

## Assimilation experiments in the Gdansk Basin (the south-eastern Baltic): How useful are the sea level data from coastal tide gauges for modelling of hydrophysical fields?

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### Abstract

In the Baltic modelling research, assimilation techniques were developed with advance. They were concerned to model assimilated basic parameters and observed them directly. In present paper, the most important was the assimilation of surface information and its projection deep into temperature and salinity fields. In oceanic investigations altimetry viewed from satellite was the sea level changes projected far inside and predetermined surface-to-subsurface correlations. To obtain improved modelled hydrophysical fields, sea level variations measured at coastal gauges and efficient data assimilation were taken into account. A data assimilation algorithm has been developed and used in conjunction with a three-dimensional baroclinic model of the Baltic Sea. It was based on a time and space weighted nudging technique. The sea level data were inserted

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continuously by updating the model solution every time step. Several sensitivity experiments with different values of time and spatial weighting scales were performed. In first series of experiments, only sea level data (SL) were assimilated. In the next simulations, seawater temperature (SWT) and seawater salinity (SWS) related directly to SL were assimilated. To evaluate the effectiveness of the assimilation scheme, modelled sea level series and vertical profiles of seawater temperature and salinity in selected coastal gauges in the Gdansk Basin were examined. Evidently low but statistically essential correlation coefficients indicated nonlinear character of vertical mixing and transfer processes. Decreasing errors obtained while comparing the model results to a control case without assimilation confirmed a real transfer of surface information deep and usefulness of such approach in modelling.

## INTRODUCTION

The state and variability of coastal areas require effective studies. One of the ways to estimate the current state of the coastal waters was to use *in situ* measurements. The difficulties appeared due to the fact that in general the observing data were distributed very irregularly in space and time. Another way was to run numerical models, which use the knowledge about the governing laws of the marine system. Modelled simulations, however, give only an approximation of the real marine environment and may have other deficiencies, arising from errors in the model formulation and forcing fields. Furthermore, in this case no use was made with information contained in the measured data.

The promising approach to obtain a complete and reasonable estimation was based on assimilation of the *in situ* data into the model when available. The model extrapolates the information in space and time. So, even in regions where no data exist, a reasonable estimation of the marine environment state would be available. Data assimilation allows to provide better estimates of nature than can be obtained by using only the observational data or the dynamical model (Robinson et al. 2002).

Several assimilation methods (cf. Ghil and Malanotte-Rizzoli 1991, Kantha and Clayson 2000) were developed. Recently, some new ones, e.g. Kalman filtering (Ghil 1989) or adjoint method (cf. Le Dimet and Talagrand 1986) were created. Thus, more economical techniques appeared as desirable, e.g. successive correction method (Daley 1991) and its applications (cf. Fischer and Latif 1995). The other one, more frequently used, was based on a simple Newtonian relaxation or nudging method (cf. Malanotte-Rizzoli and Young 1992, 1995; Li et al. 2003).

Some of the above mentioned assimilation techniques were applied in the numerical models of the Baltic Sea. The adjoint method was used by Meier (cf. Meier and Krauss 1994, Meier 2004) to assimilate sea level and wind data into model to successfully optimise model forcings (e.g. wind fields) on time scales from one day up to several months.

The nudging method with iterative procedure was applied in the Data Assimilation System (DAS) (Sokolov et al. 1997) to assimilate hydrological and hydrochemical data into numerical model in order to construct 3D gridded oceanographic database of the Baltic Sea. Lehmann (2004) has been tested the nudging scheme to assimilate temperature and salinity profiles into a model which was used to reconstruct migration pathways of eastern Baltic cod. An optimal interpolation (OI) method was included in operational system for the Baltic Sea based on parallel version of HIROMB (cf. Funkquist and Pemberton 2006; Pemberton and Funkquist 2006) to assimilate temperature and salinity profiles as well as sea surface temperature (SST). Funkquist (2004) presented some results of application of successive correction method (SCM) to assimilate SST into the HIROMB system. In Larsen et al. (2007), a hybrid data assimilation scheme to assimilate satellite sea surface temperatures (SST) into an operational ocean model has been developed and validated against *in situ* observations. The scheme included a 2D optimal interpolation (OI) part to estimate background field and with simplified Kalman filter (KF) part. A steady Kalman gain formulation was used from a full ensemble Kalman filter in a dynamically simplified model (Sørensen et al. 2003).

Assimilated values of environmental parameters at one location have a certain amount of correlation with the values observed in the closest neighbourhood. The correlation length scale (CLS) was an estimate of how fast the correlation between quantity of parameter at one location and the one taken nearby decreases with distance. To estimate the CLS, the observations from each station are correlated with ones from all the other stations. The CLS is defined as the distance that the mean correlations drop below  $1/e$  (Belousov et al. 1971, She and Nakamoto 1996). Lots of work remained to improve the variable correlation length scale in the horizontal as well as in the vertical. The alongshore correlation length scales were much longer than the cross-shelf correlation length scales (Tynana et al. 2005). Kundu and Allen (1976) using moored velocity data, found alongshore and cross-shelf correlation scales of at least 30 km. The North Sea non-tidal sea surface height variations have a characteristic time scale of one day or less, and characteristic length scales of 300-500 km along the coast and 100-200 km perpendicular to the coast (Sørensen et al. 2003).

The characteristic time scale of sea level variations at Spikarna (the Baltic) was 17 days compared to 2 - 4 days in the Inner Danish Waters and one day at Esbjerg (the North Sea). In both Northern and Baltic Seas there existed a significant inhomogeneity, spatial scales larger than 200 km were found in the southern North Sea (in both x and y directions) and in the eastern Baltic Sea (in y-direction only). The largest time scales (greater than 2.5 days) were found in the northern and eastern Baltic Sea whereas medium time scales (1.5 - 2.5 days)

were noted in the southern North Sea, and small scales (lower than 1.5 days) in the western Baltic Sea - transition waters (She et al. 2007). Proper values of the horizontal length scales referred to temperature and salinity in the Baltic seemed to lie between 20 and 50 km, while the vertical ones may vary from only a few meters up to 50 m in the deepest parts (Pemberton and Funkquist 2006).

In present paper the length scales have been estimated as the average of the autocorrelation in the four directions, N, E, S and W. The time ones referred to e-folding scales of autocorrelation function ranged from 4 hours to 2 days while the spatial scales were 7 – 40 km.

The present paper comprises modelling of hydrophysical fields using of projection sea level information by surface-to-subsurface correlation into deep waters. In oceanic research, surface information was provided mostly by satellite altimetry. In this investigation gauges measured sea level variations in the Gdansk Basin were applied. Assimilation of these data by the 3D hydrodynamic model made insight into the projection of sea level into deep sea.

The aim of this paper was to estimate the influence of sea level changes on subsurface fields of temperature and salinity. The paper was organised as follows. Introduction was the first section, and then the model was described. An important part was the consideration of assimilation scheme both of sea level data as well as assimilation of tracers, temperature and salinity. The results obtained were discussed in the next paragraph and finally, conclusions were presented.

## MODEL DESCRIPTION AND MATERIALS

In this study hindcast simulations by the 3D sigma-coordinate Baltic Sea model with a three-dimensional assimilation scheme was tested using the sea level data from tide gauges located at the Polish coast<sup>2</sup>. The model results were compared with the *in situ* measurements of vertical soundings of temperature and salinity collected during the cruises of r/v 'Oceania' in September 1995<sup>3</sup>. All numerical simulations were performed using three months (August - October 1995) forcing data from the PIDCAP'95 experiment<sup>4</sup>. Assuming that all available measurements were quasi-synoptic, temperature and salinity have been merged into the model node as the inverse distance weighted mean inversely proportional to the distance from observation hydrographic station to model point.

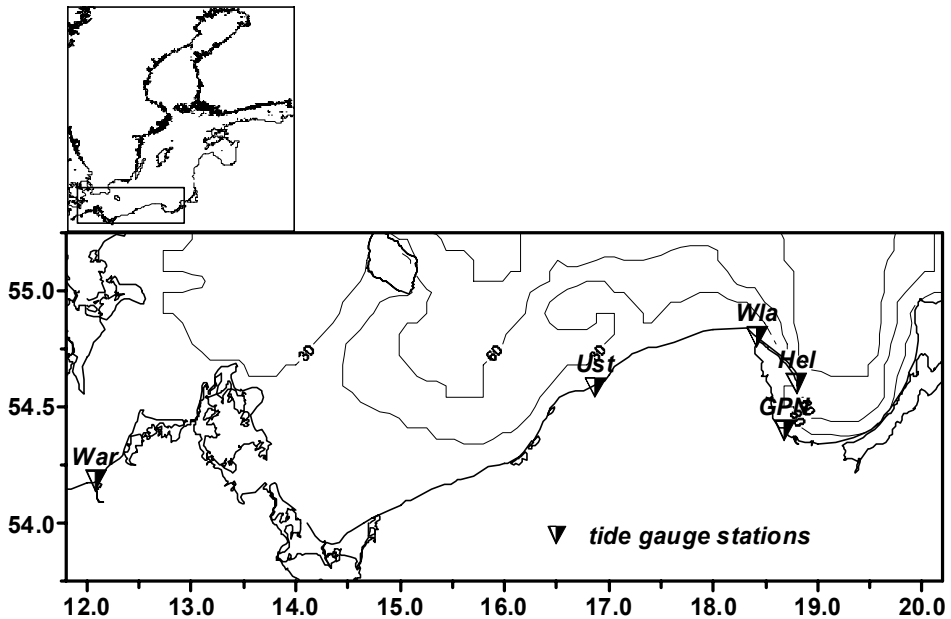
The model was based on code from the three-dimensional (3D) sigma -

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<sup>2</sup> Recorded *in situ* sea level time series at five coastal gauges: Warnemünde (War), Ustka (Ust), Hel (Hel), Władysławowo (Wla) and Gdansk\_PN (GPN) were assimilated (Fig. 1).

<sup>3</sup> Data were available from the Regional Oceanographic Database of IO PAS-<http://www.iopan.gda.pl>

<sup>4</sup> Pilot Study for Intensive Data Collection and Analysis of Precipitation



**Fig. 1.** Network of the sea level measurement stations at the eastern part of the Polish coast (Gdansk Basin) and location of five tide gauge stations used in assimilation experiments. Bottom topography based on Seifert and Kayser (1995). Numbers on isolines -depth in meters.

coordinate Princeton Ocean model (POM) of Blumberg and Mellor (1987) adapted to the Baltic Sea (Jankowski 2002a). The model area covered the whole Baltic with a horizontal resolution of ca. 5 km and 24 sigma - levels in the vertical. Simplified boundary conditions of the radiation type were applied at the open boundary of the model at the Skagerrak. The model bottom topography was elaborated on the base of the data from Seifert and Kayser (1995). The model with such horizontal and vertical resolution enabled to describe thermohaline fields and features of mesoscale current of 10 km order. To date, the model has been demonstrated as reliable for simulations of the basic features of water circulation using hydrology (Jankowski 2002a) and the upwelling events in September 1989 (Jankowski 2002b). A detail description of the POM model related to the Baltic conditions could be found in Jankowski (2002a,b).

The numerical simulations were initiated with 3D climatological fields of temperature and salinity, which were constructed on the base of the monthly mean (multi - year averaged) maps taken from Bock's (1971) and Lenz's (1971)

atlases. The annual inflows of 31 main rivers were taken into consideration. The wind fields were estimated from the atmospheric surface pressure charts. Wind stress components and surface heat fluxes were evaluated by the bulk formula (Large and Pond 1982, Launiainen and Vihma 1990).

The climatic forcing was calculated from the monthly mean surface charts of temperature and salinity from Bock's (1971) and Lenz's (1971) atlases. They have been linearly interpolated in time with an interval equal to the model internal time step. This forcing was coupled to the model by so-called method of relaxation towards climatology (cf. Lehmann 1995, Oey and Chen 1992). It has been found that these additional climatological heat fluxes (due to relaxation to climatology) were useful to drive our model in the case when calculated realistic fluxes were very small or impossible to estimate from meteorological data or absent altogether.

The model simulations were performed in two stages. The first one (20 days) - to initialise the model computations by climatological forcing. In the second stage, basing on the previous step results, a fully prognostic run was started. Real atmospheric forcing (atmospheric pressure, winds and heat fluxes) drove the model for 92 days, from 1<sup>st</sup> of August to 31<sup>st</sup> October 1995. In the simulations the salinity flux at the sea surface was neglected.

In order to reduce (suppress) these numerically induced oscillations, the meteorological forcing at all nodes of model domain was modified by its linear increase from zero to a realistic value during several days of prognostic run. In our case it was done for the first two days.

## ASSIMILATION SCHEME

Assimilation schemes are often based on time and spatial weighted nudging method (cf. Malanotte-Rizzoli and Young 1992, 1995; Li et al. 2003). In the present work a scheme by Mellor and Ezer (1994) was applied. It made use of statistical relations between sea level (SL) and seawater temperatures (SWT) as well as seawater salinity (SWS). The sea level data of different time resolution was inserted continuously updating the model solution at every time step. Preprocessed correlation between SL and subsurface SWT and SWS anomalies were used to project surface information into deeper layers. Thus, besides continuous SL insertion into the model, "temperature" and "salinity" directly related to SL were also assimilated. SL introduced into the model improved calculated seawater elevation and modified other model variables, e.g. the velocity components; that in consequence changed 3-D fields of temperature and salinity. In this way a signal from the sea surface may be transferred into deep water layers and improve calculated temperature and salinity fields.

### Assimilation of sea level data (SL)

The nudging scheme used for assimilation of sea level data from the coastal tide gauge stations was adapted from the scheme of Malanotte-Rizzoli and Young (1992), which involved spatial and time weighting factors. In our scheme sea level was assimilated by introduced nudging term on the right-hand side (RHS) of the model continuity equation:

$$\frac{\partial \eta}{\partial t} = RHS + G_{\eta}(t)G_{x,y}(x,y)(\eta - \eta_{obs}) \quad (1)$$

where:

RHS - stands for all other terms in continuity equation,  $\eta, \eta_{obs}$  - the modelled and observed sea level values,  $G_{x,y}, G_{\eta}$  - spatial and weighting functions:

$$G_{x,y} = \exp(-[(x - x_o)^2 + (y - y_o)^2] / R^2) \quad G_{\eta} = \frac{1}{\alpha_{\eta}} = const \quad (2)$$

where:

$R, \alpha_{\eta}$  - spatial and time scales which were assumed to be equal to const.  $x, y$  - station coordinates of grid point,  $x_o, y_o$  - tide gauge station coordinates

### Assimilation of "temperature" and "salinity"

The temperature and salinity were inserted into the model in a similar way as sea level, i.e. by introducing nudging term on the right-hand side (RHS) of the model transport equation for tracer (temperature or salinity):

$$\frac{dF}{dt} = RHS - G_F(x,y)G_{Fz}(\sigma,t)(F - F_{obs}) \quad (3)$$

where:

RHS - stands for all other terms in tracer equations  $F, F_{obs}$  - the modelled and observed tracer values (seawater temperature or salinity),  $G_F(x,y), G_{Fz}(\sigma,t)$  - horizontal, vertical and weighting functions calculated as follows:

$$G_F(x,y) = \exp(-[(x - x_o)^2 + (y - y_o)^2] / R^2) \quad (4)$$

$$G_{Fz}(\sigma,t) = \frac{1}{\beta_{\sigma}} + \left( \frac{1}{\alpha_{\sigma}} - \frac{1}{\beta_{\sigma}} \right) \exp\left(\sigma \frac{H}{D}\right) \quad -1 \leq \sigma \leq 0 \quad (5)$$

where:

$x, y$  – station coordinates of grid point,  $x_o, y_o$  - tide gauge station coordinates

$R$  - spatial weighting radius,  $\sigma = \frac{z - \eta}{H + \eta}$  -  $\sigma$  coordinate;  $z$  and  $H$  - vertical

coordinate and sea depth, respectively,  $\alpha_{-F}, \beta_{-F}, D = 20m$  - numerical constants, parameters in vertical- weighting function  $G_{Ftz}(\sigma, t)$

In order to define the values of tracers  $F_{obs}$  we have applied an approach based on Mellor and Ezer (1991) assimilation scheme, which used statistical links between sea level and subsurface tracer value  $F$ , i.e. temperature ( $T$ ) and salinity ( $S$ ). As it was suggested in the above-mentioned paper, these statistical relations can be estimated from the model simulation results performed prior to assimilation experiments. In our case the results of calculations for three months period were used. Time series of the modelled sea level as well as sea water temperature and salinity at all 24  $\sigma$ -levels at the grid point nearest to the location of tide gauge station were stored.

Next, the correlation factors ( $f_{-T}, f_{-S}$ ) and correlations coefficients ( $c_{-T}, c_{-S}$ ) between sea level  $\eta$  and sea water temperature  $T$  and salinity  $S$  at all sigma levels were estimated as follows:

$$f_{-T} = \frac{\overline{\delta T \delta \eta}}{(\overline{\delta \eta})^2} \quad f_{-S} = \frac{\overline{\delta S \delta \eta}}{(\overline{\delta \eta})^2} \quad (6)$$

$$c_{-T} = \frac{\overline{\delta T \delta \eta}}{[(\overline{\delta \eta \delta T})^2]^{\frac{1}{2}}} \quad c_{-S} = \frac{\overline{\delta S \delta \eta}}{[(\overline{\delta \eta \delta S})^2]^{\frac{1}{2}}} \quad (7)$$

where:

$\delta \eta = \eta - \langle \eta \rangle$ ,  $\delta T = T - \langle T \rangle$ ,  $\delta S = S - \langle S \rangle$ ,  $\langle \rangle$ - time averaging,  $\delta \eta$ ,  $\delta T$ ,  $\delta S$  - sea level, temperature and salinity fluctuations, respectively,  $\eta, T, S$  - calculated sea level, temperature and salinity,  $\langle T \rangle, \langle S \rangle$  - time averaged value of temperature or salinity.

Based on the run results without reference b0 assimilation, correlation factors ( $f$ ) and correlation coefficients ( $c$ ) between sea level ( $\eta$ ) and tracers ( $F$ ) at selected sigma levels at the grid point nearest to gauge station were calculated. All statistical parameters were estimated separately for each month of simulation period. Next, knowing monthly-mean values of each tracer ( $T, S$ ), unknown correlation factors ( $f_{-T}, f_{-S}$ ) related to sea level and tracer fluctuations as well as their values at all sigma - depth levels were estimated:

$$\delta T = f_{_T} \delta \eta \quad \delta S = f_{_S} \delta \eta \quad (8a)$$

$$T_{obs} = T_{mean} + \delta T = T_{mean} + f_{_T} \delta \eta \quad (8b)$$

$$S_{obs} = S_{mean} + \delta S = S_{mean} + f_{_S} \delta \eta \quad (8c)$$

where:

$T_{mean}$ ,  $S_{mean}$ - time averaged values for each month from control run

Exemplary values of estimated correlation factors and correlation coefficients between sea level and temperature ( $f_{_T}$ ,  $c_{_T}$ ), and sea level and salinity ( $f_{_S}$ ,  $c_{_S}$ ) for the points close to the tidal gauge in Hel were presented (Tab. 1). Positive correlation indicated that temperature and salinity increased due to the growth of sea level and negative one – opposite, temperature and salinity decreased when sea level was lowering. The best coefficients were found in September, the worse in October. In the surface layer correlation for

**Table 1**

Correlation factors ( $f_{_T}$ ,  $f_{_S}$ ) and correlation coefficients ( $c_{_T}$ ,  $c_{_S}$ ) for sea level - temperature and salinity relations calculated for data at Hel station (Tide gauge station location - Fig. 1)

Month	Sigma level	Depth [m]	$f_{_T}$	$f_{_T}$	$f_{_S}$	$c_{_T}$
<b>Aug</b>	-1	0.05	-0.0496	0.0000	-0.4557	0.0044
<b>Aug</b>	-3	0.28	-0.0456	0.0000	-0.5075	0.0019
<b>Aug</b>	-5	1.12	-0.0368	-0.0001	-0.6101	-0.0115
<b>Aug</b>	-10	8.68	-0.0090	-0.0018	-0.1975	-0.3897
<b>Aug</b>	-15	16.58	0.0165	-0.0038	0.2654	-0.5767
<b>Aug</b>	-20	24.47	0.0471	-0.0060	0.5487	-0.6271
<b>Aug</b>	-23	29.21	0.0315	-0.0057	0.3543	-0.5353
<b>Sep</b>	-3	0.28	0.0525	-0.0060	0.5496	-0.5083
<b>Sep</b>	-5	1.12	0.0520	-0.0060	0.5611	-0.5080
<b>Sep</b>	-10	8.68	0.0493	-0.0056	0.5606	-0.5031
<b>Sep</b>	-15	16.58	0.0448	-0.0059	0.4989	-0.5287
<b>Sep</b>	-20	24.47	0.0532	-0.0082	0.6684	-0.6293
<b>Sep</b>	-23	29.21	0.0569	-0.0099	0.7695	-0.6398
<b>Oct</b>	-1	0.05	-0.0088	0.0028	-0.0895	0.3114
<b>Oct</b>	-3	0.28	-0.0085	0.0028	-0.0878	0.3116
<b>Oct</b>	-5	1.12	-0.0076	0.0028	-0.0814	0.3116
<b>Oct</b>	-10	8.68	-0.0045	0.0027	-0.0545	0.3132
<b>Oct</b>	-15	16.58	0.0005	0.0021	0.0072	0.2741
<b>Oct</b>	-20	24.47	0.0154	-0.0002	0.2403	-0.0190
<b>Oct</b>	-23	29.21	0.0167	0.0003	0.2876	0.0184

temperature was negative whereas that for salinity was positive. In August, the surface layer was characterized by negative correlation for temperature and positive for salinity. On the other hand, an opposite situation was noted in the bottom layer. In September correlation for temperature was positive and for salinity – negative. Next month it happened the other way round - negative correlation was for temperature and positive for salinity. The values of correlation coefficients were not too high, however, most of them were greater than 0.15 at 95.0 significance level (Emery and Thompson 1997). This indicated that connections between sea level, and temperature and salinity were essential. Small values of correlation coefficients indicated non linear influence between sea level changes, and the changes of  $T$  and  $S$ . Knowing the evaluated factors ( $f_T$ ,  $f_S$ ) we were able to present temperature and salinity fluctuations determined by assimilated variations of sea level measured at the coastal stations (cf. eq. 8).

## RESULTS AND DISCUSSION

Several numerical experiments were performed. The first series included only sea level (SL) data assimilation, then seawater temperature (SWT) and seawater salinity (SWS) anomalies resulting from the propagation of surface information through sea levels were assimilated. The experiments with different values of time and spatial weighting scales were carried out to test the effects of the assimilation scheme parameters for modelled hydrophysical fields. Spatial weighting radii of 6, 16, 26 and 30 km were assumed. The smallest radius allowed the transfer of information into the next node of grid field, which was ca. 5 km. The next one was bigger than three grid steps. The largest ones exceeded five spatial grid steps but they were smaller than the radius of the Gulf of Gdansk. Selection of time scales was based on presently available sea level observations, which were characterized by different resolutions, from 1 hour in Warnemunde via 4 hours in Ustka to 24 hours at the rest tidal gauges. In future, higher resolution data (time step of 10 minutes) on sea level fluctuations would be available, e.g. from the IOUG network.

Information on spatial scale of R variability is necessary to estimate prognostic time scales. Therefore, the autocorrelations of the surface elevation have been calculated at several grid points in the study area. The spatial length scales were calculated as the average of the autocorrelation in the four directions. The time scale has been estimated as an e-folding scale of autocorrelation function in control run b0. For each selected point the e-folding time and space (length) scales were estimated. Time scales ranged from 4 hours to 2 days while the spatial scales were: 10 - 22 km (E), 15-30 km (N), 7-12km (S), and 10-40 km (W).

### Sea level assimilation experiments

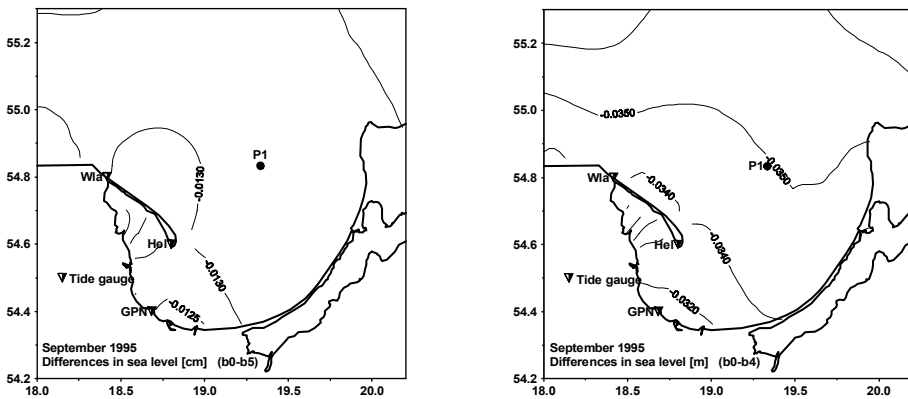
**Table 2**

We started with the sea level data assimilation from the coastal tidal stations. The assimilation scheme described above depended on two principal weighting factors – spatial  $R$  and time  $\alpha_\eta$  scales (eq. 1). Thus, in order to estimate the impact of these scales on modelled sea level, five experiments (b1 to b5) have been performed. Firstly, evaluation of time scale influence was conducted in two runs, b1 and b2. In both cases the same spatial radius was assumed ( $R = 30$  km), and the values of time scales were different ( $\alpha_\eta$  was equal to 48 and 4 hours in b1 and b2 experiments, respectively) (Tab. 2).

Basic parameters used in assimilation of sea level experiments for selecting  $\alpha_\eta$  temporal and spatial  $R$  scales

Exp.	$R$ [km]	$\alpha_\eta$ [hours]
b1	30	48
b2	30	4
b3	26	4
b4	16	4
b5	6	4

To compare the obtained results, some statistical measures were used. The run b2 showed a decrease of bias ( $ae$ ), absolute bias ( $aae$ ) and root mean square error ( $rmse$ ) from one side, and a significant growth of explanation of variance ( $exp.var$ ) from the other (Tab. 3a). Assimilation of the observed sea levels with time scale of 4 hours (relaxation time) improved the variance from 30 to 81% compared to experiment b0 (without assimilation). Moreover, the errors between model results and observations in the experiment b2 indicated that its results were closer to sea level recorded at all four tidal gauge stations than in the experiment b1 (cf. Tab. 3a). Therefore, in the other experiments on sea level data assimilation a time scale ( $\alpha_\eta$ ) of 4 hours was chosen.



**Fig. 2.** Monthly mean sea level differences between results of the runs: control one b0 and b5 (left) and b0 and b4 (right)

**Table 3**

**A.** Estimated statistical measures of the modeled with the assimilation data versus observed sea level at four stations in experiments b1 and b2 (averaged error (bias) - *ae*, absolute averaged error - *aae*, root-mean square error - *rmse* and explained variance - *exp.var*)

Exp.	Sta. [cm]	<i>ae</i> [cm]	<i>aae</i> [cm]	<i>rmse</i> [cm]	<i>exp var</i> [%]	Sta.	<i>ae</i> [cm]	<i>aae</i> [cm]	<i>rmse</i> [cm]	<i>exp. var</i> [%]
<b>b0</b>	<b>Hel</b>	3.57	11.35	14.96	36.96	<b>Ust</b>	-0.39	13.47	16.17	44.08
<b>b1</b>	<b>Hel</b>	3.15	9.12	11.99	53.70	<b>Ust</b>	-0.84	11.07	13.67	55.79
<b>b2</b>	<b>Hel</b>	0.98	5.06	7.17	81.37	<b>Ust</b>	-3.02	7.60	10.33	77.05
<b>b0</b>	<b>GPN</b>	7.56	12.81	17.51	34.04	<b>Wla</b>	0.45	13.18	15.85	35.89
<b>b1</b>	<b>GPN</b>	7.10	10.90	14.65	49.90	<b>Wla</b>	0.04	10.71	12.93	51.33
<b>b2</b>	<b>GPN</b>	4.87	6.77	9.19	79.41	<b>Wla</b>	-2.02	6.74	8.55	78.49

**B.** Estimated statistical measures of the modeled with the assimilation data versus observed sea level at four stations in experiments b3 – b5, (averaged error (bias) - *ae*, absolute averaged error - *aae*, root-mean square error - *rmse* and explained variance - *exp.var*)

Exp.	Sta. [cm]	<i>ae</i> [cm]	<i>aae</i> [cm]	<i>rmse</i> [cm]	<i>exp var</i> [%]	Sta.	<i>ae</i> [cm]	<i>aae</i> [cm]	<i>rmse</i> [cm]	<i>exp. var</i> [%]
<b>b3</b>	<b>Hel</b>	1.02	5.29	7.46	79.58	<b>Ust</b>	-2.71	7.72	10.48	75.50
<b>b4</b>	<b>Hel</b>	2.22	6.90	9.27	69.84	<b>Ust</b>	-1.72	8.83	11.61	67.40
<b>b5</b>	<b>Hel</b>	3.55	10.25	13.48	45.20	<b>Ust</b>	-0.44	12.26	14.86	49.91
<b>b3</b>	<b>GPN</b>	4.89	6.99	9.52	77.32	<b>Wla</b>	-1.78	6.93	8.81	76.69
<b>b4</b>	<b>GPN</b>	6.10	8.80	11.76	66.41	<b>Wla</b>	-0.81	8.29	10.35	66.80
<b>b5</b>	<b>GPN</b>	7.50	11.92	16.11	41.83	<b>Wla</b>	0.42	11.96	14.35	43.47

When time scale was assumed as constant, the spatial ones (the radii) were subjected to examination. In the next three experiments (b3, b4, and b5), the values of the radii ( $R$ ) were 26, 16, and 6 km, respectively, and their influence on the sea level data assimilation was evaluated. Statistical criteria described these runs univocally, the three first measures (*ae*, *aae*, *rmse*) increased with a decrease of the radii (cf. Tab. 2 and Tab. 3b). At the same time, the explained variances decreased from 75.5% – 79.58% to 41.83% - 49.91%. These results indicated that when the spatial scale decreased, propagation of information from the coast was getting smaller (Fig. 2). The run b3 with  $R$  equal to 26 km presented the influence of assimilated coastal sea level variations on the sea surface elevations quite away from shore gauges (not shown). This impact was

very limited as compared to run b0 (without assimilation) (Fig. 2).

Time and spatial scales built in assimilation scheme resolve the share of information propagated from the coastal stations on modelled domain. How was sea level changed due to the influence of both spatial and time scales during assimilation at the station P1 located at the open waters of the Gdansk Basin? The answer was given on the base of previously used statistical criteria (*ae*, *aae*, *rmse*, *exp.var*) evaluated for the differences of simulated sea level at P1 during runs b0 and b1 – b5 (Tab. 4). Small values of *ae*, *aae*, *rmse* and a great one of the *exp.var* indicated lack of assimilation influence on propagation changes from the coast. On the other hand, bigger values of errors and smaller *exp.var* reflected assimilated variations from the shore stations.

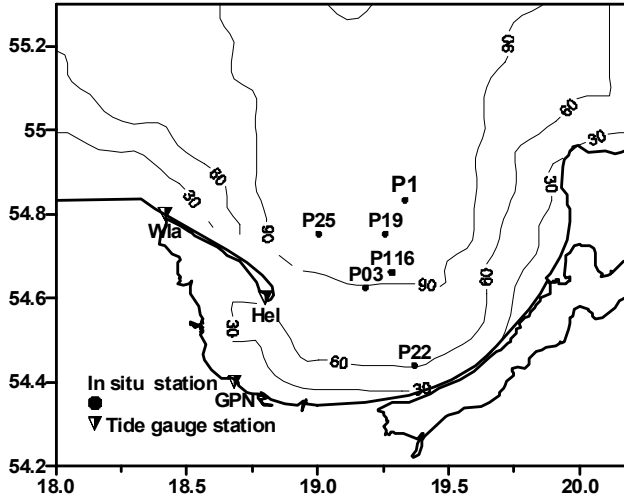
**Table 4**

Estimated statistical measures: averaged errors (bias) - *ae*, absolute averaged error - *aae*, root-mean square error - *rmse* and explained variance - *exp.var* for differences between runs b0 and b1- b5 at station P1

Exp.	<i>ae</i> [cm]	<i>aae</i> [cm]	<i>rmse</i> [cm]	<i>exp.var</i> [%]
<b>b0 - b1</b>	-0.49	3.03	3.95	94.23
<b>b0 - b2</b>	-2.76	8.59	10.98	60.26
<b>b0 - b3</b>	-2.40	8.27	10.51	62.77
<b>b0 - b4</b>	-1.38	6.06	7.79	77.88
<b>b0 - b5</b>	-0.05	1.55	1.97	98.59

The strongest influence of sea level assimilation from coastal gauges was noted during run b2 and b3, when the spatial radii were 30 and 26 km, respectively. To the contrary, the smallest influence appeared in two other runs, b1 and b5 (cf. Tab. 4 and Tab. 2). In b1 experiment  $R$  was equal to 30 km and  $\alpha_\eta$  - 48 hours, whereas in b5  $R = 6$  km and  $\alpha_\eta = 4$  hours. Therefore, the optimal scales in the assimilation of sea level seemed to be the following:  $R = 16$  km for the spatial scale and  $\alpha_\eta = 4$  hours for the time one. During assimilation, sea levels measured at the coastal stations were propagated with those scales from the coast into the open sea.

The spatial influence of assimilated sea level data from coastal tidal stations was observed far away from their locations. Impact of the assimilated information from coastal tide gauges was correlated with the value of spatial weighting radius  $R$  (Fig. 2). Statistical measures between the time series for the control run b0 and the other ones for the open sea station P1 (Fig. 3) confirmed these findings (Tab. 4).



**Fig. 3.** The study area and location of the selected stations in the Gdansk Basin (numbers on isolines - depth in meters).

It is worth noting that the evaluated bias ( $ae$ ) was negative for the stations situated in Władysławowo and Ustka, and positive for the ones located in Hel and Gdansk. The first tidal gauges were located at the coast of the open sea whereas the others – in the Gulf of Gdansk. The influence of assimilated sea level data resulted in greater horizontal gradients of calculated sea level in the western part of the Gulf of Gdansk (Fig. 2). This impact came from sea level variations at the tidal gauges in Władysławowo, Hel and Gdansk placed at the western coast of the gulf. The other feature of these results was that the influence of assimilated variable affected the whole Gdansk Basin (cf. Fig. 2).

### **“Temperature” and “salinity” assimilation experiments**

Next, we assimilated both sea level and the “observed tracer” anomaly. Assimilated variations of sea level propagated from the coast into the open sea as well as they were transferred into the natural tracers, i.e. vertically distributed temperature and salinity. As shown earlier (eqs. 8a-c), the “observed tracer” values, “temperature” and “salinity”, were related directly to the assimilated sea level elevation  $\eta$ . The assimilation scheme (eqs. 3 - 5) depended on three principal weighting factors – spatial ( $R$ ) and time ( $\alpha_F, \beta_F$ ) scales. Thus, in order to estimate the impact of these scales on modelled seawater temperature and salinity, five experiments (c1 to c3 and d1, d2) were performed. As it had been shown earlier, time scale applied in the assimilation of sea level ( $\alpha_\eta = 4$

hours) appeared to be the optimal one. To assimilate “tracers” (eqs. 3 and 5), two time scales,  $\alpha_F$ ,  $\beta_F$ , were defined. The scales used in simulation runs were depicted in Tab. 5. Thermal processes to be transferred from the surface to deeper layers need from half a day to two days, and it takes one order more time when they are initiated from the bottom. The scales referring to salinity are one order larger, even up to 200 days. The last of them were relaxation time scales based on empirical considerations (Sarmiento and Bryan 1982). In runs c1-c3, spatial scales  $R$  for “tracers” and sea level were assumed as 26 km, 16 km and 6 km whereas in runs d1 and d2 spatial scale  $R$  was equal to 16 km (cf. Tab. 5). The same spatial scales  $R$ , 26-6 km and 16 km, were applied in runs b3-b5 and b4, respectively.

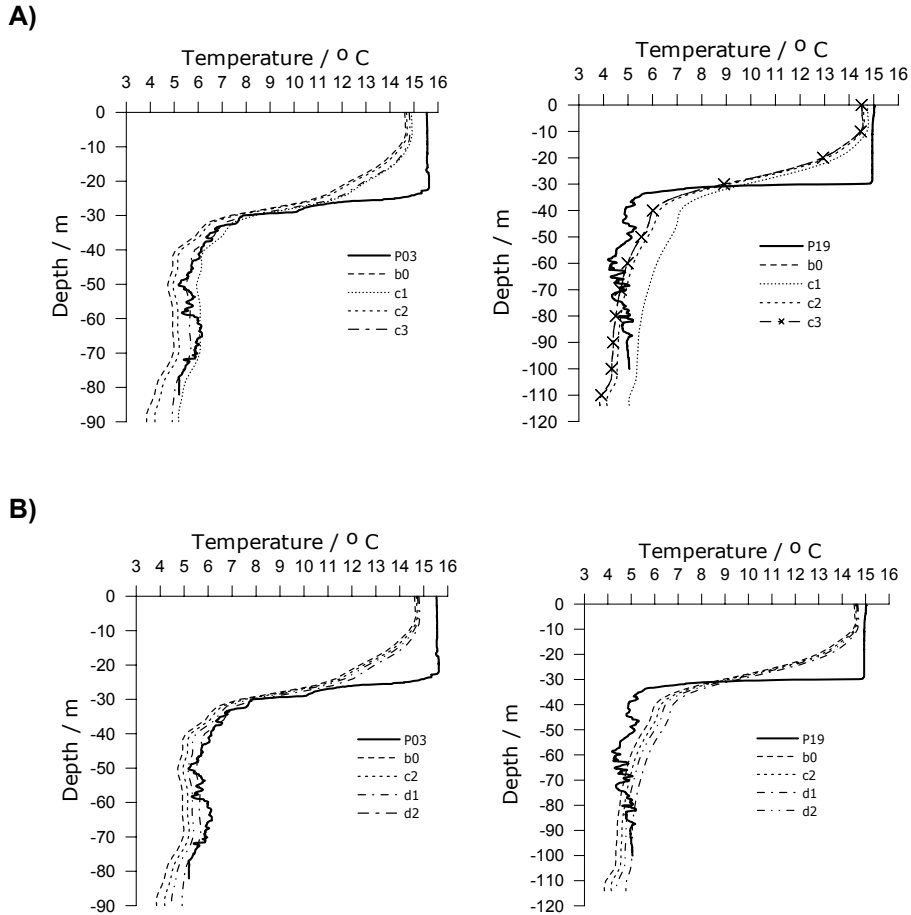
**Table 5**

Basic parameters used in the model during assimilation of temperature and salinity. ( $R$ - spatial scale;  $\alpha_\eta$ ,  $\alpha_T$ ,  $\beta_T$ ,  $\alpha_S$ ,  $\beta_S$  – temporal scales)

Exp.	$R$ [km]	$\alpha_\eta$ [hours]	$\alpha_T$ [days]	$\beta_T$ [days]	$\alpha_S$ [days]	$\beta_S$ [days]
<b>c1</b>	26	4	2	20	20	200
<b>c2</b>	16	4	2	20	20	200
<b>c3</b>	6	4	2	20	20	200
<b>d1</b>	16	4	0.5	5	1	10
<b>d2</b>	16	4	1	10	4	40

To estimate the effectiveness of the assimilation scheme, vertical profiles of modelled seawater temperature and salinity at the selected points in the Gdansk Basin have been examined. The results of simulation experiments (runs) were compared to the measurements of vertical soundings of temperature and salinity at the selected hydrographic stations in the Gdansk Basin.

Two vertical distributions of seawater temperature assimilated together with sea level according to spatial and time scale (in conformity with the scheme described in the previous section) were compared. Exemplary vertical profiles at stations P03 and P19 are shown in Fig. 4a. Spatial influence was revealed during experiments c1 – c3 when  $R$  was 26 km, 16 km and 6 km, respectively whereas all time scales were the same (4 hours) (Tabs. 6 and 7). The successive temperature distributions indicate that the largest spatial scale  $R$  have a better influence on temperature simulation above thermocline, and its mean  $R$  value (16 km) improves the course in vertical distribution below thermocline. Time scales contribution to the vertical distributions of temperature was examined during experiments c2, d1 and d2 when  $R = 16$  km (Tabs. 6 and 7, Fig. 4b). In



**Fig. 4.** A. Spatial scale experiments - modeled versus observed vertical profiles of temperature [°C] at stations: P03, P19 in September 1995; B. Temporal scale experiments - modeled versus observed vertical profiles of temperature [°C] at stations P03, P19 in September 1995

the upper part of vertical temperature distributions all spatial scales  $R$  resulted in similar course but the distributions below thermocline were described the best by the shortest time scales. However, this course did not differ too much from the others.

**Table 6**

Estimated averaged error (bias) - *ae*, absolute averaged error *aae*, root-mean square error - *rmse* and explained variance - *exp.var* for temperature and salinity model - measurements comparison at selected points: P03, P22, P25 in in Gdansk Basin

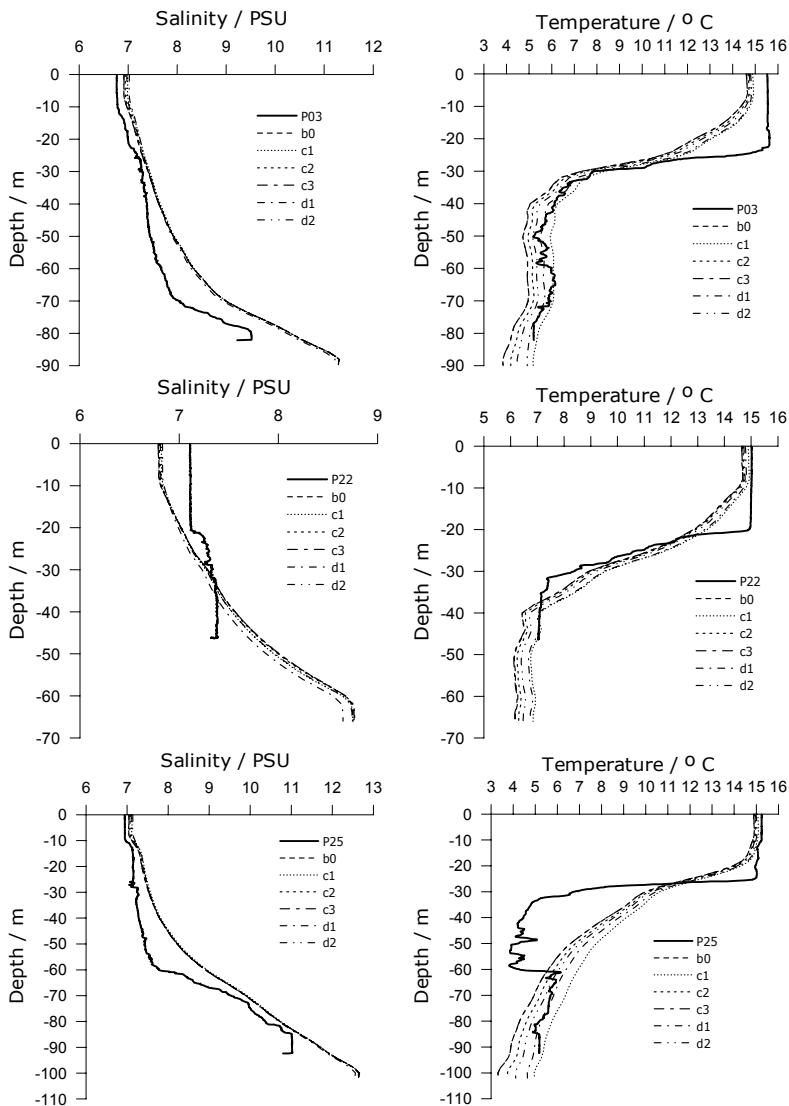
Exp.	Sta.	Salinity				Temperature			
		<i>ae</i> [PSU]	<i>aae</i> [PSU]	<i>rmse</i> [PSU]	<i>exp.var</i> [%]	<i>ae</i> [°C]	<i>aae</i> [°C]	<i>rmse</i> [°C]	<i>exp.var</i> [%]
b0	P03	-0.42	0.42	0.52	97.02	1.21	1.21	1.44	97.56
c1	P03	-0.46	0.46	0.54	97.08	0.28	0.66	0.97	98.02
c2	P03	-0.44	0.44	0.53	97.02	1.02	1.02	1.30	97.69
c3	P03	-0.43	0.43	0.53	97.02	1.22	1.22	1.45	97.56
d1	P03	-0.43	0.43	0.50	97.27	0.56	0.63	1.00	98.11
d2	P03	-0.44	0.44	0.53	97.10	0.84	0.84	1.17	97.84
b0	P22	0.09	0.20	0.23	88.71	0.34	0.66	0.81	96.15
c1	P22	0.09	0.18	0.21	88.59	-0.14	0.62	0.82	96.06
c2	P22	0.08	0.19	0.22	88.65	0.26	0.64	0.80	96.11
c3	P22	0.08	0.20	0.23	88.70	0.35	0.67	0.82	96.15
d1	P22	0.12	0.18	0.20	88.00	-0.07	0.64	0.82	96.00
d2	P22	0.09	0.19	0.22	88.50	0.14	0.63	0.78	96.10
b0	P25	-0.28	0.40	0.44	97.14	-0.51	1.60	2.01	82.08
c1	P25	-0.32	0.42	0.47	97.26	-1.45	1.75	2.46	82.95
c2	P25	-0.29	0.41	0.45	97.21	-0.71	1.54	2.06	82.37
c3	P25	-0.28	0.40	0.44	97.15	-0.50	1.60	2.01	82.11
d1	P25	-0.29	0.41	0.45	97.40	-1.12	1.51	2.24	83.01
d2	P25	-0.30	0.41	0.45	97.27	-0.87	1.51	2.12	82.63

**Table 7**

Estimated averaged error (bias) - *ae*, absolute averaged error - *aae*, root-mean square error - *rmse* and explained variance - *exp.var* for temperature and salinity model - measurements comparison at selected points P1, P19, P116 in in Gdansk Basin

Exp.	Sta.	Salinity				Temperature			
		<i>ae</i> [PSU]	<i>aae</i> [PSU]	<i>rmse</i> [PSU]	<i>exp.var</i> [%]	<i>ae</i> [°C]	<i>aae</i> [°C]	<i>rmse</i> [°C]	<i>exp.var</i> [%]
b0	P1	-0.46	0.47	0.57	96.37	0.31	0.80	1.10	96.51
c1	P1	-0.49	0.49	0.60	96.39	-0.34	1.05	1.24	96.16
c2	P1	-0.47	0.48	0.58	96.38	0.18	0.81	1.09	96.48
c3	P1	-0.46	0.47	0.57	96.36	0.31	0.80	1.10	96.51
d1	P1	-0.47	0.47	0.57	96.56	-0.11	0.87	1.11	96.59
d2	P1	-0.48	0.48	0.58	96.44	0.06	0.82	1.09	96.53
b0	P19	-0.44	0.44	0.56	94.89	0.44	0.99	1.47	92.29
c1	P19	-0.48	0.48	0.59	94.98	-0.46	1.30	1.62	92.19
c2	P19	-0.46	0.46	0.57	94.94	0.24	0.99	1.44	92.33
c3	P19	-0.44	0.45	0.56	94.88	0.44	0.99	1.47	92.29
d1	P19	-0.46	0.46	0.56	95.29	-0.17	1.07	1.47	92.48
d2	P19	-0.46	0.46	0.57	95.05	0.08	1.00	1.44	92.39
b0	P116	-0.40	0.40	0.50	93.11	0.95	1.24	2.03	90.80
c1	P116	-0.44	0.44	0.52	93.28	0.17	1.46	1.88	91.74
c2	P116	0.42	0.42	0.51	93.12	0.80	1.21	1.96	90.98
c3	P116	-0.41	0.41	0.50	93.09	0.96	1.24	2.03	90.78
d1	P116	-0.41	0.41	0.49	93.69	0.42	1.30	1.85	91.78
d2	P116	-0.42	0.42	0.51	93.30	0.65	1.22	1.91	91.24

The chosen scales significantly influenced vertical temperature distributions but their effect on salinity was hardly appreciable (cf. Fig. 5 with exemplary profiles at stations P03, P22, P25). In all runs, simulations of vertical salinity distributions for the chosen stations were represented by tightly converged

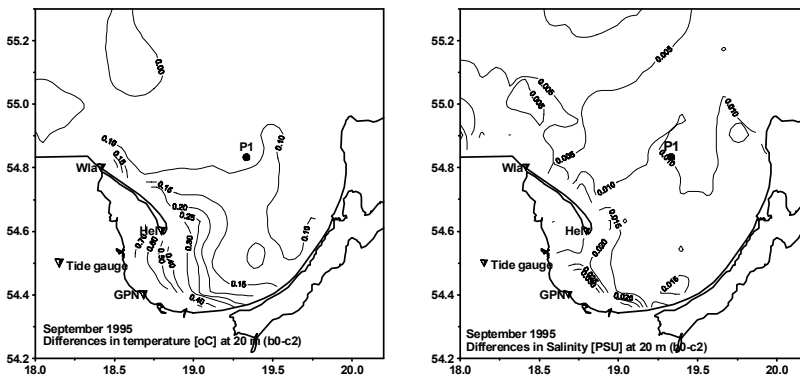


**Fig. 5.** Modeled versus observed vertical profiles of salinity [PSU] and temperature [°C] at stations P03, P22, P25 in September 1995.

running down lines (Fig. 5). Poorly differentiated values of *rmse* for salinity (Tab. 6) justified this fact. The influence of assimilation on vertical profiles of seawater salinity was very small and almost negligible. Temperature simulations were characterized by little scattered lines above thermocline whereas below thermocline the scattering was greater (Fig. 5). Similar results were obtained for deep-water stations, P1, P19 and P116.

Basic information about the impact of spatial and time scales of the nudging scheme on the vertical distribution of "temperature" and "salinity" profiles at all stations was presented in Tabs. 6 and 7 containing averaged errors. The positive bias (*ae*) of salinity was noted only at station P22 (the shallowest and close to shore station Gdansk), whereas positive bias (*ae*) of temperature was found at station P25 (nearest to shore station Hel). At the rest of deepwater stations this parameter was negative for salinity and positive for temperature. The smallest values of *rmse* were noted at station P22: 0.2 – 0.23 for salinity and 0.8 – 0.82 for temperature. The largest errors for salinity (0.57 – 0.59) and temperature (2.0 – 2.46) were found at P1 and P25 stations, respectively. High *Exp.var* were shown at P25 (82%) and P1 (96%) stations. These values indicated that the influence of assimilation on both tracers was small but it was more distinct in the case of temperature (Tabs. 6 and Tab. 7). They showed the impact of spatial and time scales in the nudging scheme on the vertical distributions of assimilated profiles of seawater "temperature" and "salinity" (Fig. 5, Tabs. 6 and 7).

Similarly as for sea level, exemplary distribution of the monthly mean differences in temperature and salinity between the results for control run b0 and c2 at 20 and 40 m depths was presented (Figs. 6 and 7). The influence of



**Fig. 6.** Monthly mean differences in sea water temperature [°C] and salinity [PSU] at depth of 20 m obtained from the runs - control b0 and c2.



seawater temperature and salinity in the Gdansk Basin. The run c3 hardly differs from b0 (without assimilation). The spatial radius in run c3 (6 km) does not differ too much from spatial grid step (5 km) in the run b0. In run c3, a threefold bigger radius caused that the differences became larger. Further enlargement of the radius (to 26 km) in run c1 gave in consequence greater differences in *rmse*. It means that the information was propagated more far away and consequently affected the distributions of modelled hydrophysical fields in the Gdansk Basin.

## CONCLUSIONS

Assimilated sea level variations measured at coastal stations influenced on the modelled field of sea surface in the whole Gdansk Basin. Their propagation depended on length scales, both spatial and time ones. Experiments b1 – b5 showed that spatial radius  $R$  of 16 km and time of 4 hours are the best among the analysed length scales. A comparison of the variations observed at the coastal gauges to the modelled results containing these scales revealed the smallest errors. To confirm this dependence in the whole area of Gdansk Basin, satellite altimetry measurements are desirable.

In the next step, to investigate the influence of the projection of surface assimilated sea level and “observed” tracers (temperature and salinity) into deeper layers, previously determined surface-to-subsurface correlations were used. The question was explored how useful those data sets were for modelling of the subsurface thermal and salinity fields. The results of experiments c1 – c3 and d1 – d2 showed that the length scales for surface and deep-seated temperature were half a day and two days, respectively, whereas those for salinity were two and twenty days, respectively. These scales were appropriate for transferring heat and salt from the surface to the bottom. If such processes were initiated from the bottom, the scales were assumed to be one order higher. Vertical length scale ( $D$ ) was equal to 20 m.

Based on the comparison of the observed and modelled sea level, temperature and salinity with their assimilated observed values, effectiveness and usefulness of assimilation were evaluated. Modelled vertical profiles of temperature displayed underestimated values of temperature in the upper layer and were close to observed values beneath thermocline (Figs. 4 and 5). The differences between modelled vertical distributions of temperature with and without assimilation reached about  $0.7^{\circ}\text{C}$  whereas those with assimilation were closer to observations.

Sea level assimilation was found to have little influence on salinity. Modelled values of salinity were overestimated. Differentiation of salinity in vertical was not so large and it caused small sensibility in transfer information

from the surface into deep layers. In the coastal zone where horizontal gradients of salinity existed, simulations with assimilation revealed bigger differentiations (Fig. 7).

The modelled vertical distribution of temperature and salinity showed similarities in homogeneous layers above thermocline and halocline. The assumed vertical correlation scale as  $D$  equal to 20 m was not also too well aimed in the algorithm. Modelled vertical distributions of temperature were much closer to the measured ones with strong temperature gradients at 30 m depth than salinity profiles (Figs. 6, 7). Vertical scale for temperature could be compared to thermocline depth limit of 50 m.

Assimilated sea level variables were transferred to subsurface structures, particularly to the field of temperatures. The experiments were run for summer season when the vertical stratification was developed. The processes of vertical heat transfer were nonlinear what was evidenced by low correlation coefficients (Tab. 1). The errors between modelled distributions with assimilation and observations were smaller than between the modelled ones without assimilation and measurements. These correlations were statistically essential. Reducing errors compared to a control case without assimilation confirmed reality of the transfer of surface information into deep and sensitivity of vertical structure (particularly the thermal one). They encourage using such approach in modelling. The answer for the question put in the title can be positive - assimilation of source surface data showed the effective reduction of errors and improvement in prediction of subsurface hydrophysical fields.

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