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ORIGINAL RESEARCH ARTICLE

On the buoyant sub-surface salinity maxima in the Gulf of Riga

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Summary Thermohaline structure in the Gulf of Riga (GoR) was investigated by a multi-platform measurement campaign in summer 2015. Stratification of the water column was mainly controlled by the temperature while salinity had only a minor contribution. Buoyant salinity maxima with variable strength were observed in the intermediate layer of the Gulf of Riga. The salinity maxima were likely formed by a simultaneous upwelling—downwelling event at the two opposite sides of the Irbe strait. The inflowing salty water did not reach the deeper (> 35 m) parts of the gulf and, therefore, the near-bottom layer of the gulf remained isolated throughout the summer. Thus, the lateral water exchange regime in the near bottom layer of the Gulf of Riga is more complicated than it was thought previously. We suggest that the occurrence of this type of water exchange resulting in a buoyant inflow and lack of lateral transport into the near-bottom layers might contribute to the rapid seasonal oxygen decline in the Gulf of Riga.

1. Introduction

The water exchange regime of semi-enclosed basins largely determines their physical and ecological nature. The classical estuarine circulation scheme in the estuaries with positive freshwater flux includes an outflow in the upper layer and inflow in the deep layer (Geyer and MacCready, 2014). The exact water exchange regime and faith of the inflowing denser water depends on the size and shape of the estuary as well as its mouth area (Valle-Levinson, 2010). Nevertheless, typically the inflowing water is in contact with the bottom of

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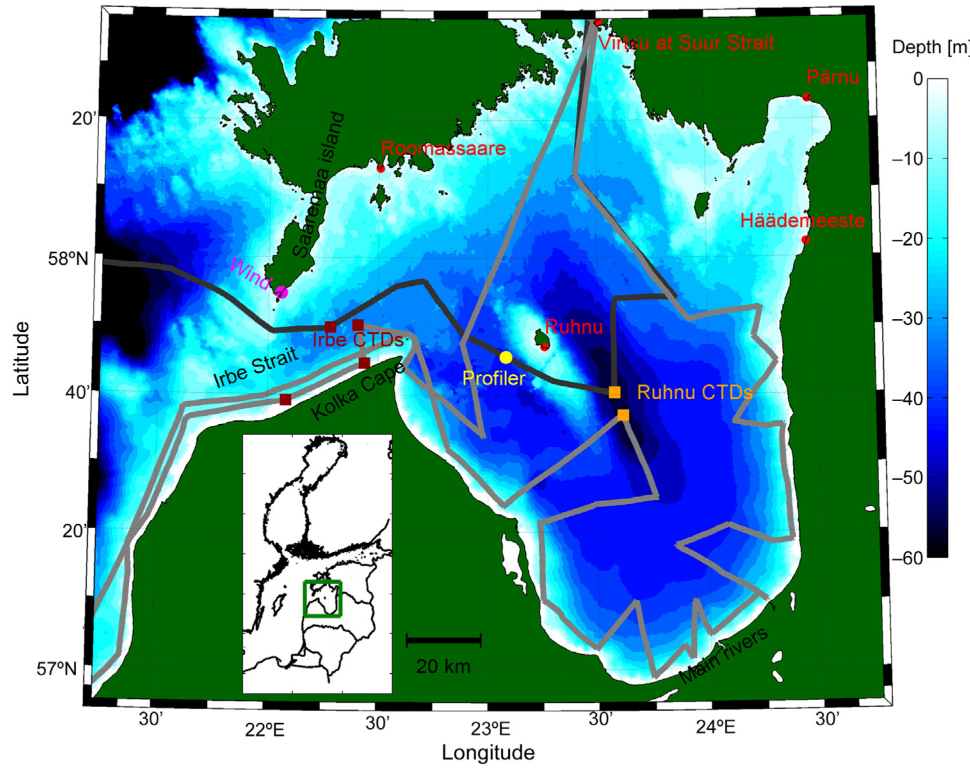

Figure 1 Map and topography of the Gulf of Riga. Color scale shows depth [m] of the study area. Lines show tracks of the RV Salme thermosalinograph surveys in July (darker gray) and August (brighter gray) 2015. Yellow circle represents the location of the moored profiler (buoy station), and red circles show locations where the coastal sea surface temperature time-series were acquired. Locations of Conductivity-Temperature-Depth (CTD) measurements at the Irbe Strait and Ruhnu Deep are shown as dark red and orange squares, respectively. The location of wind measurements at the Sõrve Cape station is shown as a magenta circle. The green box in the inlay map shows the location of the study area (Gulf of Riga) in the Baltic Sea.

the estuary. Exceptions might appear for the estuaries, which are separated from the adjacent sea basin by a sill. If the deeper layers of the estuary are filled with the saline water originating from the sporadic inflows, then the quasi-continuous water exchange over the sill involves inflowing water, which is too light to penetrate to the deepest layers of an estuary. Such regime can be found for several fjords (e.g. Belzile et al., 2016) and the Baltic Sea (Feistel et al., 2004). In the present study, we show that such water exchange regime can seasonally occur in a relatively large but shallow brackish estuary, the Gulf of Riga (GoR) as well.

The Gulf of Riga is a sub-basin of the eastern Baltic Sea. The gulf covers the area of 17 900 km^2, and its mean depth is 26 m. The deepest ($>$ 50 m) area is in the central part of the gulf (Fig. 1). The gulf is connected to the Baltic Sea via two straits: the Irbe Strait in the west with the sill depth of 25 m, width of 28 km and cross-section of 0.4 km^2 and the Suur Strait in the north with the sill depth of 5 m, width of 6 km and cross-section of 0.04 km^2.

The water and salt budgets of the gulf are formed by the two sources: the saltier water from the open Baltic Sea (Baltic Proper) and the freshwater from the rivers and due to precipitation (e.g. Raudsepp, 2001; Skudra and Lips, 2016). The water from the open Baltic flows to the deep layer of the GoR (Laanearu et al., 2000) while the riverine water occupies the upper layer. Thus there is a vertical salinity gradient present in the GoR (Skudra and Lips, 2016). The average river run-off to the gulf has been estimated approximately as 1000 m^3 s^{-1} (Berzinsh, 1995) while the average net precipitation to the gulf is about 80 m^3 s^{-1} (Omstedt and Axell, 2003). The 86% of the river run-off is discharged into the southern part of the gulf (Fig. 1, "Main rivers") from the Daugava, Lielupe and Gauja rivers (Berzinsh, 1995). This river discharge, in combination with the water exchange through the two straits in the northern part of the gulf, forms the latitudinal salinity gradient (e.g. Stipa et al., 1999). Due to the strong inter-annual variability and seasonality of the river run-off, salinity in the gulf varies remarkably at the same time-scales (e.g. Raudsepp, 2001; Skudra and Lips, 2016; Stipa et al., 1999). Due to the shallowness of the Irbe and Suur Strait, water from beneath the permanent halocline of the Baltic Sea (e.g. Reissmann et al., 2009) cannot penetrate into the GoR. There are strong salinity fronts at both straits, which change their position influenced by the wind forcing and sea level differences between the basins (Astok et al., 1999; Lilover et al., 1998).

The heat budget of the gulf (like the whole Baltic) is driven by the fluxes through the sea surface. The gulf is stratified during summers when the temperature exceeds 18°C in the upper mixed layer (Skudra and Lips, 2016). In autumn—winter, the water column is mixed down to the bottom due to the thermal convection. Further, temperature falls below the temperature of maximum density (Raudsepp, 2001; Stipa et al., 1999), and the gulf is at least partly covered by ice (Seinä and Palosuo, 1996) during winters.

The available modeling studies have shown that the heat exchange with the atmosphere and the heat content of the water column can be estimated relatively well in the Baltic (e.g. Westerlund and Tuomi, 2016). However, to simulate correct salinity distributions (both, vertical and lateral) and estimate salt fluxes between the basins and between different layers using numerical models is much more complicated (Lips et al., 2016b; Omstedt et al., 2014). This difficulty might show that the internal processes responsible for transport and mixing are described not precisely enough in such stratified, but relatively shallow basins. In the present study, we analyze the observed salinity distributions and vertical structures and relate them to the forcing.

Oxygen consumption under conditions of developed vertical stratification in summer could result in poor oxygen conditions in the near-bottom layer of deep central areas of the gulf (Aigars et al., 2015; Eglite et al., 2014). It has been argued based on the analysis of seasonal dynamics of environmental parameters and biomarkers for instance that the clam *Macoma balthica* could be stressed in late summer due to lower oxygen levels in August in the near-bottom layer of the Gulf of Riga (Barda et al., 2013). Furthermore, hypoxia at the water-sediment interface alters the phosphorus flux between the water column and sediments as well as nitrogen removal due to denitrification (Yurkovskis, 2004). Due to the sediment release of phosphorus in hypoxic conditions, the main pathway of phosphorus removal from the gulf is its relatively slow export to the Baltic Proper (Müller-Karulis and Aigars, 2011). Thus, reoccurrence of such poor oxygen conditions and benthic nutrient release could counteract decreases in the external nutrient load to the gulf similarly to the deeper Baltic basins including the Gulf of Finland (Pitkänen et al., 2001).

The main aim of the present study is to present the high-resolution view on the stratification development in the Gulf of Riga. Available investigations include the data only from episodic ship surveys (Stipa et al., 1999) or are concentrated on the long-term changes in the gulf (Raudsepp, 2001; Skudra and Lips, 2016). Studies in other basins of the Baltic have shown that implementing a new type of in situ platforms, such as moored vertical profilers (Lips et al., 2016a), Argo floats (Purokoski et al., 2013; Westerlund and Tuomi, 2016) or underwater gliders (Alenius et al., 2014; Karstensen et al., 2014), can improve the understanding of thermohaline processes. High-resolution continuous temperature—salinity measurements have not been conducted so far in the Gulf of Riga. To fill this gap, an autonomous profiler was deployed in the northwestern part of the gulf from May to September 2015.

2. Material and methods

2.1. Data

The dataset analyzed in the present study was collected from May to September 2015. Altogether 202 temperature and salinity profiles were collected by an autonomous vertical profiler equipped with OS316plus CTD (Conductivity, Temperature, Depth) probe (Idronaut S.r.l.). The profiler has been successfully applied in the Gulf of Finland for several years now (Lips et al., 2016a). The profiler was set to collect measurements in a depth range from 2 to 37 m twice a day. Initial vertical resolution of measurements was 0.1 m, but after the preliminary data processing, the profiles were stored for analysis purposes with a vertical resolution of 0.5 m. The CTD probe was calibrated by manufacturer right before the deployment, and the data quality was checked against shipborne CTD profiles several times during the study.

The shipborne CTD and dissolved oxygen data (OS320plus CTD, Idronaut S.r.l.) together with the thermosalinograph (SBE45 MicroTSG, Sea-Bird Electronics; included in the flow-through system) data were used to study the spatial thermohaline fields in the area (Section 3.2). The salinity data was checked against the water sample analyses using a salinometer 8410A Portasal (Guildline). The oxygen sensor was calibrated before each survey and was checked against water sample analyses using an OX 4000 L DO meter (WWR International, LLC).

The long-term CTD dataset 1993—2012 compiled by Skudra and Lips (2016), collected under Estonian and Latvian national monitoring programs, was extended to 2015. This dataset together with HELCOM data (http://ocean.ices.dk/helcom, 25 February 2016) was used to evaluate the occurrence of the sub-surface salinity maxima in the past (Section 3.3).

Coastal temperature measurements at Ruhnu, Häädemeeste, and Roomassaare, provided by the Estonian Environmental Agency, were included in data analysis. Likewise, the level 3 SST (sea surface temperature) product over European Seas by Copernicus Marine Environment Monitoring Service (http://marine.copernicus.eu/) was used as background information. Wind measurements at the Sõrve station were obtained from the Estonian Environmental Agency. The wind speed and direction, measured at 10 m height, were available every third hour as a 10 min average.

2.2. Calculations

The upper mixed layer (UML) depth was defined as the minimum depth where the criterion $\rho_z \geq \rho_3 + 0.15\ \mathrm{kg\ m^{-3}}$ was satisfied (ρ_z is density at depth z, and ρ_3 is density at 3-m depth).

Relative contributions of temperature (ST_T) and salinity (ST_S) to the vertical stability of the water column were estimated as:

$$ST_T = \alpha \frac{dT}{dZ} \rho, \quad (1)$$

$$ST_S = \beta \frac{dS}{dZ} \rho, \quad (2)$$

where α is the thermal expansion coefficient, β is the saline contraction coefficient, ρ is the water density; dT/dZ and dS/dZ are the temperature and salinity gradients over the vertical distance dZ. Temperature and salinity profiles were both smoothed by 2.5 m window before the stability and intrusion index calculations. Stability (ST) of the water column was acquired by summation of ST_T and ST_S. Total (contribution to) stability of the water column was calculated as vertically integrated ST_T, ST_S and ST. Likewise, the total contribution to the stability of the UML was calculated as vertically integrated ST_T, ST_S, and ST within the depth range of the UML.

In order to estimate the intensity of interleaving the intrusion index as the sum of negative salinity gradients [g kg^{-1} m^{-1}] was calculated same as by Lips et al. (2016a):

$$I = \sum_{Z=h_1}^{Z=h_2} \begin{cases} 0, & \text{if} \quad \dfrac{dS}{dZ} \geq 0 \\ abs\left(\dfrac{dS}{dZ}\right), & \text{if} \quad \dfrac{dS}{dZ} < 0 \end{cases}, \tag{3}$$

where I is the intrusion index, h_1 and h_2 are the borders of the depth interval where the index was calculated. It means if salinity increases downwards throughout a profile, the index is zero. Contrary, if there is a strong sub-surface salinity maximum, the index has a higher value.

The apparent oxygen utilization (AOU) was used to show the difference between measured dissolved oxygen content and saturation level:

$$\mathrm{AOU} = DO_{100} - DO_M, \tag{4}$$

where DO_{100} is oxygen concentration at saturation level (Weiss, 1970), and DO_M is the measured oxygen concentration.

Salinity values are given as Absolute Salinity [g kg^{-1}], density as potential density anomaly to a reference pressure of 0 dbar [σ_0; kg m^{-3}] and the unit of oxygen values is mg L^{-1} in the present paper. Density values in Section 3.3 are presented as arithmetic averages at 0—5 m and 32—38 m depth ranges in the Baltic Proper and in the Gulf of Riga, respectively.

3. Results

3.1. Temperature, salinity, density and stratification time-series at mooring station

3.1.1. General description

Water column structure was determined by the seasonal thermocline at the beginning of the study period (Fig. 2). Temperature decreased almost linearly from the base of the upper layer (9.3—10.0°C) at 15—20 m depth to the near-bottom layer (4.3—5.5°C). Vertical distribution of salinity was relatively even and ranged between 5.9 and 6.0 g kg^{-1}. Temporal development of the vertical thermohaline structure was characterized by a seasonal increase in temperature due to the atmospheric heat flux and a slight decrease in salinity in the upper layer in June. By the end of June, temperature (salinity) gradient from the upper layer to the near bottom (old winter water) layer was from 15.6 to 5.8°C and from 5.8 g kg^{-1} to 6.0 g kg^{-1}, respectively.

Advection of fresher water was observed at the buoy location in the upper 25 m in the first week of July. Salinity values below 5.4 g kg^{-1} were registered. At the same time, water temperature increased in the upper layer, which, however, cannot be related to the advection. First, the temperature maxima and salinity minima did not match exactly in time. Secondly, a simultaneous temperature increase and decrease after that were registered at all coastal stations around the gulf, which implies that the atmospheric heat flux was responsible for this temperature change.

Further developments in July included a general increase in the upper layer temperature and intermittent appearances of fresher and saltier patches in the sub-surface layer at the depth range from 14 to 24 m. Less saline water in the upper layer and thick sub-surface intrusion of saltier water after that were observed at the end of July. Upper layer temperature reached its seasonal maximum of 20.4°C in mid-August. Upper layer salinity varied around 5.5 g kg^{-1} in the first half of August and was close to 5.7 g kg^{-1} in the second half of the month.

Vertical structure in September was characterized by heat loss to the atmosphere and by advection of fresher water to the buoy location. A temperature and salinity decrease in the upper layer, as well as thickening of the mixed layer, were observed during that period. Thus, it was the advection of fresher water that dominated over the vertical transport of salt in the upper layer salt flux. Interestingly, temperatures below 4.5°C were occasionally registered at the deepest measured horizon of 37 m even in September.

3.1.2. Stratification

Time series of the water column stability are presented in Fig. 3a. The higher (red) values mark the location of the pycnocline while the dark blue color indicates that water is vertically homogeneous. The seasonal thermocline was the main contributor to the stability of the water column (Fig. 3b). There were only a few occasions when the temperature gradients had an opposite effect and caused weakening of the vertical stratification.

Negative stability values due to the vertical salinity gradient occurred in July and August, and they were especially strong in the second half of July (Fig. 3c). Those layers were located in the thermocline and were compensated by the vertical temperature gradient. It has to be noted that the layer with inverse salinity gradient was always surrounded by the layers above and below it where the salinity was increasing with the depth (Fig. 3c). The explanation for such salinity derived stability distribution was the existence of sub-surface buoyant saltier water intrusion. Starting from the sea surface, the following gradients/layers were associated with the salinity maxima and could be distinguished: (1) layer with positive salinity gradient from the ambient water in the thermocline to the core of maxima layer (red color); (2) values close to zero that show the location of the core (white color); (3) layer with negative salinity gradient between the core and ambient deep layer water (blue color); (4) values close to zero that show the point (range) where salt intrusion water merged with the ambient water (white color); (5) positive salinity gradient in deep water (red color).

The integrated stability over the water column (Fig. 3d) showed that the thermal buoyancy dominated while salinity had a minor importance in the strength of stratification at the buoy location. The minor role of salinity can be at least partly related to the fact that the highest salinity (intrusion) was located in the intermediate layer and not in the bottom layer. The salty water near the bottom would have increased the stability over the water column.

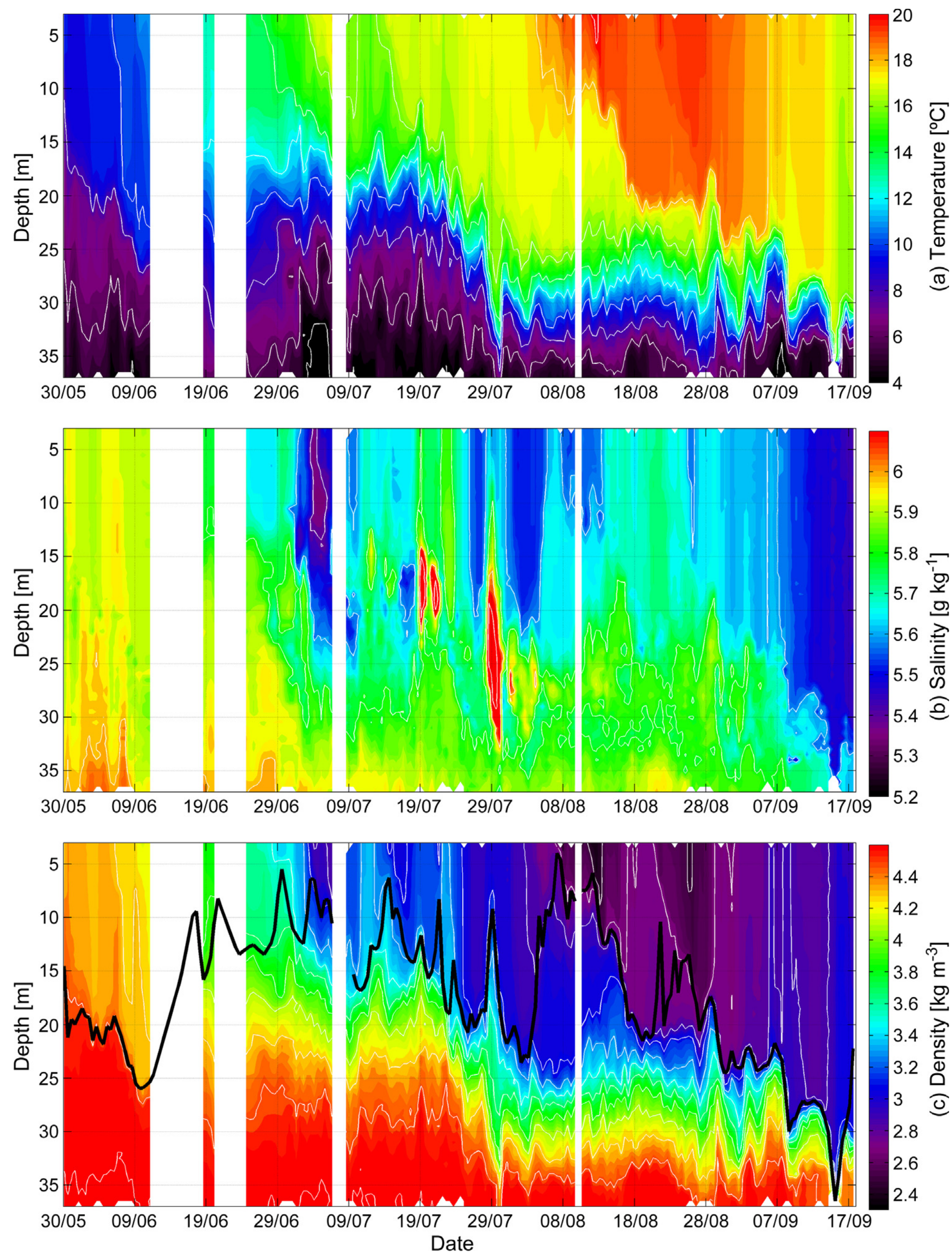

Figure 2 Time series of temperature [°C], salinity [g kg^{-1}] and density anomaly [kg m^{-3}] at the buoy station from 30 May 2015 to 18 September 2015. Black line in the bottom panel is the upper mixed layer (*UML*) depth. The location of the mooring is shown as yellow circle in Fig. 1.

The *UML* depth ranged from a few meters to 35 m during the study period (Fig. 2c). The *UML* was mostly stabilized by the temperature while salinity had occasionally (at the beginning of July) an opposite effect (Fig. 3e). Salinity and temperature had a similar contribution to the stability at the beginning of August.

Time-series of intrusion index (Fig. 3f) show relatively low values (<0.05) until the beginning of July. It is noteworthy that intrusion index values differed from zero, most probably, due to the observed sub-surface freshwater patches (Fig. 2b). Occasional peaks of intrusion index up to 0.1 were observed until the beginning of September. The highest values exceeded 0.2 and occurred in the second half of July.

3.1.3. Salinity fluctuations

The period with highest intrusion peaks is below described in more detail. It is noteworthy that immediately before the appearance of saltwater patches a clear signature of fresher water was registered at the 15—20 m depth (Fig. 4, 19 Jul 12:00). The core of this fresher layer had the salinity of 5.55—5.60 g kg^{-1}, and it lied in between the upper layer with the salinity of 5.75—5.80 g kg^{-1} and deep layer with the salinity of 5.95 g kg^{-1}.

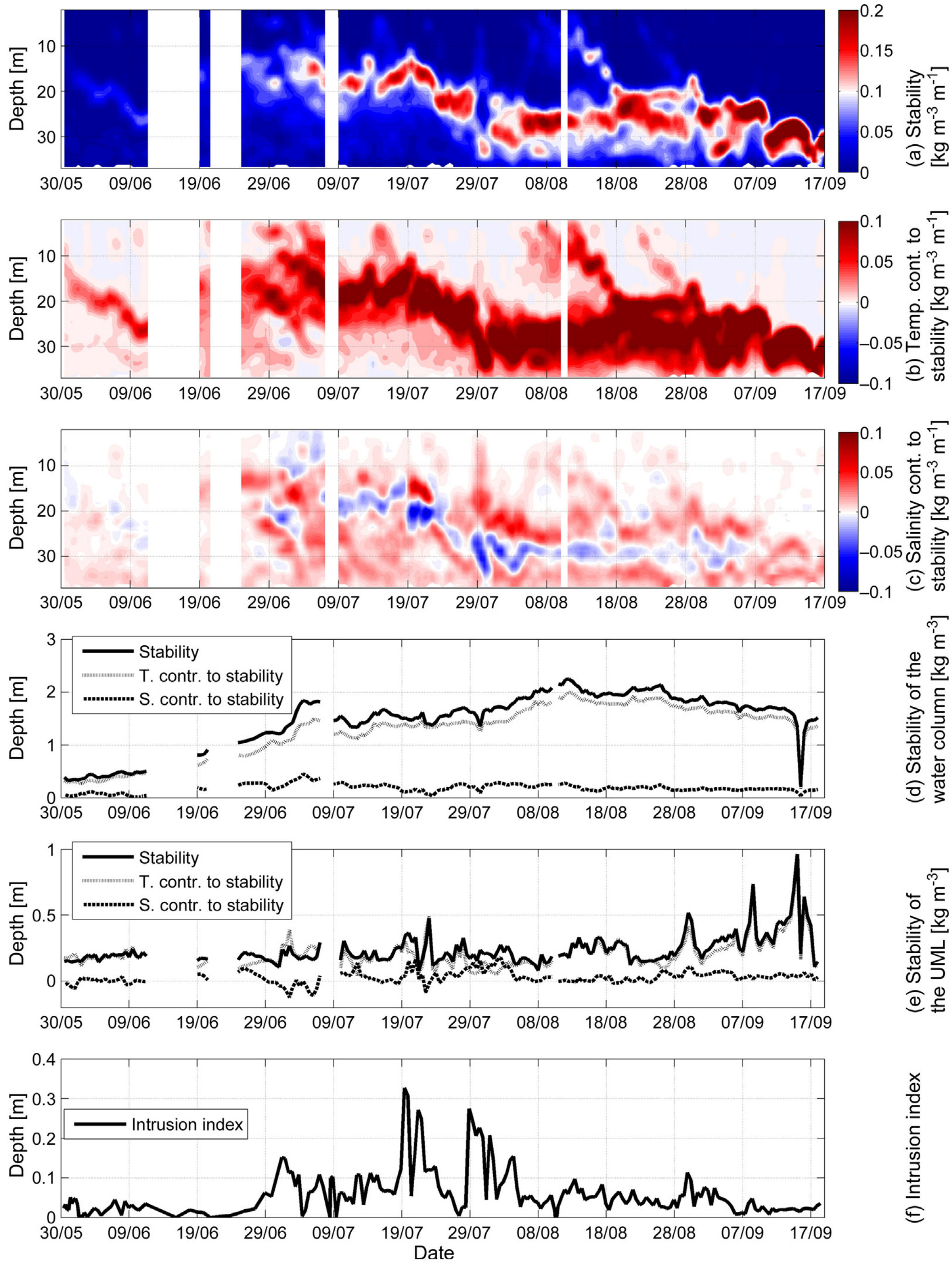

Figure 3 Temporal development of the following vertical stratification parameters at the buoy station from 30 May to 18 September 2015: stability (a; [kg m^{-3} m^{-1}]) and contribution of temperature (b) and salinity (c) to stability; integrated stability parameters [kg m^{-3}] for the whole water column (d) and for the upper layer (e); intrusion index (f).

Next, the saltier water (5.80 g kg^{-1}) appeared in the upper layer and rapidly (within our profiling interval of 12 h) the fresher water in the sub-surface layer was replaced by the salinity maximum (Fig. 4, 20 Jul 00:00). The saltier sub-surface layer was located in the depth range of 11—25 m and its core with the salinity of 6.45 g kg^{-1} was located at 17 m depth. The temperature of the core was 14.55°C while 3.5 m above the core it was 13.75°C. Thus, the salinity gradient above the core neutralized the negative buoyancy due to the vertical temperature distribution. It was opposite below the core: strong thermocline stabilized the water column despite the negative salinity gradient. The sub-surface salinity maximum disappeared on 22 July. Fresher sub-surface water was observed from 23 July to 28 July while salinity in the upper layer decreased from 5.80 to 5.55 g kg^{-1}. The saltier water reappeared in the sub-surface layer on 29 July (Fig. 4 29 July 12:00). The core of the sub-surface layer was located at 22 m depth and had salinity and

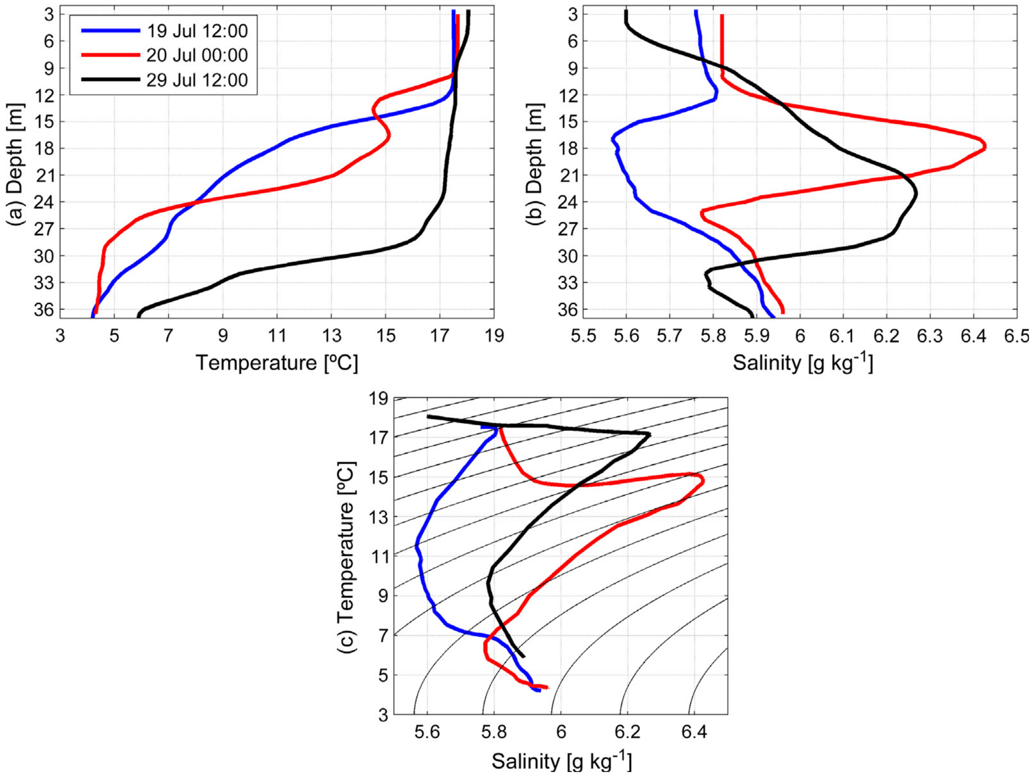

Figure 4 Selected profiles of temperature ([°C]; a) and salinity ([g kg^{-1}]; b) and respective temperature—salinity diagram with density anomaly ([kg m^{-3}]; c) as contour lines in profiling station (the location is shown as a yellow circle in Fig. 1).

temperature of 6.30 g kg^{-1} and 17.15°C, respectively. The core was only slightly (0.15°C) warmer than the layer above. That temperature change had a minor contribution to the vertical stratification as salinity increased from 5.70 g kg^{-1} to 6.30 g kg^{-1} in the same depth range. The sub-surface salinity maximum, though much weaker, was observable until the end of August.

3.2. Origin of the salt intrusion

The observed saltier water could potentially penetrate to the GoR from the Baltic Proper via either the Irbe Strait or Suur Strait. The core of the observed salt maxima layer had the salinity of 6.45, 6.30 g kg^{-1} and temperature of 14.55°C, 16.40°C on 20 July and 29 July, respectively. We can expect that if the source water has been modified, then it occurred because of mixing of it with the GoR ambient waters. The temperature—salinity curves (TS-curves) of the sub-surface salinity maxima cores indicate that this saltier water was slightly warmer than the surrounding GoR waters. Thus, if the observed core has been a subject of mixing, the source water of the salt maxima layers must have been saltier and slightly warmer than the observed core at the buoy station. Unfortunately, there were no continuous measurements available in the straits. Thus, we were not able to capture the entering process of the saltier waters directly. However, we collected some shipborne CTD profiles in the Irbe Strait area (Fig. 5). The other data source was made available by the thermosalinograph that autonomously recorded temperature and salinity of the water pumped from 2 m depth during the surveys on board RV SALME. The two surveys that were included in the analysis were conducted on 14—15 July and 7—9 August.

The first survey was conducted through the central part of the Irbe Strait on 14—15 July while the vessel visited more the southern part during the second survey; though, one station in the central part was sampled as well. During the second survey, the area was visited twice: the first passage on 7—8 August and the second passage on 9 August.

During the both surveys, a strong salinity front was found in the upper layer in the Irbe Strait area. The front was captured near the longitude of 22°E by the survey on 14—15 July (Fig. 5). Salinity values below 6.20 g kg^{-1} were observed toward east from the longitude of 22°E. The front was located further in the east on 7—9 August. However, due to different tracks of the two surveys (the latter was conducted along the southern coast) we cannot confirm, whether it was a spatial feature or temporal displacement of the front.

The most remarkable feature in the upper layer temperature dynamics was the colder water observed in the strait on 14—15 July survey. The temperature of the upper layer was around or exceeding 17°C in the Baltic Proper and the gulf, but below 13°C inside of this cold feature (Fig. 5). Since the Irbe Strait is a topographic barrier between the open Baltic and the gulf, upwellings in both basins can potentially bring cold water to the upper layer in the Irbe area. Strong SW winds (>10 m s^{-1}) occurred during the period from 5 to 10 July, i.e. favorable forcing for an upwelling in the GoR along the coast of the Saaremaa Island. Weaker winds from NW, i.e. from the favorable direction for the upwelling along the Latvian coast in the Baltic Proper, prevailed from 11 to 15 July (Fig. 6). The satellite-derived sea surface temperature, as well as temperature measurements at the coastal stations and by the onboard thermosalinograph, indicated that the upwelling occurred in the NW part of the GoR in mid-July (Fig. 7). Since salinity in the whole water column was higher in the Baltic Proper than in the core of the upwelled cold water in the Irbe Strait ($6.15\ g\ kg^{-1}$), this upwelled water cannot be a pure Baltic Proper water. On the other hand, the upwelled water was saltier than the GoR water. Thus, the upwelled water might be a mixture of waters from the both basins.

The only available profile in the Irbe Strait on 14—15 July was acquired eastward from the core of the upwelling (Fig. 5a). The profile revealed three layers: upper layer (15.00°C and $5.90\ g\ kg^{-1}$), deep layer (4.70°C and $5.95\ g\ kg^{-1}$) and salt maxima layer (12.15°C and $6.35\ g\ kg^{-1}$) between the former two (Fig. 8). The temperature profile, as well as the satellite-derived sea surface temperature, showed that the station was in the upwelling zone but not in its coldest core area. Relatively low salinity of the upper layer (upwelled water) and deep layer at the station indicated that those waters originated from the gulf. The salt maxima layer at the station and coldest water in the upper layer observed by the thermosalinograph were similar. Though, the salt maxima water was slightly colder and saltier than in the upwelled water at the CTD station. Thus, the observed salt maxima layer water was likely mixed with warmer and fresher surface water before upwelled to the surface as registered by the thermosalinograph (Fig. 8).

The structure of the salt maximum was similar to the one observed at the buoy station, which suggests that the salt impulse entered the gulf likely via the Irbe Strait. The salt maxima at the thermocline were much saltier than the water below and above and the water in those layers was slightly

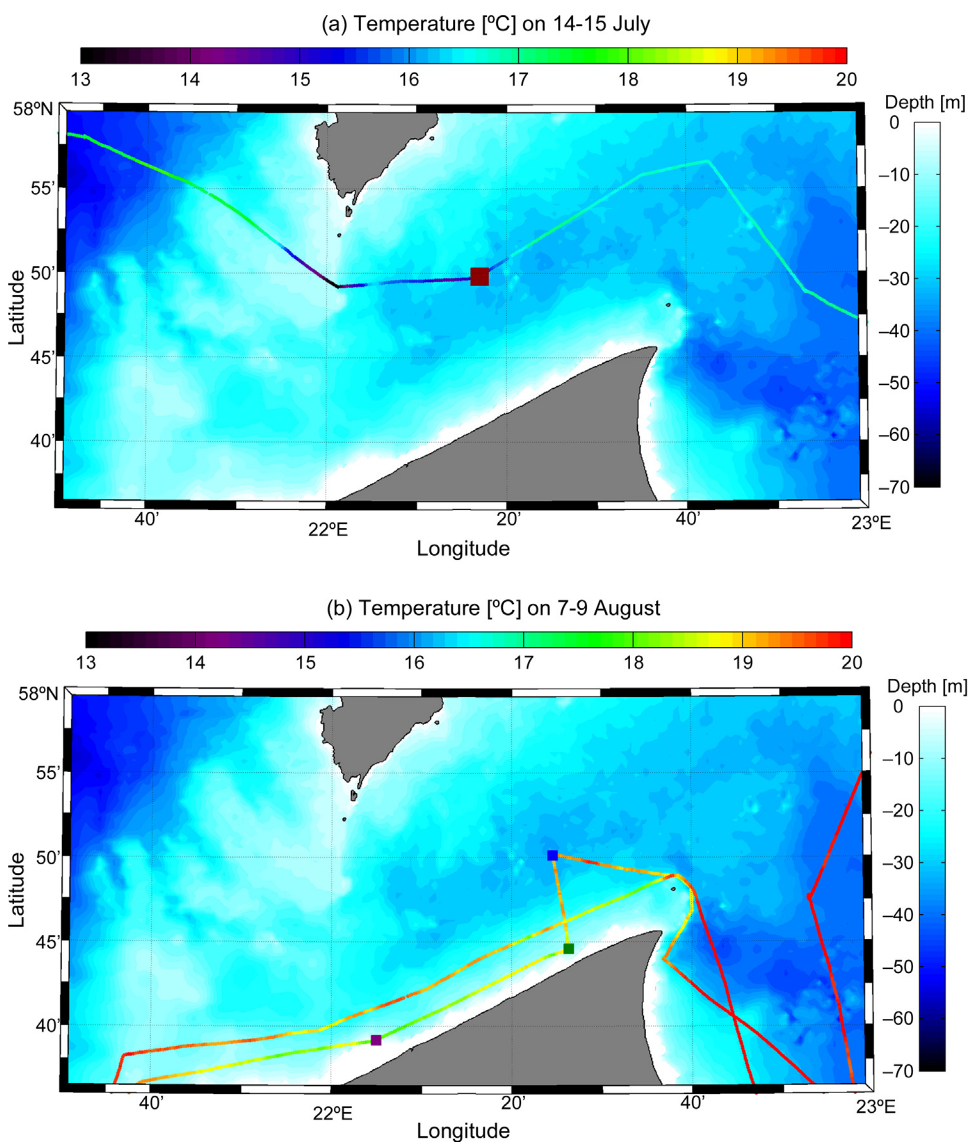

Figure 5 Temperature ([°C]; a, b) and salinity ([$g\ kg^{-1}$]; c, d) at 2 m depth acquired by the thermosalinograph on board RV SALME on 14—15 July (a, c) and 7—9 August (b, d). On vertical color bar sea depths [m] are shown.

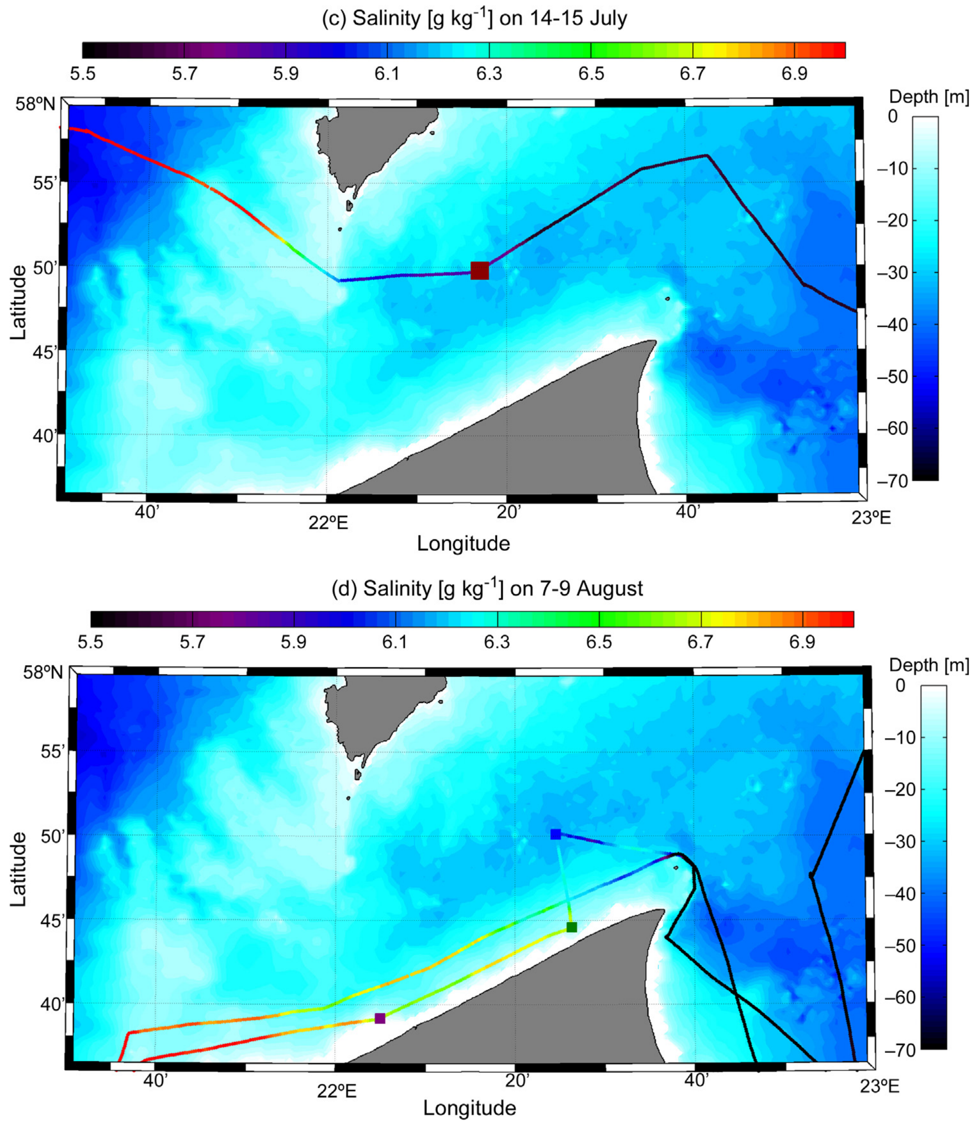

Figure 5 (*Continued*).

warmer than in the thermocline above. The latter indicates that the salt maxima water originates from the warmer layer above.

The similar structure of salt intrusion was found in the central part of the Irbe Strait on 9 August (Fig. 8). At the same time, high saline >6.7 g kg^{-1} and warm water was observed in the upper layer in the southern part of the strait (Figs. 5 and 8). TS-curves suggest that the salt maxima layer observed in the central part of the strait was a mixture of this high saline warm water and thermocline water of the GoR. The possible mechanism that can penetrate the warm and high saline water deeper is downwelling along the southern coast of the Irbe Strait. Indeed, SW wind that generates upwelling along the shore of the Saaremaa Island also causes downwelling along the southern coast of the strait. Moreover, the SW wind likely creates a coastal boundary current along the eastern coast of the Baltic Proper. It can be seen that at the station marked in green in Fig. 8, this warm, salty water has pushed over the fresher water. Thus, it likely caused some vertical mixing there as well.

In an upwelling cell along a boundary, the upwelled water is typically compensated by an onshore flow in the deeper layer. In the present case, upwelling occurred in the strait, which is relatively narrow and shallow. Thus, simultaneously with the upwelling, a downwelling occurred along the opposite coast, and the downwelled waters (originated from the Baltic Proper) rather fast reached the seabed. We suggest that the upwelled water was (at least partly) compensated by this warm and saltier downwelling water in the Irbe Strait. This hypothesis is supported by the TS-curves: one source water for the salt maxima layer observed in the central strait on 9 August (blue curve in Fig. 8) had almost the same TS-characteristics as the warm and saltier water in the southern part of the Irbe Strait (magenta and green curves in Fig. 8). Moreover, the same suggestion can be made on the basis of *AOU* vs. *S* curves (Fig. 8). This saltier and warm water was formed in the Irbe Strait and later spread to the gulf as upwelling—downwelling system relaxed.

Measurements at the coastal stations and buoy station reveal the temporal development of the upper layer

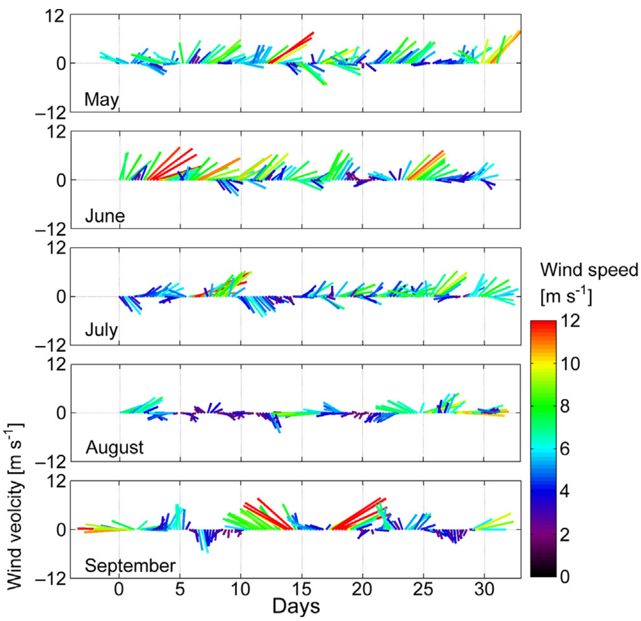

Figure 6 Wind vectors at the Sõrve meteorological station in 2015. The location of the station is shown in Fig. 1 as a magenta circle.

temperature in different parts of the gulf. There were several synoptic events of cooling (at the beginning of August) and warming (at the beginning of July) that simultaneously occurred at all stations (Fig. 9). Those were atmospheric heat flux events. The seasonal temperature increase at Ruhnu and, especially, at the buoy lagged the other two stations until the end of June. This delay was related to the slower warming of the open gulf waters comparing to the shallow coastal waters. The same tendency was in September, but this time, the slower cooling in the open sea resulted in a higher temperature at the buoy station than at the coastal stations. One can note that the lowest temperatures mostly occurred at Roomassaare in mid-July during a period of two weeks. Upwelling along the north-western coast of the gulf prevailed during this period. The Roomassaare station is not the best location to catch the upwelling in the Irbe Strait but as satellite derived sea surface temperature showed (Fig. 7), at least in some cases, upwelling events occurred in the Irbe Strait and at Roomassaare at the same time. Moreover, the period of upwelling coincided with the period of strongest salt maxima observed at the buoy station (intrusion index in Fig. 3f). This coincidence supports the suggestion that the salt pulses entered the gulf during the simultaneous upwelling in the northwestern gulf (along the shore of the Saaremaa Island) and downwelling along the southern coast of the Irbe Strait.

3.3. Earlier observations

In the previous subchapter, we suggested that the sub-surface saltwater maxima originated from the upper layer of the Baltic Proper and likely entered the gulf via the Irbe Strait. The pre-condition for the sub-surface buoyant salt maxima establishment is a specific density range of the saltwater. The inflowing salty water (or a product of its mixing with the ambient Gulf of Riga water) should be lighter than the deep layer water and denser than the upper layer water in the gulf. The upper layer density (presented as potential density anomaly σ_0) in the Baltic Proper in January—April is in a range of 5.5—6.0 kg m^{-3} (Fig. 10). Water cools down to the temperature of maximum density (2.6—2.7°C) every winter in the Gulf of Riga. Salinity over 6.9 g kg^{-1} (at the temperature of

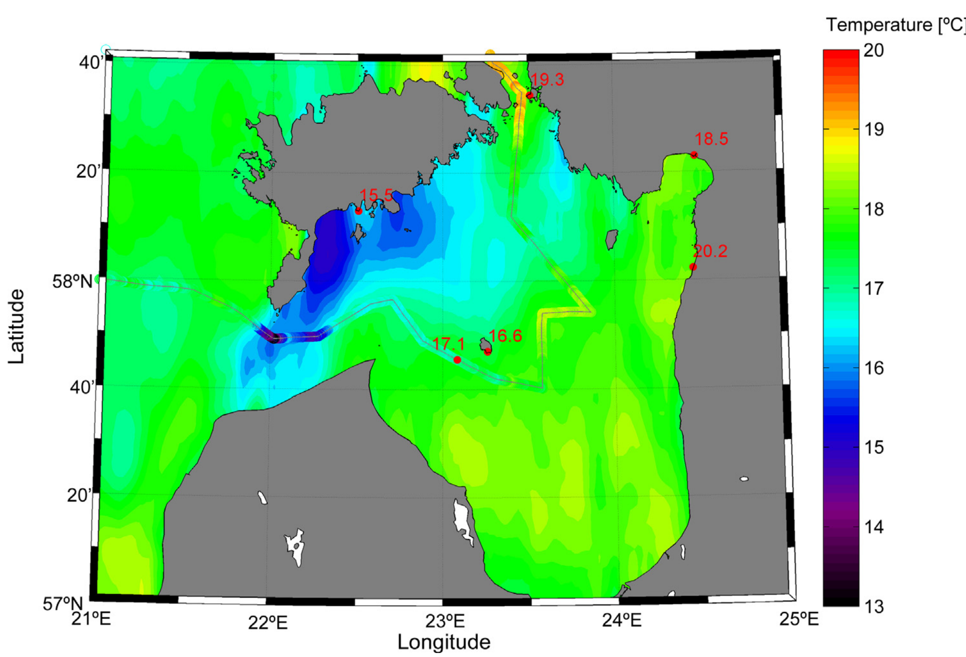

Figure 7 Satellite-derived sea surface temperature; upper layer temperature registered along the RV SALME track by thermosalinograph, at the buoy station and selected coastal stations on 15 July 2015.

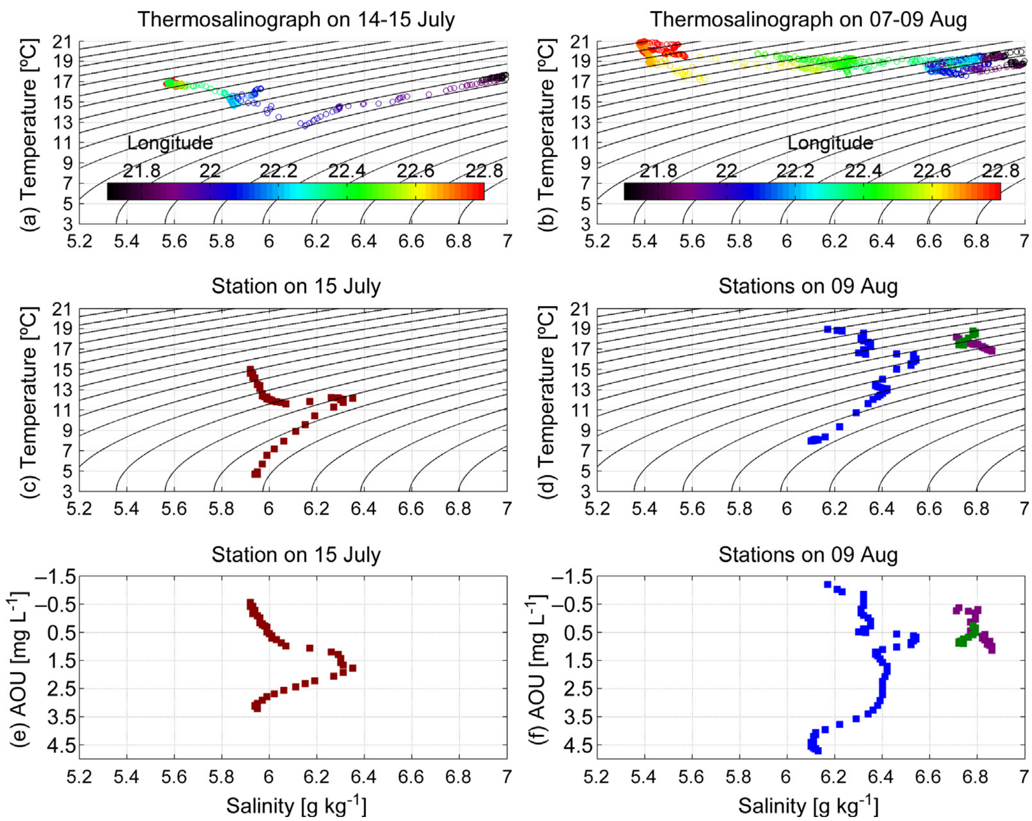

Figure 8 Temperature—salinity (TS) and apparent oxygen utilization (AOU) vs. salinity curves on 14—15 July and 7—9 August. TS-curves along the vessel track acquired by the Salmebox (a, b) and at selected Conductivity-Temperature-Depth (CTD) stations during the two analyzed surveys. Thermosalinograph tracks and locations of CTD stations are shown in Fig. 5 (station dots on the map have same colors as TS dots here). Color bars in the uppermost panel show the longitude.

2.6—2.7°C) would be necessary to reach the density of 5.5 kg m^{-3}. Salinity in the deepest layers of the gulf does not typically exceed 6.5 g kg^{-1} (Raudsepp, 2001). As shown in Fig. 10a, average density anomaly at the 32—38 m depth in

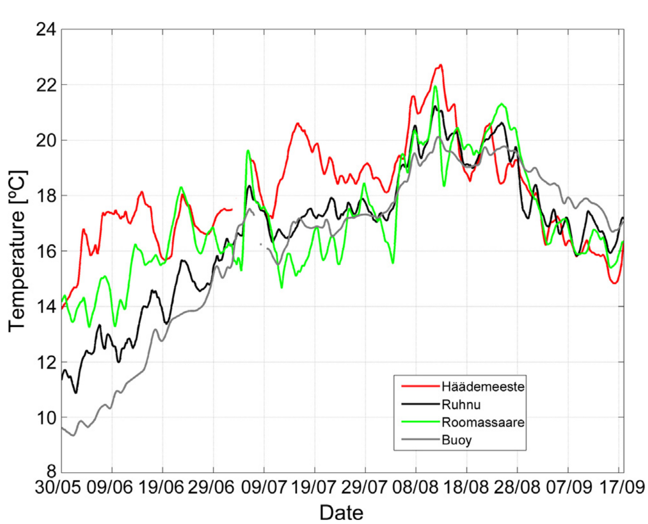

Figure 9 Upper layer water temperature at various locations in the Gulf of Riga. Coastal stations are shown as bright red circles and buoy location as a yellow circle in Fig. 1.

the Gulf of Riga is mostly below 5 kg m^{-3}. Thus, if the Baltic Proper water enters the Gulf of Riga in January—April, it must dive to the near-bottom layer of the gulf and sub-surface maxima as such cannot exist during winters.

As soon as the surface water in the Baltic Proper gets lighter due to the seasonal warming in June, the formation of sub-surface salinity maxima is feasible. In autumn, convection and wind stirring mix the denser deeper layer water and lighter upper layer water thoroughly. As a result, the deep layer temperature increases while salinity and density decrease. Therefore, sub-surface salinity maxima formation in October—November is not feasible anymore, and the dense Baltic Proper water flows to the near-bottom layer of the gulf. The upper layer water in the Baltic Proper tends to be lighter than water in the Gulf of Riga at 32—38 m depth during the timeframe from mid-July to the beginning of September (Fig. 10a). Thus, if the Baltic Proper water flows into the gulf during that time, it likely does not reach the bottom layers of the gulf and rather forms buoyant salt maxima. The temporal variability of the intrusion index, calculated based on the available full-resolution (0.5 m) CTD casts from the years 1993—2014, confirms latter and shows that the saltwater intrusions have been observed from mid-July to mid-September (Fig. 10b).

Comparison of the upper layer density in the Baltic Proper and the Gulf of Riga shows that latter water is always lighter. Even during the autumn-winter cooling period, when the

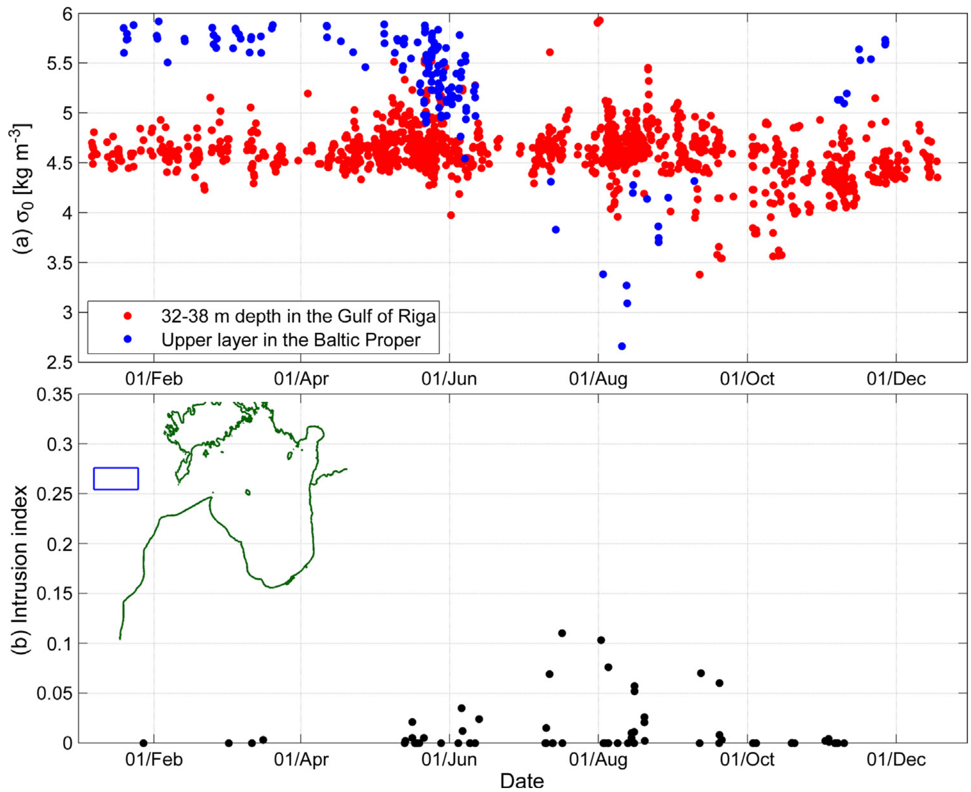

Figure 10 (a) Annual cycle of average potential density anomalies (σ_0; [kg m^{-3}]) in the 0—5 m layer of the Baltic Proper (1979—2014) and at the 32—38 m depth in the Gulf of Riga. (b) Intrusion index in the Gulf of Riga at ≥35 m deep stations (1993—2015); only high-resolution profiles were included. The blue box indicates the location of the analyzed profiles from the Baltic Proper. In the Gulf of Riga, all available deep enough measurements were included.

temperature in the Gulf of Riga drops faster than in the Baltic Proper, latter has still heavier upper layer water. This density difference shows that the water coming from the Baltic Proper cannot lie above the Gulf of Riga water.

In conclusion, we suggest that two types of Baltic Proper inflow regimes exist in the gulf. The near-bottom layer salt wedge regime is the only inflow pattern from October to May/June. In summer, from June/July to September, the regime leading to the formation of the buoyant salt intrusion (or saltier water patches) if the inflowing water originates from the upper layer or the regime resulting in the near-bottom salt wedge if the inflowing water originates from the layer beneath the seasonal thermocline can occur.

4. Discussion

The year 2015 was special for the Baltic Sea in two senses. First, strong barotropic Major Baltic Inflow (Matthäus et al., 2008) occurred in December 2014 (Mohrholz et al., 2015). Secondly, the winter 2014/2015 was exceptionally mild in the Baltic Sea area, and the Gulf of Riga was almost ice-free in winter (Uotila et al., 2015). However, we do not expect that these two events had a major impact on the thermohaline state of the gulf in summer 2015. It takes years to see the impact of a Major Inflow in the upper layer of the Baltic Proper (Reissmann et al., 2009), which is a salty water source for the Gulf of Riga. Despite an exceptionally warm winter, the gulf was ventilated down to the bottom layer in the deepest part of the gulf in the Ruhnu Deep (Fig. 11). The mild winter might have an impact on the river discharge (Apsite et al., 2013) and, thereby, to the salinity of the gulf though (Raudsepp, 2001).

High-resolution profiling showed that stratification of the water column during the study period was mainly controlled by the temperature while salinity had only a minor impact. Likewise, the stability of the upper mixed layer was mostly controlled by the temperature. Nevertheless, the vertical stratification in spring is mostly created by the freshwater fluxes, at least in the southern part of the Gulf as suggested by Stipa et al. (1999). It is worth to perform high-resolution profiling in spring, to observe the development of stratification.

Distinctive sub-surface salt maxima were observed in the thermocline in July and August 2015. Due to the lack of data, spatial view on the salt maxima is quite limited. Anyhow, all three available CTD profiles (not shown here) from July to August registered in the Ruhnu Deep revealed the occurrence of salt maxima there. Thus, the maxima can reach the central or southern part of the gulf. It can be expected that the patches of sub-surface salt maxima only occasionally reached or passed the profiling station. In other words, the rapid disappearance of maxima layer observed at the buoy station was not a result of vertical mixing.

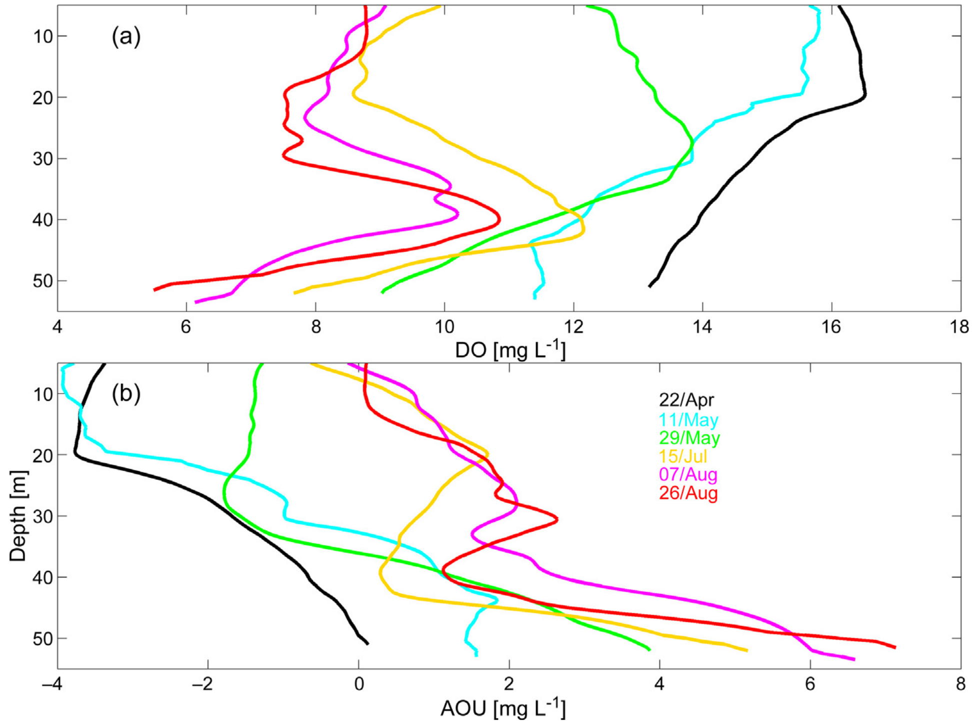

Figure 11 Dissolved oxygen content (a) and apparent oxygen utilization (AOU, b) profiles in the Ruhnu Deep stations in 2015. The location of stations is shown as orange circles in Fig. 1.

Likely, the salt maxima are intrusions of waters of different origin, e.g. as has been observed in the halocline of the Bornholm Basin in the vertical temperature distribution (e.g. Mohrholz et al., 2006). The only evidence of similar salt maxima in the GoR has been reported and measured earlier at two stations in the eastern part of the Irbe Strait by Stipa et al. (1999) on 29 August—1 September 1993. This observational evidence supports our suggestion that the patches of salt maxima enter the gulf via the Irbe Strait.

Episodic current measurements have revealed that the flow regime in the Irbe Strait is two-layered with an inflow in the deeper layer near the southern coast and an outflow along the northern coast (Lilover et al., 1998; Talpsepp, 2005). SW winds evoke upwelling in the Gulf of Riga side of the Irbe Strait along the Saaremaa Island and downwelling along the southern coast of the Irbe Strait. The upwelled Gulf of Riga water is replaced by the waters from the deeper layers of the Irbe Strait, which can be partly the salty and warm downwelling waters. The downwelling water mixes with the ambient Gulf of Riga sub-surface water and the outcome is the sub-surface salt maxima layer. The similar suggestion was drawn by Stipa et al. (1999) as well: downwelling depresses the seasonal pycnocline on the Baltic Proper side of the strait and creates a baroclinic pressure gradient along the strait, which might drive the Baltic Sea surface water, instead of the water below the seasonal pycnocline, into the Gulf of Riga. Stipa et al. (1999) suggested that downwelling occurred in the Baltic Proper along the Latvian coast in 1993 when they registered the salinity maxima in the area. The CTD-stations located closely to each other are to be sampled to check if only the baroclinic pressure gradient is needed (Stipa et al., 1999) for the maxima layer generation, or also the cross-strait downwelling as we suggested in the present study plays a significant role here.

The salt maxima-favorable coupled upwelling—downwelling situation is evoked by SW winds. Comparison of wind conditions in July—August 2015 with the climatic averages (in July—August 1966—2015) shows that summer 2015 was not a very special year. The favorable SW winds (sector 180—270° was taken into account) occurred 39% of the time in July—August while the climatic occurrence of the wind from the same direction in July—August has been 41%. Likewise, the occurrence of strong SW winds (if only the winds with the speed of >5 m s^{-1} or >10 m s^{-1}) in 2015 was close to climatic average.

The sub-surface salt maxima formation via the Suur Strait cannot be excluded. However Väinameri Archipelago is very shallow (mean depth 5 m) and large area between the open Baltic and GoR. Distance from 20 m isobath in the Baltic Proper through Väinameri to 20 m isobath in the GoR is ca 100 km. Thus, only very strong and long-lasting northerly wind impulse (Otsmann et al., 2001) could cause transport of saltier water via Väinameri Archipelago to the deeper parts of the Gulf of Riga. Temperature—salinity profiles during/after a northerly wind impulse are needed to check the possibility of this pathway.

The existence of fresher sub-surface waters found at the buoy station has also been predicted by numerical modeling (Lips et al., 2016c) in the western part of the gulf as a result of convergence between the low-salinity riverine waters transported from the southern part and waters from the Baltic Proper. The low salinity waters from the southern part of the Gulf of Riga can be transported as far as the buoy station in the present study and even further to the north due to the prevailing whole-basin anticyclonic gyre with enhanced intensity of currents in the western part of the gulf during the summer period (Lips et al., 2016b).

Buoyant intrusions of inflowing waters are well-known features in the Western Baltic Sea. It is the water that has flown through the Danish Straits to the Baltic and which is not dense (light) enough to reach the bottom layers (to stay in the surface layer) of the Baltic. Buoyant temperature maxima form at the depths of the halocline in the Western Baltic if the baroclinic inflows occur during summer (e.g. Feistel et al., 2004). Note that in the Western Baltic case, the halocline (vertical salinity distribution) stabilizes the water column at the depth range of the temperature maxima while, in the Gulf of Riga case, it is the thermocline (temperature), which compensates the effect of the salinity decrease with the depth. Since the maxima formation is sensible to wind forcing, one might expect that the salt and water balance in the gulf will be impacted by the suggested wind regime changes in the future (Christensen et al., 2015).

Such buoyant inflow regime could be found in many semi-enclosed basins where a sill restricts the water exchange with the open sea with higher salinity/density. Currently, such regime is well known only in fjords (e.g. Belzile et al., 2016). In summer when the seasonal thermocline is present, certain forcing conditions could cause deepening of the upper mixed layer and the inflowing water originating from the surface (or thermocline) layer forms a buoyant salinity maximum in a basin.

The comparison of historical density data in the Gulf of Riga and Baltic Proper showed that the salt water maxima formation is feasible from June to September. Otherwise (in October—May), the Baltic Proper water dives to the near-bottom layer of the gulf and such intrusions do not appear. The seasonality in the water exchange regime was confirmed by an analysis of intrusion index: higher values have occurred from May to September. Thus, likely the maxima layer is a common feature of the gulf in summer and has not been noticed earlier (except at two stations by Stipa et al., 1999 in the Irbe Strait) due to very sparse data, which have been available.

The salinity measurements together with current measurements and high-resolution modeling are needed to estimate the role of sub-surface salinity maxima in the salt balance of the gulf. We checked the outputs of the existing operational models HIROMB-EST (Lagemaa, 2012), which uses the HIROMB-SMHI (Funkquist, 2001) model outcome at the open sea boundary, and the HBM (Berg and Poulsen, 2012). Both models did not capture the salt maxima formation and constantly underestimated salinity, by 0.6 g kg^{-1} on average, at least at the profiling location. It is worth to identify if the incapability of reproducing salt intrusions in the models might lead to a salinity underestimation in the whole gulf.

The AOU vs. salinity curves (Fig. 8) suggested that source water for the salt intrusion had almost saturated oxygen content. This result is not a surprise as the water was originating from the upper layer of the Baltic Proper. The Gulf of Riga has a high production (Seppälä and Balode, 1999) and high oxygen consumption (HELCOM, 2009) due to decomposition of organic material in the near-bottom layer. Lateral advection might be an important source of oxygen for the deep layers in the gulf, especially in summer, when strong stratification impedes vertical mixing. However, if the inflowing saltier water does not reach the deeper bottom layers, the bottom remains isolated from this lateral advection of oxygen. Thus, the oxygen conditions in the near-bottom layer of the gulf strongly depend on the water exchange regime in the Irbe Strait. In summer 2015, the regime resulting in buoyant sub-surface saltwater intrusions prevailed, and the near-bottom layer of the gulf did not receive additional oxygen through the lateral advection. This suggestion can be confirmed by the six dissolved oxygen content and AOU profiles (Fig. 11) acquired from the Ruhnu Deep (Fig. 1) from April to August 2015.

Very high dissolved oxygen values were registered on 22 April. The oversaturation in the upper layer due to the spring bloom was up to 125% or 3.8 mg L^{-1} (shown as AOU negative values in Fig. 11b). Water was saturated in the deep layer indicating that winter convection reached the bottom. Oversaturation remained high in the upper layer, but already 1.4—1.9 mg L^{-1} of oxygen was consumed in the deeper layer by 11 May. A continuous oxygen decline occurred from May to August in the deep layers and by the measurements on 26 August, the oxygen level was 5.5 mg L^{-1} at the temperature and salinity of 3.9°C and 6.0 g kg^{-1}, respectively. The estimated oxygen consumption (based on the AOU) from spring to the end of August had been 7.1 mg L^{-1} in the near-bottom layer. According to the HELCOM (2009), the oxygen decrease might lead to hypoxia in the Gulf of Riga but not every year.

5. Conclusions

- Occasionally very strong salinity maxima were observed in the sub-surface layer.
- The sub-surface salt maxima are intrusions of saline and warm water from the Baltic Proper.
- The potential mechanism of the maxima formation likely is a simultaneous upwelling and downwelling event in the Gulf of Riga side and open sea side of the strait, respectively.
- Due to the buoyant nature of the inflowing water, the near-bottom layer of the Gulf of Riga remains isolated from the lateral flows during summers.
- Latter favors the hypoxia formation in the Gulf of Riga.

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HELCOM data (http://ocean.ices.dk/helcom, 25 February 2016) and the level 3 SST (sea surface temperature) product over European Seas by Copernicus Marine Environment Monitoring Service (http://marine.copernicus.eu/) were used in the paper.

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