Winter stratification phenomenon and its consequences in the Gulf of Finland, Baltic Sea

Taavi Liblik, Germo Väli, Inga Lips, Madis-Jaak Lilover, Villu Kikas, Jaan Laanemets

Department of Marine Systems at Tallinn University of Technology, Tallinn 12616, Estonia

Correspondence to: Taavi Liblik (Taavi.liblik@taltech.ee)

Abstract. Stratification plays an essential role in the marine system, with a shallow mixed layer being one of the preconditions for enhanced primary production in the ocean. In the Baltic Sea, the general understanding is that the upper mixed layer (UML) is well below the euphotic zone in winter. In this study, we demonstrate that wintertime UML stratification is common in the Gulf of Finland. Shallow haline stratification forms at a depth comparable to the euphotic zone in late January–early February. Stratification is invoked by a positive buoyancy flux created by the westward advection of riverine water along the northern coast of the gulf after the relaxation of westerly winds. Fresher water and haline stratification occurs approximately one month later in the southern part of the gulf. The onset of restratification is likely associated with the annual cycle of westerly winds, which ease off in late January–early February. Winter restratification can occur in the whole gulf and in the absence of ice, thus, it is a regular seasonal feature in the area. Interannual variations in the wintertime UML correspond with variations in the North Atlantic Oscillation. Chlorophyll *a* concentrations in winter can be comparable to mid summer; the limiting factor for phytoplankton bloom in winter is likely insufficient solar radiation.

1. Introduction

Upper layer stratification is an important characteristic in the dynamics of the pelagic ecosystem. However, to our knowledge, the formation of wintertime haline stratification in the upper layer of the whole Gulf of Finland has not been investigated; the present study focuses on the formation of wintertime haline stratification caused by freshwater inflow and wind forced circulation, and the observed haline stratification explains early phytoplankton dynamics. The Baltic Sea is shallow and brackish, has limited water exchange with the North Sea and is characterized by strong seasonality and gradients of oceanographic parameters (e.g. Leppäranta and Myrberg, 2009). The upper mixed layer (UML), with a typical depth of 10–20 m, forms in spring and is separated from the rest of the water column by a seasonal thermocline. The mixed layer warms up to 15–24 °C (e.g. Stramska and Białogrodzka, 2015; Tronin, 2017) and thermal stratification strengthens until August. The thermocline is eroded by thermal convection, wind stirring and current shear induced mixing, and the mixed layer deepens down to the sea bottom or the halocline at 40–80 m depth in autumn–winter (e.g. Lass et al., 2003; Liblik and Lips, 2017; Väli et al., 2013). This annual stratification cycle has substantial implications for physical, biogeochemical and biological processes in the sea. Characteristics of the pycnocline (e.g. strength) determine vertical fluxes between the surface and sub-thermocline layer. Moreover, the vertical structure of currents is strongly linked to pycnoclines (Suhhova et al., 2018). The annual cycle in stratification, together with solar radiation, mainly determines seasonality in primary production and nutrient consumption. Vertical mixing from the deeper layers, and low production in winter, allows nutrients to accumulate in the upper layer (e.g. Lilover and Stips, 2008; Nehring and Matthäus, 1991). The water column becomes stable in spring and the mixed layer is shallower than the euphotic zone, so that the spring bloom is triggered when solar radiation is sufficiently strong (Fleming and Kaitala, 2006; Jaanus et al., 2006; Lips et al., 2014; Wasmund et al., 1998).

Stratification in the northeastern Baltic Sea is particularly strong and variable. The largest river in the Baltic Sea catchment area, the Neva, discharges into the eastern end of the Gulf of Finland with a mean runoff of 3700 $m^3 s^{-1}$ (Johansson, 2018). Since river discharge is concentrated in the east, and the gulf is connected to the Baltic Proper in the west, there is a mean longitudinal salinity gradient in the upper layer from virtually 0 g kg^{-1} at the easternmost end to 6 g kg^{-1} in the west (Alenius et al., 1998). Also, mean salinity in the upper layer is lower on the northern coast than the southern due to the mean cyclonic circulation in the upper layer and prevailing westward current along the northern coast (Palmen, 1930; Rasmus et al., 2015; Stipa, 2004), (e.g. Kikas and Lips, 2016). Free water exchange between the gulf and the Baltic Proper means that there is a quasi-permanent halocline and saltier deep layer in the gulf. This lateral and vertical structure can be strongly modified by wind forcing; westerly winds drive accumulation of saltier upper layer water, deepen the UML (Liblik and Lips, 2017) and cause weakening of the halocline (Elken et al., 2003). This process can lead to the complete mixing of the water column in the gulf in the winter (Elken et al., 2014; Liblik et al., 2013; Lips et al., 2017). In contrast, easterly winds encourage westward transport of riverine water and strengthen haline stratification in the whole water column (Liblik and Lips, 2017). Wind-driven processes also generate considerable across-gulf inclination of the pycnoclines (Liblik and Lips, 2017) and upwelling and downwelling events along the southern and northern coasts (Kikas and Lips, 2016; Lehmann et al., 2012; Lips et al., 2009).

The northeastern part of the Baltic Sea is ice covered every winter (e.g. Uotila et al., 2015), although ice extent has high interannual variability. The brackish nature of Baltic Sea water means that the maximum density temperature T_{md} (2.2–3.3 °C) is higher than the freezing temperature (from –0.4 to –0.1 °C), unlike most of the world ocean. Thus, when the temperature of the surface layer is below T_{md}, warming increases water density and causes convection and vertical mixing, while cooling stabilizes the water column. Water temperature typically exceeds the T_{md} in northern and eastern parts of the Baltic Sea during winter (Karlson et al., 2016; Liblik et al., 2013), but it is not always the case in offshore areas in the southern Baltic Sea (e.g. Stepanova et al., 2015). Lateral haline buoyancy flux can compensate the thermal convection and stabilization of the shallow upper layer in spring and can occur at temperatures already below T_{md} (Eilola, 1997; Eilola and Stigebrandt, 1998; Stipa et al., 1999). One reason for the latter is the relatively low thermal expansion at temperatures around T_{md}, i.e. the impact of temperature on density is relatively small compared to the impact of salinity. Thus, onset of the seasonal pycnocline is not necessarily initiated by thermal buoyancy but could be related to haline buoyancy. Temperature below T_{md} in the cold intermediate layer after establishment of the seasonal pycnocline provides direct evidence of the latter (Chubarenko et al., 2017; Eilola, 1997; Liblik and Lips, 2017). Haline stratification creates favourable conditions for spring phytoplankton bloom (Kahru and Nõmmann, 1990; Lips et al., 2014); without haline stratification, warming would cause mixing until T_{md} is reached.

Haline stratification under ice has been observed in a number of locations including in the vicinity of River Siikajoki mouth in Bothnian Bay (Granskog et al., 2005), at Tvärminne in the northwestern Gulf of Finland (Merkouriadi and Leppäranta, 2015) and Himmerfjärden bay in the western Baltic Proper (Kari et al., 2018). Ice coverage prevents wind mixing so that even relatively low river runoff can form a plume of fresher water and stratification that can reach 10–20 km from the river mouth. A number of studies (Granskog et al., 2005; Kari et al., 2018; Merkouriadi and Leppäranta, 2015) have investigated winter and early spring haline stratification locally in nearshore regions and near relatively small freshwater sources.

The Gulf of Finland has favorable preconditions for haline stratification in the upper layer in winter. The gulf receives large amounts of fresh water, and it is at least partly covered by ice during winter. The present study hypothesizes that haline stratification occurs at a depth comparable to the euphotic zone in the Gulf of Finland, and potentially in the northeastern Baltic Proper, during wintertime. This means that the general understanding that the water column is mixed down to the halocline in the open Baltic Sea in winter (Leppäranta and Myrberg, 2009) might not be valid in the northeastern Baltic Sea. To test this hypothesis, we analyzed data from research vessel measurement campaigns, autonomously acquired Ferrybox data and historical sources, along with model simulation data.

2. Data and methods

In situ and remote sensing data

We arranged two measurement campaigns in winters 2011/12 and 2013/14 abord RV *Salme* to investigate estuarine circulation reversals in the Gulf of Finland (Fig. 1); six along the gulf surveys were conducted in each winter, with full details of survey and data processing in Liblik et al. (2013) and Lips et al. (2017). In the present study, we utilized temperature, salinity and chlorophyll *a* (Chl *a*) data from cruises in 2011 (21 December), 2012 (24–25 January, 7–8 February, 29 February, 15–16 March) and 2014 (9–10 January, 3–4 February, 4–5 March). Vertical profiles of temperature, salinity and Chl *a* fluorescence were recorded using an Ocean Seven 320plus CTD probe (Idronaut S.r.l.) equipped with a Seapoint Chl *a* fluorometer. The salinity data was calibrated against water sample analyses using a high precision salinometer 8410A Portasal (Guildline). The mean difference and standard deviation of salinity measured by CTD and salinometer was –0.022±0.014 g kg^{-1} in 2011/2012 and –0.009±0.009 in 2013/2014. Thus, after removal of offsets, the accuracy of salinity data was 0.02 g kg^{-1}. Temperature sensors were calibrated before and after surveys in the Idronaut factory and the differences with the calibration device were smaller than the initial accuracy (0.001 °C) of the Ocean Seven 320plus temperature sensor.

Chl *a* fluorescence data was compared and calibrated against water samples on the selected cruises. The linear regression between Chl *a* fluorescence sensor values and Chl *a* acquired from water samples was: Chl *a* = Fl × 1.42 (r2 = 0.90, n = 33), where Fl is the Chl *a* fluorescence recorded by the Seapoint Chl *a* fluorometer. The Chl *a* concentration in the water samples was determined using Whatman GF/F glass fibre filters following extraction at room temperature in the dark with 96% ethanol for 24 h. The Chl *a* content from the extract was measured spectrophotometrically (Thermo Helios g) in the laboratory (HELCOM, 1988). Phytoplankton biomass was determined from water samples from a selection of stations in winter 2014 (Fig. 1). Sub-samples (100 ml) were preserved and analyzed following HELCOM recommendations and EVS-EN 15972:2011 standard. Phytoplankton carbon (C) content was calculated using the C: biovolume factors method of Menden-Deuer and Lessard (2000) and for photosynthetic naked ciliate Mesodinium rubrum according to the method of Putt and

Stoecker (1989).

Wind data were recorded at Tallinnamadal and Kalbådagrund lighthouses (Fig. 1) at heights of 36 m and 32 m above sea level and 1-h and 3-h intervals, respectively. A height correction coefficient of 0.91 (neutral atmospheric stratification) was applied to convert wind speed measurements to 10 m height equivalent (Launiainen and Saarinen, 1984). Wind measurements from Tallinnamadal for winters 2011/12, 2013/14 and 2015/16 were used in the oceanographic data analysis. The Kalbådagrund dataset for the period 1981–2015 was used to illustrate the annual cycle of the along-gulf component of wind stress.

Ferrybox measurements of temperature and salinity between Tallinn–Helsinki for January–March 2012, 2014 and 2016 were also used in the study. Details of the Ferrybox system and data processing methods are given in Kikas and Lips (2016). Analyses have shown that a correction of 0.08 g kg^{-1} (the value has been stable over the years) must be added to the recorded salinity (Kikas and Lips, 2016). The standard deviation of the difference in salinity measured by Ferrybox and a high precision Portasal salinometer was 0.01 g kg^{-1} after bias correction. The accuracy of the Ferrybox temperature sensor is 0.04 °C (Kikas and Lips, 2016).

Historical data collected by the Department of Marine Systems at Tallinn University of Technology and the ICES HELCOM dataset (https://ocean.ices.dk/helcom/) were used to determine past stratification conditions. Quality assurance and data processing were in accordance with the HELCOM Monitoring Manual (Anon, 2017).

OSTIA (Donlon et al., 2012; Good et al., 2020) daily mean sea surface temperature (SST) data for the period 2010–2019 were obtained from the Copernicus Marine Environment Monitoring Service. The mean difference between the OSTIA SST product and in situ measurements is 0.01–0.03 °C and the standard deviation is 0.4–0.5 °C (Worsfold et al., n.d.). Daily mean SST along the thalweg in the Gulf of Finland (Thalweg GoF in Fig. 1) and in the Gotland Deep (box in Fig. 1) was calculated to determine if and when SST was above or below T_{md}. Salinities of 6 g kg^{-1} and 7 g kg^{-1} were used in T_{md} estimation for the Gulf of Finland and Gotland Deep, respectively.

Time series of the large-scale North Atlantic Oscillation (NAO) index was used to explain the interannual variability of wintertime upper layer stratification in the Gulf of Finland. Long-term observations of the sea level pressure differences between Reykjavik (Iceland) and Gibraltar (Spain) constitute the NAO index, which is available from the Climatic Research Unit, University of East Anglia (Jones et al., **1997**; **https://crudata.uea.ac.uk/cru/data/nao/nao.dat**).

Density is given as a potential density anomaly (σ_0) to a reference pressure of 0 dbar (Association for the Physical Sciences of the Sea, 2010). The UML depth was defined as the minimum depth where $\rho_z \geq \rho_3 + 0.15$ kg m^{-3} was satisfied. The density at 3 m depth is ρ_3 and ρ_z is the density at depth z.

Modeling

The study used the General Estuarine Transport Model (GETM, Burchard and Bolding (2002)) to obtain UML parameters in the Gulf of Finland and the Eastern Baltic Sea. GETM is a primitive equation 3-dimensional, free surface hydrostatic model with a built-in vertically adaptive coordinate scheme (Hofmeister et al., 2010). The latter has been shown to significantly reduce numerical mixing in the simulations(Gräwe et al., 2015).

Vertical mixing in the GETM is calculated using a General Ocean Turbulence Model (GOTM, Umlauf and Burchard (2005)). For the current study, the eddy diffusivity and eddy viscosity parameters were found using a two-equation k-ε model coupled with an algebraic second-moment closure (Burchard et al., 2001; Canuto et al., 2001).

A horizontal grid spacing of 0.5 nautical miles (approximately 926 m) was established for the setup domain of the whole Baltic Sea (Fig. 1), with 60 vertically adaptive layers. Parameters controlling the vertical resolution of the model during simulations were taken from Hofmeister et al. (2010) and Gräwe et al. (2015). The digital topography of the Baltic Sea was taken from the Baltic Sea Bathymetry Database (http://data.bshc.pro/, last accessed 1 April 2020), with additional data for the Gulf of Finland from Andrejev et al. (2010). Surface boundary conditions (wind stress and surface heat flux components) were calculated using bulk formulae from data generated by the operational forecast model HIRLAM (High-Resolution Limited Area Model). HIRLAM is used and maintained by the Estonian Weather Service and has a spatial resolution of 11 km and a daily forecast interval of 1 h for a total forecast length of 54 h (Männik and Merilain, 2007). All meteorological parameters were interpolated to the model grid. Model simulation was performed from 1 April 2010 to 31 December 2019.

Open boundary conditions were used in the Danish Straits. Inflow and outflow from the model is barotropically controlled using sea surface height measurements from Gothenburg station and, more specifically, Flather (1994) radiation. In terms of temperature and salinity, the model is relaxed towards climatological profiles along the open boundary using sponge layer factors according to the method of Martinsen and Engedahl (1987). The simulation used freshwater input from the 54 largest Baltic Sea rivers, together with their interannual variability as reported in HELCOM (Johansson, 2018). The riverine input is treated as a rise in the sea surface height and each river has a prescribed constant salinity of 0.5 g/kg that is diluted in the corresponding grid cell. River water temperature is assumed to be the same as that in the target cell.

The initial thermohaline field was taken from the Copernicus reanalysis of the Baltic Sea for the period 1989–2014. As the product provides a horizontal resolution of 3 nautical miles (approximately 5.56 km), and a vertical resolution from 5 m at the surface up to 50 m in the near-bottom layers, it was interpolated to the target grid. Model simulations started from a motionless state, that is with initial sea surface height and current velocity set to zero. Previous studies (e.g. Lips et al., 2016) have shown that wind-driven circulation in the Baltic Sea adjusts to forcing within 5 days.

Model validation used available Ferrybox data (2011–2016) along the Tallinn to Helsinki transect (see Fig. 1 for location). The model captures the observed variability of temperature and salinity reasonably well (Fig. 3). Standard deviations of simulated temperature and salinity for the overall (1 November 2011–1 June 2016) and wintertime (December to March 2011–2016) periods are close to observations. The standard deviation of simulated salinity is smaller than the observed for winter 2016 (January–March) and larger for 2012, while for 2014 it is close to the observed. The variability of temperature is captured well – standard deviations from the simulations are at least 0.8 of the observed for all time periods, although the model slightly overestimated temperature variability for the winter of 2012.

The overall correlation coefficient for salinity is 0.62, while it is over 0.74 for both the whole wintertime period and single years as well. There is a higher correlation for temperature (as expected); overall correlation, which includes seasonal variability, is 0.99, and for wintertime it is 0.95. Very high correlation (>0.94) for temperature is also shown for individual winters. Root mean squared differences between model and observed values are slightly larger for salinity but do not exceed the observed variability. In general, the model captures wintertime changes in the surface layers of the Gulf of Finland well. More details about model setup and validation in the Baltic Proper are given in Zhurbas et al. (2018).

3. Results

3.1. Onset of stratification and its link to wind forcing

To demonstrate the link between wind forcing, onset of stratification and increase in Chl *a*, we analyzed temperature, salinity, density and Chl *a* distributions along the gulf together with wind data for winters 2011/12 and 2013/14. Prior to the survey of 21 December 2011, there was a strong westerly wind with a maximum along gulf wind stress of 1.3 N m^{-2} (Fig. 3a). Cumulative wind stress increased by 6 N m^{-02} d from 1 November to 21 December, resulting in a warm (>5 C°, Fig. 4a), relatively salty (>6.3 g kg^{-1}, Fig. 4b) and well-mixed water column in the gulf (Fig. 4c). Chl *a* concentrations were very low, below 1 mg m^{-3} (Fig. 4d). Prior to the survey on 24–25 January 2012, weaker easterly winds had prevailed since mid January (Fig. 3a). Lower temperature (3–4 C°, Fig. 4e) in the upper 20 m coincided with slightly fresher water on 24–25 January 2012 (Fig. 4f). A salinity minimum (down to 5.8 g kg^{-1}) caused stratification in the upper layer (Fig. 4g) at a distance of 80–110 km in the section; this location was also characterized by slightly higher Chl *a* concentration (up to 1.5 mg m^{-3}) (Fig. 4h). Variable and relatively weak winds prevailed in late January and early February (Fig. 3a). On 7–8 February 2012, temperature of the upper layer was below T_{md} (2.7 °C) (Fig. 4i), salinity was low (4.8–6.0 g kg^{-1}, Fig. 4j) and there was a marked stratification and shallow UML (Fig. 4k). Higher Chl *a* concentration, occasionally >2 mg m^{-3}, was seen in the fresher and colder water along the section (Fig. 4l). Lateral Chl *a* extent was closely linked to the salinity (density) structure, with higher Chl *a* concentration associated with lower salinity and vice versa. Westerly winds prevailed in the period before the next survey at the end of February (Fig. 3a), resulting in well mixed conditions and relatively high salinity (6.0–6.7 g kg^{-1}) in the western part of the section on 29 February (Fig. 4m–n). Lower salinity, stronger stratification and slightly higher Chl *a* in the upper layer were observed in the central part of the section (Fig. 4n–p). The eastern part of the section was not visited on 29 February due to ice conditions. In the middle of March (15–16 March) the water temperature was still well below T_{md} and strong haline stratification was observed along the whole transect (Fig. 4r–t). Chl *a* concentrations in the upper layer were within the range 2–4 mg m^{-3} (Fig. 4u).

Similar trends in wind forcing and spatiotemporal patterns of temperature, salinity, density and Chl *a* were observed in winter 2013/14. Strong westerly winds dominated until early January 2014, with an increase in cumulative wind stress of 10 N m^{-2} d from 1 November 2013 (Fig. 3b). The 9–10 January 2014 survey shows a well-mixed water column and low Chl *a* (5a–d). Fresher and colder water, but only slightly higher Chl *a*, was found in the upper layer on 3–4 February (Fig. 5e–h). By 4–5 March, the area of fresher water had expanded (salinity <6 g kg^{-1}) and the UML extended over most of the section (Fig. 5j, k). The cold and fresher upper layer showed higher Chl *a*, especially in the eastern part of the section (Fig. 5i, j, l).

Thus, haline stratification and elevated Chl *a* concentration was observed in both winters (2011/12 and

2013/14) from the beginning of February. A shallow UML (<20 m) was absent after prevailing westerly winds and when SST was >T_{md}. Stratification formed as fresher water occupied the upper layer.

To examine temporal trends in haline stratification in more detail, we analyzed across the gulf changes in temperature and salinity using measurements acquired by the Ferrybox system along the Tallinn–Helsinki transect for January–March 2012, 2014 and 2016 (Fig. 6). Generally, temporal changes in salinity and temperature along the transect were quite similar for each of the study years, as was wind forcing (Fig. 3). Strong westerly winds dominated until early or mid January, and after relaxation of wind forcing, fresher water was recorded in the transect.

Based on observations at the longitudinal sections (Figs. 4 and 5), the highest sea surface salinity at which stratification and relatively shallow UML can form was assumed as 6 g kg^{-1}. Similar to the along-gulf observations (Fig. 4a, b), salty and warm water occupied the transect at the beginning of January 2012 (Fig. 6a, b). The northern part of the transect was covered in fresher water (< 6 g kg^{-1}) by the end of January, although salinity slightly increased in the southern part of the section at this time. Since the main sources of freshwater are in the east, water must have flown westward along the northern coast. The area covered by fresher water widened to almost the entire section by mid February. Water temperature declined below T_{md} in the northern part in the first half of January, while in the central and southern part of the section temperature dropped below T_{md} by the end of January. A similar spatiotemporal pattern in sea surface salinity was observed in 2014 and 2016 (Fig. 6c–f). Fresher water first appeared in the northern part in the first half of January in both 2014 and 2016. The onset of haline stratification occurred slightly earlier in 2016 due to wind forcing – the westerlies had eased off by the end of December 2015 (Fig. 3c). The segment covered by fresher water widened during January and most of the transect was occupied by water with salinity <6 g kg^{-1} at the end of January 2016 and in mid February 2014. A pulse of strong westerly wind occurred at the end of January–beginning of February 2016 (Fig. 3c). We suggest that the lighter, less saline water that originates in the east flowed westwards along the northern coast and was later transported to the southern coast in the central and western part of the gulf. The latter is likely related to the Ekman transport induced by the westerly wind impulse (Fig. 3). Thus, stratification related to the spreading of fresher water forms about one month earlier in the northern part of the gulf than in the southern part.

3.2. Spatiotemporal patterns of restratification

Here, we examine the spatiotemporal pattern of the restratification process using model simulation data and statistics of historical observations. As noted from the in situ observations, haline stratification forms after the relaxation of westerly winds. The annual cycle of the along-gulf component of wind stress shows higher monthly mean values (>0.04 N m^{-2}) and higher variability from October–January (Fig. 7); this means that strong westerly winds are more frequent and storminess is higher in these months. As a consequence, UML depth <20 m was infrequent and mean UML depth varied between 40–60 m in the western and central gulf in November, December and January 2010–2019 (Fig. 8). As an exception, the probability of UML depth <20 m was 30–40% in the northern part of the eastern area in January. Winds from the west are weaker and storms are less frequent in February and March (Fig. 8). In February, the occurrence of UML depth <20 m increased to 50–60% (Fig. 8), although in the southern and western parts of the gulf, mean UML depth was 30–40 m. The statistics from model simulation data agree well with our observations of westward advection of fresher water from the northern coast (Fig. 6). Mean UML depth was 20 m or lower in the central part of the gulf in March, and thicker at the gulf entrance (Fig. 8); the occurrence of UML depth <20 m was >60% in the central part, around 50% at the gulf entrance and much lower to the west of longitude 22° E (Fig. 8). A similar pattern is shown in the mean occurrence of the density difference between 40 m depth and the sea surface of >0.5 kg m^{-3}, based on in situ measurements for the period 1904–2020 (Fig. 9): occurrence was 40–75% in the central part of the gulf, 30–50% in the entrance of the gulf and <5% further to the west. Thus, wintertime upper layer stratification extends to 23°E in the western gulf.

Model simulation data (2010–2019) were used to examine the development of UML depth from October to March along a transect from the northern Baltic Proper to the central Gulf of Finland and in the Gotland Deep. The time series of mean UML depth for transects in the gulf (Thalweg GoF, Fig. 1) and Gotland Deep (box, Fig. 1) show that the maximum mean UML depth in the gulf mostly occurred in December, well before SST decreased to T_{md} (Fig. 10a). The onset of restratification occurred at temperatures below T_{md} (Fig. 10b). Temperature dropped below T_{md} later and rose above T_{md} earlier in the Gotland Deep compared to the gulf. In five winters out of ten SST did not fall below T_{md} in the Gotland Deep (Fig. 10b). However, whether the temperature was below T_{md} or not, restratification phenomenon were absent from the upper layer in the Gotland Deep in January–March; this means that buoyancy, created by slight thermal stratification at $<T_{md}$, is overshadowed by vertical mixing in the Gotland Deep. Vertical mixing also dominated in the Gulf of Finland in November–December. Still, from late January or early February, the advection of fresher water (Fig. 6) creates a shallow mixed layer (Fig. 10a-c).

Time series of simulated UML depth along the transect from the northern Baltic Proper to the central Gulf of Finland from October to March in 2010–2019 showed considerable synoptic and interannual variability (Fig. 10c). The deepest UML occurred in the gulf in winters 2011/12 and 2013/14, i.e. precisely the years when measurements along the thalweg also showed deep UML in the gulf (Figs. 4 and 5). The estuarine circulation reversal caused by strong westerly winds gave rise to a deep UML, while restratification occurred after prevailing easterly winds (Figs. 3–5). The frequency of westerly (easterly) winds over the Gulf of Finland in winter is positively (negatively) correlated to the NAO index (Jaagus and Kull, 2011). The strong reversal event and deep UML in winter 2011/12 were accompanied by an anomalously positive NAO index (Liblik et al., 2013). The mean December to February NAO index in 2011/12 was 2.18. Likewise, the mean NAO index in the other three winters (2013/14, 2014/15, 2015/2016) when mean UML depth in the gulf reached 60 m or deeper was > 2 in December–February (Fig. 10b). Winters 2010/11 and 2012/13, which stand out in the time series with early onset restratification in early January, had the lowest December–February averaged NAO indices during the period 2010–2019: –1.06 and 0.47, respectively (Fig. 10 a and -b). Thus, large scale atmospheric forcing alters the restratification. Low NAO index and easterly winds support restratification while high NAO index and westerly winds have the opposite effect.

2. Discussion

Positive net buoyancy flux is required for the onset of stratification in the upper layer. Processes causing negative buoyancy fluxes include vertical mixing caused by wind stirring, current shear and convection. Positive buoyancy fluxes result from advection (arrival) of lighter water to the sea surface or of denser water to the subsurface. Likewise, warming of the surface layer at temperatures above T_{md} or cooling below T_{md} strengthens stratification. The magnitude of positive buoyancy imparted from the cooling of water below T_{md} is rather small. If we consider salinity of 6 g kg^{-1}, the density difference between waters at T_{md} (2.8 °C) and freezing temperature (–0.3 °C) is 0.07 kg m^{-3}. This is the maximum density change if the water temperature is below T_{md} and salinity is 6 g kg^{-1}. We get the same density difference if we keep temperature constant (1 °C) and vary salinity by 0.09 g kg^{-1}. Our data show that changes in sea surface salinity in winter are of the order of 1–2 g kg^{-1} (Fig. 6), so the effect of salinity change to the density and buoyancy flux is about 10–20-fold higher than the effect of temperature change in the gulf. We can conclude that fresher water advection from the east is the primary source of buoyancy for the development of the stratification. Freshwater transport is controlled by wind forcing; easterly winds support advection of fresher water to the west while westerly winds impede it (Liblik and Lips, 2012; Pavelson et al., 1997). To exemplify the processes, two cases from the Gulf of Finland are illustrated in Fig. 11. When westerlies dominate, riverine water transport to the west is blocked, vertical mixing is strong and the UML is deep (Fig. 11a), while easterly or weak winds are associated with westwards advection of fresher water along the northern coast, formation of haline stratification and phytoplankton growth (Fig. 11b).

The role of the wind forcing in stratification depends on preexisting conditions. Westerly winds generally deepen the mixed layer depth due to transport of the surface layer water from the northern Baltic Proper to the gulf (Liblik and Lips, 2017). However, if fresher water is already present along the north coast, westerly winds spread it to the south and create stratification there (Figs. 3 and 6), as noted in summer also by Pavelson et al. (1997).

Wintertime stratification phenomenon in nearshore regions, extending 10–20 km from the coast, have been reported in several locations in the Baltic Sea (Granskog et al., 2005; Kari et al., 2018; Merkouriadi and Leppäranta, 2015). However, these studies were concerned with stratification under the ice, whereas in our study of the Gulf of Finland, we have shown that wintertime stratification may also occur at the basin-scale (along-gulf extent 400 km) and in the absence of considerable ice coverage. During the onset of restratification in late January 2012, the Gulf of Finland was virtually ice-free. In both winters (2011/12 and 2013/14), only the eastern part of the Gulf of Finland and the adjacent northern shore of the gulf were ice covered at the end of January; thus, winter stratification phenomenon occurred even when most of the Gulf of Finland was not covered by ice. It should be noted that, along with the frequency of easterly winds, low NAO index is also associated with increased ice coverage (Jaagus, 2006). The landfast ice zone would be expected to prevent vertical mixing and therefore supports lateral advection of riverine fresher water (Granskog et al., 2005).

The western border of the stratification phenomenon is around 23° E, i.e. at the entrance to the gulf between Hiiumaa Island and the Finnish coast. This means vertical mixing dominates over lateral buoyancy fluxes in the Baltic Proper and shallow stratification is not a common feature. The absence of the phenomenon in the Baltic Proper can be explained by the long distance from rivers, due to its larger size and topography. Riverine input per unit area in the Gulf of Finland is 7–8 times larger than in the Baltic Proper (Leppäranta and Myrberg, 2009). As the wintertime stratification phenomenon vanishes at the wider entrance area to the Gulf of Finland, it is likely that the elongated, narrow shape of the gulf accounts contributes to the formation of stratification as well as high freshwater input. In the northern part of the Gulf of Finland, occurrence of the shallow (<20 m) halocline reached over 50% in February, and in the southern part it reached over 50% in March. The high synoptic-scale and interannual variability of UML depth can be related to the wind regime and NAO index (Janssen et al., 2004), respectively. High NAO index is associated with high wind stress, low ice cover, strong upwelling/downwelling (Janssen et al., 2004) and extreme estuarine circulation reversal events (Liblik et al., 2013; Lilover et al., 2017; Lips et al., 2017; Suhhova et al., 2018). Enhanced vertical transport by upwelling/downwelling, wind stirring, and reversal events cause vertical mixing, deepening of the UML, and upward transport of nutrients from the deeper layers (Janssen et al., 2004; Lilover and Stips, 2008; Lips et al., 2017). Conversely, low NAO index supports the consumption of riverine nutrients in the Gulf of Finland while the vertical mixing of nutrients from the deeper layer is modest.

We observed Chl *a* concentration up to 3.0 and 4.5 mg m^{-3} from February to the first half of March, respectively in 2012 and 2014, i.e. occasionally comparable with the mean values in summer in the Gulf of Finland (Kononen et al., 1998; Suikkanen et al., 2007). The higher Chl *a* concentration coincided with a cold and fresher upper layer and stronger stratification. The distribution of phytoplankton biomass concentration (Fig. 12) generally follows Chl *a* structure (Fig. 5) in winter 2014. The observed winter biomass concentrations were much lower compared to summer values in the Gulf of Finland (Kononen et al., 1998, 1999). This discrepancy is probably related to the biomass/ Chl *a* ratio being higher in summer, as shown for instance in the southern Baltic (Lyngsgaard et al., 2017). The dominant species in the phytoplankton community in more elevated biomass patches in February was the photosynthetic ciliate *Mesodinium rubrum*, which is often the dominant primary producer in the post-spring bloom period (Lips and Lips, 2017). In March, *M. rubrum* dominated in the western part of the study area, whereas at other stations the spring bloom dinoflagellates were equally abundant as the photosynthetic ciliate.

Spring bloom is instigated when phytoplankton growth exceeds losses in the upper layer due to grazing or vertical mixing downwards (Smetacek and Passow, 1990). Necessary conditions for spring bloom are a stabilized upper layer that is thinner than the depth of the euphotic zone, available nutrients and strong enough solar radiation (Fennel, 1999). UML depth was 10–20 m in most of the Gulf of Finland in early March 2012 and 2014. Euphotic layer depth, estimated according to Luhtala and Tolvanen (2013) from our Secchi depth measurements, was 15–19 m in both winters, i.e. comparable with the UML depth. Also, there were sufficient nutrients available in the upper layer in February and March 2014 (Lips et al., 2017). We do not have a reference for the nutrients data in 2012. Thus, the limiting factor for phytoplankton growth is likely insufficient solar radiation. The mean downward shortwave radiation doubles in the area from February (40–50 Wm^{-2}) to March (90–100 Wm^{-2}) and quadruples in April (160–200 W m^{-2}) (Rozwadowska and Isemer, 1998; Zapadka et al., 2020). The onset of spring bloom typically occurs in April in the Gulf of Finland (Groetsch et al., 2016; Lips et al., 2014; Lips and Lips, 2017).

3. Conclusions

Using in situ measurements and model simulation, we have demonstrated wintertime occurrence of haline stratification at a depth comparable to that of the euphotic zone in in the Gulf of Finland, well before the onset of thermal stratification in spring. Stratification forms in late January–early February as a result of the westwards advection of riverine water along the northern coast of the gulf. Stratification is maintained by the positive buoyancy flux created by the advection, which is stronger than the negative flux resulting from vertical mixing. The advection of riverine water occurs after easing of westerly winds; relaxation of westerly winds is a part of the annual cycle of the local wind regime and thus the mid winter restratification is a regular seasonal feature in the area. Fresher water and haline stratification occurred approximately one month later along the southern coast of the gulf. Earlier observations of a local stratification phenomenon in the Baltic Sea in winter were under conditions of ice coverage. Our observations show that haline stratification can occur in the whole Gulf of Finland and in the absence of ice cover. Therefore, we can assume that wintertime stratification is a common phenomenon in the Gulf of Finland and its western boundary is at the entrance area to the gulf, between Hiiumaa Island and the Finnish coast.

Elevated Chl *a* and phytoplankton biomass was registered in the UML in the Gulf of Finland before the spring bloom. The limiting factor for phytoplankton growth in winter is likely insufficient light radiation. The exact role of wintertime stratification in the nutrient cycle and phytoplankton dynamics in the Gulf of Finland needs further investigation.

Code availability. Scripts to analyze the results are available upon request. Please contact TL.

Author contributions. TL led the analyses of the data and writing of the manuscript with contributions from GV, JL, M-JL and IL. TL was responsible for the measurements and GV for the modelling activities. VK was responsible for gathering and processing of the Ferrybox data. IL arranged phytoplankton biomass measurements.

Competing interests. We declare that no competing interests are present.

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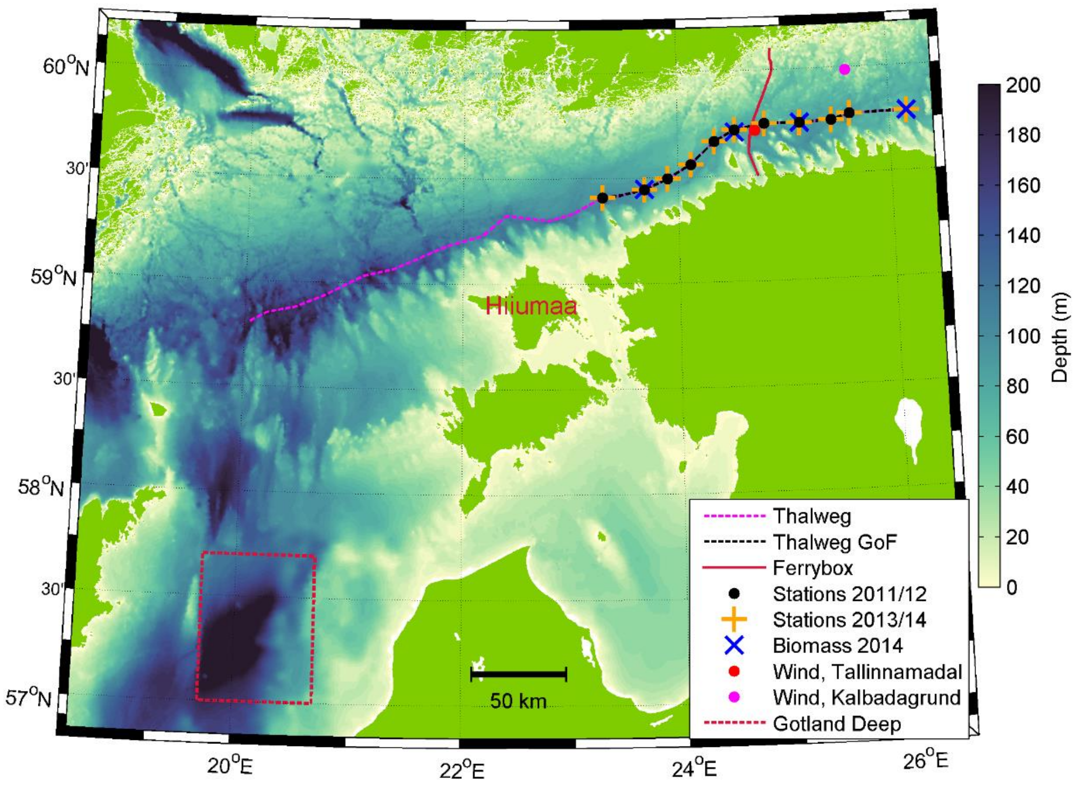


Fig. 1. Bathymetric map of the Baltic Proper and the Gulf of Finland. Thalweg from the Central Gulf of Finland, CTD stations visited in 2011/12 and 2013/14, phytoplankton biomass sampling stations, Tallinn–Helsinki Ferrybox line, Tallinnamadal and Kalbådagrund wind measurements locations and Gotland Deep area are shown.

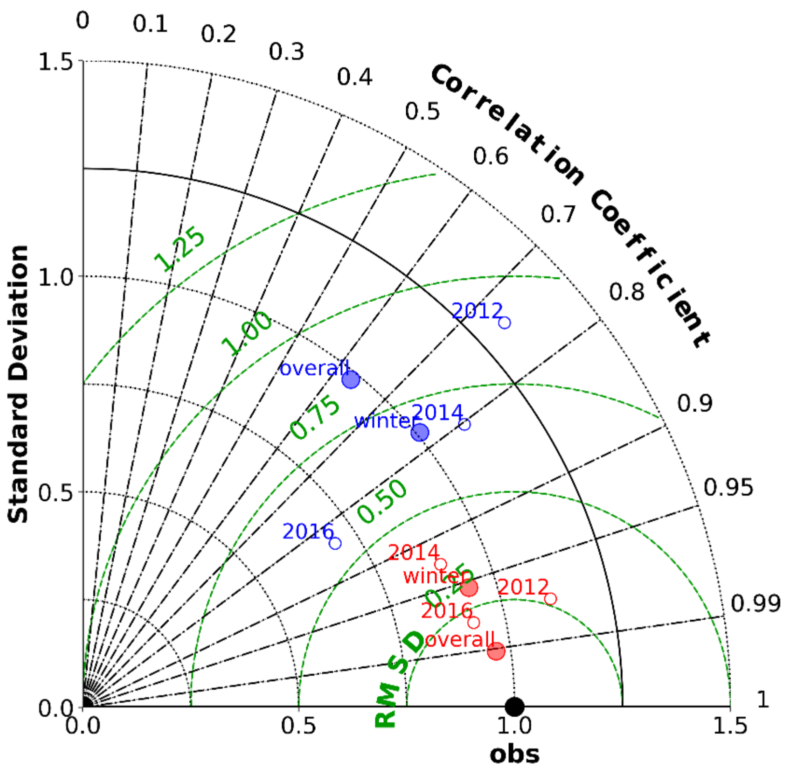


Fig. 2. Taylor diagram of simulated and measured temperature (red) and salinity (blue) along the Ferrybox transect from Tallinn to Helsinki. Overall – all available observations from 1.11. 2011 to 1.06. 2016 (filled circles); winter – all available observations from December to the end of March, 2011–2016 (filled circles) and winter observations from January–March in 2012, 2014 and 2016 (open circles).

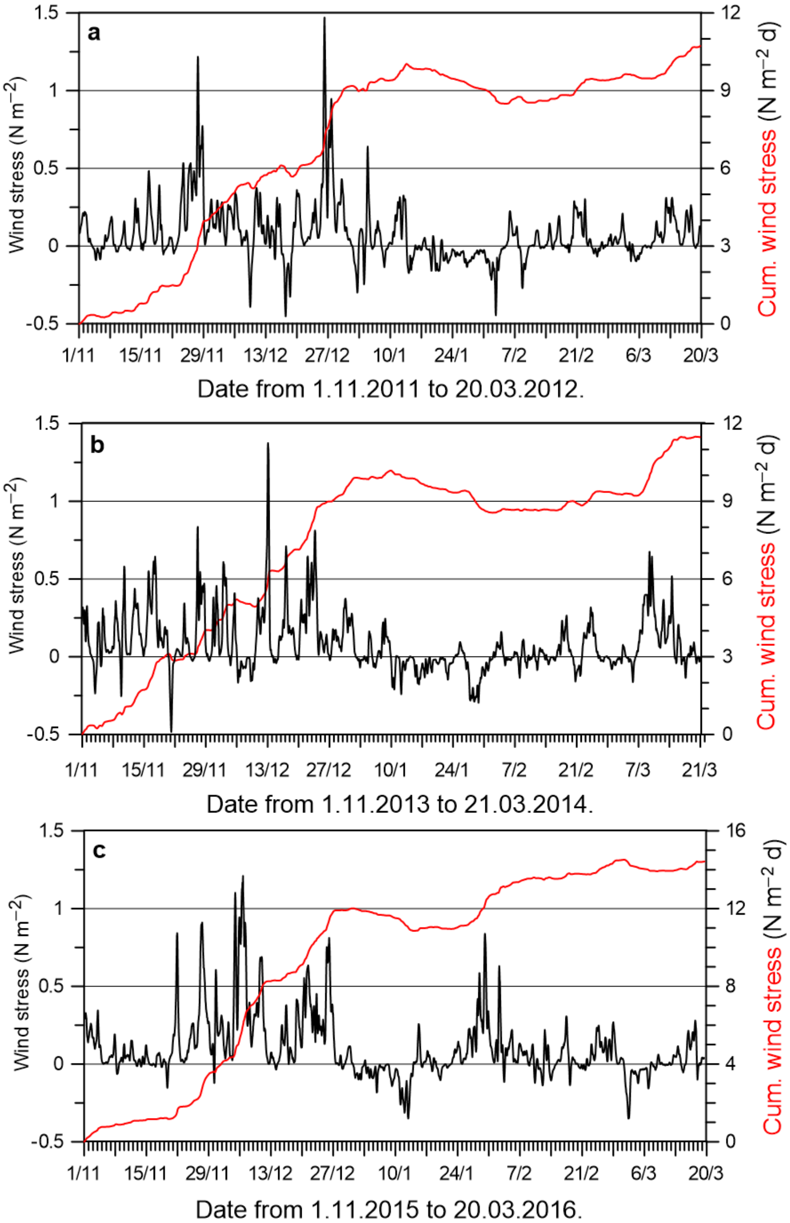


Fig. 3. Time series of an along-gulf component of wind stress (black curve, positive eastward) and cumulative along-gulf wind stress (red curve), based on wind data measured at Tallinnamadal Lighthouse in the Gulf of Finland. (a) 1 November 2011 to 20 March 2012; (b) 1 November 2013 to 21 March 2014; (c) 1 November 2011 to 20 March 2016.

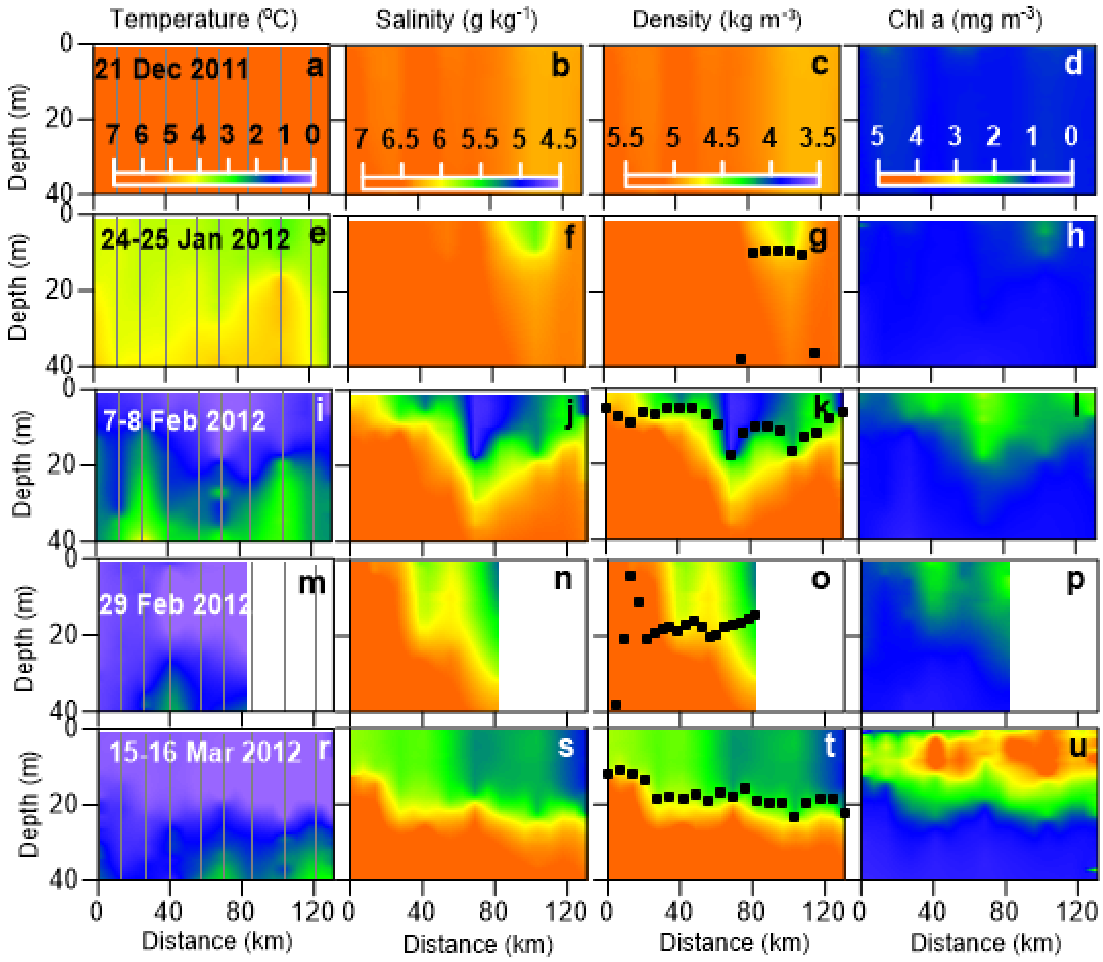


Fig. 4. Vertical sections of temperature, salinity, density anomaly and Chl *a* along a west to east profile in the Gulf of Finland (black dots in Fig. 1) in winter 2911/12.. Vertical gray lines mark the location of CTD-casts. Black dots on density anomaly panels mark the depth of the upper mixed layer.

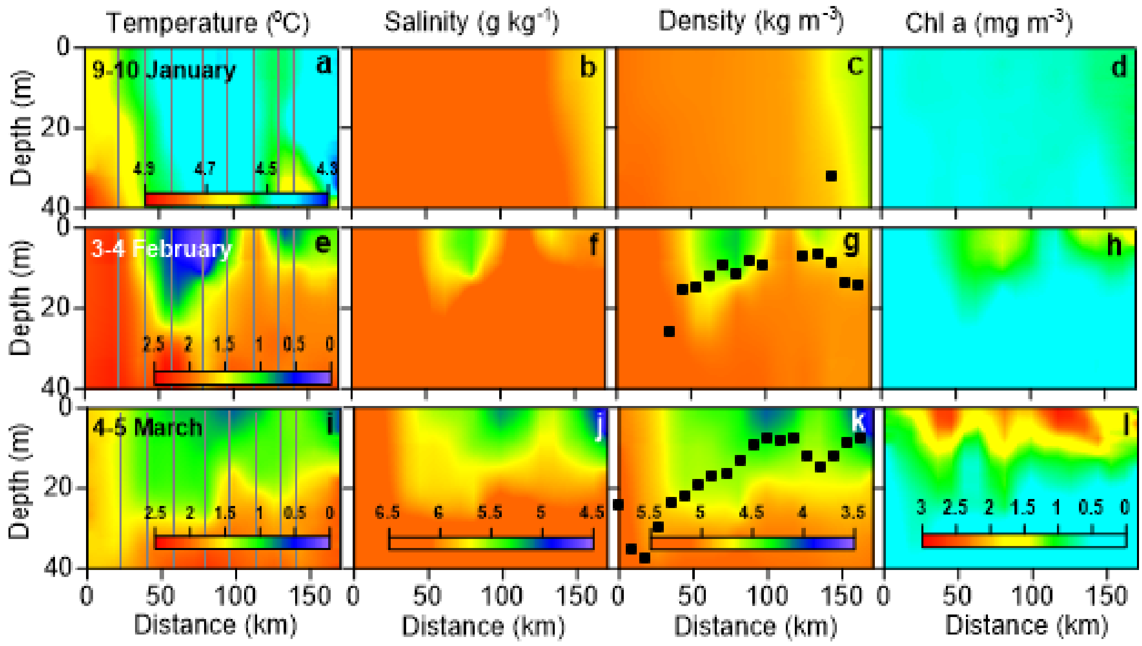
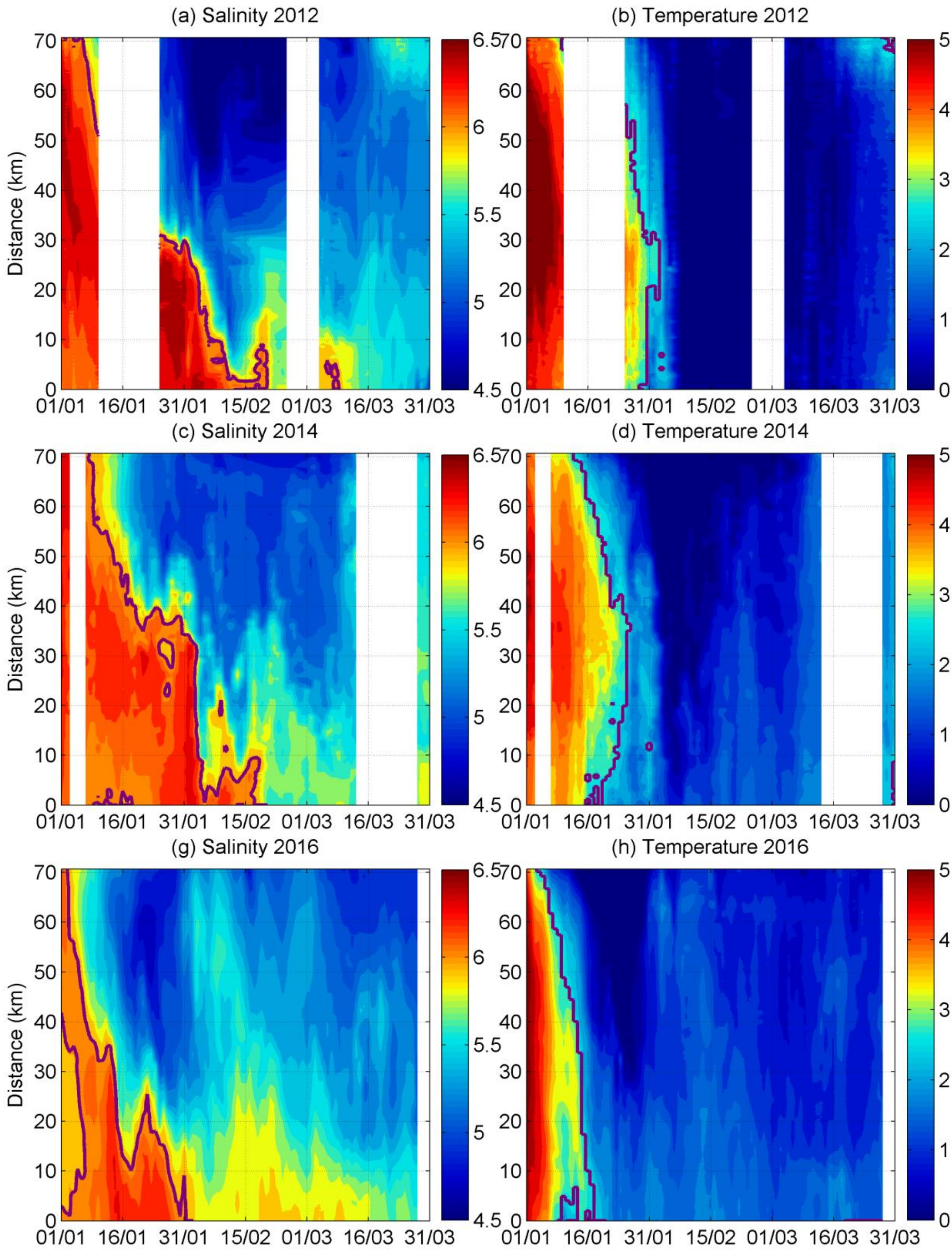


Fig. 5. Vertical sections of temperature, salinity, density anomaly and Chl *a* along a west to east profile in the Gulf of Finland (orange crosses in Fig. 1) in winter 2013/14. Vertical gray lines mark the location of CTD-casts. Black dots on density anomaly panels mark the depth of the upper mixed layer.



645 **Fig. 6. Salinity (g kg^{-1}) and temperature (°C) of the upper layer along a transect from Tallinn to Helsinki (red line in Fig. 1) in January–March 2012, 2014 and 2016. The isoline 6 g kg^{-1} is marked on the salinity plots and the maximum density temperature T_{md} on the temperature plots. The starting point of the transect (x = 0 km) is in the Bay of Tallinn at 59.500° N and 24.752° E.**

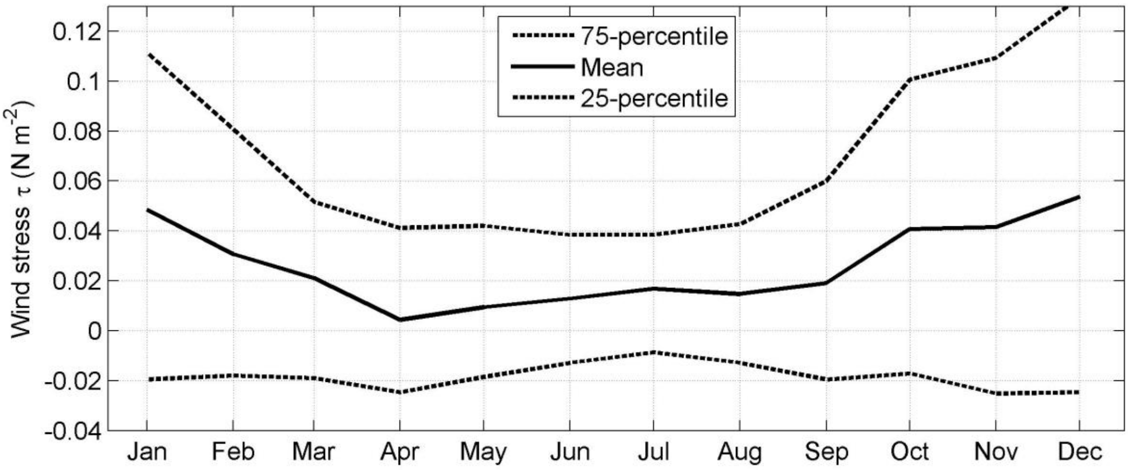


Fig. 7. Annual cycle of an along-gulf component of wind stress. Mean (solid black curve, positive eastward), 75- and 25-percentiles (dashed lines) are based on data from 1981-2015, measured at Kalbådagrund Lighthouse, Gulf of Finland.

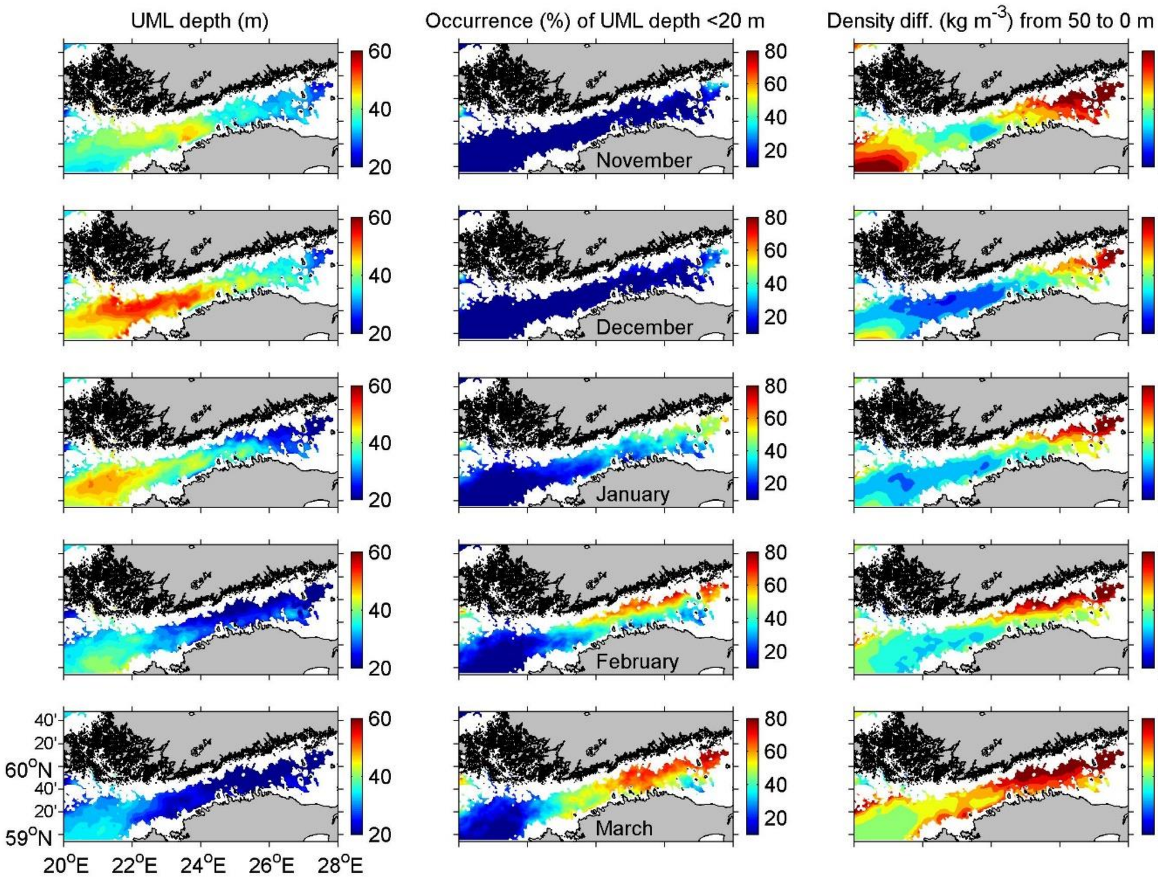


Fig. 8. Mean simulated upper mixed layer (UML) depth, percent occurrence of UML depth <20 m and density difference between 50 and 0 m depth from November to March, 2010 to 2019.

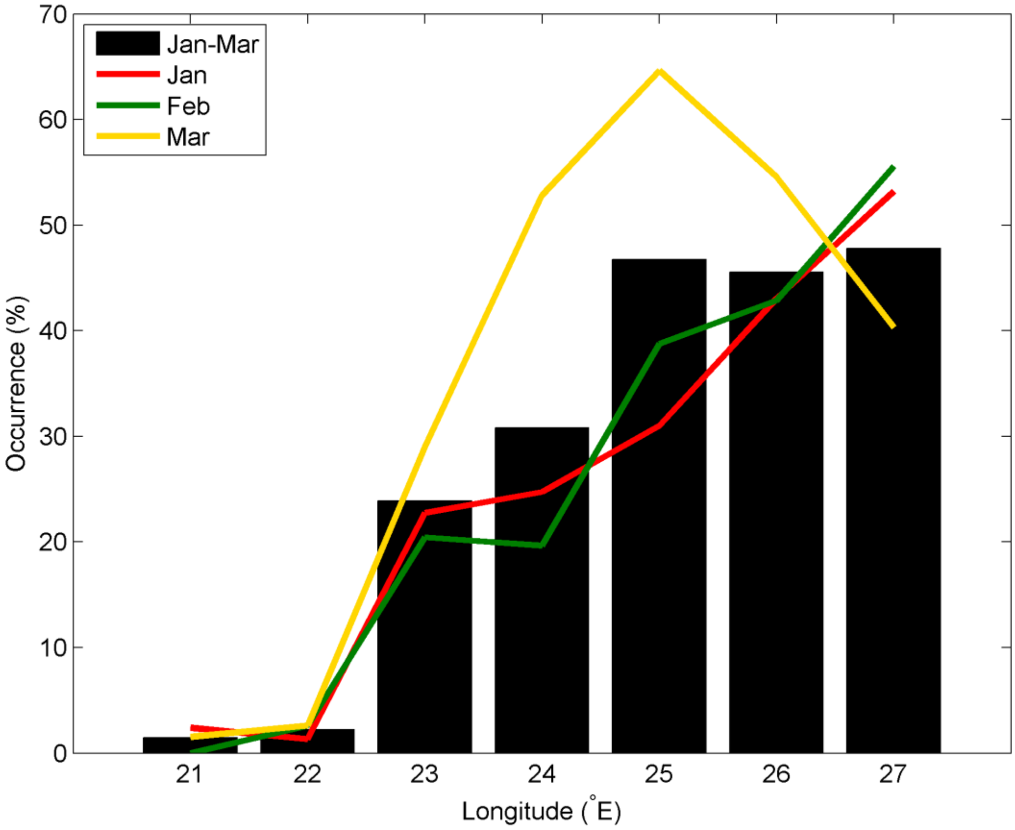


Fig. 9. Occurrence of >0.5 kg m^{-3} density difference between 40 m depth and the sea surface in the Gulf of Finland from January to March 1904–2020. A total of 2560 temperature-salinity data pairs for the surface layer and 40 m depth are included.

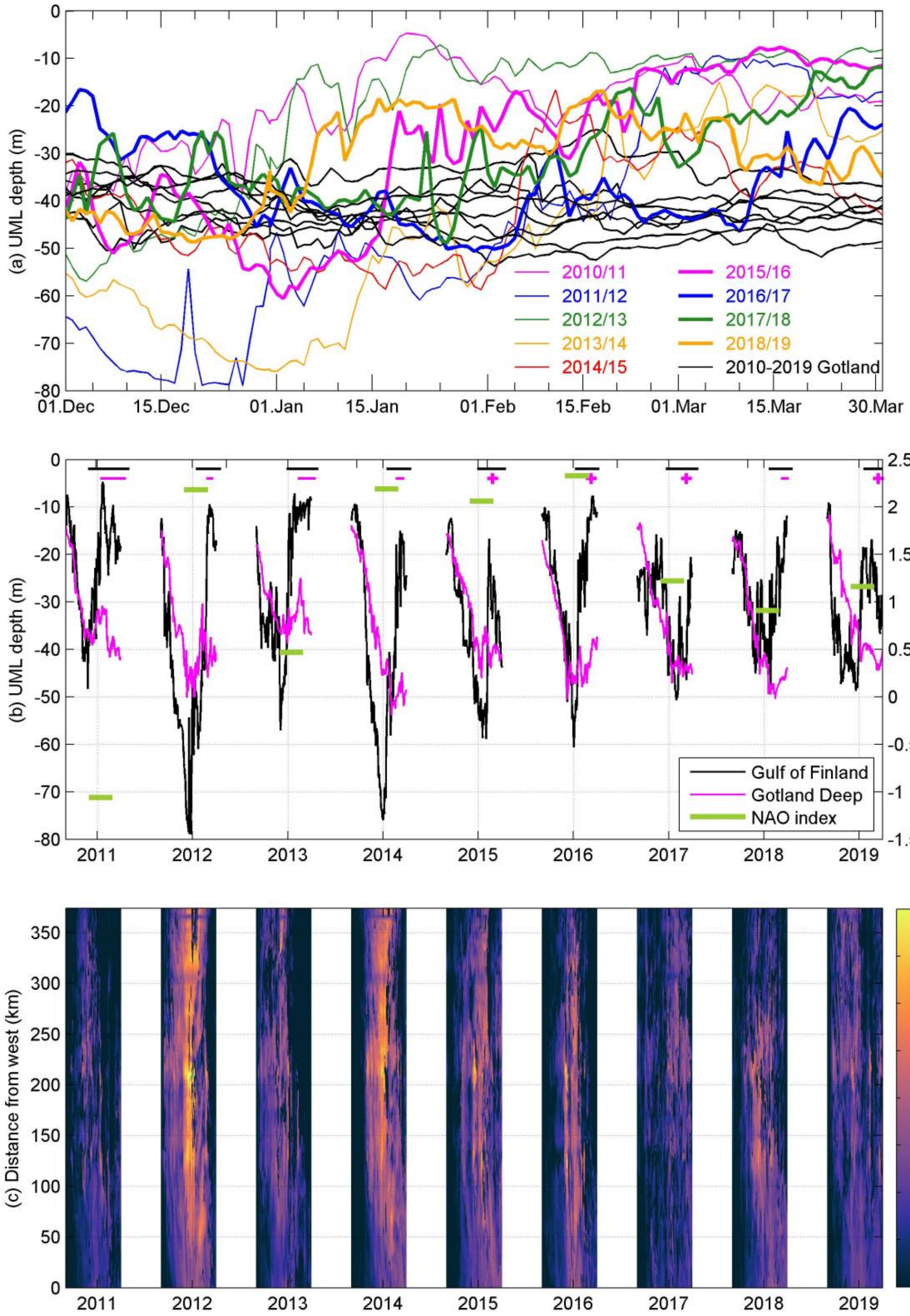


Fig. 10. (a-b) Time series of mean upper mixed layer depth in the Gulf of Finland and Gotland Deep based on model simulation data. The areal mean in (a) is calculated for the transect in the Gulf of Finland (Thalweg GoF, Fig. 1) and the box in the Gotland Deep (Fig. 1). In (b), green horizontal lines mark the mean December–February NAO index (from Jones et al., 1997); black and pink horizontal lines mark where the OSTIA SST is below T_{md} and pink crosses indicate where the minimum winter temperature did not fall

below T_{md} in the Gotland Deep. (c) Depth of the upper mixed layer along the transect from the northern Baltic Proper to the central Gulf of Finland (Thalweg in Fig. 1) for October to March, 2011 to 2019.

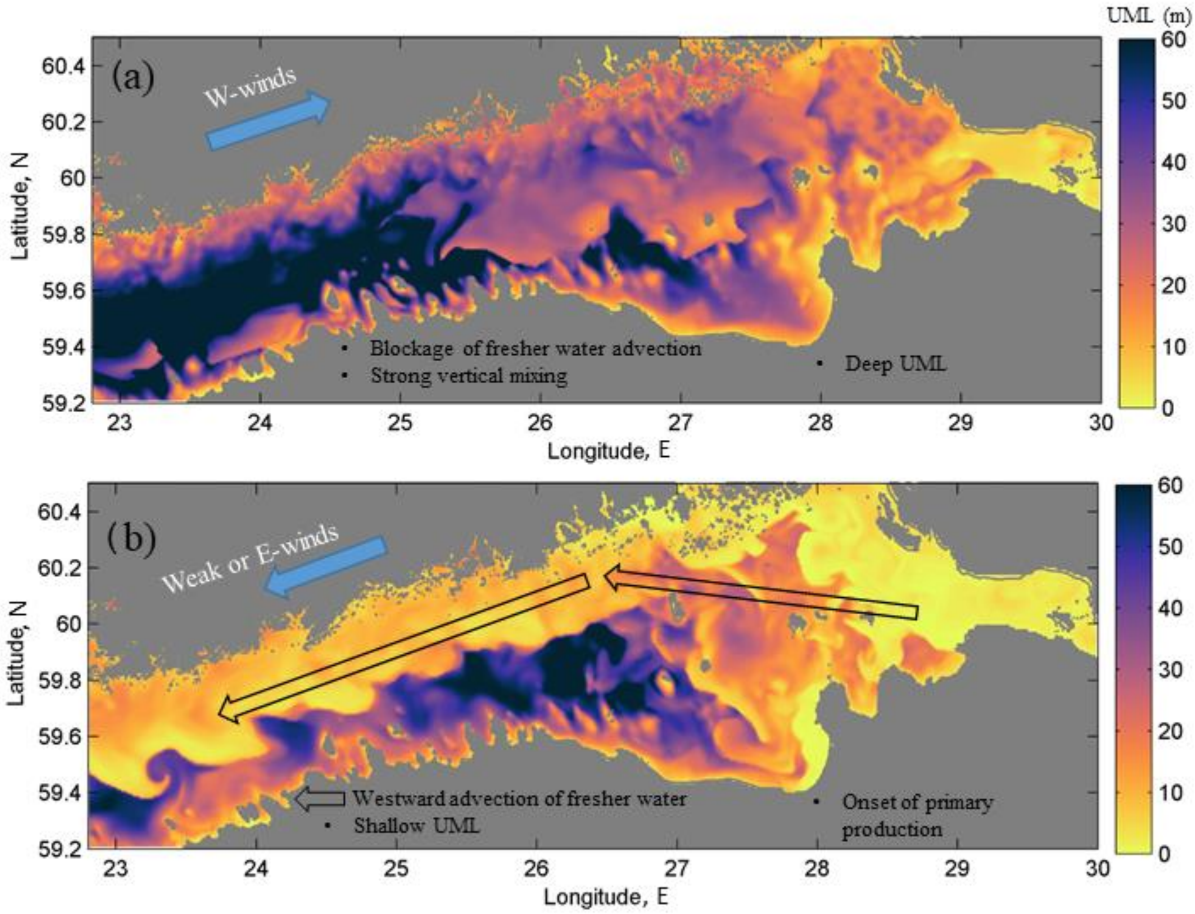
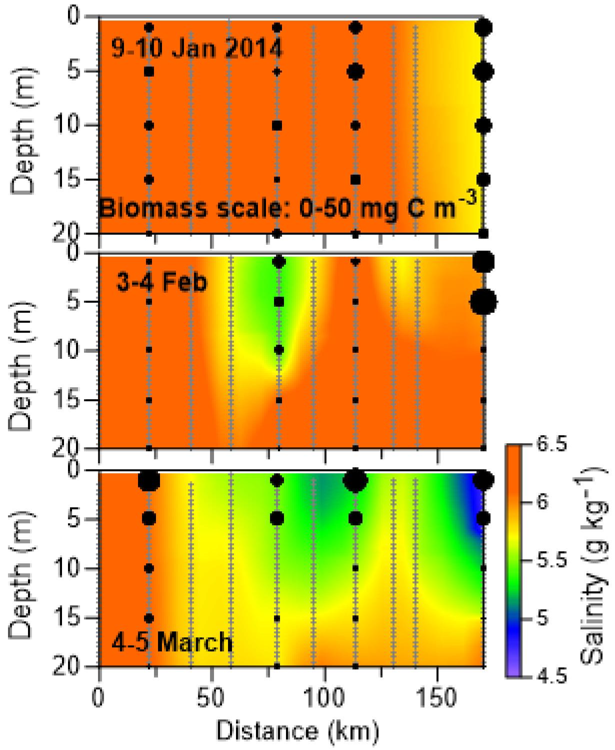


Fig. 11. Upper mixed layer depth (m) based on model simulation data in the Gulf of Finland on (a) 29 December 2011 and (b) 8 February 2012. Strong westerly winds dominated before the 29 December survey while variable and weaker wind occurred before the 8 February survey.



675 **Fig. 12. Vertical distributions of salinity (color scale) and phytoplankton biomass (black dots) along the Gulf of Finland in winter 2014. Phytoplankton biomass (mg C m^{-3}) scale is shown in the upper panel. Biomass sampling locations are shown in Fig. 1 (blue crosses). Vertical gray lines mark the location of CTD-casts.**