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Anticyclonic Eddies in the northeastern Japan/East Sea: simulation, observation and formation mechanisms

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Abstract

The Japan/East Sea is a semi-enclosed basin divided by the Subpolar Front into subtropical and subarctic regions. While eddy activity has been extensively studied in the subtropical zone and the northwestern subarctic area, this paper presents a detailed investigation of mesoscale eddies in the previously understudied northeastern region, west of Hokkaido Island. The findings show that this area is characterized by the regular formation of both mesoscale anticyclonic (AE) and cyclonic eddies (CE). The characteristics and evolution of an AE observed in this region during the summer and fall of 2015 are examined using the three-dimensional Regional Ocean Modeling System (ROMS) with a horizontal resolution of 2 km. The simulation results reveal the eddy's three-dimensional structure, including its formation, movement, and dissipation. Analysis of Ertel's potential vorticity suggests that inertial instability is the primary energy source driving the growth of this eddy. Model outputs are validated against surface current patterns derived from satellite altimetry and multi-year CTD measurements, which allow for current calculations at various depths using a linear diagnostic model. Additionally, surface drifter trajectories from multiple years provide clear evidence of AEs, consistently captured in the altimetry-derived, diagnostic, and ROMS-simulated velocity fields within the research area.

Keywords: northeastern Japan/East Sea; eddy activity; ROMS simulation; CTD measurements; Ertel's potential vorticity

1. Introduction

The Japan/East Sea (JES) is a semi-enclosed marginal sea bordered by the Russian Far East, the Korean Peninsula, and the Japanese islands. It is connected to the adjacent basins through four straits: the Tsugaru Strait (to the Pacific Ocean), the Tsushima (Korea) Strait (to the East China Sea), and the Soya (La Pérouse) and Nevelskoy straits (to the Sea of Okhotsk) (Fig. 1a). The basin is bisected by the Subpolar Front, a quasi-stationary boundary that separates the warm, saline subtropical waters in the south from the cold, fresh subarctic waters in the north (Park et al., 2004; Talley et al., 2006; Chang et al., 2016; Wagawa et al., 2020). The Tsushima Warm Current, a northeastward extension of the

Kuroshio, enters the JES through the Tsushima Strait and forms a complex, seasonally dependent system of two to three branches, including one that flows along the western coast of Japan (Watanabe et al., 2006; Yabe et al., 2021).

The JES is globally recognized as one of the most eddy-rich marginal seas. Over the past two decades, mesoscale eddy activity has been extensively investigated in its subtropical sector (Chang et al., 2004; Shin, 2009; Lee & Niiler, 2005, 2010) and, separately, in the northwestern subarctic domain (Lobanov et al., 2001; Talley et al., 2004; Prants et al., 2011, 2015; Ponomarev et al., 2018; Lee et al., 2019; Udalov et al., 2025). Long-term altimetry-based tracking has revealed areas of elevated probability for long-lived AEs and CEs across much of the basin (Udalov et al., 2025). Despite this wealth of knowledge, the northeastern sector of the JES, the region west of Hokkaido Island, remains a gap in our understanding of mesoscale dynamics. Systematic spatiotemporal analyses of eddy activity in this area are virtually absent.

First, the northeastern JES is not a passive periphery but a dynamic crossroads where the Subpolar Front impinges upon a complex bathymetric setting. The region is characterized by abrupt depth changes, seamounts, and ridge-like features that are known to constrain and amplify mesoscale instability. Unlike freely propagating open-ocean eddies, the vortices in this area appear to be topographically anchored, recurring in preferred locations with remarkable regularity (Udalov et al., 2025). This topographic control offers a unique natural laboratory to investigate the fundamental mechanisms of eddy generation, maintenance, and decay, processes that remain poorly understood in the presence of steep bottom relief.

Second, and perhaps most importantly, this region serves as a critical gateway for cross-frontal exchange between subtropical and subarctic water masses. In the northeastern JES, AEs (particularly those shed from the Tsushima Warm Current system) are hypothesized to entrain warm, saline subtropical waters and inject them deep into the subarctic domain. Such transport directly modulates the regional thermohaline structure, stratification, and even the formation of intermediate water masses. Yet, the magnitude, frequency, and mechanisms of this eddy-mediated intergyre exchange have never been quantified in this sector.

Third, the northeastern JES is biologically productive and commercially valuable fishing grounds in the Northwest Pacific (e.g., Ishizaka et al., 2019; Wang et al., 2022). Mesoscale eddies exert a profound influence on marine ecosystems by driving localized upwelling, enhancing vertical nutrient supply, and creating heterogeneous habitats that aggregate plankton and fish (Park et al., 2016). Understanding the spatial and temporal distribution of eddies in this region is therefore not merely an academic exercise, it has direct implications for the sustainable use of marine living resources. Without a robust physical framework describing eddy activity, any attempt to link physical forcing to biological response in this area remains speculative.

In light of these motivations, this study presents a multi-method investigation of mesoscale eddies in the previously understudied region west of Hokkaido. We combine high-resolution numerical simulations with the ROMS, satellite altimetry, multi-year CTD casts, and diagnostic current calculations to achieve a multi-faceted view of eddy structure, life cycle, and dynamics. As a representative case, we examine in detail an AE observed during the summer and autumn of 2015 (AE2015), which is robustly captured across all independent datasets.

The paper is structured as follows. Section 2 describes the ROMS setup, the diagnostic modeling methodology based on CTD data, and the altimetry-based eddy detection algorithm. In Section 3.1, we identify hotspots of AE and CE activity in the study region and examine the vertical distribution of eddy centers. Section 3.2 presents examples of the background circulation derived from diagnostic simulations. Sections 3.3 and 3.4 detail the lifecycle of AE2015 and its three-dimensional thermohaline structure. Section 4 discusses the dynamical mechanisms, particularly the role of inertial and baroclinic instabilities, and validates the simulation results against independent observational datasets. Section 5 summarizes the key findings and outlines their implications for regional oceanography and beyond.

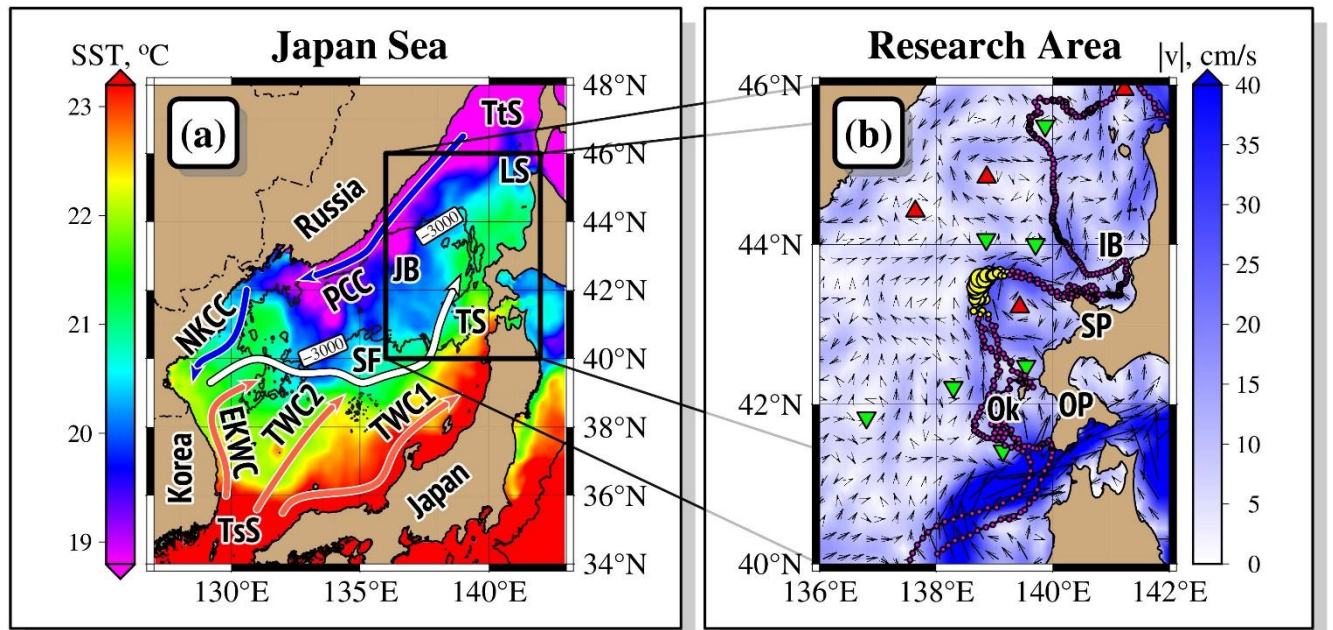


Figure 1. a) Map of sea surface temperature in the Japan/East Sea with a superimposed schematic of the main currents. Abbreviations: JB – Japan Basin; TsS – Tsushima (or Korea) Strait; TS – Tsugaru Strait; LS – La Pérouse (or Soya) Strait; TtS – Tatar Strait; EKWC – East Korean Warm Current; NKCC – North Korean Cold Current; PCC – Primorsky Cold Current; TWC1 and TWC2 – the coastal and offshore branches of the Tsushima Warm Current, respectively; SF – Subpolar Front. The 3000 meters bathymetry contour is shown in black. b) Current pattern based on AVISO data in the research area on October 13, 2015, showing the location of eddy AE2015 centered at (43°N, 139.5°E). The long trajectory of a drifter temporarily trapped by the eddy is shown in pink.

Abbreviations: Ok – Okushiri Island; OP – Oshima Peninsula; IB – Ishikari Bay; SP – Shakotan Peninsula. Red and green triangles indicated elliptic points (centers) of anticyclonic and cyclonic mesoscale eddies, respectively

2. Methods

2.1. ROMS modelling

For a detailed study of eddies generated in the research area, the regional ROMS circulation model was applied to the central and northern parts of JES. The model domain extends from 126.5°E to 142.2°E and from 38°N to 52°N. The horizontal grid spacing is 2 km, while the vertical grid comprises 48 s-layers with enhanced resolution in the surface and bottom layers. The bottom topography used in the model is based on the ETOPO2v2 global relief dataset. The surface boundary conditions for radiative balance and freshwater flux were derived from precipitation, atmospheric pressure, solar radiation flux, air temperature, air humidity, and cloud cover data taken from the NCEP-DOE AMIP-II meteorological reanalysis (Kanamitsu et al., 2002). Wind forcing was obtained from satellite scatterometer measurements provided by the Global Ocean Wind Analysis – Blended ASCAT (Bentamy et al., 2012). Sea surface temperature was reconstructed from the Global Ocean OSTIA Sea Surface Temperature and Sea Ice Analysis satellite product (Donlon et al., 2012). To adjust surface salinity, monthly climatological data from the World Ocean Atlas 2023 (Reagan et al., 2024) were employed. Open boundary conditions (southern and eastern boundaries) were prescribed using daily-mean data from the Japan Coastal Ocean Predictability Experiment (JCOPE2) reanalysis with a horizontal resolution of 1/12° (Miyazawa et al., 2017). The simulation period covers 1 January 2014 to 31 December 2016, and the model output is recorded every hour.

2.2. Diagnostic modelling

For diagnostic calculations, we used a linear diagnostic model, the theoretical basis of which was formulated by Sarkisyan and Sündermann (2009). The model includes initial simplified equations of motion (advection and horizontal turbulent exchange are excluded) and continuity equation in the hydrostatic and Boussinesq approximations. The boundary conditions at the sea surface are wind friction stress and “rigid lid” with a sticking at the bottom. After transforming the initial equations, the resulting set of partial differential equations includes the elliptic sea level equation, the continuity equation and explicit formulas for calculating the horizontal components of the current vectors. When numerically solving the elliptic sea level equation, the method described by Kochergin et al., (1991) was applied. The method consists in reducing the Neumann problem to the Dirichlet problem, i.e., the level surface is calculated by successive approximations at the nodes of the closed boundary of the computational domain, and then at the internal points of the domain. In the numerical approximation of the second derivatives on horizontal coordinates, the scheme of central differences is used. Successive approximations of the obtained system of linear equations for the level values in the grid nodes were

performed using the Gauss-Seidel method. Central differences were used to discretize the equations of horizontal components of the velocity vector. The complete system of numerical equations has been then solved by the iterative method.

The difference of the proposed model from the linear diagnostic Sarkisian's D1 model is in a more correct numerical scheme of solving the boundary condition problem for the sea level. This allows to agree the flow field with the relief and shoreline by the iterative method. The second advantage is the exclusion of the overestimated influence of sharp bottom relief gradients. Such a model has been successfully applied to calculate currents from CTD measurements (Fayman & Ponomarev, 2018) and from climatic data (Fayman, 2015).

The temperature and salinity fields were taken from the CTD measurements conducted by various scientific organizations from different countries in the central and northern parts of the JES from 1960 to 2024. These CTD data are available in the World Ocean Database 2023 (WOD23) (Mishonov et al., 2024). CTD measurements were grouped separately for different months of each year for the above period. Then, those years, where the grid of CTD stations was sufficiently dense, were selected to reconstruct the sea water temperature field.

The number of salinity measurements was several times smaller than the number of temperature measurements and all available measurements were made in the upper 500 m layer. Therefore, the CTD measurements from WOD23 were combined with World Ocean Atlas 2023 (WOA 2023) climate data (Reagan et al., 2023). Salinity from the surface to 1500 m and temperature from 500 m to 1500 m were taken from monthly mean data from WOA 2023. Salinity and temperature from 1500 m to bottom ~~depth~~ were taken from seasonal average data from WOA 2023. Temperature from the surface to 500 m was taken from the World Ocean Database 2023 for specific years, namely: 1967, 1971, 1972, 1974, 1977, 1978, 1979, 1980 - 1986, 1989, 1990, and 1999. Thus, only temperature in the upper 500 m layer was taken from WOD23, while the temperature at the underlying horizons and all salinity values were taken from the WOA 2023 climate data. The modeling domain was taken within 41°N - 45°N, 137°E - 142°E with the 0.05° horizontal grid spacing. The depth fields were taken from the 2-minute Gridded Global Relief Data (ETOPO2) v2 (2006). All data were interpolated onto the computational grid using the Linear Kriging method.

2.3. Eddy detection algorithm and satellite data

The Angular Momentum Eddy Detection and tracking Algorithm (AMEDA), with the settings recommended by Le Vu et al. (2018), was applied to detect mesoscale eddies. The algorithm identifies eddy centers based on the extrema of the local normalized angular momentum. To compute the occurrence frequencies of AEs and CEs, we overlaid a regular $0.1^\circ \times 0.1^\circ$ grid onto the research area. For each grid point, we counted the number of days on which it was located inside any detected eddy contour over the entire observation period from January 1, 1993 to January 1, 2025. This count was

normalized by the total number of days to obtain the relative occurrence frequency. The resulting values are color-coded on geographic maps, highlighting regions where mesoscale eddies preferentially form and circulate. The 0.1° grid was chosen for post-processing and statistical analysis as it provides an optimal balance between map detail and computational efficiency for the multi-year dataset.

The absolute dynamic topography maps are produced by Salto/Duacs (aviso.altimetry.fr) and distributed by the Copernicus Marine Environment Monitoring Service (CMEMS) with a spatial resolution of $1/8^\circ \times 1/8^\circ$ and a daily time step. The trajectory data for surface floats were obtained from the website: <https://www.aoml.noaa.gov/phod/gdp/data.php>.

3. Results

3.1. Eddy activity in the research area

The modeling domain is presented in Fig. 1b. The key actor in this region is the branch of the Tsushima Warm Current (see the current schematic in Fig. 1a). Upon reaching the western coast of Hokkaido Island, this current separates into two branches: a coastal branch, which continues toward the Soya Strait, and an offshore branch, which turns in the northwest direction and follows toward the coast of Primorsky Krai (Russia) over the northern boundary of the Japan Basin (Fig. 1a).

The plots with calculated occurrence frequencies of AEs and CEs (see Sec. 2.3) with lifespans exceeding 30 days are shown in Figs. 2a and b, respectively. These figures demonstrate the domains where mesoscale eddies have been regularly observed in the period from January 1st, 1993 to January 1st, 2025. The values shown in this figure represent the eddy occurrence frequency, the fraction of time that mesoscale eddies were present at each location over the 32-year period beginning in 1993. The persistent long-term signals observed here may originate from two sources: long-lived stationary eddies that remain in the area, and numerous transient eddies that pass through this location in succession. AEs appear regularly over the central Japan Basin, along its northern boundary and to the west of the Ishikari Bay (Fig. 2a). As for CEs, they appear regularly along the continental slope along the continental coast (Fig. 2b) and in the eastern part of the Japan Basin. CE hotspots are also observed at the locations with AE hotspots, in Ishikari Bay and to the west of the Shakotan Peninsula, but the CE occurrence frequency there is smaller as compared with that of AEs.

When circulating, both AEs and CEs exhibit an inhomogeneous distribution over bathymetry in the region. Figure 3 presents the daily positions of AE and CE centers for the total observation period. The majority of eddies of both polarities are observed to circulate either along the coast or within the deepest portions of the Japan Basin. Additionally, numerous AEs are found circulating over regions with depths of 1 km, located to the north of the Japan Basin, as indicated in Fig. 2a.

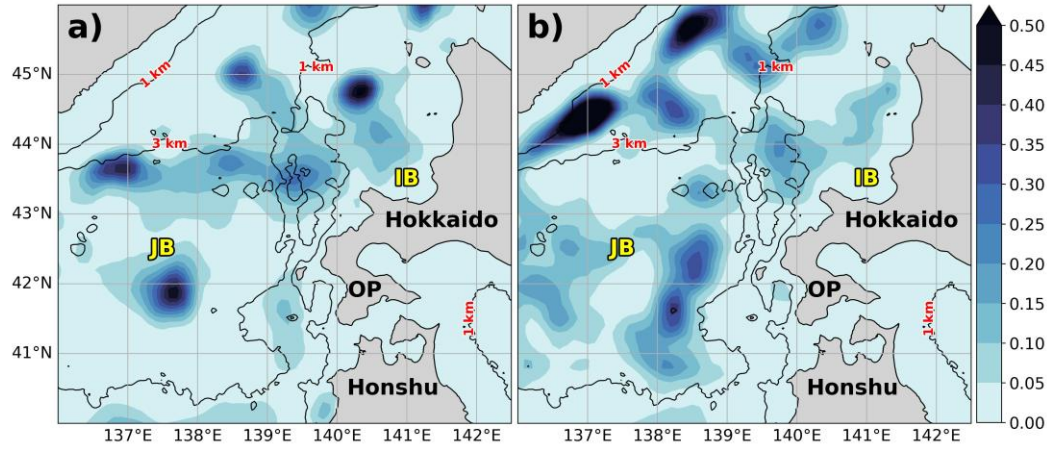


Figure 2. Altimetry-based occurrence frequency of the detected mesoscale a) anticyclones and b) cyclones with the lifespan > 30 days in the research area from January 1, 1993, to January 1, 2025. JB, OP and IB stand for the Japan Basin, Oshima Peninsula and Ishikari Bay

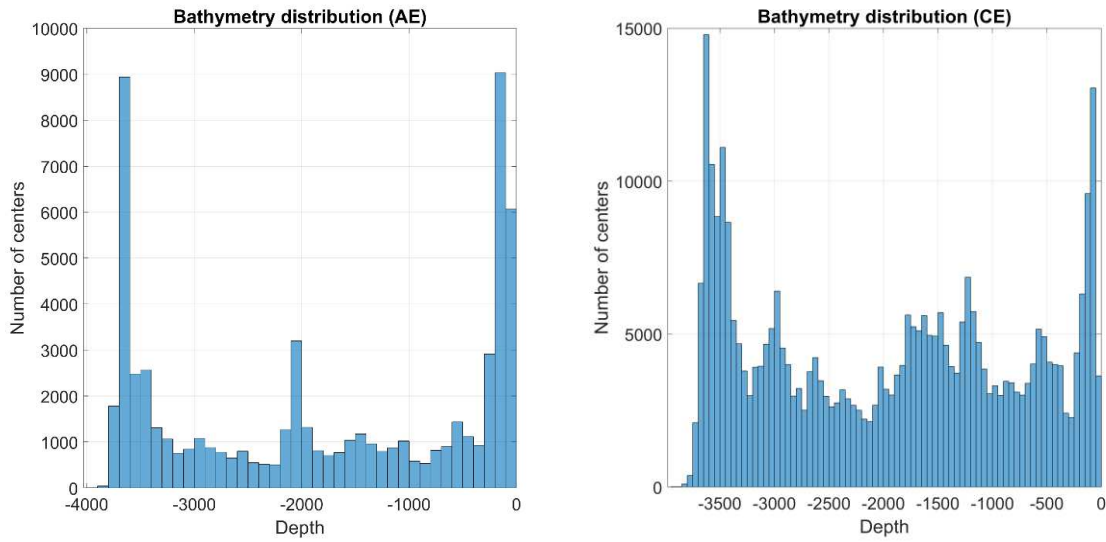


Figure 3. Distribution of the daily center locations of (a) anticyclones (AEs) and (b) cyclones (CEs) over bathymetry in the region shown in Fig. 2

3.2. Background circulation based on diagnostic simulation

The maps in Fig. 2 highlight previously unexplored domains in the northeastern JES, where surface-intensified, long-lived mesoscale eddies have not been studied before. In particular, a mesoscale AE is regularly observed to the southwest of the Shakotan Peninsula and Ishikari Bay, as reflected in the AVISO velocity field (Fig. 1b) and the trajectories of surface drifters (Fig. 4). Additional episodes of drifters trapped by this eddy are presented in the Supplementary material (Fig. S1). In this section, we present the results of diagnostic calculations (see Sec. 2.2) based on available CTD measurements from various years. The input data for the diagnostic model consist of the depth field, sea surface wind, and three-dimensional fields of temperature and salinity on a horizontally uniform grid. Typical circulation patterns in the research area featuring pronounced AEs, are shown in Fig. 5.

The diagnostic calculations in Fig. 5 show pronounced AEs with diameters in the range of 70–100 km west of Ishikari Bay in September of 1974, 1977, 1980, 1981–1983, 1990, and 1999 (see also Fig. S2). In 1971, 1972, 1978, 1985, 1986, and 1989, AEs were less pronounced, with diameters not exceeding 50 km. In the remaining years with available CTD data, specifically in 1967, 1979, and 1984, no AEs were detected in the research area. Together with the AVISO and drifter data, the diagnostic calculations clearly confirm the regular presence of intense AEs in this region.

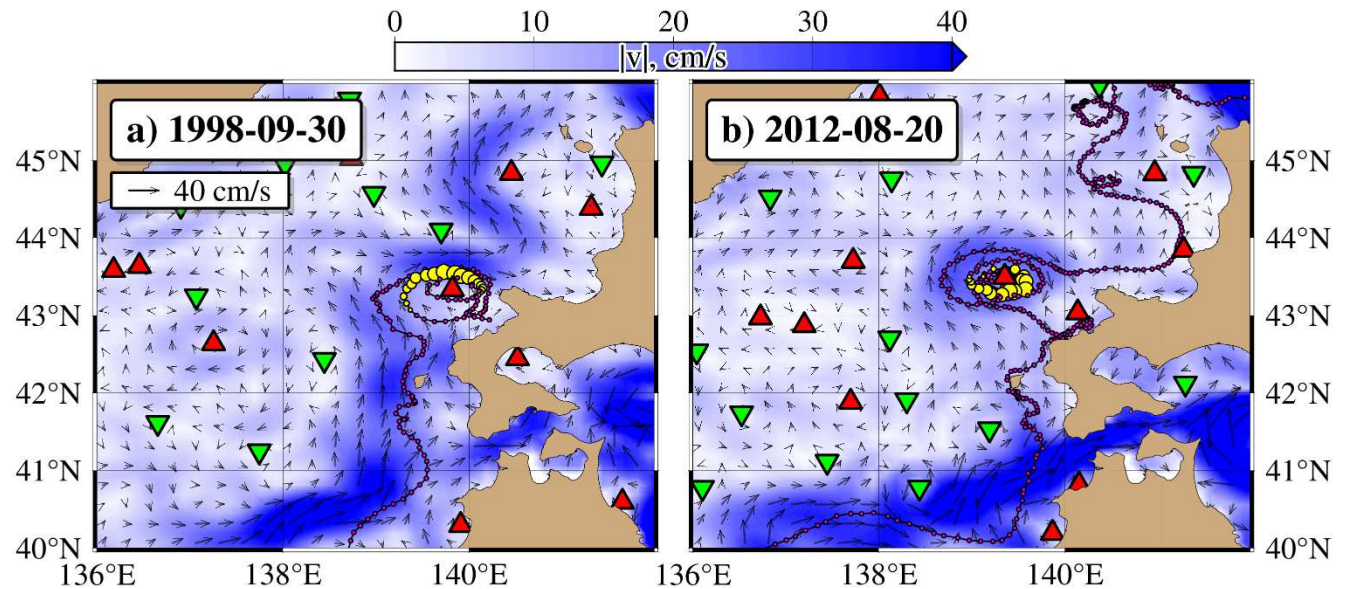


Figure 4. Trajectories of northward-moving drifters temporarily trapped by an anticyclonic eddy west of the Shakotan Peninsula and Ishikari Bay, with the AVISO velocity field superimposed on (a) 30 September 1998 and (b) 20 August 2012 (see also Fig. S1). Red and green triangles indicated elliptic points (centers) of anticyclonic and cyclonic mesoscale eddies, respectively

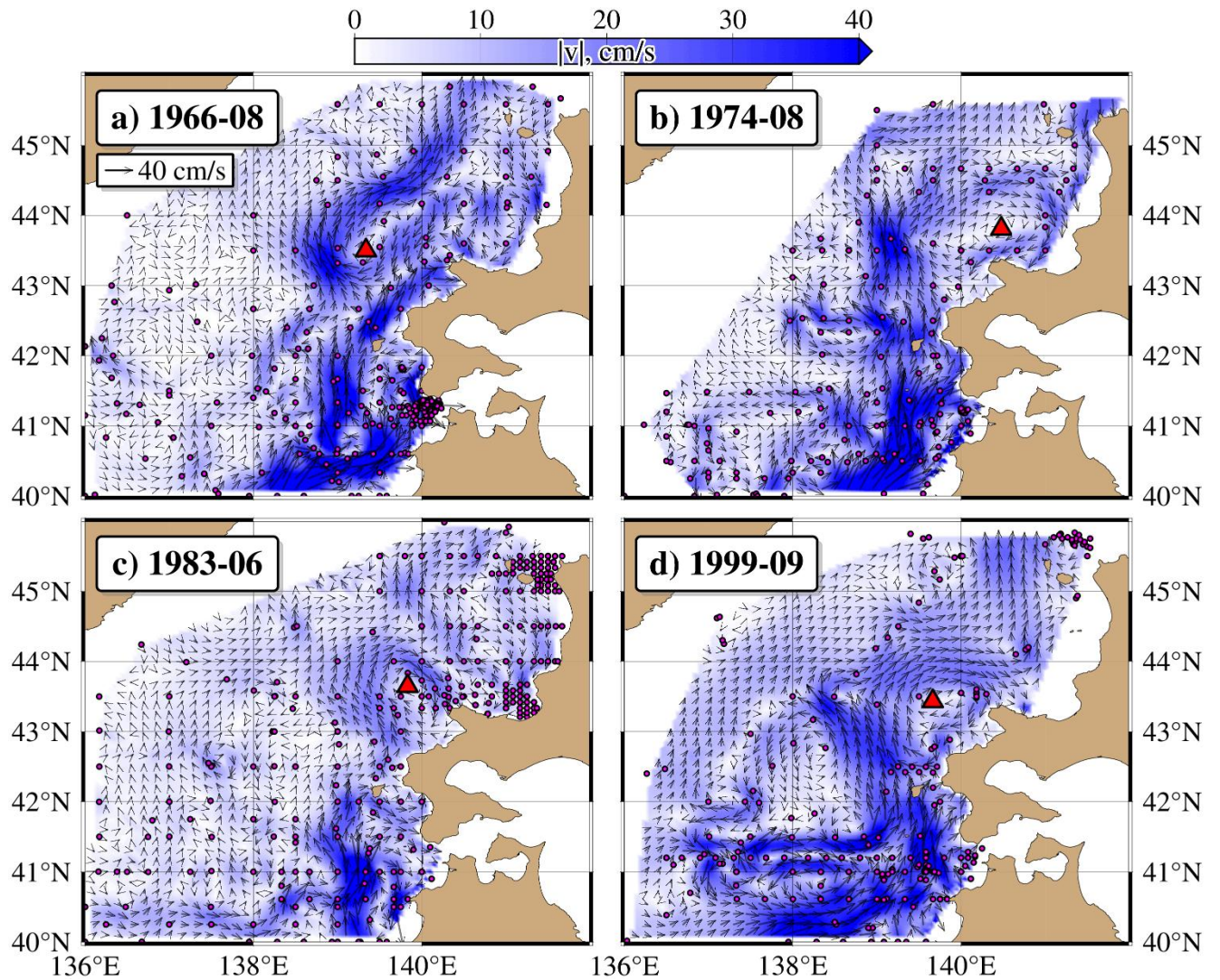


Figure 5. Surface currents obtained from diagnostic calculations for specific years show the presence of a mesoscale anticyclonic eddy in the research area, with its center indicated by a red triangle (see also Fig. S2). Dots indicate CTD stations

3.3. The anticyclonic eddy under study: life cycle and vertical structure

3.3.1. The life cycle of AE2015

The ROMS-based simulations for 2015 reproduced a circulation pattern consistent with the schematic of currents in the study region (see Fig. 1a). In late July, the model captures the Tsushima Warm Current flowing along the western coast of Hokkaido Island, with velocities reaching up to 50 cm/s at both the surface and at 50 m depth (Fig. 6). This current splits into two branches: a coastal branch in the near-surface layer that flows along the shoreline as far as the Soya Strait, and an offshore branch that turns northwestward and runs along the northern boundary of the Japan Basin toward the continent (Watanabe et al., 2006; Yabe et al., 2021). The life cycle of the studied AE comprises several phases, each lasting about one month. Starting in late July, an AE with a diameter of up to 50 km forms to the south of the Shakotan Peninsula (Fig. 6). This eddy is vertically coherent from the surface to the seabed. By early September, it migrates northwestward. At the surface, its diameter extends

approximately 100 km zonally and 50 km meridionally (Fig. 7). Throughout September, AE2015 continues to grow by merging with smaller AEs that form in the bays south and north of the Shakotan Peninsula. In late September to early October, AE2015 remains stationary but expands to a diameter of about 150 km. However, at depth of 200 m and below, its diameter does not exceed 50 km. During October, the eddy drifts further northwestward and enlarges to a diameter of up to 200 km (Fig. 8). Below 200 m, its diameter remains under 100 km, and at depths of 1 km and greater, the eddy is no longer detectable. In late October, AE2015 merges with a topographically constrained AE centered over a seamount at approximately (44.2°N, 138.3°E).

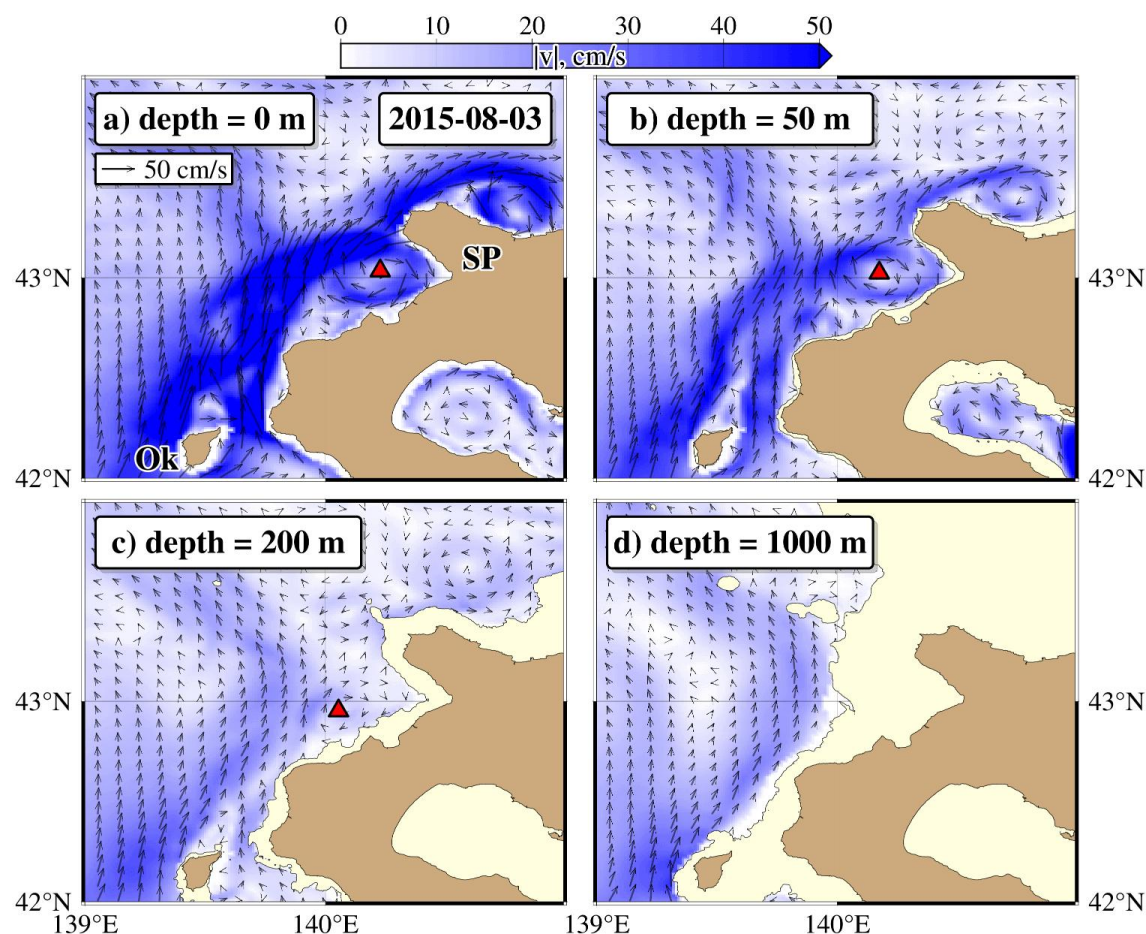


Figure 6. Late-July generation of eddy AE2015 in the ROMS-simulated velocity field. The red triangle marks the eddy's center at different depths. Ok – Okushiri Island, SP – Shakotan Peninsula

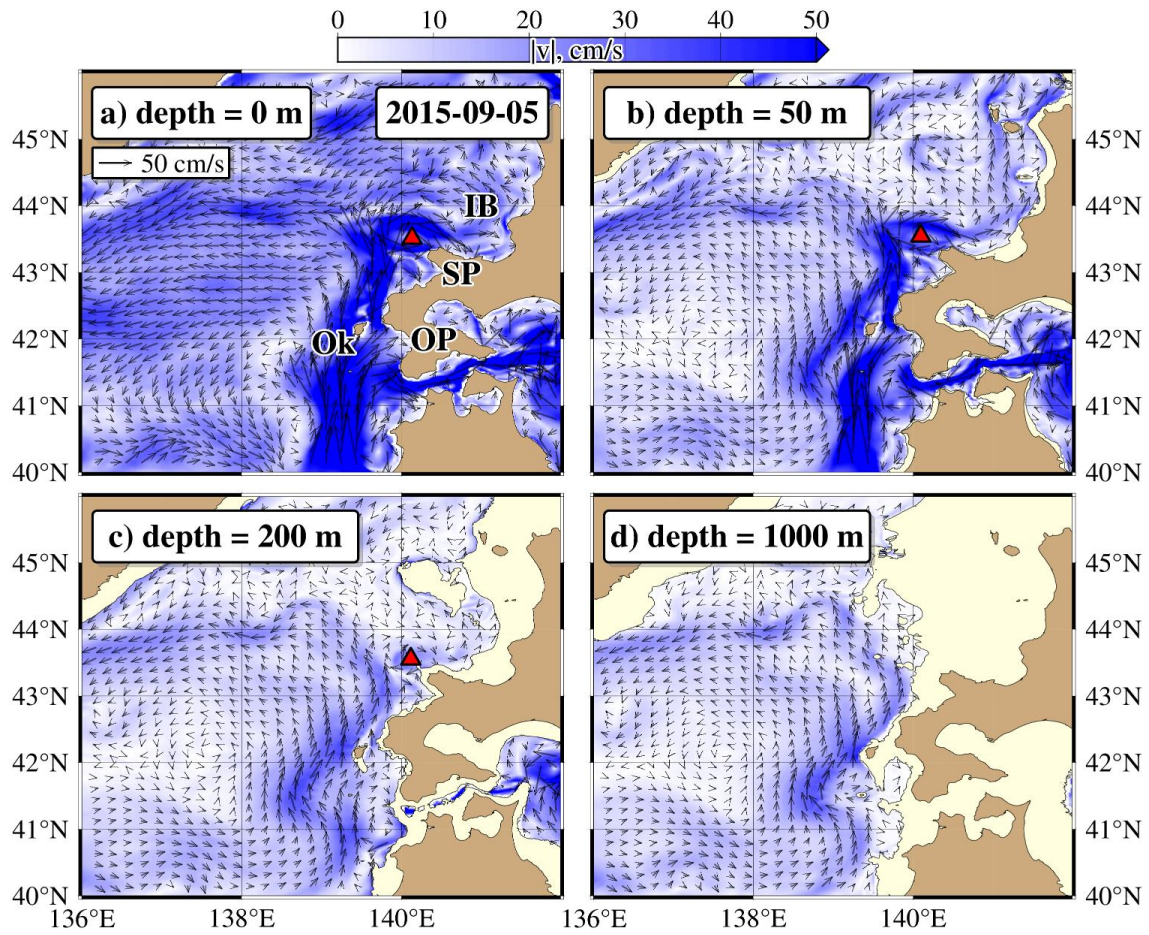


Figure 7. AE2015 in September. Ok – Okushiri Island, OP – Oshima Peninsula, IB – Ishikari Bay, SP – Shakotan Peninsula. The red triangle marks the eddy's center

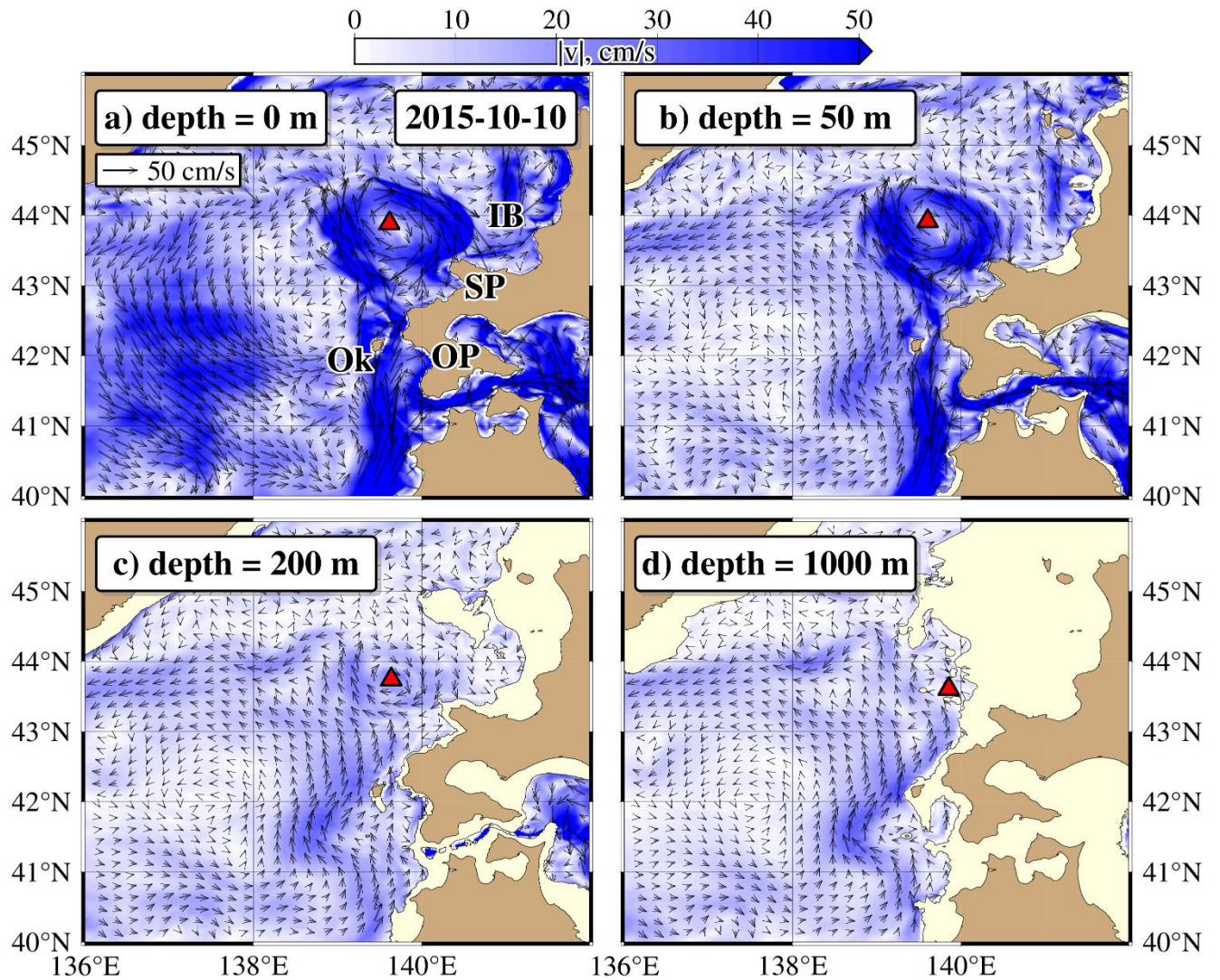


Figure 8. The mature phase of AE2015 in October. The red triangle marks the eddy's center

3.3.2. Eddy vertical structure and properties

To study the vertical structure of the studied eddy, the following was done. The three-dimensional structure of the eddy was visualized at various stages of its life: first stage - the formation of eddies (Fig. 9a), second stage - development (Figures 9b, 9c), third stage - movement (Fig. 9d). Vertical sections of temperature through the eddy center for various dates were also made (Fig. S3).

The model results show the following: during the eddy formation period (late August – early September), its diameter reaches 50 km (Fig. 9a). Vertical sections show a characteristic temperature structure, when the eddy manifests itself in the form of isotherms with a parabolic shape. The upper layer warms up to 22°C. The 10°C isotherm descends in the eddy core to a depth of 100 meters, the 5°C isotherm descends to a depth of 230 m, and the 2°C isotherm descends to a depth of 440 m (Fig. S3a). By mid-September, the eddy size increases to 100 km in diameter (Fig. 9b), and the water temperature in the upper layer drops to 20°C. The 10°C, 5°C, and 2°C isotherms occur at depths of 100 m, 200 m, and 380 m, respectively (Fig. S3b). In early October, the eddy increases in diameter to 150 km (Fig. 9c),

and the water temperature in the upper layer drops to 18°C. The 10°C, 5°C, and 2°C isotherms occur at depths of 120 m, 240 m, and 420 m, respectively (Fig. S3c).

The vertical sections of temperature based on CTD measurements were made for comparison with the model results. The choice of vertical sections was determined by the following conditions. First conditions - the diagnostic current patterns calculated based on CTD data should show a pronounced AE in the study area. Second conditions - CTD section should pass through the eddy center or should be in close proximity to it. Third conditions - the CTD measurements should be made in period August - October, i.e., during the study period. Such conditions apply to measurements conducted in August 1983, September 1974, 1981, and October 1983 (Fig. S4). Almost all measurements were conducted by the Hydrographic Division of Japan Maritime Self-Defense Force. In October 1983, measurements were made by the Hokkaido National Fisheries Research Institute. The results of visual comparison of vertical sections showed that the model results and measurement results are in good agreement with each other (Fig. S3, S4).

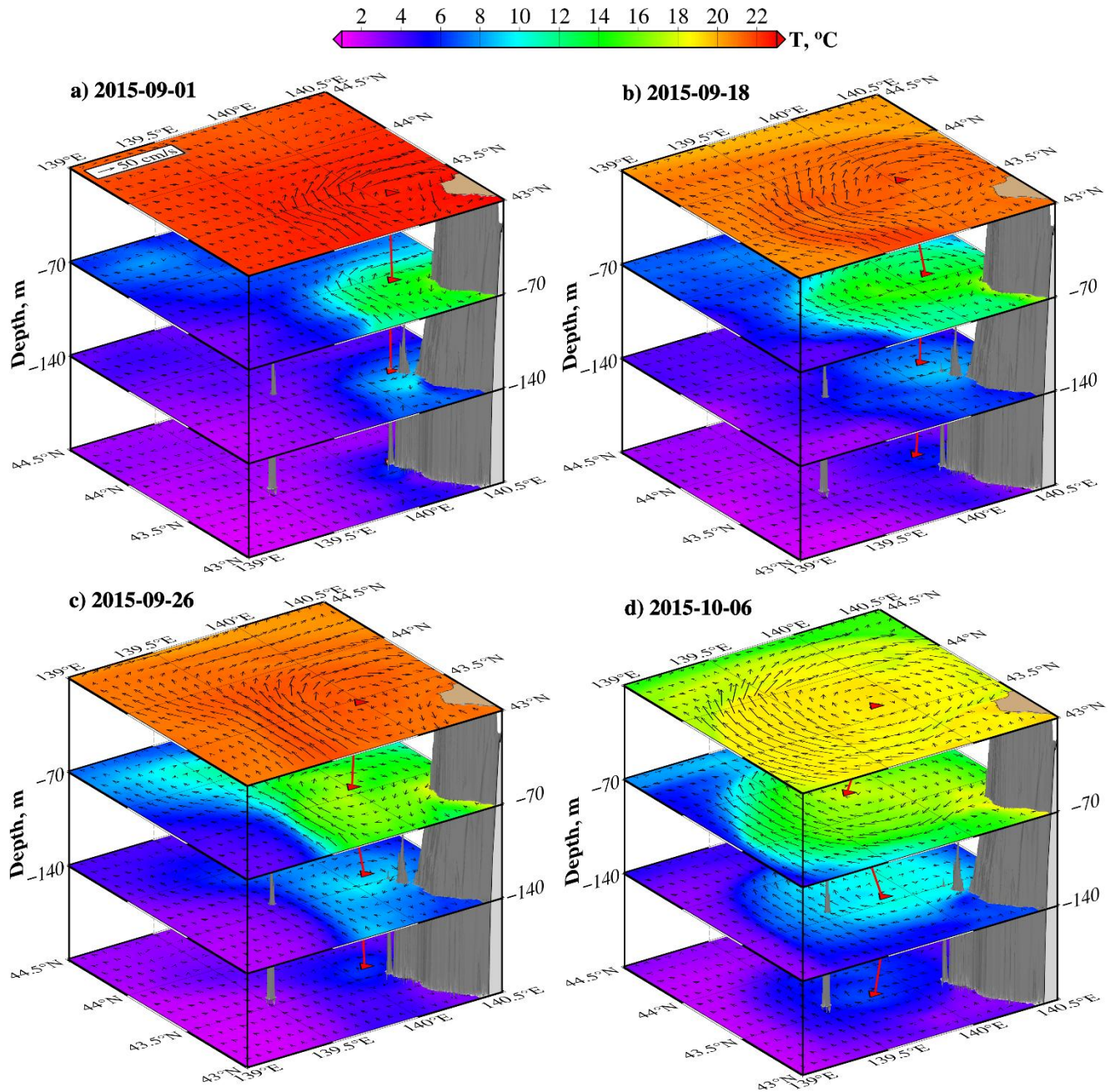


Figure 9. Vertical thermohaline structure during the period of eddy generation (September 1), development (September 18 and 26), and movement stages (October 6). The ROMS-simulated velocity field at different depths is shown by arrows. The red triangle marks the eddy's center at different depths

3.4. Estimation of the components of Ertel's potential vorticity

This section assesses the contribution of various components of Ertel's potential vorticity (EPV). Such an assessment is necessary to identify the primary mechanisms driving eddy generation in the research area. The work has been carried out in three stages. At the first stage, eddy kinetic energy (EKE) was calculated at different depth horizons separately for each month of the second half of 2015 in order to precisely delineate the boundaries of eddy-active zones. At the second stage, the contribution of the geostrophic balance to the water circulation was evaluated within these zones. At the third stage, EPV

was computed, and the contributions of individual terms responsible for inertial (barotropic) and symmetric (baroclinic) instabilities were assessed. In our context, the terms ‘inertial’ and ‘symmetric’ instabilities are more appropriate.

EKE was calculated for the surface and at depths of 50 m and 100 m separately for July, August, and September using the well-known formula:

$$EKE = \frac{1}{2} \left((u - \bar{u})^2 + (v - \bar{v})^2 \right) . \quad (1)$$

As a result, a region bounded by (42°N–44°N, 139°E–141°E, black rectangle in Fig. 10) with enhanced EKE was identified.

The contribution of the geostrophic balance to the general circulation was computed using the following algorithm:

1. Based on the ROMS temperature and salinity fields, density was calculated using the equation of state for seawater.
2. Using the seawater density, currents were computed via the dynamic height method

$$\begin{cases} u_{din} = \frac{g}{f \rho_0} \int_H^z \frac{\partial \rho}{\partial y} dz \\ v_{din} = -\frac{g}{f \rho_0} \int_H^z \frac{\partial \rho}{\partial x} dz \end{cases} . \quad (2)$$

3) Contribution of the geostrophic balance was calculated as follows:

$$C = \left(\frac{U_{din}}{U_{ROMS}} \right), \quad U_{ROMS} = \sqrt{u_{ROMS}^2 + v_{ROMS}^2}, \quad U_{din} = \sqrt{u_{din}^2 + v_{din}^2}, \quad (3)$$

where (u_{ROMS}, v_{ROMS}) are the velocity components from the ROMS model, and (u_{din}, v_{din}) are the velocity components derived from the dynamic method. The averaging in formula (3) was performed over the upper 50-meter layer. Calculations showed that the geostrophic contribution to the equations of motion does not exceed 60%.

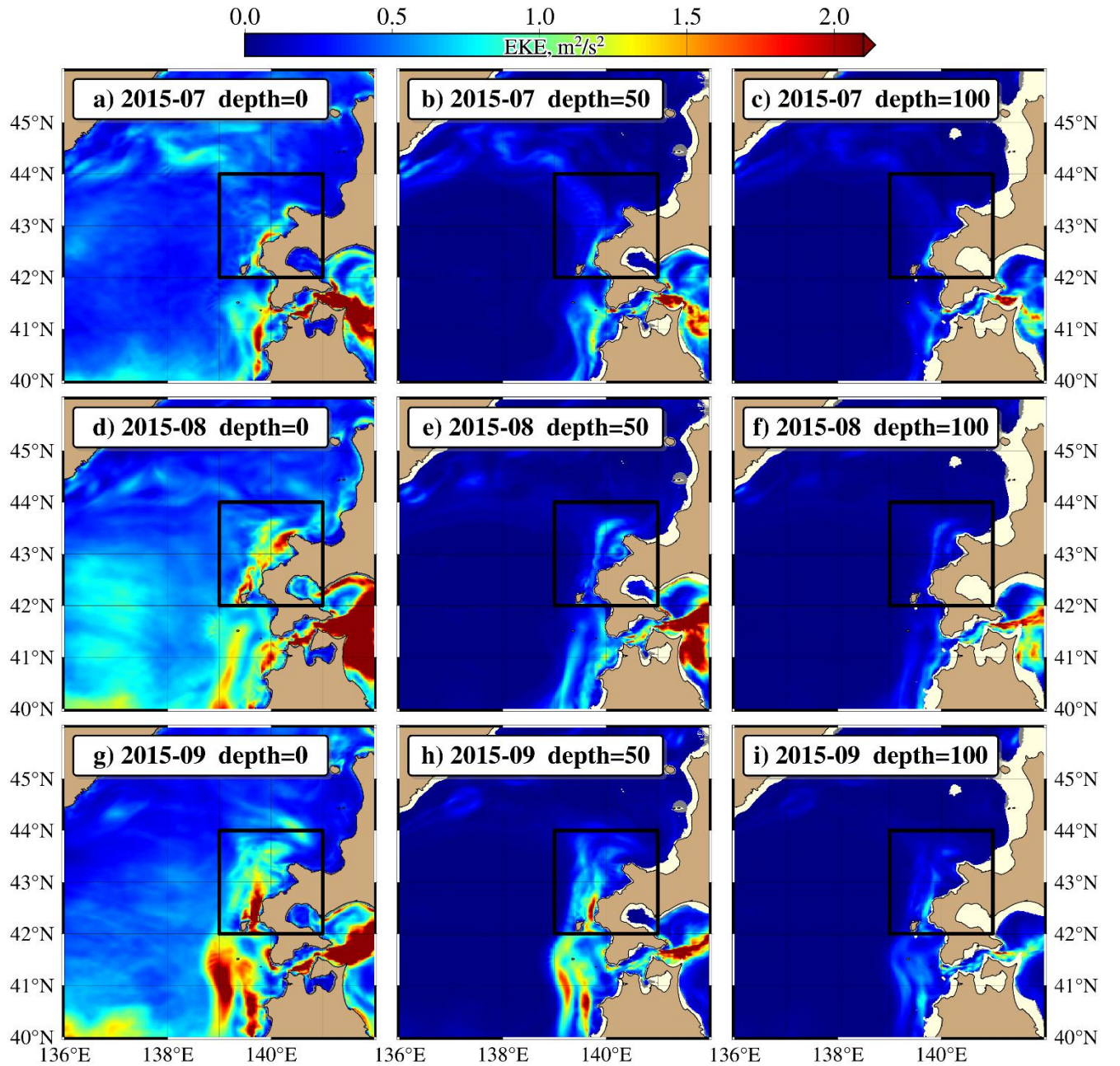


Figure 10. Eddy Kinetic Energy (m^2/s^2) for July, August, and September at the surface, 50 m, and 100 m depth. The rectangular box highlights the research area

Next, the evolution of EPV was computed for the region whose boundaries were identified in the first step (Fig. 10). For a continuously stratified fluid, the EPV formula is (Schubert et al., 2004):

$$q = \frac{(f \cdot \vec{k} + \vec{\nabla} \times \vec{u}) \cdot \vec{\nabla} b}{g}, \quad (4)$$

where $b = -g \frac{\rho}{\rho_0}$ is the buoyancy, f is the Coriolis parameter, ρ is the density, $\vec{\nabla}$ is the nabla operator,

$\vec{u} = (u, v, w)$ are the velocity components, and g is the gravitational acceleration. Formula (4) can be expressed as a sum:

$$q = q_N + q_{bc}, \quad (5)$$

with terms accounting for stratification and absolute vorticity

$$q_N = \frac{1}{g} \left(f + \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) N^2, \quad (6)$$

and for baroclinicity arising from horizontal buoyancy gradients

$$q_{bc} = -\frac{1}{g} \left(\frac{\partial v}{\partial z} \frac{\partial b}{\partial x} - \frac{\partial u}{\partial z} \frac{\partial b}{\partial y} \right). \quad (7)$$

Different types of instabilities occur when EPV (4) takes the sign opposite to that of the Coriolis parameter (i.e., becomes negative in the Northern Hemisphere). This can happen if:

$|q_{bc}| > q_N, \quad q_{bc} < 0$ - symmetric instability;

$\zeta^a < 0, \quad N^2 > 0$ - inertial instability;

$\zeta^a < 0, \quad N^2 > 0, \quad |q_{bc}| > q_N, \quad q_{bc} < 0$ - inertial/symmetric instability;

$N^2 < 0, \quad |q_{bc}| < |q_N|$ - gravitational instability.

Calculations using formulas (5), (6), and (7) showed that EPV averaged over the entire volume within the rectangular region outlined in Fig. 10 is positive during the summer–autumn period of 2015. This is inconsistent with the fact that active eddy formation is observed during this period (see Fig. 10). Thus, negative EPV may exist only in small patches within this region. To identify them precisely, the model results were linearly interpolated vertically onto a uniform grid with a step of 5 m from the surface to a depth of 1 km. The rectangular region in Fig. 10 was divided into many elementary volumes of size 2×2 km horizontally and 5 m vertically. Next, we counted the number of elementary volumes in which EPV was negative, followed by the number of elementary volumes in which the conditions for different types of instabilities were satisfied.

The time series of wind components and volume transport through the southern boundary of the research area are shown in Figs. 11a and b for the period from July 1 to October 30. Figure 11c shows the period (mid-July to October) during which volumes with negative potential vorticity exist in the research area. Periods of sharp growth of such volumes, specifically, mid- and late August, mid-September, and early October — coincide with periods of sharp intensification of the Tsushima

Warm Current. Figure 11c also indicates that eddy generation is primarily determined by inertial instability.

Figure 12 (see also Fig. S5) shows the spatial distribution of elementary volumes with negative potential vorticity, where the counts for the period from July to October 2015 in the upper 1 km layer are summed at each grid point. Regions of negative potential vorticity correspond to regions of inertial instability. Figure 12 demonstrates that inertial instability arises from sharp horizontal gradients in the current field, which develop after the flow passes Okushiri Island (Fig. 1b) in the coastal zone of Hokkaido Island and along the traverse of the Shakotan Peninsula. Symmetric instability is observed in already formed eddies and plays a secondary role by enhancing the effect of inertial instability, with exceptions on September 10, September 26, October 2, and October 9 (see Fig. 11c). From mid-July to October, specific wind patterns induce upwelling/downwelling and contribute to the manifestation of symmetric instability, which becomes the main contributor to negative potential vorticity during those exceptions.

In summary, the intensification of the Tsushima Warm Current leads to increased horizontal velocity gradients in specific areas: the traverses of Okushiri Island and the Shakotan Peninsula, and the coastal zone of Hokkaido Island. Large gradients trigger inertial instability, which in turn leads to the generation of AEs. Symmetric instability plays a secondary role by amplifying already formed eddies. Increased wind speed over the research area also plays a secondary role. This is confirmed by the correlation coefficients between the time series in Fig. 11, presented in Table 1, where:

- q – volume in which potential vorticity is negative (Fig. 11c, red plot)
- q_N – volume in which the condition for inertial instability is satisfied (Fig. 11c, blue plot)
- q_{bc} – volume in which the condition for symmetric instability is satisfied (Fig. 11c, green plot)
- V – volume transport through the southern boundary of the research area in the upper 200 m, in Sverdrups (Fig. 11b)
- U_{wind} – U component of wind speed (Fig. 11a, red plot)
- V_{wind} – V component of wind speed (Fig. 11a, blue plot)

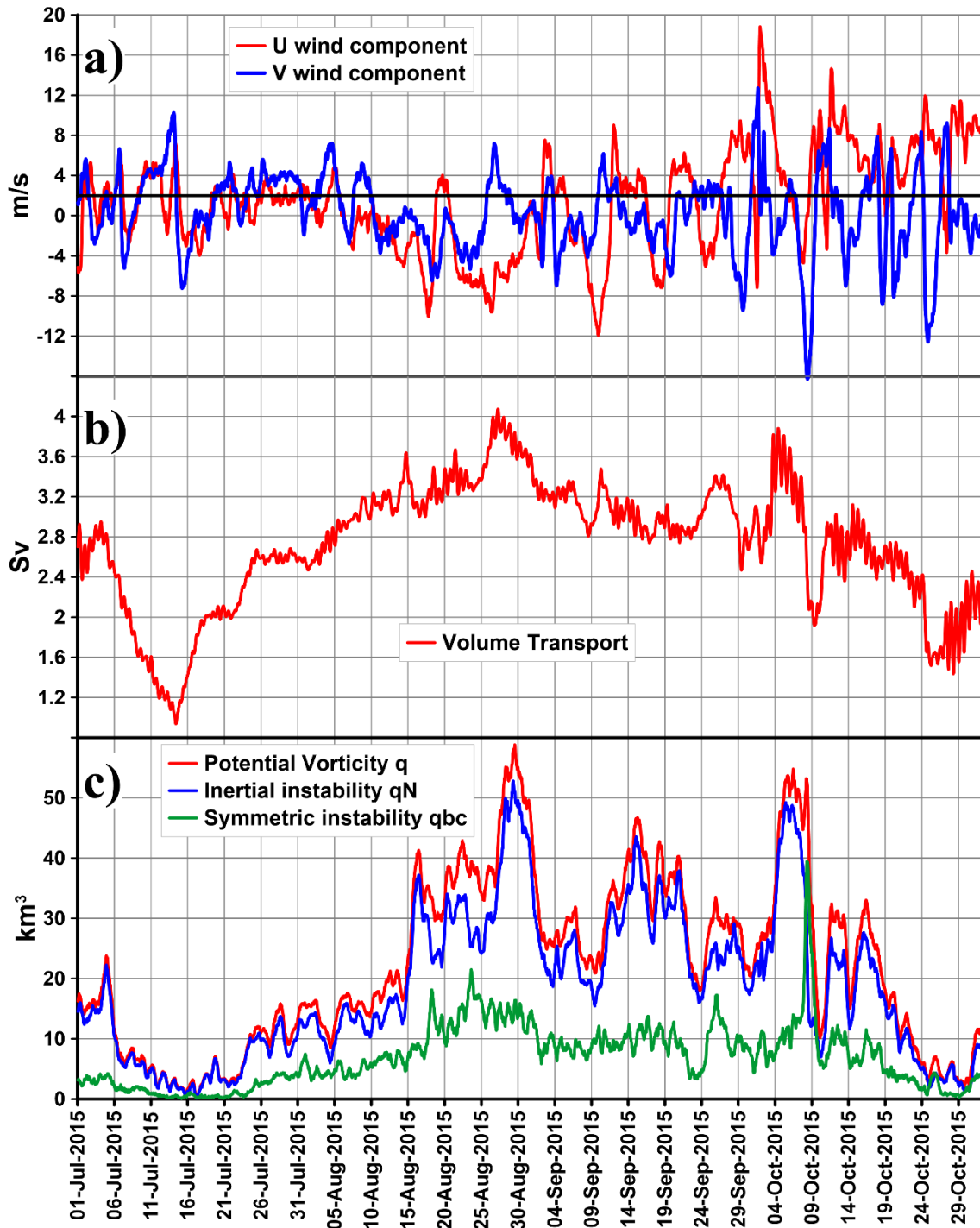


Figure 11. a) U and V wind components over the research area (m/s). b) Volume transport (Sverdrups) through the southern boundary of the research area within the upper 200 meters. c) Evolution of the water volume (km^3) characterized by negative Ertel potential vorticity (red), and volumes where conditions for inertial (blue) and symmetric (green) instability are satisfied

The correlation coefficients of importance are highlighted in bold in Table 1. Based on these, the following conclusions can be drawn, which are consistent with the above-mentioned observations. Inertial instability is primarily determined by the Tsushima Warm Current with $r(q, V) = 0.79$ and depends only weakly on the wind with $r(q, U_{\text{wind}}) = -0.27$ and $r(q, V_{\text{wind}}) = -0.21$. Nevertheless, the wind plays a significant, though not dominant, role in the generation of instability, because the intensification of easterly winds accelerates the flow along the west coast of Hokkaido Island with r

($V_{\text{wind}}, U_{\text{wind}}$) = -0.38 , resulting in negative correlation coefficients with the wind. One particularly interesting conclusion should be noted: symmetric instability, in contrast to centrifugal instability, is less influenced by currents (0.65 vs. 0.78) but is substantially more influenced by the wind ($|-0.34| > |-0.15|$; see Table 1).

Table 1. Correlation coefficients r

	q	q_N	q_{bc}	V	U_{wind}	V_{wind}
Q						
q_N	0.98					
q_{bc}	0.84	0.73				
V	0.79	0.78	0.65			
U_{wind}	-0.27	-0.27	-0.27	-0.38		
V_{wind}	-0.21	-0.15	-0.34	-0.06	0.01	

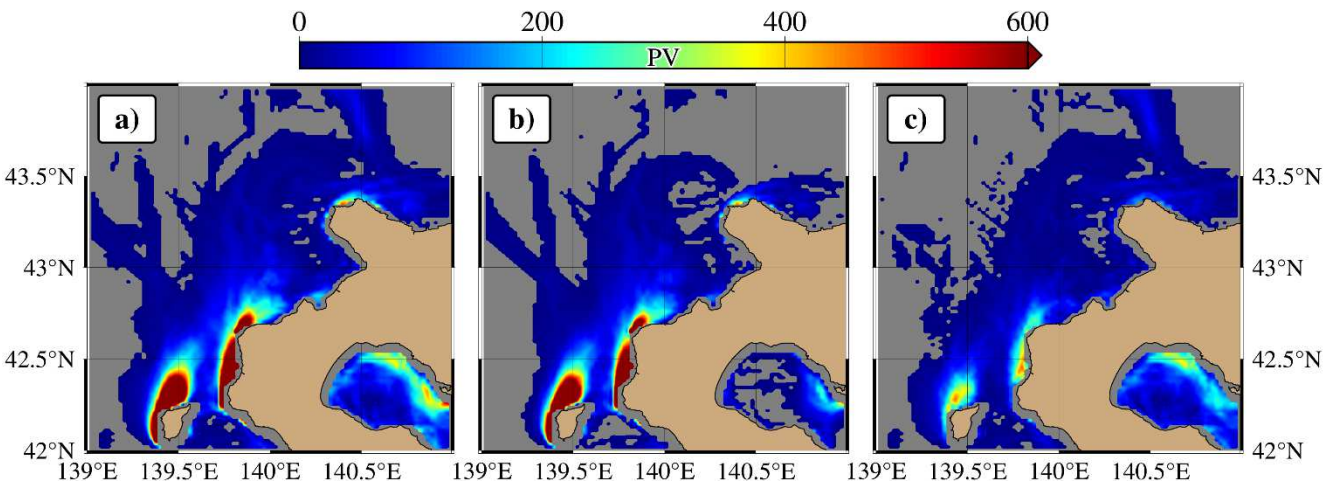


Figure 12. a) Distribution of the number of elementary volumes in which the Ertel's potential vorticity is negative, b) the number of volumes in which the condition for inertial instability is satisfied, and c) the number of volumes in which the condition for symmetric instability is satisfied

Discussion

The validation of the model results was made in several directions. To identify the main object of this study, the AE2015, we used surface current patterns derived from satellite altimetry data and multi-year CTD measurements. Satellite altimetry provides sea surface height fields, which allow the computation of surface geostrophic currents. CTD measurements provide three-dimensional density fields, enabling the calculation of currents at various depths, including the surface, using a linear diagnostic model. Thus, to validate the ROMS model outputs, we compared current data obtained through different methods based on fundamentally different types of measurements and assessed their consistency.

Over a long period of time, from January 1, 1993 to January 1, 2025, we selected seven periods during which the CTD station network was sufficiently dense to reconstruct the three-dimensional density field for subsequent diagnostic calculations using interpolation methods. These periods include May and August 1993, June and August 1994, August 1995, September 1999, and November 2000. A comparison of the diagnostic currents with the geostrophic currents (see Fig. S6) derived from AVISO data for the mid-date of each corresponding month showed good agreement between them. (The structure of geostrophic currents from AVISO data varies only slightly over the course of a month, so the mid-month date was used). Several surface drifters were identified within the research area. The trajectories of these drifters clearly demonstrate the presence of the AEs, consistently captured in altimetry-derived (Figs. 4 and S1), diagnostic (Fig. 5), and ROMS-simulated velocity fields (Figs. 6, 7 and 8).

It has been established that AE2015 was formed mainly due to the meandering of the Tsushima Warm Current (Figs. 4, 5, and 6). In the central northern part of the sea, over the slope of the Japan Basin, a cyclonic circulation is observed, apparently driven by the prevailing winds. Calculations and analysis presented in Sec. 3.4 showed that from mid-August onward, the absolute value of negative EPV increased (Fig. 11). This indicates that eddy formation was primarily determined by inertial (barotropic) instability resulting from sharp horizontal gradients of the velocity field, which arise after the coastal branch of the Tsushima Warm Current rounds the coastline of southwestern Hokkaido. The manifestation of symmetric (baroclinic) instability is observed within the already-formed eddy and plays a secondary role, amplifying the effect of inertial instability.

4. Conclusions

We calculated the occurrence frequency of mesoscale anticyclones and cyclones based on altimetry data in the previously under-studied region west of Hokkaido. Over a long period from, January 1, 1993 to January 1, 2025, we found that this area is characterized by the regular formation of both anticyclonic and cyclonic mesoscale eddies. Most of these eddies are topographically controlled by the Japan Basin and the complex Hokkaido coastline, which includes numerous capes and bays. The characteristics and evolution of an anticyclonic eddy (AE2015), observed in this region during the summer and fall of 2015, were examined using the Regional Ocean Modeling System (ROMS) with a horizontal resolution of 2 km. The simulation results revealed the eddy's three-dimensional structure and its full life cycle, including formation, movement, and dissipation. Calculation and analysis of Ertel's potential vorticity showed that inertial (barotropic) instability was the primary energy source driving the growth of AE2015. Symmetric (baroclinic) instability was observed within the already formed eddy and played a secondary role, amplifying the effect of inertial instability. Model outputs were validated against surface current patterns derived from satellite altimetry and multi-year CTD measurements. Furthermore, multi-year surface drifter trajectories provide clear evidence of anticyclonic eddies, which are

consistently captured in the altimetry-derived, diagnostic, and ROMS-simulated velocity fields within the research area.

CRedit authorship contribution statement

Pavel A. Fayman: Writing – original draft, Investigation, Validation. Sergey V. Prants: Writing – original draft, Investigation, Writing – review & editing, Funding acquisition. Alexander A. Didov: - artwork

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Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

The data are available upon request.

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