The role of nearshore slope on cross-shore surface transport during a coastal upwelling event in Gulf of Finland, Baltic Sea

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1. Introduction

Wind-driven coastal upwellings serve as a core mechanism responsible for the vertical exchange of water masses and different substances in the World Ocean. Recent studies have also indicated that such phenomena (e.g., upwellings in the Southern Ocean) may even influence our climate and ecosystem globally (Anderson et al., 2009; Morrison et al., 2015). Among several interesting aspects of the upwelling phenomena, this study focuses on the role of the nearshore slope on the presence of distinct surface features during an upwelling event in the Gulf of Finland, Baltic Sea (Fig. 1). To isolate and quantify these features, it is important to recapture that an upwelling is generally represented by two stages: an active phase and a relax phase (Gurova et al., 2013). The active phase develops when persistent winds stimulate offshore Ekman transport of surface waters. This subsequently generates an upward movement of denser and usually cooler water from deeper layers and an associated tilt of isopycnals. If the wind lasts long enough, cooler water surfaces and forms an elongated patch in the nearshore or an alongshore jet (Fig. 2a) (Bakun, 1990).

If the wind weakens or rotates so that offshore Ekman transport diminishes, the upwelling enters into so-called relax phase. The strong temperature and/or density gradients on the sea surface produced in the first phase, persist for some time (Zhurbas et al., 2008; Gurova et al., 2013). The nearshore patch of upwelled water (or the associated alongshore jet) eventually becomes unstable (Zhurbas et al., 2008) and usually develops various (sub)mesoscale phenomena such as jets, filaments, eddies, etc. (named differently by different scientists). Evidence of an upwelling event is often captured by satellite measurements. This decrease in sea surface temperature (SST) tends to have a strong signature compared to that of surrounding waters. As satellite images are usually contaminated by cloud cover, the appearance and evolution of the different phases may not always be captured.

The Gulf has two specific features that may impact the course and appearance of upwellings. Firstly, it is a shallow sea area with a mean depth of about 37 m. Secondly, its temperature regime has a strong seasonal variation. This area regularly hosts clearly detectable upwellings during the summer and autumn months when the water masses are relatively strongly stratified (Zhurbas et al., 2008; Gurova et al., 2013). Many studies have observed the dynamics of the different phases of upwellings in the gulf (Zhurbas et al., 2008; Laanemets et al., 2011). This process is crucial for the gulf ecosystem. During strong thermal stratification the surface layer is depleted of nutrients. On the one hand, upwellings supply the euphotic zone with nutrients sourced from deeper layers. Whilst, on the other hand, this process also supports the

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onset of cyanobacterial blooms (Laanemets et al., 2004).

Owing to the importance of this process for the Gulf of Finland, numerous studies have examined the upwelling phenomenon in this region using various methods (models, in situ, satellite) (Myrberg and Andrejev, 2003; Suursaar and Aps, 2007; amongst others). Due to the prevailing south-westerly and westerly winds the northern nearshore of the gulf hosts upwellings during some 30% of the time whereas the southern coasts often (during about 25% of the time) has downwellings (Myrberg and Andrejev, 2003).

Even though upwellings in the southern nearshore are less frequent and not as persistent as near the northern coast, strong upwellings also occur near the southern coast of the gulf. The changing wind patterns (Soomere et al., 2015) apparently have increased their frequency in the recent past. This change may substantially modify the functioning of the ecosystem of the gulf. In particular, it may cause a major shift of the feeding area of fish and associated wildlife.

This study focuses on an intense upwelling event near the south-western coast of the gulf that was triggered by a relatively infrequent system of persistent easterly winds. This event has an unprecedented coverage by in situ drifters that followed surface currents in the area directly affected by the upwelling (Delpeche-Ellmann et al., 2017). As it developed during a time interval with a reasonable number of at least 75% cloud free days, this event was also captured in a sequence of high-resolution satellite images.

This unusually wide coverage made it possible to establish several specific features and interesting developments of the upwelling events in the region. The analysis of satellite-derived SST and in situ drifters’ motions near the entrance of the gulf (Fig. 1) has demonstrated that on several occasions the upwelling process may contain an additional well-defined mid-phase as depicted in Fig. 4 of Delpeche-Ellmann et al. (2017). This phase becomes evident between the classic active phase (when cooler water upwells near the coast and optionally forms an alongshore jet at the surface) and the relaxation phase (when the jets start to disintegrate). This phase is characterised by the presence of jets of cooler water that gradually migrate offshore. The intensity of lateral mixing is low during this phase. After some time, in the relaxation phase, after the winds have decreased, these cross-shore jets lose their identity and develop into several filaments and mesoscale features. The most intense mixing occurs in this phase (Fig. 3b) (Zhurbas et al., 2008; Delpeche-Ellmann et al., 2017).

The focus of this study is on the characteristic feature of upwellings in this region. Namely, during the mid-phase the cross-shore jets tend to migrate offshore from distinct locations (Fig. 2b). This feature has been identified by several studies of upwelling events in this area (Suursaar and Aps, 2007; Laanemets et al., 2011; Rikas and Lips, 2016). It has an interesting (but not necessarily dynamically related) analogue in model simulations of Lagrangian transport of surface waters in the gulf. Namely, such simulations suggest that an intense net-transport of surface waters in the north-south direction systematically occurs in the vicinity of the two distinct locations where such cross-shore jets have been observed (Fig. 3a, c). This (albeit qualitative) match of the outcome of the two independent studies signals that two locations near the southern coast of the gulf may host certain permanent (e.g., geographic) features that enhance intense surface cross-shore transport.

The idea that bathymetry may have a strong influence on the appearance of the cross-shore jets is not new and has been repeatedly mentioned in previous studies (Zhurbas et al., 2004; Laanemets et al., 2009). However, this connection has not been quantified and the actual verification of an interrelation of the bottom slopes and the behaviour of cross-shore jets has remained unexplored. One of the main objectives of this study is to quantify, at least to a first approximation, the role of the sea bed slopes at certain locations on the development of cross-shore jets by utilising simple statistical analysis of satellite SST and in situ drifter data.

1.1. Basic features of upwellings

The evolution and transport of water masses and different substances in coastal upwellings are generated and influenced by a multitude of drivers. These drivers can be grouped into two categories. Firstly, permanent (temporally and spatially unchangeable) conditions such as geographic features (the shape of the nearshore, bathymetry, coastline undulations, etc.) that can potentially steer upwelling events.

Fig. 1. Location scheme of the Baltic Sea and the study area (red box) near the entrance of the Gulf of Finland. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)
Secondly, a particular upwelling is shaped by several features that vary greatly on weekly to annual scales such as properties of winds and characteristics of water masses such as stratification, the associated internal Rossby radius, the local water level and interactions with other oceanic processes (waves, eddies etc.) (Preller and O'Brien, 1980; Narimousa and Maxworthy, 1985, 1987a; 1987b; Lentz and Chapman, 2004; Chen et al., 2013).

Extensive variability of some of these drivers affects the appearance of the upwelling and its transport properties both spatially and temporally (Miranda et al., 2013; Tim et al., 2015; Wang et al., 2015). Upwellings generally tend to appear in similar locations. Due to the short-term changes in wind and ocean properties, the appearance and intensity of single upwellings usually differ from each other (Suursaar and Aps, 2007). Globally and locally changing wind patterns evidently enhance such differences (Bakun, 1990; Soomere et al., 2015). In this light it is increasingly important to understand which properties of upwellings are invariant with respect to the listed changes. Solving this task obviously contributes to further clarification of the functioning of the ecosystem of the gulf. Ecosystem studies usually utilise a vast amount of resources from massive in situ data collection up to extensive computing facilities. As the upwelling events have usually intermittent nature, it is usually not possible to timely concentrate the necessary resources and simpler solutions may at times be required to study their properties. A feasible way is to exploit the satellite data in combination with limited in situ data.

Many studies have examined the role of coastline features and local bathymetry in the dynamics of upwellings since the 1980s (Preller and O'Brien, 1980; Narimousa and Maxworthy, 1985, 1987a; among others). It is likely that some concealed features of bathymetry such as the nearshore slope may steer, stabilise and even trap upwelling events (Kämpf, 2012). While coastline features seem to have very little effect on the upwelling dynamics, bottom topography tends to steer the upwelling maximum (Kämpf, 2012).

The upwelling dynamics is often examined using a steady two-dimensional theory (Lentz and Chapman, 2004; Choboter et al., 2011, among others). Its outcome was supported by observations of depth averaged velocity profiles and by numerical simulations with primitive equations. It is convenient to examine the cross-shelf momentum flux...
divergence relative to the wind stress using a dimensionless parameter – the Burger number. It indicates where the upwelled water may have originated from in the deeper layers (e.g., Fig. 2, $V_i$ or $V_b$). The knowledge of interrelations between $V_i$ (the onshore return flow) and $V_b$ (the flow in the bottom boundary layer) makes it possible to a certain extent conclude whether nutrients, carbon, sediment parcels, or oxygen may be present in the upwelled waters.

Most of the relevant studies have focused on the onshore–cross-shore transport dynamics related to the deeper layers. Their main question has been whether the source of the cooler water is the intermediate layer ($V_i$) or the bottom layer ($V_b$).

A different viewpoint was adopted by Narimousa and Maxworthy (1985). They addressed the parameters that could be used to calculate the cross-shore extent of the upwelled water on sea surface (Figs. 2b and 3a). They utilized the geostrophic balance between the Coriolis force and the horizontal pressure gradient forces. In this paper, we attempt to utilize the theory developed by Lentz and Chapman (2004) to determine where within the deeper layers the cooler surface water originated from. Further on, we apply the approach of Narimousa and Maxworthy (1985) to determine the extension of the upwelled water patch.

Even though the presence of the cooler upwelled water is evident in satellite SST data, its dynamics is largely unknown. Our intention is to combine SST data and information about bathymetry with in situ data about currents to: (1) identify a possible threshold for bottom slopes that may serve as a trigger for the cross-shore jets, (2) utilize the Burger number (Lentz and Chapman, 2004) to identify the depth (or layer) where the cooler surface water may have originated from, (3) exploit the concepts developed by Narimousa and Maxworthy (1985) to quantify the cross-shore extension of upwelling jets and (4) finally, combine an estimate of this extension and in situ data about stratification to approximately solve a sort of inverse problem – to derive an estimate of unknown parameters such as the thickness of the surface layer. These results would assist in identifying the transport of nutrients in the surface layer and their link to cyanobacterial blooms in the study area. As we mostly rely on nondimensional parameters, the outcome outlines a simple method for identifying some of the characteristics of the upwelling process, especially when limited resources are available.

2. Study area, methods and data

2.1. Study area

The Gulf of Finland is an elongated bay at the north-eastern end of the Baltic Sea. It stretches from the main basin of the Baltic Sea between Finland and Estonia to the major city of Saint Petersburg in Russia. The gulf is about 400 km long and its width varies between 48 and 125 km (Fig. 1). The impact of saltier water from the Baltic Proper combined with a large amount of fresh water input from the eastern gulf results in a strong gradient and extensive spatio-temporal variability in salinity and temperature both in vertical and horizontal directions (Alenius et al., 1998). The brackish environment (surface salinity 6–6.5) in the western part of the gulf changes to an almost freshwater one in its easternmost section. The western gulf has an almost permanent halocline (bottom salinity of 7–10) whilst the eastern part is often fully mixed.

The overall motion of the water masses in the gulf is traditionally viewed as mostly cyclonic (Alenius et al., 1998; Leppäranta and Myrberg, 2009). The motion only represents the long term average and not necessarily becomes evident in short-term measurements or modelling efforts (Myrberg and Soomere, 2013).
The surface layer of seas and oceans exhibits Ekman drift for moderate (6–10 m/s) and strong (> 10 m/s) winds. The observed speed of wind-driven surface currents is usually 2–3% of the wind speed in the study area. The observed direction of such currents is 20°–30° to the right of the wind direction (Alenius et al., 1998; Leppäranta and Myrberg, 2009). Importantly, the circulation pattern varies remarkably in different layers of this water body (Andrejev et al., 2004a, 2004b) apparently because of strong stratification of its water masses. The motions in the uppermost layer (with a thickness of a few metres) are at times (during stronger wind events) largely disconnected from the motions in the deeper layers (Soomere et al., 2011). During lower wind speeds the uppermost layer may be entrained into the motions in the deeper layers and often moves almost independently of the wind properties (Gästgivars et al., 2006; Delpeche-Ellmann et al., 2016).

Wind-driven upwellings may easily modify this subtle balance, especially during the above-described mid-phase when wind weakens but still persists for some time. This phase is characterised by the formation of clearly defined offshore directed cooler surface water jets and associated cross-shore surface transport. The intensity of mixing is limited and the transport of water offshore predominates (Figs. 1b and 3a).

Importantly, these surface cross-shore jets may override the classic Ekman transport and modify the properties of Lagrangian transport of the surrounding surface waters (Delpeche-Ellmann et al., 2017). This feature apparently reflects the presence of a very thin uppermost layer and overall strong stratification. This combination may weaken the entrainment of deeper layers into the Ekman transport and may lead to a specific interaction of a thin layer of cooler upwelled water with the surrounding surface water.

The instantaneous patterns of motions in the surface layer of the gulf thus strongly depend on the wind speed, direction and persistence. The most frequent wind directions in the north-western Baltic Proper are south-west and north-north-west (NNW) (Soomere and Keervalliik, 2003). The wind system in the gulf itself contains occasional relatively strong eastern and north winds. This pattern is not static: there is almost no evidence about strong NNW winds in the older data (Soomere and Keervalliik, 2001) whereas the direction of the geostrophic air flow (represented by a vector consisting of the average zonal and meridional wind components) over the southern Baltic Sea has rotated by ∼40° at the end of the 1980s. This rotation is most pronounced in winter and spring months and most probably represents a signal of climate change (Soomere et al., 2015).

As mentioned earlier the predominant winds from the south-west generate frequent upwellings near the northern (archipelago) coast of the gulf (Leppäranta and Myrberg, 2009). Eastern winds are less frequent but still often cause major upwellings along the southern (more regularly shaped) coast of the gulf. As the uppermost mixed layer is very thin (just a few metres) in spring and early summer, even relatively weak and short-term alongshore winds may cause drastic changes in the SST during upwelling events. This feature makes it possible to track the onset and development of upwellings in this area in relatively great detail and during long time intervals.

The strongest upwelling phenomena near the southern coast of the gulf are usually observed in the middle part and near the entrance of the gulf next to the north-western (NW) shores of Estonia (Fig. 1) and have an alongshore extension of 30–150 km. The width of the nearshore upwelled cooler water patch is often related to the internal Rossby radius. This quantity is usually in the range of 2–10 km in the Baltic Sea (Fennel et al., 1991). It varies over seasons in the gulf and often is as small as 1.3 km (Alenius et al., 2003). Thus the expected cross-shore width of such patches is 5–20 km. Filaments of upwelled water at times may spread much farther. They often reach several tens of kilometres out into the sea and in some occasions cross the entire gulf (Zhubas et al., 2008).

The cooler surface cross-shore jets often develop at two distinct locations of the southern coast of the gulf. These locations roughly match the areas where intense cross-shore Lagrangian transport is likely (Fig. 3). Delpeche-Ellmann et al. (2017) argue that these offshore-directed jets are relatively stable during the mid-phase of upwellings. The resulting surface currents may provide, on average, intense Lagrangian transport on time scales of a few days to a week across the gulf (Soomere et al., 2011). Similarly, they may contribute to the drift of water (or pollution) parcels from offshore areas to the nearshore (Andrejev et al., 2011).

An explanation for the preferred location of such cross-shore jets may be the presence of certain features of the seabed. The deepest areas of the gulf (80–100 m) are located in the western and southern parts of the gulf within or near the study area (Fig. 4). The central part of the gulf is over 60 m deep. The seabed has greatly different appearance in the northern (Finnish) and southern (Estonian) side. The northern nearshore is shallower (depths typically 20–40 m). It has an extremely irregular coastline, rugged bathymetry and contains extensive archipelago areas.

The southern coast seems irregular on large-scale maps and contains a few major islands and several bays deeply cut into the mainland. It is, however, fairly regular on scales of a few kilometres. The overall shape of the seabed is gently sloping towards the deepest part of the gulf (that is located at a distance of about 20 km from the shoreline). Typically, the average slope of the nearshore seabed is on the order of 0.005 but some sections are fairly steep, with a slope up to 0.025. This variability in the seabed slope may be a potential trigger for the onset of cross-shore cooler surface water jets during upwellings. In this paper we test the hypothesis that the shape of the seabed of the study area may largely govern the locations from where such offshore-directed jets start. The areas often covered by such jets naturally serve as regions in which intense cross-gulf transport of various substances is likely.

2.2. Theoretical width of the upwelling front

To quantify the properties of the cross-shore jets we apply a simple model developed by Narimousa and Maxworthy (1985) based on laboratory experiments with stratified two-layer fluid in a rotating cylindrical tank and sloping bottom. Their approach makes it possible to estimate the cross-shore width (λs) of the upwelled flow via the balance of the impact of the Coriolis force on the surface layer (that influences the migration of the front away from the coast) and the horizontal pressure across the front due to the distortion of the interface (that tends to halt the front). Thus the greater the horizontal pressure gradient, the less are the chances for the front to migrate seawards. Also, the greater the Coriolis force, the more likely the front starts to move seawards.

The balance of these two constituents implies the following relationship for the seaward front of the upwelled cooler waters (Narimousa and Maxworthy, 1985, Sec. 5):

$$-frAΩ = f^2 \frac{dη_s}{dr}$$

(1)

Here rAΩ is the plate velocity, f = 2Ω sin ϕ is the Coriolis parameter, ϕ represents latitude and $Ω = 0.7292 \times 10^{-4} \text{s}^{-1}$ represents the rotation rate of the Earth, $g' = g Δ \rho / \rho_0$ is the reduced gravity between two layers, $Δ \rho$ is the density difference between the two layers, $\rho_0$ is the average density of the fluid, $g = 9.81 \text{m s}^{-2}$ is acceleration of gravity, and $η_s$ is the depth of the front from the surface. The general solution of Eq. (1) is:

$$-fAΔΩr^2 = 2g' \left( η_s + c_1 \right)$$

(2)

where $c_1$ is the integration constant. If R is the distance of the front (at the sea surface) from the centre of the tank, then the thickness of the patch of the cooler upwelled water $η_s = 0$ at the distance $r = R$ from the shoreline. The solution (2) with this boundary condition becomes:
where $s$ indicates strong stratification, $h_0$ is the bottom slope, and thus the Burger number and $\mu$ is the tank radius. Solving (4) for $R$ gives:

$$R = \left( \frac{4g h_0 R_0^2}{\mu \Delta \Omega} \right)^{1/2}. \tag{5}$$

Since the width of the upwelled flow at the surface is $\lambda_s = R_0 - R$, we obtain from (5) the following expression for $\lambda_s$:

$$\lambda_s = 1 - \left( \frac{4g h_0}{\mu \Delta \Omega} \right)^{1/4} \quad \text{or} \quad \lambda_s = R_0 \left[ 1 - \left( \frac{4g h_0}{\mu \Delta \Omega R_0^2} \right)^{1/4} \right]. \tag{6}$$

We focus on two parameters that govern the width of the upwelling front (Eq. (6)) and that can be evaluated from the existing information. These are: i) an estimate of the spatial extent of the upwelled patch derived from satellite SST data and ii) in situ data of the stratification from which we can calculate $g'$. Eq. (6) is used below to assess the thickness of the top layer and the associated depth from which the water is brought to the sea surface from other measurable quantities for the Gulf of Finland conditions.

### 2.3. The Burger number in the upwelling theory: Bathymetry and stratification

Lentz and Chapman (2004) demonstrated how different stratifications and properties of bathymetry can influence the vertical structure of the wind-driven cross-shelf flow even in the absence of alongshore variations. The cross-shelf momentum flux divergence relative to the wind stress depends on the Burger number

$$S = \frac{a N f}{\mu}, \tag{7}$$

where $a$ is the bottom slope, $N = \sqrt{\frac{g \beta T}{\rho}}$ is the buoyancy frequency, $\rho$ represents the average density and $f = 0.000125 \text{ s}^{-1}$ is the Coriolis parameter for the latitude of the Gulf of Finland. The Burger number is often used to express the relationship between stratification, Coriolis force and properties of bathymetry. It indicates from which depth the upwelled water may have originated from. The values $S < 1$ characterise relatively weak stratification and gentle slopes indicating that the cross-shelf momentum flux divergence is small, the bottom stress balances the wind stress and the onshore return flow is mainly in the bottom boundary layer. The values $S > 1$ indicate strong stratification and/or steep bottom slopes. In such occasions the self-sustaining momentum flux divergence balances the wind stress, the bottom stress is small and the onshore return flow is in the interior of the water column.

Charney (1955) and Allen (1980) indicated that the cross-shelf scale for the region of sloping isopycnals during upwelling should be related to the baroclinic Rossby radius $R_o = Nh/\alpha$, where $h$ is the local water depth. It is easy to see that $R_o = Sh/\alpha$ and thus the Burger number and Rossby radius are proportional to each other in this framework. Therefore, the isopycnal slopes should be proportional to the ratio of the water depth $h$ and the Rossby radius (Eq. (9) in Lentz and Chapman, 2004)

$$\frac{\partial \rho/\partial y}{\partial \rho/\partial z} \approx \frac{h}{R_o} \approx \pm \frac{\alpha h}{Nh/\alpha} = \pm \frac{f}{N}, \tag{8}$$

where $\alpha$ is a proportionality constant. We now attempt to utilise the Burger number (Eq. (7)) for a deeper insight into the dynamics of the upwelling event in the gulf.

### 2.4. Bathymetry

The definition of the Burger number (Eq. (7)) displays that the slope of the nearshore seabed is one of the core parameters that govern the stability of the upwelled water jet. The concept of the Burger number ignores small-scale inhomogeneity of the seabed. As the baroclinic Rossby radius is usually well above 1 km in the study area (Alenius et al., 2003), we employ parameters of the seabed that are averaged over about 1 km.

Bathymetric data for the study area was provided by the Estonian Maritime Administration. This data with a spatial resolution of 40 m was collected to the requirements of the International Hydrographic Surveying standards (S44). The data were gridded using a linear interpolation to a resolution of 1 km along latitudes and 1.7 km along longitudes. From this data set we constructed 80 cross-sections of the nearshore separated by 1.2 km from each other and oriented roughly perpendicularly to the isobaths and to the coastline (Fig. 5).

The water depth along the cross-sections exhibits partially uneven bottom features. Their general appearance, however, is fairly regular in the water depth from about 20 m to about 90–110 m. An approximation to the average bottom slope was calculated using the maximum and minimum depths and the relevant locations for the gradually deepening parts of the profiles. Small-scale features (such as a local elevation at a depth of about 50 m in the profile in Fig. 6) were ignored. For profiles that contained islands or extensive nearshore shoals only the gradually deepening part was used.

### 2.5. Satellite derived and in-situ SST data

As mentioned earlier numerous studies have examined the upwelling phenomena in the Gulf of Finland (Zhurbas et al., 2008; Laanemets et al., 2009; Suursaar and Aps, 2007). In this study we focus on a remarkably well documented strong upwelling event with a strong signature of the cross-shore jet on 28 May – 5 June 2013 (Delpeche-Ellmann et al., 2017). For this event, additional to satellite SST in situ data about surface currents and stratification was available. The SST data are derived from the Earth Observation System (EOS) Moderate Resolution Imaging Spectrometer (MODIS) Aqua level 2 (thermal bands 31 (11 µ) and 32 (12 µ)) data. The relevant data set with a spatial resolution of 1 km was retrieved from the open access NASA OceanColor
Fig. 6. Examples of the shape of seabed in the study area. The dotted line provides an example of the slope calculation.

website (http://oceancolor.gsfc.nasa.gov/). An estimated accuracy/reliability of 0.4–0.5 °C of the SST data is expected (Brown and Minnett, 1999; Kozlov et al., 2014).

Our focus is on the relationship of the properties of intense cooler cross-shore jets and the local bathymetry. To shed light on this issue, a further examination was performed for the day that had the most intense cooler cross-shore jets. The SST data captured on 31st May 2013 (Fig. 3a) was linearly interpolated to cover the above-described 45 km long cross-sections (Section 2.4, Fig. 5). These interpolated data along each cross-section were used to evaluate a mean SST value for each profile. Such a procedure highlights the locations of the cross-shore cooler water jets. The mean SST for each cross-section was compared with its average slope.

3. Results

3.1. Stratification and chemical parameters

The upwelling in question was apparently triggered by persistent easterly winds on 21–24 May 2013 with an average speed of 7.6 m/s according to the data from the Kalbådagrund weather station. The SST decreased from 8 °C on 24 May to 4 °C on 28 May 2013. This event contained a clearly identifiable mid-phase (about 28 May–2 June) between the active and relax phases. During this phase the wind speed was still high, the upwelled water travelled almost straight across the gulf to a distance of 40–45 km from the coast (Fig. 3a and b). Two persistent and distinct areas of the formation of such jets were (i) the NW coast of Estonia near Dirhami and (ii) the vicinity of Tallinn (Fig. 3a and b). These areas were separated by more than 100 km with no cross-shore jets (Fig. 3a and b).

The data sampled at two stations of the Estonian Marine Institute provide information about temperature, salinity, density (so-called CTD data), and chlorophyll fluorescence and dissolved oxygen. The measurements were performed in 31 May 2013, that is, in the middle of the upwelling mid-phase. Sampling station S1 (59.483°N, 22.95°E) was located outside the cooler water jet and reflected the properties of surrounding waters. Another station S2 (59.325°N, 23.267°E) was located about 20 km from S1 in the cooler water jet near the western border of the impacted area (Fig. 3a). Both stations were located at a depth of 85–95 m in the thalweg area of the gulf. As the properties of water masses and particularly the main features of stratification along the thalweg are usually fairly homogeneous over many tens of kilometres, it is likely that station S1 to a first approximation represents the water properties that existed in station S2 prior to the upwelling event.

The density profiles (Fig. 7c) reflect the typical strong stratification in the deeper parts of the western gulf (Alenius et al., 1998; Leppäranta and Myrberg, 2009). This structure of water masses is characteristic during strong intrusion of saltier waters of the northern Gotland Basin into the gulf (Elken et al., 2003) during easterly winds. The water density has a three-layer appearance with jump layers at 20 m and 60–80 m. The three-layer structure is even more clearly present in profiles of salinity and temperature. A particularly strong thermocline is located at a depth of 20 m and a characteristic inversion at depths exceeding 60 m (Fig. 7a and b).

The upwelling evidently modifies the properties of the entire water column in station S2. The largest changes occurred for the temperature, chlorophyll and dissolved oxygen of the uppermost 20 m thick layer (Fig. 7). The temperature profile in S2 shows almost homogenous surface water of 6 °C with a thickness of 5 m whereas the undisturbed surrounding waters had a 20 m thick uppermost layer with an almost constant temperature > 8 °C. Beneath this depth the water column in S1 has about 15 m thick layer with a moderate (for the gulf conditions) temperature gradient of the order of 0.2 °C⋅m⁻¹. The lower border of the thermocline (where the temperature drops from 6 °C to 2 °C) lies at a depth of about 30 m.

The SST in coastal stations (Delpeche-Ellmann et al., 2017) first dropped down to 4 °C but by the 31st May it increased to 6 °C. Thus it is likely that the cooler surface water of ~4 °C may have originated between the 15–30 m depth. This conjecture is supported by the salinity data. The salinity in the surface layer only shows a slight variation between the two stations. It is thus likely that the upwelled water did not originate from saltier waters below the halocline (and almost certainly not from the bottom layers).

The chlorophyll fluorescence data represents light emitted and absorbed by chlorophyll molecules. Whilst it does not directly represent the concentration of chlorophyll, it gives a reasonable indicator of its relative amount whereas higher fluorescence readings represent higher chlorophyll content. Fig. 7d shows that in the uppermost 0–20 m the waters in S2 had lower values of chlorophyll than those in S1. The greatest difference between the two recordings was in the uppermost 10 m thick layer. All these observations together indicate that the cooler upwelled water was likely brought up to the surface layer from the 15–30 m depth, that is, from within the vicinity of the pycnocline where the water was only slightly saltier than in the surface layer.

3.2. Seabed slope and average temperature

To ascertain the role of bathymetry slopes in creating and/or enhancing the surface cross-shore jet we examined the relationship of the seabed slope and average SST along 80 cross-shore profiles. Their orientation roughly matched the normal to the overall shoreline whereas the single cross-sections were slightly tilted from the north–south direction (Section 2.4). As mentioned above, the overall shape of the bathymetry of the study area, even it is to some extent rough on larger scales (> 5 km), is still much more regular than in the northern part of the gulf. Comparatively steep slopes are mostly located in the vicinity of islands but also along certain sections of relatively straight segments of the mainland coast.

All profiles have slopes larger than 0.0025 (a slope of 0.005 means that the seabed deepens, on average, 5 mm for every 1 m along the profile). More than half of profiles have the average slope between 0.0025 and 0.005. The typical range (~80% of all slopes) is from 0.0025 to 0.006 (Fig. 8). The frequency of occurrence of different slopes mostly (except those < 0.0035) follow a normal distribution (Fig. 9). The steeper (> 0.0075) slopes contains several outliers (cross-sections with exceedingly steep slopes up to 0.0225, Fig. 9). Such outliers are localized in distinct areas (Fig. 10). Importantly, their locations match the areas from where the cooler cross-shore jets of surface water stemmed (Fig. 11).
To identify a possible relationship between the slopes of seabed and variations in SST, the average temperature for each cross-section was calculated as described in Section 2.5. Interestingly, some cross-sections with very large slopes were located in the area with low average temperatures during the upwelling event. A scatter plot of the two quantities (Fig. 12) reveals that the lowest average temperatures occurred exclusively along cross-sections with large average slopes. Another view to the relationship in question offers a quantile-quantile plot of the average temperature and slope (Fig. 13). The sub-population of slopes > 0.0075 and average SST < 7 °C have a clearly different distribution than the rest of the data set. Importantly, all profiles with the average slope > 0.0075 and average SST < 7 °C belong to the same area from where the cross-shore jets origin.

Even though the provided evidence is heavily generalised (as we only consider the average SST over a 40 km wide stretch), it is remarkable that all signatures of cross-shore cooler water jets are co-located with nearshore areas with steep slopes. It is likely that, under the existing stratification and driving forces of the upwelling, the minimum slope that may have triggered the cross-shore surface jet is about 0.0075. This co-location is apparently a general feature for many upwellings along the study area. In particular, it provides a rough estimate of the geographical location of where the cooler water jet may be brought up from (Fig. 14).

3.3. The extension of the cooler water

To ascertain the cross-gulf extension of the upwelling event, we utilise Eq. (6). The values of the main parameters involved into Eq. (6) can be evaluated using a combination of satellite and in-situ data for 31 May 2013. The quantity $\rho_1 - \rho_1$ expresses the difference in water
density across the main pycnocline (between the two upper layers on density profile in Fig. 7). If the upper layer with a density \( \rho_1 \) is interpreted to be from the 0–20 m, the lower layer with \( \rho_2 \) from 20 to 50 m and \( \rho_0 \) as the average density for depths of 0–90 m, an estimate for the reduced gravity is \( g' = g(\rho_2 - \rho_1)/\rho_0 = 0.0088 \text{ m s}^{-2} \). The Coriolis parameter is \( f = \Omega \sin \phi = 0.000125 \text{ s}^{-1} \), where \( \phi = 59.4^\circ \text{N} \) is the approximate latitude of the study area. The thickness of the surface layer \( h_0 = 1 – 60 \text{ m} \) is considered as a variable quantity.

In laboratory experiments of Narimousa and Maxworthy (1985) the upper layer was spun up to \( r\Delta \Omega \) (plate velocity). It is natural to interpret this quantity as the horizontal speed of the upper layer in the study area. This speed was estimated based on the drifter’s relocation for 31 May 2013. Differently from some other days, the surface current (about 0.1 m s\(^{-1}\)) was in the range of the usual value of 2–3\% of the wind speed (Delpeche-Ellmann et al., 2016). Similarly to the thickness of the surface layer, we allowed \( r\Delta \Omega \) to vary within the range of \( r\Delta \Omega = 0.1 – 0.5 \text{ m s}^{-1} \).

Another parameter that has to be estimated from the geometry of the study area is \( R_0 \) (the radius of the tank in experiments). As the effective width of the gulf varies between 48 and 125 km, we have simulated the extension of the upwelling filaments for the values of \( R_0 \) in the range from 40 to 100 km to represent the half-width of the gulf.

The estimates of the reach of the cross-shore jet based on the available data, described assumptions and arguments of Narimousa and Maxworthy (1985) for the surface width of the patch of the nearshore cooler water are presented in Fig. 15 as a function of the thickness of the surface layer. This representation is useful for rough estimates of the parameters of the dynamical regime and properties of water masses that produced the observed cross-shore jets.

Not surprisingly, from Fig. 15 it follows that the surface layer thickness before an upwelling event strongly influences the reach of the upwelled cooler water and the location of its seaward front. It is heuristically clear that the greater the surface layer thickness, the narrower is this band. The SST data indicates that the cross-shore jet during the upwelling in 2013 reached a distance of up to 40–45 km from the coast. This range corresponds to the surface layer thickness of...
5–22 m, depending on the speed of the surface current (Fig. 15). Since the relocation speed of water in the jets was about 0.1 m s\(^{-1}\) (Delpeche-Ellmann et al., 2016), the reach of a cross-shore jet to a distance of 40–45 km would require an upper layer thickness between 3 and 5 m. The theory thus seems to underestimate the upper layer compared to the density profiles in stations S1 and S2 (Fig. 7).

Narimousa and Maxworthy (1985) showed that only when the maximum lowering of the interface (caused by the Ekman transport) is greater than twice the surface layer thickness will the front migrate seawards. In the opposite case, although the cooler water appears near the wall at smaller depths, it does not necessarily arrive to the surface. Thus the thinner the surface layer, the more likely the front starts to migrate seawards.

The reach of the upwelling front from the shoreline is mostly governed by the joint impact of the surface layer thickness (that influences the horizontal pressure gradient force) and varying velocity (the influence the Coriolis force). The above-described pattern of upwelled water in the study area may thus stem from a thinner surface layer in the areas of steep slopes and/or from a greater horizontal velocity in these areas compared to that occurring over the gentler slopes. It is also likely that in the 100 km stretch where the cross-shore jet did not exist (Fig. 3) the surface layer was deeper.

The varying bottom slopes may add several nuances to this picture. They may lead to different temporal scales for the upwelled waters to climb up the slope. In theory, for steep slopes the time taken for the upwelled waters to surface would be less than for water with similar properties along gentler slopes (Chen et al., 2013). Thus the slope of the nearshore seabed combined with the changing wind speed may also influence whether the cross-shore jets are formed.

3.4. The Burger number

The Burger number (Eq. (7)) gives an indication of where within the deeper layers the cooler upwelled waters may have originated from. The typical buoyancy frequency in the study area was \( N = 0.025 \text{ s}^{-1} \) (approximated from the maximum value at the 20–30 m depth in stations S1 and S2, Fig. 7). For the average nearshore slopes, not surprisingly, relatively high Burger numbers (> 1.5) occur specifically in the areas where cooler surface cross-shore jets were observed (Fig. 16). These values indicate, similarly to the results described in Section 3.1, that most likely the upwelled cooler water was sourced not from the bottom waters but from an intermediate layer in the vicinity of the pycnocline.

Remarkably, in most locations the Burger number is < 1 (Fig. 16). In these areas it is likely that the upwelled water originated from the bottom layers. This does not necessarily mean larger physical depth of the nearshore; instead, it indicates that the upwelled water stems from nearshore areas where the depth is smaller than the depth of pycnocline, the water column (that is not very deep) is relatively homogeneous and the (average) stratification is weak. This sort of interdependence of the properties of the average stratification and local depth complicates estimates of the source of waters and the width of upwelled cooler patch on the sea surface. However, this complication does not alter the conjecture that the presence of cooler cross-shore surface water jets at the steeper slopes was more likely than along the gentler slopes.

4. Discussion

Previous studies in the Baltic Sea have suggested that a strong relationship exists between the appearance of cooler water jets during upwelling events and bottom inhomogeneity (Zhurbas et al., 2004; Laanemets et al., 2009; Gurova et al., 2013). One of the central topics
addressed in this paper was to demonstrate and quantify whether a robust connection exists between the shape of the seabed (in particular, its cross-shore slope) and the location of the cross-shore jets of cooler upwelled waters on the sea surface on the southern coast of the Gulf of Finland, Baltic Sea. The most interesting outcome of the co-analysis of satellite derived SST data and high resolution bathymetry data is that coastal segments with average slopes of > 0.0075 seem to be particularly prone to the development of cross-shore jets of cooler water. These areas were also characterised by relatively large Burger numbers (> 1.5) which hints that the source of the upwelled water most likely originated not from the deep bottom layer but from an intermediate layer. This conjecture extends the outcome of Gurova et al. (2013) who found, using satellite data and numerical models in the Baltic Sea, that a strong connection with coastal jets was controlled by vorticity dynamics related to depth variations in the direction of flow. Our results show a similar connection with the bathymetry whereas the governing parameter seems to be the average slope of the seabed.

The width of the nearshore area covered by cross-shore surface jets is estimated using the classic arguments by Narimousa and Maxworthy (1985) and satellite derived SST. In the gulf, for a cross-shore jet to reach the distance of 40–45 km, the surface layer thickness (or the depth from which the water is upwelled to the surface) should be in the range of 5–22 m for relatively large surface speeds but only 3–5 m for speeds on the order of 0.1 m s⁻¹. The density profiles at two sampling stations in the vicinity of the upwelling confirm that this is a realistic estimate. The local values of the Burger number and detailed temperature and salinity profile data in a jet and in an undisturbed area indicate that the cooler water of ~ 4 °C most probably originated from an interior depth of 15–30 m. Thus it is likely that in the locations of steep slopes, the upwelled water originates from certain mid-depths of the water column (Fig. 17).

These results match well those of previous studies that examine the source of nutrients in the gulf. For instance, both field studies and model simulations have hinted that upwelling events trigger the most intensive offshore exchange of upwelled nutrients in the gulf (Laanemets et al., 2004, 2011; Zhurbas et al., 2008). It has also been suggested that an excess of phosphorus (one of the major sources for the formation of cyanobacteria blooms) tends to exist in the surface layer during the upwelling event for some time and that the main reserve of phosphorus tends to be located in shallower depths (in the range of the thermocline) than the nitrogen (Laanemets et al., 2009).

More generally, our results provide a possible key for better

Fig. 15. The reach of the patch of cooler upwelled water from the shoreline as a function of thickness of surface layer for surface current speed 0.1 m s⁻¹ and $R_e = 40–100$ km (left) and the same reach for $R_e = 100$ km and different values of surface current speed (right).

Fig. 16. Burger numbers for cross-shore profiles in Fig. 5. The profiles are numbered from west to east. Red squares represent the Burger number ≥ 1.5 within the cooler jets. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

Fig. 17. Schematic representation of the upwelling event in the western Gulf of Finland. The upwelled waters most likely originated from the 15–30 m depth. The event produced a cooler cross-shore surface jet that extended 40–45 km from the coast.
understanding and quantification of the location(s) (in terms of both depth and coastal segment) where the main transport of these nutrients (phosphorous) may originate from. The relevant fluxes apparently strongly depend on the appearance of the bathymetry (represented here as the average slope of the nearshore). These slopes play an important role in the development of cross-shore jets of cooler surface waters. It is important to note that the nearshore slope is not uniform within the area covered by the upwelling process. The cross-shore jets seem to mostly originate from distinct locations with steep slopes > 0.0075. These results can be very useful for mitigation of onset of cyanobacteria blooms. They also provide an option for rapid (albeit rough) estimates of the surface layer thickness required to produce the cross-shore jet and the depth of the source of the cooler water.

Even though the analysis focuses on a single upwelling event, it highlights several intriguing aspects. Firstly, the theory of Narimousa and Maxworthy (1985), originally developed to identify the width of the sea area covered by the patch of cooler water jet, can be used for express estimates of a solution to an inverse problem, namely, how thick should be the surface layer to produce a widely spreading (up to 40–45 km) system of upwelling jets and filaments. Such a method can be conveniently used with satellite data as a hint for the dynamics of the upwelling events. This approach is useful when limited data are available as it is basically sufficient to have adequate information about the surface layer thickness, stratification and surface current speed. Our analysis suggests that one cannot assume that the formal surface layer thickness (the depth from where the cooler water reaches the sea surface) is homogenous over longer coastal segments. In contrary, this thickness varies depending on the average slope of the nearshore.

Even though the theory does not take into account the actual shelf slope, it produces a fair estimate of the offshore reach of the cross-shore jets in segments that have steep slopes. It is highly interesting that specifically the areas with steep bottom slopes seem to be prone for hosting surface cross-shore jets that reach far offshore. Although this conjecture reflects quite limited observational material, it is remarkable that the locations in question match the regions identified in previous studies as potential start or end points of rapid cross-gulf Lagrangian transport patterns (Soomere et al., 2011). The relevant simulations were based on modelled currents in a very thin (3 m) surface layer in a model where the bottom slopes were involved only implicitly. These results imply that the bathymetry slope is usually not the only (or governing) parameter that affects the development of cross-shore surface jets and that the thickness of the surface layer and the horizontal velocity normally play a major role. However, the presented material suggests that the areas where the average slope of the nearshore seabed varies largely may significantly influence the location of cross-shore jets. It is highly desirable to include the potential alongshore variations in the nearshore slope into the relevant theory. Also, it is crucial to employ high resolution bathymetry in studies of coastal upwelling.

The analysis also once more highlights the classic issue of the importance of high resolution bathymetry and stratification data in the understanding of the course and consequences of the upwelling process. In particular, the shape of the seabed seems to be crucial for the development and persistence of the mid-phase of the upwelling during which the most intense cross-shore surface transport of upwelled waters and substances occurs. The role and consequences of the presence of this mid-phase (which lasts a few days) remain largely unclear; however, it is likely that it may play a major role in quantifying and understanding the impact of upwellings on a global scale, especially in the Southern Ocean.

Finally, we mention that the observed and forecast climate changes (in particular, rotation of strong wind direction and changes in the persistence of winds, Soomere et al., 2015; Soomere and Pindsoo, 2016; Kudryavtseva and Soomere, 2017) may considerably influence the occurrence, intensity and appearance of upwelling events (Bakun, 1990; Wang et al., 2015). Using the resources available from satellite data, the method proposed in this study can be used as a simple express method to quantify some properties of upwellings based on fairly limited resources. This method can be used for other areas in the Baltic Sea and also globally.

5. Conclusions

- A simple method that uses satellite derived sea surface temperature data and limited in situ data is developed to evaluate several properties of upwellings.
- Steep bottom slopes of > 0.0075 have a strong influence on the location and development of cross-shore jets observed during upwelling events in the Gulf of Finland.
- The upwelled cooler water was most likely sourced from the interior of the water column during the upwelling event in May–June 2013. This feature suggests that the dynamics of nutrients and their connection to cyanobacteria blooms may substantially depend on the depth from where the upwelled water is actually sourced.
- Whilst bathymetry plays a major role in the onset of cross-shore jets, other elements of the system such as the wind stress, currents, stratification and thickness of the surface layer also influence the presence of these cross-shore jets and the depth from where the upwelled water is actually sourced.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.ecss.2018.03.018.

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