Vertical water circulation in the southern Baltic and its environmental implications

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Southern Baltic
Gdańsk and Bornholm Deeps
Vertical water circulation
Environmental consequences

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Abstract

A simple method of estimating the vertical current vector velocity component w based on the water shear stress vorticity equation is presented and briefly discussed. Spatial distributions of the mean climatic values of w for selected months calculated from the averaged multi-year observations of atmospheric pressure and water density (diagnostic approach) in the southern Baltic Sea are investigated. The results show the existence of 'permanent' (during the year) zones of upwelling and downwelling in the region of the Baltic Deeps related to the clockwise and anticlockwise gyres observed on the charts of calculated horizontal currents. These characteristic features of the horizontal and vertical water movements in the region of the Gdańsk and Bornholm Deeps may be important for the ecological, biological and geological understanding of the region as well as in the context of the pollution problem, especially during stagnation periods.

1. Introduction

The paper deals with a linear model of estimating the vertical velocity component of sea currents. The main aim of this study is to complete the results of a diagnostic model of the wind- and density-driven steady water circulation in the Baltic Sea (Jankowski and Kowalik, 1980; Staśkiewicz, 1988; Kowalik and Staśkiewicz, 1976; Sarkisyan *et al.*, 1975) and to obtain a more detailed spatial representation of seawater motion. The vertical velocity component can be estimated from calculations of the horizontal currents using the continuity equation. This estimate, however, is non-unique, owing to differences in the orders of magnitude of the terms in the continuity equation and to truncation errors in numerical models of water circulation (Sarkisyan *et al.*, 1986; Semenov, 1981). In this study another method was used, based on the solution of the equation for the z-th vorticity component of the water shear stress and is modified version of the method proposed in Jankowski (1984).

Only a few estimates of the vertical velocity under Baltic Sea conditions have been made (Staśkiewicz, 1974, *cit.* after Jankowski (1988); Kuznetsova and Tyuriakov, 1981), and in both studies the standard method of calculating w on the basis of the continuity equation was applied. It is also very difficult to find the values of w directly from *in situ* measurements because of its order of magnitude ($\pm 10^{-4}$ cm s^{-1}). Hitherto, numerical modelling has been the simplest way of obtaining an adequate estimate of w.

A recognition of the vertical water circulation is interesting not only from the point of view of seawater dynamics in order to obtain a complete picture of the spatial current structure, but also for an understanding of the variability in the ecological parameters. The water circulation in the Baltic Sea, which is controlled by the complex coast lines and bottom relief and driven by the wind and the density gradient (thermohaline) forcing, may have some influence on the distributions of ecological, biological or pollution fields. It is believed that the results of the diagnostic model calculation of w may be useful, at least to explain the occurrence in the Baltic Deeps of the so-called 'dead' zones with low concentrations of oxygen (*cf.* Melvasalo *et al.*, 1981; Andersin and Sandler, 1988; Trzosińska, 1994).

2. Methods

The method of estimating the vertical current velocity component proposed in this paper, based on the solution of the equation for the z-th vorticity component of the water shear stress, is a modified version of the model presented in Jankowski (1984). The general form of the vorticity equation depends on the set of equations of the diagnostic model of steady wind- and density-driven water circulation, *i.e.* water movements due to the climatic fields of forcing (multi-year averaged fields of wind and water density).

2.1. Equations and boundary conditions

Thus the basic set of equation and boundary conditions of the diagnostic model can be written in the following form (Ramming and Kowalik, 1980; Jankowski, 1988):

$$\underline{\frac{\partial u}{\partial t}} - fv = -\frac{1}{\rho_0}\frac{\partial p_a}{\partial x} - g\frac{\partial \zeta}{\partial x} - \frac{g}{\rho_0}\frac{\partial}{\partial x}\int_0^z \rho d\eta + \frac{1}{\rho_0}\frac{\partial \tau_x}{\partial z}, \tag{1}$$

$$\underline{\frac{\partial v}{\partial t}} + fu = -\frac{1}{\rho_0}\frac{\partial p_a}{\partial y} - g\frac{\partial \zeta}{\partial y} - \frac{g}{\rho_0}\frac{\partial}{\partial y}\int_0^z \rho d\eta + \frac{1}{\rho_0}\frac{\partial \tau_y}{\partial z}, \tag{2}$$

the continuity equation:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \tag{3}$$

the boundary conditions at the free sea surface ($z = -\zeta$):

$$\tau_x = \rho_0 A_z \frac{\partial u}{\partial z} = -\tau_x^s, \qquad \tau_y = \rho_0 A_z \frac{\partial v}{\partial z} = -\tau_y^s, \tag{4}$$

$$w_\zeta = -\left(\frac{\partial \zeta}{\underline{\partial t}} + u_\zeta \frac{\partial \zeta}{\partial x} + v_\zeta \frac{\partial \zeta}{\partial y} \right), \tag{5}$$

the boundary conditions at the sea bottom ($z = H$):
for the non-slip condition:

$$u_H = 0, \quad v_H = 0, \quad w_H = 0, \tag{6}$$

or for the slip condition with bottom friction:

$$\tau_x = \rho_0 A_z \frac{\partial u}{\partial z} = -\tau_x^H, \qquad \tau_y = \rho_0 A_z \frac{\partial v}{\partial z} = -\tau_y^H, \tag{7}$$

$$w_H = u_H \frac{\partial H}{\partial x} + v_H \frac{\partial H}{\partial y}, \tag{8}$$

where

u, v, w – components of the current velocity vector along the the axes of the Cartesian system of coordinates, the origin of which is located on the free sea surface, the x-axis being oriented to the East, the y-axis to the North and z-axis vertically downwards,

$f = 2\,\omega \sin \varphi$ – Coriolis parameter ($\omega = 0.729 \times 10^{-5}$ rad s^{-1} – angular velocity of the Earth's rotation about its own axis; φ – latitude),

ρ_0, ρ – the mean seawater density and the deviation of seawater density from its mean value ($\rho \ll \rho_0$) respectively,

A_z – vertical eddy viscosity vertical exchange coefficient (assumed constant with z-axis),

τ_x, τ_y – water shear stress components,

τ_x^s, τ_y^s – wind stress components at the free sea surface,

τ_x^H, τ_y^H – bottom stress components,

u_ζ, v_ζ, w_ζ – components of the current velocity at the free sea surface,

u_H, v_H, w_H – components of the current velocity at the sea bottom,

ζ, H – sea level and the sea bottom depth respectively,

g – acceleration due to gravity,

p – hydrostatic pressure,

p_a – atmospheric pressure at the free sea surface.

The terms underlined in eqs. (1), (2) and (5) are omitted in calculations of the steady–state components of the current velocity vector. Eqs. (1)–(3) and (5) in their full form are applied to calculate mass transport components and sea level.

By integrating eqs. (1)–(2) and continuity eq. (3) with respect to z from the free sea surface $z = -\zeta \approx 0$ to the bottom $z = H$, taking into account boundary conditions (4), (5), (7) and (8), and assuming that $H \gg \zeta$ and linear bottom friction in eq. (7), we derive the equations of mass transport and sea level (Ramming and Kowalik, 1980; Jankowski, 1988):

$$\frac{\partial M_x}{\partial t} - fM_y = -H\frac{\partial p_a}{\partial x} - \rho_0 gH\frac{\partial \zeta}{\partial x} - g\int\limits_{-\zeta}^{H}\frac{\partial}{\partial x}\int\limits_{-\zeta}^{z}\rho d\eta dz + \tau_x^s - RM_x, \tag{9}$$

$$\frac{\partial M_y}{\partial t} + fM_x = -H\frac{\partial p_a}{\partial y} - \rho_0 gH\frac{\partial \zeta}{\partial y} - g\int\limits_{-\zeta}^{H}\frac{\partial}{\partial y}\int\limits_{-\zeta}^{z}\rho d\eta dz + \tau_y^s - RM_y, \tag{10}$$

$$\frac{\partial \zeta}{\partial t} + \frac{1}{\rho_0}\left(\frac{\partial M_x}{\partial x} + \frac{\partial M_y}{\partial y}\right) = 0, \tag{11}$$

where

$M_x = \int\limits_{-\zeta}^{H} \rho_0 u dz$, $M_y = \int\limits_{-\zeta}^{H} \rho_0 v dz$ – respective mass transport components along the x-axis and y-axis,

$R = \frac{\pi A_z}{4H^2}$ – bottom friction coefficient.

The relevant lateral boundary and initial conditions for the mass transport and sea level eqs. (9)–(11) can be written in the form (Ramming and Kowalik, 1980; Jankowski, 1988):

$$M_n = \begin{cases} 0 & \text{at the closed (solid) boundary,} \\ \Phi(L) & \text{at the open (liquid) boundary,} \end{cases} \tag{12}$$

for $\quad t = 0, \quad \zeta = 0, \quad M_x = 0, \quad M_y = 0,$ (13)

where

M_n – mass transport component normal to the shoreline L,

$\Phi(L)$ – water balance at the open boundary (water mass exchange with adjacent basin or river water inflows).

Marchuk and Kagan (1977) proved that the set of eqs. (9)–(11) subject to the boundary (12) and initial (13) conditions possesses a unique solution; therefore, after the fields of the external forces have been computed from the stationary set of data, the set is integrated through time until a steady-state occurs.

The finite difference scheme used by Hansen (1962); Ramming and Kowalik (1980) was applied to the system of eqs. (9)–(11), and the numerical calculations were carried out with a time step of 120 seconds and respective space steps of $\delta\varphi = 10'$ (about ten nautical miles) and $\delta\lambda = 15'$.

Calculated in such a manner, the mass transport components M_x, M_y and sea level variations ζ are subsequently inserted into the steady-state form of the equations and boundary conditions (1)–(8) in order to derive the components of current velocity u, v, w.

In the case of a vertical eddy viscosity coefficient A_z independent of z, it is easy to obtain an analytical solution for horizontal current velocity components u, v from the steady-state form of the set of eqs. (1), (2) and boundary conditions (4), (6) (Jankowski, 1988; Ramming and Kowalik, 1980).

Assuming that the atmospheric pressure, sea level and wind stress are known, eqs. (1), (2) resolve into a second-order differential equation for the complex velocity $D = u + iv$ (where $i = \sqrt{(-1)}$) (Jankowski, 1988; Ramming and Kowalik, 1980):

$$\frac{\partial^2 D}{\partial z^2} - p_1^2 D = \frac{1}{\rho_0 A_z}\left(\frac{\partial p_a}{\partial x} + i\frac{\partial p_a}{\partial y}\right) + \frac{g}{A_z}\left(\frac{\partial \zeta}{\partial x} + i\frac{\partial \zeta}{\partial y}\right) + G(z), \quad (14)$$

where

$$D = u + iv, \qquad p_1 = \sqrt{\frac{if}{A_z}},$$

$$G(z) = \frac{g}{\rho_0 A_z}\left(\frac{\partial}{\partial x}\int_0^z \rho d\eta + i\frac{\partial}{\partial y}\int_0^z \rho d\eta\right). \quad (15)$$

In complex notation the boundary conditions (4), (6) take the form:

for $\quad z = -\zeta; \qquad \rho_0 A_z \frac{\partial D}{\partial z} = -\tau^s = -(\tau_x^s + i\tau_y^s), \quad (16)$

for $\quad z = H; \qquad D = u + iv = 0 + i0. \quad (17)$

The solution of the above eq. (14) with boundary conditions (16), (17) can be written in the form (Ramming and Kowalik, 1980; Kowalik and Staśkiewicz, 1976):

$$D = \frac{\tau^s}{\rho_0 A_z p_1}\frac{\sinh[p_1(H-z)]}{\cosh(p_1 H)} + \frac{g}{if}\left(\frac{\partial \zeta}{\partial x} + i\frac{\partial \zeta}{\partial y}\right)\left(\frac{\cosh(p_1 z)}{\cosh(p_1 H)} - 1\right) + $$
$$+ F_1(p_a, \rho), \quad (18)$$

where

$$
\begin{aligned}
F_1(p_a,\rho) &= -[B_1(-H)e^{-p_1H}+B_2(-H)e^{p_1H}]\frac{\cosh(p_1z)}{\cosh(p_1H)}+\\
&+ B_1(z)e^{p_1z}+B_2(z)e^{-p_1z}+\frac{1}{\rho_0 if}\times\\
&\times \left(\frac{\partial p_a}{\partial x}+i\frac{\partial p_a}{\partial y}\right)\left(\frac{\cosh(p_1z)}{\cosh(p_1H)}-1\right),
\end{aligned}
\tag{19}
$$

$$
B_1(z)=\frac{1}{p_1}\int_0^z G(\eta)e^{-p_1\eta}d\eta,\qquad B_2(z)=-\frac{1}{p_1}\int_0^z G(\eta)e^{p_1\eta}d\eta,
$$

$$
\tau^s=\tau_x^s+i\tau_y^s. \tag{20}
$$

2.2. Model for estimating the vertical velocity component

Knowing the horizontal components of the current velocity and using continuity eq. (3), the vertical current velocity component w can be estimated.

It is rather difficult, however, to use the continuity eq. (3) directly for estimating the vertical velocity in the circulation sea models. One reason for this is that the continuity eq. (3) is a first-order differential equation and it is impossible to discover an appropriate solution fulfilling both boundary conditions, *i.e.* at the free sea surface (5) and at the bottom (6). This is necessary in order to fulfil the integral mass conservation law for stationary flows following from the continuity equation (Semenov, 1981; Sarkisyan *et al.*, 1986). This can be derived by integrating the continuity eq. (3) with respect to z from the free sea surface $z=-\zeta$ to its bottom $z=H$, and taking into consideration the boundary conditions (5) and (6) (or (8)) and the dependence of the integral limits of x and y:

$$
\begin{aligned}
w_H-w_\zeta=\frac{\partial}{\partial x}\int_{-\zeta}^{H}ud\eta &+ \frac{\partial}{\partial y}\int_{-\zeta}^{H}vd\eta+u_H\frac{\partial H}{\partial x}+v_H\frac{\partial H}{\partial y}+\\
&+ u_\zeta\frac{\partial\zeta}{\partial x}+v_\zeta\frac{\partial\zeta}{\partial y},
\end{aligned}
\tag{21}
$$

or after taking into consideration the expressions for w_ζ (5) and w_H (6) or (8) in their final form:

$$
\frac{\partial}{\partial x}\int_{-\zeta}^{H}ud\eta+\frac{\partial}{\partial y}\int_{-\zeta}^{H}vd\eta=\frac{1}{\rho_0}\left(\frac{\partial M_x}{\partial x}+\frac{\partial M_y}{\partial y}\right)=0. \tag{22}
$$

The expression obtained has to be related to the sea level eq. (11) in the steady-state form ($\partial\zeta/\partial t = 0$) in order for the vertical velocity component to be estimated correctly.

The second reason for not applying the continuity equation directly to calculate w is because the difference between the order of magnitude of the small term $\partial w/\partial z$ in comparison with the other large terms $\partial u/\partial x$ and $\partial v/\partial y$ leads to considerable errors in the calculated values of w. In order to avoid the above problems it is better to take into consideration the water shear vorticity equation instead, as the problem of estimating the vertical velocity component is formulated in atmosphere circulation models (Thompson, 1961), and to increase the order of the continuity equation with respect to z (Semenov, 1981; Sarkisyan *et al.*, 1986; Demin and Yakovlev, 1985) .

Thus in order to complete the calculation of water currents in the Baltic Sea using the diagnostic model, the modified version of a simple model for estimating the vertical current velocity component based on the equation for the z-th vorticity component of water shear stress was applied (*cf.* Jankowski, 1984, 1988). The most important steps of the method providing the basic formulae for calculating w are now described.

2.2.1. Vorticity equation

Differentiation of steady-state eq. (1) with respect to y and eq. (2) with respect to x and subtraction of eq. (1) from (2) yields

$$\beta v + f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = \frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right), \tag{23}$$

where $\beta = df/dy$ (β – plane approximation). Taking eq. (16) and continuity eq. (3) into consideration we obtain:

$$\frac{\partial w}{\partial z} = -\frac{1}{\rho_0 f}\frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right) + \frac{\beta}{f}v = -\frac{1}{\rho_0 f}\frac{\partial}{\partial z}(rot_z\vec{\tau}) + \frac{\beta}{f}v, \tag{24}$$

where $\vec{\tau}$ is the vector of the water shear stress.

In order to find a unique solution to eq. (24) we increase the order of eq. (24) (*cf. e.g.* Sarkisyan *et al.*, 1986; Demin and Yakovlev, 1985) by differentiating it side by side with respect to z

$$\begin{aligned}\frac{\partial^2 w}{\partial z^2} &= -\frac{1}{\rho_0 f}\frac{\partial^2}{\partial z^2}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right) + \frac{\beta}{f}\frac{\partial v}{\partial z} = \\ &= -\frac{1}{\rho_0 f}\frac{\partial^2}{\partial z^2}(rot_z\vec{\tau}) + \frac{\beta}{f}\frac{\partial v}{\partial z}.\end{aligned} \tag{25}$$

The solution of eq. (25) with boundary conditions at the free sea surface (5) and at the sea bottom (6) (or (8)) can be written in the form

$$
\begin{aligned}
w(x,y,z) &= -\frac{1}{\rho_0 f} rot_z \vec{\tau} + \frac{\beta}{f}\int\limits_{-\zeta}^{z} v d\eta + \frac{z+\zeta}{H+\zeta} \times \\
&\times \left(w_H - \frac{1}{\rho_0 f} rot_z \vec{\tau}^H - \frac{\beta}{f}\int\limits_{-\zeta}^{H} v d\eta \right) + \\
&+ \frac{H-z}{H+\zeta}\left(w_\zeta - \frac{1}{\rho_0 f} rot_z \vec{\tau}^s \right). \qquad (26)
\end{aligned}
$$

Furthermore, in order to calculate w with the aid of eq. (26), it is necessary to estimate the components of the water shear stress vector $\vec{\tau}$ *i.e.* τ_x, τ_y. In order to attain this, one can find the analytical form to calculate τ ($\tau = \tau_x + i\tau_y$) from the analytical solution for the horizontal current velocity (18). The complex form of the solution for water shear stress $\tau = \tau_x + i\tau_y$ takes the following form:

$$
\begin{aligned}
\tau = \rho_0 A_z \frac{\partial D}{\partial z} &= -\frac{\tau^s \cosh[p_1(H-z)]}{\cosh(p_1 H)} + \frac{\rho_0 g A_z p_1}{if} \times \\
&\times \left(\frac{\partial \zeta}{\partial x} + i\frac{\partial \zeta}{\partial y} \right) \frac{\sinh(p_1 z)}{\cosh(p_1 H)} + F_{11}(p_a, \rho), \qquad (27)
\end{aligned}
$$

where

$$
\begin{aligned}
F_{11}(p_a,\rho) &= -\rho_0 p_1 A_z [B_1(-H)e^{-p_1 H} + B_2(-H)e^{p_1 H}] \times \\
&\times \frac{\sinh(p_1 z)}{\cosh(p_1 H)} + \rho_0 p_1 A_z [B_1(z)e^{p_1 z} - B_2(z)e^{-p_1 z}] + \\
&+ \frac{p_1 A_z}{if}\left(\frac{\partial p_a}{\partial x} + i\frac{\partial p_a}{\partial y} \right) \frac{\sinh(p_1 z)}{\cosh(p_1 H)}. \qquad (28)
\end{aligned}
$$

The components of the water shear stress τ_x, τ_y and their values at the sea bottom $\tau_x^H = -\tau_x(x,y,H)$, $\tau_y^H = -\tau_y(x,y,H)$ can be estimated from the formulas

$$
\tau_x(x,y,z) = \frac{1}{2}(\tau + \tau^*), \qquad \tau_y(x,y,z) = -\frac{1}{2i}(\tau - \tau^*), \qquad (29)
$$

$$
\begin{aligned}
\tau_x^H &= -\frac{\rho_0 A_z}{2}\rho_0 \left[\frac{\partial}{\partial z}(D + D^*) \right]_{z=H}, \\
\tau_y^H &= -\frac{\rho_0 A_z}{2i} \left[\frac{\partial}{\partial z}(D - D^*) \right]_{z=H}, \qquad (30)
\end{aligned}
$$

where D^*, τ^* – complex conjugate values of the current velocity D and water shear stress τ respectively.

3. Results

The calculations were performed for wind and density fields estimated from the averaged multi-year observations of atmospheric pressure over the Baltic, and water temperature and salinity (from Lenz's (1971) and Bock's (1971) atlases) for selected months of the year.

First, the geostrophic wind was determined from the distribution of atmospheric pressure

$$W_{ags} = C_r W_{ag} = C_r B_T \frac{\partial p_a}{\partial n}, \tag{31}$$

where

W_{ags} – module of geostrophic surface wind velocity vector,
W_{ag} – module of geostrophic (gradient) wind vector velocity,
$C_r = 0.7$ – reduction factor, taking into account the weakening of wind velocity due to friction effects in the vicinity of the free sea surface,
$\frac{\partial p_a}{\partial n}$ – horizontal gradient of atmospheric pressure (in hPa/100 km),
B_T – coefficient including the stratification conditions at the sea – atmosphere interface (assumed constant and equal to 4.69, *i.e.* the mean value of data taken from Garbalewski and Malicki (1971)).

Next, the wind stress components are estimated by standard formulas taking into account the deflection angle between the geostrophic surface wind W_{ags} and the 'real' surface wind W_a *i.e.* $\alpha = 15°$ (Ramming and Kowalik, 1980; Jankowski, 1988):

$$\tau_x^s = \rho_a C_D W_a W_x, \qquad \tau_y^s = \rho_a C_D W_a W_y, \tag{32}$$

where

$W_a = \sqrt{W_x^2 + W_y^2} = W_{ags}$ – module of the 'real' surface wind velocity vector,
W_x, W_y – the 'real' surface wind velocity vector components along the x-axis and y-axis respectively,
ρ_a, C_D – air density and drag coefficient respectively.

In calculations the following standard values are used: for air density $\rho_a = 1.29 \times 10^{-3}$ g cm^{-3} and drag coefficient $C_D = 2.5 \times 10^{-3}$.

The eddy viscosity coefficient A_z was estimated by Felsenbaum's theory of wind–driven currents, in which values of A_z, dependent on the wind velocity, Coriolis parameter and basin depth can be estimated from the expressions (Ramming and Kowalik, 1980; Jankowski, 1988)

$$A_z = \alpha_1 W_a H, \text{ if } H < H_{cr} \quad \text{or} \quad A_z = \alpha_2 \frac{W_a^2}{f}, \text{ if } H \geq H_{cr}, \tag{33}$$

where

$H_{cr} = \alpha_3 W_a/f$ – critical depth,

$\alpha_1, \alpha_2, \alpha_3$ – coefficients equal to 0.54×10^{-4}, 4.7×10^{-8} and 8.7×10^{-4} respectively.

Seawater density ρ_w fields were calculated on the basis of seawater temperature and salinity from the Lenz and Bock atlases (Bock (1971); Lenz (1971)), applying the equation of state in the Mamayev form (Mamayev, 1970):

$$\rho_w = a_1 + a_2 T + a_3 T^2 + a_4 T \times S + a_5 S, \tag{34}$$

where

ρ_w, T, S – seawater density [g m^{-3}], temperature [°C] and salinity [PSU],

a_1, a_2, a_3, a_4, a_5 – constant coefficients equal to 1.000082, -3.5×10^{-6}, -4.69×10^{-6}, -2.0×10^{-6} and -8.02×10^{-4} respectively.

Numerical calculations were performed for the closed Baltic Sea on a numerical grid of modified spherical coordinats (λ, φ, z) with steps $\delta\lambda = 15'$ and $\delta\varphi = 10'$ respectively and with a constant Coriolis parameter.

The computation results are presented in the form of schematic charts of the vertical velocity components at selected depths for May, July and September, representing the conditions in spring, summer and autumn. Distributions of w along a vertical longitudinal (zonal) section complete the spatial pictures of vertical water circulation.

Fields of w at selected depths

The results of calculations of the vertical current velocity component at two depths $z = 30$ m and $z = 50$ m for May, July and September in the southern Baltic are plotted in Figs. 1 and 2.

For each month the characteristic distribution of up- and downwelling regions in the vicinity of the Gdańsk, Bornholm and Gotland Deeps can be seen both in the surface layer (Fig. 1, $z = 30$ m) and in a deeper one (Fig. 2, $z = 50$ m). The intensity of the vertical movements is lowest in July (summer conditions) and highest in September (autumn conditions), when w reaches values of the order of 5×10^{-3} cm s^{-1}. These spatial patterns are easily correlated with the clusters of gyres in the horizontal current fields, as can be seen in Fig. 3, which illustrates examples of horizontal currents at a depth of $z = 30$ m for May (for current fields in other months – see *e.g.* Jankowski and Kowalik, 1980; Staśkiewicz, 1988; Jankowski and Skałka, 1995). The upwelling (downwelling) regions in Fig. 1 and 2 are related to the anticlockwise (clockwise) gyre-like structures in the horizontal current fields (*cf.* Fig. 3).

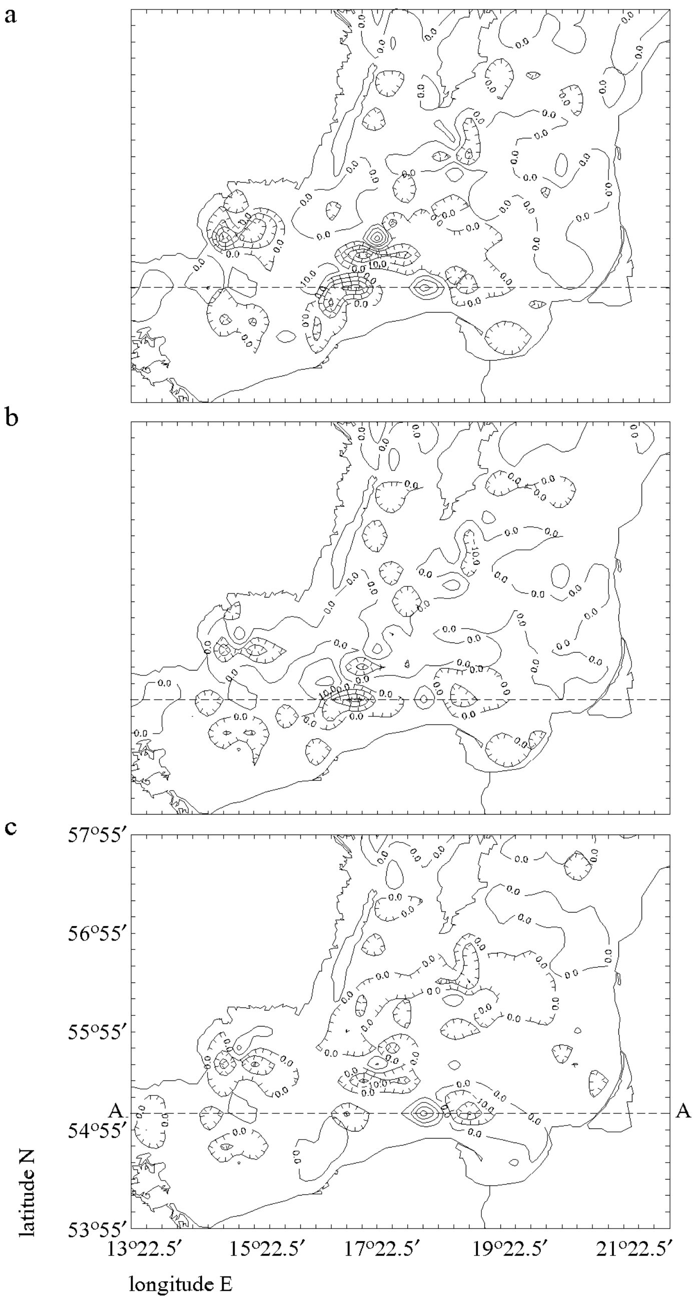

Fig. 1. Vertical current velocity component w [10^{-4} cm s^{-1}] at depth $z = 30$ m in the southern Baltic for selected months: May (a), July (b) and September (c). Line A–A indicates the location of the longitudinal (zonal) section ($\varphi = 55°05'$). Hatched areas indicate regions of upwelling (negative values of w). Contour interval $\triangle w = \pm\, 10{\times}10^{-4}$ [cm s^{-1}]

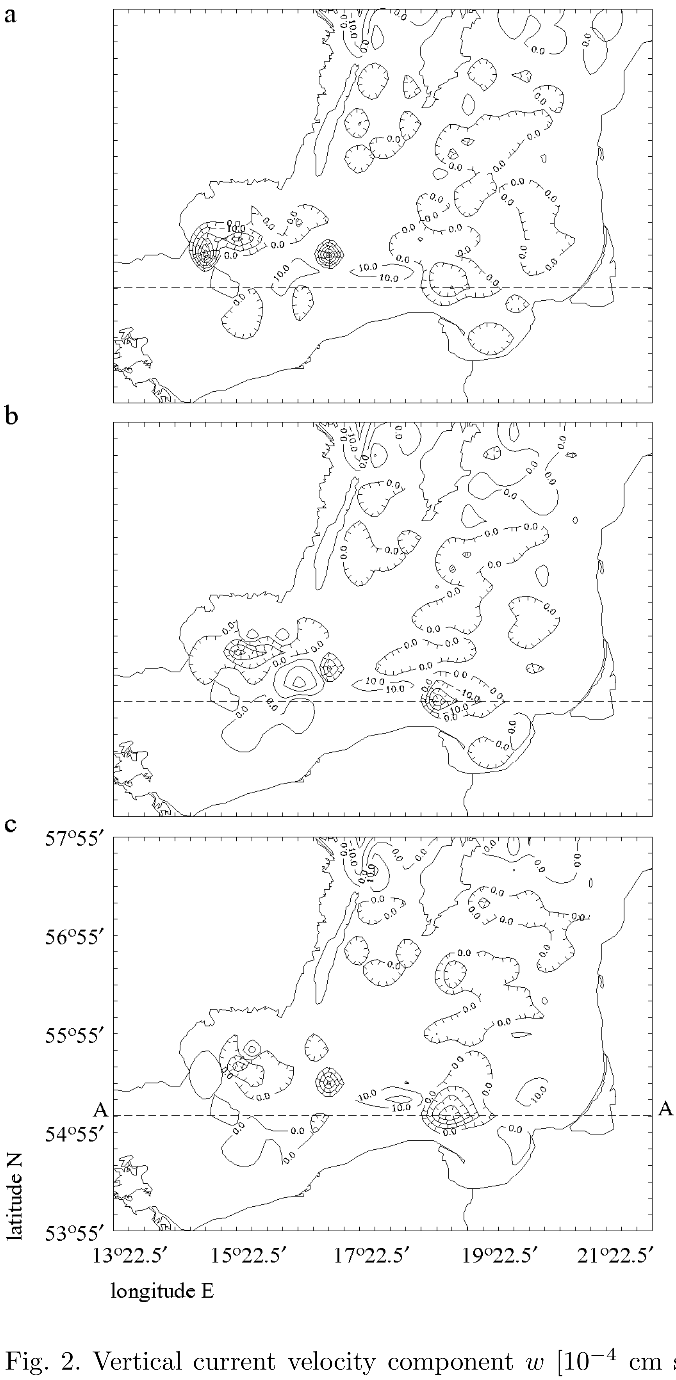

Fig. 2. Vertical current velocity component w [10^{-4} cm s^{-1}] at depth $z = 50$ m in the southern Baltic for selected months: May (a), July (b) and September (c). Line A–A indicates the location of the longitudinal (zonal) section ($\varphi = 55°05'$). Hatched areas indicate regions of upwelling (negative values of w). Contour interval $\triangle w = \pm\, 10{\times}10^{-4}$ [cm s^{-1}]

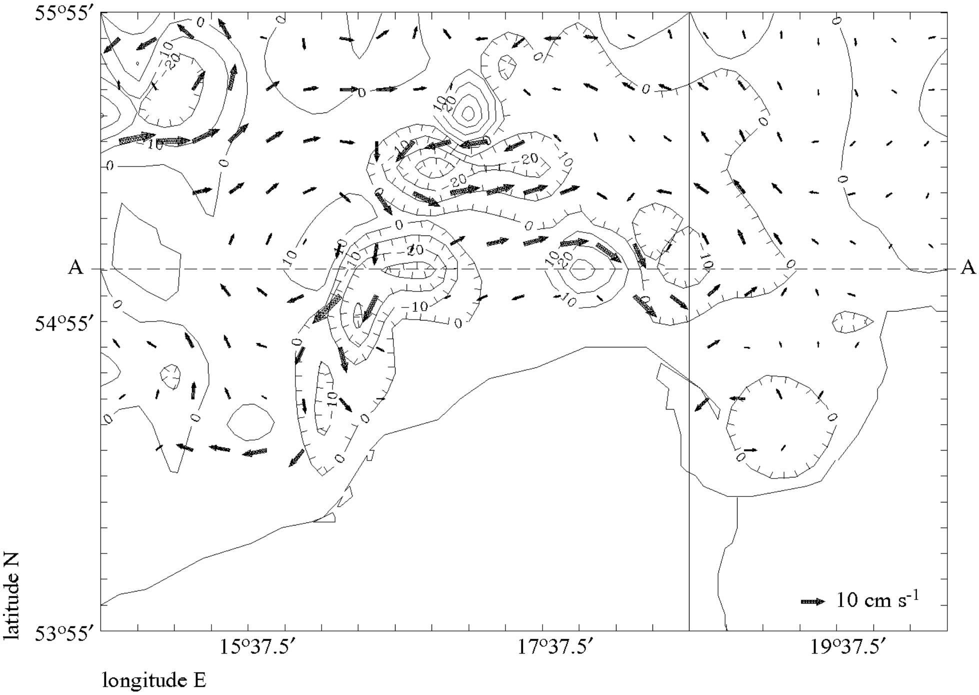

Fig. 3. Horizontal current vectors [cm s^{-1}] at depth z = 30 m and isolines of the vertical current velocity component w [10^{-4} cm s^{-1}] in the vicinity of the Gdańsk and Bornholm Deeps in May. Line A–A indicates the location of the longitudinal (zonal) section ($\varphi = 55°05'$). Hatched areas indicate regions of upwelling (negative values of w). Contour interval $\triangle w = \pm\, 10 \times 10^{-4}$ [cm s^{-1}]

Longitudinal vertical section of w

In order to complete the spatial structure of the vertical water circulation the distributions of w along a longitudinal vertical section ($\varphi = 55°05'$) from Bornholm to the east coast of the Baltic for May, July and September were plotted in Fig. 4 (the location of the section is shown in Figs. 1 and 2 – line A–A).

The occurrence of an area of upwelling in the vicinity of the Gdańsk Deep (Fig. 4, λ = 18–19.5°E) during each of the months of the year is one of the characteristics of the local water dynamics controlled by the bottom topography and the hydrological conditions of the region.

It is worth noting the occurrence of 'lenses' with up- and downwelling currents observed in the water layers close to the sea bottom.

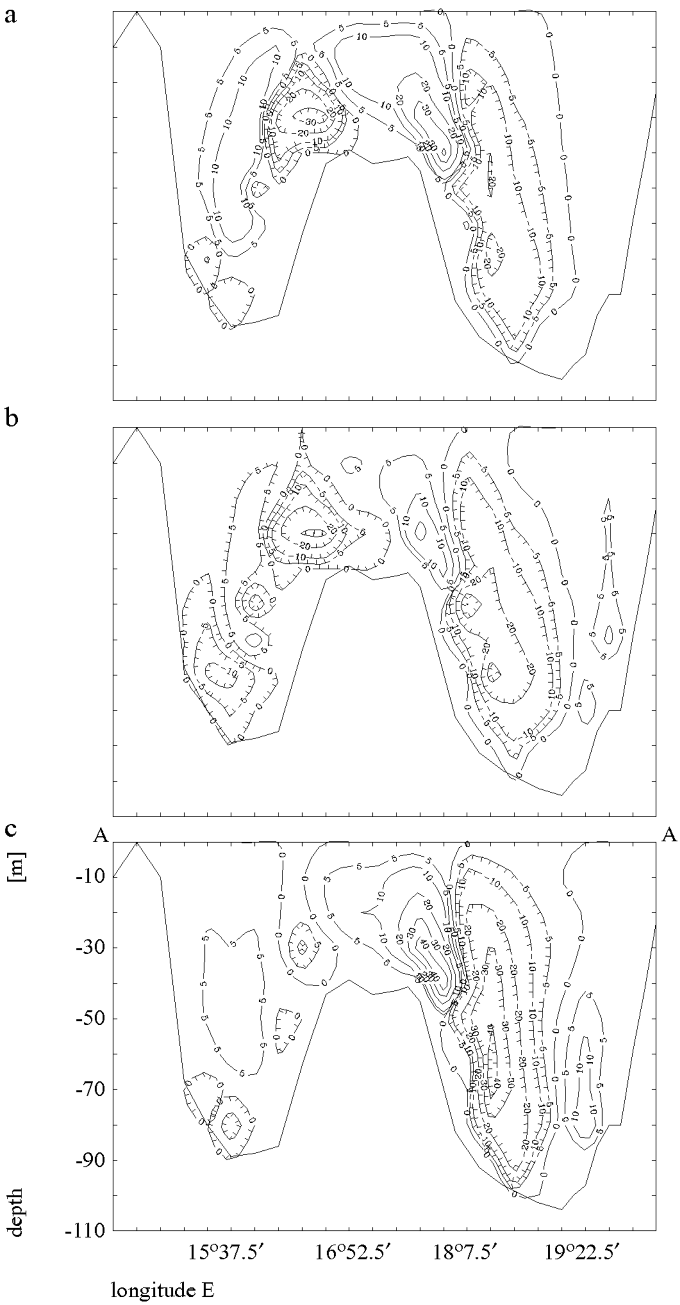

Fig. 4. Vertical current velocity component w [10^{-4} cm s^{-1}] along the longitudinal vertical section A–A ($\varphi = 55°05'$; its location is shown in Figs. 1 and 2) for selected months: May (a), July (b) and September (c). Isolines were plotted for the following values of w: $\pm\,(0, 5, 10, 20, 30, 40, 50, 75, 100)\times10^{-4}$ [cm s^{-1}]. Hatched areas indicate regions of upwelling (negative values of w)

4. Discussion

Maximal (minimal) values of the vertical velocity component in the Baltic are in the range of $\pm(10^{-3}$–$10^{-2})$ cm s^{-1}. Over large areas of the Baltic Sea its values do not exceed $\pm(10^{-4}$–$10^{-5})$ cm s^{-1}. From these results it follows that the simple model of estimating the vertical velocity presented in this paper gives realistic values of w. They are consistent with similar estimates of w for the Baltic Sea reported by other authors (*cf.* Staśkiewicz, 1974, *cit.* after Jankowski (1988); Kuznetsova and Tyuriakov, 1981) and with the results of calculations of this component of current velocity in other sea basins (see for example – Sarkisyan *et al.*, 1986).

In Fig. 5 the spatial distribution of the seawater density (multi-year averaged data) for May along the same longitudinal vertical section as in the case of w in Fig. 4 was plotted. The layout of isolines of the seawater density indirectly confirms the occurrence of upwelling in the vicinity of the Gdańsk Deep.

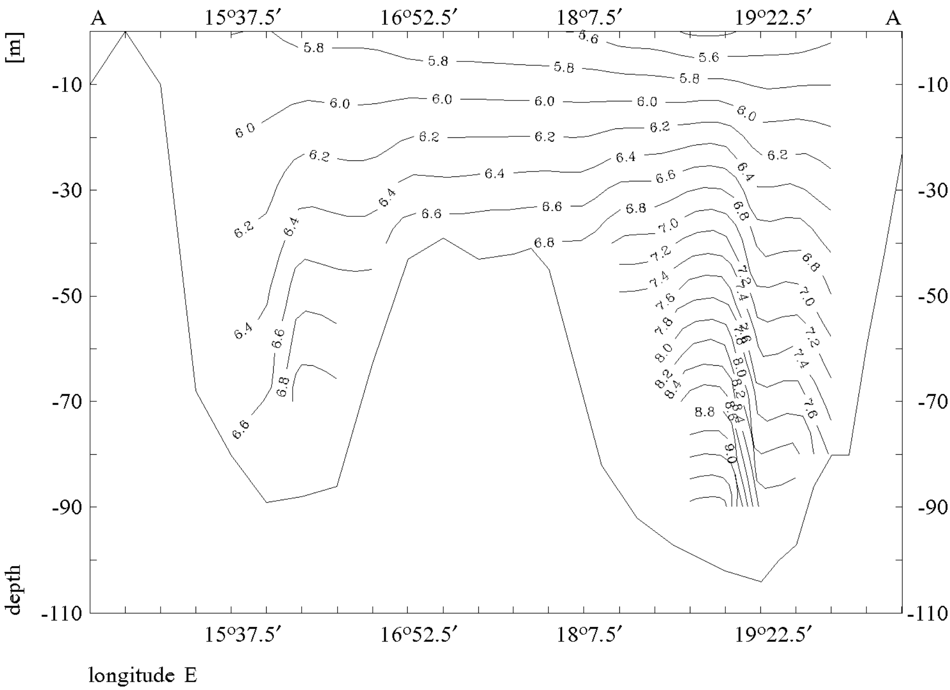

Fig. 5. Seawater density σ_t along the zonal longitudinal section A–A ($\varphi = 55°05'$; its location is shown in Figs. 1 and 2). Contour interval $\triangle\sigma_t = 0.2$

The calculated values of w allow the duration of the vertical movements in upwelling (downwelling) areas to be estimated. The mean vertical velocity

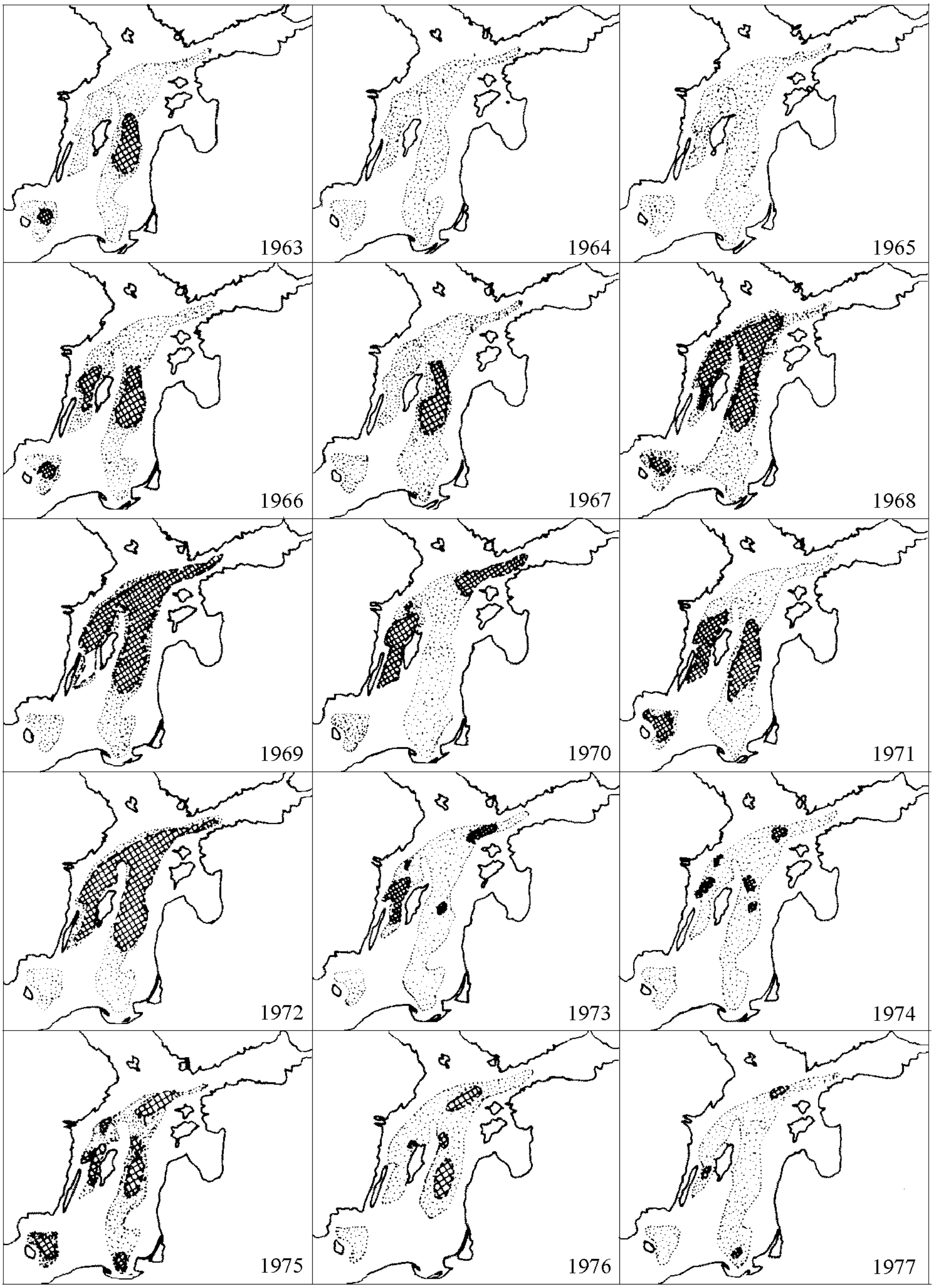

Fig. 6. The spatial distribution of bottoms covered by hydrogen sulphide-containing waters (hatched areas) or with water containing less than 2 cm^3 oxygen per dm^3 (dotted areas) in the Baltic Sea (Melvasalo *et al.*, 1981) for the years 1963–1977. Each map shows the sum of all situations recorded during the year in question

is calculated at about $\pm 10^{-3}$ cm s^{-1}, resulting in $w = \pm 0.864$ m day^{-1}. Thus, a period of about 116 days is necessary for a water molecule to move along a vertical distance equal to the local basin depth of 100 m.

The results of calculations of the vertical velocity component suggest that the regions in the vicinity of the Gdańsk and Bornholm Deeps have some characteristc features in the local water dynamics (circulation pattern). During all the months of a year gyre-like structures are observed in the current fields (Staśkiewicz, 1988; Jankowski and Skałka, 1995) and the areas of upwelling (downwelling) connected with them (Figs. 1–4). The complicated bottom topography in combination with external forcing (wind and density fields) are significant factors in controlling the dynamics of water in the region.

This gyre-like structure of the water circulation is important not only for the ecological, biological and geological understanding of the region, but also in pollution problems. By comparison with adjacent water masses, conditions are created in which chemical, biological or geological substances can accumulate more readily or in which the water can act as a sink for all passive substances like pollutants. This increases the residence time of these substances within gyres and may intensify *e.g.* oxygen deficiencies or the redistribution of the sediments in bottom layers.

The spatial distribution of so-called 'dead zones' (low concentrations of oxygen or the occurrence of hydrogen sulphide) (see Fig. 6 after Melvasalo *et al.* (1981) for the period from 1963 to 1977; and *cf.* also Andersin and Sandler (1988) for the years 1963–1987 and Trzosińska (1994) for the years from 1969–1989) are nearly identical with the areas of the gyre-like currents and are related to the 'permanent' upwelling (downwelling) regions in the Gdańsk and Bornholm Deeps.

5. Conclusions

This model for estimating the vertical current velocity component w provides a unique solution to eq. (25), fulfilling not only the kinematic boundary conditions at the free sea surface (5) and the sea bottom (6) (or (8)) but also, within the set of the assumptions made in the diagnostic model, the integral continuity eq. (22).

The results of calculations of w yield reasonable estimates of the order of magnitude of the vertical current velocity component consistent with the results reported by other investigators. Areas of intensified vertical water movements are well correlated with ecological parameter distributions and with the gyre-like structures in the current fields.

The spatial distributions of the vertical velocity component w and the horizontal currents estimated by the diagnostic model for selected months may provide useful information for elucidating the hydrological background necessary to diagnose the state of the Baltic Sea marine environment on a seasonal (and annual) time scale, especially during stagnation periods.

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