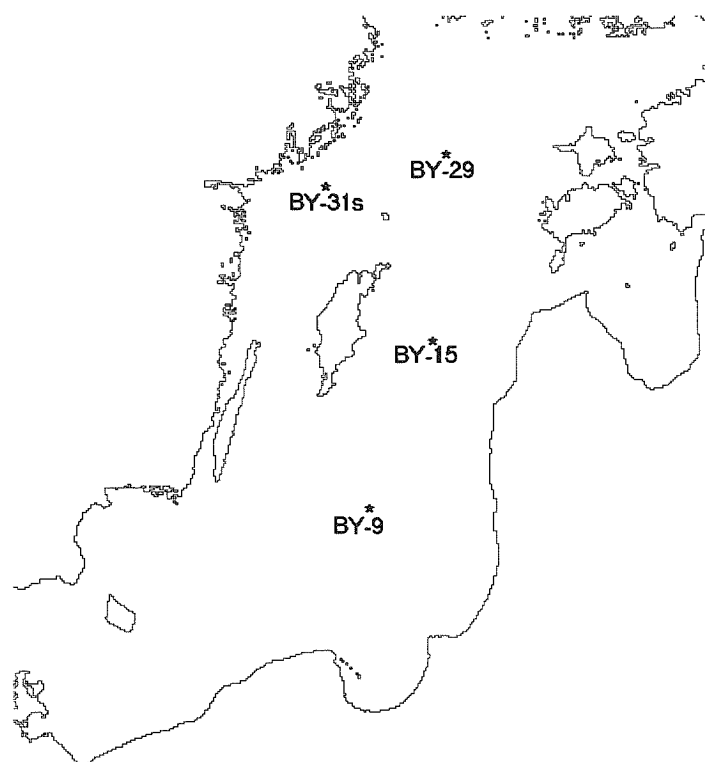


Fig. 1. Location of sediment sampling sites (after Carman & Rahm 1997).

Since these samples were taken during the long stagnation period that started in the Baltic Sea in the late 70's, the bottom water at the two deepest stations BY 15 and BY 29 was anoxic, while hydrogen sulphide was already present in the uppermost centimetre of sediments at site BY 9 with oxic bottom water. At BY 31s hydrogen sulphide was absent only from the upper three cm of sediments. This site was characterized by coarser grain size and lower porosity compared to other sites. However, no signs of macrofauna were found either at this site or at the others.

Table 1. Near-bottom (nbw1) and pore water (pw2) concentrations of nutrients, oxygen and hydrogen sulphide (μM) compiled from Tables 1 and 2 in Carman & Rahm (1997).

Station	Depth (m)	$\text{NH}_4\text{-N}$		$\text{NO}_3\text{-N}^3$		$\text{PO}_4\text{-P}$		$\text{O}_2/\text{H}_2\text{S}^4$	
		nbw	pw	nbw	pw	nbw	pw	nbw	pw
BY 9	91	0.5	67.7	10.95	-	2.34	12.1	109.3/-	14.0
BY 15	248	14.7	424.5	-	-	8.34	18.4	-/88.8	223.3
BY 29	181	0.31	22.5	7.87	-	3.33	4.4	-/6.0	9.0
BY 31s	125	0.2	30.0	9.33	-	2.78	1.2	81.2/-	0.0

¹Nutrient samples were collected about 3-5 m above the bottom, oxygen/hydrogen sulphide were sampled about 0.5 above the sediment surface.

²Pore water concentrations from the topmost cm of sediments.

³No data on nitrate concentration in the pore water were published.

⁴For pore waters, only hydrogen sulphide concentrations were published.

According to the model assumptions, $N_{org}(z)$ and $P_{org}(z)$ in Equations (6) and (10) represent the “reactive” fractions of organic nitrogen and phosphorus. To estimate their vertical distribution from the data on total amounts of organic nitrogen and phosphorus presented by Carman & Rahm (1997), we used a multi-G approach (Berner 1980, Westrich & Berner, 1984) as modified by Carignan & Lean (1991) to account for variable porosity and compaction within the upper layers of sediments:

$$G(m) = G_0 + \sum_1^n G_j e^{-r_j m / R} \quad (16)$$

In Equation 16 $G(m)$ is the concentration of organic nitrogen or phosphorus ($\text{mmol N (P) kg}^{-1}$) as a function of the cumulative mass of dry sediment m (kg m^{-2}), G_0 is the concentration of stable fraction, G_j are the concentrations of n fractions with different specific mineralization rates r_j (yr^{-1}), R is the sedimentation rate ($\text{kg m}^{-2} \text{ yr}^{-1}$). Thus, implementation of Equation 16 to the measured distributions of organic nutrients allows simultaneous evaluation of both vertical profiles for fractions with different reactivity and corresponding specific rates of their degradation.

The initial set of other constants was estimated from the available literature. The diffusion coefficients of the nutrients were estimated from the data in Li & Gregori (1974) with the following generalising assumptions: a) the values at 0°C are half those at 25°C (Lerman 1977), b) the values for ammonium and nitrate are approximately equal, c) the value for phosphate is about a third of those for ammonium and nitrate. The formula to calculate the diffusion coefficient of oxygen as a function of temperature was taken from Ruardij & Van Raaphorst (1995). Under low ambient temperature conditions the diffusion coefficients of ammonium, nitrate, and oxygen were taken as equal to $1 \times 10^{-4} \text{ m}^2 \text{ day}^{-1}$, while the diffusion coefficient of phosphate was taken as $0.33 \times 10^{-4} \text{ m}^2 \text{ day}^{-1}$.

The adsorption coefficient of ammonium K_A was taken as 1 in anoxic conditions and set to zero in oxic conditions (Mackin & Aller 1984). Taking into account the proportionality between adsorbed phosphate and porosity (Ruardij & Van Raaphorst 1995) as well as the rather high porosity of the prototype sediments (0.80–0.96), the adsorption coefficient of phosphate was taken as 500 for oxic conditions and as 2 for anoxic conditions (Krom & Berner 1980).

The specific nitrogen and phosphorus mineralization rates were estimated from the field data as described in the next Section. The specific nitrification and denitrification rates discussed in the literature (Raaphorst & al. 1990, Blackburn & Blackburn 1993, Ruardij & Van Raaphorst 1995, and references therein) vary within several orders of magnitude. In the present study their values were derived from calibration experiments: $k_{\text{NO}} = 0.1 \text{ day}^{-1}$, $d_{\text{NO}} = 10.0 \text{ day}^{-1}$. The oxygen “threshold” for denitrification was set to $\varepsilon_2 = 10 \text{ mmol O}_2 \text{ m}^{-3}$. The oxygen consumption equivalents were taken (according to Gundersen & Mountain 1973, Klump & Matrens 1983) as: $a_m = 6.6$, $a_n = 2.0$, $a_d = 1.3 \text{ mmol O}_2 \text{ mmol}^{-1} \text{ N}$.

The numerical experiments were performed for the upper 10 cm of sediments with a 1-mm space step and a 1-day time step.

3. RESULTS

3.1. Labile fractions and mineralization rates

Equation 16 was fitted by a non-linear curve fitter (Microcal™ Origin™, v.5.0) to the observed organic nitrogen and phosphorus concentrations both at each station individually and pooled together (Fig. 2). The best fits to both individual and combined distributions were obtained with a simple two-component model ($n = 1$), implying the existence of stable and one reactive fractions. The assumption on several reactive fractions ($n = 2, 3$) resulted in a splitting of a curve corresponding to $n = 1$ into several curves with nearly equal parameters ($0.5 G_{1,n=1} \approx G_{1,n=2} \approx G_{2,n=2}, k_0 \approx k_1 \approx k_2$).

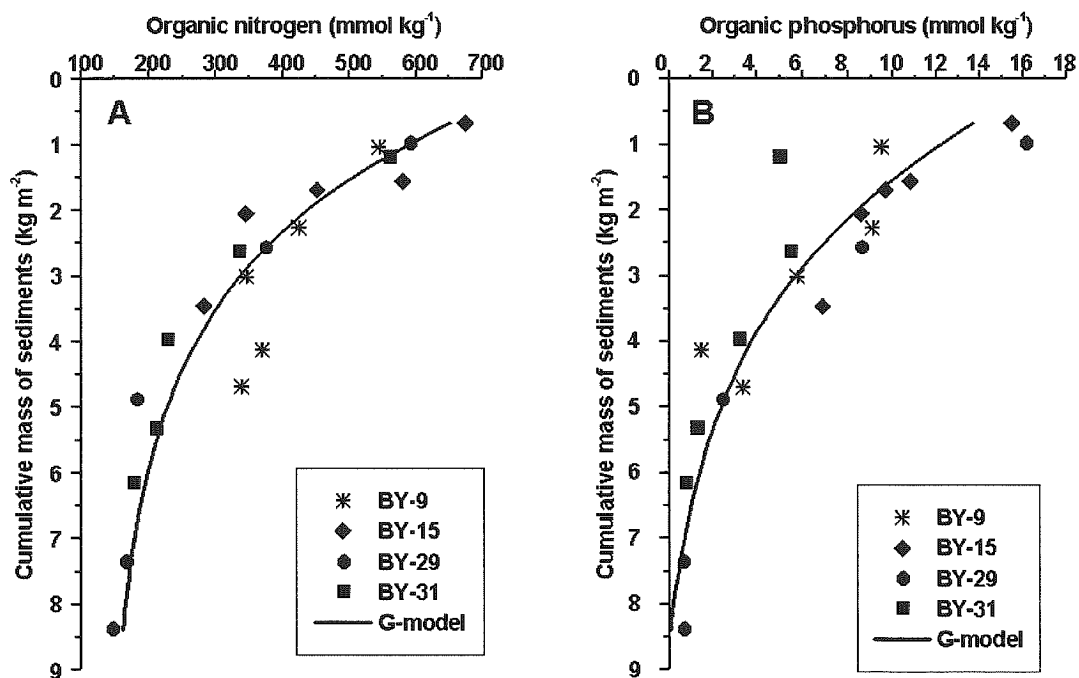


Fig. 2. Organic nitrogen (A) and phosphorus (B) concentrations vs. cumulative sediment dry mass distribution. Curves show a fit of the mineralization model (Eq. 16) to combined data pools.

As seen from Equation 16, the mineralization rate derived with this procedure depends on the sediment accumulation rate. For the combined curves shown in Figure 2 the reported range of accumulation rates of 0.5–4.0 mm yr⁻¹ (Jonsson & al. 1990, Carman & Rahm 1997) yields a range 0.0001–0.001 d⁻¹ for nitrogen mineralization and a slightly narrower range 0.0001–0.0007 d⁻¹ for phosphorus mineralization. In the absence of the site-related information on accumulation rates, the specific mineralization rates $r_N = 0.0005$ d⁻¹, $r_P = 0.0004$ d⁻¹ were taken as a first approximation.

However, the obtained estimates of the G-fractions for nitrogen mineralization differ qualitatively from those for phosphorus. The stable and reactive organic nitrogen concentrations are quite comparable (Table 2), while the “stable” organic phosphorus values are generally close to zero.

Table 2. Labile and stable organic nitrogen concentrations (mmol m^{-3}) estimated with the G-model (Eq. 16).

<i>Parameter</i>	BY-9	By-15	BY-29	BY-31s	Combined
G_o	336	108	129	165	143
G_l	515	797	728	820	677

Correspondingly, the vertical profiles of $N_{org}(z)$ at each site were calculated by subtraction of a “stable” fraction from the measured distributions, while the vertical profiles of $P_{org}(z)$ were taken as being the same as the measured ones.

3.2. Model calibration

With the site-related pore water nutrient and oxygen concentrations derived from the data in Carman & Rahm (1997), the vertical distributions of the labile organic nutrients as prescribed from the evaluation above and all the model constants assumed to be invariable between sites, the model was run for one time-step at each location. In other words, the model was used as an “intelligent” calculator of the transport flows and the biogeochemical fluxes both corresponding to given distributions of concentrations and mutually interrelated according to the model formulations, rather than as a dynamic simulation tool.

The resulting profiles are shown in Figure 3 together with the measured concentrations.

The mass balance analysis of water-bottom diffusive flows and biogeochemical fluxes integrated over the whole modelling domain, i.e. over the upper 10 cm of sediment, shows significant misbalances that would affect inorganic nutrient distributions (Table 3). For example, at sites BY 9 and BY 15 the negative ammonium balances compared to the ammonification and ammonium release from the sediments imply ongoing depletion of inorganic nitrogen in these sediments, while at sites BY 29 and BY 31 ammonium accumulation can be expected.

Because no information is available about either the existence or the absence of long-term trends in pore water nutrient concentrations, the next set of calibration experiments was run assuming that these deep-water sediments were in a quasi-steady-state. In these experiments the sets of constants were tuned by a trial-and-error method under the following regulations: a) to yield an approximate integral conservation of nutrients, b) to keep a reasonable fit to measured vertical distributions, c) to change parameters describing similar processes (such as diffusion and mineralization) in mutual concordance. The last requirement allows one to exploit the advantage of coupled formulations, where state variables are interrelated both through oxygen/hydrogen sulphide distributions and, in the nitrogen case, by nitrification. Qualitative “geographical” support for the parameter tuning is discussed below. As a result, the quasi-steady-state site-related values in Table 3 were obtained, with some constants changed compared to their values in the “standard set” described above.

Parameter tuning at site BY 9 indicates high values for the mineralization rates. Assuming that at the shallowest site BY 9 organic matter is less “diluted” with old re-deposited material and, consequently, is more labile, the specific mineralization rates for this site were increased roughly three-fold: $r_N = 0.0015 \text{ d}^{-1}$, $r_P = 0.001 \text{ d}^{-1}$, keeping the diffusion constants unchanged.

The site BY31 is typically suboxic. The redoxcline, if it occurs, usually lies very deep, implying better ventilation than at the anoxic sites BY15 and BY29. Accordingly, the diffusion coefficients here were increased, without changing the mineralization rates: $1.5 \cdot 10^{-4} \text{ m}^2 \text{ day}^{-1}$ for inorganic nitrogen species and oxygen, $0.5 \cdot 10^{-4} \text{ m}^2 \text{ day}^{-1}$ for phosphate.

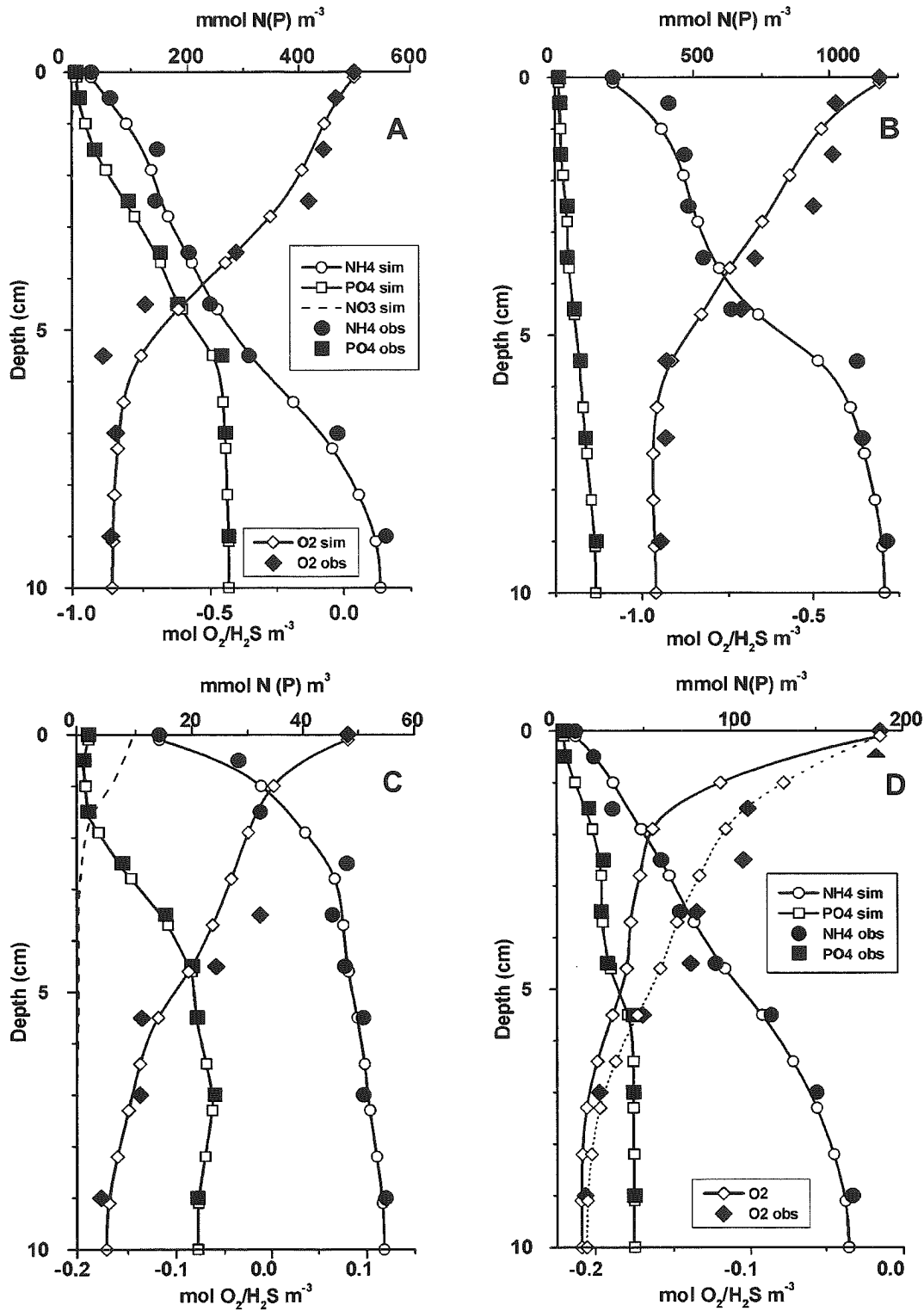


Fig. 3. Vertical distributions of nutrients and oxygen measured (symbols) and calculated (curves) with the invariable parameter set at sites BY 9 (A), BY 15 (B), BY 31s (C), and BY 29 (D). The dotted curve shows the simulation with decreased mineralization at BY 29. Hydrogen sulphide is considered as "negative" oxygen.

Table 3. Transport flows, integral transformation fluxes, and resulting mass balances calculated with invariable parameters (inp) and assuming quasi-steady state (qss) distributions. Amm – ammonification, Nit – nitrification, Den – denitrification, Pmin – phosphorus mineralization. Negative/positive flows correspond to output/input from/into the sediments.

	BY 9		BY 15		BY 29		By31	
	inp	qss	inp	qss	inp	qss	inp	qss
<i>Diffusion flows across water-sediment interface (mmol m⁻² d⁻¹)</i>								
NH ₄	-0.65	-0.77	-2.31	-1.39	-0.30	-0.22	-0.27	-0.34
NO ₃	0.24	0.24	-	-	-	-	0.03	0.05
PO ₄	-0.040	-0.040	-0.027	-0.015	-0.011	-0.010	0.005	0.008
O ₂ /H ₂ S	1.43/	2.41/	/-1.17	/-0.75	/-0.76	/-0.41	1.27/	1.55/
<i>Biogeochemical fluxes within sediments (mmol m⁻² d⁻¹)</i>								
Amm	0.21	0.62	1.05	1.05	0.62	0.25	0.51	0.51
Nit	0.02	0.02	-	-	-	-	0.16	0.16
Den	0.23	0.23	-	-	-	-	0.19	0.20
Pmin	0.022	0.054	0.027	0.027	0.017	0.009	0.012	0.012
<i>Integral balances within sediments (mmol m⁻² d⁻¹)</i>								
NH ₄	-0.47	-0.17	-1.26	-0.34	0.33	0.03	0.07	0.01
NO ₃	0.03	0.03	-	-	-	-	0.01	0.01
PO ₄	-0.019	0.014	0.000	0.012	0.006	-0.002	0.017	0.020

At the anoxic site BY 29 in the invariable-parameter case, the calculated ammonium accumulation was nearly equal to the ammonium release from sediments (Table 3), while the simulated H₂S concentration was much higher than that measured (compare the solid and dotted curves in Figure 3). As revealed by simulations, intensifying the diffusion process turned out to be less effective than a decrease in hydrogen sulphide production. It may also be less “geographically” justified, since there is no information about possibly increased diffusivity at this site. Correspondingly, the decision was made not to alter the diffusion coefficients and to reduce the mineralization rates to $r_N = 0.0002 \text{ d}^{-1}$, $r_P = 0.0002 \text{ day}^{-1}$.

In contrast to site By 29, the simulated ammonium release at the anoxic site BY 15 was more than twice the integral ammonification flux. However, an increase in mineralization would lead to excessive hydrogen sulphide production in the model, resulting in a deviation of the simulated vertical profile from that measured towards that shown for site BY 29 (Fig. 3). Therefore, the approximation of steady state for this deepest site was achieved by a two-fold reduction in the diffusion coefficients: $0.5 \cdot 10^{-4} \text{ m}^2 \text{ day}^{-1}$ for inorganic nitrogen species and oxygen, $0.17 \cdot 10^{-4} \text{ m}^2 \text{ day}^{-1}$ for phosphate.

Similar balance considerations were also involved in the tuning of the specific nitrification and denitrification rates using the data from BY 9 and BY 31s. An over-/underestimation of the nitrification flux would lead to ammonium depletion/accumulation, while a miscalculated denitrification flux causes an unbalanced nitrate distribution. The values finally found provide for an approximate balance at both sites, although the ammonium concentrations at BY 9 (redoxcline at 3mm) are about 5–10 times higher than those at BY 31s (redoxcline at 30 mm).

In accordance with the changes of parameters, the changes of transport flows and transformation fluxes were also comparatively small – within a factor of three (Table 3).

3.3. Model implementation

The approach described above might be used when the data on both organic and inorganic dissolved nutrient distributions are available. However, much more “historical” information on only organic fractions is stored in different laboratories around the Baltic Sea. One possibility for using such data for the evaluation of biogeochemical fluxes is explored below.

First step is to reconstruct the vertical profiles of the labile fractions and to estimate the mineralization constants with Equation 16 similarly to the procedure described in the previous Section. Correspondingly, in further demonstration numerical experiments the initial distributions of organic nitrogen and phosphorus were taken to be the same as in Figure 2; all the constants were given as in the "standard set". The initial distributions of nutrients dissolved in pore waters were set to zero. Thus, "oxic" and "anoxic" cases differ only in the boundary conditions at the surface (Table 4).

Table 4. Nutrient and oxygen concentrations (mmol m^{-3}) at the sediment surface given as the boundary conditions for "oxic" and "anoxic" cases.

Experiment	NH_4	NO_3	PO_4	$\text{O}_2/\text{H}_2\text{S}$
Oxic	1	10	1	200
Anoxic	10	-	10	-100

With this set-up the model was run until the steady state was achieved regarding both vertical distributions of variables and transport flows as well as integral biogeochemical fluxes. For the nitrogen variables it was necessary to let the model run to simulate about one year, while for phosphate, with lower values of both diffusion coefficient and of mineralization flux about three years was required. Since in the oxic case the redoxcline was situated at a depth of 4 mm (Fig. 4), the resulting profiles in the anoxic case differ from those for the oxic case only in the difference in boundary conditions.

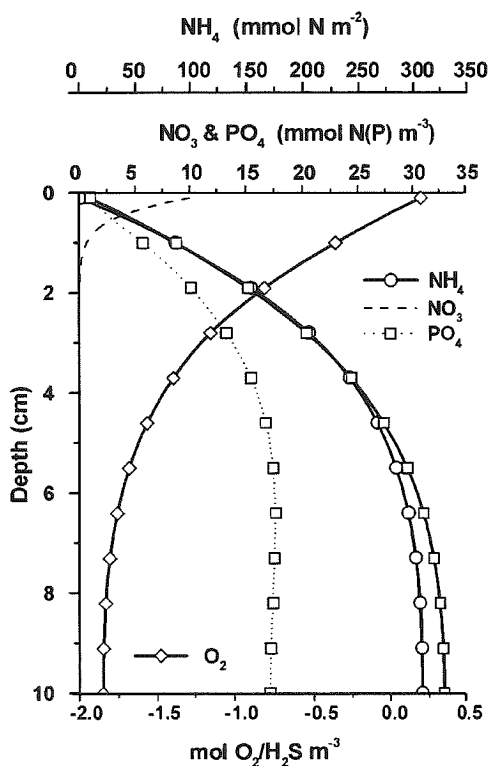


Fig. 4. Vertical distributions of nutrients and oxygen under the oxic scenario. Solid curves correspond to the steady-state solution on the 300th day for ammonium, nitrate, and oxygen, and on the 1000th day for phosphate. The dotted curve shows the phosphate distribution on the 300th day. Correspondingly, the differences in flows and fluxes are also small (Table 5).

Table 5. Bottom-water exchange ($\text{mmol m}^{-2} \text{ day}^{-1}$) and integral biogeochemical fluxes ($\text{mmol m}^{-3} \text{ day}^{-1}$) in the upper 10 cm of sediments.

	Oxic case	Anoxic case
Ammonium release	1.01	1.02
Nitrate intake	0.21	-
Phosphate release	0.03	0.03
O ₂ intake/H ₂ S release	6.47	-3.37
<i>Ammonification</i>	1.02	1.02
Nitrification	0.01	-
Denitrification	0.22	-
Phosphorus mineralization	0.03	0.03

4. DISCUSSION

With certain restrictions, the model appears to give reasonable estimates of the sediment-water interface processes.

The validity of the “standard” parameter set found in this study is limited by two conditions, at least. All data sets are from rather deep accumulation bottoms with either quasi-permanent anoxia or suboxia, and thus without macrofauna and, hence, supposedly with relatively low diffusivity. All samples were taken during a stagnation phase, hence the rather shallow redoxcline within the sediments, high subsurface ammonium concentrations, and low phosphate concentrations being depleted during anoxic conditions.

The first step, i.e. implementation of the G-approach, might be suitable for the estimation of the mineralization rates. Westrich & Berner (1984) found three fractions (one stable and two labile) but were dealing with freshly-sedimented organic matter, which might explain the difference from the results presented here (one stable and one labile), derived from real sediments with thickness of 5 cm, i.e. accumulated over several decades. Similar results with a single-G decomposition model were obtained, e.g. by Carignan & Lean (1991) with a specific rate for nitrogen of 0.0052 d^{-1} .

The possible ways of implementation of the model depend on data availability:

- a) Only organic nutrients available: run to steady-state as in the last set of experiments
- b) Both organic nutrients in the solid phase and inorganic in pore water available: run as a calculator assuming a steady state
- c) Actual flux information available: use as additional constraints during calibration.

5. CONCLUSION

A simple mathematical model of nutrient transformation in sediments has been developed to serve as a tool to utilize a large amount of sediment data. When calibrated against measurements, the model proved to be reasonable enough to be used for the evaluation of nutrient fluxes both across the water-bottom interface and within sediments. Further calibration for a wider spectrum of ambient conditions is needed. Inclusion of silica into the model is under consideration.

Acknowledgements

This work is a part of the EU-funded research project BASYS (Baltic System Study) (Contract MAS3-CT96-0058). In addition, financial support from the Nordic Council of Ministers is gratefully acknowledged.

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DISTRIBUTION OF SEDIMENTS AND SUSPENDED MATTER IN THE GULF OF FINLAND – THE CENTRAL BALTIC PROPER TRANSECT WITH A PARTICULAR EMPHASIS OF TRANSPORTATION: MINEROGENIC EVIDENCE

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ABSTRACT

To elucidate sediment particle behaviour in the Baltic Sea sediments and suspended matter, several sediment and suspended sediment samples were collected in the Gulf of Finland and Baltic Proper in 1997 and 1998. Results for two flux models (total and mineralogical), based on hypothetical approach, and *in situ*-model are presented. According to total flux model, less than 2 % of the suspended material originated in the central Gulf of Finland reaches the Gotland deep area. This is 7 % of the total load from that direction. Approximately 0.5 % of the Gotland deep total accumulation is of central Gulf of Finland origin. The mineralogical model suggests 1 %, 8 % and 0.3 % values. The potential mobile fraction is 23 % (total) and 18 % (mineralogical) indicating that almost all particulate matter accumulates permanently in the adjacent bottom deposits. In certain areas there is a clear correlation between aluminosilicate and carbonate coexistence in the water column. Amorphous ferromanganese particles occur frequently in the water column while they are almost absent in the adjacent sediments. There are both temporal and non-temporal fluxes present in the studied areas. The textural maturity of sediments is poor. All these observations strongly suggest that *in situ*-type fluxes prevail in the northern Baltic Sea environment. The above-mentioned numerical values are probably overestimates of the real values.

1. INTRODUCTION

The Baltic Sea area is surrounded by distinct different bedrock features. A precambrian crystalline basement prevails almost all over Finnish and Swedish coastal areas while younger Phanerozoic formations mostly occur in eastern and southern parts of the Baltic Proper. The contact between Precambrian and Postcambrian formations mostly lies in the line of the Gulf of Finland, the Northern Baltic Proper and the Landsort Deep in the bottom areas of the Baltic Proper. These rocks and sediments are mainly covered by younger Quaternary deposits which are weathering products of older material. Especially areas deeper than 50-70 m are continuously active sedimentation basins (Niemistö & al. 1978, Winterhalter 1980).

The Gulf of Finland is a direct continuation of the Baltic Proper area; there are no topographic barrier which prevents water interaction between these areas. Easternmost areas of the Gulf of Finland are polluted by heavy metals (Leivuori 1998) which have a tendency to accumulate in the deepest part of sedimentation basins where tranquil conditions prevail (Emelyanov 1986). These metals are accumulated in the basin areas either by redox-changes of supernatant water or transportation of suspended material from less stable areas.

As suspended matter, metals can be transported into deeper areas either by clay minerals or organic particles. Clay minerals are the chemical products of older materials which have been weathered over a time interval of some hundreds of millions of years. Despite of different bedrock features in the Baltic Proper area, many recent sediments have been mixed by the last glaciations which obscure the interpretation of origin of clayey material. Crystalline bedrock influences the recent deposits above Phanerozoic formations (Klagish & Goldfarb 1994; Kadastik 1994, 1995; Gaigalas 1996). The uniformity of clay mineral distribution has been established in the bottom sediments of the Baltic Sea

area (Boström & al. 1978, Emelyanov 1992, Gingele & Leipe 1997). The suspended sediments have been, however, analyzed rather geochemically than mineralogically (e.g. Bernard & al. 1989, Boström & al. 1981, Ingri & al. 1991, Brüggmann & al. 1992, Leivuori & Vallius 1998).

Several worldwide resuspension modelling attempts based on physical, biogeochemical as well as laboratory experiments have been made (see Hawley 1982, Amos & Mosher 1985, Gibbs 1985, Eisma 1986, McLaren & Little 1987, Vale & Sundby 1987, Uncles & al. 1988, Bevers & Yeats 1989, Grabemann & Krause 1989, Madsen & al. 1993, Dupont & al. 1993, Statham & al. 1993, Lindsay & al. 1996, Betty Ng & al. 1996). There are also some previous attempts to model resuspension in coastal areas of the Baltic Sea area (Virtanen & al. 1988, Blomqvist & Larsson 1994, Heiskanen & Leppänen 1995, Juntura & al. 1996). The aim of this study is to clarify material behaviour in the Gulf of Finland – the central Baltic Proper transect using mineralogical and supported geochemical tracers. Due to the main method used (X-ray diffraction, XRD) and long distances inside the study area, the approach of this study is hypothetical. This study is part of the international BASYS (Baltic System Study) project which is part of the EU/MAST research programme with special emphasis on basin-basin fluxes (Subgroup 3B).

2. MATERIALS AND METHODS

Sampling – All samples were collected from the Gulf of Finland (stations GF2, JML), the northern (stations LL17, NCB) and the central (Fårö Deep, BY15) Baltic Proper (Appendix, Fig. 1). Bottom sediments were collected using a Gemini corer (the second generation type of Niemistö corer, Niemistö 1974). Sediment traps were moored at 5 m and 20–40 m above the bottom level (Appendix, Fig. 2). The upper traps were French Technicap models P.P.S. 5/2 with a collection area of 1.0 m² and P.P.S. 3/3 with a collection area of 0.125 m², respectively. The former is equipped with a rotating unit for 24 and the latter for 12 collection bottles. The lower trap (Eila-trap) consists of three separate, non-rotatable cylindrical units with a 0.008 m² collecting area for each tube. There was no clear indications of mineral alterations by any chemical pretreatments (e.g. preservation liquid, formaldehyde 2 v/v).

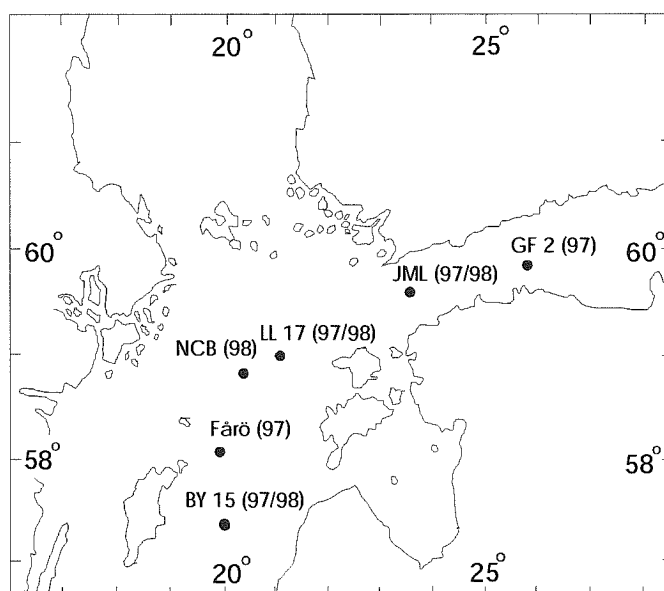


Fig. 1. Sampling sites. (97), (98) denote sampling year at given station.

Mineralogy – Part of all raw samples was dried and ground on an agate mortar for visual description and immersion studies. The colour scale is based on the Munsell chart HUE 10 YR values. The immersion liquid was a diluted CH₃I solution (refraction index ≈ 1.55). A semi-quantitative morphology and mineral content determination was recorded using a polarizing microscope (see

Appendix). The grain size analyses was made by a pipette method. The porosity of ferromanganese rich samples was estimated by comparing the water volume with and without samples.

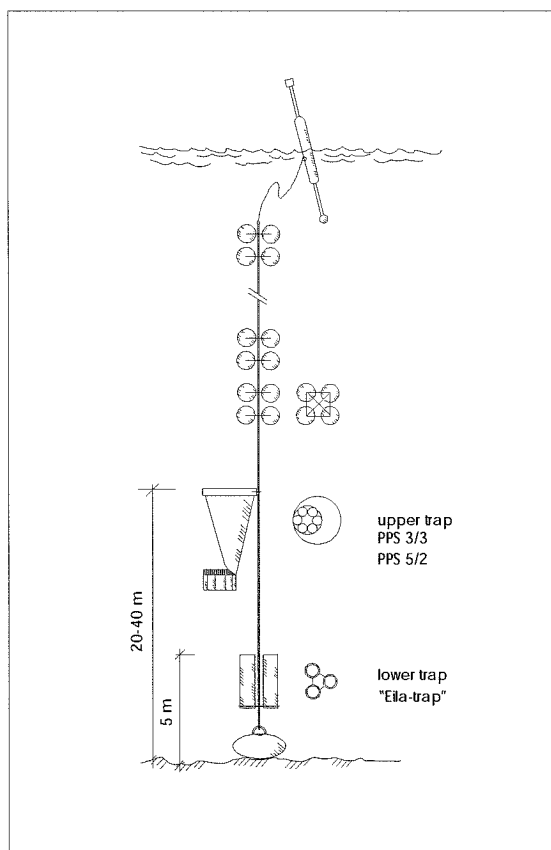


Fig. 2. The arrangement of sediment trap system.

All wet samples were subjected to ca. 15 % hydrogen peroxide (H_2O_2) treatment (15 min boiling) to remove reactive organic matter before XRD studies. Oriented specimens were prepared using the glass slide method (Moore and Reynolds 1989). The procedure was repeated twice for every sample. The diffractometer was a Philips PW3710 with a graphite curved monochromator and Cu tube as an X-ray source. For total fraction, $5-55^\circ$ (2θ) were scanned at a step size of 0.05 (2θ) and 1.0 sec. per step using a voltage of 50 kV and a current of 30 mA. The reproducibility for peaks was better than 15 %. Only $5-35^\circ$ (2θ) are presented. The values of crystalline mineral particles (CMP; 10 % – 20 % – 30 % – 40 % – 50 %; [1]-[5], see Fig. 3 a-f) were estimated using raw data from Table 1 and Fig. 4 and are assumed to represent their precise values. All data were weighed by collection day numbers. Clay mineralogy is described elsewhere (Mälkki 2001).

Chemistry – A total aluminium content (Fig. 4) is based on HF – $HClO_4$ extraction, see Vallius & Leivuori 1999. Some partially leached, microwave digested (60 % HNO_3) data from the 1997 samples are presented. The recovery for Fe and Mn was ≈ 100 % and for Al 60 % relative to total dissolution. Using the crustal “background” values presented by Berry & al. (1983) the HNO_3 leachable background ratios (Fe/Al = 1; Mn/Al = 0.017) were estimated. Also some individual leaching tests were made for samples rich in ferromanganese concretions (1 M hydroxylamine hydrochloride – 1 M HCl) and for carbonate rich samples (1 M HCl), see the Appendix.

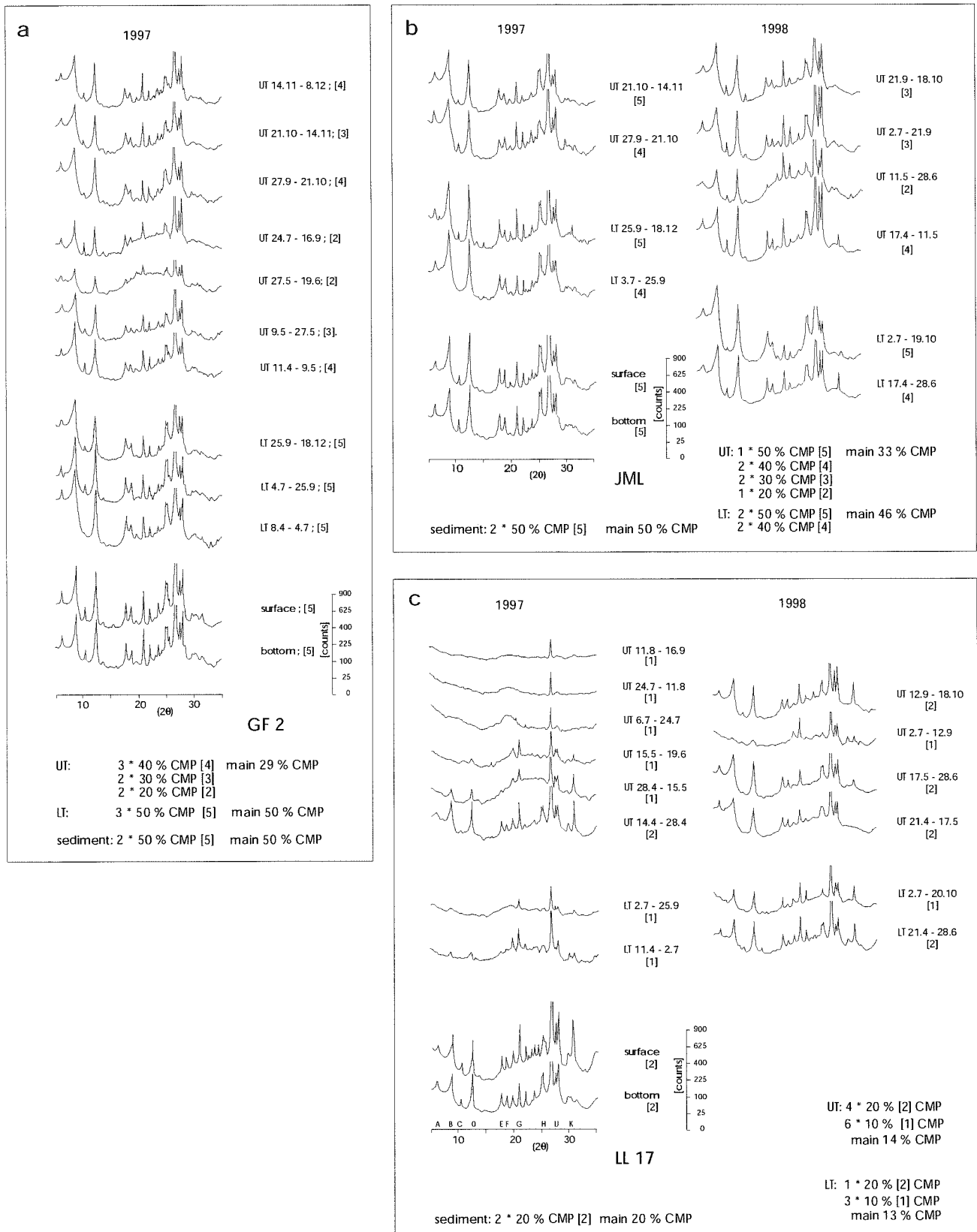


Fig. 3. XRD patterns 5-35 °(2θ). GF2 (a), JML (b), LL17 (c), NCB (d), Fårö Deep (e) and BY15 (f). Main peak description: A = Chlorite (14 Å); B = illite (10 Å); C = amphibole (8.4 Å); D = chlorite/kaolinite (7 Å); E = illite (5 Å); F = chlorite (4.7 Å); G = quartz 4.24 Å; H = kaolinite/chlorite doublet (3.58/3.54 Å); I = (K)feldspar (3.25 Å); J = (Na,Ca) feldspar (3.19 Å); K = carbonate (2.89 Å); L = unknown (2.71 Å), only BY 15. All diffractograms are drawn on the same scale. The vertical bar was drawn on the basis of the lowest scan. Note that carbonate XRD peak can be totally absent despite a high presence of carbonate crystals (Appendix).

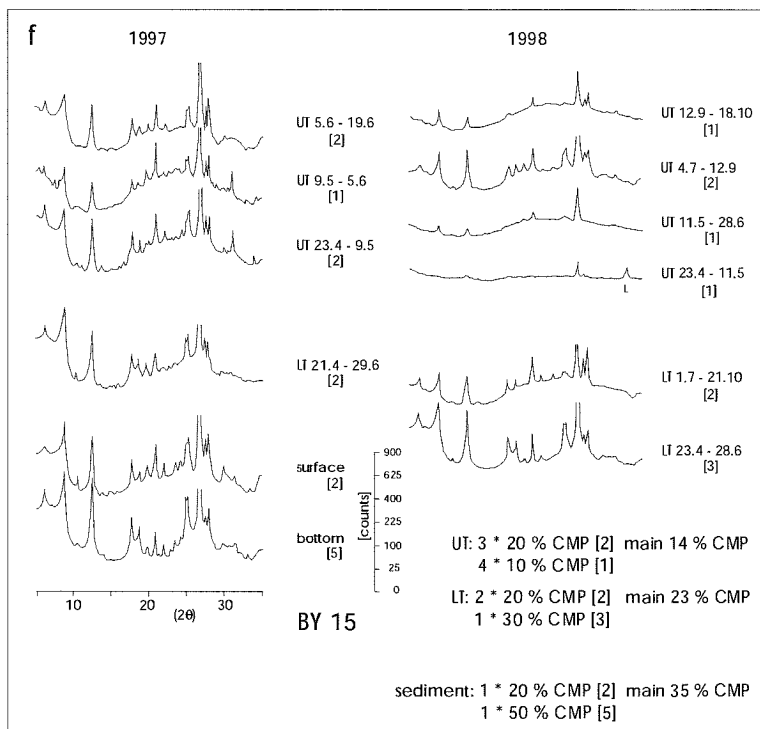
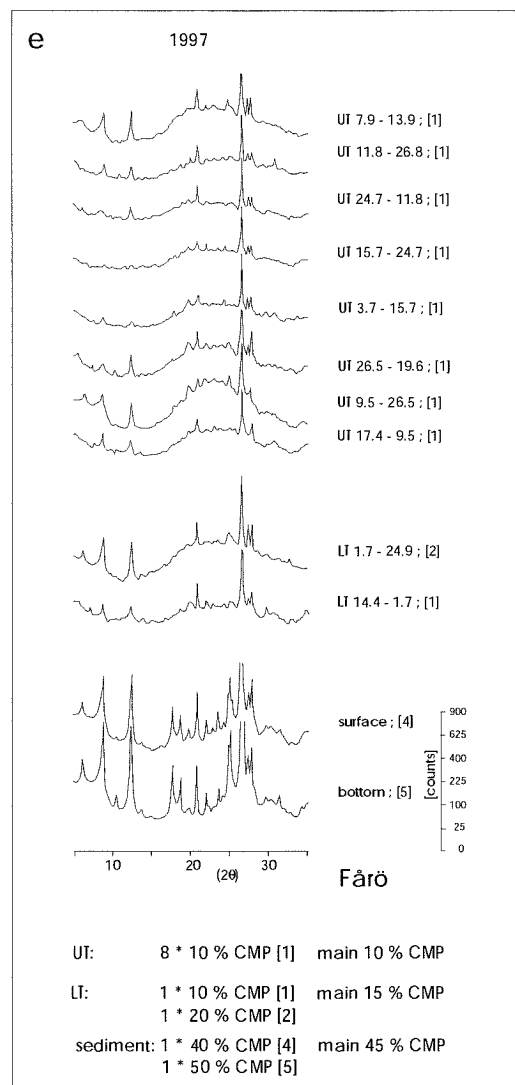
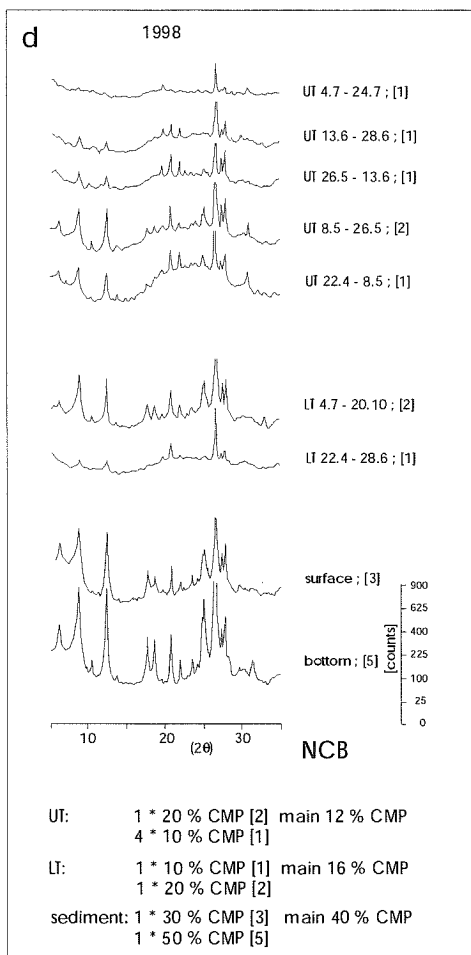


Table 1. Collection periods and accumulation rates ($\text{mg m}^{-2} \text{d}^{-1}$) in 1997 (a) and 1998 (b). Note that all collection periods (upper trap) are presented as 6-interval cycles in order to make comparison easier and attenuate short-range fluctuations. The extreme case (7.6 – 19.6.97 in LL17) is probably due to uneven collection caused by trap itself. These cycles do not correspond to analysed fractions due to strong differences in accumulation rates.

GF2 date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	JML date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	LL17 date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	Fårö date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	BY15 date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$
11.4. - 22.4.	795			14.4. - 22.4.	3504	17.4. - 28.4.	273	23.4. - 2.5.	293
22.4. - 4.5.	1560			22.4. - 4.5.	2281	28.4. - 9.5.	1294	2.5. - 11.5.	591
4.5. - 15.5.	1527			4.5. - 15.5.	2797	9.5. - 20.5.	582	11.5. - 20.5.	292
15.5. - 27.5.	92			15.5. - 27.5.	1195	20.5. - 30.5.	228	20.5. - 30.5.	828
27.5. - 7.6.	603			27.5. - 7.6.	1270	30.5. - 9.6.	89	30.5. - 9.6.	564
7.6. - 19.6.	1430	5.6. - 19.6.	148	7.6. - 19.6.	7932	9.6. - 19.6.	103	9.6. - 19.6.	316
mean	1001			mean	3176	mean	442	mean	481
6.7. - 18.7.	856	6.7. - 18.7.	16	6.7. - 18.7.	2567	3.7. - 15.7.	300		
18.7. - 30.7.	614	18.7. - 30.7.	16	18.7. - 30.7.	2472	15.7. - 27.7.	425		
30.7. - 11.8.	184	30.7. - 11.8.	13	30.7. - 11.8.	2044	27.7. - 8.8.	469		
11.8. - 23.8.	223	11.8. - 23.8.	15	11.8. - 23.8.	434	8.8. - 20.8.	352		
23.8. - 4.9.	271	23.8. - 4.9.	228	23.8. - 4.9.	493	20.8. - 1.9.	79		
4.9. - 16.9.	99	4.9. - 16.9.	14	4.9. - 16.9.	102	1.9. - 13.9.	537		
mean	375	mean	50	mean	1352	mean	360		
27.9. - 9.10.	1314	27.9. - 9.10.	959						
9.10. - 21.10.	2430	9.10. - 21.10.	1032						
21.10. - 2.11.	997	21.10. - 2.11.	595						
2.11. - 14.11.	624	2.11. - 14.11.	1060						
14.11. - 26.11.	1061	14.11. - 26.11.	922						
26.11. - 8.12.	1042	26.11. - 8.12.	482						
mean	1245	mean	842						

GF2 date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	JML date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	LL17 date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	Fårö date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	BY15 date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$
8.4. - 4.7.	4406	4.6. - 3.7.	1431	11.4. - 2.7.	3860	14.4. - 1.7.	3190	21.4. - 29.6.	1619
4.7. - 25.9.	8449	3.7. - 25.9.	4003	2.7. - 25.9.	3983	1.7. - 24.9.	863		
25.9. - 18.12.	8269	25.9. - 18.12.	1984						

(a)

JML date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	LL17 date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	NCB date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$	BY15 date	upper trap $\text{mg m}^{-2} \text{d}^{-1}$
17.4. - 23.4.	1297	21.4. - 29.4.	870	22.4. - 29.4.	1104	23.4. - 29.4.	827
29.4. - 11.5.	513	29.4. - 11.5.	300	29.4. - 11.5.	524	29.4. - 11.5.	580
11.5. - 23.5.	403	11.5. - 23.5.	602	11.5. - 23.5.	238	11.5. - 23.5.	912
23.5. - 4.6.	173	23.5. - 4.6.	239	23.5. - 4.6.	270	23.5. - 4.6.	319
4.6. - 16.6.	232	4.6. - 16.6.	351	4.6. - 16.6.	770	4.6. - 16.6.	284
16.6. - 28.6.	294	16.6. - 28.6.	733	16.6. - 28.6.	635	16.6. - 28.6.	331
mean	485	mean	495	mean	552	mean	516
2.7. - 20.7.	214	2.7. - 20.7.	560	4.7. - 24.7.	606	4.7. - 22.7.	337
20.7. - 7.8.	328	20.7. - 7.8.	628			22.7. - 9.8.	267
7.8. - 25.8.	512	7.8. - 25.8.	625			9.8. - 26.8.	145
25.8. - 12.9.	232	25.8. - 12.9.	581			26.8. - 12.9.	269
12.9. - 30.9.	459	12.9. - 30.9.	622			12.9. - 30.9.	146
30.9. - 18.10.	747	30.9. - 18.10.	1068			30.9. - 18.10.	399
mean	415	mean	681			mean	261

JML date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	LL17 date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	NCB date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$	BY15 date	lower trap $\text{mg m}^{-2} \text{d}^{-1}$
17.4. - 28.6.	1655	21.4. - 28.6.	5221	22.4. - 28.6.	1452	23.4. - 28.6.	2822
2.7. - 19.10.	3173	2.7. - 20.10.	2828	4.7. - 20.10.	1256	1.7. - 21.10.	453

(b)

Modelling – The transportation models (total and mineralogical) assume that the material collected from the lower trap represents inside-basin flux with no capability of further horizontal transportation. It is further assumed that there is an instant contact between the lower trap and adjacent sediment. The material from the upper trap was attributed to be a mobile fraction and is capable of moving to the next basin without a material loss. The sediment-suspension normalizing factor 0.4 was evaluated by comparing the main accumulation rates of suspended material and sediments at GF2 and BY15 stations. The sizes of basin areas are based both on topographical and literature values.

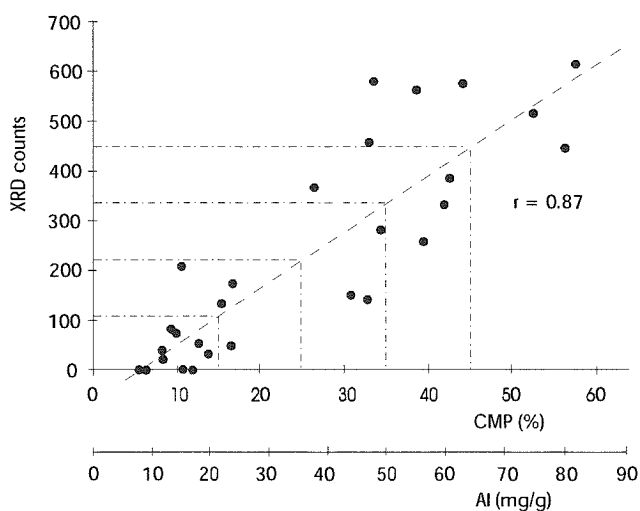


Fig. 4. The correlation of illite 10 Å (001) peak height (total peak count number – background count number) as a function of total Al-content of the sediment. Chemical data from unpublished sources and from Leivuori & Vallius (1998).

3. RESULTS AND INTERPRETATIONS

General

XRD patterns (both total and $< 2 \mu\text{m}$ fraction) are almost identical throughout bottom cores. Although the inflowing salt water pulse can affect material distribution to some extent (Bernard & al. 1989), it is more likely that the homogeneity is mainly due to ancient glaciations. Only a slight drop of peak intensities occur toward the Gotland Deep (Fig. 3 a-f) because of stronger enrichment of organic material in the central Baltic Proper (Emelyanov 1988).

The $< 2 \mu\text{m}$ fraction consists of 70 % illite, 20 % kaolinite and 10 % chlorite thoroughly. Due to a weak presence of the “real” iron minerals it is probable that illite is partly trioctahedral. The contribution of minerals in typical “background” sediment (Al-content is ca. 7 %) is 30 % (clay minerals), 10 % (feldspars) and 10 % (quartz). Using these data it can be estimated that 1 % of total Al corresponds ca. 7 % of crystalline mineral particles (CMP). The correlation ($r = 0.87$) is presented in Figure 4. Only the 10 % CMP intervals have been recorded.

Silicates

Neither temporal nor spatial significant changes of XRD patterns occur in the suspended matter of Gulf of Finland (Fig. 3 a,b). The only significant peak dilution was caused by spring-summer blooms during the collection period between 27.5 – 19.6 (GF2). According to Kankaanpää & al. (1997a), diatoms (opal) contribute almost 100 % of the organic matter during that time. Diatoms radially developed

algae with crystalline outer shell structure, occur as 30-50 μm diameter. The effect of blooms is weaker in JML samples indicating almost an even CMP distribution. The CMP content of the bottom and upper trap samples of GF2 and JML were ca. 50 %. The upper trap values only reached 20-50 % CMP values (Fig. 3 a,b).

Complete different XRD patterns occur in the northern Baltic Proper. During the first stage of collecting periods (late spring – early summer) there are substantial pulse of suspended matter from adjacent sedimentary regions, which are visible in LL17, NCB and Fårö Deep areas (Fig. 3 c,d,e). The melting of ice cover and increasing Neva runoff during spring-time generate an opposite inflow into the Gulf of Finland in the deeper layers (Haapala & Alenius 1994, Myrberg 1998) which probably explains the pulses observed. Because of the limited number of collection periods it is unknown whether this pulse is a single event or multiplied series of individual pulses. It is, however, evident that the pulse generation or attenuation is rather gentle and not a violent, abrupt event (Fig. 3 c,e). The maximum CMP values (20 %) generation as well as CMP attenuation to minimum values (10 %) last 1-2 months. The pulses after calm summer periods are probably caused by autumn storms.

The maximum of CMP during late spring – early summer and middle autumn (20 %) is of the same magnitude as that found in adjacent bottom sediments in the northern Baltic Proper. During these periods total particulate matter (TPM) is probably governed by sediment-originated material with only minor effects of non-sediment originated authigenic processes. In contrast, almost tranquil conditions prevail most of the summer period. TPM fluxes (for a single station) are somewhat even during the whole collection period in LL17, NCB as well as the Fårö Deep (Table 1) suggesting that a scarcity of XRD peaks in that time is not explained by a “diluting effect” of an unusual intensive organic detritus accumulation.

Ferromanganese concretions

A major part of summer-seasonal suspended matter (northern Baltic Proper) contain substantial amounts (even > 75 %, calculated as MnO_2) of 5-30 μm reddish brown Fe-Mn-bearing concretionary material which are amorphous and almost completely absent in the sediment surfaces in the studied areas (Appendix). Bernard & al. (1989) found that 88 % of suspended particles are Mn-Si-Fe-type in near-bottom waters in the northern Baltic Proper during the non-productive season. Porosity was estimated to be 80 v-% which yields an apparent density of 0.21 – 0.28 g cm^{-3} ($n=3$) for total sediment. Fuerstenau and Han (1977) reported ca. 60 % porosity of deep-sea ferromanganese concretions (FMC) and Baltic concretions closely resemble deep sea FMC (Winterhalter and Siivola 1967). Although the density value is only indicative, it possibly explains the almost total absence of FMC on adjacent bottom sediments. The HNO_3 leachable Mn/Al ratio is 16.0 ± 23.1 ($n=23$) (Fig. 5 a,b,c) in the Baltic Proper area; that is, the estimated total Mn/Al is ca. 10. Ingri & al. (1991) found a higher Mn/Al main value (27.5) in the suspended matter of the Baltic Proper. By contrast, Mn/Al values in the sediments are close to the typical background values (0.01). The enrichment of many trace metals in the TPM relative to adjacent sediments has also been established (Brügmann & al. 1992).

FMC also occur in the suspended matter in the Gulf of Finland, occasionally even at high concentrations (Appendix). Exceptional Mn-enrichments may occur in certain time periods in the Gulf of Finland area (Fig. 5e and 6b). The enrichment of Mn-rich particles is, however, generally lower in the Gulf of Finland than the Baltic Proper which is also seen in more light coloured specimens of the Gulf of Finland area as well as lower HNO_3 -leachable Mn/Al ratios (0.86 ± 0.92 , $n = 18$; Fig. 5d,e). Mn-rich particles are seasonally abundant above the anoxic near-bottom level in the basins of the Baltic Proper (Ingri & al. 1991). In contrast, near-bottom water is usually oxidized in the Gulf of Finland area (Pertilä & al. 1995).

There is a tendency for an increase in the Mn/Fe ratio toward the Gulf of Finland area (Fig. 5). If the estimated HNO_3 leachable crustal background Mn/Al (0.017) and Fe/Al (1.0) ratios were eliminated, the proportions of FMC bound manganese (as MnO_2) and iron (as Fe_2O_3) can be evaluated. The relative proportion of MnO_2 is 69-86%, (BY15, $n = 3$), 93-95% (Fårö, $n = 4$) and 96-98% (LL17, $n = 3$). The FMC component is very close to the pure Mn-component in the Gulf of Finland area since it is not

possible to detect a non-background Fe component. By comparison, the dissolved O₂ content were 0-0.2 ml l⁻¹ (BY 15), 0.1-0.3 ml l⁻¹ (Fårö), 0.2-1 ml l⁻¹ (LL 17), 1.2-9.2 ml l⁻¹ (JML) and 3.3-7.1 ml l⁻¹ (GF 2) [O₂ data from the FIMR database]. Evidently a more reducing environment inhibits authigenic Mn-precipitation relative to Fe precipitation allowing lower Mn/Fe values in the Baltic Proper area.

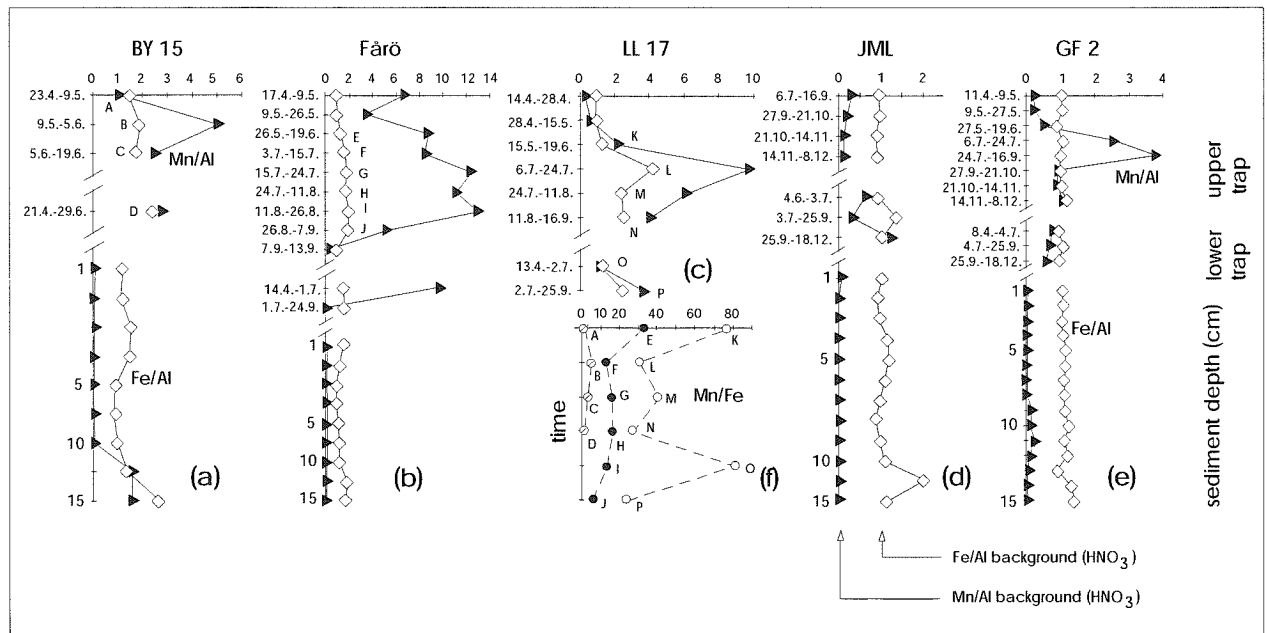


Fig. 5. HNO₃ leachable Fe/Al and Mn/Al ratios in the sediment and suspended samples in 1997 (1998 data not available): (a) BY 15, (b) Fårö, (c) LL 17 (Mn/10*Al), (d) JML, (e) GF 2. Fig. 5 (f) shows calculated Mn/Fe ratios of FMC in BY 15, Fårö and LL 17 area (it was not possible to detect FMC-Fe in GF2 and JML samples). Symbols A-P refer to trap locations and time periods. Chemical data from unpublished sources.

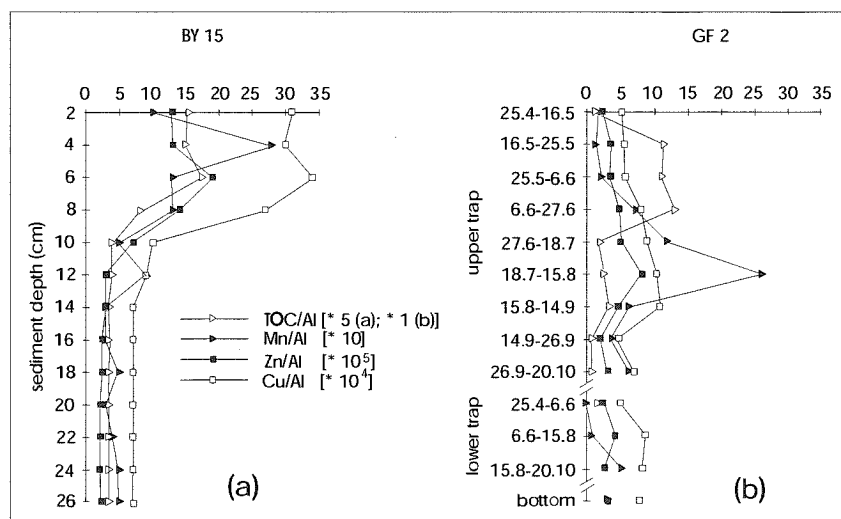


Fig. 6. Al-normalized data from the Gotland deep sediment column “final sinking area” (a) and from GF2 traps and surface sediment (“starting flux material”) (b). Average near-shore mud data (not given) from Chester & Aston (1976). Average shale data (not given) from Wedepohl (1991). GF2 raw data from Leivuori & Vallius (1998) [with permission from Elsevier Science]. BY15 data was redrawn after Emeis & al. (1998) [with permission from Elsevier Science].

Unlike most cases of suspended FMC rich TPM, a clear 2.71 Å peak ($32.95^\circ 2\theta$) was found in a single sample of the Gotland Deep (Fig. 3f). It is not clear whether or not this peak is linked to the FMC. According to the data of Burns and Burns (1977) the only candidate from the FMC group could be hematite (α -Fe₂O₃), $d_{(1=100)} = 2.69$ Å which is supported by lower Mn/Fe values (Fig. 5f) in that area. This is, however, speculative since it is not possible to detect $d_{(1=60)} = 1.69$ Å.

Carbonate

The abundance of authigenic Mn-rich carbonate particles in sediments (Ca-rhodocrosite, kutnahorite) are well known in some basins in the Baltic Proper (Suess 1979, Manheim 1982, Jakobsen & Postma 1993, Salonen & al. 1995, Neumann & al. 1997, Sternbeck and Sohlenius 1997). Using the generalised formula Mn_{0.7}Ca_{0.3}CO₃ (Neumann & al. 1997) and Mn-content in laminated sequence of BY15 (Appendix, Fig. 5a) ca. 10 % carbonate content is reached which is consistent with the study of Jakobsen & Postma (1993). Both biogenic (rounded, diameter 5-10 µm) and inorganic (rhombohedral, 10-100 µm) carbonate particles were observed. The morphology of carbonate particles in suspension are clearly reminiscent of the carbonates visible in adjacent sediments. Also some angular, 100-200 µm detrital carbonate clasts (BY15 upper trap) were detected indicating the origin of Phanerozoic bedrock formations. There are some Ordovician outcrops on the slopes of some basins in the Baltic Proper area and these formations contain limestones, dolomites, marls as well as shales and sandstones (Emelyanov & Kharin 1988, Tuuling & al. 1995). The d-value of strongest X-ray peak (≈ 2.89 Å) is uniform in all samples and excludes the presence of limestone but not the possibility that dolomite and Ca-rhodocrosite/ kutnahorite coexist in the suspended matter. Authigenic Mn-carbonates with a detrital dolomitic core has also been reported (Jakobsen & Postma 1993).

Grain size and morphology

No systematic variations in grain size distribution of fine fraction occur toward the deeper basins (Table 2). The observed main portion of “clay-fraction” (22 %; 4 µm limit) is somewhat lower than measured by Emelyanov (1995) from surface muddy sediments in the Gulf of Finland (34 %; 1 µm limit). Vallius (1999) estimated that there was a 65 % clay fraction in the GF2 sediments with only a weak vertical variation. Different methodology as well as different pretreatments can distort interpretation considerably (Barbanti & Bothner 1993, Konert & Vandenberghe 1997). The portion of detrital clasts (> 20 µm) appears to be slightly higher in buried deposits (5-20 %) rather than in surface sediments (< 5 %) in the Gulf of Finland (Appendix). The textural maturity of these clasts is weak indicating single-pulsed stormy events rather than continuous stress during deposition/reworking. Eddy-like short life-time currents occur in the near bottom layer in the Gulf of Finland (Alenius & al. 1998). Despite the general absence of clasts in the northern and central Baltic Proper sediments, there are no clear sign of relative clay mineral enrichments toward the depths (Table 4). Fine quartz particles in the < 2 µm fraction can represent about 20 % (Konert & Vandenberghe 1997).

Table 2. Clay/silt+sand ratios (4 µ limit) for sediment samples. There are no grain size data for suspended material due to low sample amounts.

	GF2 (n=1)	JML (n=3)	LL17 (n=1)	NCB (n=3)	Fårö (n=1)	BY15 (n=1)
1-3 cm	0.16	0.23 (0.17-0.28)	0.19	0.32 (0.28-0.33)	0.25	0.23
3-5 cm	0.25	0.28 (0.23-0.30)	0.28	0.33 (0.32-0.37)	0.23	–
20-22 cm	0.28	0.32 (0.32-0.33)	0.28	0.28 (0.26-0.33)	0.28	0.27
22-24 cm	–	–	–	–	–	0.27

Heavy minerals

Pyrite crystals were detected in almost every sediment core (Appendix). The size of these framboids vary from 10 to 50 microns. Elevated Fe/Al values measured from lower sedimentary sequences of BY 15 (Fig. 5a) indicate ca. 5 % FeS₂ content which is consistent with earlier studies (Lein 1983, Boesen & Postma 1988, Sternbeck & Sohlenius 1997) and evidently corresponds to pyrite maximum observed from the laminated sequence (Appendix). There are also some indications of pyrite framboids in suspension. These observations are, however, partly obscure and not clearly connected with their environmental conditions. Resuspended pyrite framboids have been reported earlier by Skei 1988, Middelburg & al. 1988. Sulfidic metal colloids can occur in the water column of the Baltic Sea (Brügmann & al. 1997).

Less than 1-2 % of the amphibole crystals (< 50 μ) occur in shallower basinal areas (GF2, JML). Apparently part of these particles are fine grained because a characteristic X-ray peak (8.4 Å) is visible in almost every sample. Despite the higher density, the fine-grained nature probably allows mineral existence even in the upper parts of the water column. Belmans & al. (1993) found “(Ca,Fe) Al-silicate” existence in Baltic Sea sediments with 2 % abundance that apparently denotes amphibole group minerals.

4. DISCUSSION

Sedimentation rates

To evaluate the role of basin-basin fluxes some simplifications must be made. Only limited published data of total sediment accumulation rates in the Baltic Proper area are available. Niemistö & Voipio (1981) measured a value of 2.2 mm a⁻¹ in the superficial sediments of the northern Baltic Proper (Teili 1) using ²¹⁰Pb datings. The dry weigh content of these sediments are ca. 10 % which yields values of 700 – 900 mg m⁻² d⁻¹. The sedimentation rates in the Gotland Deep area are very variable even in slightly different localities (Niemistö and Voipio 1974). Salonen & al. (1995) estimated the sedimentation rate of 1.0 – 1.6 mm a⁻¹ in BY15 based on varve counting. Emeis & al. (1998) measured 300 – 900 mg m⁻² d⁻¹ and 2.1 – 2.5 mm a⁻¹ at different locations of the Gotland Deep area based both on ¹³⁷Cs and ²¹⁰Pb methods. The sedimentation rate varies also widely in the Gulf of Finland, even within the basin (Kankaanpää & al. 1997b, Vallius 1999). Based on several ¹³⁷Cs measurements, Vallius and Leivuori (1999) estimated a sedimentation rate of 2400 mg m⁻² d⁻¹ in the whole Gulf of Finland area.

Basin areas

Also the rational extent of basin areas need to be evaluated. This approach omits the “real” parameters of basins. For example, the “real” basin size of GF2 called here as a provenance basin is ca. 10 km² (Vallius 1999). By contrast, the surface area of the Gotland Deep (“final accumulation”) basin is at least two orders of magnitude larger. The basin area evaluation is based on topographical features in the Baltic Proper; in the Gulf of Finland these values are based on accumulation bottom data presented by Vallius & Leivuori (1999). The LL17 and NCB stations’ results are combined because of the same basin area location (northern Baltic Proper). The sizes of the sedimentation areas are defined here as 2000 km² at GF2 station, 2000 km² at JML station, 2000 km² at LL17 and NCB combined station, 1000 km² in the Fårö Deep and 3000 km² in the Gotland Deep. Because the flux can proceed both to the Landsort Deep and the Gotland deep direction, the total flux from the stations of LL17 and NCB is multiplied by a factor 2/3. This value is only a formal selection. The Gulf of Finland originated material represents only a minor fraction of the central Baltic Proper (BY 15) accumulation. The arriving angle at station BY 15 from the Gulf of Finland direction was estimated to be 5/360 (degrees). It is used 800 mg m⁻² d⁻¹ at BY 15 instead of 300 mg m⁻² d⁻¹ (see Emeis & al. 1998). The latter case probably represents the surface fluffy layer (Perttilä & Niemistö 1993) which is capable of making back-and-forth movements along bottom (Winterhalter, pers.comm.). Thus, the estimated accumulation in the Gotland Deep is 12 mg m⁻² d⁻¹ from the Gulf of Finland direction.

Transportation mechanism

The third aspect is the transportation mechanism generation. Leivuori & Vallius (1998) used the material collected from the upper trap as a “primary, sinking particulate matter” and further calculated that the “excess” material collected from the lower trap can use to represent a resuspension parameter. This kind of approach, however, is valid only during spring and autumn blooms. In the open Baltic Proper the water column carbon content can reach even 50 % during the productive season (Blomqvist & Larsson 1994). Organic matter was estimated to be $2.1 * C_{org}$ by weight (Widerlund 1996), that is, almost 100 % of material occur in the biogenic form during intensive blooms. In general, the suspended load of biogenic matter can vary considerably (40-80 %) in the Baltic Sea area (Emelyanov & Pustelnikov 1975, Boström & al. 1981). However, primary productivity during late autumn-winter-early spring is only 2 % compared to the bloom periods (Bernard & al. 1989) and the gross carbon sedimentation is often dominated by resuspended organic residue (Blomqvist & Larsson 1994). In addition, approximately 20 % of the total organic carbon can be resuspended during bloom periods in the coastal area of the Gulf of Finland (Heiskanen & Leppänen 1995). The eolian originated mineral particles are probably of minor significance relative to current transported near-bottom particles. This mechanism also applies to many metals (Boström & al. 1983) and there seems to be a clear correlation between selected metals and Al in the suspended matter of GF2 (Leivuori & Vallius 1998). It is, thus logical to assume that there is stronger bottom-trap interaction of material rather than primary productivity controlled trap accumulations.

To justify previous arguments it is possible to establish the rough relative content of CMP and organic part in the upper trap collection of GF2 (data based on Leivuori & Vallius 1998). The accumulation rate is ca. $300 \text{ mg m}^{-2} \text{ d}^{-1}$ of C_{org} during late spring – early summer period and $75 \text{ mg m}^{-2} \text{ d}^{-1}$ for the rest period. It is here assumed that the latter content can also be applied to the unknown winter-period. The main Al flux is ca. $40 \text{ mg m}^{-2} \text{ d}^{-1}$ annually. Calculated amounts of organic matter ($230\text{-}240 \text{ mg m}^{-2} \text{ d}^{-1}$) and CMP ($280 \text{ mg m}^{-2} \text{ d}^{-1}$) indicate that the minerogenic flux slightly prevails relative to biogenic counterpart. Puls & al. (1997) reported that even 80 % of the suspended matter consist of mineral particles in the German Bight in wintertime. In addition, a considerable part of the organic carbon appears to be detrital and sediment-originated authigenic minerals occur in suspension (e.g. Mn-carbonate). Hence, the sediment-driving suspension mechanism can be established more easily. It is here assumed that the lower trap material is not capable of basin-basin movement and the upper trap material represents a mobile fraction.

Sediment-suspension relationship

The final approximation concerns the relationships between the annually accumulated sediment and the adjacent suspended matter. The main accumulation ratios of the upper trap, the lower trap and surface sediments are $0.44 : 3.5 : 1$ at the GF2 station and $0.47 : 1.68 : 1$ at the BY 15 station, respectively. In the northern Baltic Proper (LL17 and NCB stations), this ratio reaches the values of $1.4 : 3.8 : 1$ if estimated sediment accumulation rate $800 \text{ mg m}^{-2} \text{ d}^{-1}$ is used. There appears to be rather constant “excess” of material accumulated in lower traps relative to adjacent sediments. Assuming that the material in the lower traps is rather sediment than primary production controlled, it is reasonable to assume that this “excess” is mainly due to stronger deposition during measuring periods relative to winter-time periods when ice cover exists. All trap flux values (Fig. 7a,b) were multiplied by a factor 0.4 in order to normalize this effect.

Model A: Transportation flux without any limitations

The flux chart for TPM is presented in Fig. 7a. After correction the initial “starting” flux of the GF2 station ($2800 \text{ mg m}^{-2} \text{ d}^{-1}$) is still 1.4 fold relative to the sediment value ($2000 \text{ mg m}^{-2} \text{ d}^{-1}$). The estimated sediment flux for the whole Gulf of Finland varies considerably ($1400 \text{ mg m}^{-2} \text{ d}^{-1}$, Kankaanpää & al. 1997a; $2400 \text{ mg m}^{-2} \text{ d}^{-1}$, Vallius & Leivuori 1999) which neglects the initial flux error estimate. Using a 2000 km^2 influencing area for the station GF2 the estimated amount of potential mobile fraction (GF2 upper trap) is $700 * 10^6 \text{ g d}^{-1}$. According to this model, the “final” flux from the Fårö upper trap will be

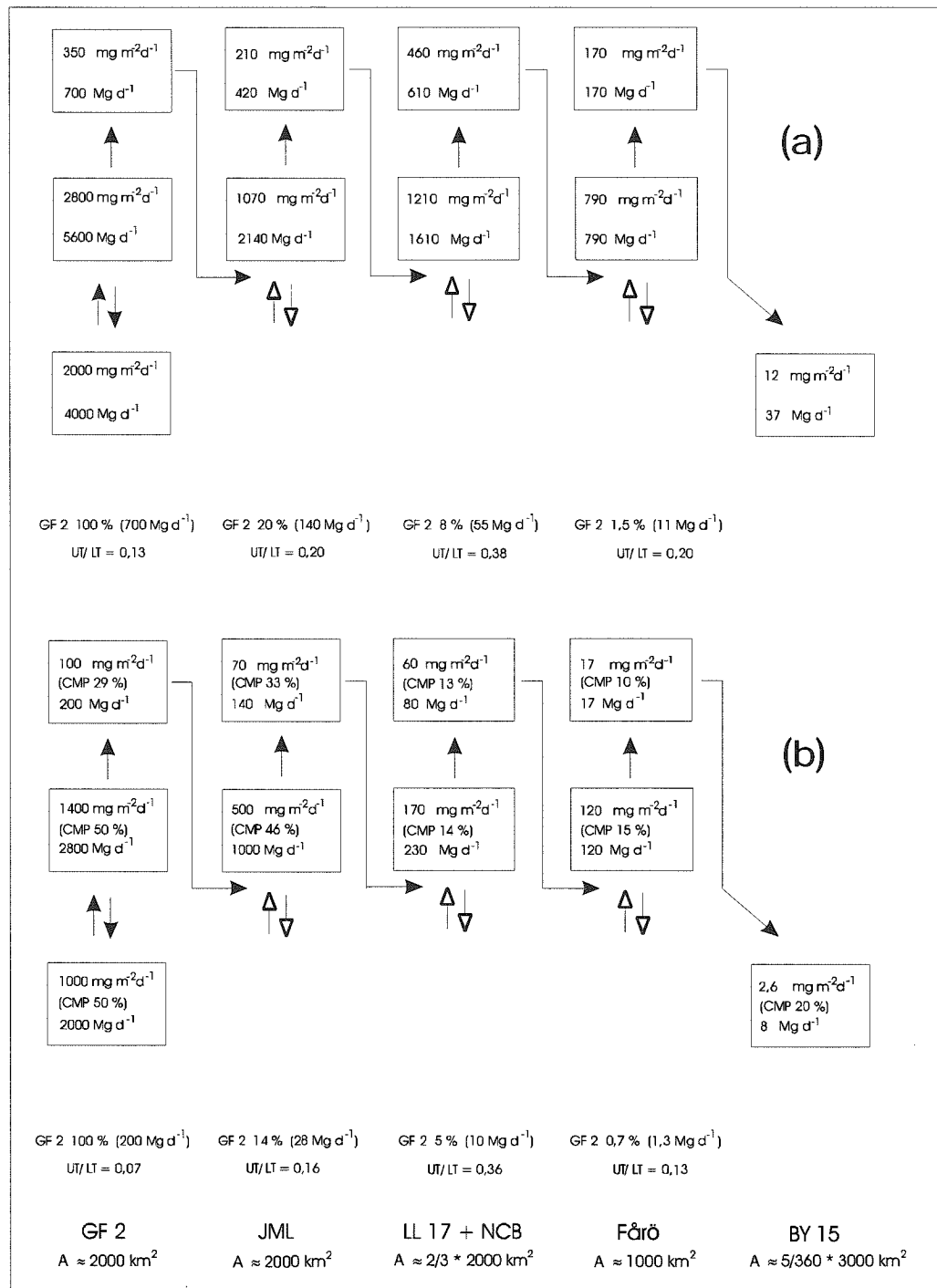


Fig. 7. Assumed total (a) and CMP (b) flux patterns from the GF2 to the BY15. Surface sediment values are original, all trap data were multiplied by a factor of 0.4. UT = upper trap, LT = lower trap. The parameters of UT in the Fårö area represent “final flux” to the Gotland Deep. GF 2 %-value indicates the proportion of GF2 originated material in the UT. Note that the model ignores trap parameters measured from BY15. GF2 approximated sediment data from Kankaanpää & al. (1997a); Vallius & Leivuori (1999). BY15 sediment data after (Erems & al. 1998).

A calculation example: If 200 Mg d⁻¹ of CMP (GF2 UT) reaches the JML bottom which has active interaction between LT, the JML LT and thus UT contain 20 % of GF2-originated material. This is ca.15 % of original GF2 UT value.

170 * 10⁶ g d⁻¹ which overestimates the estimated accumulation rate (40 * 10⁶ g d⁻¹) of the BY15 station four-foldly. This model takes not into account the real nature of particulate matter. Suspension contains variable amounts of material which has formed authigenically in the deep water column

(FMC). Bacterially mediated *in situ*-formed Mn-particle enrichments have been reported in deep ocean waters (Cowen & Bruland 1985, Bruland & al. 1994, Kuss & Kremling 1999). Manganese oxidizing microorganisms are able to survive in deep waters due to a lack of photoinhibition (Sunda & Huntsman 1988). The enrichment of Mn in suspension can reach a magnitude of 10^2 - 10^3 relative to adjacent sediments (Fig. 5). In addition, different forms of Mn-particles in suspension (mainly FMC) and sediments (mainly carbonates) exclude the possibility for vigorous interaction between surface Mn-particles and suspended Mn-particles. The suspended FMC cannot be linked with shallow water located FMC because of the clearly crystalline character of the latter (Glasby & al. 1996, Zhamoïda & al. 1996). The total organic carbon (TOC) content can vary considerably even in the near-bottom – bottom transition (Leppänen 1988). The higher content of organic matter contributes strongly to the sedimenting values in the Gotland Deep area. It is evident that Model A contains many shortcomings for particulate matter flux behaviour which make this interpretation less useful.

Model B: Transportation flux only by CMP

This model (Fig. 7b) takes only into account resistant mineral particles (aluminosilicates + quartz). This fraction remains practically unaffected by any biogeochemical processes, e.g. redox-variations. Also the mass errors due to organic matter degradation and formation/destruction of authigenic particles will be neglected. The CMP flux which is capable of moving toward the depths is $200 * 10^6$ g d^{-1} , which is only 7 % of total accumulation in GF2 area. The “final flux” from Fårö upper trap is $16 * 10^6$ g d^{-1} which differs from estimated accumulation in the BY15 ($8 * 10^6$ g d^{-1}) only 50 %. The main shortcomings are (i) clear carbonate-CMP correlation in TPM especially in the northern Baltic Proper indicating local suspension processes rather than a longer-range transportation mechanism, (ii) the lack of particle abrasion and (iii) the almost constant ratios among aluminosilicates. Weak currents capable of transporting organic-rich loose sediment material to deeper areas (Winterhalter 1992) cannot abrade or segregate silicate particles to a great extent. The enrichment of finer-grained feldspar or quartz grains in deeper areas is almost purely due to calm sorting by eddy-like short life-time currents. Despite the advantages of Model B relative to A, the overestimation of fluxes has also been established for this model.

Model C: An *In situ*-model with only slight ability of transportation

Models A and B are, in fact, supported *in situ* due to low upper/lower trap -ratios (0.1 – 0.4). Models A and B suggest that only 1-2 % ($11 * 10^6$ g d^{-1}) and < 1 % ($1.3 * 10^6$ g d^{-1}) of “finally accumulated” material is derived from the GF2 station. Also the final accumulation portions from GF2 are small compared to total accumulation from that direction, i.e. $susp_{(GF2)} / susp_{(JML, LL17+NCB, Fårö)}$ are only 7% and 8%. Due to shortcomings of models A and B, the real final accumulations are even smaller. Unlike northern Baltic Proper there are no significant temporal fluctuations of material behaviour in the Gulf of Finland. Assuming that significant basin-basin mass transfer processes occur, local temporal features would probably be mixed more efficiently by intruding fluxes. It is probable that there will be detectable XRD peaks throughout the collection periods in the northern Baltic Proper area if significant material transfer from the Gulf of Finland occur. Typical near-bottom currents in the northern Baltic Proper (3-5 $cm s^{-1}$, Leppänen & Alenius 1988) allow material transportation from the Gulf of Finland to the northern Baltic Proper during 1-2 months if the flux is continuous.

There are no evidence for relative enrichments of CMP in the sediment (Table 3) indicating a poor reworking capability by currents. Another proof of *in situ* dominance is a weak sign of feldspar and quartz enrichment relative to clay particles in the upper traps compared to adjacent sediments. If this was the case the mobile flux would increase feldspar and quartz enrichment to the deeper areas which is not observed.

Although the *in situ* type processes seem to prevail in the studied areas, some precautions must be taken. Despite the clear carbonate-CMP relationship in the northern Baltic Proper it is unknown why this ratio is more obscure in the JML area. Carbonate occurrence in the traps is partly connected to the vernar period (April-May); in addition, carbonate is almost absent in the JML sediments (Fig. 3 b,c,d;

Appendix). The breakdown of organic matter and ice cover during the winter can lower redox level in the surface sediments in the Baltic region (Ingri & Ponter 1986) favouring carbonate formation. Carbonate crystals cannot persist for long periods in the water columns of the Baltic area due to undersaturation (Perttilä, pers.comm.). If this mechanism explains sporadic carbonate occurrence in the water column (JML), it does not explain carbonate-CMP correlation in the northern Baltic Proper area. The other problem is carbonate occurrence in the upper traps and the sediment but absence in the lower traps in the Gotland Deep. At least part of carbonate particles must be sediment-originated (see description earlier) suggesting within-basin lateral movements. Lateral particle transport occur at the halocline of the Baltic Sea (Boström & al. 1988, Glasby & al. 1996). The upper trap locations were, however, ca. 100 m below the halocline at BY15 station.

It is also possible that there is an active bottom benthic nepheloid and a flyffy layer capable of material transport (Leipe, pers.comm.). In that case the lower trap adjustments (5 m above bottom) result an underestimation of true transportation capability. The lower sediment accumulation rates relative to lower traps may reflect this process (Leipe, pers.comm.).

This study strongly supports the earlier observations (Winterhalter 1972, Boström & al. 1983) that a submarine erosion of older deep sediment deposits by postglacial land up-lift generates the main part of present day accumulation to the basins in the Baltic Sea.

Table 3. XRD-statistics for studied samples. The ratios are based on relative peak heights. Clays/feldspars = $10\text{Å}+7\text{Å} / 3.19\text{Å}+3.24\text{Å}$; clays/quartz = $10\text{Å}+7\text{Å} / 1.82\text{Å}$; feldspars/quartz = $10\text{Å}+7\text{Å} / 1.82\text{Å}$; clay/clay = $10\text{Å} / 7\text{Å}$.

	clays/ feldspars		clays/ quartz		feldspars/ quartz		clay/clay	
GF2 UT	1.08 ± 0.44	(n=12)	4.94 ± 1.31	(n=12)	5.07 ± 1.83	(n=12)	1.05 ± 0.09	(n=12)
GF2 LT	2.16 ± 0.54	(n=3)	8.17 ± 1.06	(n=3)	3.90 ± 0.75	(n=3)	1.21 ± 0.18	(n=3)
GF2 SED	2.46 ± 1.36	(n=4)	11.3 ± 11.7	(n=4)	4.05 ± 2.33	(n=4)	1.09 ± 0.13	(n=4)
JML UT	1.60 ± 0.96	(n=11)	8.90 ± 3.96	(n=11)	6.50 ± 3.25	(n=11)	0.92 ± 0.15	(n=11)
JML LT	2.47 ± 1.15	(n=6)	12.7 ± 3.60	(n=6)	5.62 ± 1.56	(n=6)	1.08 ± 0.12	(n=6)
JML SED	2.78 ± 1.48	(n=4)	13.1 ± 11.6	(n=4)	4.10 ± 1.71	(n=4)	0.91 ± 0.13	(n=4)
LL 17 UT	1.02 ± 0.31	(n=8)	4.79 ± 2.23	(n=8)	4.71 ± 1.49	(n=8)	0.96 ± 0.13	(n=8)
LL17 LT	1.02 ± 0.29	(n=4)	2.74 ± 1.23	(n=5)	3.03 ± 0.79	(n=4)	0.89 ± 0.19	(n=4)
LL17 SED	1.15 ± 0.18	(n=2)	3.30	(n=1)	3.30	(n=1)	0.75 ± 0.11	(n=2)
NCB UT	0.85 ± 0.36	(n=5)	4.42 ± 3.06	(n=6)	5.10 ± 2.26	(n=5)	0.92 ± 0.12	(n=5)
NCB LT	1.57	(n=1)	8.90	(n=1)	5.70	(n=1)	1.00	(n=1)
NCB SED	1.85 ± 0.53	(n=5)	(no data)		(no data)		0.94 ± 0.11	(n=5)
Fårö UT	0.94 ± 0.29	(n=7)	3.01 ± 1.38	(n=7)	3.32 ± 0.96	(n=5)	0.96 ± 0.35	(n=7)
Fårö LT	0.83 ± 0.24	(n=2)	1.70	(n=1)	2.60	(n=1)	0.84 ± 0.17	(n=2)
Fårö SED	2.33 ± 0.75	(n=4)	9.18 ± 7.46	(n=4)	3.58 ± 1.62	(n=4)	1.01 ± 0.21	(n=4)
BY15 UT	1.29 ± 0.35	(n=7)	4.95 ± 1.61	(n=8)	4.39 ± 2.15	(n=7)	1.01 ± 0.29	(n=7)
BY15 LT	2.06 ± 0.97	(n=4)	6.38 ± 1.11	(n=4)	4.13 ± 3.12	(n=4)	0.94 ± 0.11	(n=4)
BY15 SED	2.02 ± 0.84	(n=3)	3.70 ± 2.52	(n=3)	1.70 ± 0.62	(n=3)	0.93 ± 0.12	(n=3)
Mean	1.56 ± 0.90	(n=92)	6.61 ± 5.10	(n=88)	4.58 ± 2.20	(n=83)	0.98 ± 0.18	(n=92)

Geochemical aspect

Fig. 6a and b present some Al-normalized geochemical features between “starting material” (GF2 surface and traps) and “finally accumulating material” (BY15 sediment core). The increase of TOC/Al values at GF2 station are independent of Cu/Al and Zn/Al ratios indicating that TOC does not effect Cu

and Zn behaviour at GF2 station. This behaviour can be generalized to many other trace metals as well (Vallius & Leivuori 1999). By contrast, there is a striking correlation between TOC, Zn and Cu at BY15 station. The “background” Cu/Al ratio is 5.5 – 8.5 ($\times 10^4$) i.e. the enrichment factor (EF) is always < 2 indicating natural origin. The Zn/Al background EF of 2-4 can also be natural origin since EF values from unity to an order of magnitude may still indicate crustal material source due to uncertainties in fractionation during the weathering processes (Duce & al. 1976). Whether or not the CMP fraction is mobile, it does not prove enrichments of Cu and Zn in the central Baltic Proper. The slight enrichment of Cu and Zn with Mn-precipitates in the GF2 suspension cannot be established in the central Baltic Proper where Mn mainly occur as carbonates. This is easily seen in high background values (Mn/Al = 1-5 in the central Baltic Proper, see also Fig. 5a) which exceed the “real” background value by a factor of 100-500. Substantial Mn/Al enrichments in GF2 suspension do not enrich deeper sediments systematically. Decreasing Mn/Fe ratios in suspension toward deeper areas are due to dissolved oxygen content in the water column rather than alterations during transportation. The discussion above excludes the Cu and Zn transportation via TPM from the Gulf of Finland to the central Baltic Proper. It is probable that only a few inflows of salt pulses from the Danish straits over any period of time are capable of transporting enough metal-rich particulates into the basins (Suess & Erlenkeuser 1975). Also transportation in the dissolved form by redox changes seems to be unlikely alternative since the K_d values (“leachable fraction”/“dissolved fraction”) are very high (magnitude of 10^5) for many metals in the Baltic Sea (Brügmann & al. 1992). Dissolved Cu and Zn in the anoxic water column are ca. 20 % and 10 % compared to the oxic values (Brügmann & al. 1997, 1998). The possible dissolved migration under oxic environments, on the other hand, allows neglect of sediment data in the Gulf of Finland.

5. CONCLUSIONS

The study, based on a novel approach, clearly shows that the transportation of the particulate matter from the shallower basins to the deeper areas in the northern Baltic Proper is negligible relative to direct basin infillings from adjacent land areas. A submarine erosion of older deep sediment deposits by postglacial land up-lift generates most of present day accumulation to the basins in the Baltic Sea. Only a few percent of material originated from central parts of the Gulf of Finland finally reaches the Gotland Deep basins. According to parameters used in this study, there are no indication that the metal enrichments seen in the Gotland Deep are due to long range particle transport control in the northern side of the Baltic Sea. The minor transported fraction is efficiently masked by local processes which makes tracing of particle movement almost impossible in the Baltic Sea area.

Acknowledgements

The author would like to thank Dr. Martti Lehtinen of the University of Helsinki, Department of Geology and Mineralogy, who performed the X-ray scans. Dr. K. Boström, Dr. T. Leipe, Prof. Matti Perttilä, Dr. Eeva-Liisa Poutanen and M-L. Räisänen gave valuable comments on an earlier drafts of this paper. The author is grateful to Elsevier Science for use of copyrighted data. Donald Smart kindly improved the language before the final version of this paper.

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Appendix

Sediment (S) and trap (T) stations, their coordinates (based on the Finnish KJ system with GPS positioning), water depth (wdepth) at given station and sampling date/collection periods. Sampler: Niemistö gravity corer (NGC), Eila-trap (Eila), French Technicap traps (PPS 3/3, PPS 5/2). Sampling depth (sdepth): cm below sediment surface and meters above sediment surface.

Visual description: ox., oxic; sox., suboxic; aox., anoxic; lam., laminated; fluffy, flyffy layer; br., brown; gr., gray; p.br., pale brown; l.br.gr.; light brownish gray; l.gr.br., light grayish brown; gr.br., grayish brown; re.br., reddish brown; ye.br., yellowish brown; d.br., dark brown; v.d.br., very dark brown; d.gr.br., dark grayish brown; v.d.gr.br., very dark grayish brown.

Relative contribution of microscopically determined minerals (note that the visual scale is not proportional to real values, see notes below): clasts (quartz and feldspars, $> 20 \mu$); amph., amphibole; carb., carbonate; FMC, Fe-Mn-concretions; ND, not defined; ?, uncertain observation. Symbols: – absent or slightly visible, + visible, ++ moderately visible, +++ clearly visible, ++++ very clearly visible. Enrichment factor for Fe and Mn are calculated from HNO_3 soluble fraction: $(\text{Fe}/\text{Al})_{\text{sample}} / 1$, $(\text{Mn}/\text{Al})_{\text{sample}} / 0.017$.

Notes: Microscopically estimated maximum fractions: 10-20 % for clasts (a), 1-2 % for amphibole (b), 4-8 % for pyrite (c). Location correction: 0-2 cm (d), 2-10 cm (e), 10-15 cm (f) in Fig. 5a due to different positions. 1 M HCl test for carbonate-rich sediments: 22 % (g), 18 % (h). 1 M Hydroxylamine hydrochloride – 1 M HCl leach test for FMC-rich suspension: 67 % (i), 39 % (j) and 61 % (k).

station	latitude	longitude	wdepth	date	sampler	sdepth	description	class	amph.	pyrite	carb.	FMC	Fe (EF)	Mn (EF)
GF2 S	59.50.36 N	25.51.57 E	82 m	8.4.97	NGC	0-3 cm 3-5 cm	ox., gr.br. sox./aox., l.br.gr.	++ ++	+ (b) +	- +	- -	ND ND	1.1 1.2	2.7 2.1
GF2 S	59.50.30 N	25.51.58 E	84 m	15.4.98	NGC	0-1 cm 20-22 cm	aox., l.br.gr. ox., p.br. aox., l.br.gr.	+++ (a) ++ ++	+ + -	+ + +	- - -	ND + -	ND ND ND	ND ND ND
JML S	59.35.01 N	23.38.00 E	79 m	9.4.97	NGC	0-2 cm 4-5 cm	ox./sox., p.br. aox., l.br.gr.	+ +	- -	- +	- -	ND ND	1.0 1.1	1.8 0.9
JML S	59.35.28 N	23.38.92 E	79 m	16.4.98	NGC	0-1 cm 20-22 cm	sox./aox., lam., p.br. aox., l.br.gr.	++ ++	+ +	+ +	+ -	- -	ND ND	ND ND
LL17 S	59.01.91 N	21.03.96 E	159 m	20.4.98	NGC	0-1 cm 3-5 cm 20-22 cm	aox., lam., p.br. aox., lam. gr.br. aox., gr.br.	+ + +	+ + -	- + ++	++++ (e) ++++ ++	+ - +	ND ND ND	ND ND ND
NCB S	58.49.10 N	20.15.16 E	174 m	21.4.98	NGC	0-1 cm 20-22 cm	aox., lam., p.br. aox., gr.	+ +	- -	+ ++	++ -	- +	ND ND	ND ND
NCB S	58.46.00 N	20.13.37 E	224 m	21.4.98	NGC	0-2 cm	aox., fluffy, p.br.	-	-	-	+++	-	ND	ND
Fårø S	58.05.40 N	19.50.59 E	180 m	14.4.97	NGC	0-5 cm 20-22 cm	aox., gr.br. aox., l.br.gr.	- +	- -	+ +	- -	ND ND	1.1 ND	0.9 ND
Fårø S	58.00.01 N	19.53.99 E	194 m	22.4.98	NGC	0-1 cm 20-22 cm	aox., fluffy, gr.br. aox., l.br.gr.	+ -	- -	- -	++ +++	- -	ND ND	ND ND
BY15 S	57.18.98 N	19.50.59 E	240 m	21.4.97	NGC	0-1 cm (d) 1-3 cm (e) 3-5 cm (f)	aox., lam. (?), re.br. aox., fluffy, gr.br. aox., lam., gr.br.	- + -	- + -	+ + +++ (c)	++ + +++	ND ND ND	1.2 1.2 2.1	3.7 1.3 94
BY15 S	57.17.00 N	20.07.98 E	241 m	22.4.98	NGC	0-1 cm 20-22 cm	aox., l.br.gr. aox., lam. (?) br. aox., l.br.gr.	- + -	- + +	- + ++	+++ +++ (h) +	- -	ND ND ND	ND ND ND
GF2 T	59.50.55 N	25.51.57 E	82 m	8.4. - 4.7.97	Eila	+ 5 m	gr.br.	+	+	-	-	+++	1.0	47
GF2 T	59.50.54 N	25.51.60 E	84 m	4.7. - 25.9.97	Eila	+ 5 m	gr.br.	++	+	-	-	+++	1.1	44
GF2 T	59.50.54 N	25.51.60 E	84 m	25.9. - 18.12.97	Eila	+ 5 m	gr.br.	+	+	-	-	++	1.0	35
GF2 T	59.50.55 N	25.51.57 E	82 m	11.4. - 9.5.97	PPS 3/3	+ 20 m	l.br.gr.	+	+	-	-	+	1.0	14
GF2 T	59.50.54 N	25.51.60 E	84 m	9.5. - 27.5.97	PPS 3/3	+ 20 m	p.br.	+	+	-	-	+	1.1	15
GF2 T	59.50.54 N	25.51.60 E	84 m	27.5. - 19.6.97	PPS 3/3	+ 20 m	l.br.gr.	-	-	-	-	-	0.9	32
GF2 T	59.50.54 N	25.51.60 E	84 m	6.7. - 24.7.97	PPS 3/3	+ 20 m	ND	ND	ND	ND	ND	ND	1.1	148
GF2 T	59.50.54 N	25.51.60 E	84 m	24.7. - 16.9.97	PPS 3/3	+ 20 m	d.gr.br.	-	-	-	-	+++	1.0	222
GF2 T	59.50.54 N	25.51.60 E	84 m	27.9. - 21.10.97	PPS 3/3	+ 20 m	d.gr.br.	+	+	-	-	+++	1.1	54
GF2 T	59.50.54 N	25.51.60 E	84 m	21.10. - 14.11.97	PPS 3/3	+ 20 m	d.gr.br.	-	-	-	-	+++	1.0	56
GF2 T	59.50.54 N	25.51.60 E	84 m	14.11. - 8.12.97	PPS 3/3	+ 20 m	d.gr.br.	+	+	-	-	+++	1.2	65
JML T	59.35.20 N	23.37.88 E	81 m	4.6. - 3.7.97	Eila	+ 5 m	ND	ND	ND	ND	ND	ND	0.9	37
JML T	59.35.21 N	23.37.77 E	81 m	3.7. - 25.9.97	Eila	+ 5 m	gr.br.	+	-	-	+	++	1.4	20
JML T	59.35.21 N	23.37.51 E	81 m	25.9. - 18.12.97	Eila	+ 5 m	gr.br.	+	-	-	+	++	1.0	73
JML T	59.35.20 N	23.37.88 E	81 m	5.6. - 19.6.97	PPS 3/3	+ 20 m	ND	ND	ND	ND	ND	ND	ND	ND
JML T	59.35.21 N	23.37.77 E	81 m	6.7. - 16.9.97	PPS 3/3	+ 20 m	ND	ND	ND	ND	ND	ND	1.0	19
JML T	59.35.21 N	23.37.51 E	81 m	27.9. - 21.10.97	PPS 3/3	+ 20 m	d.gr.br.	-	-	-	-	++	1.0	12
JML T	59.35.21 N	23.37.51 E	81 m	21.10. - 14.11.97	PPS 3/3	+ 20 m	d.gr.br.	-	-	?	+	++	0.9	7
JML T	59.35.22 N	23.37.73 E	79 m	14.11. - 8.12.97	Eila	+ 5 m	ND	ND	ND	ND	ND	ND	0.9	8
JML T	59.34.89 N	23.37.78 E	80 m	17.4. - 28.6.98	Eila	+ 5 m	gr.br.	+	-	+	+	+++	ND	ND
JML T	59.34.89 N	23.37.78 E	80 m	2.7. - 19.10.98	Eila	+ 5 m	gr.br.	+	-	-	-	++	ND	ND
JML T	59.35.22 N	23.37.73 E	79 m	17.4. - 11.5.98	PPS 3/3	+ 20 m	p.br.	+	-	-	++	-	ND	ND
JML T	59.34.89 N	23.37.78 E	80 m	2.7. - 12.9.98	PPS 3/3	+ 20 m	br.	+	-	-	+	+	ND	ND
JML T	59.34.89 N	23.37.78 E	80 m	12.9. - 18.10.98	PPS 3/3	+ 20 m	gr.br.	-	-	-	-	+	ND	ND

station	latitude	longitude	wdepth	date	sampler	sdepth	description	clasts	amph.	pyrite	carb.	FMC	Fe (EF)	Mn (EF)
LL17 T	59.02.16 N	21.05.99 E	215 m	11.4. - 2.7.97	Eila	+ 5 m	v.d.gr.br.	-	-	-	+	++++ (i)	1.2	726
	59.02.14 N	21.06.08 E	223 m	2.7. - 25.9.97	Eila	+ 5 m	v.d.br.	-	-	-	-	++++	2.5	2176
LL17 T	59.02.16 N	21.05.99 E	215 m	14.4. - 28.4.97	PPS 3/3	+ 30 m	v.d.gr.br.	+	-	-	++	++++	0.9	146
				28.4. - 15.5.97			v.d.gr.br.	-	-	-	+	++++	0.9	337
				15.5. - 19.6.97			v.d.br.	-	-	-	-	++++	1.3	1279
LL17 T	59.02.14 N	21.06.08 E	223 m	6.7. - 24.7.97	PPS 3/3	+ 30 m	v.d.br.	-	-	-	-	++++	4.2	5699
				24.7. - 11.8.97			v.d.br.	-	-	-	-	++++	2.4	3607
				11.8. - 16.9.97			v.d.br.	-	-	-	+	++++	2.5	2425
LL17 T	59.01.98 N	21.04.14 E	158 m	21.4. - 28.6.98	Eila	+ 5 m	v.d.gr.br.	+	-	-	+	++++	ND	ND
	59.02.02 N	21.05.02 E	173 m	2.7. - 20.10.98	Eila	+ 5 m	v.d.gr.br.	-	-	-	+	++++	ND	ND
LL17 T	59.01.98 N	21.04.14 E	158 m	21.4. - 17.5.98	PPS 3/3	+ 30 m	ye.br.	+	-	-	-	++	ND	ND
				17.5. - 28.6.98			d.br.	-	-	-	-	++++	ND	ND
	59.02.02 N	21.05.02 E	173 m	2.7. - 12.9.98	PPS 3/3	+ 30 m	v.d.br.	-	-	-	+	++++	ND	ND
				12.9. - 18.10.98			d.br.	-	-	-	+	+++	ND	ND
NCB T	58.50.22 N	20.14.50 E	175 m	22.4. - 28.6.98	Eila	+ 5 m	v.d.br.	-	-	-	-	++++	ND	ND
	58.49.13 N	20.15.20 E	175 m	4.7. - 20.10.98	Eila	+ 5 m	v.d.br.	-	-	-	-	+++	ND	ND
NCB T	58.50.22 N	20.14.50 E	175 m	22.4. - 8.5.98	PPS 5/2	+ 30 m	d.gr.br.	-	-	-	-	+++	ND	ND
				8.5. - 26.5.98			d.gr.br.	-	-	-	-	+++	ND	ND
				26.5. - 13.6.98			v.d.br.	-	-	-	-	++++	ND	ND
				13.6. - 28.6.98			v.d.br.	-	-	-	-	++++	ND	ND
NCB T	58.49.13 N	20.15.20 E	175 m	4.7. - 24.7.98	PPS 5/2	+ 30 m	v.d.br.	-	-	-	+	++++ (j)	ND	ND
Fårö T	58.05.53 N	19.50.58 E	182 m	14.4. - 1.7.97	Eila	+ 5 m	v.d.gr.br.	-	-	-	+	++++ (k)	1.4	553
	58.05.56 N	19.50.61 E	192 m	1.7. - 24.9.97	Eila	+ 5 m	br.	-	-	-	-	-	1.6	1.5
Fårö T	58.05.53 N	19.50.58 E	182 m	17.4. - 9.5.97	PPS 5/2	+ 40 m	v.d.gr.br.	-	-	-	+	++	0.9	394
				9.5. - 26.5.97			gr.br.	-	-	-	+	+	1.0	207
				26.5. - 19.6.97			v.d.gr.br.	-	-	-	-	++++	1.3	531
Fårö T	58.05.56 N	19.50.61 E	192 m	3.7. - 15.7.97	PPS 5/2	+ 40 m	v.d.br.	-	-	-	-	++++	1.6	480
				15.7. - 24.7.97			v.d.br.	-	-	-	-	++++	1.8	755
				24.7. - 11.8.97			v.d.br.	-	-	-	-	++++	1.7	670
				11.8. - 26.8.97			v.d.br.	-	-	-	-	++++	2.1	791
				26.8. - 7.9.97			v.d.br.	-	-	-	-	++++	2.0	310
				7.9.97 - 13.9.97			p.br.	-	-	-	-	-	1.1	21
BY15 T	57.17.02 N	20.13.47 E	231 m	21.4. - 29.6.97	Eila	+ 5 m	gr.br.	-	-	-	-	+++	2.4	170
BY15 T	57.17.02 N	20.13.47 E	231 m	23.4. - 9.5.97	PPS 5/2	+ 40 m	gr.br.	+	-	?	++	++	1.4	67
				9.5. - 5.6.97			v.d.gr.br.	-	-	?	+	+++	1.9	291
				5.6. - 19.6.97			d.br.	-	-	-	+	+++	1.7	144
BY15 T	57.17.04 N	20.13.51 E	237 m	23.4. - 28.6.98	Eila	+ 5 m	d.br.	-	-	?	+	++++	ND	ND
	57.17.02 N	20.13.65 E	237 m	1.7. - 21.10.98	Eila	+ 5 m	br.	-	-	-	-	-	ND	ND
BY15 T	57.17.04 N	20.13.51 E	237 m	23.4. - 11.5.98	PPS 3/3	+ 40 m	v.d.br.	-	-	-	+	++++	ND	ND
				11.5. - 28.6.98			v.d.gr.br.	-	-	-	-	++++	ND	ND
	57.17.02 N	20.13.65 E	237 m	4.7. - 12.9.98	PPS 3/3	+ 40 m	ye.br.	-	-	-	-	-	ND	ND
				12.9. - 18.10.98			br.	-	-	-	-	-	ND	ND

LONG-TERM CHANGES OF ZOOBENTHOS BIOMASS IN THE BARENTS SEA

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INTRODUCTION

Long-term variability of marine ecosystems has recently become one of the most topical hydrobiological study subjects. Such an interest in the problem can be explained by the lack of strict interpretation of global climatic fluctuations and other related phenomena. It is still not clear what the causes of the local and global ecosystem fluctuations are and whether the influence of the natural factors on biota prevails over the influence of the human factors.

Marine macrozoobenthos indicates and integrates long-term environmental changes very well. Most benthic organisms are slow moving and their life cycles are long in comparison to e.g. pelagic organisms. The temperature dependence of the Barents Sea zoobenthos was shown by many researches (Galkin 1998), but human influence on bottom fauna was considered only as a hypothesis (Denisenko & Denisenko 1998).

The aim of our study was to reveal the long-term temperature and fishery dependence of zoobenthos biomass in the Barents Sea.

MATERIALS AND METHODS

Zoobenthos. A lot of local studies were carried out in the Barents Sea, but only two extensive quantitative zoobenthos surveys have taken place there (in 1926-1935 and 1968-1970). The sampling networks and the schedules of these extensive surveys were not identical (Fig. 1A, 1B). The most extensive sampling program was performed in 1968-1970. However, the data on the southern and central parts of the Barents Sea for the year 1968 are lacking because the materials of the 1st cruise of R/V "Maslov" (1968) have been lost. The sampling network in the period 1924-1932 was very irregular. Most of the sampling in the open sea area was carried out in 1927-1932. In the southeastern part of the Barents Sea, the Pechora Sea, 4 detailed benthic surveys have been carried out during the last 60 years.

For our research we used all the material, which had been kept in archives and were accessible.

The biomass mapping and the computation of average biomass values for the whole Barents Sea in different periods of benthic surveys were based on the estimated values. Estimated biomass values have been calculated by Kriging method (Cressie 1991) for the fixed vertexes of interpolation grid on the basis of the real biomass values and geographical positions of corresponding stations.

Temperature regime. It is generally accepted, that observations in the 0-200 m layer of the Kola transect reflects the interannual changes in the heat content of the Coastal, Murman and Central Branch of the Nordcap currents and characterise the thermal regime of the whole Barents Sea. The Kola meridians transect has been studied for 98 years starting in 1900. For our study we used all this series of observations.

Bottom trawling. More than 600 monthly fish-trawling maps for the Barents Sea in 1946-1997, kindly placed at our disposal by "Sevrybpromrazvedka" (Murmansk Service of Fish Stocks Searching), were digitised and processed by computer. For the period before WWII trawling activity was quite precisely restored on the basis of literature sources (Shorygin 1933, Maslov 1939, 1944).

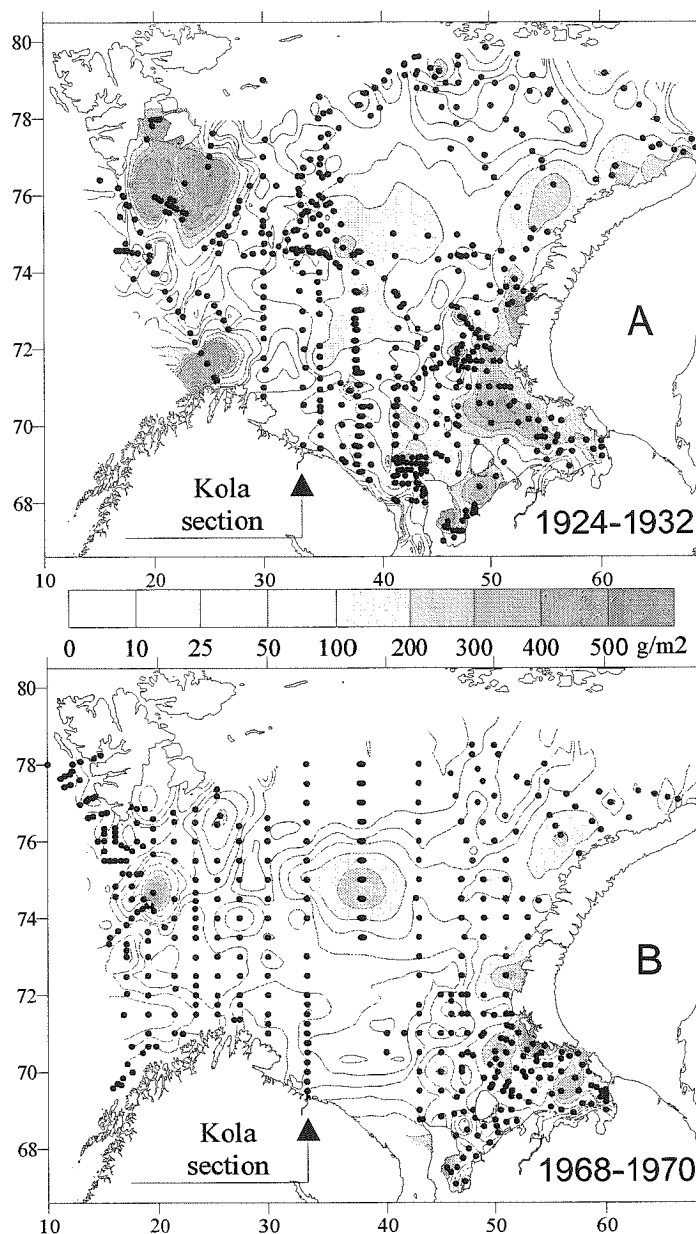


Fig. 1. Station network of quantitative zoobenthos sampling and distribution of zoobenthos biomass (g/m^2 of wet weight) in 1924-1932 (A) and 1968-1970 (B).

RESULTS AND DISCUSSION

Previous analysis of species composition and spatial distribution of zoobenthos communities during the different research periods along the Kola transect has not shown any clear connection of biomass changes with the temperature variations. The values of zoobenthos biomass in the coastal areas (up to 72° N) were approximately the same during all periods, except the late 1940s, when the biomass reached its maximum values along the whole transect. Maximal range of the values variations was registered in the late 1960s (Fig. 2). In general, the similarity of biomass distribution near the shore was the greatest in the middle 1930s and 1990s and in the offshore area of transect – in the late 1940s and in the middle 1990s.

In the Pechora Sea area the calculated average zoobenthos biomass during 1924, 1959 and 1992-1993 was 251, 220 and 258 g/m^2 respectively. After strong lowering of the temperature in 1969, the value of biomass was by one third lower (176 g/m^2) than in other three periods. This fact can be completely

explained by primary production variations in the Barents Sea caused by temperature changes in the same periods (Slagstad & Stokke 1994).

The main conclusion, made by Antipova (1975) on the basis of the results analysis of two extensive benthic surveys, was that the decrease of zoobenthos biomass in the Barents Sea in 1968-1970 in comparison with 1924-1932, was probably caused by the previous cold period. Our attempts to compare an average zoobenthos biomass for the whole Barents Sea in 1926-1935 and in 1968-1970 have shown that the simple comparison of statistically average values is not correct, because the spatial distribution of stations at the first period was far from regular.

The results of analysis based on the estimated values have shown that the decrease of zoobenthos biomass in the whole Barents Sea in the late 1960s was about 60% in contrast to 20%, shown by Antipova (1975). The variability in the primary production values during the warm and cold years in the Barents Sea (Slagstad & Stokke 1994) explains the changes in zoobenthos biomass only by 30% and only in the unlikely cases of food deficiency for the animals. Differences in sampling methods, according to our calculations, explain in addition not more than 4-5% of the biomass variation.

What was the actual reason of the notable decrease of zoobenthos biomass after the late 1960s? Earlier the bottom trawling activity in different fishing areas of the Barents Sea was mapped (Denisenko & Denisenko 1991). The similarity of this activity and the decline of zoobenthos biomass, considered in the present paper (Fig. 3A, 3B), gave an idea of possible reasons of this phenomenon.

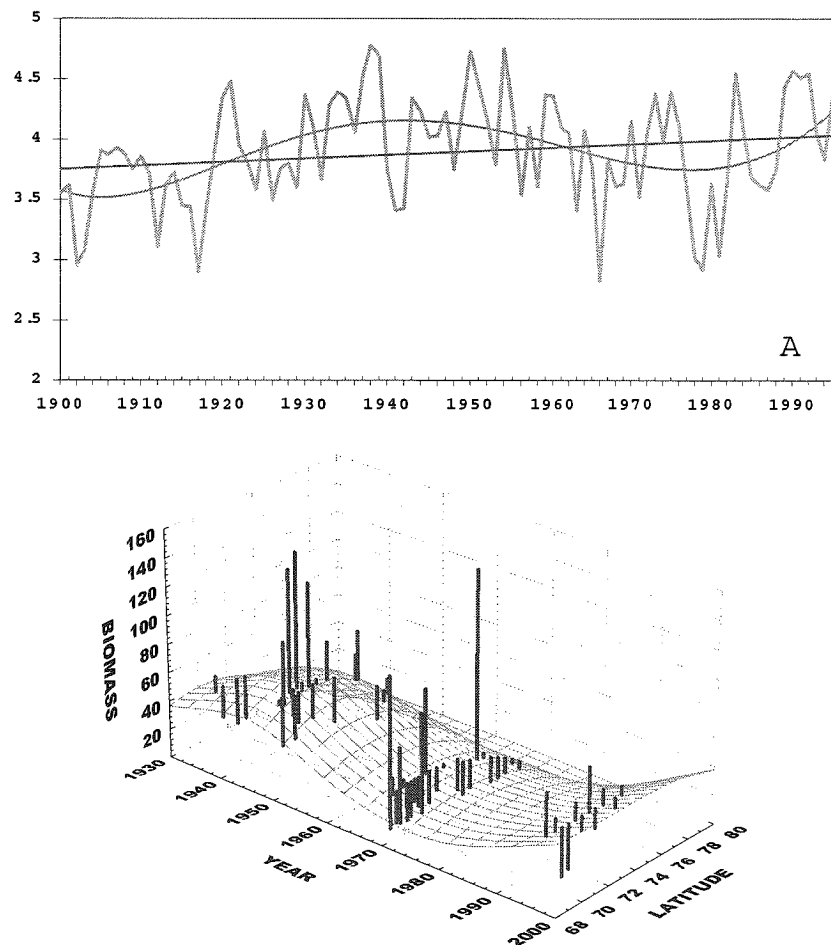


Fig. 2. Average annual water temperature, its linear and spline trends in the layer 0-200 m along the Kola transect (A); trend surface (spline smooth) and absolute values deviations of zoobenthos biomass (g/m^2 of wet weight) at the sampling stations along the Kola transect in different years (B).

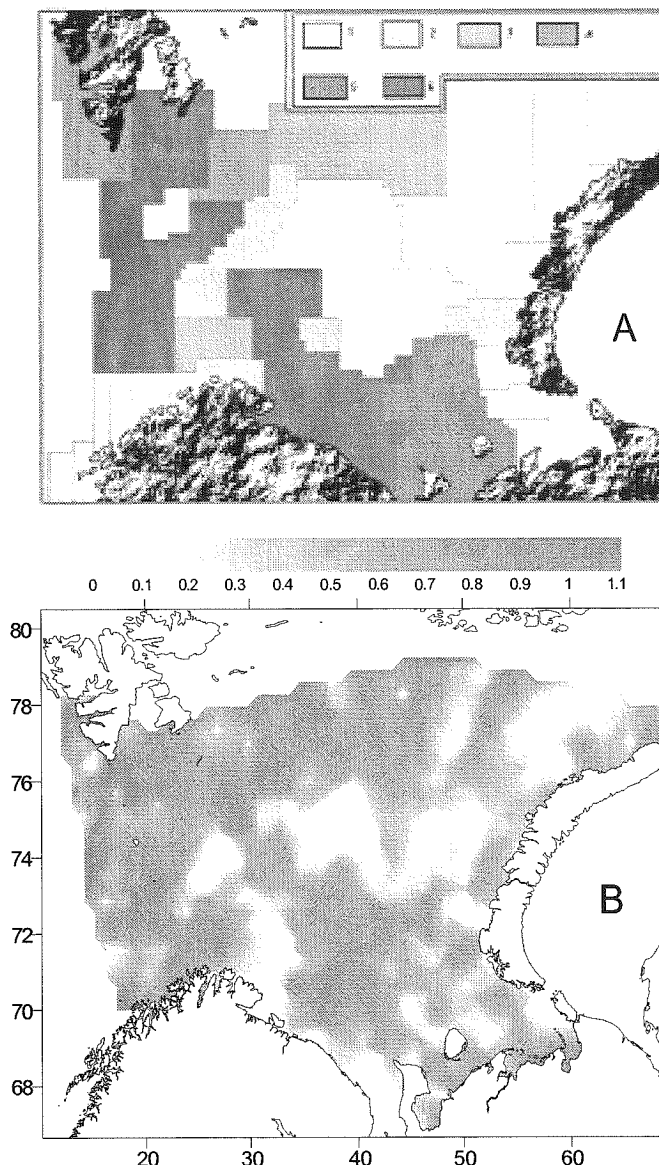


Fig. 3. Number of demersal fishery years in different regions of the Barents Sea in 1956-1985 (A): 1 – 5, 2 – 6-10, 3 – 11-15, 4 – 16-20, 5 – 21-25, 6 – 26-30; decrease (B) of zoobenthos biomass (1 unit = $1 - b_2/b_1$) in 1968-1970 (b_2) in comparison to 1926-1932 (b_1).

Data available on trawling activity and zoobenthos biomass values disagree in time and space considerably. This restricted, at the present research stage, the analysed area of biomass change to the area of the Kola transect. The results of the analysis performed are shown in Figure 4 and allow formulating the following conclusions. The high correlation ($r^2 = 0.987$) between biomass values (with a 4-year delay) and intensity of trawling activity along the Kola meridian specifies that this relation can be considered as functional. The calculation of non-linear regression (negative exponent) for the same data gives lower correlation ($r^2 = 0.929$), but this form of the regression seems more natural and appropriate to reality.

The plotted points of data (Fig. 4B) show that an increase of bottom trawling intensity is accompanied by the decline of zoobenthos biomass almost by 70 %. That estimation is close to that received via comparison of maps illustrating the distribution of zoobenthos biomass in the Barents Sea in 1926-1932 and in 1968-1970.

The obtained results lead to the obvious conclusion that the fish trawling is one of the main causal factors of long-term fluctuations of bottom communities in the Barents Sea. The absence of any significant fishery in the Barents Sea during WWII (Fig. 4A) explains the high values of zoobenthos

biomass along the Kola section at the end of the 1940s, when the trawling activity only began its regaining.

The four-year delay in reaction of zoobenthos to the trawling intensity is probably caused by the average life expectancy of bottom organisms. Apparently, it is correct, as P/B (production/biomass) ratio for the whole zoobenthos in the Barents Sea varies within 0,2-0,3 (Brotskaya & Zenkevich 1939, Zenkevich 1963). The basic role thus is played not by the destruction of the adult animals, but by destruction of their potential posterity.

These conclusions make doubtful the opinion of some researchers, who interpret moderate bottom trawling as a stimulus for zoobenthos development or as a factor, negative influence of which is short and comparable to seasonal environmental influence (Rijnsdorp & Leeuwen 1996, Kaiser & al. 1998, Hansson & al. 1998, etc.). This may be true, but only for cases of very low trawling activity, which is not common in traditional fishing areas. Therefore, we share the viewpoint of other researchers, who consider bottom trawling as one of the major impacts on benthic communities in the areas of intensive fishery (e.g. Bergman & Hup 1992, Philippart 1998).

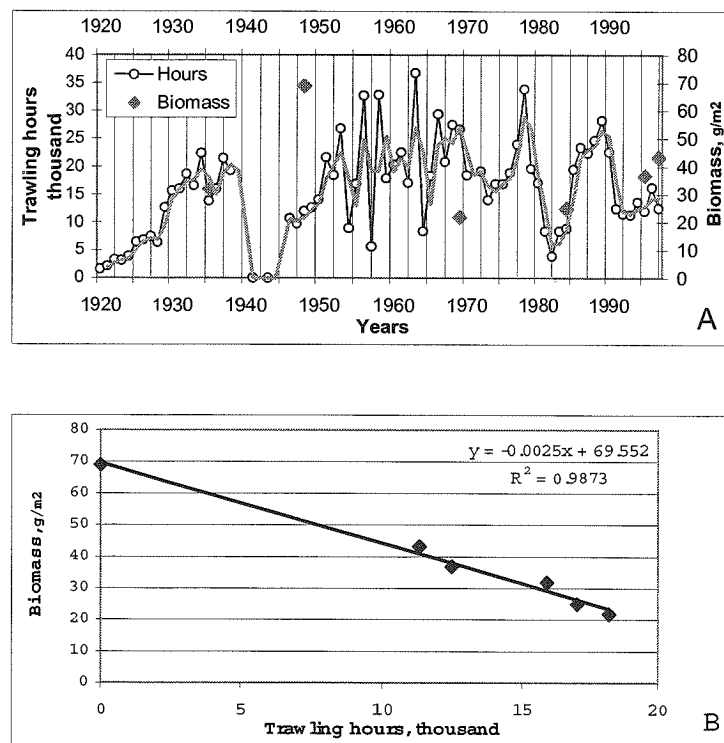


Fig. 4. Annual trawling activity (left axis, hours, solid line – 3-year running average) and zoobenthos biomass (right axis, g/m^2) along the Kola transect during 1920-1997 (A); opposite relation between biomass (4-year delay) and trawling activity (B).

Taking into account the obtained results, it is necessary to point out that more comprehensive programs of marine zoobenthos monitoring, than those now existing, should be done. This is important for recognizing possible influence of fishery on the long-term fluctuations and natural successions of bottom communities.

The number of the Barents Sea regions not used in demersal fishing is relatively low. One of them is the Pechora Sea, where mainly scientific bottom trawling and short-term fishing of the polar cod by the bottom trawls and floating trawls have taken place. The benthic studies carried out there earlier could provide a sound basis for further monitoring of the bottom fauna. Together with the observations along the Kola transect this would be helpful for revealing not only climatic reasons, but also anthropogenic factor responsible for the long-term zoobenthos fluctuations in the Barents Sea.

Acknowledgements

The author thanks the Academy of Finland for financing the project and the Finnish Institute of Marine Research for the possibility to work at the Institute. The author also wants to thank the specialists of the Institute H. Sandler, A.-B. Andersin, L. Grönlund, H. Kuosa and A. Laine for organisational help and consultations.

Учет, подсчет

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No. 45

Simple sediment model: a tool to estimate nutrient regeneration in Baltic Sea sediments
Distribution of sediments and suspended matter in the Gulf of Finland – the central Baltic Proper transect
with a particular emphasis of transportation: minerogenic evidence
Long-term changes of zoobenthos biomass in the Barents Sea

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ISSN 1238-5328