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Internal Solitary Waves in the White Sea: Hot-Spots, Structure, and Kinematics from Multi-Sensor Observations

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Abstract: A detailed picture of internal solitary waves (ISWs) in the White Sea is presented based on an analysis of historical spaceborne synthetic aperture radar (SAR) data and field measurements. The major hot-spot of ISW generation locates in the southwestern (SW) Gorlo Strait (GS), characterized by the presence of strong tides, complex topography, and two distinct fronts. Here, pronounced high-frequency isopycnal depressions of 5–8 m were regularly observed during flood and flood/ebb slackening. Other regions of pronounced ISW activity are found near Solovetsky Islands and in the northwestern Onega Bay. The spatial and kinematic properties of the observed ISWs are linked to water depth, with larger wave trains and higher propagation speeds being observed over the deep regions. Direct estimates of ISW propagation speeds from sequential and single SAR images agree well, while theoretical ones obtained using a two-layer model overestimate the observed values by 2–3 times. This is explained by the effective modulation of ISW propagation speed during the tidal cycle by background currents that are not accounted for in the model. Enhanced values of diapycnic diffusion coefficient in the pycnocline layer were registered near the frontal zones, where intense 14–17 m high ISWs were regularly observed.

Keywords: internal solitary waves; hot-spots; tidal dynamics; frontal zones; wave kinematics; SAR imaging; vertical mixing; White Sea; Arctic Ocean



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1. Introduction

The White Sea (WS) is one of the most isolated and most commonly studied basins of the Arctic Ocean [1,2]. It is known for very strong tides and the most intense tidal energy dissipation in the Eurasian Arctic [3,4]. Powerful tidal dynamics create strong quasi-periodic semidiurnal tidal currents reaching 1.5 m/s [2]. Under such conditions, internal tides (ITs) of the dominant semidiurnal period should arise in areas where stratification is well expressed. As the models predict [5,6], ITs can have amplitudes of up to 10 m at generation sites, and due to the proximity to the critical latitude for the M_2 tide, they should dissipate locally, eventually evolving into wave trains of short-period internal solitary waves (ISWs) [7].

Oceanographic observations of the last decade have made it possible to detect significant short-term variabilities of currents, temperature, and salinity in various regions of the White Sea [8–12]. Strong thermocline oscillations associated with the propagation of short-period ISWs of 15–17 m height and periods of 15–20 min were repeatedly recorded in the southwestern part of the sea [13,14]. The formation of large-scale ISW trains generated due to tide-topography-frontal zone interactions in the eastern WS crossing the entire sea basin, and dissipating on the shallow western shelf were also reported in [15], based on an analysis of spaceborne synthetic aperture radar (SAR) data.

However, at the moment, there is no detailed information regarding the spatio-temporal distribution of ISWs in the White Sea that has been confirmed by long-term observations, nor their generation hot-spots, spatial and kinematic properties, or influence

on water mixing. In this paper, building on historical SAR data and in situ measurements acquired in 2007–2014, we aim to fill this gap and present an updated record of the ISW field in the White Sea.

Materials and Methods for the ISW identification of SAR data and the in situ measurements are introduced in Section 2. In Section 3, we present the results of ISW observations from satellite and in situ data, including the identification of their hot spots, horizontal and vertical properties, peculiarities of their generation and kinematics in the key hot spot, and impact on vertical mixing. In Section 4, we discuss and summarize the obtained results.

2. Materials and Methods

To investigate the field of internal solitary waves in the White Sea, here, we use satellite and in situ measurements made during extended summer periods from May to September in 2007–2014. Satellite observations were primarily used to infer hot-spot regions of ISW activity, and their spatial and kinematic properties, while in situ data were used to characterize background hydrology, understand the vertical structure of ISWs, and assess their influence on water mixing. Below, we describe in more details each dataset and the methods used to extract the corresponding ISW properties.

2.1. Satellite Observations

Here, we used a collection of 282 C-band SAR images acquired by ENVISAT Advanced SAR (ASAR) and RADARSAT-1,2 from May to September of 2007–2013 (Table 1). In particular, we used 7 images in 2007, 2 images in 2008, 65 images in 2009, 131 images in 2010, 48 images in 2011, 21 images in 2012, and 2 images in 2013. The data of ENVISAT ASAR, 256 images, were used for 2007–2011; they were acquired at vertical polarization in Wide Swath Mode (WSM) and Image Mode Precision (IMP), with a viewing area of $400 \times 400 \text{ km}^2$ and $100 \times 100 \text{ km}^2$, and a spatial resolution of 150 m and 25 m, respectively. For 2012 and 2013, we used 26 RADARSAT-1,2 images with spatial resolutions of 25 m and 6 m, respectively. The summary of the SAR data availability for each month of the study period is given in Table 1. As seen, the portion of high-resolution SAR images ($<30 \text{ m}$) is about 30% of the total number of images used. Most of the data were from 2009 (65 images) and 2010 (131 images), while the maximum number of images for all years was available in July (79 images) and August (74 images).

Table 1. Number of SAR images by month, covering the White Sea in 2007–2013.

Month	Number of SAR Images		
	Envisat ASAR		RADARSAT-1,2
	IMP	WSM	
May	0	43	0
June	0	45	0
July	20	48	11
August	17	43	15
September	19	21	0
Total	56	200	26

Figure 1 shows the total spatial coverage of the study site using SAR data during the study period. As seen, the sea area was covered pretty well. Over 120 SAR images covered the Basin, Onega, and Dvina bays (OB and DB, respectively); and two major straits around Solovetsky Islands (SI), namely, Zapadnaya (west of SI) and Vostochnaya (east of SI). Solovetskaya Salma straits, hereinafter, are ZSSS and VSSS, respectively. A slightly lower number of data points, 80–100 images, was available over the periphery of the study region, i.e., in the northwestern part of the Kandalaksha Gulf, in the northeastern Gorlo Strait (GS), and in the southeastern OB.

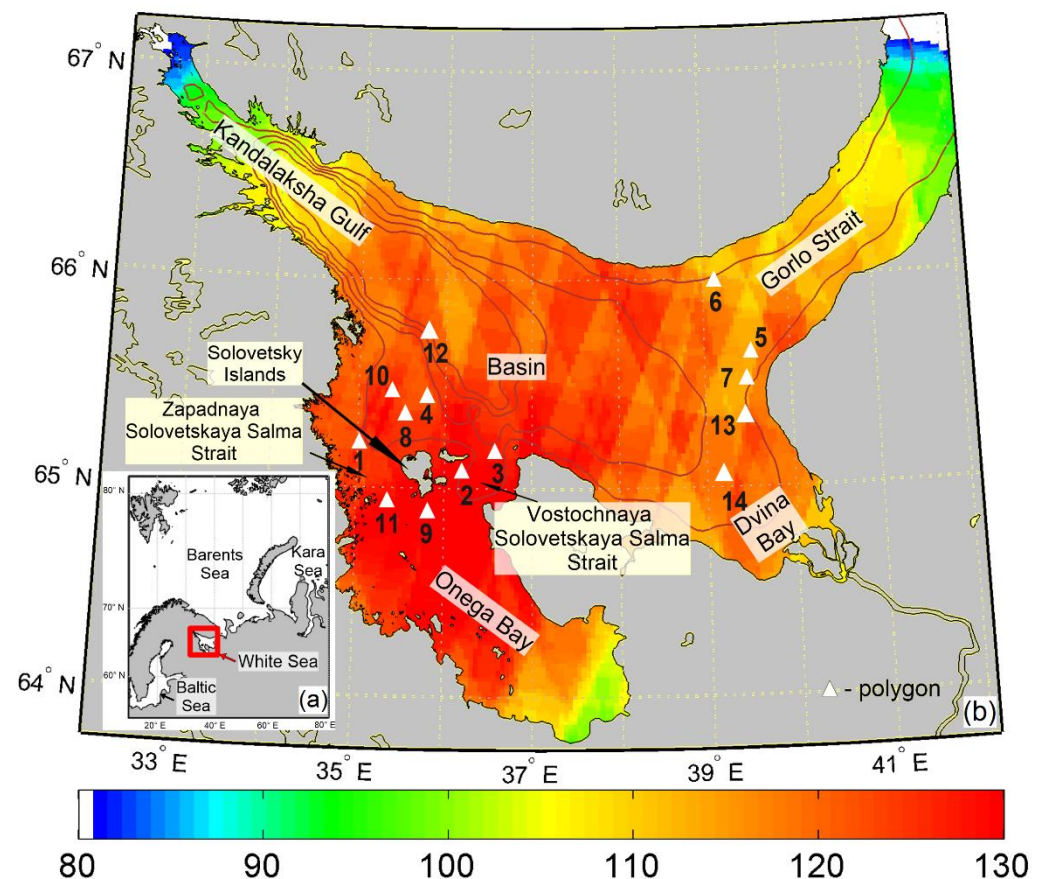


Figure 1. (a) Map of the Northern Europe with the location of the White Sea marked by red box. (b) Spatial coverage of the study site using satellite SAR data. The number of available SAR images is shown in color. White triangles and numbers mark the locations of polygons of contact measurements made in 2009–2014. Overlaid grey lines are isobaths drawn with a 50 m step taken from IBCAO v.3.0 [16].

In SAR images, ISWs usually have very recognizable patterns of alternating bright and dark bands of enhanced and depressed radar backscatter of concentric form (see Figure 2), as ubiquitously demonstrated for both the Arctic regions [17–22] and the rest of the World Ocean [23–27].

In this work, the surface manifestations of oceanic ISWs were visually detected in SAR images, and further analysis of their properties followed the methodology described in [28]. Following [28], once clear signatures of ISWs were detected, a curve matching the leading wave of each ISW train and a transect crossing the wave train in the propagation direction were drawn to estimate a number of spatial ISW properties. To reduce the speckle noise typical for SAR images of the sea surface, the values of normalized radar cross-section (NRCS or σ^0) along the transect were calculated as a mean of 10 parallel transects, as has often been performed in similar studies (e.g., [24]). Then, using a mean σ^0 profile along the transect, its orientation, and the image itself, we determined the following ISW properties: crest length (L_c) of the ISW train, its length (L), area (A), and propagation direction (d), the wavelength of the leading wave (λ_0), and the interpacket distance (λ_g) between sequential wave trains (Figure 2). The latter was used to estimate the propagation speed (C_p) of ISWs, assuming that the generation of every new packet occurred every tidal period, with the lunar semidiurnal M_2 tide being the dominant tidal component in the White Sea [6].

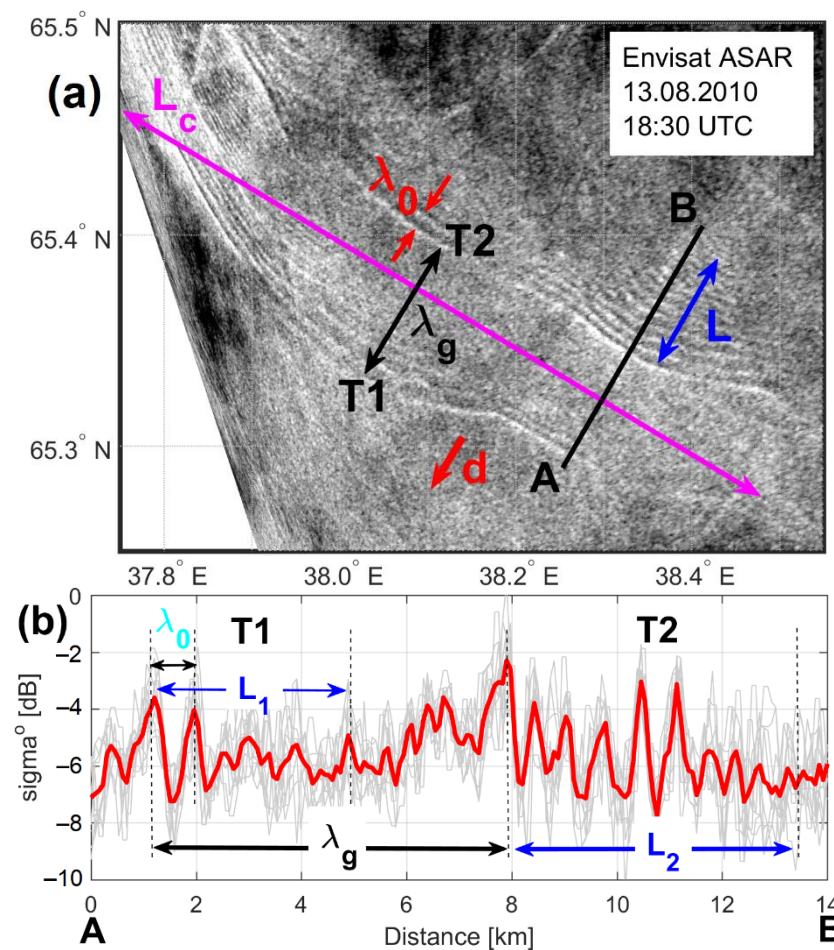


Figure 2. Detection of the ISW parameters from (a) an enlarged Envisat ASAR image acquired on 13 August 2010 at 18:30 UTC, and (b) SAR signal variations along the transect AB shown in (a).

Figure 2 shows an example of an enlarged Envisat ASAR image acquired on 13 August 2010 (18:30 UTC) depicting the clear signatures of two sequential ISW trains, T1 and T2, propagating southwest from the shallow GS toward the deep Basin. It also schematically shows how the different ISW properties are estimated from the 2D SAR image (Figure 2a) and from the variation of the SAR signal backscatter along the transect AB crossing the wave trains T1 and T2 (Figure 2b). The interpacket distance λ_g between the wave trains T1 and T2 is about 7 km, which equals to $C_p = \lambda_g / T_{M2} = 7000 \text{ m} / 12,43600 \text{ s} \approx 0,16 \text{ m/s}$.

The spatial fields of an ISW probability and the mean values of their properties were then calculated on a horizontal grid of 40×40 cells with a cell size equal to about 110 km^2 . The mean values of interpacket distances and propagation speeds were obtained only for the data of 2010 (comprising approximately half of all the SAR data used). The information on ocean depth corresponding to the center of each detected ISW train was derived from IBCAO version 3.0 [16].

2.2. In Situ Measurements

In situ observations were made in July–August, 2009–2014, in the two straits around the SI (the ZSSS and the VSSS), the Basin, the Onega Bay, the Gorlo Strait, and the Dvina Bay, at 14 polygons shown by white triangles in Figure 1. The observations included the installation of autonomous buoy stations and frequent oceanographic stations (scans) made from an anchored vessel. All works were carried out with the data binding to the tidal phase. Measurements were made when the wind speed was below 8 m/s .

Systems of two or three buoy stations approximately 1–2 km apart were installed at each polygon. The duration of the observations lasted from one to three M_2 tidal cycles.

Doppler flow profilers ADP SonTek-500 and ADCP Workhorse Sentinel 300 equipped with pressure and temperature sensors were installed near the bottom. The discreteness of the measurements was 2 min. In addition, JFE ALEC (<https://www.jfe-advantech.co.jp/eng/products/ocean-ryusoku.html#aem1da>, accessed on 28 August 2022) current meters were installed at 10, 13, 16, and 19 m depths. CTD instruments SBE-19plus (<https://www.seabird.com/sbe-19plus-v2-seacat-profiler-ctd/product?id=60761421596>, accessed on 14 July 2022) and CTD-90M (<https://www.sea-sun-tech.com/product/multiparameter-probe-ctd-90-memory/>, accessed on 22 July 2022) were installed in the pycnocline layer. Data from the buoy stations were used mainly to obtain additional information about the hydrological conditions in the study area, and to assess the propagation velocity and the direction of the ISWs.

Basic information regarding the ISWs was obtained by scanning the water column from surface to bottom with CTD-90M and Cast Away (<https://www.ysi.com/castaway-ctd>, accessed on 17 August 2022) sensors from the anchored vessel. One “descent-ascent” scan cycle took about 1–2 min. The vertical properties of the ISWs were estimated from the observations of the thermocline depth that usually coincided with the pycnocline depth. Therefore, we also used the water temperature data for the analysis of the ISW properties. As shown in Figure 3, the time variations in the position of the thermocline allowed us to determine the height and the period of each wave passing through the polygon. The wave height was estimated as the average value between the height of the front and the rear slopes of a soliton, and the period of each wave was estimated as the time interval between the adjacent maxima of the isotherm (Figure 3). Oscillations with the periods from 3 to 60 min and a height of more than 1 m were considered as ISWs. The former was largely dictated by the technical constraints of the measurements, i.e., the lower limit (3 min) was determined by the shortest period of the repeated CTD casts (over 1–2 min), and the upper limit was determined by the minimal interval of the continuous measurements (~2 h).

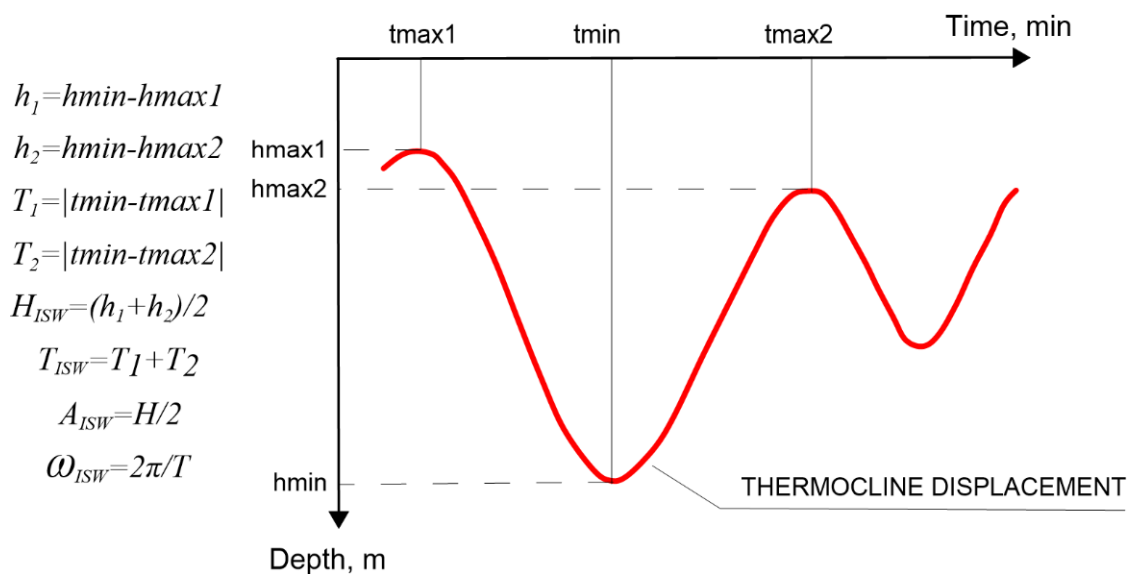


Figure 3. Schematic of the ISW height and period retrieval from vertical CTD casts.

To assess the role of the ISWs in vertical mixing we used temperature, salinity, and current speed measurements at a 0.5 m vertical grid to calculate the turbulent diffusivity or the diapycnic diffusion coefficient [29]:

$$K_z = 0.2\varepsilon / N^2 \quad (1)$$

where $N^2 = -(g/\rho_0)\partial\rho/\partial z$ is the squared buoyancy frequency (mean over tidal cycle), N —the Brunt–Väisälä frequency, and ε is the dissipation rate of turbulent kinetic energy defined from the current meter data following [30]:

$$\varepsilon = 7.5\nu \overline{\left(\frac{\partial u}{\partial z}\right)^2} \quad (2)$$

where ν is the dynamic viscosity of seawater, and $\partial u/\partial z$ is the vertical shear of horizontal velocity. All calculations were performed for the water layer where the ISWs were registered, typically from 5 to 35 m (the actual depths ranged at various polygons) using two time-averaging intervals. In the first case, we used a 2 min interval to keep the effects of the ISWs in the data. In the second case, we applied a 2 h averaging to fully eliminate the effects of the ISWs. Further, the results of both cases were averaged during the tidal cycle to assess the contributions of the short-period processes to the vertical turbulent exchange. For the long-lasting polygons where the ISWs were registered, the calculations were repeated for 2–3 M_2 tidal cycles, and then averaged.

3. Results

3.1. Hot-Spots of the ISWs and the Background Properties

In total, 516 surface signatures of the ISWs were registered in 282 SAR images in May–September of 2007–2013. A seasonal distribution of the number of the available SAR images is shown in Figure 4. Note that the number of the medium-resolution (WSM) SAR images (black bars) is given separately from the higher-resolution SAR data (grey bars). The number of WSM images was nearly similar from May to August, and around half as many in September. The higher resolution SAR data were available only during July–September, with a smaller number of images in September. Despite the relative number of the high-resolution images during these months being 2.5 times smaller than that of the WSM images, the number of identified ISW signatures was nearly similar, i.e., 244 ISW signatures in the high-resolution images and 272 in the medium-resolution images.

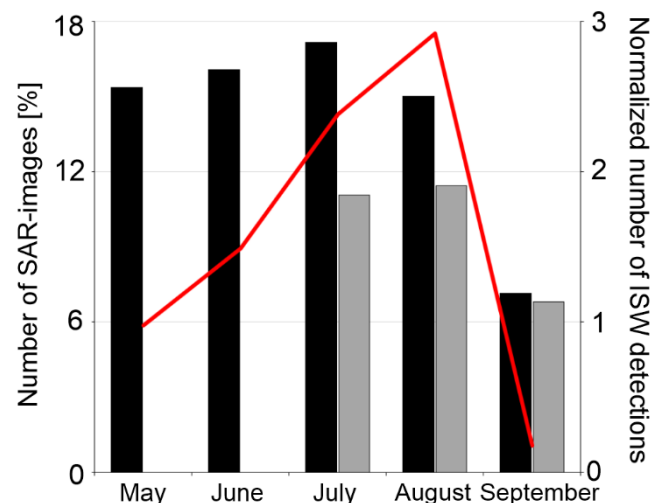


Figure 4. Histogram distribution of the number of available SAR images (black and grey bars, left axis) and the normalized number of ISW detections (red line, right axis) per month during May–September of 2007–2013. Black (grey) bars indicate the number of medium (high)-resolution SAR images.

As the data availability strongly varied from month to month, we used a normalized number of ISW detections (red curve in Figure 4) calculated as a ratio of the total number of ISW detections per month to the number of SAR images available per given month. As seen, the highest number of ISW detections was registered in August, followed by July, i.e., during the months when the vertical stratification conditions were the most suitable for ISW generation in the WS [13].

Figure 5a shows a map with the plotted leading crests of all ISWs detected in the White Sea during the study period. As seen, most waves are concentrated in the northeastern part of the sea, i.e., in the southwestern (SW) part of the Gorlo Strait (GS) and at the outer boundary of the Dvina Bay (DB). Pronounced ISW activity is also seen in the eastern part of the Basin, in the straits around the Solovetsky Islands archipelago, and in the northwestern Onega Bay (OB). Apparently, no ISW activity is seen over the deep-sea regions in the Kandalaksha Bay (KB) and the western Basin.

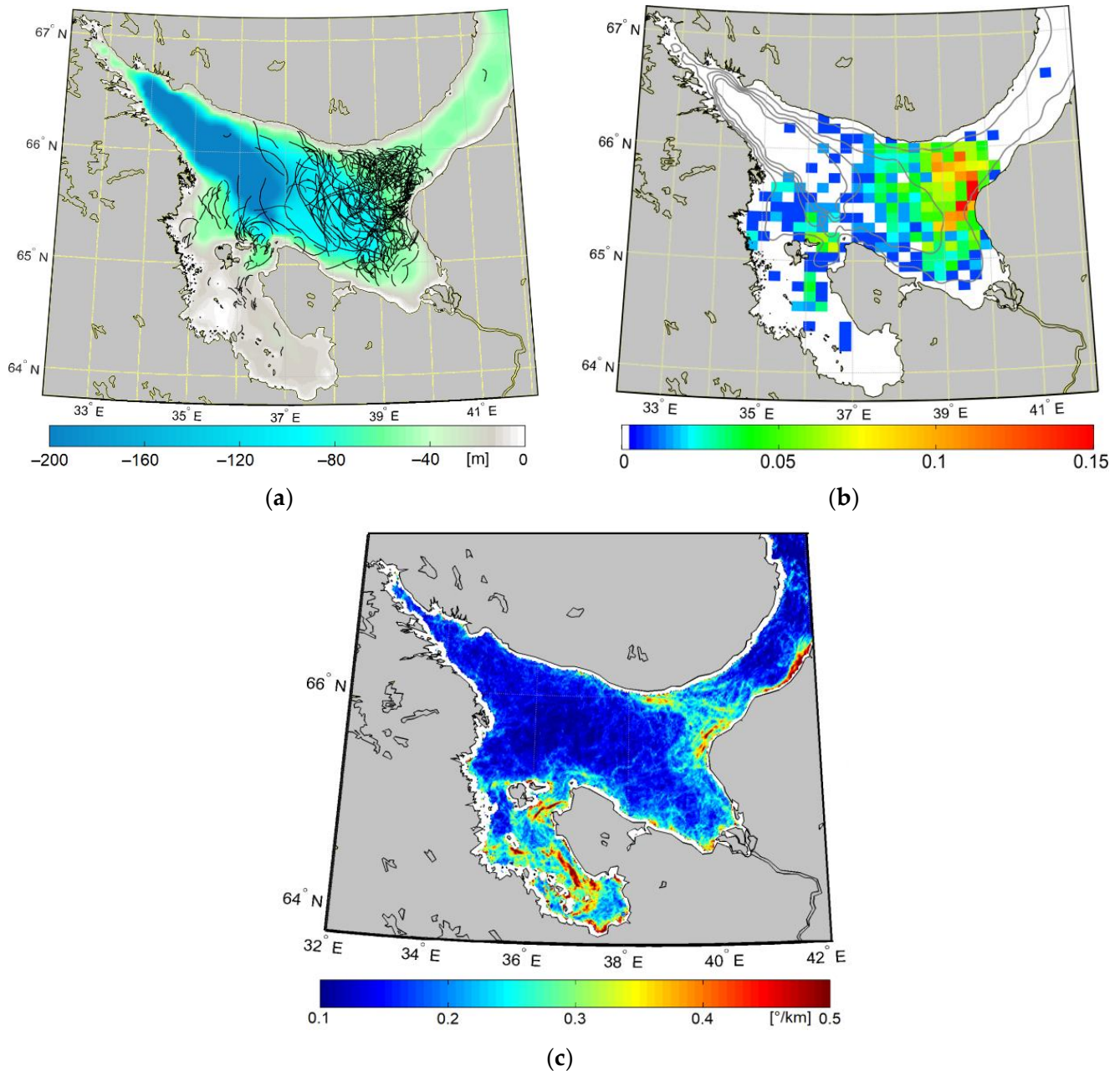


Figure 5. Summary of the ISW observations in the White Sea during 2007–2013: (a) Map with location of leading waves of all ISWs detected during the study period, (b) spatial variability of the ISW probability in the study region, (c) mean SST gradient calculated based on MODIS Aqua data acquired in May–September 2010.

Figure 5b further shows a spatial distribution of the ISW probability calculated on the horizontal grid of 40×40 cells by dividing the total number of ISW detections per grid cell by the total number of SAR observations of that grid cell. The highest probability values

(0.1–0.2) are found at the boundary between the western Basin, the SW GS, and the DB—the major hot-spot of regular ISW generation in the White Sea. Other regions of regular ISW activity are located in the vicinity of the SI, particularly east of it (in the VSSS), where the ISW probability is about 0.08. In the rest of the sea, it does not exceed 0.05 (Figure 5b).

Figure 5c shows a spatial distribution of a mean sea surface temperature (SST) gradient obtained from the daily MODIS Aqua images acquired in May–September 2010. The SST gradient field clearly depicts the main frontal zones of the White Sea, many of which are found in the SW Gorlo Strait, in the VSSS, and in the Onega Bay. An intercomparison of the SST gradient field with that of the ISW probability shows a considerable correspondence between the regions of the high ISW probability and the high SST gradients, suggesting that many of the observed ISWs are generated in the vicinity of strong frontal zones.

To illustrate certain differences of hydrological conditions in the various WS regions, Figure 6 shows the vertical distributions of the Brunt–Väisälä frequency averaged over the tidal period from the field measurements made at polygons 1, 4, and 7 in the summer of 2010 (polygon 1) and 2012 (polygons 4 and 7).

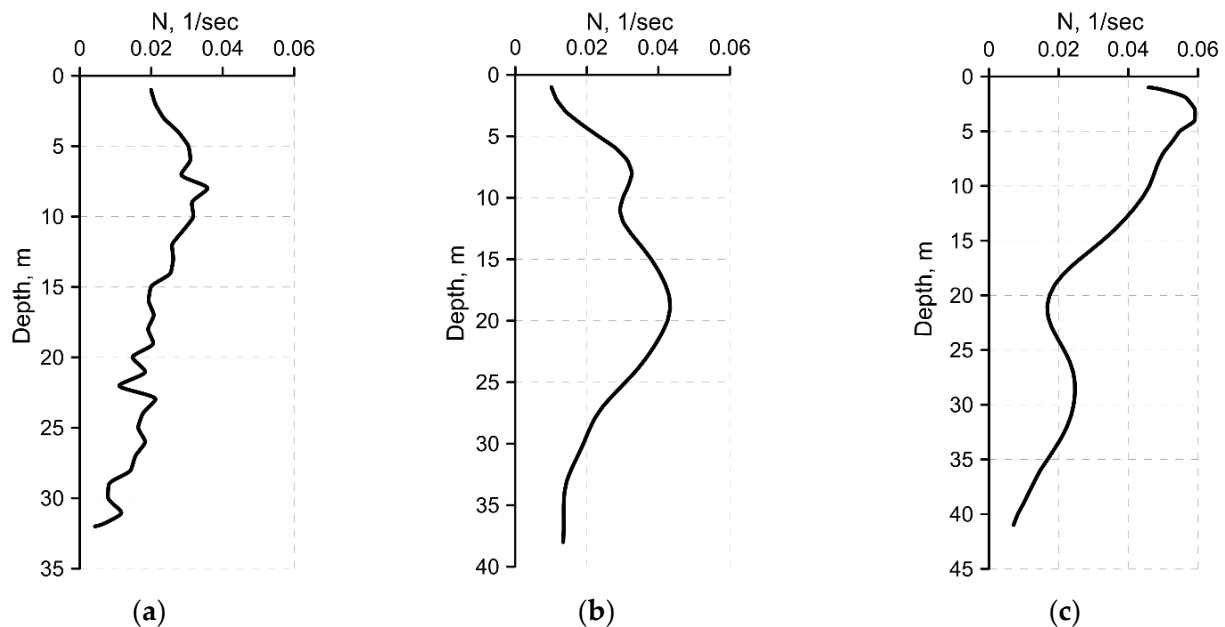


Figure 6. Vertical profiles of buoyancy frequency at polygons: (a) one in summer, 2010; (b) 4 in summer, 2012; and (c) 7 in summer, 2012.

At polygon 1 located on the shallow western shelf (at the boundary of the western Basin and the ZSSS, see Figure 1), the maximal $N(z)$ value $\sim 0.04 \text{ s}^{-1}$ is found at an 8–9 m depth, with the typical thickness of the pycnocline layer being about 5 m (Figure 6a). Polygon 4 is located at the shelf–deep water boundary. The maximal $N(z)$ value here ($\sim 0.043 \text{ s}^{-1}$) is similar to polygon 1, but it is found deeper, at 15–25 m. The pycnocline layer is about 10 m thick here (Figure 6b).

Polygon 7 is located in the coastal area of the eastern WS where the occurrence of ISWs is highest (Figure 5b). Here, the buoyancy frequency profile strongly differs from other polygons, due to a strong impact of the Dvina River runoff, described in more detail below. Its main peak is found at about a depth of 5 m, and reaches $\sim 0.06 \text{ s}^{-1}$. The impact of the riverine water extends down to 20 m, while another minor peak of $\sim 0.025 \text{ s}^{-1}$ at 25–30 m depth is associated with the subsurface water originating from the Basin (Figure 6c).

At all polygons, the water structure corresponds to the so called “Basin type”, characterized by well-expressed vertical density stratification. The upper layer, extending to 5–20 m depth, is formed by surface waters warmed during the summer and desalinated by the river runoff and sea ice melt. Subsurface water is saltier, with the temperature gradually decreasing with depth.

The overall analysis of measurements at the different WS polygons shows a significant variability of thermohaline fields in two ranges of periodicity, with a characteristic semi-diurnal (tidal) period and periods of tens of minutes (associated with ISWs). In different regions characterized by different hydrological and morphometric conditions, it manifests itself in different ways:

- (i) In the deep Basin and bays of the White Sea, the most intense fluctuations have a semi-diurnal period;
- (ii) In the shallow parts of the WS (e.g., in shelf areas with a well-pronounced two-layer stratification) tidally induced sharp changes in the vertical water structure are regularly observed, with the main contribution being made by short-period fluctuations related to ISWs;
- (iii) In the tidally active regions near frontal zones (at the boundary of the Basin and the GS), first mode ISWs are observed every tidal cycle, and their heights often exceed the upper mixed layer depth, and the period is about 10–20 min;
- (iv) In the Onega and Dvina Bays, weakly pronounced short-period fluctuations are observed against the background of well-pronounced semidiurnal variations in the pycnocline. Stronger fluctuations associated with ISWs are observed occasionally.

3.2. Spatial and Kinematic Properties of ISWs from SAR Observations

In this subsection, we present the results addressing the spatial and kinematic properties of ISWs derived from SAR data. Figure 7 shows a map and a histogram of ISW propagation direction. As seen, two main directions prevail—western/southwestern and eastern. While some regions have a single well-defined ISW direction, the others have a mixture of them, due to a local complex bottom emanating the waves in many possible directions.

The wavelength of the leading wave in ISW packets ranges from 0.1 to 2.6 km (Figure 7c,d). The wavelength of ISWs increases with the water depth (Figure 7c), being around 0.5 km on average over the shallow regions, and above 1 km in the deeper parts of the sea. The smallest values (0.1–0.3 km) are seen east of the SI, in the OB, and in some parts of the GS.

Figure 8 further shows the spatial distribution and histograms of other spatial properties of ISW packets—their crest length, packet length, and area.

The crest length values are strongly varying in space, with the smallest values of 5–10 km being observed east of the SI in the VSSS and the OB, moderate values of 15–20 km—in the SW GS and the DB, and the highest values above 40 km in the central and eastern parts of the Basin (Figure 8a), known for the development of the large-scale ISWs propagating westward with crests exceeding 100 km [28]. The overall range of the crest length values is 1.5–143 km, with a mean value of about 17 km (Figure 8b).

The length of ISW packets varies from 0.1 to 11.5 km, with a mean value of 1.3 km (Figure 8c,d). More than 75% of all packets have length values of below 2 km that are broadly found all over the eastern White Sea (Figure 8c). Small values of 1 km and below are registered in the GS, around the SI, and in the OB. The longest packets (>8 km) are seen over the deeper regions in the northwestern part of the sea.

Knowing the crest length and the packet length of ISWs, it is easy to obtain the area of the ISW packets by multiplying the former two. As obtained, the area of about 80% of the detected ISW trains does not exceed 50 km². As seen from Figure 8e, smallest wave trains ($A < 20$ km²) are observed in the SW GS, the OB, and around the SI. The largest ISW packets with the area exceeding 150–200 km² are again observed over the deeper (>100 m) parts of the Basin. The overall range of the ISW packet area values spans three orders of magnitude from ~0.2 km² to ~854 km², with a mean value of ~30 km² (Figure 8f).

During the analysis of SAR images, sequential wave packets having identical patterns and propagation direction were often registered. As has been already mentioned above, and as widely applied in literature [28,31,32], the distance between such sequential wave trains, λ_g , can be used to assess the propagation speed of ISW trains, C_p (Figure 9).

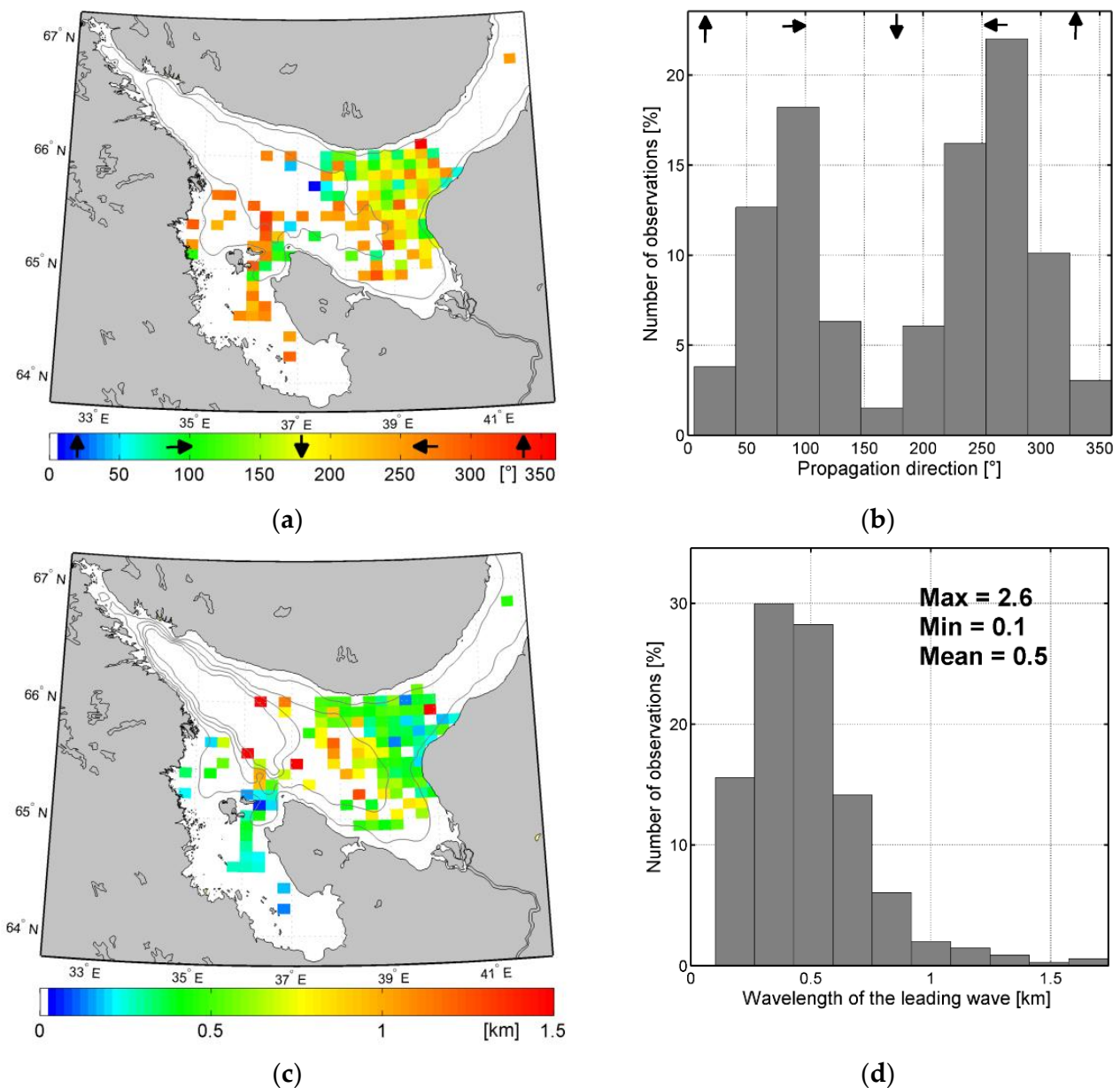


Figure 7. Spatial maps (a,c) and histograms (b,d) of ISW propagation direction (a,b) and wavelength of the leading wave (c,d).

Interpacket distances were obtained in 67 cases. As seen from Figure 9a, higher λ_g values correspond to higher water depths, and vice versa. The overall range of λ_g values is between 0.7 and 46 km, with a mean value of about 7 km (Figure 9b).

Assuming that generation of every new ISW train happens each M_2 tidal cycle (i.e., every 12.4 h), the ISW propagation speed values corresponding to the observed λ_g values equal 0.02–1.03 m/s, with a mean value of 0.15 m/s (Figure 9c,d). The values of $C_p < 0.1$ m/s do not seem very convincing, because of the large relative error in measuring the distance between consequent ISW trains (± 150 m). In turn, the range of values of 0.1–1 m/s is rather realistic. In similarity to λ_g values, C_p increases as the waves propagate from shallow to deep water. The different shallow sea regions have similar C_p values of 0.1–0.2 m/s, while the highest propagation speed is attributed to the largest and perhaps most intense ISWs.

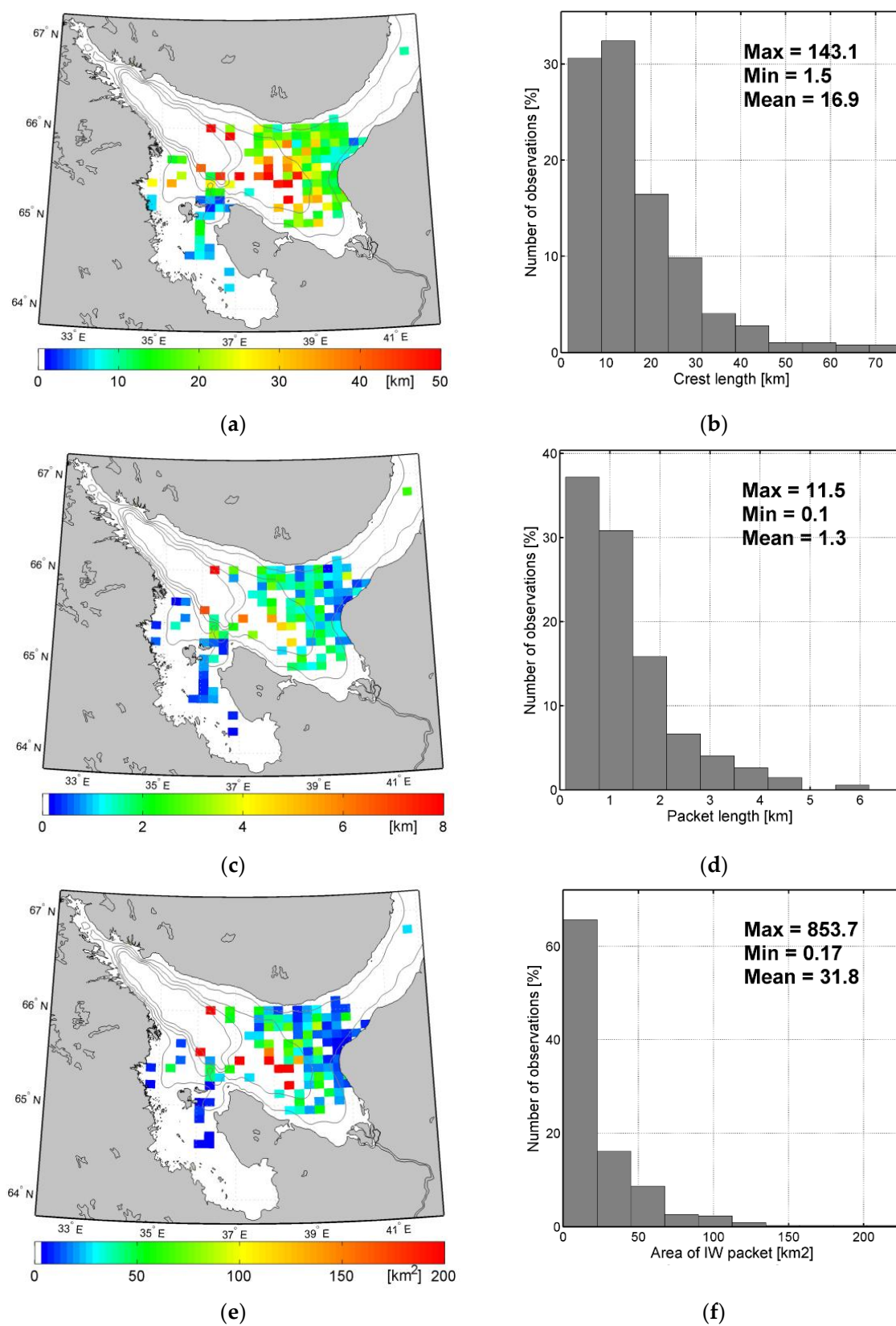


Figure 8. Spatial maps (a,c,e) and histograms (b,d,f) of ISW crest length, packet length (c,d), and area (e,f).

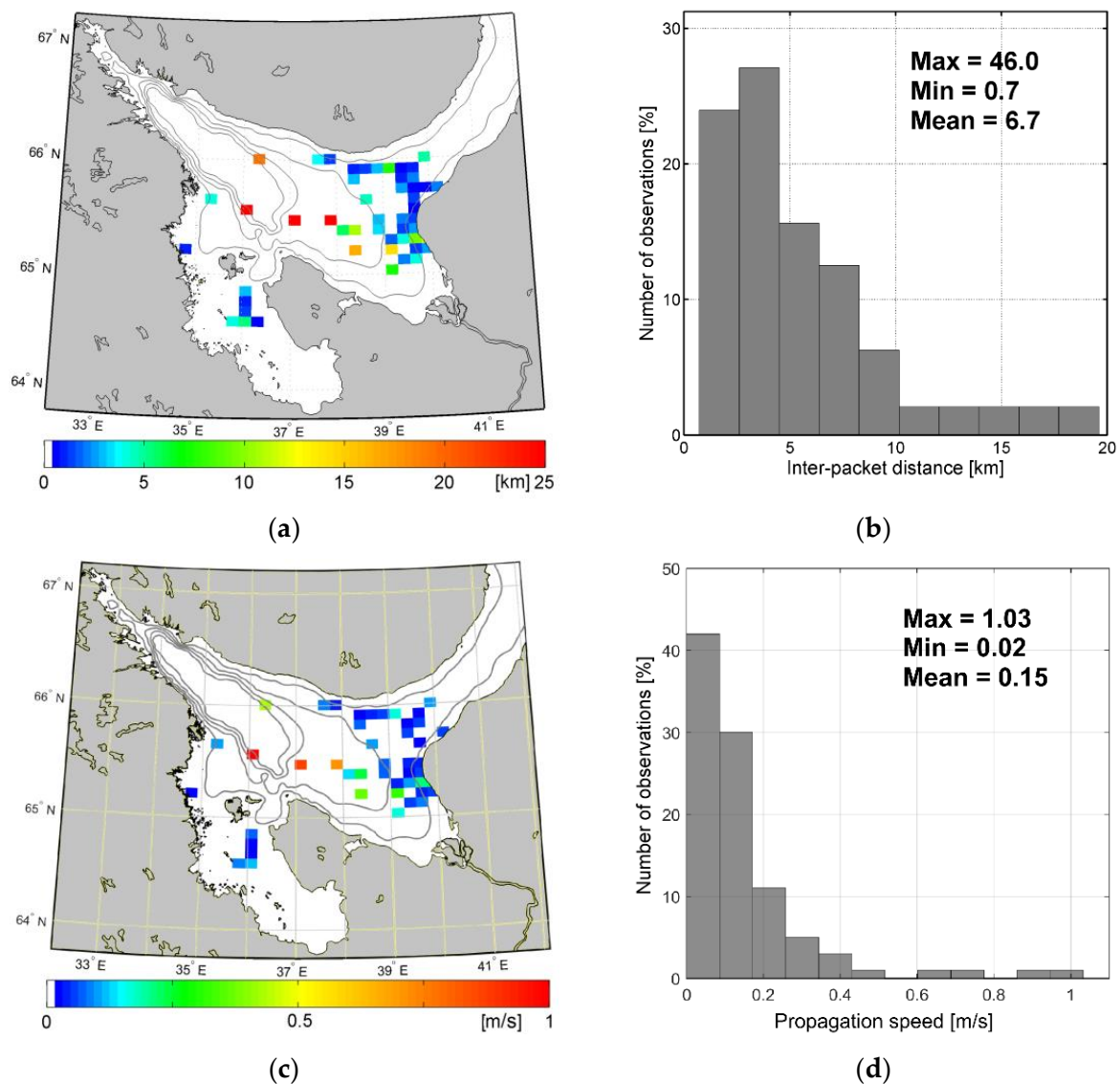


Figure 9. Spatial maps (a,c) and histogram (b,d) of interpacket distance (a,b) and propagation speed of ISWs (c,d).

3.3. ISW Generation in the SE Gorlo Strait

As already shown above in Figure 5a,b, the region including the western part of the Basin, the SW GS, and the DB is the main hot spot of ISW generation in the White Sea. It has a very dense and complex ISW pattern. This region is known for intense tides, and the formation of sharp density fronts and complex bottom topography, all resulting in the generation of the large-scale ISWs that are able to cross the entire Basin and dissipate on the western sea shelf [12]. Below, we consider the ISW formation in this region in more detail.

The SST map acquired on 4 August 2010 (Figure 10a) shows a typical picture of SST distribution in the eastern part of the WS, where two very pronounced frontal features permanently exist. One of them, hereinafter the Gorlo Front (GF), is found in the SW GS, where the relatively cold, saline, and mixed Barents Sea water entering the GS from the northeast meets the warmer water of the WS Basin. Another one is formed in the Dvina Bay by the Northern Dvina River discharge. This river plume water, hereinafter the Dvina Front (DF), is a couple of degrees °C warmer than the surrounding water of the Basin, and much warmer (>10 °C) than the GF water. It fills the southern part of the DB, and then spreads northward along the Winter Coast (see Figure 10a for its location).

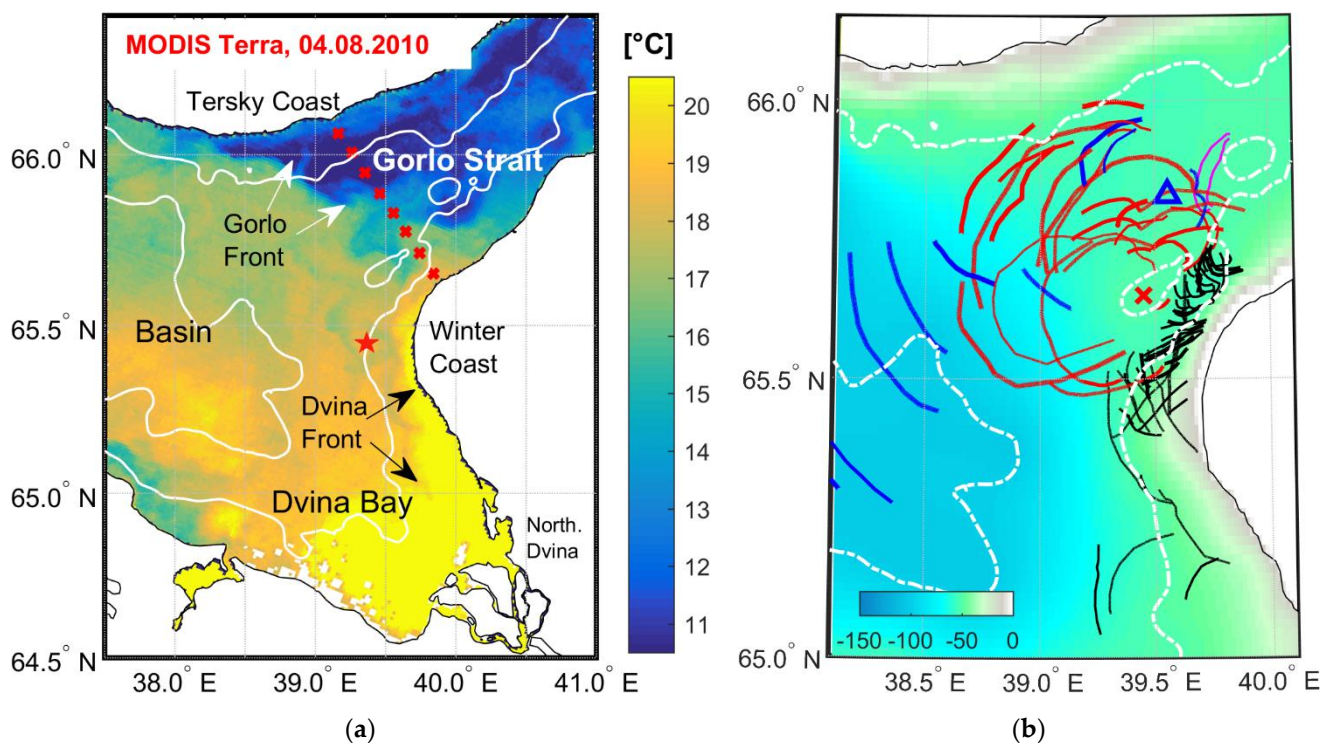


Figure 10. (a) MODIS Terra SST map acquired on 4 August 2010 over the eastern part of the White Sea, showing two major fronts formed by the Barents Sea water entering through the Gorlo Strait and the Northern Dvina River outflow to the Dvina Bay. Red crosses and red star show locations of the transect crossing the GS and polygon 7, respectively. (b) Locations of ISW crests detected on 29 July–5 August 2012 in Radarsat-1/-2 data; red curves show ISW signatures detected on 29 July 2012 (see Figure 10). Red cross and blue triangle mark underwater banks in the GS. Solid and dashed white lines in (a,b) show 50 m and 100 m isobaths from IBCAO v.3 [16].

To obtain a better understanding of the ISW generation and structure in the SW GS, a 4 day experimental study was carried out in this part of the sea on 29 July–1 August 2012. The experiment included measurements along the section crossing the GS, and fixed-point measurements at polygon 7 (see Figure 10a for their locations), all collocated with the Radarsat-1/-2 overpasses.

Figure 10b shows a bathymetry map covering the SW GS, and adjacent regions with overlaid ISW crests detected in the Radarsat-1/-2 data on 29 July–5 August 2012. The overall ISW pattern is rather complex; hence, we show the ISW signatures in different colors depending on their presumed source of origin. Red color denotes ISW signatures most often having circular shapes and presumably originating from the interaction of the DF, with an underwater bank located to the northwest of the Winter Coast (marked by a red cross in Figure 10b). Blue color shows ISW signatures typically directed southwest from the GS toward the deep Basin. These wave trains usually have rather long crests (up to 200 km), and are presumably generated by the complex interaction of tidal currents, the Gorlo Front, and a system of underwater sills (marked by a blue triangle in Figure 10b). The last group of the ISWs shown in black represents the wave trains found in the immediate proximity to the Dvina Front. Below, we consider the overall hydrology and the ISW properties in this region in more detail.

Figure 11 shows a vertical distribution of water temperature along the transect crossing the GS made during the ebb and flood tides on 30–31 July 2012. The leftmost part of the transect shows a well-mixed GS water of Barents origin having a temperature of about 7 °C and salinity of about 28 psu (not shown). Stratified water of the northern Basin fills the entire central part of the transect, with the temperature varying from 13.5 °C near the surface to 2 °C near the bottom (the salinity varies from 25.3 to 27.5 psu). In both graphs, a

temperature front outcropping to the surface is found at about 10 miles from the Tersky Coast; it marks a cross-strait position of the Gorlo Front. The along-strait position of the GF also depends on the width and intensity of the relatively warm and fresh Dvina Front that is usually attached to the Winter Coast (Figures 10a and 12a). The DF is seen in the rightmost part of the transect during the ebb tide as a warm ($>14^{\circ}\text{C}$) 5–7 m thick surface layer (Figure 11a).

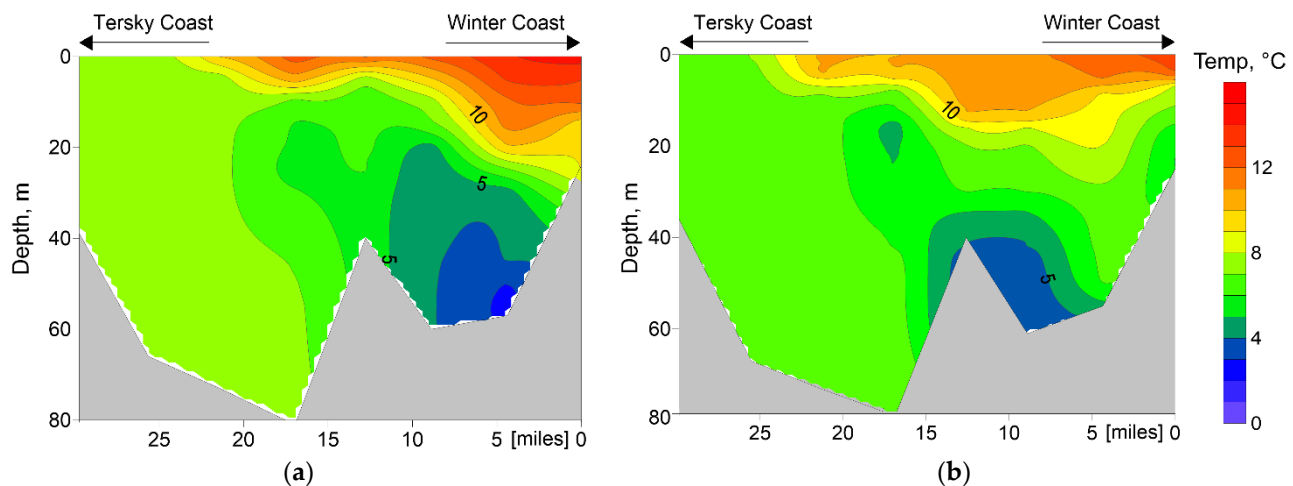


Figure 11. Temperature variations along the section crossing the Gorlo Strait during the ebb (a) and flood (b) tides on 30–31 July 2012.

The tidal currents are very strong in the Gorlo Strait, reaching 1.5 m/s [2], and can easily influence the vertical and horizontal location and intensity of frontal zones in the strait (Figure 11). Tidal currents are directed southwest (northeast) toward the WS Basin (Barents Sea) during the flood (ebb) tide. During the ebb tide, stratified water of the Basin and the DF comes into the strait along the Winter Coast, forming a strong vertical temperature gradient of up to $1^{\circ}\text{C}/\text{m}$; the thermocline lies very close to the surface, at 5–12 m depth (Figure 11a). During the flood, less stratified water partly mixed in the strait returns back, the vertical temperature gradient weakens, and the thermocline deepens to about 18–20 m (Figure 11b).

Figure 12 shows a time plot of vertical distribution of density anomaly calculated using EOS-80, together with currents and the tidal range at polygon 7. The polygon was occupied during ~24 h, starting from 19:30 UTC on 31 July to 19:00 UTC on 1 August 2012.

The total water depth at the polygon was ~60 m, but the CTD casts were made down to 40 m to shorten the sampling intervals to 2–3 min. Frequent CTD-sampling resolved the high-frequency oscillations of the upper 40 m quite well, revealing a strong variability of isopycnal depths, both in the pycnocline layer (~5–15 m) and below (Figure 12a–c).

As seen from Figure 12, low-frequency variations of isopycnals are consistent, with a dominant semidiurnal signal of tidal range. Pronounced isopycnal depressions of 5–8 m are observed on 31 July at 21:50, 23:50 UTC and on 1 August at 01:30, 02:40, 06:50, 8:40, 10:10, and 11:30 UTC. Coherent isopycnal displacements are also seen below the pycnocline layer down to 30 m depth. Observed high-frequency internal waves are nonlinear, have a packet structure (3–5 individual waves each) with the most pronounced leading wave and decaying oscillations toward the rear of the packet. The period of wave trains is about an hour, while the period of individual waves is 10–20 min.

The most intense waves are observed during the flood (21:50, 8:40, and 10:10 UTC), with depth-integrated southward currents reaching 0.5 m/s (Figure 12). Some wave trains (23:50, 06:50, and 11:30 UTC), were also observed during the flood/ebb slackening and reversal. A maximum tidal current speed up to 0.98 m/s was observed during the ebb tide, when two wave trains were recorded (01:30 and 02:40 UTC). Strong tidal currents

during the flood and ebb tides were recorded over the entire water column, peaking in the pycnocline layer.

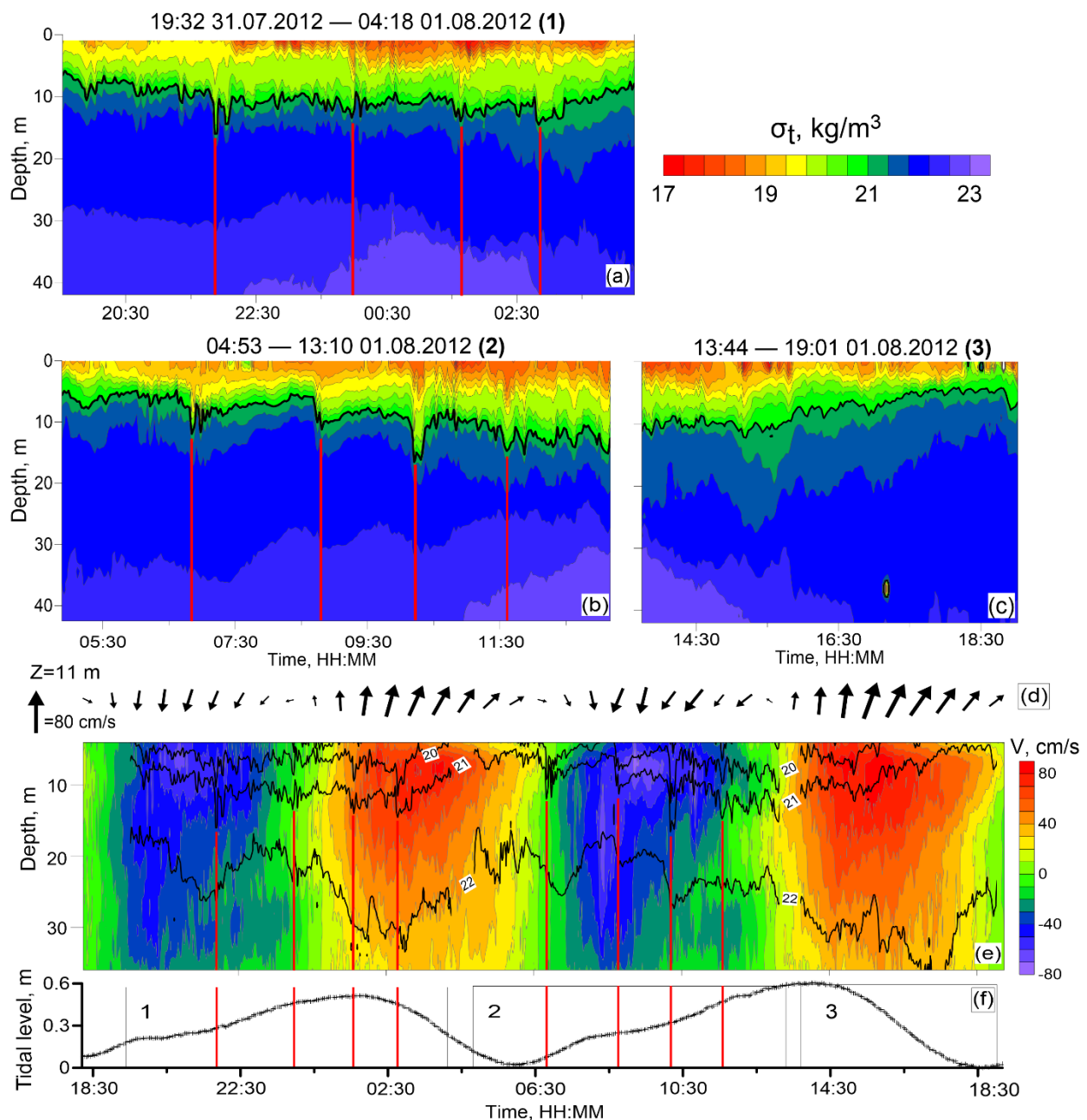


Figure 12. Hydrographic measurements at polygon 7 during a 24 h station on 31 July–1 August 2012: (a–c) Vertical distribution of density anomaly from CTD casts made during time periods 1–3; ADCP measurements of (d) current speed and direction at 11 m depth. (e) North–south current component; (f) tidal level. Red vertical lines mark the time intervals when ISW trains are clearly present in the data.

A detailed 2D picture of the ISW field in the study region is captured by a series of Radarsat-1/-2 images. Figure 13a shows an enlarged part of a Radarsat-2 HH-polarization image acquired on 29 July 2012 (14:46 UTC) over the SW GS. It shows a number of distinct hydrodynamic patterns—signatures of surface current front (the Dvina Front, marked by yellow dashed line in Figure 13a) along the Winter Coast, large semicircular ISW trains spreading offshore, and smaller wave trains located near the current front. A wider view of ISW signatures seen in this SAR image are shown in Figure 10b by red curves.

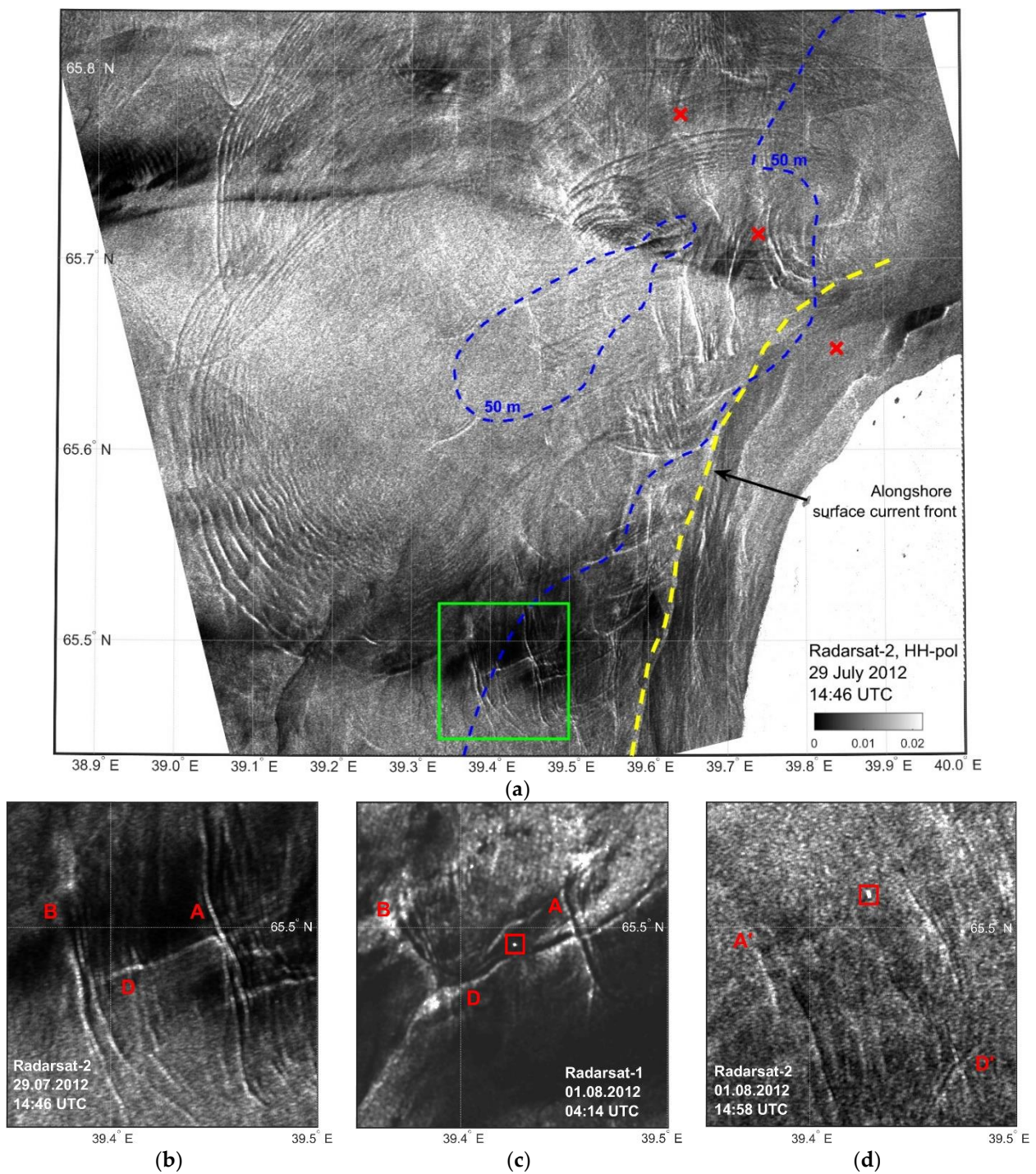


Figure 13. Examples of ISW manifestations in the southwestern part of the Gorlo Strait. (a) Radarsat-2 image acquired on 29 July 2012 at 14:46 UTC. Green box marks the location of an enlarged fragment shown in (b). Blue dashed line shows 50 m isobath. Yellow dashed line shows the location of the alongshore surface current front associated with the Dvina Front. Red crosses mark a station crossing the Gorlo Strait. (b) Enlarged fragment of (a) with marked locations of the ISW trains A, B, and D. (c) Enlarged fragment of a Radarsat-1 image acquired on 1 August 2012 at 04:14 UTC with marked locations of the ISW trains A, B, and D. Red square marks the ship location. (d) Enlarged fragment of Radarsat-2 image acquired on 1 August 2012 at 14:58 UTC with marked locations of the ISW trains A' and D'. Red square marks the ship location.

Bathymetry contours at a 50 m isobath taken from IBCAO v.3 [16] show the existence of an elongated bank centered at 65.67°N, 39.5°E and oriented along the strait (Figure 13a). In general, the observed ISW pattern fits well with the noted bathymetry feature, assuming its generation via the common lee-wave mechanism [33,34]. Indeed, ISWs were usually registered during the flood, and their locations as observed in the SAR data are found on the lee side of the topographic obstacle, with the waves propagating downslope.

A green box in Figure 13a marks the region of polygon 7 that is captured in two other SAR images of 1 August depicting the ship location and the ISWs nearby (Figure 12c,d). An enlarged crop of the Radarsat-2 image of 29 July shows a system of two pairs of consequent ISW trains traveling west-southwest (WSW, wave trains A, B) and south-southeast (SSE, wave train D), i.e., approximately at right angles to each other (Figure 13b).

On the next Radarsat-1 image of 1 August (04:14 UTC) one can see that the ISW pattern is exactly similar to that in Figure 13b. Note that the locations of wave trains A, B, and D are almost identical (Figure 13c). The time lag between the two images of 29 July and 1 August is equal to 61 h 28 min, i.e., almost exact five M_2 tidal periods, suggesting an M_2 periodicity of ISW arrival to the location of polygon 7. On the same day, a Radarsat-2 image was made at 14:48 UTC, i.e., 10 h 44 min later than the preceding Radarsat-1 image. It shows a similar configuration of the ISW trains (Figure 13d). We presume that the ISW signatures A' and D' denoted in the second image (14:48 UTC) are linked to the wave trains A and D seen in the first image (04:14 UTC). In such a case, their horizontal shifts in space and the time gap (10 h 44 min) between the two sequential observations (SO) can be used to assess their propagation speed (C_p), as is usually performed from such observations [32,35,36].

For simplicity, we consider only the leading waves of each wave train. An estimation of the horizontal shift of the ISW signatures in the SO is made with a 1 km step along each wave crest. Figure 14 shows a resulting map of the propagation speed of four ISW trains present in both images. First, we consider the wave trains A and B. In all fragments of SAR images shown in Figure 12b–d, the interpacket distance (ID or λ_g) between them is about 3.6 km, which, assuming their generation every M_2 period (12.4 h), equals to a propagation speed of 0.08 m/s. An analysis of the SO gives C_p equal to 0.087 m/s.

The wave train D has a curved shape in both images of 1 August 2012 (Figures 13c and 14). The interpacket distance for D can be estimated only for its southwestern part, i.e., between D and E₁, and equals 4.5–4.7 km or ~0.12 m/s, in terms of C_p . From the sequential images, C_p is ~0.1 m/s for the southwestern part and ~0.6 m/s—for the northeastern. For the wave train E, consisting of two parts E₁ and E₂ clearly visible in the first image (04:14 UTC, not shown), the SO-based mean propagation speed is equal to 0.11 m/s for E₁ and 0.13 m/s for E₂, respectively. The alternative estimate of C_p for E can be made from the IDs between wave trains D' and E'. In the vicinity of E₁ and E₂, the IDs are 4.8–5 km and 6–6.5 km, respectively, yielding C_p values of 0.12–0.13 m/s and 0.16–0.17 m/s.

In general, the ISW propagation speed estimates obtained from SOs and IDs agree quite well with the ID-based estimates providing somewhat higher values. Some variations in the propagation speed values along the crests of individual ISW trains (e.g., wave trains D and E) are observed, which are most probably linked to differences in sea depth and vertical stratification along their trajectories. Higher C_p values are seen over the deeper regions and vice versa, in accordance with the results obtained in Section 3.2 above (Figure 9c) and the results of previous studies [32,37].

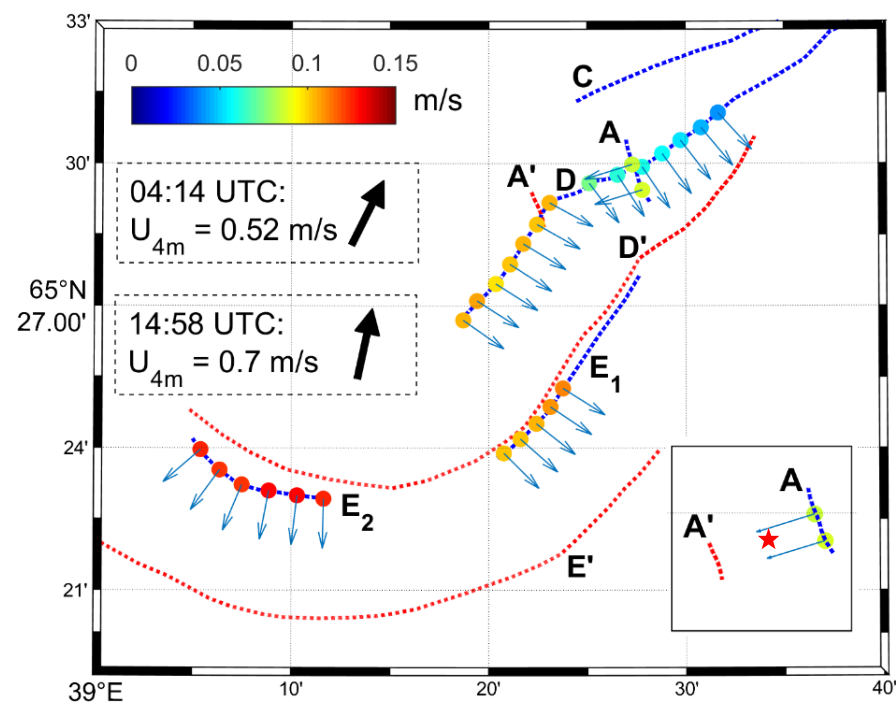


Figure 14. Propagation speed of ISWs derived from sequential Radarsat-1/-2 images on 1 August 2012. Blue (red) dotted lines show locations of ISW leading waves identified in Radarsat-1 (Radarsat-2) image at 04:14 (14:48) UTC. Colored circles with arrows denote ISW propagation speed and direction, respectively. Letters mark particular ISW trains. Two dashed boxes show information on tidal speed and direction at 4 m depth at times of SAR acquisitions. Black box in the bottom right corner shows propagation speed and direction for the waves A and A', and the location of polygon 7 (shown by red star).

Another option to estimate C_p is to use a theoretical approach and the vertical CTD measurements made at polygon 7. For this purpose, a convenient option is to use a dispersion relation for the lowest internal mode in a two-layer approximation, often used in similar applications [22,31,38,39]. In this case, the phase speed of the linear internal waves is [40]:

$$c_0 = \left(\frac{\delta\rho}{\rho_0} \frac{g}{k} \right)^{1/2} \{ \coth kd + \coth k(D-d) \}^{-1/2} \quad (3)$$

where $\delta\rho$ is the density difference between the upper and lower layers, ρ_0 —average density below the pycnocline, g —gravity acceleration, k —wave number defined as $k = 2\pi/\lambda$, where the wavelength λ is taken as a wavelength of the leading waves in ISW trains from SAR observations, d —mean pycnocline depth, and D —full depth. In the case of nonlinear internal solitary waves, the phase speed would be equal to:

$$C = c_0 + \alpha\eta_0/3 \quad (4)$$

where η_0 —amplitude of ISWs, and α —coefficient of quadratic nonlinearity, defined as:

$$\alpha = 3/2c_0(h_1 - h_2)/h_1h_2 \quad (5)$$

where h_1 , h_2 —upper- and lower-layer depths, respectively. Using the average over the tidal cycle characteristic values of $D = 60$ m, $d = 8$ m, $\delta\rho = 2$ kg/m³, $\rho_0 = 1022$ kg/m³, $\lambda = 250$ –400 m, and $\eta_0 = 2$ –8 m, we obtain $C = 0.2$ –0.32 m/s.

The theoretical C values are about two times higher than C_p obtained directly from the SAR images, ranging from 0.08 to 0.17 m/s. The possible explanation of such a difference could be related to the fact that the two-layer model does not account for background

currents that can modulate the effective propagation speed of ISWs. As Figure 14 shows, the surface (and depth-averaged) currents were quite strong (0.5–0.7 m/s) and opposing relative to the ISW travel directions at the moments of the SAR observations. Strong counterclockwise tidal currents superposed with wind drift and the alongshore current of the Dvina Front altogether should modulate the effective ISW propagation speed during the tidal period, explaining the observed difference. In our case, the northward ebb current is stronger than the southward flood current (Figure 12d) possibly explaining the lower SAR-based propagation speed values of southward traveling ISWs, compared to the theoretical ones.

3.4. ISW Heights and Associated Vertical Mixing

In situ data collected at 14 polygons enabled some statistics about the ISW properties to be obtained over the different WS regions. Table 2 provides a summary of ISW detection frequency per M_2 cycle, and their periods and heights recorded at polygons.

Table 2. Statistics of ISW properties in the White Sea recorded at polygons 1–14 in 2009–2014.

Polygon (Year)	N of ISWs Per M_2 Cycle	ISW Period: Average (Min–Max), Min	ISW Height: Average (Min–Max), m
1 (2009)	33	11 (3–36)	3.1 (1–7.8)
1 (2010)	46	18 (4–72)	5.6 (1–17.9)
1 (2011)	40	12 (3–32)	3.7 (1–13.9)
1 (2012)	57	7 (4–34)	2.2 (1–7.7)
1 (2013)	48	12 (3–36)	2.2 (1–7.6)
2 (2012)	81	7 (3–25)	2.6 (1–9)
3 (2012)	68	9 (3–24)	2.5 (1–8.3)
3 (2014)	36	14 (3–30)	4.5 (1–15)
4 (2012)	12	8 (4–26)	1.5 (1–3)
5 (2012)	12	9 (5–21)	1.8 (1–4.9)
6 (2012)	0	–	–
7 (2012)	57	13 (3–42)	2.5 (1–7.6)
8 (2012)	11	8 (4–23)	1.6 (1–3.5)
9 (2009)	25	14 (4–58)	3.9 (1–14.7)
10 (2011)	12	25 (6–68)	1.6 (1–3.3)
11 (2011)	26	6 (2–20)	1.7 (1–4.9)
12 (2013)	3	30 (18–41)	1.2 (1–1.8)
13 (2014)	21	16 (6–32)	1.8 (1–7.0)
14 (2014)	7	26 (18–56)	1(1)

The most intense ISWs in terms of wave height were recorded in the vicinity of the frontal zones (polygons 1–3, 7, 9, and 13). Their heights were often comparable to the thickness of the upper mixed layer (see example in Figure 15). In the deep Basin and far from the frontal zones and bottom irregularities (polygons 4, 8, 10, 12, and 14), the ISWs were recorded less often, usually having relatively small heights.

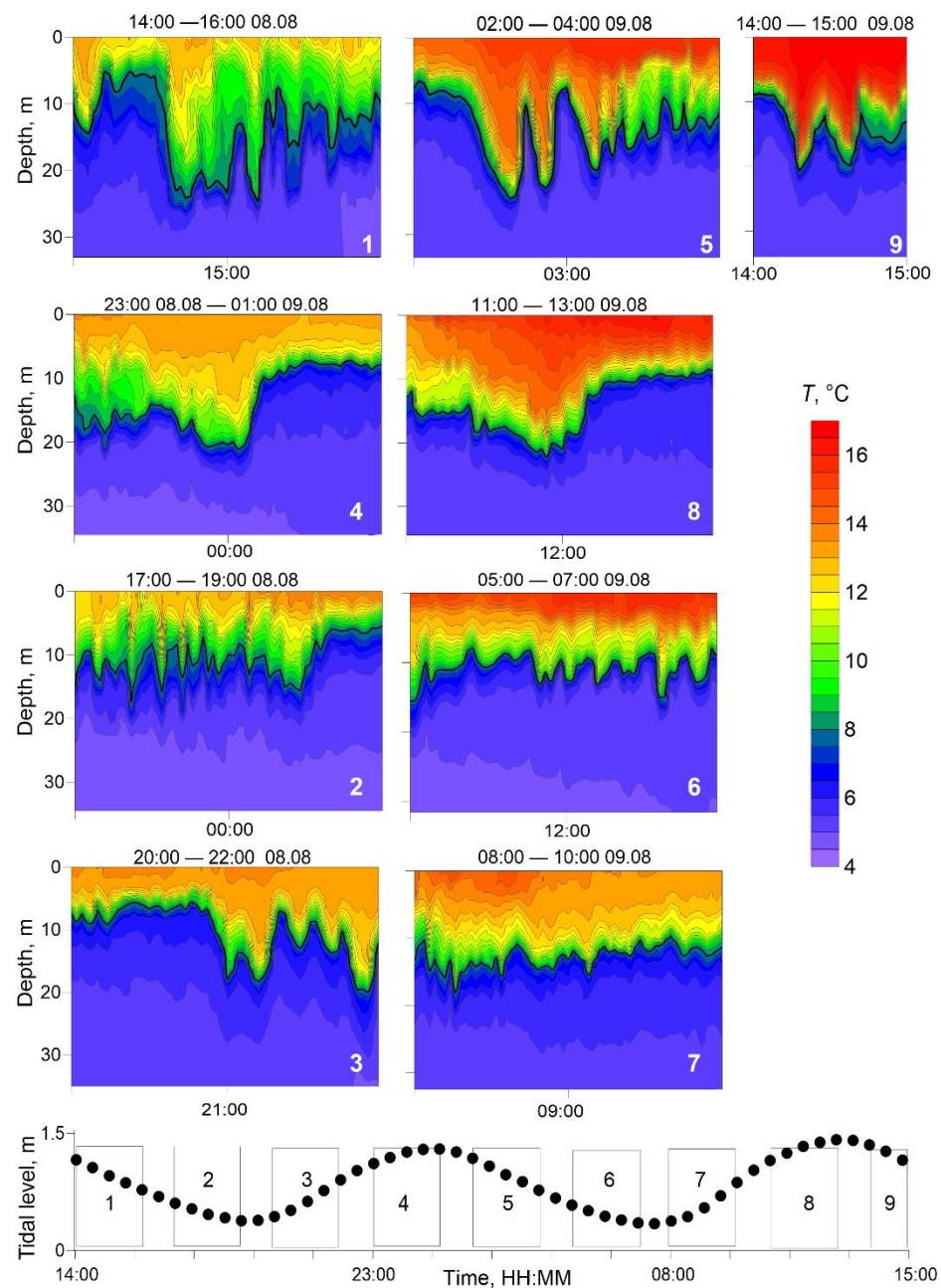


Figure 15. Vertical distribution of water temperature from repeated CTD casts made at polygon 1 during a 25 h station on 8–9 August 2010. Bottom graph shows tidal level variations and periods of observations marked by numbered boxes.

Rather intense ISWs 8–15 m high were registered south of the SI (polygon 9) and in the VSSS (polygons 2, 3)—a very dynamic region with strong tidal currents and very complex bottom topography. Here, the number of ISWs registered per tidal cycle was the highest. In the eastern WS close to the Dvina Front (polygons 7, 13), the maximum wave heights were typically not exceeding 7–8 m.

The most intense ISWs, 14–18 m high, were recorded in summer 2010 and 2011 at polygon 1 located near the frontal zone on the border of the Basin and the ZSSS. In other years (2009, 2012–2013), their heights were typically below 8 m. The mean values during 2009–2013 for polygon 1 are 3–5 m for the ISW heights and 12–17 min for their periods. The intermittency of ISW events was about 50% in the record, with at least 2–3 intense ISW trains observed each tidal cycle.

Figure 15 shows the observations of intense ISW trains made at polygon 1 in August 2010. The polygon was located in the vicinity of an underwater bank Severnaya, with an average depth of 10 m surrounded by regions of 35–50 m depth. Shipborne measurements were completed south of the bank over a 35 m depth during 25 h from 14:00 UTC on 8 August to 15:00 UTC on 9 August 2010. As shown earlier in Figure 6a, the buoyancy frequency profile at polygon 1 has maximal $N(z)$ value ($\sim 0.04 \text{ s}^{-1}$) at 8–10 m depth. The time variations of sea level and water temperature show distinct semidiurnal variability.

The data analysis clearly shows the formation of a southward propagating internal bore with a very sheer, 10–14 m high front during the peak flood (episodes 4 and 8 in Figure 15). Before it comes, the water structure has three distinct layers: a 8–12 m thick warm surface layer, a 8–10 m thick thermocline layer, and a nearly homogeneous cold bottom layer. After the bore passing, the upper layers become thinner (about 4–5 m each) and the vertical temperature gradient across the thermocline rises to $2 \text{ }^{\circ}\text{C m}^{-1}$.

About 2.5 h later (during the ebb), the tidal current slackens and gradually changes direction eastward. At this moment, the ISW trains, consisting of 3–5 strongly nonlinear waves with maximal heights of 14–17 m and periods of 10–20 min, are seen in the record (episodes 1, 5, and 9 in Figure 15), followed by a tail of smaller waves of shorter period (episodes 2 and 6). The propagation of such intense ISW trains observed each tidal cycle is accompanied by the surface signatures in the form of slicks (not shown), commonly observed elsewhere (e.g., [24]).

Notably, the first two waves in the ISW train shown in episode 1 are less clear than those in episode 5, due to possible wave breaking. Indeed, the values of Richardson number $R_i = N^2/S^2$, where $S^2 = (\partial u/\partial z)^2 + (\partial v/\partial z)^2$ is the squared current shear, are strongly below 1 at the time interval from 14:41 to 15:06 UTC, indicating favorable conditions for hydrodynamic instability, wave breaking, and enhanced vertical mixing in the water column.

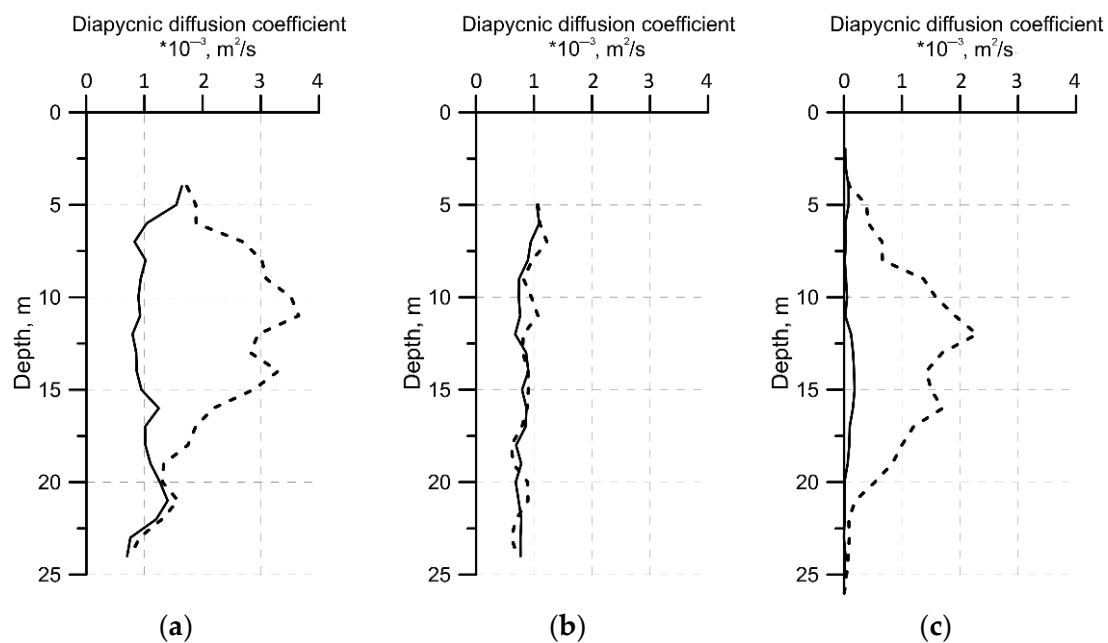
Averaged over the full period of observations (25 h, see Section 2.2 for details), the values of the diapycnic diffusion coefficient K_z in the pycnocline layer (8–14 m) are $12.1 \cdot 10^{-4}$ and $21.5 \cdot 10^{-4} \text{ m}^2\text{s}^{-1}$ when applying 2 h and 2 min time averaging steps, respectively (Table 3). The difference indicates the contribution of the ISWs to the vertical turbulent exchange. The vertical K_z profile obtained in summer 2011 over the same location shows a similar situation (Figure 16a). The K_z values resolving high-frequency variability in the water column are 3–4 times higher than the background values of $\sim 10^{-3} \text{ m}^2\text{s}^{-1}$ obtained with the 2 h averaging. The peak values of $33\text{--}37 \cdot 10^{-4} \text{ m}^2\text{s}^{-1}$ are registered in the pycnocline layer at 10–15 m depths.

A summary of the K_z estimates obtained at polygons 1–14 using the two time-averaging steps is given in Table 3. The values of K_z vary from $3 \cdot 10^{-4}$ to $23.5 \cdot 10^{-4} \text{ m}^2\text{s}^{-1}$. In most cases, the K_z values are decreasing with depth (Figure 16). However, the character of vertical K_z variability could differ significantly, depending on the time-averaging interval of the initial data. The pronounced maxima of the K_z values immediately below the pycnocline were observed at polygons 1, 3, 7, and 13 for the estimates made with the 2 min step, and absent in the data with the 2 h averaging (see Table 3 and examples in Figure 16a,c).

It is important to note that enhanced K_z values in the pycnocline were registered mostly at polygons near the frontal zones where intense ISWs of higher than 5 m were regularly present (Table 3). To the contrary, the regions with absent ISWs activity (polygons 4, 5, 10, and 14) are characterized by low K_z values that do not depend on the choice of the averaging interval. This is well illustrated in the vertical K_z profiles at polygon 4 (Figure 16b), where both profiles have identical values and average around $10^{-3} \text{ m}^2\text{s}^{-1}$. All these facts allow us to conclude that ISWs strongly influence the hydrology and water mixing in the White Sea, leading to enhanced vertical turbulent exchange in the coastal and strait regions in the vicinity of strong frontal zones.

Table 3. Averaged K_z estimates at different polygons supplemented with statistics of ISW observations in summer 2009–2014.

Polygon (Year)	Layer, m	Diapycnic Diffusion Coefficient, $\text{m}^2\text{s}^{-1} * 10^{-4}$		N of ISWs Per M_2 Cycle with Height > 5 m
		Averaging Period		
		2 h	2 Min	
1 (2009)	4–15	9.7	10.5	5
1(2010)	8–14	12.1	21.5	13
1 (2011)	4–24	15.9	23.5	5
10 (2011)	4–38	7.5	7.7	0
1 (2012)	3–27	8.5	10.4	3
2 (2012)	6–35	11.8	13.1	2
3 (2012)	5–35	16.5	19.9	5
4 (2012)	4–25	8.2	8.4	0
5 (2012)	4–30	11.6	11.7	0
7 (2012)	2–26	15.1	19.8	2
8 (2012)	4–25	5.4	6.0	0
1 (2013)	4–24	9.2	10.0	1
3 (2014)	5–35	17.0	20.9	8
13 (2014)	5–24	9.0	10.8	4
14 (2014)	3–32	3.0	3.0	0

**Figure 16.** Vertical profiles of the diapycnic diffusion coefficient K_z averaged during the tidal cycle from measurements made: (a) at polygon 1 in summer, 2011; (b) at polygon 4 in summer, 2012; (c) at polygon 7 in summer, 2012. Solid (dashed) lines show calculations made with a 2 h (2 min) time step.

4. Discussion

In this work, we have attempted to build a most detailed picture of ISW properties in the White Sea, based on an analysis of satellite and field observations, including the definition of the main ISW hot-spots, their horizontal (from SAR data) and vertical (from in

situ data) properties, the peculiarities of generation and kinematics in the main hot-spot region; and assess a basin-wide influence on the water mixing.

As obtained, the spatial and kinematic properties of the observed ISWs are linked to water depth/distance from the generation regions, with larger wave trains and higher propagation speeds being observed over the deeper regions. The maximal values of ISW propagation speeds attain ~ 1 m/s and correspond to the largest wave trains observed in the deep Basin.

For the summer of 2010, the basin-wide ISW propagation speeds were estimated from the distance between the consequent wave trains seen in a single SAR image, assuming their generation once per M_2 tidal period. Obviously, the true periodicity of ISW generation is not well-known, though it is natural to suppose a semidiurnal periodicity in the region where the M_2 is the dominant tidal component. However, some of the observed ISWs could be generated twice more frequently, e.g., in the transcritical regime during the acceleration and deceleration phases of the ebb tidal current [25,34]. For such waves, the resulting propagation speeds would be two times higher than the obtained values.

Nevertheless, the obtained ISW propagation speeds of 0.1–1 m/s seem to be quite realistic. First, this was because a similar range of values (0.1–0.2 m/s) was derived from sequential SAR observations. Second, because in some regions, the background currents are very strong, up to 0.8–1 m/s. The superposition of the current velocities with the phase speed of ISWs can easily result in the instant propagation speeds of ISW signatures in the order of 1 m/s.

A detailed analysis of the ISW field in the SW Gorlo Strait, the major hot spot of ISW generation in the sea, from concurrent satellite and field observations, showed rather complex ISW signatures resulting from multiple generations of sources, due to a local complex bottom. Here, the observed ISW patterns fit the bathymetry features well, assuming their generation via the common lee-wave mechanism. Measurements at a 24 h station revealed pronounced isopycnal depressions of 5–8 m due to passing of high-frequency ISW trains with periods of 10–20 min. The most intense waves were observed during the flood, while some others occurred during the flood/ebb slackening and reversal, similarly as reported in [41].

Unfortunately, concurrent field measurements were made southeast of the large semi-circular ISW pattern depicted in Figure 13a. These wave trains have rather strong radar signal contrast [19], and appear to be the intense high-amplitude ISWs. However, historical field campaigns [8,10,13–15] were not able to catch them and prove this, and it remains a subject for future studies.

An analysis of sequential Radarsat-1/-2 images over the SW GS enabled us to directly assess the propagation speeds of consequent ISW trains traveling to different directions. As obtained, estimates derived from the two data sources—sequential images and interpacket distances (from a single image), agreed quite well, with the latter providing somewhat higher values. However, as the literature suggests, sequential observations usually provide more accurate propagation speed values [32,35,37].

Based on well-expressed vertical density stratification at polygon 7 in the SW GS, with a single major peak of buoyancy frequency in the surface layer, we assumed that a two-layer approximation could be satisfactorily applied to theoretically assess the phase speed of the observed ISWs. However, the comparison of SAR-based estimates with those derived from theory showed that the latter were about two times higher than the observed values. Such a difference is presumably related to the fact that the two-layer model does not account for background currents that can effectively modulate the propagation speed of ISWs during the tidal cycle. Thus, SAR-based estimates appear to be more useful for practical applications where the actual propagation speeds of ISWs are needed.

5. Conclusions

The results of internal wave identification in the satellite SAR data of various spatial resolution clearly show that the latter is important in such studies. The number of identified

ISW signatures was nearly similar for medium- (150 m) and high-resolution (6–25 m) SAR data, though the number of the latter was 2.5 times smaller during the same time period.

The major hot-spot of permanent ISW generation in the White Sea characterized by the highest ISW probability is the region in the SW Gorlo Strait, where a superposition of two strong fronts (the Dvina Front and the Gorlo Front), active tidal dynamics, and complex topography takes place. Other regions of pronounced ISW activity are found in straits near the Solovetsky Islands and in the NW Onega Bay, all characterized by strong tides and the presence of frontal zones. An important role of frontal zones in favoring the stratification conditions for ISW generation is also demonstrated from a very good correspondence between the MODIS-based SST gradients and the ISW probability field in the White Sea.

An analysis of field measurements at polygons shows a significant variability of thermohaline fields in two ranges of periodicity—semidiurnal (12.4 h) and high-frequency (tens of minutes). The former is primarily observed in the deep Basin, away from rough topography and frontal zones, while the latter or superposition of the two is predominantly found over the shallow coastal regions and near frontal zones.

Field measurements at polygons reveal that the most intense ISWs are recorded in the vicinity of frontal zones, where their heights often exceed the upper mixed layer thickness. The most intense ISWs were detected on the western sea shelf. Here, packets of strongly nonlinear internal waves with maximal heights of 14–17 m and periods of 10–20 min were repeatedly recorded immediately after the flood–ebb transition, when the tidal current slackened and gradually changed its direction.

An analysis of diapycnic diffusion coefficient values obtained at polygons across the White Sea showed that enhanced K_z values in the pycnocline are registered mostly at polygons near frontal zones where intense ISWs (higher than 5 m) are regularly present. This result is in agreement with previous findings from other Arctic regions showing enhanced turbulent kinetic energy dissipation rates associated with the propagation and breaking of nonlinear ISWs [41,42]. In our work, the influence of ISWs on vertical mixing was assessed in a simplified manner, while future studies should definitely address it in a more sophisticated way using contemporary microstructure observations.

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