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THE INFLUENCE OF THE ADVECTION ON THE WATER TEMPERATURE DISTRIBUTION IN THE GULF OF GDAŃSK; NUMERICAL STUDY

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Key words: Gulf of Gdańsk, numerical study, a three-dimensional hydrodynamic model, water temperature, advection

Abstract

The influence of advection on spatial distribution of temperature in the Gulf of Gdańsk is analysed. In this analysis heat fluxes are completely omitted. The advection is brought about mainly by winds and depends particularly on their direction. Eastern winds curry the surface waters (usually contaminated) out of the Gulf and cause the inflow of cool bottom waters from the Gdańsk Deep. The same winds induce the movement of the oceanic salty and oxidized water from the Bornholm to the Gdańsk Deep via the Słupsk Channel. The advection during eastern winds is a desirable phenomenon for restoring the Gulf of Gdańsk waters. The temperature distribution depends also on the advection influenced by meridional winds.

INTRODUCTION

Recently, two new trends in scientific research of the Baltic Sea have emerged. The first is connected with studies on spreading of surface contaminated waters (Fennel and Neumann 1996, Suursaar and Astok 1996, Tamsalu 1996); the second concentrates on the inflows of salty, oceanic waters into the Baltic through the Danish Straits (Krauss and Brügge 1991, Gidhagen and Håkansson 1992, Håkansson and Dahlin 1993, Lehmann 1995, Elken 1996). In papers by Krauss

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and Brügge (1991) and by Jankowski (1997) it has been reported that in consequence of eastern winds, the bottom waters are translocated from the Bornholm to the Gdańsk Deep via the Słupsk Channel. These salty and oxidized waters are advancing also to the basins farther off the Danish Straits.

The aim of this project is the investigation of flows in the Gulf of Gdańsk connected with the outflow of polluted surface waters and inflow of salty and oxidized ones originating from deeper layers. The wind driven circulation causes the outflow of surface waters out of the Gulf and the compensatory inflow of deep waters, cool and oxidized. To investigate such a possibility of restoring the Gulf waters, a three-dimensional, baroclinic, hydrodynamic model has been used (Blumberg and Mellor 1987). Water temperature has been chosen as an indicator of the mentioned flows. To confirm directions of the flows, the cross-sections of temperature and velocity at the region of the Gulf of Gdańsk open boundary is analysed. In this study, the atmospheric heat fluxes have not been taken into account, which in consequence excluded processes of the vertical diffusion of heat from consideration. Both spatial and temporal distributions of temperature are thus consequently formed due to advection dependent on wind forcing. The character of temperature fields allows one to confirm the existence of the inflows of restoring waters.

To achieve in such a way formulated purpose, a few steps based on using the hydrodynamic model were undertaken. The three-dimensional, baroclinic hydrodynamic model used was subjected to simulated wind conditions. The winds assumed for this modelling represented multi-year averaged velocities and directions encountered over the Baltic Sea in August, that is gale winds (20 $m\,s^{-1}$) blowing from N, S, E. These wind stresses were ascribed to each note of numerical grid. Upon these winds, a diurnal periodicity was superpositioned and the model's behaviour was investigated. The initial density field for whole Baltic Sea was based on August data.

Preliminary computations designed to test the simulation process were done using the five-layer model. The further computations were continued with the aid of eleven-layer model. The results of the vertical model of temperature for chosen stations in the Gulf of Gdańsk are presented in Fig. 1. Temporary changes of the mean temperature of layers enabled the description of water outflows or inflows into the Gulf while the temperature and velocity in the open boundary region (cross-section AA'), showed directions of advection which was responsible for the obtained thermal distributions.

A DESCRIPTION OF THE MODEL

In the presented investigations, the three-dimensional, baroclinic Blumberg and Mellor's model was used (Blumberg and Mellor 1987). This model was developed for the oceanic coastal waters as well as for the description of dynamic

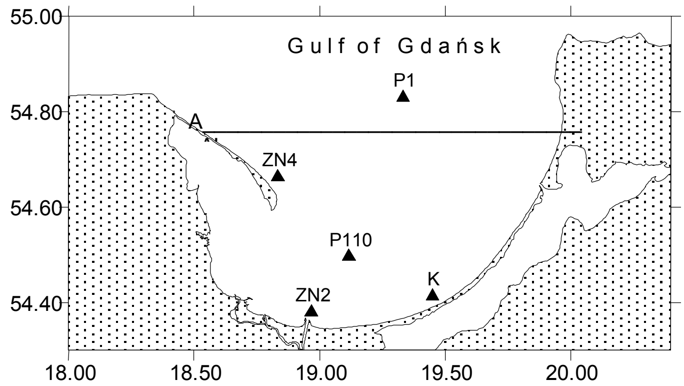

Fig. 1. The localisation of stations and the cross-section AA' in the Gulf of Gdańsk

processes in estuaries, *e.g.* the Gulf of Mexico (Oey 1996), Hudson-Raritan estuary (Oey *et al.* 1985). The equations of this circulation model describe the velocity, surface elevation, temperature and salinity fields. The Reynolds equations of motion were following:

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z}-fv=-\frac{1}{\rho_0}\frac{\partial p}{\partial x}+\frac{\partial}{\partial z}\left(K_{\mathrm{M}}\frac{\partial u}{\partial z}\right)+ + A_{\mathrm{M}}\left(\frac{\partial^2 u}{\partial x^2}+\frac{\partial^2 u}{\partial y^2}\right), \tag{1}$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+w\frac{\partial v}{\partial z}+fu=-\frac{1}{\rho_0}\frac{\partial p}{\partial y}+\frac{\partial}{\partial z}\left(K_{\mathrm{M}}\frac{\partial v}{\partial z}\right)+A_{\mathrm{M}}\left(\frac{\partial^2 v}{\partial x^2}+\frac{\partial^2 v}{\partial y^2}\right), \tag{2}$$

where:

u, v and w, components of the horizontal velocity along the axes x, y, z oriented towards east, north and upward, respectively; f, Coriolis parameter; ρ, ρ_o, *in situ* and reference density; p, pressure; K_M and A_M, vertical and horizontal eddy diffusivity of turbulent momentum.

The continuity equation was

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

The conservation equation for temperature was written as

$$\frac{\partial T}{\partial t}+u\frac{\partial T}{\partial x}+v\frac{\partial T}{\partial y}+w\frac{\partial T}{\partial z}=\frac{\partial}{\partial z}\left(K_{\mathrm{H}}\frac{\partial T}{\partial z}\right)+A_{\mathrm{H}}\left(\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial y^2}\right)+\theta_{\mathrm{T}}, \tag{4}$$

where:

T, temperature of water; K_{H} and A_H, vertical and horizontal heat diffusivity for turbulent mixing of heat; θ_{T}, source of heat function.

Using the temperature and salinity, the density was computed according to an equation of state

$$\rho=\rho(T,S), \tag{5}$$

given by Mamayev (1975) in a form

$$\rho=1+\sigma_{\mathrm{t}}, \tag{6}$$

where:

$\sigma_t = 28.152-0.0735T-0.00469T^2+(0.802-0.002T)\cdot(S-35)$; S, salinity; T, water temperature.

The above equations contain parameterized Reynold's stresses and flux terms which were obtained according to the second order turbulence model (Mellor and Yamada 1982). They characterise the turbulence kinetic energy and turbulence macroscale ℓ:

$$\begin{aligned}\frac{\partial q^2}{\partial t}+u\frac{\partial q^2}{\partial x}+v\frac{\partial q^2}{\partial y}+w\frac{\partial q^2}{\partial z}&=\frac{\partial}{\partial z}\left(K_{\mathrm{q}}\frac{\partial q^2}{\partial z}\right)+2K_{\mathrm{M}}\left[\left(\frac{\partial u}{\partial z}\right)^2+\left(\frac{\partial v}{\partial z}\right)^2\right]+\\&+\frac{2g}{\rho_0}K_{\mathrm{H}}\frac{\partial\rho}{\partial z}-\frac{2q^3}{B_1\ell}+A_{\mathrm{M}}\left(\frac{\partial^2 q^2}{\partial x^2}+\frac{\partial^2 q^2}{\partial y^2}\right),\end{aligned} \tag{7}$$

$$\begin{aligned}\frac{\partial q^2\ell}{\partial t}+u\frac{\partial q^2\ell}{\partial x}+v\frac{\partial q^2\ell}{\partial y}+w\frac{\partial q^2\ell}{\partial z}&=\frac{\partial}{\partial z}\left(K_{\mathrm{q}}\frac{\partial q^2\ell}{\partial z}\right)+\ell E_1K_{\mathrm{M}}\left[\left(\frac{\partial u}{\partial z}\right)^2+\left(\frac{\partial v}{\partial z}\right)^2\right]+\\&+\frac{\ell E_1 g}{\rho_0}K_{\mathrm{H}}\frac{\partial\rho}{\partial z}-\frac{q^3}{B_1}\widetilde{W}+A_{\mathrm{M}}\left(\frac{\partial^2 q^2\ell}{\partial x^2}+\frac{\partial^2 q^2\ell}{\partial y^2}\right),\end{aligned} \tag{8}$$

where:

$q^2\ell$, turbulent kinetic energy; ℓ, macroscale of turbulence; K_{q}, the vertical coefficient of turbulence kinetic energy; κ, Karman's constant; H, depth of the sea; B_1, E_1, E_2, empirical constants.

The boundary conditions at the free surface $z = \xi(x, y)$ included

- wind stresses of momentum

$$\rho_0 K_M \frac{\partial u}{\partial z} = \tau_{ox} , \tag{9}$$

$$\rho_0 K_M \frac{\partial v}{\partial z} = \tau_{oy} , \tag{10}$$

- heat fluxes

$$\rho_0 K_H \frac{\partial T}{\partial z} = H_0 , \tag{11}$$

- energy fluxes

$$q^2 = B_1^{2/3} U_{*0}^2 , \tag{12}$$

$$q^2 \ell = 0 , \tag{13}$$

- kinematic condition

$$w = u \frac{\partial \xi}{\partial x} + v \frac{\partial \xi}{\partial y} + \frac{\partial \xi}{\partial t} . \tag{14}$$

At the bottom $z = H$, the bottom friction stresses

$$\rho_0 K_M \frac{\partial u}{\partial z} = \tau_{bx} \tag{15}$$

and

$$\rho_0 K_M \frac{\partial v}{\partial z} = \tau_{by} , \tag{16}$$

were parameterized as

$$\tau_{bx} = \rho_0 C_D \sqrt{u_b^2 + v_b^2}\, u_b \tag{17}$$

and

$$\tau_{by} = \rho_0 C_D \sqrt{u_b^2 + v_b^2}\, v_b \, . \tag{18}$$

The energy fluxes at the bottom are expressed as

$$q^2 = B_1^{2/3} U_{*b}^2 \, , \tag{19}$$

$$q^2 \ell = 0 \, , \tag{20}$$

and the kinematic condition as

$$w_b = -u_b \frac{\partial H}{\partial x} - v_b \frac{\partial H}{\partial y} \, , \tag{21}$$

where:

τ_{ox}, τ_{oy}, surface wind stresses; H_0, atmospheric heat flux; τ_{bx}, τ_{by}, bottom stresses; C_D, drag coefficient ($C_D = 0.0025$); $u_{*_b} = \sqrt{\frac{\tau}{\rho_0}}$, velocity friction; u, u_b, v, v_b, w, w_b, components of velocities at the surface (no index) and at the bottom (index b) respectively.

Boundary condition at the lateral boundaries (rivers) was assumed to be

$$\frac{\partial T}{\partial t} + u_n \frac{\partial T}{\partial n} = 0 \, . \tag{22}$$

The initial conditions that is those at the 0 moment ($t = 0$) were taken as follows:

- for velocities

$$u\,(x, y, z) = 0, \tag{23}$$

$$v\,(x, y, z) = 0, \tag{24}$$

$$w\,(x, y, z) = 0, \tag{25}$$

- for the sea level

$$\xi(x, y) = 0 \, ; \tag{26}$$

- for temperature

$$T = T(x, y, z). \tag{27}$$

The present experiment excludes the atmospheric heat fluxes which means that vertical heat diffusion does not occur at any layer. In this way, the factors responsible for either maintaining or destroying the stratification have been also removed (Rowe and Wells 1985). Spatial and temporary temperature distributions are thus caused only by the advective waters. Omitting the surface fluxes gave the insight into the fields of temperature induced solely by dynamics.

The model was tested for different wind forcing. Initially, the multi-year averaged wind field (including velocities and directions in August) over the whole Baltic Sea area was adopted. The maximal mean velocities reached 5 m s^{-1} with wind direction from W as dominant. Next runs of the model contained greater (20 m s^{-1}) wind velocities for the chosen directions: N, S or E. Both N and S directions seemed to be important for the meridional exchange between the Gulf waters and the Southern Baltic. Eastern winds were considered because of the possibility of the surface outflow of contaminated waters, according to Ekman's law. On the other hand, it was supposed that the compensating inflow of salty, cool and oxidized waters from the Gdańsk Deep should occur. The latter waters move from the Słupsk Channel, mainly during easterly winds, too.

In the next step, the superposition of diurnal periodicity upon the previously assumed winds was tested. Zonal and meridional components of wind stresses were taken as

$$\tau_{x,y} = \tau_{\mathrm{o}}[1 + 0.05 \cdot \sin(\frac{2\pi}{T_{\mathrm{d}}})t], \tag{28}$$

where:

τ_o, wind stress equivalent to climatic wind in August or constant velocity 20 m s^{-1};

T_d, the diurnal period.

Dynamic responses of the sea and changes of the temperature distribution were expected.

The equations contained transformed from (x, y, z, t) to (x^*, y^*, σ, t^*) coordinates where: $x^* = x$, $y^* = y$, $t^* = t$ and σ was defined as

$$\sigma = \frac{z - \xi}{H + \xi}. \tag{29}$$

This model uses a staggered finite difference grid (Arakawa C grid).

RESULTS

The primary scope of this project has been the investigation of the model behaviour at the southern Baltic. The responses of the model subjected to constant or periodical southerly wind stress are presented in Figs. 2 and 3. During constant wind from S (Fig. 2 a) and the surface stress amounting to 0.63 N m^{-2} (for 20 m s^{-1}), the sea level oscillations fluctuated in the range from -5 cm to 29 cm (Fig. 2 b). After 10 days, the bottom friction brought about the decrease of 1 cm oscillations. The sea level was set up 18 cm above the mean level. The volume of the basin was slightly decreasing over 25 days, despite of the fluctuations resulting in the changes of the sea level, which was confirmed by positive balance of outflows (Tab. 1, Fig. 2 c). When the values of the balance (ΔV) are positive this means that the outflow towards the Southern Baltic took place. It can be seen as a little downward slope of the mean value of the total volume (Fig. 2 c). The averaged, for the whole basin, temperature appeared to be constant (Fig. 2 d). Additional imposing of diurnal periodicity resulted in the oscillations of the sea level (comp. Fig. 2 a, b and Fig. 3 a, b), their amplitudes followed the wind conditions. Such factors like the volume and temperature remained the same. The results obtained for constant and periodical winds were expected and were in accord with the law of conservation.

Table 1

Balance of flows through the cross-section AA' [Sv[1]] for 25 days with 20 m s^{-1} winds from directions: South (S), North (N), East (E) and VIII[2]

	Inflows [–Sv]				outflows [+Sv]				ΔV [+–Sv]			
Balance after	VIII	S	N	E	VIII	S	N	E	VIII	S	N	E
1 day	2.3	18.3	17.8	10.6	2.3	19.3	17.1	10.8	–0.03	+0.97	–0.79	+0.16
10 days	7.5	48.4	19.8	25.3	7.6	50.6	17.3	24.5	+0.15	+2.29	–2.51	–0.84
15 days	7.6	48.9	19.7	23.2	7.4	50.4	17.2	21.6	–0.24	+1.52	–2.48	–1.61
20 days	7.6	37.8	17.1	14.2	7.0	38.6	15.1	13.5	–0.60	+0.77	–2.00	–0.69
25 days	8.8	37.9	17.2	12.8	8.1	39.2	14.8	13.3	–0.75	+1.21	–2.34	+0.51

The wind impact on the vertical temperature distribution was studied also in relation to wind duration (Fig. 4) (black stars in Figs. 2 and 3). The initial profile was based on August data. The evolution of the profile shapes was occurring at the changing rate. During the first five days, the quickest changes of the profile took place, mainly at the mixing layer. After subsequent 10 days the vertical gradients of temperature decreased leading to homogeneity. At station P1, after

[1] Sv = 10^6 m^3 s^{-1} [sverdrup]

[2] VIII, climatic field of wind based on the average velocities and directions in August over the Baltic Sea

20 days waters became completely mixed and the temperature dropped to 10.5 °C. It does not indicate homogeneity in the whole basin, however, which is confirmed by the mean temperature (Figs. 2 d and 3 d). For the short time, the vertical profile appeared to reverse. This points out to the influence of wind also on shaping the temperature distribution by the horizontal diffusion and advection. Comparison between the initial and final temperature profiles clearly showed the change of the thermal structure at station P1.

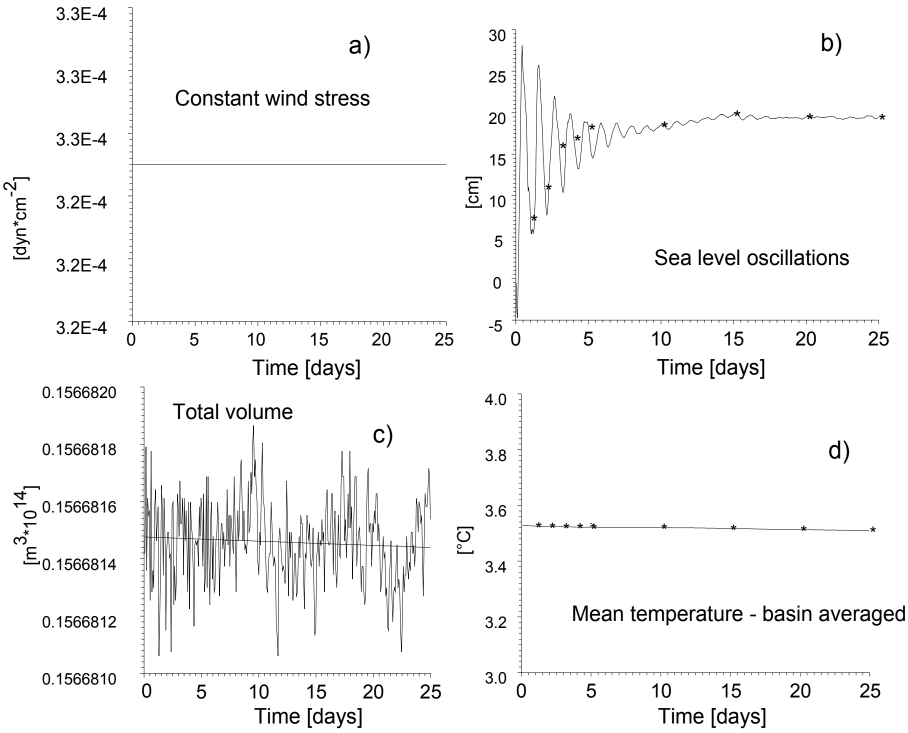

Fig. 2. The constant wind stress from S (a), the sea level oscillations (b), changes of volume (c) and mean temperatures (d) within the period of 25 days and assumed August density field in the Baltic Sea; (the black stars denote the days when the vertical temperature profiles were taken)

When the model was subjected to easterly wind stress, the amplitudes of sea level oscillations were a little smaller while temperatures decreased markedly (Fig. 5). The mean volume increased slightly (Tab. 1) due to certain overbalance of the inflow (Fig. 5 a). This fact suggests that easterly winds induce a frictional Ekman's flow away from the Gulf. Uniform vertical profile was reached at temperature of 5 °C (Fig. 5 c), which means that there must have occurred a clearly defined cold deep inflow. The same wind stress but from different directions resulted in different vertical profiles and dissimilar fields of temperatures. For instance, the simulations with climatic field of the wind in August gave unsteady oscillations of the sea level (Fig. 6) while vertical temperature profiles did not change considerably after successive time steps. The sea level fluctuations were unsteady and exhibited damping, periodical pulsations. Such fluctuations reflect long-term oscillations induced by fields of climatic winds. The small amplitudes resulted from gentle winds. The vertical profiles during the entire simulation essentially resembled the initial ones. The results of the baroclinic model depended on wind stresses. Mild winds caused mainly oscillations of the stratified waters but strong winds lead to the homogeneity brought about by advection and eddy viscosity. It seems that the used model can be employed for prognostic evaluations.

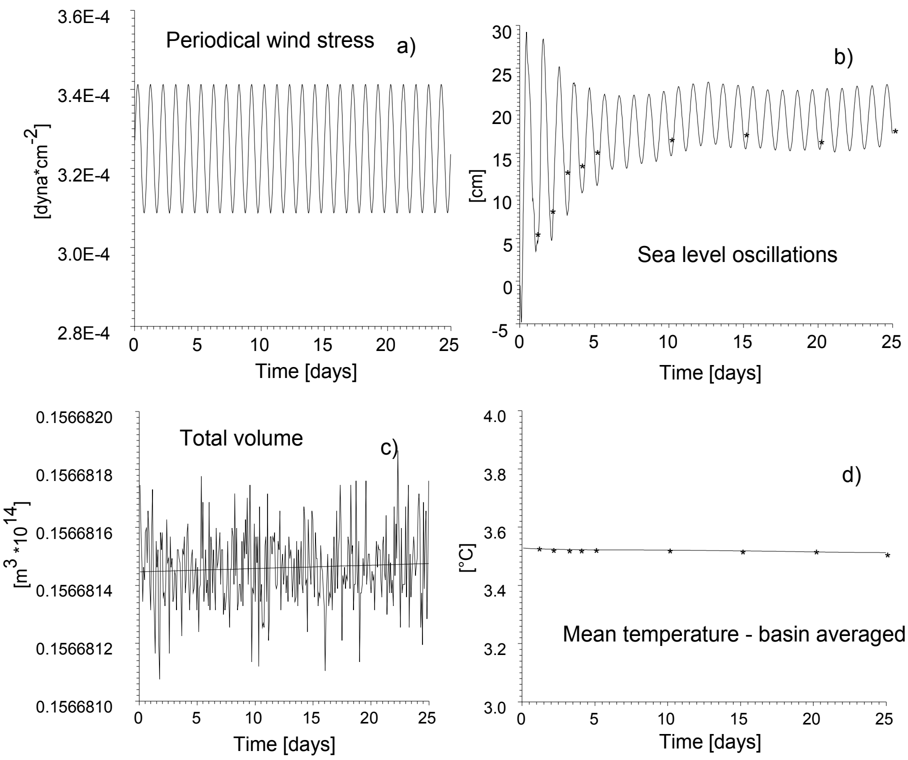

Fig. 3. The oscillating wind with diurnal period (a), the sea level oscillations (b), changes of volume (c) and mean temperatures (d), within the period of 25 days and assumed August density field in the Baltic Sea; (the black stars denote the days when the vertical temperature profiles were taken)

THE EVOLUTION OF TEMPERATURE PROFILES IN VERTICAL MODEL

The evolution of profiles derived from the vertical model was simulated by the eleven-layer model. The water temperature was followed at all chosen stations (P1, ZN4, P110, ZN2 and K) during periodical easterly wind stress. At station

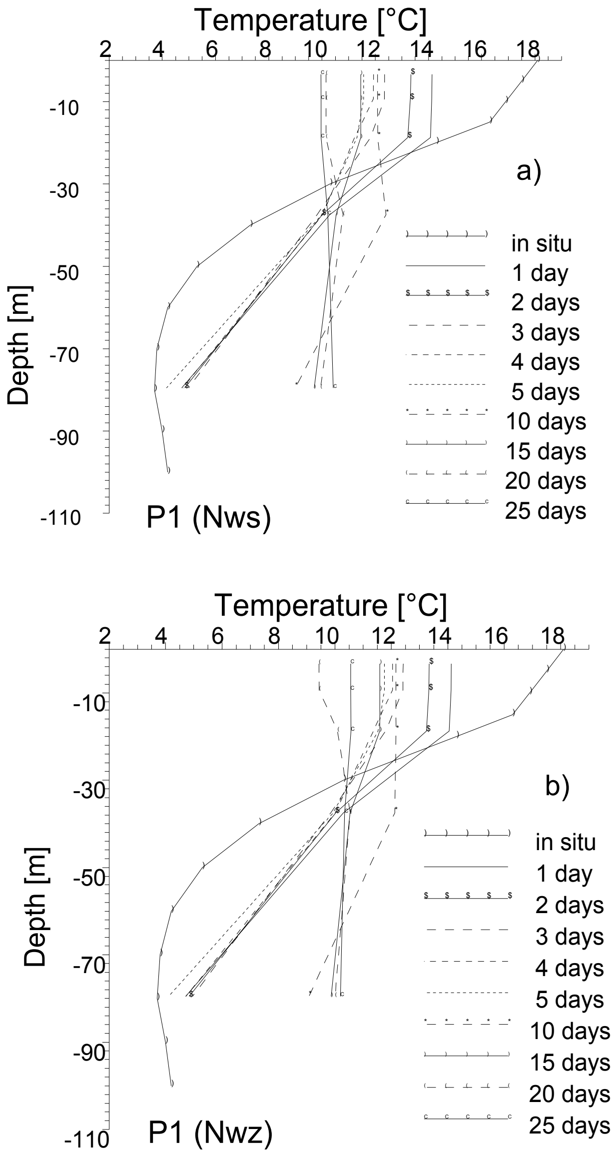

Fig. 4. Vertical temperature profiles simulated by the five-layer model for station P1 (the Gdańsk Deep) influenced by a constant (a) and periodical (b) wind (20 m s^{-1}) from N

P1, the profiles were changing intensively for the first 10 days (Fig. 7 a). During the rest of simulation time, they became homogeneous or were approaching uniformity. At station ZN4, localised close to the Hel Peninsula, a shallow thermocline moved down from a depth of 20 m to 60 m, then this weak stratification almost disappeared (Fig. 7 b). More pronounced course of the temperature profiles emerged in the case of station P110 (Fig. 7 c). Generally, at the three analysed stations localized farther off the shoreline, the vertical stratifica-

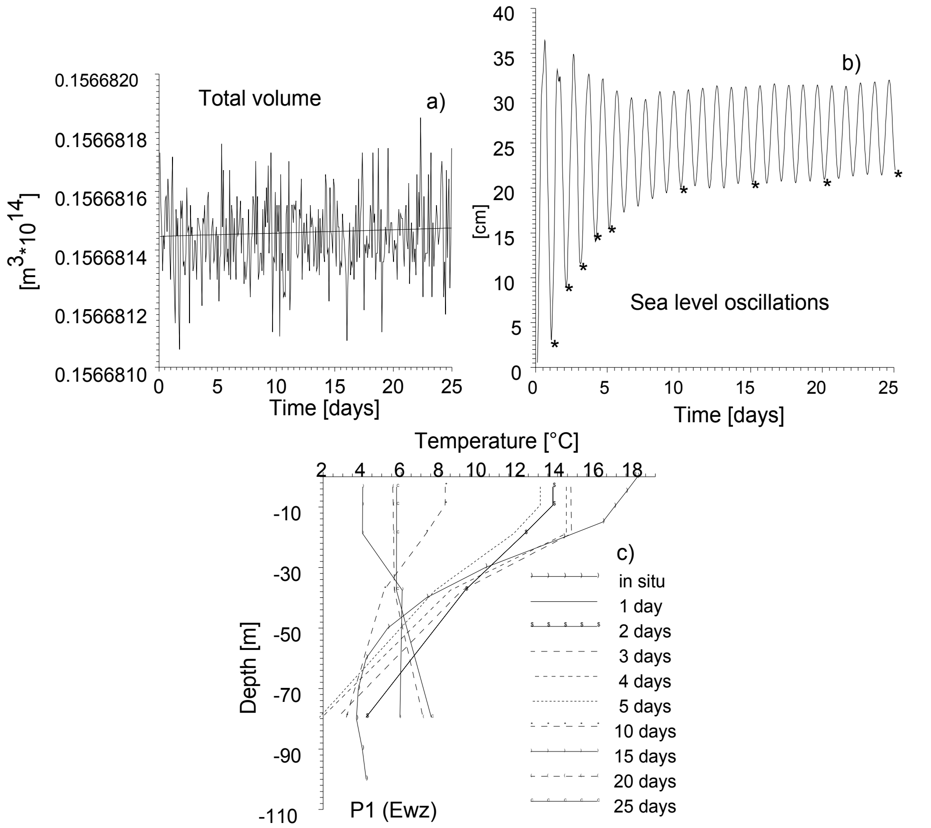

Fig. 5. Changes of volume (a), oscillations of the sea level (b) and vertical temperature profiles (c) induced by periodical wind (20 m s^{-1}) from E at station P1

tion was developing rather at the beginning of the simulations, then it was becoming less obvious.

The temperature profiles derived from modelling at stations localised close to the shore (ZN2, K) were a little different (Figs 7 d, c). At station ZN2, a maximal vertical gradient of temperature moved 20 m downward from 15 m of depth, became upturned, then moved 30 m upward and finally disappeared leaving behind homogeneity (Fig. 7 d). At station K (Fig. 7 e), the distribution of temperature was characterised by close-to-uniform profile with only just marked vertical gradients. The vertical temperature distributions, as a result of acting meridional (northerly and southerly) and August winds, were chosen for shallow station ZN2 (Fig. 8).

The response of the vertical distribution of temperature subjected to different directions of wind is presented for station ZN2. At this station, the temperature profiles were computed for the circulation driven by August field of wind and by gale winds from N, E and S. Simulation for August wind gave the distinct summer stratification. The northerly wind turned out to be the most mixing one and after three days the water temperature was uniform. Strong wind from E induced clearcut vertical distribution of temperature and the formation of a warming vertical front with time (Fig. 7 d). The field of temperature evaluated for southerly wind distinguished existing thermocline at 30 m to 40 m (Fig. 8 a).

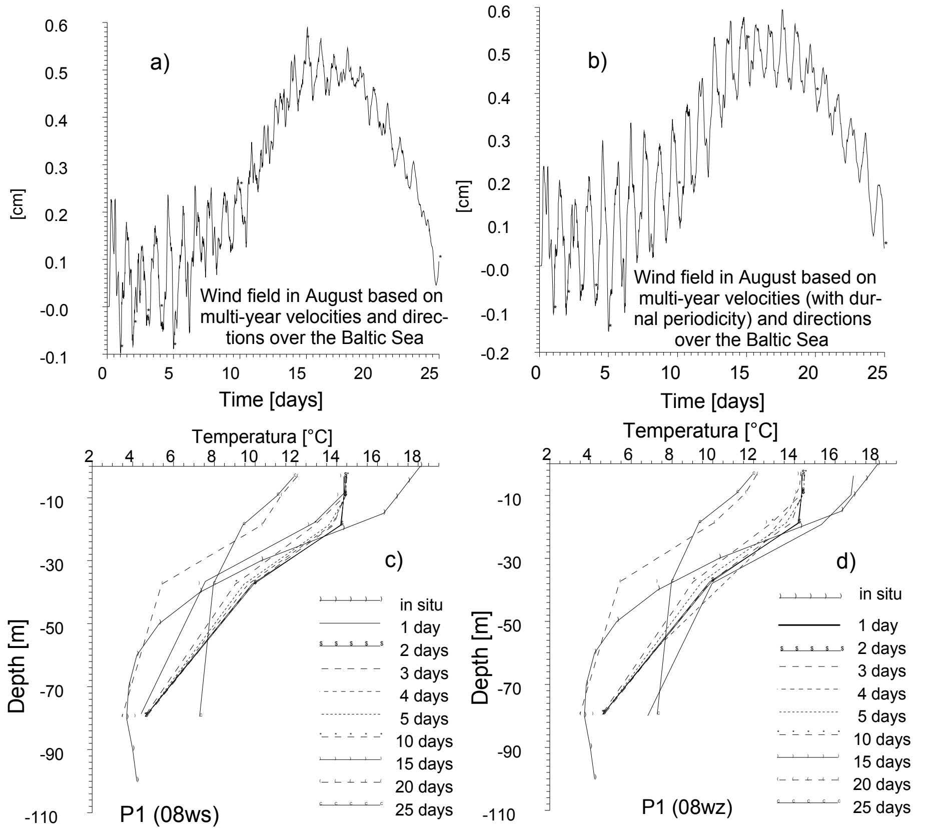

Fig. 6. The sea level oscillations and vertical temperature profiles induced by climatic wind (August field) in the Baltic Sea; simulation for station P1

MEAN TEMPERATURE OF LAYERS

Winds from different directions brought about different fluctuations of layer temperatures. During gentle winds in August (when the velocity did not exceed 5 m s^{-1}), the thermal stratification was not destroyed (Fig. 9). Time series of the eleven mean layer temperatures was nearly parallel for 25 days of the simulation. At station ZN4 localised close to the shoreline, the stratification was stable, particularly in the surface layers. The same layers, farther off the shore were homogeneous. The vertical temperature profiles covered the range temperatures from 3.5 °C to 18 °C (Fig. 9). The diagrams obtained, revealed that mild winds maintained the summer stratification.

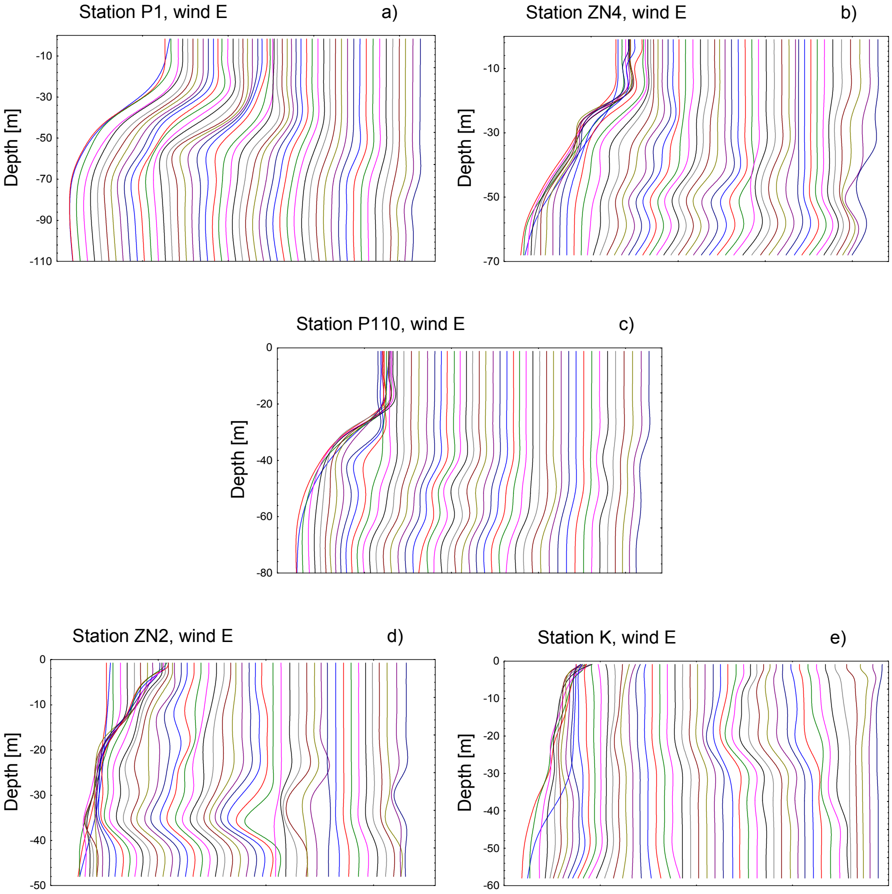

Fig. 7. Time series of vertical temperature profiles simulated for chosen stations in the Gulf of Gdańsk

Opposite behaviour of model results was observed when advection was influenced by winds from N and E. Northerly wind caused strong surface inflows and the rise of water temperature (Fig. 10). The most evident impact of these inflows was seen as the homogeneous profile (after 3 days) at shallow station ZN2 (48 m), with a high value of temperature exceeding 18 °C (Fig. 10). Similar situation occurred at station P110 (80 m of depth), however, homogeneity at this profile was achieved after 17 days of the simulation. The smallest changes of temperature were noticed in the region of the Gdańsk Deep (station P1)

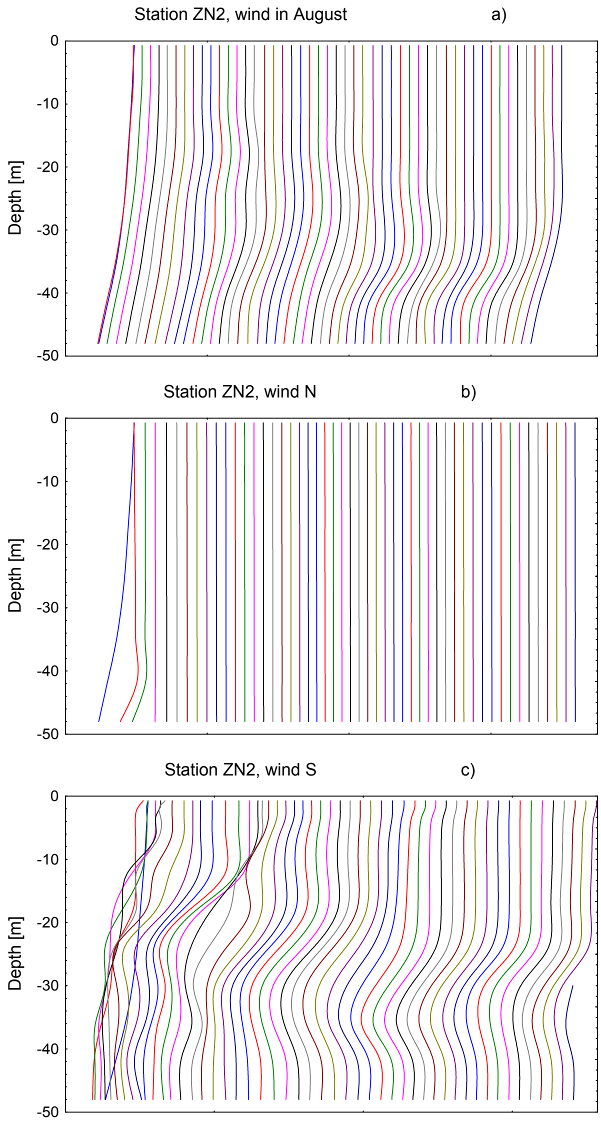

Fig. 8. Time series of vertical temperature profiles induced by winds from different directions; simulation for station ZN2

where the surface temperature slightly decreased and the bottom one increased. In the case of coastal waters (stations ZN4, K), the temperatures fluctuated (Fig. 10), but generally the water temperature was increasing.

Easterly winds had the opposite effect, that is they caused the outflow of surface waters from the Gulf and the inflow of cool waters from the deeper levels (Fig. 11). Series of mean temperature resulting from easterly winds were just opposite to the ones described above. At all stations (P1-K), the layer temperatures were fluctuating but decreasing. The initial profiles covered the temperature range from 3.5 °C to 18 °C, the final ones the range from 3 °C to 8 °C.

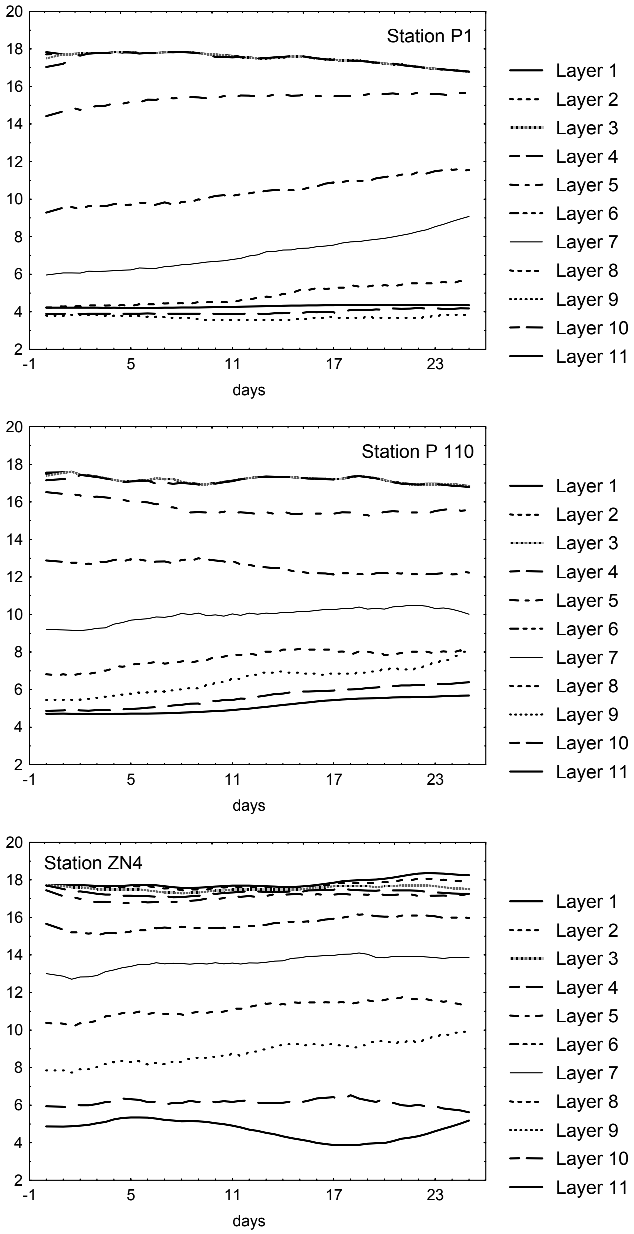

Fig. 9. Modelled mean temperature of water layers at chosen stations in the Gulf of Gdańsk shaped by the field of climatic wind in August

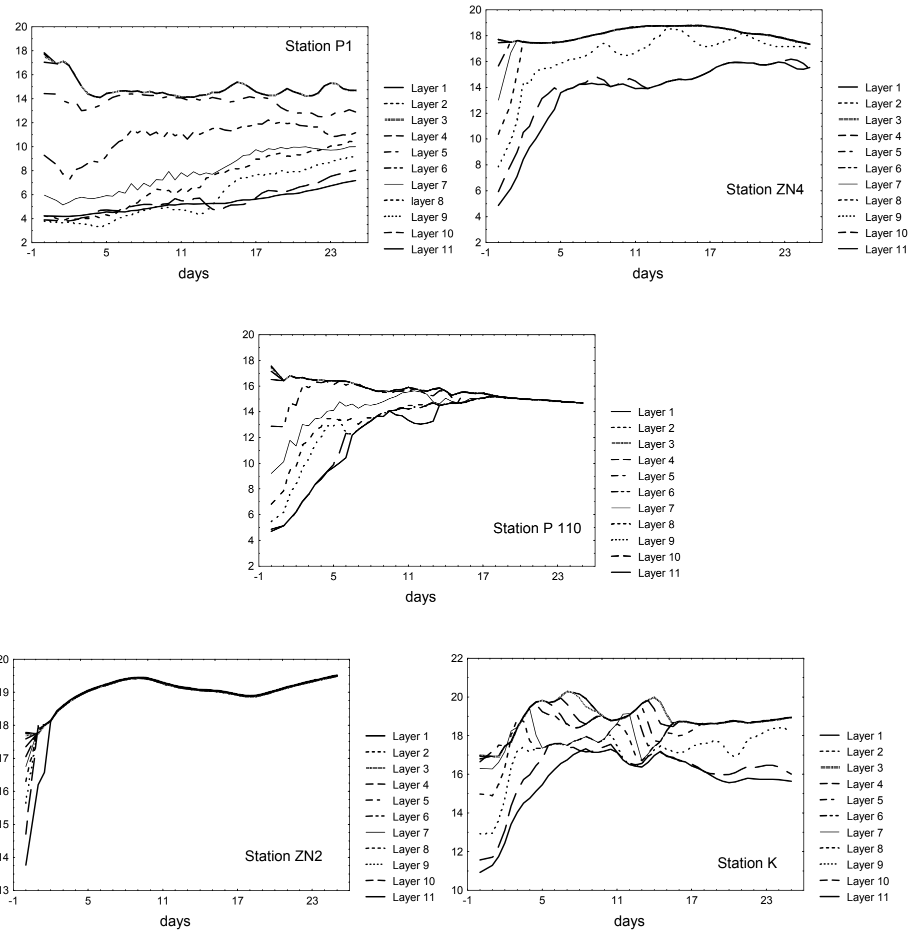

Fig. 10. Modelled mean temperature of water layers at chosen stations in the Gulf of Gdańsk shaped by the field of winds from N

THE TEMPERATURE AND VELOCITY IN THE REGION OF THE OPEN BOUNDARY OF THE GULF OF GDAŃSK

As it has been mentioned before, the presented model considers neither atmospheric heat fluxes nor river runoff nor outflow of warmer surface water out of the Gulf, therefore the spatial distribution of temperature depends only on advection of cool near-bottom waters from the Gdańsk Deep. The main area of such exchanges is found in the open boundary region of the Gulf of Gdańsk (Fig. 1), depicted as the cross-section AA' (Figs. 12-14).

To illustrate the character of the temperature and velocity distributions, two situations were analysed: one after the first and the other after the 25th day of the simulations with imposed wind from N (Fig. 12), S (Fig. 13), and E (Fig. 14). In

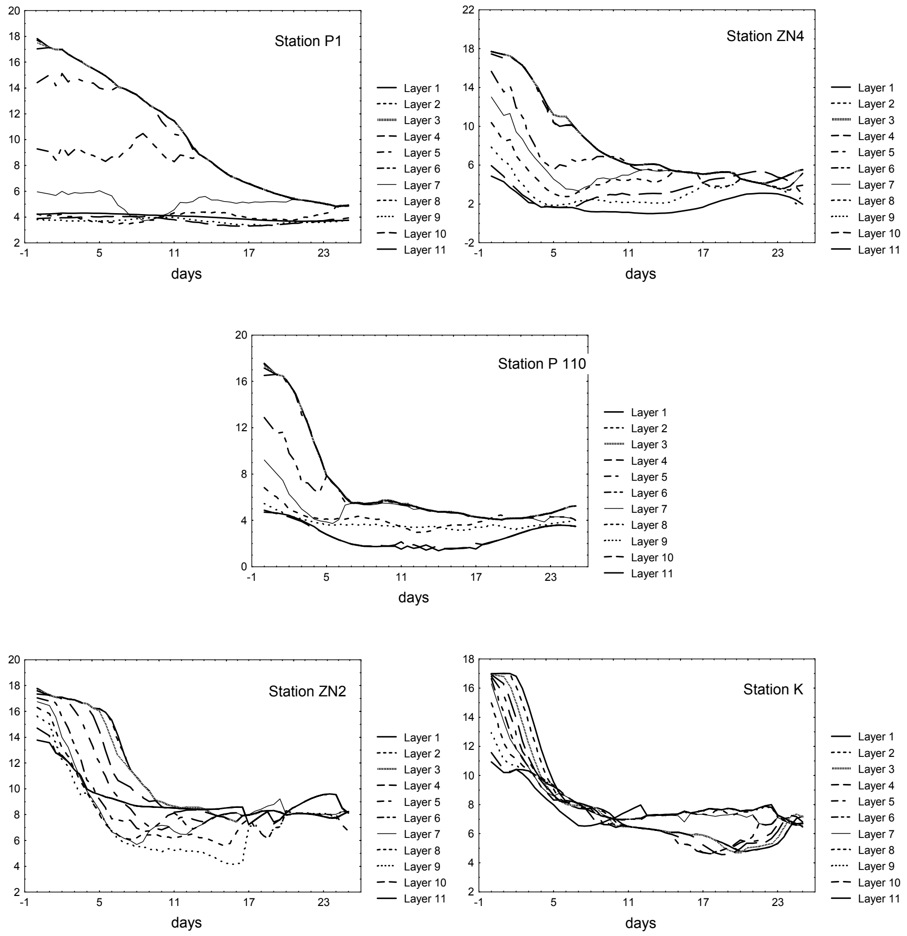

Fig. 11. Modelled mean temperature of water layers at chosen stations in the Gulf of Gdańsk shaped by the field of winds from E

Figures 12–14, left panels refer to thermal cross-sections, right ones to velocity distributions. Solid velocity lines denote positive values of flows toward the North, this means the flow of waters out of the Gulf of Gdańsk. Broken velocity lines represent the flow into the Gulf.

For the first analysed situation (Fig. 12 a), a set of isotherms was obtained indicating the existence of a thermocline. After first day this thermocline was found at the medium depths then dropped considerably to a depth of 80–90 m 25 days later (Fig. 12 c). These temperature cross-sections were associated with velocity cross-sections (Fig. 12 b, d). Both of them exhibited distinct inflows of surface waters into the Gulf (broken lines) and outflows of near-bottom waters (compare with Fig. 10) as well as lateral boundary layers, particularly against the eastern wall.

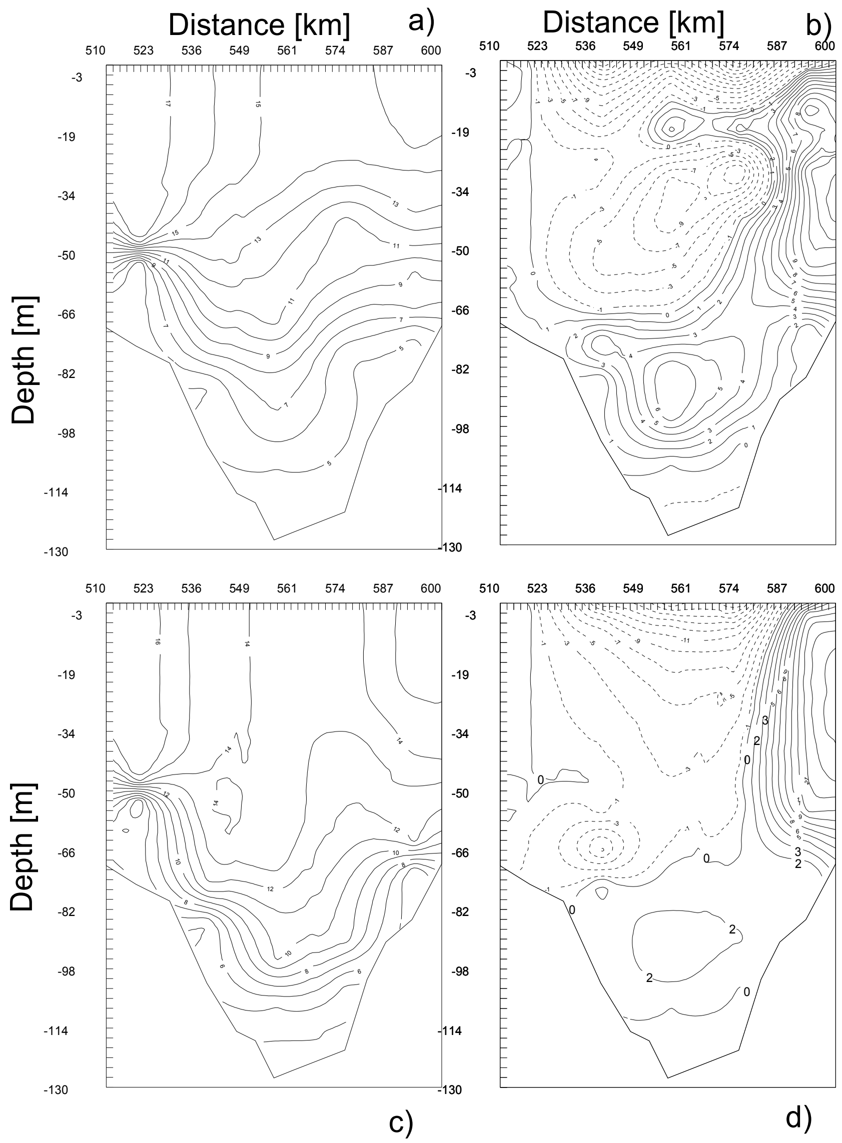

Fig. 12. The cross-sections AA' for the temperature (a), (c) and velocity (b), (d) formed by wind from N after 1 (a), (b) and 25 days (c), (d)

The cross-section flows generated by southerly winds displayed rather opposite features, that is the outflow of surface waters and inflow of bottom ones (Fig. 13 b, d). This led to different temperature distributions compared to simulations involving northerly winds. The beginning of the thermocline was found at 30 m of depth and moved down to 60 m 25 days later (Fig. 13 a, c).

Exceptionally different temperature and velocity cross-sections were

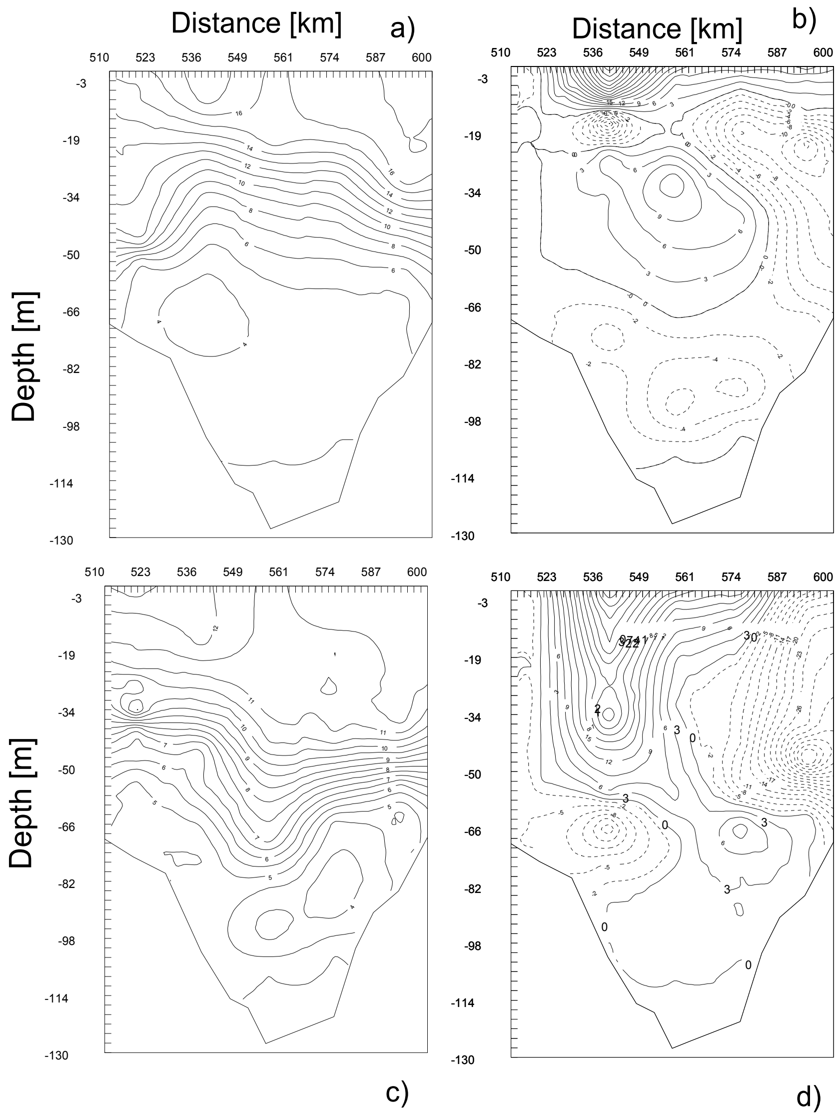

Fig. 13. The cross-sections AA' for the temperature (a), (c) and velocity (b), (d) formed by wind from S after 1 (a), (b) and 25 days (c), (d)

obtained during the action of easterly winds. The warm surface layer moved out of the Gulf through the entire width of the cross-section AA'(Fig. 14b). Through the lower part of this cross-section from 25 m to the bottom, came the strong inflow of cool waters (Fig. 14 a, b). The velocity cross-sections confirmed the water temperature decline at all points taken into account (Fig. 11). These intensive flows, modified over 25 days, formed the outflow ranging from the middle part of the section (from surface to the bottom) and two boundary fluxes

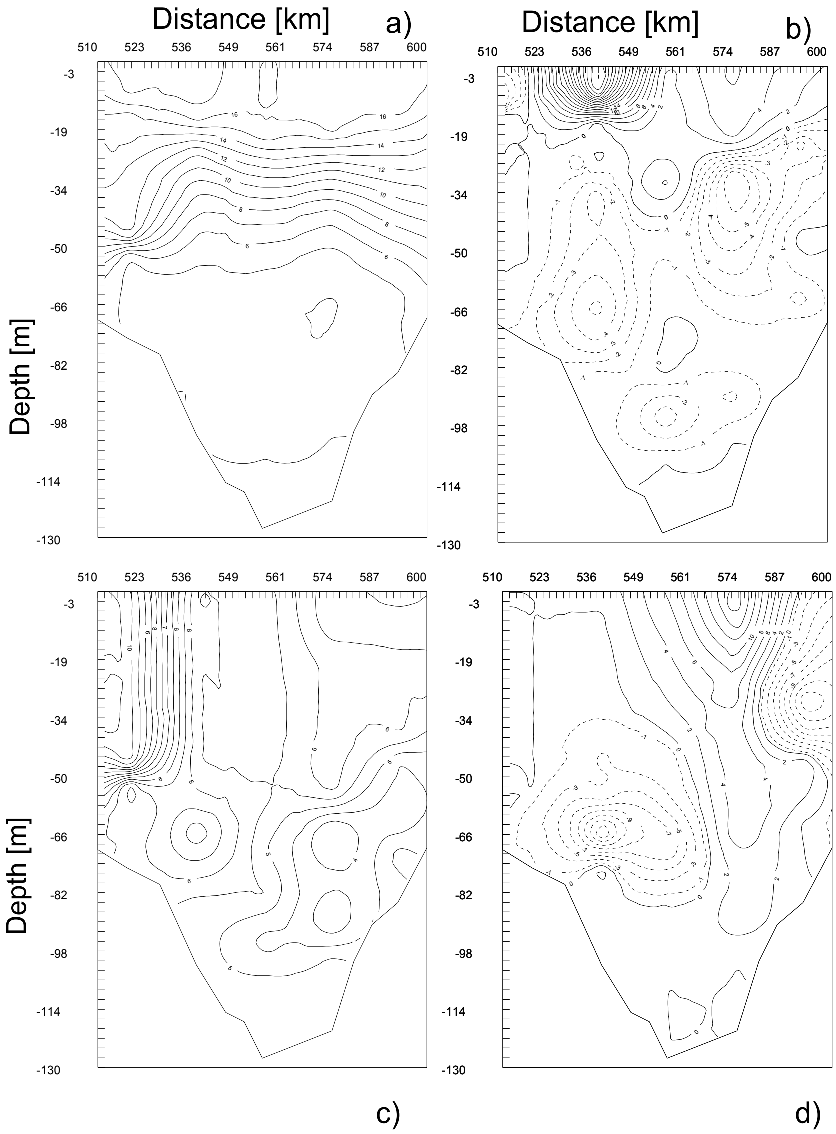

Fig. 14. The cross-sections AA' for the temperature (a), (c) and velocity (b), (d) formed by wind from E after 1 (a), (b) and 25 days (c), (d)

with cores at a depth of 35 m and 65 m (Fig. 14 d). Due to this exchange of waters, the vertical stratification was transformed into horizontal one with clear thermal front (Fig. 14 c).

DISCUSSION

Generally, two groups of results have been obtained. The first one was derived from five-layer model, the second from using eleven-layer model. The results illustrate how the model works when differential wind forcings are imposed. Responses of the model to constant winds from chosen direction appeared as decreasing sea level oscillations (Fig. 2) and fluctuations of total volume around the mean value. The total volume tends to decrease in the case of southerly winds and to increase during easterly winds (compare Figs. 2, 5 and Tab. 1). Considering the openness of the Gulf and the time of simulation shorter than the whole cycle of horizontal water exchange, the mentioned tendencies agree with the law of conservation and unsteadiness of processes.

Southerly winds induced the overbalance outflow out of the Gulf (Tab. 1). In contrast, northerly and easterly winds increased the inflow into it. For the later two cases, the vertical temperature profiles were homogeneous but with temperature 10.5 °C and 5.5 °C for winds from N and E, respectively (Figs. 4, 5). Basin averaged temperature of water confirmed the law of conservation for both constant and diurnal periodical wind stress (Figs. 2, 3). Differential wind forcing caused different evolution of temperature profiles. Gentle climatic August wind (less or equal to 5 $m\,s^{-1}$) did not generate any advection able to destroy the summer stratification (Figs. 6, 8), whilst storm winds can change the spatial distribution of temperature (Figs. 4, 5).

The results of the eleven-layer model described much clearer the vertical thermal structure. They include the calculations of vertical distribution, the mean layer temperature and the cross-sections of velocity. The vertical temperature profiles were correlated with the field of depth. The stratification became stronger and longer in deeper parts of the Gulf (Fig. 7). The structure of vertical temperature distributions depended essentially on the direction of wind. At station ZN2, northerly wind homogenized the waters but easterly maintained the stratification (Figs. 8, 10, 11). Time series of mean layer temperatures were opposite in the case of northerly and easterly winds. When advection caused by winds from N, led to the increase of the temperature of water (Fig. 10), the decrease because of easterly winds took place. Northerly winds brought about the lowering of thermocline (Fig. 12 a, c), easterly allowed to rebuilt the vertical stratification. Only southerly winds maintained the summer stratification at depths of permanent halocline (Fig. 13 a, c).

CONCLUSIONS

The model, which was imposed to the climatic and storm winds, both constant and periodical, showed that its functioning obeys the law of conservation. In studies on temperature fields in the Gulf of Gdańsk, the influence of advection caused by wind is noteworthy. The results presented elsewhere (Krauss and

Brügge 1991, Jankowski 1997) indicated that the oceanic waters can advect from the Bornholm to the Gdańsk Deep via the Słupsk Channel due to winds from E. The results described in this paper provide the evidence that the character of exchange of waters in the Gulf of Gdańsk depends on the speed and direction of wind forcing. The temperature distribution used as indicator of circulation appears to be a sensible choice for research on outflows of surface as well as inflows of bottom waters. Excluding the atmospheric heat fluxes made it possible to demonstrate the role of the advection in forming the temperature field. The surface waters were shown to move out of the Gulf while deep waters to be directed into it. The analyses of thermal changes demonstrated also the essential influences of advection on spatial distributions of temperature in the Gulf. It is revealed that the eastern winds represent a natural factor causing the renewal of the Gulf of Gdańsk waters, which is important from ecological point of view.

At this stage, it may be concluded that it would be reasonable to extend tracing to the salinity and oxidation from the Słupsk Channel to the Gulf of Gdańsk. It seems also important to evaluate the anemobaric conditions which may favour restoring the Gulf waters.

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