

Review of oceanic mesoscale processes in the North Pacific: physical and biogeochemical impacts

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ABSTRACT

Physical transport dynamics occurring at the ocean mesoscale (~ 20 km – 200 km) largely determine the environment in which biogeochemical processes occur. As a result, understanding and modeling mesoscale transport is crucial for determining the physical modulations of the marine ecosystem. This review synthesizes current knowledge of mesoscale eddies and their impacts on the marine ecosystem across most of the North Pacific and its marginal Seas. The North Pacific domain north of 20°N is divided in four regions, and for each region known, unknowns and known-unknowns are summarized with a focus on physical properties, physical-biogeochemical interactions, and the impacts of climate variability and change on the eddy field and on the marine ecosystem.

1. Introduction

The role that the ocean circulation plays on the living marine resources is widely recognized, but often not coherently integrated. This review aims at synthesizing the physical knowledge across a large portion of the North Pacific and its marginal Seas, focusing on mesoscale eddies and their impacts on the marine ecosystem. Oceanic mesoscale eddies with radii between *ca.* 20 and 200 km are ubiquitous features of the world ocean (e.g., Robinson, 1983). They are typically produced by instability processes and are in approximate geostrophic balance in the horizontal and in hydrostatic balance in the vertical (e.g., McGillicuddy, 2016). High-resolution sea surface height (SSH) fields constructed by merging measurements from two or more simultaneously operating altimeters have revealed that SSH variability throughout most of the world ocean is dominated by westward-propagating nonlinear mesoscale eddies that trap fluid at their core (Chelton et al., 2007; 2011b). Recent progress on eddy detection and tracking algorithms has helped clarify eddy formation mechanisms and location, their propagation, amplitude, radius, and evolution (Chelton et al., 2007; 2011b).

Mesoscale eddies influence the marine ecosystems in many ways; foremost by altering phytoplankton (chlorophyll) concentrations in the euphotic layer. This is achieved through at least six mechanisms (McGillicuddy, 2016). The first two are related to eddy advection. On time scales longer than 2–3 weeks, the dominant impact on surface chlorophyll variability is that of eddy-induced horizontal advection by the rotational velocities of the eddies (Chelton, 2011a). These rotational motions redistribute the chlorophyll creating a dipole pattern but likely no net increase in primary productivity; following Gaube et al. (2014), in this paper, this mechanism is referred to as eddy stirring. The second advective mechanism is eddy trapping. Most extratropical eddies are formed because of nonlinear instabilities that occur in presence of strong currents and/or bathymetric features and therefore relatively close to coastal, nutrient rich areas (Chelton et al., 2011b). The eddies in which the nutrients are trapped inside are transported away from their formation region, creating patches of high chlorophyll in otherwise less productive areas. This mechanism does not augment primary productivity globally either, but contributes significantly to regional budgets. The following four mechanisms increase primary productivity, rather than just moving nutrients and chlorophyll around horizontally. Eddy pumping, the third mechanism described by Falkowski et al. (1991), occurs when a cyclonic eddy intensifies and upwelling occurs in its core, resulting in enhanced chlorophyll growth. The fourth mechanism is eddy-induced Ekman pumping and is controlled by the surface stress differences around the eddy, by

the interaction of the surface stress with the surface current vorticity gradient, and, to a lesser extent, by the eddy-induced spatial variability of sea surface temperature (SST) (Gaube et al., 2015). The fifth mechanism is associated with the change of mixed-layer depth due to the presence of mesoscale eddies, with shallow mixed layers tending to have higher chlorophyll levels and vice versa. The mixed layer depth in eddies is influenced directly by isopycnal displacement and indirectly by changing propagation characteristics of near-inertial waves and by SST, as summarized in McGillicuddy (2016). This mechanism is important in the Southern Ocean, where it has been linked to changes in light and micro-nutrient availability that impact the near-surface chlorophyll concentration (Song et al., 2018). The sixth mechanism is due to processes that occur at scales of few kilometers (submesoscales) around the eddy periphery and include streamers and local upwelling, which may enhance chlorophyll production (McGillicuddy et al., 1995; McGillicuddy, 2016; Zhong et al., 2017; Liu et al., 2021).

Eddies further impact zooplankton, fish, marine mammals, and seabirds. The zooplankton modulation was first supported by observations in the Gulf Stream region, where Wiebe et al. (1976) reported that in a cold core cyclonic eddy or “ring” a zooplankton assemblage persisted longer than a phytoplankton assemblage. Mesoscale eddies can have both positive and negative impacts on fish recruitment via retention of fish larvae at their interior. For example, in the Kuroshio region, Kasai et al. (2002) found that eddies contribute to fish recruitment because they tend to recirculate in the coastal nursery area. On the other hand, in the Leeuwin Current system in the Indian Ocean the entrainment of larvae and eggs of teleost populations in a warm core eddy had a negative impact on recruitment (Gaughan, 2007). Mesoscale eddies may also influence higher trophic levels. For example, in the Gulf of Mexico the catch per unit effort (CPUE) of Atlantic bluefin tuna is significantly higher in cyclonic eddies (Teo and Block, 2010). In the Southern Ocean, mesoscale eddies and (sub-)mesoscale features ensure that king penguins and elephant seals have access to localized areas with high concentrations of prey (Cotté et al., 2007; 2015).

In the North Pacific, which is a source of “food, economic benefit and recreation, primarily through the abundant living marine resources” for all the Nations that surround it (PICES, 2004) and the target region for this paper, many recurrent eddies have been observed and have been shown to impact the ecosystems. Warm-core and cold-core rings in the Kuroshio and Oyashio Extension (KOE) Region are among the most extensively studied mesoscale eddies because their peripheries are excellent fishing grounds for pelagic fish, such as skipjack, mackerel, flying squid

and saury (e.g., Saitoh et al., 1986; Sugimoto and Tameishi, 1992; Prants et al., 2014a). In the California Current System (CCS) larval abundance of Pacific sardine is greater in offshore eddies relative to the inshore, slope and surrounding offshore waters (Logerwell et al., 2001) due to high survival rate supported by enhanced primary production in their interior (Logerwell and Smith, 2001). The Gulf of Alaska is recognized as a high-nutrient low-chlorophyll region (Nishioka et al., 2001) due to limited iron availability, and eddies formed along its coastlines contribute to the transport of iron from the coastal to the offshore ocean (Johnson et al., 2005). These eddies are so transformative that they are named based on their formation area; Haida, Sitka, and Yakutat eddies form off Haida Gwaii, Sitka and Yakutat, respectively (Crawford, 2002; Tabata, 1982; Ladd et al., 2005a).

Mesoscale eddies in the North Pacific have been investigated mainly from the perspective of their formation area, that is, many regional studies have been written on the properties and impacts of eddies formed in a given region. Few inter-comparisons have been conducted, but within specific areas (e.g., Henson and Thomas, 2008 in the Gulf of Alaska; Itoh and Yasuda, 2010a in the Kuroshio/Oyashio and their extension regions; Cheng et al., 2014 in the offshore North Pacific) and with a narrow focus, usually on physical properties. Existing inter-comparisons are also often limited to the analysis of satellite data (e.g., Kouketsu et al., 2015). This paper, on the other hand, synthesizes physical knowledge on mesoscale eddies and their impacts on the marine ecosystem across a range of scales and data types across most of the North Pacific focusing on the region north of 20°N (next referred simply as North Pacific). First, we divide the North Pacific area of interest into four regions, and for each region briefly discuss eddy formation (type and timing) and how long they last using eddy track data (Chelton et al., 2011b) (Section 2). Then we review the impact of mesoscale eddies on the marine ecosystem as well as their physical characteristics within each region (Section 3), summarizing findings in four tables that focus on known and unknowns in each region. Finally, we discuss the similarities and differences across regions, highlighting where observations/studies are mostly needed (Section 4). Although the target of this paper is mesoscale processes and their impacts, we introduce submesoscale processes as well whenever possible (see McWilliams, 2016 for a review of submesoscale circulations). Additionally, we provide information, when useful and available, about climate trends that may directly or indirectly, i.e. through changes in mesoscale activity, impact the marine ecosystem.

2. General characteristics of mesoscale eddies in the North Pacific and their impact on chlorophyll-a distribution

In this synthesis, the North Pacific has been divided into 4 regions (Figs. 1 and 2). Region 1 covers the CCS. Its northern boundary was determined by evaluating the correlation between SSH and surface chlorophyll-a (Chl-a) anomalies (Fig. 1) and selecting the latitude at which negative correlations appear (up to 45°–50°N) (Gaube et al., 2014; Kouketsu et al., 2015). It includes the west coast of Vancouver Island, that is part of the upwelling dominated CCS (e.g., Denman et al., 1981; Crawford and Thomson, 1991). Region 2 consists of the coastal region to the north of the CCS that extends to the northeastern North Pacific and the Bering Sea, where downwelling due to along-shore wind is prevalent (e.g. Henson and Thomas, 2008). Region 3 covers the western boundary current and includes the marginal seas in the western Pacific where physical and biological processes are strongly influenced by the Kuroshio and Oyashio currents (e.g. Yasuda, 2003). Region 4 covers the open North Pacific with the KOE. Strong eddies are seldom observed in the eastern part of Region 4 (e.g. Cheng et al., 2014).

The most noticeable features in the correlation between SSH and surface Chl-a anomalies in Fig. 1 are the strong negative correlations along the Kuroshio and Kuroshio extension and in the CCS. Gaube et al. (2014) presumed that the negative correlation in the Kuroshio area, which is found for most western boundary currents, results from higher chlorophyll in cyclonic eddies due to upwelling accompanied by eddy intensification and also due to the trapping of coastal waters in these regions, which have higher chlorophyll. In the CCS, on the other hand, eddy stirring and trapping play an important role and the correlation results from a chlorophyll response to the numerous cyclonic eddies (Gaube et al., 2014). Positive correlations are observed along the Alaskan Stream and Alaska Current in Region 2 and along the East Kamchatka and Oyashio currents in Region 3. Here the ambient cross-current Chl-a gradient is negative (i.e. lower Chl-a to the left of the currents) which explains the positive sign of the 4 weeks lag correlation between Chl-a and SSH anomaly (Figs. 5b and 6a in Gaube et al., 2014). Regional studies have shown that micronutrients transported from the shelf region offshore by eddy stirring and trapping support primary production in the Gulf of Alaska (e.g., Crawford et al., 2007; Ladd et al., 2009), as further reviewed in Section 3.2.

Maintaining a dynamic view of the North Pacific system is essential as mesoscale activity may change over time. The last four decades have seen a non-uniform increase in SST over

most of the North Pacific (Fig. 3). The Kuroshio and Oyashio currents and their extensions and adjacent regions experienced the greatest warming (Region 3 and 4), approaching 0.5°C/decade in the marginal seas between 1980 and 2018 (Fig. 3). The SST trend over the CCS (Region 1) and portions of the Alaskan coastlines (Region 2) has been more moderate, but these areas have been hit by extreme warming events in recent years, as we will discuss in Section 3. Warming, whether long term or episodic, and changes in stratification and water masses can modify or modulate the mesoscale distribution, and/or change how the ecosystem responds to the presence of mesoscale eddies.

To begin our review of mesoscale dynamics, we first discuss the relative contribution of each eddy formation area in the North Pacific from the perspective of the frequency of eddy ‘spawning’ and/or eddy lifetime using Chelton et al. (2011b) version 4 eddy dataset (http://wombat.coas.oregonstate.edu/eddies/AVISO_announcement.html). This dataset covers the period from January 1, 1993 to January 18, 2018 (~25 years) and includes eddies with lifetimes of 4 weeks or longer. To statistically quantify the relative contribution of each eddy formation area, we estimate the "eddy yield", that is, the average yearly lifetime integral of the eddies formed in each 2°×2° box over 25 years.

Figs. 4a and 5a show the formation rate of anticyclonic and cyclonic eddies in each 2°×2° box per year, respectively. While not uniformly, eddies form throughout the open North Pacific, consistent with results by Chelton et al. (2011b) for a shorter 16-year analysis. This agrees with the conclusions that nearly all of the World Oceans are baroclinically unstable (Gill et al., 1974; Robinson and McWilliams, 1974; Stammer, 1998; Smith, 2007b among others). Eddies are formed most frequently (*ca.* twice per year in the 2°×2° boxes) along the eastern boundary (Region 1 and eastern part of Region 2), along island chains between the open North Pacific and marginal seas (Regions 2 and 3), and near the Hawaiian Islands (~20°N, 155°W) and seamounts west-northwest of Hawaiian Islands (Region 4) (Figs. 4a and 5a). Meandering boundary currents, interactions between currents and topography (e.g., Swaters and Mysak, 1985 in the northeastern Gulf of Alaska), wind forcing effects (e.g., Thomson and Gower, 1998; Ladd and Cheng, 2016 in the northeastern Gulf of Alaska), and tidal mixing (e.g., Ohshima et al., 2005 in the southeastern Sea of Okhotsk) contribute to eddy formation. Eddies form slightly less frequently (*ca.* more than once per year but less than two in the 2°×2° boxes) in the open subtropical gyre west of the eastern boundary (around 20°–40°N, 140°–170°W: eastern part of Region 4). In this area the meridional flow is effective at generating strong eddies (Spall, 2000; Smith, 2007a). Within the western

boundary currents and their extensions (Regions 3 and 4), the eddy formation frequency is high (Chelton et al., 2011b). Eddies seldom form in the center of the Alaskan gyre ($\sim 50^{\circ}\text{N}$, 160°W). This area is known as an eddy desert (Henson and Thomas, 2008); our analysis indicates that the eddy desert is a region of both rare eddy formation as well as rare eddy propagation.

Overall, the spatial distribution of eddy formation is similar for both anticyclonic and cyclonic eddies (Figs. 4a and 5a), but differs seasonally (Fig. 6). Along the eastern boundary of the North Pacific (northern portion of Region 1 and eastern part of Region 2), anticyclonic eddies form more frequently in winter and spring while cyclonic eddies form more frequently in boreal summer and fall. In the northern part (eastern part of Region 2), anticyclonic eddies are generally thought to be generated in boreal winter, when downwelling-favorable southerly winds are at their maximum (Tabata, 1982; Henson and Thomas, 2008). In the southern part of Region 1 the eddy kinetic energy (EKE) is stronger in summer/fall and has been associated with local upwelling-favorable northerly winds in boreal summer (Marchesiello et al., 2003). Around 20° – 30°N and 130° – 170°W in the southeastern portion of Region 4, both anticyclonic and cyclonic eddies form frequently in boreal winter and spring. The seasonal variation of eddy formation in the 20° – 30°N band of Region 4 is consistent with that of vertical velocity shear and stratification, pointing to baroclinic instability as generation mechanism (Qiu, 1999). Near islands/seamounts/island-chains in Regions 2 and 3, marked seasonal variations in eddy formation are not observed. In the western boundary currents and their extensions in Regions 3 and 4, eddies form slightly more frequently in winter and spring but differences between seasons are small.

The distribution of eddy lifetime is quite different from that of eddy formation rate (Figs. 4 and 5); areas of frequent eddy formation - e.g., eastern boundary of the North Pacific - do not correspond to areas where long-lived eddies are formed. Figs. 4b and 5b show anticyclonic and cyclonic eddy lifetimes averaged within their formation area, respectively; colors in each grid indicate average lifetimes of eddies formed in the grid. The distributions of anticyclonic and cyclonic lifetimes are similar everywhere except in the Gulf of Alaska, and eddies that last more than 100 days preferentially form in the open ocean (Region 4). In the Gulf of Alaska, along the Aleutian Islands in Region 2, to the east off Hokkaido and the Kamchatka peninsula in Region 3, on the other hand, the average lifetime of anticyclonic eddy is longer than that of cyclones, as found in previous studies (Isoguchi and Kawamura, 2003; Henson and Thomas, 2008; Itoh and Yasuda, 2010a; Lyman and Johnson, 2015; Prants et.al., 2018; 2020).

Using the information on eddy formation rate and lifetime, we evaluated eddy yield (sum of lifetime divided by 9149 days (analysis period: 01Jan1993–18Jan2018)) within each $2^\circ \times 2^\circ$ formation area (Figs. 4c and 5c), providing a quantitative measure of the importance of the eddies to the circulation (over a 25-yr period). The geographical distributions of eddy yield (Figs. 4c and 5c) are mostly similar to eddy formation rate (Figs. 4a and 5a), except for few areas. Along the eastern boundary of the North Pacific, both anticyclonic and cyclonic eddies formed south of 45°N (in Region 1) can be traced for a long time (averaged lifetimes $> \sim 100$ days, Figs. 4b and 5b). In the area north of 45°N , on the other hand, only anticyclonic eddies are traced for long periods, while the lifetime of cyclonic eddies is relatively short (Fig. 5b). Figs. 4c and 5c also indicate that Region 1 is the most important from the perspective of eddy yield, followed by the eastern part of the Gulf of Alaska and the area along the Aleutian Islands in Region 2, especially for anticyclonic eddies. Eddy yield is high also in Region 4 around $20^\circ\text{--}40^\circ\text{N}$ and $160^\circ\text{E}\text{--}140^\circ\text{W}$, away from the eddy desert, because moderately long-lived eddies are formed there relatively frequently. Near islands and seamounts, many eddies are formed but their lifetime is generally short, except for those adjacent to the Hawaiian Islands. In the western boundary currents and their extensions (Region 3 and 4), eddy yield is relatively high especially in the Oyashio front region due to frequent eddy formation (Figs. 4a and 5a). Since eddy yield is just a sum of eddy lifetime divided by analysis days, more detailed studies are required to quantify the contribution of each eddy formation area to heat and freshwater transport, primary productivity, etc.

3. Regional summaries

3.1 Region 1: California Current System region

3.1.1. Introduction

Region 1 is occupied by the CCS (Figs. 1, 2 and 7) extending from the North Pacific Current (NPC) (about $\sim 50^\circ\text{N}$) to Baja California, Mexico (about 15°N). Its circulation consists in an equatorward flow that is forced by winds and by the pressure system along the west coast of the North American continent. The CCS is one of the four Eastern Boundary Upwelling Systems (Capone and Hutchins, 2013) and sustains a productive and diverse fishery that relies on the upwelling of nutrient-rich waters into the euphotic layer. More than 60 years ago it was noted that high biomass in the CCS is associated with low water temperatures (Reid et al., 1958). The

source of this cold water is coastal upwelling driven by the longshore component of local wind that results in a divergence of surface water at the coast in turn replaced by cold, nutrient-rich water from greater depths. Near the coast a wind-driven southward flowing jet dominates the dynamics, while further off-shore curl-driven upwelling is predominant (Checkley and Barth, 2009). Underneath the CCS, between 150 and 250 m of depth, the California Undercurrent transports water along the North America continental slope from the equatorial Pacific to Vancouver Island which is warm, salty and nutrient-rich relative to the interior of the subtropical gyre (Fig. 7).

Meandering characterizes both the CCS and the coastal jet, and is accompanied by the generation of vorticity filaments and mesoscale eddies with size comparable to the Rossby radius of deformation (order 100 km) and smaller (10–50 km). The generation of mesoscale eddies is generally higher between boreal late summer to early fall and results from the accumulation of wind-forced energy in the mean flow, as shown by Strub and James (2000) in the first complete assessment of the seasonal evolution of this region. Summer and fall are also, and by far, the most active seasons for California Undercurrent eddy generation (Kurian et al., 2011). The seasonal cycle has stronger amplitude in the California Undercurrent than in the CCS.

3.1.2. Physical properties

In the CCS, the presence and relevance of cyclonic (cold) and anticyclonic (warm) mesoscale eddies has been recognized through observations and modeling since the 1980s (Owen, 1980; Lynn and Simpson, 1987; Batteen, 1997; Huyer et al., 1998; Garfield et al., 1999; Chereskin et al., 2000; Strub and James, 2000; Barth et al., 2005). Owen (1980) enumerated the eddies that occurred in the CalCOFI Atlas (Wyllie, 1966) in the 25°–38°N latitudinal band, finding a predominance of large (> 100 km) cyclones over large anticyclones. In the subsurface, on the other hand, observations in the late 1990s' revealed that warm anticyclonic eddies prevail (Huyer et al., 1998; Garfield et al., 1999; Chereskin et al., 2000). They originate from the California Undercurrent and have been often referred as Cuddies (California Undercurrent Eddies; Garfield et al., 1999; Collins et al., 2013; Pelland et al., 2013). Cuddies are smaller than surface eddies and their representation in ocean model requires submesoscale-resolving meshes (grid sizes below 3 km in the horizontal) to capture the generation process.

Information on the formation areas, propagation, variability, and general physical properties of mesoscale eddies in the CCS has become increasingly available through the satellite altimeter records and the increasing resolution of numerical experiments. Systematic eddy identification and tracking have been conducted using maps of SSH, either observed or modeled (Stegmann and Schwing, 2007; Chaigneau et al., 2009; Kurian et al., 2011; Chelton et al., 2011b). The first satellite-based demography of persistent mesoscale eddies was compiled by Stegmann and Schwing (2007), using altimeter data from 1995 to 2004 over the area 30°–50°N and 130°W to the continent. They showed that both cyclonic and anticyclonic eddies with lifetime >70 days are generated predominantly in areas seaward of the 1000 m isobath, and propagate westward with a speed up to 3.5 km day⁻¹. The sizes (based on close contours) of cyclones and anticyclones are comparable, with mean $\pm$ std of 163 ± 65 km and 166 ± 59 km, respectively, but the mean generation rates are ~ 13 cyclonic and ~ 6 anticyclonic eddies per year, which is consistent with the analysis by Owen (1980). Long-lived eddies and mesoscale eddy activity in general are greatest roughly between latitudes 32°–40°N, and there is a distinct minimum to the north of approximately 42°N. The main formation area for mesoscale cyclones is 32°–34°N, while for anticyclones eddies is between 38° and 40°N. These contrasts have been confirmed by Chaigneau et al. (2009) and Chelton et al. (2011b), and more recently by Chenillat et al. (2018) for the southern part of the CCS off the Baja California Peninsula.

Using both satellite altimetry and a high-resolution numerical model, Kurian et al. (2011) identified eddies both at the surface and at the subsurface. At the surface, long-lived cyclones occur approximately twice as frequently as long-lived anticyclones, whereas at the subsurface, the former are less than 25% of the latter. The difference in dominance of cyclonic/anticyclonic eddies with depth is consistent with previous studies (e.g., Owen, 1980; Huyer et al., 1998; Garfield et al., 1999; Chereskin et al., 2000; Stegmann and Schwing, 2007; Chaigneau et al., 2009; Chelton et al., 2011b). The composites of temperature and salinity profiles clearly show cold/fresh (warm/saline) anomalies for both surface and subsurface anticyclonic (cyclonic) eddies.

As mentioned, subsurface anticyclonic eddies generated by the California Undercurrent are classified as submesoscale coherent vortices and named Cuddies (Garfield et al., 1999; Collins et al., 2013; Pelland et al., 2013). The analysis of a submesoscale-resolving simulation suggests that the subsurface mesoscale anticyclonic eddies with a scale close to the baroclinic radius of deformation emerge from the generation of submesoscale negative-vorticity along the inshore

side of the California Undercurrent (Molemaker et al., 2015) through shear instability. In this region elevated submesoscale variance has indeed been observed (Itoh and Rudnick, 2017).

3.1.3. Physical-biogeochemical interactions

Biological activity along the CCS has been investigated since the late 1940's following the creation of the California Cooperative Oceanic Fisheries Investigation (CalCOFI) by Harald U. Sverdup and others. The eddies formed near shore along the CCS path are primary contributor to long-distance cross-shelf transport and to the redistribution of upwelled water, and they are characterized by enhanced nutrients and therefore elevated biological activity (Marchesiello et al., 2003; Gruber et al., 2011; Chenillat et al., 2018).

Early recognition of spatially correlated mesoscale patterns of cold, upwelled water and enhanced surface pigment concentration goes back decades in this region (e.g., Strub et al., 1990; Abbott and Barksdale, 1991). Current understanding of the biological response to jets, filaments, meanders, fronts and eddies in the CCS has been reviewed by Checkley and Barth (2009). Correlated temperature and surface chlorophyll features are associated with upwelling filaments with widths of ~10 km and lengths of 100s of km extending offshore from the coastline, for example, Heceta Bank (44°N), Cape Blanco (42.8°N), Cape Mendocino (40.4°N) and Point Conception (34.5°N) (Fig. 8). Other large features on the order of 100s of km in the along-coast direction are associated with flow-topography interaction at capes and submarine banks and canyons (e.g., Barth et al., 2005; Stegmann and Schwing, 2007). Mesoscale eddies, both cyclonic and anticyclonic with diameters of 30–50 km show correlated variation of temperature and surface chlorophyll. Cyclonic eddies are regions of surface convergence and therefore enhanced surface chlorophyll. At the finest scales, submesoscale instabilities (Barth, 1994) and eddies with cold water and high chlorophyll are present on the flanks of the meandering jets and filaments. All these features are apparent in Fig. 9.

A recent analysis based on satellite-derived measurements has shown that the longer-lived cyclonic eddies located offshore the CCS but generated near the coast are characterized by higher concentrations of particulate organic carbon than cyclones of similar amplitude generated offshore, and that this particulate organic carbon enrichment can be tracked for 1000 km from the coast (Amos et al., 2019).

3.1.4 Variability

In Region 1 the relation between large-scale climate variability and ocean transport, and therefore marine ecosystems variability, has been widely documented through long-term observing systems (see Di Lorenzo et al., 2013a and citation therein). Relevant large-scale climate modes are the El Niño Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997; Mantua and Hare, 2002) and the North Pacific Gyre Oscillation (NPGO) (Di Lorenzo et al., 2008; 2009). The warm phase of ENSO, El Niño, has large and often negative consequences on the CCS ecosystem through both local (see e.g., Alexander et al., 2002) and advective mechanisms (see e.g., Frischkencht et al., 2015). Impacts on the marine ecosystem may occur through the direct influence of ENSO on temperature, dissolved O₂, and pH, via food web changes and through changes in advection and therefore transport of organisms (predators, prey, parasites, or pathogens) poleward and/or onshore. The mechanism by which the PDO and NPGO influence the marine ecosystem of the CCS is established through their atmospheric forcing counterparts, the Aleutian Low and the North Pacific Oscillation, respectively, which control the upwelling and downwelling 3-dimensional cells at interannual to decadal scales (Chhak and Di Lorenzo 2007; Chhak et al., 2009; Song et al., 2011; Jacox et al., 2018). The NPGO is key to the upwelling in the CCS, as shown through numerical simulations (Combes et al., 2013). Through its link to the coastal wind stress, the NPGO explains much of the decadal variability of salinity and key biological variables in the CCS (Di Lorenzo et al., 2008). In the last decades the variance of these modes and especially of the NPGO has grown over time in the winter season, as evidenced by both sea level pressure and surface temperature anomalies (Liguori and Di Lorenzo, 2018). It remains to be investigated if and how this increase in variance in the modes impacts mesoscale transport dynamics.

Advances in regional physical-biological modeling in the past twenty years (see the seminal paper by Marchesiello et al., 2003) have allowed better resolution of how the eddy-scale processes in the CCS modulate the water mass distribution and in turn the marine ecosystem. In terms of ecosystems, transport dynamics control primary productivity through nutrient fluxes associated with upwelling and eddy-pumping, ecosystem connectivity and species distribution (e.g., zooplankton) through changes in alongshore currents, and exchanges of biological properties between the shelves and the open ocean (e.g., iron, phytoplankton, fish larvae, and carbon) via meso- and submeso-scale eddies and fronts (see for a review Di Lorenzo et al., 2013b).

Jets, eddies and filaments are responsible for a high production that is either consumed, is transported horizontally or sinks, at times in a subsiding water mass (Fig. 9). According to model simulations (Plattner et al., 2005; Gruber et al., 2006), mesoscale circulations cause a decoupling between new and export production along the central California coast through the offshore transport of small particles. Cyclonic (cold core) eddies that form offshore of the upwelling jet are usually areas of high primary and secondary production (Huntley et al., 2000). For example, off the coast of Oregon late-stage larvae of Pacific sardines are more abundant offshore in the cyclonic eddies than elsewhere, with eggs being in larger concentrations inshore (Logerwell and Smith, 2001; Fig. 8). This observation has been corroborated by a related modeling study by Logerwell et al. (2001) that attributed the high survival of sardine larvae in offshore, cyclonic eddies, to enhanced production of their planktonic food. Similarly, the strong mesoscale activity in the core of the California Current off central and southern California (Lynn and Simpson, 1987) affects chemical and biological properties (Hayward and Venrick, 1998), and support sardine spawning (Checkley et al., 2000).

With a 10-year long integration of a coupled physical-biological model run at 5 km horizontal resolution, Chenillat et al. (2016) showed that cyclonic eddies in the CCS propagate westward across the shelf up to 800 km, transporting coastal planktonic organisms and maintaining locally elevated primary productivity for up to 1 year. Modeled anticyclones, on the other hand, covered about half of the distance during their ~ 6 months lifetime and had a more limited impact on the ecosystem. In this simulation eddies occupied about 8% of the modeled CCS domain at any given time but were responsible for approximately 50% of the nitrate transport. The influence of mesoscale eddies is not limited to primary productivity. Recent in situ and modeling studies have indicated that mesoscale features enhance carbon export in the CCS (Stukel et al., 2017) and are important for understanding trends in ocean acidification in the CCS (Gruber et al., 2012).

The role of submesoscale variability and overall contribution of submesoscale circulations to the CCS ecosystem remains uncertain. The vertical velocities associated with surface frontogenesis and nonlinear Ekman pumping have been shown to downwell phytoplankton, effectively removing them from the euphotic layer. Gruber et al. (2011) found that the overall CCS productivity decreased in submesoscale permitting simulations compared to mesoscale resolving runs because phytoplankton contained in submesoscale filaments formed at the mesoscale upwelling front were subducted out of the euphotic zone before the nutrients were

fully consumed. An investigation focused on Monterey Bay used numerical simulations together with bio-optical and physical observations to show that the negative impact of submesoscale downwelling is strong in early summer, but the high vertical velocities at the edge of the CCS front have a positive effect on chlorophyll concentrations later in the season, when the mixed-layer depth is shallower than the euphotic depth (Shulman et al., 2015).

In terms of trends, over the last decade the region has experienced dramatic changes in its climate with the occurrence of several record-breaking temperature extremes between 2013–2021, such as the multi-year 2013–15 Northeast Pacific marine heatwave (Bond et al. 2015; Di Lorenzo and Mantua, 2016; McCabe et al., 2016), and the strong 2015-16 El Niño (Jacox et al., 2016). These changes in ocean temperatures influence the metabolic rate and oxygen demand of marine species (e.g., Deutsch et al., 2015), and displace habitats (e.g., Pinsky et al., 2013).

It is known that the eddy circulation statistics are influenced by climate forcing (Davis and Di Lorenzo, 2015). The regional model simulations by Combes et al. (2013), for example, have shown that in the CCS the low-frequency variability of coastal upwelling and the cross-shelf transport of the upwelled water mass depend on the alongshore wind stress in both the northern and southern portions. However, the offshore surface transport is modulated by the mesoscale activity which differs in the two regions. In the south, in particular, cyclonic eddies entrain water masses of southern origin that are then advected poleward by the California Undercurrent, and Kahru et al. (2012) have shown that fronts have increased in frequency over the last couple of decades. This likely happened in response to climate-driven changes in wind and stratification. Building upon these results, Chenillat et al. (2013) investigated the influence of intrinsic and deterministic physical forcing on the ecosystem response in the CCS, again using a circulation model coupled to the North Pacific Ecosystem Model for Understanding Regional Oceanography (NEMURO; Kishi et al., 2007). In their simulation the timing of the onset of the upwelling season in the CCS varies as function of the alongshore winds. The onset, in turn, affects not only the coastal ecosystem during the upwelling season but also the offshore ecosystem year-round. This offshore response can be as much as a factor of two stronger than that at the coast in relative amplitude.

On the marine ecosystem side, drastic changes in the chemistry have been projected, with ocean acidity expected to rise dramatically by 2050 (Gruber et al., 2012). These model projections are limited by resolution and an oversimplified representation of biogeochemical

cycling but have motivated further observational and modeling work. Chan et al. (2017), for example, have been able to describe a remarkable level of spatial variability in the penetration of acidified waters into nearshore habitats across the CCS using a novel coastal ocean acidification observing network. These authors found that in some hotspots, suboptimal conditions for calcifying organisms already extended to up to 56% of the summer season and were accompanied by a great level of pH variability. In other areas, however, persistent refuges were also found, suggesting that local adaptation may persist on longer time scales than predicted by climate models.

3.1.5 Summary

The CCS occupies Region 1 and is a productive Eastern Boundary Upwelling System where eddies are commonly observed. Equatorward winds produce offshore surface Ekman transport and upwelling along the coast, especially during the summer, and bring deep, cold, nutrient-rich waters to the surface. Mesoscale eddies, both cyclonic and anticyclonic, are formed from the CCS meandering as it flows southward along the coast. These eddies redistribute nutrients, transporting the upwelled water hundreds of kilometers offshore from the coastal area into the oligotrophic interior of the Pacific Ocean. Observational and modeling studies have quantified how and how much coastal water is transported by the eddies. The cyclones are areas of elevated primary and secondary production and are preferred nursery ground for late-stage sardine larvae. ENSO, PDO and NPGO, all contribute to the variability of the coastal Ekman upwelling and strongly influence the characteristics of the upwelled waters (temperature, salinity, pH, nutrient and oxygen concentrations), therefore impacting the marine ecosystem of the CCS. The main characteristics of the eddies in this region, knowns and unknowns are summarized in Table 1. Overall, research is needed to explore the intrinsic and forced variability of the mesoscale eddy field in the CCS, given its effect on ecosystem productivity. In particular, it is critical to better quantify the intrinsic variability of the eddy field and the relation between climate forcing and this intrinsic variability, which is likely nonlinear. This is particularly urgent considering the increasing frequency in marine heat waves recorded in the last decade (Oliver et al., 2018). Additionally, the contribution of subsurface eddies to the productivity of the area remains unconstrained, and targeted field campaign would improve understanding of the role of these mesoscale circulations on the productivity of the region.

499 Table 1: Characteristics and impact on the marine ecosystem of eddies in Region 1. Priorities
500 are underlined and doubly underlined (higher priority).

Eddy type	Surface cyclone	Surface anticyclone	Subsurface anticyclone
Formation area	30°–45°N (most formed at 32°–34°N) seaward of 1000 m isobath (Stegmann and Schwing, 2007)	30°–50°N (most formed at 38°–40°N) seaward of 1000 m isobath (Stegmann and Schwing, 2007)	Along California Undercurrent. Frequently formed at ~34°N and ~40°N (Kurian et al., 2011)
Formation mechanism	Baroclinic instability and flow-topography interaction of surface-intensified California Current and coastal jets (Checkley and Barth, 2009)	Baroclinic instability of surface-intensified California Current and coastal jets (Checkley and Barth, 2009)	From the generation of submesoscale negative-vorticity along the California Undercurrent through shear instability (Molemaker et al. 2015).
Polarity	Cyclonic	Anticyclonic	Anticyclonic
Diameter	163 ± 65 km (Stegmann and Schwing, 2007)	166 ± 59 km (Stegmann and Schwing, 2007)	59 km (median) (Collins et al., 2013) ; 20.4 km (Pelland et al., 2013)
Propagation	Westward		

Lifetime (years)	Up to 0.8 (Stegmann and Schwing, 2007)	Up to 1 (Stegmann and Schwing, 2007)	Up to 1.4 (Collins et al., 2013)
Heat/freshwater transport	<u>Not yet fully quantified. Observations are needed to validate model estimates over the whole region and to quantify overall submesoscale impacts.</u>		Heat and salt fluxed seaward, out of poleward undercurrent (Pelland et al., 2013). More observations are needed for a seasonal quantification.
Nutrient transport	Offshore transport (Nagai et al., 2015); Offshore transport of carbon (surface cyclone, Barth et al., 2002)		
Impact on chlorophyll	Seaward transport of coastal planktonic organisms (e.g. Strub et al., 1990; Abbott and Barksdale, 1991; Logerwell and Smith, 2001; Checkley and Barth, 2009; Chenillat et al., 2016)		<u>Unknown. Likely less important than for surface eddies but could transport nutrients.</u>
Impact on zoo/ichthyoplankton			<u>Unknown. Observational studies to survey zooplankton assemblages inside, around, and outside mesoscale eddies are recommended.</u>
Impact on higher trophic levels	Positive impact on sardine survival (surface cyclone: Logerwell and Smith 2001) and linked to fur seals migration (Pelland et al.,		

	2014, Sterling et al., 2014). <u>Studies targeting other species are recommended. Historical fishery data could be used to this end.</u>
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502 **3.2 Region 2: Northeastern North Pacific and the Bering Sea**

503 *3.2.1 Introduction*

504 Region 2 covers the subarctic North Pacific (Figs. 1, 2 and 10). The northern part of the NPC
505 enters the western edge of this region and flows northwestward along the coast of Canada and
506 US as the Alaska Current, which, in turn, flows west-southwestward as the Alaskan Stream
507 along the southern coast of the Alaska Peninsula and the Aleutian Islands. Part of the Alaskan
508 Stream enters the Bering Sea, flows eastward just north of the Aleutian Islands, turns
509 northwestward and flows along the shelf break of the eastern Bering Sea shelf as the Bering
510 Slope Current. Mesoscale eddies in this region are mostly associated with these three boundary
511 currents (see Fig. 10).

512 In Region 2, the existence of mesoscale eddies was recognized in the first half of 20th century
513 through analysis of currents flowing in the opposite direction to that of the prevailing flow in
514 the Gulf of Alaska. Based on hydrographic sections across the continental slope, McEwen et al.
515 (1930) reported an eastward current near the continental slope across a section where the
516 westward Alaskan Stream and Alaska Current were dominant. Bennet (1959) identified
517 numerous eddies from hydrographic surveys, and Kirwan et al. (1978) found anticyclonic
518 eddies in the eastern edge of the Gulf of Alaska using satellite tracked drifters. The first
519 comprehensive characterization of the eddy field in the region was realized by Tabata (1982).
520 Analyzing hydrographic data from 1954 to 1967 and drifter observations from 1977, he
521 suggested that the anticyclonic eddies observed off Sitka (Alaska), which he named Sitka
522 eddies, were formed locally and persisted for six months or more. Years later, Gower and
523 Tabata (1993) tracked the motion of anticyclonic eddies through the Gulf of Alaska using
524 observations of the exact repeat mission of the Geosat satellite altimeter, from November 1986
525 to January 1989. Their analysis showed that more eddies were formed in the ENSO winter of
526 1986/87 than in the following non-ENSO winter, as summarized by Crawford (2002). Since
527 1992, when satellite observations by at least two operating altimeters became available, many
528 studies have been published on eddies in the Gulf of Alaska. Among them, Ladd (2007) and

Henson and Thomas (2008) investigated their seasonal and interannual variability, and found that EKE was high in winter/spring but large eddies (diameter > ~80 km) were fewest in winter.

West of the Gulf of Alaska and south of the Aleutian Islands, mesoscale eddies were first detected after the launch of the Geosat satellite altimeter. Okkonen (1992) first reported the observation of a meander which subsequently separated from the Alaskan Stream as an anticyclonic eddy. More recently, Rogachev et al. (2007) and Saito et al. (2016) investigated eddies in the western subarctic gyre, which mainly form at 170°–175°E south of the Aleutian Islands and supply the East Kamchatka Current with heat, while Ueno et al. (2009) studied the propagation of anticyclonic eddies southwestward in the Alaskan Stream and indicated that eddies form not only off Sitka and Yakutat, but also south of the Alaska Peninsula and Aleutian Islands and then propagate westward beyond 180°.

In the Bering Sea, mesoscale eddies have long been studied. Based on hydrographic data, Kinder et al. (1975) described a mesoscale eddy in the southeastern corner of the Bering Sea basin along the Bering Slope Current system flowing northwestward along the shelf break. Satellite altimetry data further motivated studies to characterize Bering Sea eddies (e.g., wave field satisfying the dispersion relation for topographic planetary waves: Okkonen, 1993) and discuss their spatio-temporal variation along the shelf break, which is related to wind field as well as topography (Okkonen, 2001; Ladd et al., 2012).

Studies on the impact of mesoscale eddies on chemical and biological fields in the subarctic North Pacific started in 1990s. In the southeastern Bering Sea, Schumacher and Stabenro (1994) found an association between high concentrations of pollock larvae and eddies. Springer et al. (1996) summarized how physical processes at the shelf edge, such as intense tidal mixing and eddies in the Bering Slope Current, bring nutrients into the euphotic zone and contribute to enhanced primary and secondary production and elevated biomass of phytoplankton and zooplankton. After Martin et al. (1989) hypothesized that iron is an essential micronutrient that controls phytoplankton growth, iron observations within and around eddies were conducted (e.g., Johnson et al., 2005; Rovegno et al., 2009; Ladd et al., 2009; Tanaka et al., 2012; 2015; 2017) finding that indeed eddies were responsible for transporting both macro- and micronutrients from the shelf region into the basin. Eddies have been shown to contribute to distributions of chlorophyll also in the Gulf of Alaska (e.g., Crawford et al., 2005), south of the Aleutian Islands (Ueno et al., 2010) and in the Bering Sea (e.g., Mizobata et al., 2002), and to

be linked to concentrations of zooplankton in the Gulf of Alaska (Mackas and Galbraith, 2002) and south of the Aleutian Islands (Saito et al., 2014), and to concentrations of fish in the Gulf of Alaska (Kline, 2010) and in the Bering Sea (Andreev et al., 2018). Relationships between eddies and marine mammals and seabirds have been also investigated (Ream et al., 2005; Pelland et al., 2014; Paredes et al., 2014; Santora et al., 2018). In most cases, the impact of mesoscale eddies on the marine ecosystem is via micronutrient transport to the high nutrient low chlorophyll (HNLC) area in the middle of the gyre and local enhancement of primary production (e.g., Ladd et al., 2007; Lippiatt et al., 2011), but horizontal advection by mesoscale structures can also influence larval fish assemblages over the shelf and slope (Atwood et al., 2010).

3.2.2 Physical properties

In the subarctic North Pacific, eddy yield is relatively high in the Alaska Current region, at the western edge of the Alaskan Stream region and in the Bering Slope Current region (Figs. 4, 5 and 11). This is consistent with the EKE map. In the Alaska Current and Alaskan Stream region, the formation frequency of anticyclonic and cyclonic eddies is similar (Figs. 4 and 5) but anticyclones are longer-lived. As a result, Lyman and Johnson (2015) estimated that only ~37% of all detected eddies were cyclonic in the Gulf of Alaska, and even less (~15%) were identified by Henson and Thomas (2008) using the Okubo-Weiss parameter (Okubo, 1970; Weiss, 1991). Anticyclonic eddies are not only longer-lived, but also substantially larger (Fig. 11), and more nonlinear (Fig 2. b and d in Lyman and Johnson, 2015 showing azimuthal velocity divided by translation speed). Cyclonic eddies formed in the Alaskan Stream region west of the Gulf of Alaska have not been discussed in the literature (Henson and Thomas, 2008; Lyman and Johnson, 2015), while in the Alaska Current region, cyclonic eddies form more frequently but have shorter lifetime than anticyclones (Figs. 4b and 5b). Overall, in Region 2 cyclonic eddies have received relatively less attention than anticyclonic eddies and should be further investigated.

In the Alaska Current region, anticyclonic eddies form frequently off Haida Gwaii, Sitka, and Yakutat (Fig. 11) (Crawford, 2002; Whitney and Robert, 2002). Haida eddies form off the southern tip of Haida Gwaii around 53°N and move mostly westward to the central Gulf of Alaska (Crawford, 2002; Whitney and Robert, 2002). Sitka eddies form off Sitka around 57°N and move mostly westward (Gower, 1989; Crawford, 2002), with few moving northwestward

and becoming embedded in the Alaskan Stream (Crawford et al., 2000; Ueno et al., 2009). Yakutat eddies form at the head of the Gulf of Alaska (141°–144°W) near Yakutat, Alaska, and propagate southwestward along the Alaskan Stream (Okkonen et al., 2003; Ladd et al., 2005a; 2007; Janout et al., 2009; Ueno et al., 2009). These three groups of eddies are primarily distinguished by their origin and share common features, e.g., anticyclonic rotation and ~200 km diameter, but their formation mechanisms differ. The formation of Haida eddies is associated with the mean advection of warmer and fresher water masses around the cape from Hecate Strait between Haida Gwaii and North America (Crawford et al., 2002; Di Lorenzo et al., 2005), while Sitka and Yakutat eddies form via baroclinic instabilities in the northward flowing currents along the shelf associated with coastally trapped Kelvin waves and alongshore downwelling winds that destabilize the Alaska Current by enhancing the velocity shear in the vertical (Thomson and Gower, 1998; Melsom et al., 1999; Murray et al., 2001). The tightly defined Sitka eddy formation region is associated with topographic interactions (Swaters and Mysak, 1985). Ladd et al. (2009) suggested that Yakutat eddies form in shallow shelf water with riverine input, while the Sitka and Haida eddies appear to form in deeper waters. Shore et al. (2008) suggested that barotropic instability (both baroclinic and barotropic instabilities) is (are) important as an energy source for eddies in the north (east) coast of the Gulf of Alaska. Ladd and Cheng (2016) and Ladd et al. (2016) suggested that gap-winds – strong offshore-directed winds channeled through mountain gaps – may play a role in the formation of Sitka and Yakutat eddies and may modify the eddies near Kodiak Island.

Haida, Sitka and Yakutat eddies transport heat/freshwater from the coastal region to the deep-sea region of the Gulf of Alaska via advection along their outer rim while also trapping coastal water in their center. The latter mechanism has been evaluated based on in-situ and satellite observations. Based on 12 hydrographic sections from 1995 to 2001, the offshore heat and freshwater transport in the top 500 m by a Haida eddy are about 3×10^{19} J and 50 km³ in a typical winter, with a range of $1\text{--}10 \times 10^{19}$ J and 0–70 km³ (Crawford, 2005). Ladd et al. (2007) estimated that a Yakutat eddy formed in 2003 contained 5.7×10^{18} J of anomalous heat and 21.1×10^{10} kg of anomalous salt between the 25.2 and 27.0 kg m⁻³ isopycnals when it was first sampled in May 2003; the anomalies eroded over time as the eddy was sampled 4 more times through October 2004. Lyman and Johnson (2015) conducted a more comprehensive study on heat and freshwater transport to the basin using eddy trajectories (Chelton et al., 2011a) and Argo profiling float hydrographic data and found $16.2 (\pm 5.6) \times 10^{18}$ J of anomalous heat and –

5.6 (± 3.8) km³ of freshwater per eddy and $\sim 48.6 \times 10^{18}$ J and ~ -16.8 km³ per year, respectively, between 0 and 900 m. The positive heat flux and negative freshwater flux presumably derive from the Alaska Current and the Alaskan Stream, which both carry subsurface water that is warmer and saltier than that within the center of the gyre (Lyman and Johnson, 2015). Xiu et al. (2011) further suggested the importance of submesoscale processes in the Haida eddy region using numerical simulations. Their model experiment indicates that submesoscale features at the periphery of eddy cores dominate the vertical velocity field, and that the vertical velocity at the eddy centers is about one to two orders of magnitude lower than that in the submesoscale structures.

In the Alaskan Stream region, at least three formation sites of long-lifetime anticyclonic eddies have been documented, and the respective eddies have been named Kenai, Alaskan Stream and Aleutian eddies (Fig. 12). The Kenai eddy formation site is near the Kenai Peninsula ($\sim 150^\circ\text{W}$; Rovegno et al., 2009). One well-studied Kenai eddy, containing warm core water with a uniform temperature-salinity relationship near the eddy center, originated there and initially propagated along the Alaskan Stream (Rovegno et al., 2009) and then far to the west ($\sim 175^\circ\text{E}$) with a lifetime of 3.6 years (Ueno et al., 2012). The original warm core water was preserved for 2.7 years but strongly modified by eddy-eddy interaction after that (Ueno et al., 2012). Although at least six eddies formed near the Kenai Peninsula over 16 years (Rovegno et al., 2009), their formation mechanism, interannual variation, and heat/freshwater transport remain to be investigated.

The Alaskan Stream eddy formation site is south of the Alaska Peninsula and Aleutian Islands between 157° and 169°W (Ueno et al., 2009). Alaskan Stream eddies have been observed to cross the 180° meridian and reach the western subarctic gyre and they contribute warm water to the western subarctic gyre just south of the Aleutian Islands (Ueno et al., 2009). Five Alaskan Stream eddies formed during 1992–2006 and lasted up to five years. Four of the 5 eddies formed under negative or weakly positive wind stress curl, which possibly caused the Alaskan Stream to separate from the coast and become unstable (Ueno et al., 2009). A comparison of eddy propagation speeds in the Alaskan Stream with the bottom slope showed that eddies propagated faster over steeper slopes. Finally, the anticyclonic Aleutian eddy formation site is south of the Aleutian Islands at the western edge of the Alaskan Stream between 170° and 175°E (Rogachev et al., 2007; Saito et al., 2016; Budyansky et al., 2022), and Aleutian eddies separate from the main current when it turns north through Near Strait into the

Bering Sea. Using an automatic eddy detection algorithm and altimetry-based daily Lagrangian maps, all long-lived and large-scale anticyclonic Aleutian eddies with size in the range of 100 – 300 km have been identified, tracked and documented during the period 1993 – 2020 (Budyansky et al., 2022). A strong asymmetry between the number of cyclones and anticyclones has been found. After detaching from the Alaskan Stream, the anticyclones propagate southwestward due to the beta-effect. The observations, based on Argo float profiles (Budyansky et al., 2022) and ship measurements (Rogachev et al., 2007; Saito et al., 2016), reveal typical subarctic vertical structure with warm and saline mesothermal water in the intermediate layer. The Aleutian eddies transport warm and saline water under a cold upper layer that eventually contributes to the mesothermal layer of the Western Subarctic Gyre.

In the Bering Slope Current region, frequent eddy generation is associated with the Bering slope canyons (Bering, Pribilof, Zhemchug, Navarin; Okkonen, 2001; Kinney et al., 2009; Ladd et al., 2012, Fig. 13). Eddies in this region have horizontal scales ranging from 10–200 km (Paluszkiwicz and Niebauer, 1984; Stabeno et al., 1999) and vertical scales reaching at least 1000 m (Roden, 1995; Mizobata et al., 2002). The eddy activity is strongest in spring, and instabilities in the Bering Slope Current, wind forcing, topographic interactions and flow through the eastern Aleutian passes have all been suggested as possible generation mechanisms (e.g., Paluszkiwicz and Niebauer, 1984; Schumacher and Stabeno, 1994; Mizobata et al., 2008).

3.2.3 Impact on chemical and biological fields

In the Alaska Current and Alaskan Stream regions, Haida, Sitka, Yakutat and Kenai eddies provide deep-sea waters with macro/micro-nutrients -especially iron - and biota from the coastal areas. Haida (Johnson et al., 2005; Cullen et al., 2009), Sitka (Brown et al., 2012), Yakutat (Crusius et al., 2017) and Kenai eddies (Lippiatt et al., 2011) all contribute to the iron transport, as also indicated by numerical simulations (Combes et al., 2009; Fiechter and Moore, 2012). Johnson et al. (2005) further suggested that upward transport along isopycnals and upwelling within the eddy due to eddy decay (eddy pumping) provide steady fluxes of iron into the euphotic zone from the iron-rich subsurface waters of the eddy cores. Ladd et al. (2009) showed that iron concentrations measured in a Yakutat eddy were higher than any previous Haida eddy measurements (Johnson et al., 2005), including those from an eddy at its formation time, when

biological drawdown had presumably not yet occurred. Therefore, eddies formed on the shelf around Yakutat are key in providing iron to the Gulf of Alaska basin.

In the Bering Sea, eddies can be important for transporting iron and macronutrients to the shelf break as well (Hurst et al., 2010). Eddy induced cross-shelf transport may supply macronutrients (Springer et al., 1996; Mizobata et al., 2006; Mizobata et al., 2008) and iron (Hurst et al., 2010), and cause the highly productive ecosystem found along the edge of the continental shelf in the Bering Sea, in the so-called Green Belt. Additionally, Ladd et al. (2012) showed evidence that eddies trap deeper water from the outer shelf in their core at formation time and carry it with a high nitrate signature off-shelf. All in all, eddies in Region 2 enhance production on-shelf and as well as off-shelf (e.g., Ladd et al., 2005b; Mizobata et al., 2006).

Fig. 1 indicated that chlorophyll concentration is positively correlated with SSH (high chlorophyll associated with anticyclonic eddies) in the Alaska Current, Alaskan Stream and Bering Slope Current regions. This is consistent with previous studies. For example, Crawford et al. (2007) showed that in the Gulf of Alaska more than half of all surface chlorophyll was inside the 4 cm SSH contours of anticyclonic mesoscale eddies (Haida, Sitka and Yakutat eddies), yet these contours enclosed only 10% of the total surface area of offshore waters. They also indicated that macro- and micronutrient transport by eddy trapping, stirring and pumping contribute to the offshore primary production. Lippiatt et al. (2010; 2011) indicated that mesoscale anticyclonic eddies in the Gulf of Alaska are key to the offshore transport of iron-rich coastal waters, enriched by the runoff from glacial rivers and streams. Ueno et al. (2010) confirmed that Alaskan Stream eddies contribute significantly to the Chl-a distribution in the deep-sea region of the subarctic North Pacific based on the climatological Chl-a distributions averaged over space and time. They also suggested that nutrient-rich coastal water is transported offshore via a southward meander of the Alaskan Stream due to Alaskan Stream eddy trapping in the core and eddy stirring in the outer ring. Vertical transport of iron via eddy pumping or eddy-induced Ekman pumping may be another mechanism by which nutrients are supplied to the euphotic layer (Ueno et al., 2010), although Dobashi et al. (2021) indicated that it is not always effective in the offshore region of the western subarctic gyre.

In the Bering Slope Current region, Mizobata and Saitoh (2004) and Mizobata et al. (2008) found that the variability in the eddy field and the primary production along the Bering Sea shelf edge are positively correlated. Increase in Bering Slope Current transport and eddy activity

contributes to vertical nutrient supply to the subsurface layer and to shelf–slope exchanges, thereby maintaining high primary productivity along the shelf edge of the southeastern Bering Sea. Ladd et al. (2012) indicated that the horizontal exchange of water within an eddy may result in the excess nutrients and fresher water within the eddy core, influencing Chl-a distributions throughout the summer months. Tanaka et al. (2012; 2015; 2017) further proposed that tide-induced vertical mixing promotes the iron supply to the euphotic zone in the Bering Slope Current region and sustains local summertime high biological production.

Eddy enhanced primary production and mesoscale advection further impact the marine ecosystem in Region 2. Haida eddies affect zooplankton populations in the offshore regions through supply of shelf- and slope- origin species to offshore regions. Nearshore tracer species are often the dominant or sub-dominant zooplankton within eddies in the first summer after they leave the coast, but their abundances decline rapidly after that (Mackas and Galbraith, 2002). Shelf species of diatoms and calanoid copepods were recorded in Continuous Plankton Recorder (CPR) samples within or near Haida and Sitka eddies and they persisted through the sampling period (Batten and Crawford, 2005). The zooplankton in the Haida eddies are mostly a mixture in both abundance and community composition between the continental margin and offshore source regions, but in about 1/3 of them abundances within-eddy are higher than in either source region (Mackas et al., 2005). Zooplankton assemblages in the Haida and Sitka eddies differ in a statistically significant way, but this is not the case in the Sitka and Yakutat or Yakutat and Haida eddies (Ladd et al., 2009). In the Alaskan Stream region, Saito et al. (2014) reported that large oceanic copepods were more abundant, and some species had accumulated more lipids inside the eddies than outside, a likely reflection of the greater primary production in the eddies.

Mesoscale eddies further impact marine ecosystems by affecting fish distributions. Atwood et al. (2010) indicated that the richness of ichthyoplankton correlates with the distance from eddy centers, and assemblages within eddies are significantly different from those in surrounding basin and shelf waters in the eastern Gulf of Alaska. They also suggested that mesoscale eddies propagating along the continental shelf-break influence larval fish assemblages. In particular, eddies may influence the recruitment of sablefish by increasing nutritional availability and affecting transport (Shotwell et al., 2014), and individual based models indicate that eddies may be key to larval transport of some species (e.g., Stockhausen et al., 2019). Furthermore, Prants et al. (2019) proposed that the observed positive correlations between wind stress curl and

salmon catches in the eastern Bering Sea may result from changes in mesoscale eddy activity. A better understanding of the impact of mesoscale eddies on larval abundance, survival, and transport would contribute to more reliable estimates of recruitment success, potentially improving management decisions in the Gulf of Alaska.

3.2.4 Variability

Mesoscale eddies in Region 2 show seasonal and interannual variations. For example, for the Haida, Sitka and Yakutat eddies EKE and eddy numbers are large in winter and spring, due to their stronger downwelling winds, and low in summer and autumn (Ladd, 2007; Henson and Thomas, 2008). The interannual variability is related to the variability in downwelling winds (Ladd, 2007; Henson and Thomas, 2008) and also to the number of gap-wind events (Ladd and Cheng, 2016). Long-term trends in eddy formation have not been investigated yet in the Alaska Current, Alaskan Stream or Bering Slope Current regions. For the Alaskan Stream, however, Prants et al. (2013) indicated the water flux across the Near Strait, connecting the Pacific Ocean with the Bering Sea, is highly variable and controlled by mesoscale and submesoscale eddies. Interannual variations of heat and freshwater transport for Haida, Sitka, Yakutat and Kenai eddies have been estimated, but solely owing to fluctuations in the number of eddies that are observed in SSH to cross 150°W or offshore of 200 km from the 1000 m isobath (Lyman and Johnson, 2015). Lyman and Johnson (2015) also indicated that heat and freshwater content anomalies of cyclonic eddies are an order of magnitude smaller than those of anticyclonic eddies.

In the Bering Slope Current region, eddy activity is negatively correlated with the North Pacific Index, a measure of the strength of the Aleutian Low (Trenberth and Hurrell, 1994), and positively correlated with Pacific North America Pattern (Barnston and Livezey, 1987). North Pacific Index and Pacific North America Pattern are associated with the wind stress curl over the North Pacific (Ishi and Hanawa, 2005). Indeed, it is suggested that the wind-driven gyre spin-up during the preceding winter may lead to increased eddy activity in the Bering Sea in boreal spring (Ladd et al., 2012). Furthermore, Prants et al. (2019) have shown that eddy activity in the eastern Bering Sea and Alaskan Stream is related to wind stress curl in the northern North Pacific in November-March. Finally, Ding et al. (2018) pointed out that eddy activity increased between 1993 and 2011 across most of the Northeast Pacific. However, in

Region 2 there was longitudinal variability in the increase/decrease of eddy activity with a positive trend around 160°W and negative trend around 170°W (see Ding et al., 2018's Fig. 2).

The effects of mesoscale eddies on ecosystem variability have been studied mainly for upper trophic levels. For example, eddy activity influences the distribution of adult female northern fur seals during their migration between the Bering Sea and the California Current region (Pelland et al., 2014; Ream et al., 2005). Fur seal behavior (foraging) has also been associated with eddy activity, but the quantitative relationship remains unclear and requires further examination of subsurface mesoscale oceanic structures along with behavioral responses or lack thereof. When it comes to bird foraging, black-legged kittiwakes, breeding at the Pribilof Islands in the Bering Sea, often fed near anticyclonic, or inside cyclonic eddies (Paredes et al., 2014). Paredes et al. (2014) also showed that in 2010 high eddy activity in the Bering Sea, as measured by EKE, coincided with a 63% increase in foraging in the basin by birds from St. Paul compared to 2008, when the EKE was low.

3.2.5 Summary

In the subpolar North Pacific, mesoscale eddies form frequently in the Alaska Current region, at the western edge of the Alaskan Stream region and in the Bering Slope Current area. They typically form by baroclinic instability associated with winds, coastal water outflow, or local topography. In the Alaska Current and Alaskan Stream regions, anticyclonic eddies have longer lifetime and greater size than cyclonic eddies, and most studies to date have focused on them (Table 2). They are named Haida, Sitka, Yakutat, Kenai, Alaskan Stream, and Aleutian eddies, according to their formation area. In the Bering Slope Current region, eddies of both polarities have been investigated. Haida, Aleutian, Bering Slope Current, and some Sitka eddies propagate offshore from the shelf into the basin and transport coastal water to the ocean interior. Yakutat, Kenai, Alaskan Stream and some of the Sitka eddies mostly propagate along the Alaska Current and Alaskan Stream, without detaching from the Alaskan Stream, and therefore do not propagate offshore. All these mesoscale eddies contribute to the transport of iron from the coastal/shelf area to the basin interior and support biological production in the iron-limited HNLC North Pacific subarctic and Bering Sea basins. Mesoscale eddies in Region 2 further impact zooplankton and higher trophic level species such as fish, marine mammals and birds mainly through elevated primary production due to macro- and micro-nutrient trapping in their core and advection into the basin or to the shelf boundaries. We summarize known/unknown

characteristics and impacts of mesoscale eddies in Table 2. Further observational and modeling studies are needed especially in relation to the ecological impacts of the Aleutian eddies. Recently, it was found that the seasonal and interannual variation of the halocline, which affects primary production through pycnocline/mixed layer depth, differ between the western and eastern parts of the subarctic North Pacific (Katsura et al., 2020; Ueno et al., 2022). This east-west contrast in background physical environment should be considered to better understand the different impacts that mesoscale eddies have on primary production within Region 2.

Table 2: Characteristics and impact on marine ecosystem of eddies in Region 2. Underlining as in Table 1.

Eddy type	Haida/Sitka/Yakutat	Kenai/ Alaskan Stream	Aleutian	BSC
Formation area	Off Haida Gwaii, Sitka and Yakutat in the northeastern Gulf of Alaska	Along the Alaskan Stream from the Kenai Peninsula to the Aleutian Islands	South of the Aleutian Islands between 170° and 175°E	Along the eastern shelf-break of the Bering Sea
Polarity	Anticyclonic	Anticyclonic	Mainly anticyclonic	Anticyclonic / Cyclonic
Formation mechanism (BC: baroclinic, BT: barotropic)	Shelf water advection (Crawford et al., 2002; Di Lorenzo et al., 2005), BC instabilities associated with waves and winds (Thomson	BT and BC instabilities (Thomson and Gower, 1998; Shore et al., 2008) and instabilities associated with winds	<u>Only limited theoretical understanding. Realistic models that allow for a direct comparison</u>	Instabilities associated with wind, topography, flow through Aleutian passes (e.g. Paluszkievicz and Niebauer,

	and Gower, 1998; Shore et al., 2008)	(Ueno et al., 2009)	<u>with in-situ and satellite observations are needed</u>	1984; Mizobata et al., 2008).
Diameter	Around 100 km on average (Henson and Thomas, 2008)	Around 150 km on average (Ueno et al., 2010; Lippiatt et al., 2011)	80–330 km (Rogachev et al., 2007; Budyansky et al., 2022)	10–200 km (Ladd et al., 2012)
Propagation	To Gulf of Alaska basin, along the Alaskan Stream (Crawford, 2002; Ladd et al., 2007)	Along the Alaskan Stream and sometimes detached from the Alaskan Stream (Ladd et al., 2007; Janout et al., 2009)	To western subarctic basin (Rogachev et al., 2007; Budyansky et al., 2022)	Off-shelf (westward/southwestward) (Ladd et al., 2012)
Lifetime (years)	Up to 5 (Crawford, 2002; Ladd et al., 2007)	Up to 5 (Ueno et al., 2009, 2012)	Up to 4 (Budyansky et al., 2022)	Up to 2 (Chelton et al., 2011b)
Heat/fresh water transport	Seaward warmer, fresher water transport (Crawford, 2005; Ladd et al., 2007)	Seaward warmer, fresher water transport, shelf-basin exchange (Okkonen et al., 2003;	<u>Unknown. Modeling and observational studies are urgently needed.</u>	Shelf-basin exchange (Kinney et al., 2009)

		Ueno et al., 2009)		
Nutrient transport	Seaward (Whitney and Robert, 2002; Ladd et al., 2009) and upward (Johnson et al., 2005) macro/micro-nutrient transport	Shelf-basin exchange (Lippiatt et al., 2011)	<u>Unknown.</u> <u>In-situ</u> <u>observations</u> <u>complemente</u> <u>d by</u> <u>modeling</u> <u>investigations</u> <u>are needed.</u>	Eddy induced cross-shelf transport (Mizobata et al., 2006; Hurst et al., 2010)
Impact on chlorophyll	Enhancing production in the Gulf of Alaska basin (Whitney and Robert, 2002; Crawford et al., 2007)	Enhancing offshore production (e.g., Ueno et al., 2010)	<u>Unknown.</u> <u>Satellite data</u> <u>analyses</u> <u>together with</u> <u>targeted field</u> <u>campaigns</u> <u>and modeling</u> <u>studies are</u> <u>recommended.</u>	Maintaining high primary productivity along the shelf edge (Springer et al., 1996; Ladd et al., 2012)
Impact on zoo/ichthyopl ankton	Affecting zoo/ichthyoplankton assemblages in Gulf of Alaska (Mackas and Galbraith, 2002, Atwood et al., 2010)	<u>Unknown.</u> <u>Field</u> <u>campaigns</u> <u>sampling</u> <u>zooplankton</u> <u>assemblages</u> <u>inside, outside</u> <u>and at the</u> <u>boundary of</u> <u>mesoscale</u>	Affecting life stage distribution of zooplankton (Saito et al., 2014)	Affecting zoo/ichthyoplankt on distribution (Schumacher and Stabeno, 1994; Springer et al., 1996)

		<u>eddies are recommended.</u>		
Impact on higher trophic levels	Fur seals (Ream et al., 2005; Pelland et al., 2014, Sterling et al., 2014), Salmon (Healey et al., 2000), Steller sea lions (Miller et al., 2005)	<u>Unknown. Ship observations, bio-logging, analysis of fishery data are recommended.</u>		Fur seals (Nordstrom et al., 2013; Sterling et al., 2014), Salmon (Prants et al., 2019), Seabirds (Paredes et al., 2014; Santora et al., 2018)

818

819 3.3. Region 3: Western boundary of the North Pacific and marginal seas

820 3.3.1 Introduction

821 Two strong boundary currents flow along the western edge of the North Pacific: Oyashio and
822 Kuroshio (Fig. 14). The Oyashio Current starts along the west coast of the Kamchatka Peninsula
823 as the East Kamchatka Current, flows along the Kuril Islands and turns to the east near
824 Hokkaido. The Kuroshio Current flows east of Luzon Island and Taiwan, entering the East
825 China Sea through the East Taiwan Channel, flows along the shelf edge of the East China Sea
826 continental shelf, enters the North Pacific Ocean off Kyushu Island, flowing along the south of
827 Japan, and turns to the east near the Boso Peninsula in the southeastern portion of the Honshu
828 Island. Mesoscale eddies are frequently generated along the Kuroshio/Oyashio and their branch
829 currents such as the Tsushima Warm Current. These eddies can influence the volume transports
830 and pathways of the current systems (e.g., Kawabe, 1995, Yin et al., 2019, Sugimoto and
831 Tameishi, 1992). The Tsushima Warm Current diverges into the northward Tsushima Warm
832 Current nearshore branch close to the east coast of Korea and the eastward Tsushima Warm
833 Current offshore branch close to the west coast of Japan. The Tsushima Warm Current
834 nearshore branch is called the East Korea Warm Current, which forms the subpolar front and
835 the Ulleung Warm Eddy when it collides with the North Korea Cold Current.

In one of the first studies in Region 3, Kitano (1975) identified 154 warm-core eddies in the Kuroshio – Oyashio frontal zone over the period 1957–1973 from shipboard observations. Some eddies moved northward and eastward along the Japan Trench from offshore of the Tohoku region and along the southern Kuril-Kamchatka Trench (Yasuda et al., 1992 and references therein). Later studies confirmed that a large number of strong eddies move poleward along the trenches or, less frequently, on a path along the Subarctic Front (e.g., Bulatov and Lobanov, 1983; Yasuda et al., 2000; Rogachev, 2000a; Isoguchi and Kawamura, 2003; Itoh and Yasuda, 2010a, b; Kaneko et al., 2015; Prants et al., 2017a). Additionally, mesoscale eddies have been studied in the marginal seas for more than 30 years. Wakatsuchi and Oshima (1990) observed trains of eddies in the Sea of Okhotsk and suggested they form via barotropic instability of the Soya Warm Current.

Dynamic processes vary across Region 3 and the impact of physical phenomena on the marine ecosystem differs depending on the area. Southward western boundary currents such as the East Kamchatka Current, Oyashio, East Sakhalin Current and Primorye (Liman) Current contribute to the eddy formation and ecosystem variability north of ~40°N, while changes in the Kuroshio and Tsushima Warm Current are large contributors to the mesoscale and marine ecosystem south of ~40°N. The Tsushima Warm Current also affects the mesoscale variability in the Sea of Okhotsk as the Soya Warm Current and the Kuroshio Warm Core Rings can at times reach the area east of Hokkaido.

Eddies associated with the Oyashio and Kuroshio currents modulate the saury fishing grounds which are located preferentially along fronts where the productive cold waters of the Oyashio Current, the warmer waters of the southern branch of the Soya Warm Current, and waters of warm-core Kuroshio eddies converge (Prants et al., 2014a; Prants et al., 2021). In the Kuroshio region, Kasai et al. (2002) suggested that eddies are important for fish recruitment because they are quasi-stationary and remain in the coastal nursery area.

3.3.2 Physical properties

In the northwest Pacific, east of the Kamchatka Peninsula, Kuril Islands and Japanese Islands, the Oyashio and/or Kuroshio Currents contribute to the generation of a variety of mesoscale eddies. These eddies, also known as rings, can be quasi-stationary or propagate away from their formation regions, transporting mass, momentum, salt, heat, chemical and biological constituents over large distances. Some of them form in the Tohoku region, where the waters of

the Tsugaru Warm Current mix with Oyashio and Kuroshio waters. The Tohoku eddies slowly drift along the Japan Trench to the north against the mean flow (Isoguchi and Kawamura, 2003; Itoh and Yasuda, 2010a, b; Prants et al., 2014b; Kaneko et al., 2015; Prants et al., 2017a). Some rings that split off from the western meander of the Kuroshio Extension also drift along this path. Those that split off from the eastern meander, propagate westward under the influence of the planetary beta effect, reach the Japan Trench, and then turn northward to Hokkaido. The eddies that reach the eastern coast of Hokkaido Island, so-called Hokkaido anticyclonic eddies often stagnate for a while and then drift along the Kuril-Kamchatka Trench to the northeast (Fig. 15) reaching as far as Bussol' Strait, which is the deepest strait connecting the Sea of Okhotsk with the ocean (Bulatov and Lobanov, 1983; Yasuda et al., 2000; Itoh and Yasuda, 2010a, b; Prants et al., 2014a; Prants et al., 2018).

Mesoscale eddies are frequently observed along the Japan and Kuril–Kamchatka Trenches, either quasi-stationary or propagating along the trenches against an opposing mean surface flow (e.g., Kitano, 1975; Yasuda et al., 2000; Itoh and Yasuda, 2010a, b; Kaneko et al., 2015; Prants et al., 2018). Using a Lagrangian eddy monitoring technique (Prants et al., 2017a, d, 2018), it has been possible to investigate in detail life histories of individual eddies including time and locations of their formation, splitting, merging, decay, as well as retention and release of water due to interactions with currents and other eddies. Figs. 15a and 15b show altimetry-based velocities and the Lagrangian origin map, respectively, for August 1, 2010 to exemplify the characteristics of the eddy activity in the area. The warm-core rings with red core in Fig. 15b pinched off from the meanders at the northern flank of the Kuroshio Extension. These rings then propagate westward due to the planetary beta effect (KR1 in Fig. 15b) and once they reach the steep slope of the Japan Trench, they generally turn northward (KR2 to KR3 in Fig. 15) along the Trench (Qiu and Chen, 2005; Sugimoto and Hanawa, 2012; Prants et al., 2014b).

The planetary beta effect should induce a southwestward advection of eddies. Nevertheless, the trench eddies often move northeastward against the southwestward mean flow. The underlying physical mechanisms are complex and are not yet well understood, but the topographic beta effect within the trench appears to dominate the planetary beta effect, and the deep northeastward current at the offshore trench edge guides the eddies poleward (Yasuda et al., 2000).

Kuril eddies form along the Kuril–Kamchatka Trench on the oceanic side of the islands and are characterized by a cold and fresh core. The most energetic among them are the long-lived Bussol’ eddies (e.g., Bulatov and Lobanov, 1983; Rogachev, 2000b; Yasuda et al., 2000; Rabinovich et al., 2002; Prants et al., 2016), formed near the Bussol’ Strait. Lagrangian maps indicate that the waters in the Bussol’ eddy consist of a mixture of cold and fresh water from the Sea of Okhotsk and warmer and more saline water from the East Kamchatka Current (Prants et al., 2016) (shown by green and blue colors, respectively, in Fig. 15b). Eddies formed to the north of the Bussol’ Strait drift southwestward with the Oyashio Current and eventually decay, while large-scale long-lived warm-core eddies can reach the area, moving from Hokkaido to the northeast along the trench (Yasuda et al., 2000; Prants et al., 2018). The Bussol’ Strait area is therefore a meeting point for eddies drifting from the south and the north, as suggested by earlier studies based on SST, current meter mooring and model data (Bulatov and Lobanov, 1983; Isoguchi and Kawamura, 2003).

The northern Kuril–Kamchatka Trench is a catching area for the so-called Kamchatka eddies (eddies KE1 and KE2 in Fig. 15) that move here from the north following the path of the East Kamchatka Current (e.g., Solomon and Ahlnäs, 1978; Stabeno et al., 1994, 1999; Rogachev and Shlyk, 2019). These eddies form due to the intricate coastline of the Kamchatka Peninsula and its irregular nearshore bathymetry that trigger instabilities in the East Kamchatka Current (L’Her et al., 2021). The shedding of anticyclonic eddies occurs mostly between the current and the coast and that of cyclonic eddies offshore of the current. The Kamchatka eddies occasionally merge with quasi-stationary trench eddies and persist from a few months to more than a year, until they eventually shed from the slope and are dispersed into the mean flow (Prants et al., 2020).

In the Kuril Basin located in the southwestern Sea of Okhotsk (Fig. 15a), large (100–200 km) anticyclonic eddies can be seen as spiral structures in ice floe patterns, surface drifters and in satellite SST, altimetry, and imagery (e.g., Wakatsuchi and Martin, 1990; Bulatov et al., 1999; Ohshima et al., 2002; Mitnik and Dubina, 2019; Zhabin and Andreev, 2019). These eddies (see Fig. 16) have lifetimes from several months to more than a year, being quasi-stationary or slowly moving northwestward or westward (Bulatov et al., 1999). Ohshima et al. (2005) indicated, through numerical model experiments, that the eddies originating from the area adjacent to the Kuril Straits are generated by baroclinic instability at the front that form between the water around the straits, which is relatively uniform due to the strong tidal mixing, and the

offshore water. In the southwestern Kuril Basin, the negative wind stress curl, which prevails during the summer season, intensifies the anticyclonic eddies (Zhabin and Andreev, 2019). In the nearshore zone of the Kuril Islands, mesoscale eddies between 35–70 km in diameter are generated by the interaction of large-scale and tidal currents, which are very strong, with topography (Mitnik and Dubina, 2019; Nakamura et al., 2012; Zhabin and Andreev, 2019).

The Soya Warm Current flows into the Sea of Okhotsk through the La Perouse (Soya) Strait and then along the northeastern Hokkaido coast (Fig. 14). Eddies often interact with the Soya Warm Current, and their water turns warmer than ambient (Wakatsuchi and Martin, 1990; Uchimoto et al., 2007; Zhabin and Andreev, 2019). Trains of eddies, with diameters of 45–50 km, have been observed at the boundary between the Soya Warm Current and the colder and less saline offshore water. Barotropic instability of the Soya Warm Current has been suggested as a mechanism for their formation (Wakatsuchi and Oshima, 1990). Saline water can be advected by the offshore Soya Warm Current streamers along the eddies' edges towards the deep Kuril Basin (Zhabin and Andreev, 2019). The colder, nutrient-rich water of the East Sakhalin Current can also be caught in these eddies (Prants et al., 2017c). Overall, eddies in the central Kuril Basin can hold both Soya Warm Current water and colder and fresher water originating from the tidal zone near the Kuril Islands, while eddies in the northern Kuril Basin usually have a cold surface core that traps tidal water, especially in summer (Bulatov et al., 1999).

Around the Subpolar Front located $\sim 40^\circ\text{N}$, between the Korean Peninsula and the Japanese Islands, frontal eddies have been found in hydrographic surveys, satellite images (Sugimoto and Tameishi, 1992, Takematsu et al., 1999) and in a circulation model (Prants et al., 2015). At least some of them may live over the deep Japan Basin for a few months and they provide an effective mechanism for cross-frontal exchanges in the Subpolar Front. The existence of meridional cross-front transport corridors has been revealed by statistical analysis of Lagrangian transport based on the AVISO satellite altimetry currents (Prants et al., 2017b).

During the cold season, a cluster of warm anticyclonic eddies is regularly detected off the northeastern coast of the Korean Peninsula (e.g., Sugimoto and Tameishi, 1992; Danchenkov et al., 2006). These eddies are forced by the local anticyclonic stress curl of the northwestern wind, which is typical during the winter monsoon (Yoon and Kim, 2009). Warm eddies are also

observed in spring and summer, likely generated by the local anticyclonic stress curl of the western and southwestern winds (Trusenкова et al., 2009).

Warm eddies form along the west coast of the Japanese Islands and move westward due to the planetary β -effect or with the intermittent westward Tsushima Warm Current branches off the Tsugaru Strait (Takematsu et al., 1999). They carry salty water from the Tsushima Warm Current that flows off the Japanese Islands. When the water mass cools, it subducts while moving westward, and contributes to the high salinity intermediate water (Danchenkov et al., 2006). Anticyclonic eddies are also frequently detected between the coast and the cold western boundary Primorye Current and North Korea Cold Current (Prants et al., 2011; Ladychenko and Lobanov, 2013). They form through shear instabilities and slowly move southwestward, with translation speeds lower than the surface velocity of the Primorye Current (Prants et al., 2011). In Peter the Great Bay, off Vladivostok, mesoscale eddies carry shelf fresh and warm water to the southwest, into the deep Japan Basin, where they become a source water for the intermediate salinity minimum of the basin (Ladychenko and Lobanov, 2013). Additionally, brine rejections during episodes of sea-ice formation in winter form a cold and saline water, which sinks to intermediate and deep layers (Kim et al., 2002; Talley et al., 2003). Mesoscale anticyclones and submesoscale cyclones then carry this cold and saline water offshore (Fayman et al., 2019a, b).

The regional circulation around the Korean Peninsula (Fig. 17) is described by (1) tide-dominant currents and seasonal northward warm currents in the Yellow Sea (Hwang et al., 2014), (2) low-frequency geostrophic flows and the subpolar front along the east coast of Korea associated with regional boundary currents with distinct water temperature difference (Danchenkov et al., 2006, Yoo et al., 2018, Lee et al., 2019), and (3) a mixture of tide-dominated, geostrophic currents, intermittent fronts and eddies to the south of the Korean peninsula (Kim et al., 2000a; Lie et al., 2015). Tidal fronts and eddies associated with islands and coastal boundaries (Lie, 1989; Hwang et al., 2014; Jeong et al., 2009), the migration of persistent and intermittent fronts over continental shelves (Son et al., 2010; Yang et al., 1998), and submesoscale frontal eddies associated with baroclinic instability (Yoo et al., 2018) have all been observed in this region. An anticyclonic eddy known as the Ulleung Warm Eddy is frequently observed east of the Korean Peninsula, centered at $37^{\circ} - 38^{\circ}$ N, with a diameter of at least 150 km and a lifetime from several months to more than one year (Chang et al., 2004). It is generated within the anticyclonic meander of the East Korea Warm Current that flows

northward as part of the subpolar front (Yoshikawa et al., 2012). The Ulleung Warm Eddy surface signature is often a cold core wrapped by a warm filament (Chang et al., 2004).

Lastly, we summarize the mesoscale processes around Kuroshio. To the east of Taiwan, Kuroshio intrusions represent an important mechanism of water exchange between the East China Sea shelf and the western North Pacific Ocean (Matsuno et al., 2009). Kuroshio intrusions are linked to the formation of a cold dome (Shen et al., 2011), a cyclonic circulation that occurs northeast of Taiwan. Kuroshio and its intrusions in this area are affected by the open ocean eddies detached from the Subtropical Counter Current (STCC) and the North Equatorial Current, as described in Sections 3.3.3. Upwelling of subsurface water occurs around the Kuroshio in the East China Sea (e.g., Wong and Zhang, 2003), and eddy-induced lateral mixing due to the Kuroshio around the Ryukyu Islands in the East China Sea has been recently investigated using a submesoscale-eddy-resolving model (Kamidaira et al., 2017).

The Kuroshio follows the shelf break of the East China Sea and the south coast of Japan, passing through Tokara Strait and Izu Ridge, and then moves offshore to the Kuroshio Extension region. In recent years, several studies focused on sub-mesoscale processes associated with flow-topography interactions around the ridges (e.g., Nagai et al., 2021, Hasegawa et al., 2021). Along the south coast of Japan, the Kuroshio Current follows two major paths, characterized by a large or small meander (e.g., Kawabe, 1995). In the small meander configuration, a cyclonic eddy propagates the Kuroshio path to the east (Nagano and Kawabe, 2004). In the large meander pathway, a large cyclonic circulation forms between the meandered Kuroshio and the coast, as captured by satellite data analysis (e.g., Ebuchi and Hanawa, 2000; Kobashi and Kawamura, 2001).

3.3.3 Impact on biogeochemical and marine ecosystem fields

In the open North Pacific, the Kuroshio rings transport subtropical water with biogeochemical properties that differ from ambient waters (e.g., Tomosada, 1986; Sugimoto and Tameishi, 1992; Kaneko et al., 2015). Some rings reach subpolar latitudes up to 46.5°N at Bussol' Strait (Bulatov and Lobanov, 1983; Rogachev, 2000a; Yasuda et al., 2000; Itoh and Yasuda, 2010a) and greatly modify fishing grounds and ecosystems (Saitoh et al., 1986; Sugimoto and Tameishi, 1992; Prants et al., 2021). For example, the ring WCR86, formed in 1986 at approximately 37°N (Sugimoto and Tameishi, 1992), moved along the Japan and Kuril–Kamchatka Trenches and reached as far as 46.5°N at the Bussol' Strait; despite having travelled

a long way from its origin, it still contained warm, salty water in its upper layer (Yasuda et al., 2000; Rogachev, 2000a). Offshore of southeastern Hokkaido, warm core rings often block the coastal branch of the nutrient-rich Oyashio Current causing a change of the migration routes of pelagic species and a shift of the richest fishing grounds offshore (Saitoh et al., 1986; Sugimoto et al., 1992; Prants et al., 2014a; Prants et al., 2021).

The large Kuril eddies on the Pacific Ocean side of Kuril Islands influence transport pathways, water masses, nutrient distributions and chlorophyll concentrations. Near the coast, they can block the coastal flow and thus intensify the offshore branches of the East Kamchatka and Oyashio Currents (see Figs. 15 and 16). Due to strong mixing, the Kuril eddies affect the distributions and vertical fluxes of dissolved oxygen, nutrients and dissolved inorganic carbon in the Oyashio region, and thus the plankton blooms. Using in situ observation and satellite data, it has been found that the boundaries of the Kuril eddies are enriched in nutrients and characterized by high biological productivity (Kusakabe et al., 2002).

The impact of meso- and submeso-scale processes on the marine ecosystem is remarkable in the northwestern Sea of Okhotsk, namely in the Sakhalin Bay, where the Amur River enters the Sea (Zhabin et al., 2010). The region northwestward of the Sakhalin Bay around the Bolshoy Shantar Island is highly productive; e.g., it is a bowhead whale feeding ground (Rogachev et al., 2008). It has been shown that submesoscale eddies in this area are generated by the interaction of Amur River freshwater, strong tidal currents, and topographic torque induced by the bathymetry, resulting in intensive vertical mixing and abundant nutrient supply to the euphotic layer (Rogachev et al., 2008; Zhabin et al., 2010).

Around the Korean Peninsula, the temporal and spatial variability of surface phytoplankton blooms has been observed using ocean color (e.g., Kim et al., 2000b; Yamada et al., 2004, 2005; Kim et al., 2007; Park et al., 2012b; Shi and Wang, 2012; Lee and Kim, 2018). Park et al. (2012b) showed that satellite Chl-a concentrations correlate with numerous eddies during the phytoplankton spring bloom in the East Korea Warm Current region. These data also indicate that along the southeast coast of the Korean Peninsula, the biological productivity is enhanced mainly by wind-driven upwelling (Yoo and Park, 2009), submesoscale eddies and shear currents (Lee and Kim, 2018). In summer, surface Chl-a is low in the Ulleung Warm Eddy interior due to the strong vertical stratification and nutrient depletion inside the eddy. However, a strong Chl-a maximum has been observed in the subsurface layer, which is attributed to an

intrathermocline eddy formed within the Ulleung Warm Eddy mixing surface and deep-ocean waters and supplying sufficient nutrients. This suggests a key ecological role of the Ulleung Warm Eddy (Kim et al., 2012). Moreover, the regional spring and fall blooms are more significant within 40 km of the coast than offshore because of more energetic submesoscale horizontal shear and vortical phenomena onshore as well as their propagation in the cross-shore direction. Specifically, the regional spring bloom starts offshore and migrates onshore with a time delay of about one month, resulting from the onshore propagation of geostrophic currents, the deepening of the mixed layer, and favorable nutrient fluxes from the subsurface (Lee and Kim, 2018).

To the northeast of Taiwan, the Kuroshio intrusions to the East China Sea shelf and its mesoscale dynamics sustain the local biogeochemical fluxes, biological productivity, spawning grounds, larval transport routes, and habitat conditions (temperature and food) of commercial fish (e.g., Liu et al., 1992; Gong et al., 1997; Liu et al., 2010; Sassa et al., 2008; Sassa and Tsukamoto, 2010). Mesoscale eddies are also likely to contribute to the indirect transport of terrigenous particles from the East China Sea shelf and slope to the southern Okinawa Trough and the Pacific Ocean, as evidenced by the observation of suspended matter collected in August 1994 in a cyclonic eddy, linked to the Kuroshio intrusion (Hsu et al., 1998). Wong and Zhang (2003) surveyed the distribution of iodine, governed by biological activities, in a transect across the southern East China Sea shelf off the northeast of Taiwan, and suggested that the exchange between the open ocean and marginal sea due to mesoscale fluctuations of the Kuroshio is an important source. Miller et al. (2002) found that the widespread distribution of marine eels in the East China Sea is linked to the transport by the Kuroshio and Tsushima Warm Current. Sassa et al. (2008) and Sassa and Tsukamoto (2010) reported that the spawning ground and habitat condition (temperature and food) of larval fish (jack mackerel and chub mackerel) vary from year to year due to the warm Kuroshio intrusion in the northern Taiwan.

The southern coast of Japan is a major spawning ground for pelagic fishes such as sardines and anchovies (e.g., Kimura et al., 2000; Nakata et al., 2000; Kasai et al., 2002), due in large part to mesoscale activity. Mesoscale eddies induced by frontal disturbances of the Kuroshio Current can stimulate local biological production that results, in turn, in an increase of the number of anchovy larvae (Kimura et al., 2000). These eddies also have implications for larval transport, distribution and food availability for anchovy in the Enshu-nada Sea (Nakata et al., 2000). Kasai et al. (2002) showed that the primary production in the subsurface layer is

enhanced when the nutrient-rich coastal water is entrained into Kuroshio frontal eddies. Furthermore, the Kuroshio Current plays a crucial role in supplying nutrient to the shelf-slope region off the southern coast of Japan (e.g., Kuroda et al., 2018). Kuroda et al. (2014, 2018) investigated through modeling the circulation of Tosa Bay and revealed that when the Kuroshio takes a stable near-shore path, nitrate was frequently uplifted around the Kuroshio front and horizontally transported along the front and into the bay via the counterclockwise circulation within the bay, and at times it was further uplifted onto the shelf.

3.3.4 Variability

The coastal areas of the western North Pacific Ocean underwent significant warming in the past three decades. Its manifestation is a decreasing trend in seasonal sea ice in the Sea of Okhotsk, since, at least, the 1970s. As reported by the Japan Meteorological Agency, the reduction of the maximum sea ice extent from 1970 onwards is equivalent to a 3.9% decrease of the ice cover per decade (https://www.data.jma.go.jp/gmd/kaiyou/english/seaice_okhotsk/series_okhotsk_e.html). The positive SST trend results from the weakening of the East Asian winter monsoon, in turn linked to the abrupt decline of the Siberian high, the cold-core high pressure system that modulates air temperatures in boreal winter with a center located in the northwestern Mongolia region (Park et al., 2012a). The SST warming during winter then persists throughout the year in the coastal areas. Since 1990, the ocean dynamical response to the Aleutian Low variability has strengthened, and the PDO influence on SST in Region 3 has increased. This increase in the relative role of ocean (PDO-mediated) versus atmospheric (East Asian winter monsoon-mediated) influences has been interpreted as a response to the weakening of the Siberian high-pressure system. To our knowledge, the impacts of these changes on mesoscale variability, eddy formation and ecosystem functioning in Region 3 have not been investigated.

To date, the interannual to seasonal variability in eddy activity in Region 3 has only been studied in the upstream Kuroshio region (Chang et al., 2015; Yin et al., 2017; Yin et al., 2019). Chang et al. (2015) concluded that the change of the Kuroshio transport is locally caused by mass convergence and divergence produced by the mesoscale eddies and is therefore non-uniform. In eddy-rich years, the Kuroshio is generally stronger upstream. The large number of warm eddies to the east of Taiwan and Luzon Island further strengthens the jet, whereas the larger number of cold eddies to the east of the Luzon Strait weakens the Kuroshio in the Luzon

Strait. Yin et al. (2017) analyzed the impact of mesoscale eddies on the Kuroshio intrusion variability northeast of Taiwan, finding that cyclonic (anticyclonic) eddy increase (decrease) the Kuroshio intrusion. Yin et al. (2019) found that the seasonality of the relative number of anticyclonic and cyclonic eddies, which strengthen the Kuroshio between June and August and weaken the Kuroshio between February and April in the East Taiwan Channel, is determined by the number of cyclonic eddies.

The Kuroshio and its intrusions northeast of Taiwan are affected by open ocean eddies from the STCC and North Equatorial Current regions. The Kuroshio volume transport at the entrance to the East China Sea has a remarkable 100-day variability of ± 10 Sv (Johns et al., 2001), which has been attributed to open ocean mesoscale eddies (Zhang et al., 2001). Anticyclonic (cyclonic) eddies enhance (weaken) the Kuroshio transport, as shown through observations (Vélez-Belchí et al., 2013) and modelling (using HYCOM, Yin et al., 2017). Additionally, in the model it has been found that the arrival of cyclonic eddies introduces a positive potential vorticity flux and weakens the cross-slope potential vorticity gradient, favoring the onshore intrusion of Kuroshio water. The modulation of the Kuroshio front is important for the cross-shelf exchange of heat, salt and nutrients. The seasonal cycle in SST in the Bohai Sea, Yellow Sea and East China Sea depends on the frontal structure and physical mechanisms responsible for the front formation, e.g., tidal mixing, water mass convergence, river runoff, and deep-water mass convection (Belkin et al., 2009; Shi and Wang, 2012).

Summer marine heatwaves have been observed during 2010–2016 in the Oyashio region, and may be responsible for an increase of yellowtail catch in the waters surrounding northern Japan (Miyama et al., 2021). The heatwaves might have been caused by a weakened southward intrusion of the Oyashio near the coast, accompanied by an increase in anticyclonic eddies from the Kuroshio Extension (Miyama et al., 2021). Interactions between the Oyashio and mesoscale eddies result in an interdecadal shift in the cold-water intrusion of the Oyashio to the area off the southwestern coast of Hokkaido, and large areas of favorable potential fishing grounds near the Hokkaido coast have disappeared in the 2010s (Kuroda and Yokouchi, 2017; Prants et al., 2021). Although the variability of the western Oyashio front may be controlled by decadal modulations of mesoscale activity in the upstream Kuroshio Extension (Qiu et al., 2017), further studies are needed to identify the driver for the warming in summer and increased anticyclonic eddies after 2010 (Miyama et al., 2021). Using 1993–2020 altimetry data, Trusenkov and Kaplunenko (2022) demonstrated the existence of quasi-biennial (in 1995–1998), interannual

(3–7 years; since 2002), and decadal (8–10 years; in 2002–2014) fluctuations of EKE in the area west of Hokkaido between the Korean Peninsula and the Japanese Islands.

3.3.5. *Summary*

Mesoscale eddies in Region 3 are influenced by coastal processes such as tidal and topographic mixing, sea ice formation and melting, typhoon upwelling, weather events and terrestrial inputs of fresh water. The interaction of eddies, often generated over the shelves, with the western boundary currents occurs frequently and characterizes the circulation of this region. The Japan and Kuril–Kamchatka Trenches tend to catch mesoscale eddies, which stagnate or propagate over the trenches. Eddies are essential to exchanging waters between adjacent enclosed basins and the open ocean, inshore to offshore, and shallow to deep. Owing to the technological developments of satellite remote sensing, advances in modeling, newly developed analytical methods such as Lagrangian eddy monitoring, and tremendous efforts in in-situ observations, a large amount of information on the eddy activities and their influences has been accumulated for Region 3. Several questions are still open due to the complex interactions of multiscale, nonlinear and highly diverse processes, including eddy-eddy interactions, eddy-merging and dipole-eddy processes.

The marine ecosystems in the western North Pacific and marginal seas are influenced by local changes in the convergence of water masses around eddies on the open-ocean fronts and upwelling and tidal mixing on the coastal fronts, as well as the interactions and exchanges between the local/coastal ecosystems and the open ocean ones. In recent years, changes attributable to global warming have become more apparent, including an increase in the frequency of extreme events in both the atmosphere and the ocean that threatens marine ecosystems. Climate projections show monotonic increases in ocean warming and frequency of extreme events, such as marine heat waves.

Again, we summarize main eddy characteristics in the Table below. In this region the physical properties have been generally sampled and quantified, while open questions remain on the overall impacts on the ecosystem, especially on zooplankton and higher trophic levels. Given the economic and social role played by fishery in this area, it is imperative to close these gaps with targeted campaigns and repeated and mooring observations, noting that simulations of zooplankton dynamics would also benefit greatly from the data collected. Constraining plankton cycling will help forecasting climate change impacts on the fishery of Region 3.

1178 Table 3: Characteristics and impact on marine ecosystem of eddies in Region 3. Underlining
1179 as in Table 1.

Eddy type	Cold dome off northeastern Taiwan	Kuroshio small meander	Ulleung Warm Eddy	Coastal eddies off the northeastern coast of Korea	Eddies along the Kuril – Kamchatka Trench*	Eddies in the Kuril Basin
Formation area	Off northeastern Taiwan	Kuroshio south coast of Kyushu	East of the Korea Peninsula	Coastal areas near the East Coast Warm Current	Local or within the Kuroshio Extension	Local
Polarity	Cyclonic	Cyclonic	Anticyclonic	Anticyclonic/Cyclonic	Anticyclonic	Anticyclonic/Cyclonic
Formation mechanism	Interaction between Kuroshio current and East China Sea shelf (Shen et al., 2011)	Advection of positive PV caused by the propagation of cyclonic eddies (Usui et al. 2008)	Potential vorticity balance (Jo et al., 2017)	Baroclinic instability (Yoo et al., 2018)	Boundary current instability (Yasuda et al., 2000)	Baroclinic / barotropic instability, tidal processes, topographic torque (Ohshima et al., 2005)
Diameter	70 km (Hsu et al., 1998)	50–100 km (Nagano and Kawabe, 2004)	86 km (average, Lee et al., 2019)	3–15 km (Yoo et al., 2018)	100–350 km (Yasuda et al., 2000)	~100 km (Ohshima et al., 2005)
Propagation	Stationary (Chuang et al., 1993)	Eastward (Nagano and	Quasi stationary	Onshore and offshore propagations	Stationary or moving towards the Bussol' Strait	Quasi stationary (Wakatsuc

		Kawabe, 2004)	(Chang et al., 2004)	(Yoo et al., 2018)	(Isoguchi and Kawamura, 2003; Prants et al., 2020)	hi and Martin, 1991; Ohshima et al., 2002)
Lifetime	Summer	1 – a few months (Nagano and Kawabe, 2004)	From several months to more than a year (Chang et al., 2004)	A few days to a few weeks (Yoo et al., 2018)	From several months to more than a year (Rogachev, 2000a; Prants et al. (2020);	From several months to more than a year (Bulatov et al., 1999)
Heat/freshwater transport	Decreasing surface temperature due to vertical mixing (Shen et al., 2011). <u>Further evaluations to verify generality are needed.</u>	Affecting winter mixed layer in the shelf-slope waters facing the Kuroshio (Kuroda et al., 2014)	<u>Unknown. Heat and salt budget analysis based on data assimilated products that capture the variability of the eddy field is recommended.</u>	Vertical transport (Yoo et al., 2018)	Resulting in latitudinal transports (Itoh and Yasuda, 2010b)	Transporting Soya Warm Current and waters around Kuril Islands offshore (Bulatov et al., 1999)
Nutrient transport	Offshore transport of terrigenous	Supplying nutrient to the shelf-slope	Supplying nutrient to the	Vertical transport (Lee and Kim, 2018)	Advecting coastal high productivity waters offshore	Offshore transport of

	material (Hsu et al., 1998)	region (Kuroda et al., 2018)	subsurface layer (Kim et al., 2012)		(Kusakabe et al., 2002)	nutrients (Zhabin and Andreev, 2019)
Impact on chlorophyll	Enhanced through upwelling (Gong et al., 1997)	Enhanced (Kimura et al., 2000; Kasai et al., 2002)	Enhanced (Kim et al., 2012)	Correlated with submesoscale horizontal shear, vortical phenomena, their cross-shore propagations (Lee and Kim, 2018)	Advecting coastal high productivity waters offshore (Kusakabe et al., 2002)	Enhanced (Zhabin and Andreev, 2019)
Impact on zoo/ichthyoplankton	Affected through upwelling (Sassa and Tsukamoto, 2010)	Impacting on the spawning ground for pelagic fishes (Kimura et al., 2000; Kasai et al., 2002)	Influencing latitudinal heterogeneity of plankton (Kang et al., 2004)	<u>Unknown.</u> <u>Expected to yield blooming of zooplankton; further studies are recommended.</u>	<u>Unknown. Field campaigns targeting zooplankton assemblages in and around mesoscale structures are recommended.</u>	
Impact on higher trophic levels	Fishing (Shen et al., 2011)	Anchovy recruitment (Nakata et al., 2000)	<u>Unknown.</u> <u>Expected to yield slow blooming of the higher trophic levels; in-situ concurrent observations</u>	<u>Unknown.</u> <u>Expected to yield rapid blooming of the higher trophic levels; in-situ concurrent observations (e.g., physical quantities and</u>	Pacific saury (Prants et al., 2014a; Prants et al., 2021)	<u>Unknown.</u> <u>Analysis of fishing data and oceanographic data is recommended.</u>

			<u>(e.g., physical quantities and fishing data) are required.</u>	<u>fishing data) are required.</u>		
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*Including warm-core ring (eddy), Kuril eddy, Bussol' eddy and Kamchatka eddy.

3.4 Region 4: Extratropical open North Pacific

3.4.1 Introduction

Region 4 covers the extratropical open North Pacific (Fig. 1 and 2). It includes the KOE, the NPC, the STCC and the area east of 170°W. While eddies in these four areas often interact with each other, here we characterize their mesoscale features separately, as their background conditions, such as flow field and stratification, are considerably different.

The KOE is the downstream part of the Kuroshio and the Oyashio currents, flowing eastward after separating from Japan at around 35°N and 42°N. It includes strong frontal boundaries between warm and salty subtropical waters, and cold and fresh subpolar waters (Yuan and Talley, 1996; Yasuda, 2003; Bishop and Watts, 2014). Intense mesoscale eddies, both warm- and cold-core rings, are often shed from these currents (Mizuno and White, 1983; Yasuda et al., 1996; Itoh and Yasuda, 2010a; Sasaki and Minobe, 2015; Dong et al., 2017). The energy of the KOE decreases as it moves eastward; east of the dateline the KOE becomes the broad NPC¹ (Qiu, 2002). Finally, the STCC is a shallow eastward current flowing between 18°N and 28°N at a speed of a few centimeters per second that is counter to the westward current predicted by wind-driven ocean circulation theory. This counter current is formed due to the combined forcing of surface wind and heat fluxes (Kobashi and Kubokawa, 2012).

¹ Although the NPC definition varies among published literature, we use here the term to indicate the downstream part of KOE.

The EKE estimated from SSH (Qiu, 2002, Fig. 18) or a high resolution numerical model (Qiu et al., 2018) is highest along the KOE, followed by the STCC. The high EKE within the KOE is partly attributed to variability of the currents themselves, but contributions from detached eddies, especially warm and cold-core rings shed from the northern and southern side of the Kuroshio Extension, are likely significant (Yasuda et al., 1992; Okuda et al., 2001; Qiu et al., 2006; Itoh and Yasuda, 2010a). The EKE level in the NPC is lower than that in the KOE, and so is the amplitude of the eddies shed by it (Itoh and Yasuda, 2010a; Dong et al., 2017). Nonetheless, moderate amplitude eddies occur frequently in the NPC (Sasaki and Minobe, 2015). They form either in this area or to the west of it (and then propagate into the region). While the formation rate of eddies generated in the NPC is low, their lifetime is long (Figs. 4 and 5).

Transport properties of oceanic systems involving strong meandering jets and associated eddies, such as KOE, NPC and the warm-/cold-core rings that detach from them, have been the subject of research and debate for more than thirty years, starting with the seminal paper by Bower et al. (1985). Do meandering jets enhance or inhibit transport in the cross-stream direction, i.e., are they “barriers or blenders”? This is an important question from both oceanographic and biological prospective because jets and eddies contribute to the redistribution of both physical and bio-geo-chemical tracers in the ocean, as well as to the transport of slowly-swimming biological organisms such as fish larvae (Rypina et al., 2014; 2016; 2019).

East of 170°W, the offshore North Pacific is largely an eddy desert (see e.g., Fig. 4 in Cheng et al., 2014). There is a long history of authors noting the large difference in eddy activity between the western and eastern North Pacific. Dantzler (1976) was among the first; he showed the amplitude of the structure function of dynamic height dropped more than one order of magnitude when comparing similar regions east and west of 170°W. Bernstein and White (1977) called the region 30°–40°N, 170°–140°W an eddy void and they also found an order of magnitude difference in eddy energy at wavelengths greater than 300 km between the eastern and western North Pacific. In subsequent years, many other authors used different datasets to demonstrate the same (e.g., Kirwan et al., 1978; Emery, 1983; Holloway, 1986). A particularly compelling instance was provided by an array of moorings showing that the EKE starts dropping east of 152°E and decreases by a factor of 4 by 165°E, and by a factor of 50 (100) by 152°W for abyssal (near-surface) data (Schmitz, 1988).

3.4.2 Physical properties

In this region, distant from shore, direct observations of mesoscale eddies have been limited compared to the other nearshore regions, with very few observations until satellite altimetry and the Argo float array became available. Since the early 1990s, the basic characteristics of eddies such as number, lifetime and formation/transit patterns have been increasingly investigated (e.g., Chelton et al., 2011b; Cheng et al., 2014).

The governing dynamics for the KOE eddies are diverse. The Kuroshio Extension is an inertial jet providing energy for large-amplitude meanders and eddies (Qiu, 2002). Using linear stability analysis Kouketsu et al. (2008) found that the unstable waves in the KOE with wavelength of ~200 km (Kouketsu et al., 2005; 2007) were caused mainly by baroclinic instability. Furthermore, baroclinic instabilities due to the intense KOE density front and barotropic instabilities due to horizontal shears north and south of the jet play important roles (Yang and San Liang, 2018; Ji et al., 2018). Within the STCC, the meridional potential vorticity gradient is positive, but down below the surface the potential vorticity gradient is negative due to the westward North Equatorial Current. Baroclinic instability occurs due to the reversal in the potential vorticity gradient, inducing high eddy activity (Qiu, 1999; Roemmich and Gilson, 2001; Kobashi and Kawamura, 2002; Chang and Oey, 2014).

Satellite-based eddy identification and tracking can be used to derive an eddy census through metrics such as number, size, amplitude, lifetime, propagation speed, and appearance/disappearance, which are often projected onto geographical/hydrographical maps (e.g., Figs. 4, 5 and 6). Such altimeter-based eddy censuses have been compiled for the KOE/NPC (Itoh and Yasuda, 2010a; Sasaki and Minobe, 2015; Ji et al., 2018) and STCC domains (Liu et al., 2012). Cheng et al. (2014) conducted a comprehensive statistical analysis on the mesoscale eddies in the North Pacific. They indicated that cyclonic eddies are more numerous than anticyclones in Region 4 and the number of eddies decreases with latitude, as well as their size, speed and travel distance in accordance with the Rossby deformation radius and the phase speed of the baroclinic Rossby waves. The lifespan of the eddies does not show latitudinal dependence. In the case of lifespan, EKE, travel distance, and amplitude, the standard deviations are as large as the mean values; therefore, the mean values have low significance (Cheng et al., 2014). These censuses update and expand earlier studies based on limited hydrographic surveys (Mizuno and White, 1983; Maximenko et al., 2001).

As mentioned, temperature and salinity profiles of the eddies in Region 4 were sparse until the development of Argo floats. Exceptions are those collected by the Japanese and Soviet Union campaigns that extended observations offshore of Region 3, in the northern part of KOE (Tomosada, 1986; Yasuda et al., 1992; Yasuda et al., 2000; Rogachev, 2000b; Maximenko et al., 2001; Komatsu et al., 2004). In the transition zone from the Kuroshio Extension to the Oyashio Extension, intense anticyclonic eddies are often observed, which have warm and saline core water with a thickness of several hundreds of meters, originating from the Kuroshio Extension (Tomosada, 1986; Yasuda et al., 1992). A Soviet hydro-physical experiment “Megapolygon-87” conducted around the subarctic frontal zone also captured several mesoscale eddies (Maximenko et al., 2001). In this area, a comparable number of warm anticyclonic and cold cyclonic eddies were observed, but the anticyclones were more distinct and persistent (Maximenko et al., 2001). Many of these anticyclonic eddies originate from the Kuroshio Extension, but there are some formed in the Sea of Okhotsk, Bering Sea and Alaskan Stream (Rogachev, 2000b; Yasuda et al., 2000; Rogachev et al., 2007), and migrate in Region 4 from Regions 2 and 3 (see Section 3.2.2 and 3.3.2 and references therein). Those from the Sea of Okhotsk or the Bering Sea are cold/fresh-core anticyclonic eddies (Rogachev, 2000b; Yasuda et al., 2000; Rogachev et al., 2007), while anticyclonic eddies originating from the Alaskan Stream have a warm and saline core (Rogachev et al., 2007).

Since the early 2000s, Argo floats have considerably increased the availability of temperature/salinity profiles within the eddies in Region 4. Combining satellite information on positioning and Argo data, it has been possible to compile profiles near the eddy center or composites of the three-dimensional structure. Itoh and Yasuda (2010b) analyzed the temperature and salinity profiles of mesoscale eddies near their center, and found anticyclonic eddies with a cold and saline core also in the KOE. In addition, some anticyclonic eddies with a distinct warm core in an upper layer have a cold and fresh core in a lower layer, which is likely formed through the interaction of eddies originating from the Kuroshio Extension and the Sea of Okhotsk.

As to how mesoscale structures impact transport in this region, evidence suggests that at the ocean surface, strong oceanic jets such as the Kuroshio act as permeable transport barriers (e.g., Rypina et al., 2011; 2013; 2018 and references therein). However, the existence of such barriers at depth is less understood and remains the subject of ongoing research. Burkholder and Lozier (2011, 2014) and Garraffo et al. (2014) showed that although jets may act as barriers at the

surface, the cross-stream transport and exchange is stronger along subsurface isopycnals. On the other hand, recent work by Cedarholm et al. (2019) suggests that along the $26.5 \sigma_\theta$ isopycnal surface (that lays deeper than 500 m between 140°E and 180°E at 30°N), fluid parcel trajectories are unlikely to cross the Kuroshio Extension west of $155^\circ\text{--}160^\circ\text{E}$, and therefore most of the simulated trajectories cross the front from the northern flank to the southern one in a limited longitudinal band east of 160°E , and then remain to the south of the front.

The suppression of the cross-jet transport near the Kuroshio Extension core, and more generally near the shearless cores of all strong oceanic jets, agrees with the strong Kolmogorov-Arnold-Moser stability arguments (Rypina et al., 2007), although the latter only rigorously applies to spatially periodic flows. It is also in line with the critical layer theory results of Ferrari and Nikurashin (2010) and Chen et al. (2014), who found limited subsurface cross-Kuroshio Extension transport west of $\sim 155^\circ\text{E}$, with enhanced meridional transport at this longitude. Eddies, therefore, play an important role in enhancing (or strengthening) cross frontal transport. Budyansky et al. (2015) demonstrated that anticyclonic Kuroshio warm core rings transported Fukushima-derived cesium, that was subducted and trapped in the subsurface core and intermediate water layers (100~500 m), northward of the subarctic front.

While EKE is weak to the east of 170°W , mooring data from 49.55°N , 138.6°W reveal that occasional eddy events still dominate the depth-averaged kinetic energy budget (Freeland, 1993). Eddies are seldom observed around the center of the eastern part of the North Pacific subarctic gyre (Cheng et al., 2014); almost all eddies generated in the North Pacific travel westward (meaning many enter this area from the coastal regions indicated as Regions 1 and 2 in this work). When this happens, eddies contribute nutrients and elevated productivity by transporting coastal, nutrient rich waters.

The exception to the east of 170°W being an eddy desert is given by the seamount generated eddies north and west of the Hawaiian Islands. As early as 1974, Bernstein (1974) confirmed that eddies existed in this area and observed them drifting westward at 4.6 cm/s using temperature observations collected northeast of the Hawaiian Islands. Shortly thereafter, Royer (1978) showed that baroclinic eddies along 158°W north of Hawaii were correlated with upstream seamounts and pointed to Huppert and Bryan's (1976) theoretical work for eddy generation by seamounts to explain their formation. Later, the analysis of mooring data from

28°N, 152°W showed that all these eddies propagated southwest-ward and derived energy baroclinically from the mean flow (Niiler and Hall, 1988).

3.4.3 Impact on chemical and biological fields

In Region 4, the relationship between eddies and the surface Chl-a concentration is spatially variable (Fig. 1). Negative correlations between SSH and Chl-a are most prominent along the stream of the Kuroshio Extension, typically west of 160°E, which corresponds to the mid-zone of the KOE. These negative correlations are primarily attributable to the eddy pumping effect. Upwelling and downwelling within cyclonic and anticyclonic eddies during their formation phase are likely responsible for the increase and decrease in primary productivity, respectively (Kouketsu et al., 2015). Because of the prominent impact of cyclonic eddies on primary productivity, their structure and contribution to nutrient transport and subsequent enhancement of primary production have been highlighted (Sasai et al., 2010; Nakano et al., 2013; Honda et al., 2018). Numerical experiments revealed that core waters of cyclonic eddies consist mainly of waters from the northern side of the Kuroshio Extension (Nakano et al., 2013), pointing to the importance of horizontal advection (eddy-trapping) in addition to vertical pumping.

Away from the center of the Kuroshio Extension, either northward, southward or eastward, the negative correlation gradually weakens (Fig. 1). This is because fewer eddies form in the areas around the Kuroshio Extension, and therefore there is less eddy-induced transport of nutrients, both vertically and horizontally (Kouketsu et al., 2015; Nakano et al., 2013). An exception is found in the southern part of the recirculation gyre (around 30°N, 140°–150°E) where slight positive correlations occur. This is partly due to horizontal advection by anticyclonic eddies that entrain nutrient-rich waters from the north (Kouketsu et al., 2015), and partly to cross-stream transport caused by quasi-steady recirculations. For the northern area of the KOE, where slight positive correlations are observed, it has been suggested that cold anticyclonic eddies that originate in the marginal seas contribute nutrient rich waters (Yasuda et al., 2000; Rogachev, 2000b; Itoh and Yasuda, 2010b; Prants et al., 2016). In the area east of 170°W, except for the region near Hawaii, the relationship between SSH anomaly and Chl-a is weak (Fig. 1). The sparsity of the eddies as well as small eddy amplitude especially in the northern part (e.g., Chelton et al., 2011b; Cheng et al., 2014) explain this weak relation.

South of the recirculation gyre (i.e. south of ~30°N), the correlation map transitions in sign near 23°–24°N and again around 17°–18°N, with correlation coefficients changing from

negative to positive, and then positive to negative again (Fig. 1). The positive band around 18°–23°N roughly corresponds to the STCC region, although the pattern is not clear at the western end. The eddy pumping mechanism does not explain the positive correlation, but it should be kept in mind that both cyclonic and anticyclonic eddies are generated along the STCC (Yang et al., 2013; Chang et al., 2017). The analysis of the cross correlation of Chl-a and eddy-induced Ekman pumping (Fig. 1d in Gaube et al., 2014) suggests that eddy current-induced Ekman pumping at the ocean surface may be responsible for the positive correlation in Fig. 1.

The eddy field near Hawaii is important biologically; decades of observations in this region suggest that anticyclonic and cyclonic eddies increase local primary productivity, change phytoplankton assemblages to favor larger species (diatoms) and increase nitrogen fixation. Venrick (1990) was among the first to point out the correlation between mesoscale patterns of dynamic topography (eddies) and chlorophyll distributions west of the Hawaiian Ridge, and postulated that eddy activity accounts for some of the variation in productivity in the Central North Pacific. Allen et al. (1996) noted that near the island of Hawaii, area-averaged photosynthetic rates were one third higher at the edge of a cyclonic eddy and 2/3 higher in the center than outside the eddy. In another study, vertically integrated nutrient and chlorophyll concentrations were shown to be enhanced in the center of a cyclonic eddy (Vaillancourt et al., 2003). The presence of eddies also impacts phytoplankton community structure, with large (>3 µm) phytoplankton enhanced within the eddy while small ones are enhanced outside of it. This effect was confirmed by Brown et al. (2008), but only for deep waters within the eddy; they observed a deep chlorophyll maximum in the center of a cold-core cyclonic eddy in the lee of the Hawaiian Islands that was comprised primarily of chain-forming diatoms. This contrasted with the upper mixed-layer populations within the eddy, which were similar to populations in the upper mixed-layer outside the eddy. The impact of anticyclonic eddies was demonstrated by Fong et al. (2008), who observed increased near-surface chlorophyll (5-fold) and increased nitrogen fixation (2–18 fold) in an anticyclonic eddy just north of Hawaii. Likewise, Guidi et al. (2012) observed increased nitrogen fixation (and high chlorophyll) caused by eddies, but at the interface between an anticyclonic and a cyclonic eddy and they attribute the increase to horizontal stirring. Eddy-eddy interaction has also been proposed for episodic injection of nutrients, perhaps due to surface frontogenesis and nonlinear Ekman pumping in the complex eddy field near Hawaii (Calil and Richards, 2010), and observed as nitrate variability by profiling floats (Ascani et al., 2013).

While cyclonic eddies in the KOE and NPC often enhance primary production through eddy pumping of nutrients, water trapping and more frequent water exchanges in the frontal structure surrounding both cyclonic and anticyclonic eddies are important for higher trophic levels as well. Warm anticyclonic eddies are often pinched off from the Kuroshio Extension and stay in the northern KOE for more than one year. They trap warm and saline water masses within their core, while interacting with the Kuroshio Extension, the surrounding Oyashio waters, and also with other anticyclonic eddies (Saitoh et al., 1986; Yasuda et al., 1992; Itoh et al., 2011; Kaneko et al., 2015). These processes are linked to the thriving of a unique community of zooplankton (Yamamoto and Nishizawa, 1986; Tsuda and Nemoto, 1992; Terazaki, 1992). As eddies in the offshore areas generally propagate westward, they are assumed to play important roles in transporting large copepods westward (Shimode et al., 2012).

Anticyclonic eddies and mesoscale fronts are also important for fish migration across the Kuroshio and Oyashio regions. Sugimoto and Tameishi (1992) revealed that various fish species, such as skipjack, mackerel, flying squid and saury use different parts of these vortical features for foraging. Since some fish species aggregates in and/or around eddies, their positions, observed through satellite altimetry, are used in forecasting fishing grounds of some species, such as Pacific saury in the northern KOE (Yasuda and Watanabe, 1994; Yasuda and Kitagawa, 1996; Prants et al., 2014a; Syah et al., 2016) and neon flying squid in the KOE/NPC (Alabia et al., 2015).

Although the trophic productivity in the STCC is generally lower than in the KOE/NPC, some migratory species use STCC eddies for their migration or larval transport. Chang et al. (2018) suggested that larval growth of Japanese eel that spawned west of the Mariana Islands is more successful if the larvae are trapped within mesoscale eddies, as they are confined within the food-rich cyclonic eddy environment. Mugo et al. (2010) found that the preferred habitat of the skipjack tuna in the western North Pacific can be successfully estimated by a geo-statistical model if information on SST, SSH and EKE from satellite remote sensing is properly accounted for. Skipjack tuna fishing sets were made in areas with low to moderate EKE, indicating that there were instances when catches were associated with eddies (Mugo et al., 2010).

In the case of the Hawaiian eddies, some observations suggest that they also impact higher trophic levels through entrapment and/or transport of zooplankton and larval fish. Several studies have found zooplankton and larval fish biomass to be enhanced inside cyclonic cold-

core eddies relative to outside around Hawaii (Lobel and Robinson, 1988; Landry et al., 2008). Lobel and Robinson (1986) used a combination of drogues and net sampling to illustrate how zooplankton and larval fish can be trapped inside an eddy; for larval reef fish, this meant that they would be retained near natal reefs long enough to complete their pelagic development phase. Later, Lobel (2011) showed that larval lizardfish were ready to settle at a younger age when a cyclonic eddy was present, indicating that the eddy influences not only retention but also larval plasticity. However, not all studies have found that cyclonic cold-core eddies positively influence fish recruitment; Fox et al. (2012) found that the presence of cyclonic cold-core eddies was negatively correlated with fish recruitment and Vaz et al. (2013) found both positive (increased connectivity between populations on the island of Hawaii) and negative (entrapping larvae and transporting them away from the coast) influences. Finally, some work has shown a positive correlation between fish catches and the presence of cyclonic eddies (e.g., Seki et al., 2002).

3.4.4 Variability

Neither altimetry data nor high resolution model outputs have long enough time series to investigate trends beyond the past two decades in Region 4. Climate models suggest that under global warming scenarios the southern half of the subtropical North Pacific Gyre will weaken while the northern one will intensify and shift northward (Cheon et al., 2012). Whether such a change would be conducive to changes in mesoscale activities has yet to be proved; to date global climate models have too coarse resolution to explicitly represent eddy processes.

Recently, Ding et al. (2018) found a significant increase in eddy activity north of the Hawaiian islands, in the Hawaiian – Emperor seamount chain (triangle with the vertices 30°N,142°W; 30°N,175°W; 48°N,175°W) over the period 1993–2011, and attributed the change to increasing eddy lifetime caused by weakening wind speed over the region, which was possibly linked to sea surface warming.

At interannual scales, the EKE in the KOE and NPC undergoes a multi-year modulation (Qiu and Chen, 2005; Nonaka et al., 2008). The EKE is inversely proportional to the strength of the Kuroshio Extension jet. When the Kuroshio Extension jet and recirculation gyre are weaker than the mean (for example during 1996–2001), the regional EKE level is high, and vice versa. This behavior results from the migration of the Kuroshio Extension jet inflow over the Izu–Ogasawara Ridge extending from Japan southward along 140°E rather than from intrinsic

baroclinic instability of the Kuroshio Extension jet (Qiu and Chen, 2005). Wang et al. (2016) also reported that baroclinic instability is not the dominant contributor, and that barotropic instability prevails.

In the STCC area, interannual variations are driven by the vertical shear between the STCC and the subsurface westward-flowing North Equatorial Current, which is due to surface Ekman temperature gradient convergence within the STCC band of 18°–28°N. Such shear induces baroclinic instability that in turn drives variability in the mesoscale eddy field (Qiu and Chen, 2010; Liu et al., 2012). In the Hawaiian Lee Countercurrent (17°N–21.7°N and 170E°–160°W), Yoshida et al. (2011) found a low-frequency modulation linked to the PDO, with greater vertical shear, enhancement of conditions favorable to baroclinic instability, and positive EKE anomalies during the positive phase of the PDO, and vice versa for periods of negative PDO.

At seasonal scales, altimetry data show that in the KOE and NPC the EKE maximum occurs in summer and the minimum in winter (Tai and White, 1990; Stammer and Wunsch, 1999; Ducet and Le Traon, 2001; Scharffenberg and Stammer, 2010; Zhai, 2017). Bishop et al. (2013) reported that the seasonal cycle in the KOE and NPC is associated with baroclinic instability based on the analysis of current- and pressure-equipped inverted echo sounders (CPIES) data. Yang and San Liang (2018) have used the HYCOM reanalysis to suggest that both baroclinic and barotropic instabilities are responsible for the observed seasonality.

In the STCC, the EKE maximum occurs in May and the minimum in December (Qiu et al., 2014; Liu et al., 2012; Chang and Oey, 2014). This is because cooling during fall and winter brings the meridional tilt of the upper thermocline at its maximum in early spring, just before warming starts; vertical shear is intensified and baroclinic instability occurs, intensifying the EKE one or two months later (Kobashi and Kawamura, 2002; Qiu, 1999; Qiu et al., 2008; Qiu and Chen, 2010; Liu et al., 2012; Chang and Oey, 2014). Furthermore, Chang and Oey (2014) proposed that the intensification of the SST front in winter and spring accelerates the eddy growth rate.

The seasonality reported from the analysis of altimetry data may be biased because its spatial resolution, order 150 km, is not enough to fully account for the mesoscale variability. High resolution modeling studies show that the timing and the strength of the seasonality signal changes among years due to the cascade of kinetic energy occurring in submesoscale eddies (Qiu et al., 2014; Sasaki et al., 2014; 2017). A recent high-resolution ocean-only model

simulation has shown that the EKE peaks around May–June for the Kuroshio Extension, around April for the STCC and in January–February for the Oyashio Extension (Sasaki et al., 2017; see their Fig. 5), with minima in fall–early winter. Considering the dominant role of baroclinic instabilities in all Region 4 subregions, the difference in the phase across the different domains may be caused by the different seasonality in stratification and/or phase speed of Rossby waves. Submesoscale permitting simulations used to investigate the interannual to decadal variations of submesoscale motions in the subtropical Northwestern and Northeastern Pacific (Sasaki et al., 2020, 2022) have shown that these variations were modulated by the PDO or ENSO.

Eddy variability in the lee of the Hawaiian Islands has been found to be strongest at seasonal and sub-seasonal scales. Calil et al. (2008) attributed it to the regional, orographic wind forcing using a regional model and modifying the wind field. Yoshida et al. (2010) confirmed this result by performing a statistical analysis of weekly satellite SSH images, and further showed that the variability of the ocean eddy population around Hawaii has distinguishable time scales at 60 and 100 days: 100-day period eddies are found west of 160°W and their variability is generated by instabilities in the shear of the mean flow (similar to the analysis by Niiler and Hall, 1988), while 60-day eddies are found close to the island Hawaii and their variability is linked to wind stress curl anomalies associated to the blocking of the trades by the island orography.

3.4.5 Summary

In the extratropical open North Pacific, the EKE is relatively strong in the southwestern part and weak in the northeastern part as suggested in Fig. 18. The eddy activity is strongest along the KOE, where an inertial jet provides energy for strong meanders and eddies, and baroclinic and barotropic instabilities due to strong fronts and shears contribute to eddy formation. Eddy activity is weaker in the NPC, the downstream end of the KOE. High eddy activity is also observed along the STCC, where baroclinic instability occurs due to the reversal in the potential vorticity gradient. East of 170°W, the offshore North Pacific is largely an eddy desert, except around the Hawaiian Islands.

Being far from the coasts, studies on the impact of eddies on biological fields in Region 4 have largely been limited to from the analysis of satellite observations. The relationship between eddies and the surface Chl-a concentration is spatially variable (Fig. 1). In the KOE and around Hawaii, there is evidence of eddies impacting species in the higher trophic level. In the last two decades, numerical experiments have become a critical tool to study the impact of

eddies on the ecosystem in many parts of Region 4. Table 4 summarizes known/unknown eddy characteristics which have large uncertainties given that in-situ data are scarcer in this Region than in the previous three. Again, knowledge of the eddy impact on zooplankton and higher trophic levels is limited, and targeted field campaigns would help constraining it.

Seasonal to interannual variations of mesoscale activity have been studied mainly by satellite altimetry data as well as numerical modeling, finding that generation mechanisms and variability differ in each subregion in Region 4. Due to the lack of long-term time series of satellite altimetry data and high-resolution model outputs, however, investigations of trends beyond the past two decades are lacking. It is recommended that the variability at longer timescales is studied using long-term high-resolution ocean simulations such as OGCM for the Earth Simulator version 2 (OFES2) (Sasaki et al., 2020). Such studies would also help investigating the relationship between mesoscale and submesoscale activities and climate modes of variability, like the PDO and ENSO. Observations of the submesoscale motions from the forthcoming Surface Water and Ocean Topography (SWOT) altimeter mission (<https://swot.jpl.nasa.gov>) will be especially valuable in this poorly sampled region.

In addition, there is a strong need to investigate the impact of eddy variability on the biological production and marine ecosystem. Most of Region 4 is far from coasts, and it is difficult to conduct repeated ship observations. The analysis of satellite and biogeochemical Argo data coupled with ecosystem modeling would be especially important to comprehensively quantify the impact of mesoscale processes on the ecosystem. For successful ecosystem modeling, field surveys that target the functional relations between the various components are crucial. Recently, Arostegui et al. (2022) showed a pervasive pattern of increased pelagic predator catch inside anticyclonic eddies relative to cyclones and non-eddy areas using a large-scale fishery dataset. Interdisciplinary comparisons are another key tool to understand the impact of eddies on higher trophic level ecosystem.

Table 4: Characteristics and impact on marine ecosystem of eddies in Region 4. Underlining as in Table 1.

Eddy type	KOE/NPC	STCC	East of 170°W	Near Hawaii
EKE	Very strong and getting weaker to the east	Strong	Weak	Strong

Polarity	Anticyclonic and cyclonic: cyclonic eddies are more numerous than anticyclonic eddies (Cheng et al., 2014)			Anticyclonic/ Cyclonic (Lumpkin, 1998)
Formation mechanism	Instabilities associated with strong front/shear (Qiu 2002; Yang and San Liang, 2018; Ji et al 2018)	Instabilities associated with meridional potential vorticity gradient (e.g. Qiu 1999)	<u>Unknown. Numerical simulations could help in this regard.</u>	Instabilities associated with seamounts (Huppert and Bryan, 1976)
Diameter	168 $\pm$ 32 km (anticyclonic eddies), 162 $\pm$ 28 km (cyclonic eddies) for KOE (Cheng et al., 2014)	180 $\pm$ 32 km (anticyclonic eddies), 174 $\pm$ 30 km (cyclonic eddies) (Cheng et al., 2014)	150–200 km (Fig. 4 in Cheng et al., 2014)	~200 km (Fig. 6 in Liu et al., 2012)
Propagation	Mostly westward but northward/southward at the crests/troughs of Kuroshio Extension, respectively (Itoh and Yasuda, 2010a)	Westward (Cheng et al., 2014)		
Lifetime	19 $\pm$ 21 weeks (anticyclonic eddies), 18 $\pm$ 16 weeks (cyclonic eddies) for KOE (Cheng et al., 2014)	17 $\pm$ 14 weeks (anticyclonic eddies), 15 $\pm$ 12 weeks (cyclonic eddies) for KOE (Cheng et al., 2014)	7–21 weeks (anticyclonic eddies), 6–18 weeks (cyclonic eddies) (Figs. 4b and 5b)	<u>Unknown. Analyses focusing on eddies near Hawaii are needed.</u>
Impact on chlorophyll (correlation between	Negative (eddy pumping and trapping, e.g.	Weakly positive (eddy-induced	Very weak	Positive and Negative (multiple

SSH anomaly and chlorophyll-a (expected mechanisms))	Kouketsu et al., 2015, Nakano et al., 2013)	Ekman pumping, Gaube et al., 2014)		mechanisms, Gaube et al., 2014)
Impact on zoo/ichthyoplankton	Zooplankton aggregation at the front of an anticyclonic eddy (Yamamoto and Nishizawa, 1986)	<u>Unknown. Field or ecosystem modeling studies are needed to characterize zooplankton assemblages.</u>	<u>Unknown. Further studies that account for eddy formation area (Region 1 or Region 4) are recommended.</u>	Zooplankton were concentrated at the eddy center (Lobel and Robinson, 1988)
Impact on higher trophic levels	Fish (Sugimoto and Tameishi, 1992), flying squid (Alabia et al., 2015)	Japanese eel (Chang et al., 2018), skipjack tuna (Mugo et al., 2010)		Larval fish (e.g., Lobel and Robinson, 1986), fish recruitment (Fox et al., 2012), fish catch (Seki et al., 2002)

4. Concluding remarks

In this work, we have synthesized current knowledge of mesoscale eddies and their impacts on the marine ecosystem across the North Pacific and its marginal Seas, across the CCS region (Region 1), the northeastern North Pacific and the Bering Sea (Region 2), the western boundary of the North Pacific and marginal seas (Region 3), and the extratropical open North Pacific (Region 4) (Figs. 1 and 2).

While not uniformly, eddies form throughout the open North Pacific, more frequently along the eastern boundary (Region 1 and eastern part of Region 2), along island chains between the open North Pacific and marginal seas (Region 2 and 3), and near the Hawaiian Islands (Region 4) and seamounts, and only seldomly in the center of the Alaskan gyre. Correlation analysis between SSH and chlorophyll shows that primary productivity is especially high in cyclonic eddies in Region 1 and in anticyclonic eddies in Region 2, as also indicated by regional studies.

In Region 3 and 4, on the other hand, the correlation coefficient varies from region to region: negative correlations are most prominent along the stream of the Kuroshio Extension, due to the eddy pumping effect (Kouketsu et al., 2015), while north of the KOE slight positive correlations are observed, likely because of cold anticyclonic eddies that originate in the marginal seas and contribute nutrient rich waters (Prants et al., 2016). In the area east of 170°W, the relationship between SSH anomaly and Chl-a is weak everywhere but around Hawaii, due to the overall weaker and sparser eddy activity. Positive coefficients are observed around 18°–23°N, and result from eddy-induced Ekman pumping (Gaube et al., 2014).

Four tables summarize mesoscale characteristics and eddy-ecosystem interactions in each Region, highlighting similarities and differences. For example, both cyclonic and anticyclonic eddies are observed in Region 1 and 4, but specific polarities are predominant in Region 2 and 3, where eddy formation is influenced by topography, or localized currents. Eddies with the longest lifetime (up to 5 years) are observed in Region 2, and quasi-stationary eddies are observed in Region 3. In Region 1 and in the eastern part of Region 2 along the west coast of US and Canada, eddies propagate offshore, where they transport coastal waters via eddy trapping. In Region 3 and 4, on the other hand, upwelling within the eddies and horizontal stirring are key to explain the biogeochemical properties of the mesoscale eddies.

The tables also indicate where more research is needed. In Region 1, the impact of subsurface anticyclones on the marine ecosystem has not been quantified, while in Region 2, the Aleutian eddies are not well characterized. In Region 3, the impact of eddies on the marine ecosystem is better constrained in the southern half of the domain than in the northern part, and in Region 4 eddy formation mechanisms and mesoscale impacts on higher trophic levels ecosystem have received little attention east of 170°W.

Major open questions differ among regions. In Region 1, a better understanding of the relationship between climate forcing and the intrinsic variability of the mesoscale eddy field is urgently needed, in consideration of the increasing frequency in marine heat waves recorded in the last decade (Oliver et al., 2018). In Region 2, it would be important to figure out how climate variability influences mesoscale activity, given recent observations pointing to an increase in the latter between 1993 and 2011 across most of the Northeast Pacific (Ding et al., 2018). Similarly, in Region 3 the impact of climatic changes on mesoscale variability, eddy formation and ecosystem functioning have not been investigated in depth, despite significant

warming in the past three decades. For example, summer marine heatwaves observed during 2010–2016 in the Oyashio region have been correlated with an increase of yellowtail catch in northern Japan (Miyama et al., 2021) and an increase in the detachment of anticyclonic eddies from the Kuroshio Extension, but a mechanistic understanding of the causal relationships among these events is missing. In Region 4, neither altimetry data nor high resolution model outputs provide long enough time series to investigate trends beyond the past two decades. Climate models suggest that under global warming scenarios the circulation in the southern part of the subtropical North Pacific Gyre will weaken while the northern one will intensify and shift northward (Cheon et al., 2012). Whether such a change would be conducive to changes in mesoscale activities has yet to be investigated; current climate models are too coarse to explicitly represent eddy processes.

Although many studies have revealed the impact of mesoscale eddies on phytoplankton, zooplankton and higher trophic level species individually, seamless understanding linking multiscale physics to chemical properties and finally to the marine ecosystem remains an aim. Recently, regional modeling studies have been used to investigate the physical forcing of eddies onto lower trophic level ecosystems (e.g., Chenillat et al., 2016 for an application to Region 1). As detailed in this review, the mechanisms impacting biological production differ from region to region, and sometime within the same region.

In this review we focused on mesoscale processes and their impact on marine ecosystem in the North Pacific, and discussed submesoscale processes when possible. Submesoscale circulations, which are ubiquitous in the world ocean, enhance vertical velocity at fronts and around mesoscale eddies (McWilliams, 2016), and their contribution should be better assessed and accounted. More studies combining high-resolution numerical models and high-resolution in-situ observations, as done for example in Zhong et al. (2017), are needed.

The North Pacific, especially its northwestern portion, is one of the world's major fishing grounds (e.g., FAO, 2019): adaptation to present and anticipated marine ecosystem changes and the development and adoption of policies that allow for a sustainable development and use of marine resources are essential to continue benefitting from these ecosystem services. To achieve these goals, policy makers need reliable information of the current and forecasted status of marine ecosystems. In this review we have argued that a better understanding and consideration of the role of mesoscale and submesoscale variability is needed to attribute the marine

ecosystem responses to natural and/or anthropogenic forcing across a wide spectrum of time and space scales. As such, investigating biophysical interactions at the ocean mesoscales and smaller scales should be a scientific priority in the North Pacific.

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Figure captions

Fig. 1. Correlation between SSH anomaly and surface Chl-a anomaly corresponding to eddies from 1998 to 2012 (after Kouketsu et al. (2015)). SSH anomaly are high-pass-filtered (< 300 days) to remove long-term changes. Chl-a anomalies are logarithmic deviations from the weekly climatology calculated from 1998 to 2012 data. Gray areas denote correlations that are insignificant at a 90 % confidence level. Figure courtesy of S. Kouketsu. Adapted by permission from Springer Nature: Kouketsu et al. (2015).

Fig. 2. Schematic representation for the circulation of the open North Pacific. The thin white contours, whose contour interval is 0.1 m, indicate the mean dynamic topography produced by CLS and distributed by Aviso+ with support from Cnes (<https://www.aviso.altimetry.fr/>), and downloaded from ftp://ftp-access.aviso.altimetry.fr/auxiliary/mdt/mdt_cnes_cls2013_global/. Acronyms used in this figure is as follows: North Pacific Current (NPC), Kuroshio and Oyashio Extension (KOE), Subtropical Counter Current (STCC), Hawaiian Lee Countercurrent (HLCC), North Equatorial Current (NEC) and California Current System (CCS).

Fig. 3. Average linear trend over 1980-2018 in $^{\circ}\text{C}/\text{decade}$ in the COBE SST2 and Sea-Ice reanalysis (COBEv2; Hirahara et al., 2014). The COBEv2 data set was developed by the Japanese Meteorological Agency, covers the period since 1850 and has a spatial resolution of 1° latitude $\times$ 1° longitude. We note that the period considered, for which we have satellite measurements, is skewed towards a slight predominance of negative PDO conditions.

Fig. 4. Distribution of anticyclonic eddy (a) formation rate (times year⁻¹), (b) averaged lifetime (day) and (c) yield (sum of lifetime divided by 9149 days (01Jan1993–18Jan2018) for those formed in each 2°×2° box over 25 years.

Fig. 5. Same as Fig. 4 but for cyclonic eddies.

Fig. 6. Distribution of anticyclonic eddy formation rate (times season⁻¹) in (a) winter, (b) spring, (c) summer and (d) fall and cyclonic eddy formation in (e) winter, (f) spring, (g) summer and (h) fall at each 2°×2° grid from January 1, 1993 to January 18, 2018.

Fig. 7. Schematic representation for the currents of the CCS region (Region 1). Gray arrows indicate the direction of mean surface geostrophic velocity with speed greater than 0.05 m s⁻¹. Colors represent bottom topography (see Fig. 2). The mean geostrophic velocity data were produced by CLS (Collecte Localisation Satellites), distributed by Aviso+ with support from Cnes data center (<https://www.aviso.altimetry.fr/>), and downloaded from ftp://ftp-access.aviso.altimetry.fr/auxiliary/mdt/mdt_cnes_cls2013_global/.

Fig. 8. Sea-surface temperature (right) and chlorophyll (left) measured by satellite along the U.S. west coast on September 26, 1998 (from Barth, 2007).

Fig. 9. (a) Vertical section of Chl a (mg m⁻³) derived from fluorescence (color) along 37.87°N on 30 June 1993 overlaid with contours of density anomaly (kg m⁻³). The locations of each SeaSoar up-down cycle are indicated by triangles along the bottom. (b) Geopotential anomaly at 100 m, $\Delta\phi_{100/200}$ (m² s⁻²) (thick curve), and 100 m north-south geostrophic velocity, v_g (m s⁻¹) (thin curve), both referenced to acoustic Doppler current profiler velocity at 200 m, along 37.87°N. From Barth et al. (2002).

Fig. 10. The same as Fig. 7 but for the northeastern North Pacific and the Bering Sea (Region 2)

Fig. 11. Map of (a) anticyclonic and (b) cyclonic eddy amplitudes (circle radius), ages (color), from January 2003 to April 2012. From Lyman and Johnson (2015).

Fig. 12. (a) Trajectories of Kenai 2006 and Kenai 2006a (adapted from Ueno et al., 2012). Colors represent SLA (cm) at the eddy and the radius of each circle in the map mostly corresponds to an Okubo-Weiss radius. (b) Trajectories of long-lived Alaskan Stream eddies propagating westward along the Alaskan Stream (after Ueno et al., 2009, © American

Meteorological Society). Shading represents sea level anomalies (cm) at the eddy center. (c) The track of Aleutian eddies, based on mesoscale altimetry (reprinted from Rogachev et al., 2007, with permission from Elsevier).

Fig. 13. (a) EKE ($\text{cm}^2 \text{s}^{-2}$) averaged over full years (1993–2009) calculated from AVISO altimetry data. Gray shading denotes shelf ($< 200 \text{ m}$). Black line shows trajectory of 1997 Pribilof Eddy (15 June–27 August 1997) calculated from drifters. Reprinted from Ladd et al. (2012), with permission from Elsevier.

Fig. 14. As in Fig. 7 but for the western boundary of the North Pacific and marginal seas (Region 3). In this figure, gray arrows indicate the direction of mean surface geostrophic velocity whose speed is stronger than 0.1 m s^{-1} .

Fig. 15. (a) Snapshot current speed (cm/s) and (b) Lagrangian origin map on August 1, 2010, based on AVISO altimetry. Centers of anticyclonic features on this date are marked by triangles. Marked in (b) are Kuroshio rings as KR1-3, Hokkaido eddies as HE, Bussol' eddies as BE, Kamchatka eddies as KE1-2. (b) Colors mark fluid particles which crossed the sections of the same color shown in (a) during two years before the observation date. Isobaths from 7 to 10 km are shown by magenta contours. The origin of particles in the white areas could not be determined.

Fig. 16. SSH field showing the Bussol' eddy (A) sampled in a cruise in 2012 (white circles, Prants et al, 2016) and 'red' anticyclonic eddies with $\text{SSH} > 45 \text{ cm}$ in the Kuril Basin. Up(down)ward oriented triangles indicate the centers of anticyclonic (cyclonic) features on June 25, 2012. Reprinted from Prants et al. (2016), with permission from Elsevier.

Fig. 17. Schematic of the regional circulation around the Korean Peninsula. Blue, red, purple, and orange colors indicate the water temperature of cold, warm, fronts, and riverine waters, respectively. Solid and dashed curves denote the persistent and seasonal currents, respectively (s and w indicate the directions of currents in summer and winter, respectively). Acronyms of primary regional currents are listed in the order of the Yellow Sea Warm Current (YSWC), West Korea Coastal Current (WKCC), Chinese Coastal Current (CCC), Jeju Warm Current (JWC), Jeju Tsushima Front (JTF)/Cheju Tsushima Front (CTF), Jeju Yangtze Front (JYF), Yangtze Diluted Water (YDW), Kuroshio Current (KC), North Korea Cold Current (NKCC), Tsushima Warm Current (TWC), Tsushima Warm Current-Nearshore Branch (TWC-NB),

Tsushima Warm Current-Offshore Branch (TWC-OB), East Korea Warm Current (EKWC), Subpolar Front (SPF), and Ulleung Warm Eddy (UWE). Adapted from Lee et al. (2019).

Fig. 18. EKE distributions in the North Pacific Ocean from the T/P SSH data of October 1992 to November 2000 after removing signals longer than the annual period. Contour intervals are $0.02 \text{ m}^2\text{s}^{-2}$ for solid lines; dashed lines denote the $0.01 \text{ m}^2\text{s}^{-2}$ contours. Reprinted by permission from Springer Nature: Qiu (2002).

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Figures

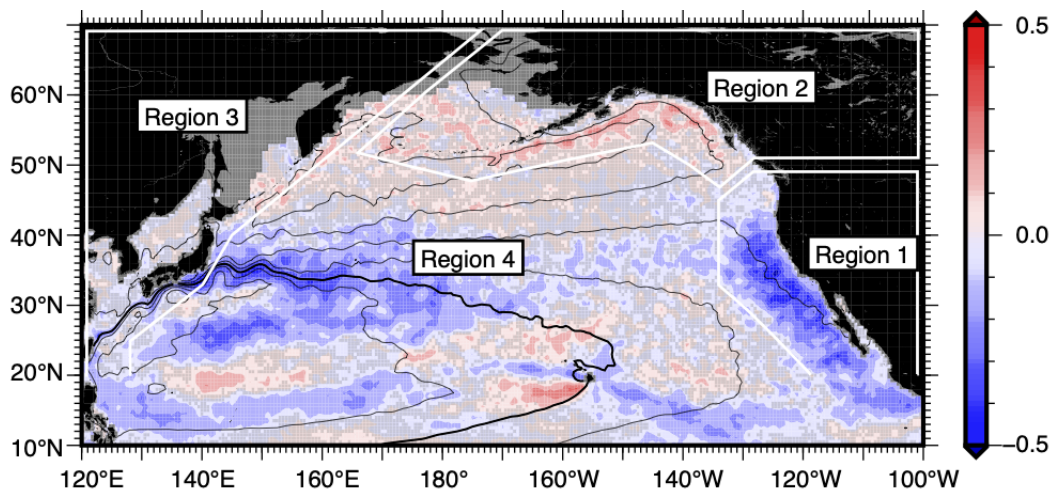


Fig. 1. Correlation between SSH anomaly and surface Chl-a anomaly corresponding to eddies from 1998 to 2012 (after Kouketsu et al. (2015)). SSH anomaly are high-pass-filtered (< 300 days) to remove long-term changes. Chl-a anomalies are logarithmic deviations from the weekly climatology calculated from 1998 to 2012 data. Gray areas denote insignificant correlations at a 90 % confidence level. Figure courtesy of S. Kouketsu. Adapted by permission from Springer Nature: Kouketsu et al. (2015).

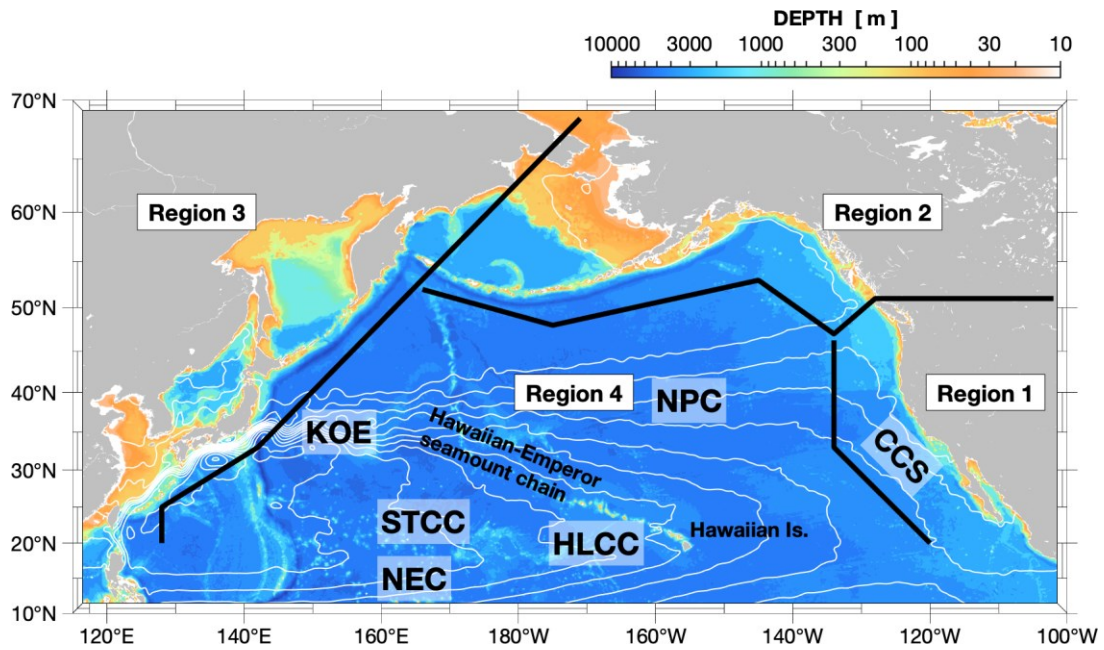


Fig. 2. Schematic representation for the circulation of the open North Pacific. The thin white contours, whose contour interval is 0.1 m, indicate the mean dynamic topography produced by CLS and distributed by Aviso+ with support from Cnes (<https://www.aviso.altimetry.fr/>), and downloaded from ftp://ftp-access.aviso.altimetry.fr/auxiliary/mdt/mdt_cnes_cls2013_global/. Acronyms used in this figure is as follows: North Pacific Current (NPC), Kuroshio and Oyashio Extension (KOE), Subtropical Counter Current (STCC), Hawaiian Lee Countercurrent (HLCC), North Equatorial Current (NEC) and California Current System (CCS).

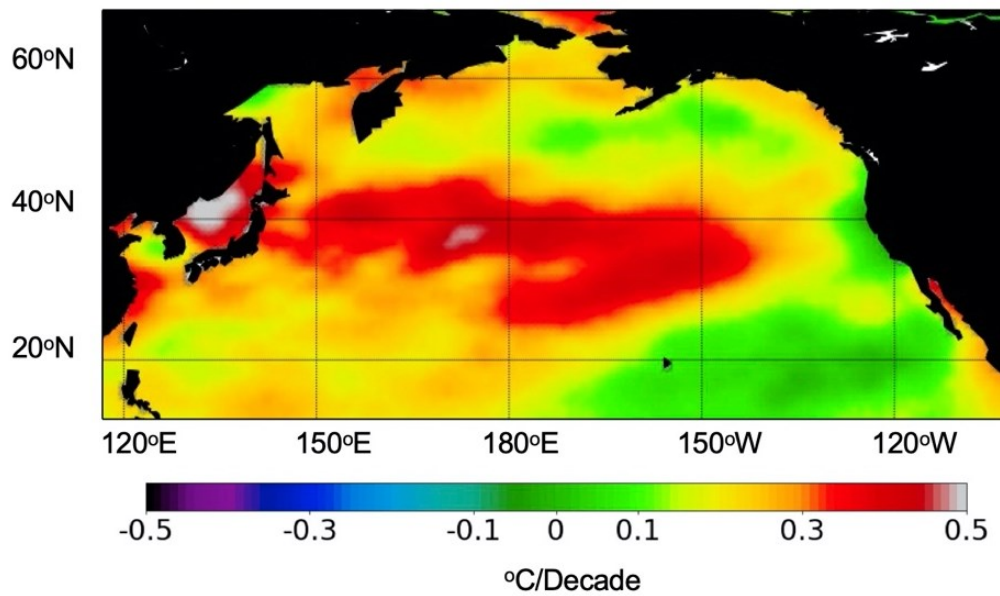
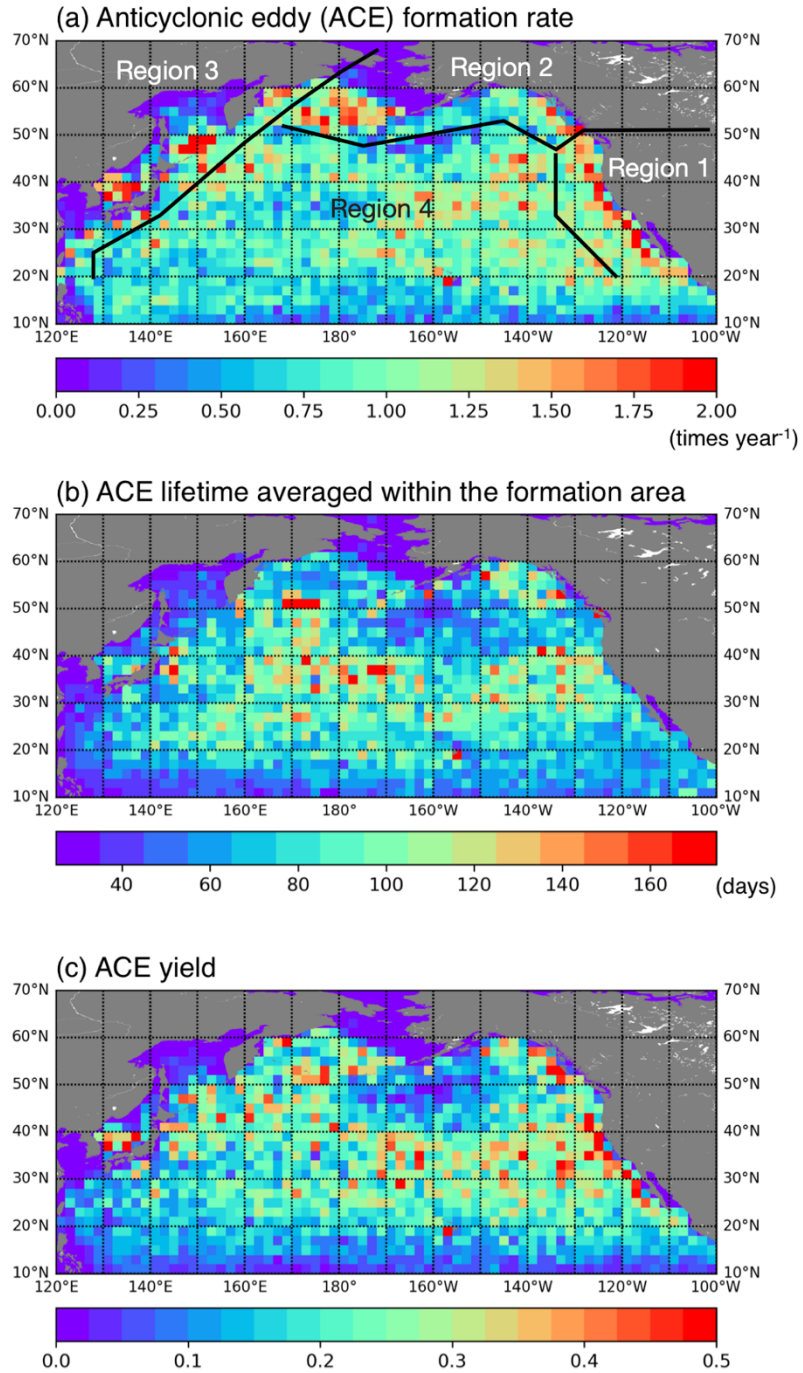
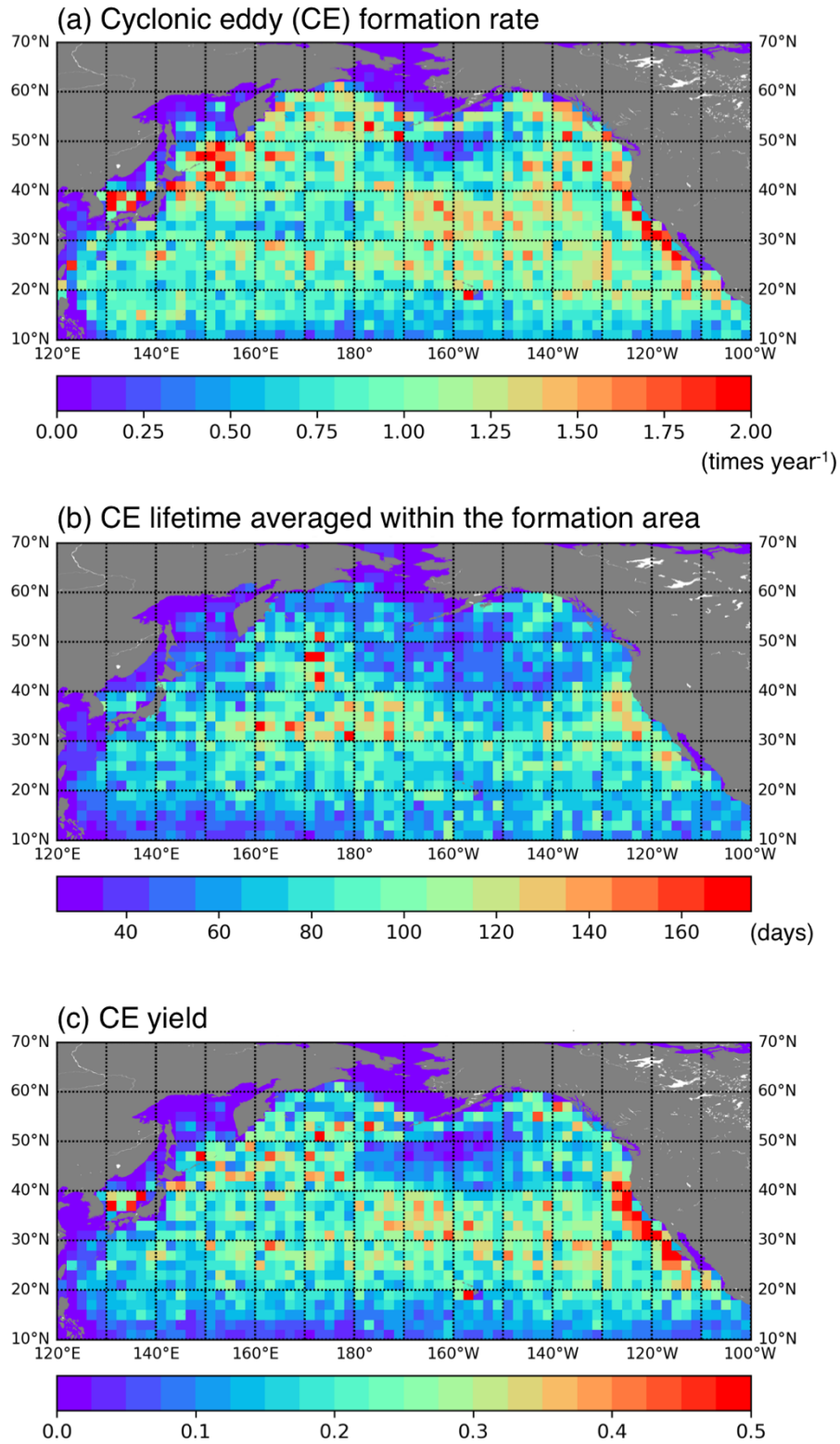


Fig. 3. Average linear trend over 1980-2018 in $^{\circ}\text{C}/\text{decade}$ in the COBE SST2 and Sea-Ice reanalysis (COBEv2; Hirahara et al., 2014). The COBEv2 data set was developed by the Japanese Meteorological Agency, covers the period since 1850 and has a spatial resolution of 1° latitude $\times$ 1° longitude. We note that the period considered, for which we have satellite measurements, is skewed towards a slight predominance of negative PDO conditions.



2982

2983 **Fig. 4.** Distribution of anticyclonic eddy (a) formation rate (times year⁻¹), (b) averaged lifetime
 2984 (day) and (c) yield (sum of lifetime divided by 9149 days (01Jan1993–18Jan2018)) for those
 2985 formed in each 2°×2° box over 25 years.



2986

2987 **Fig. 5.** Same as Fig. 4 but for cyclonic eddies.

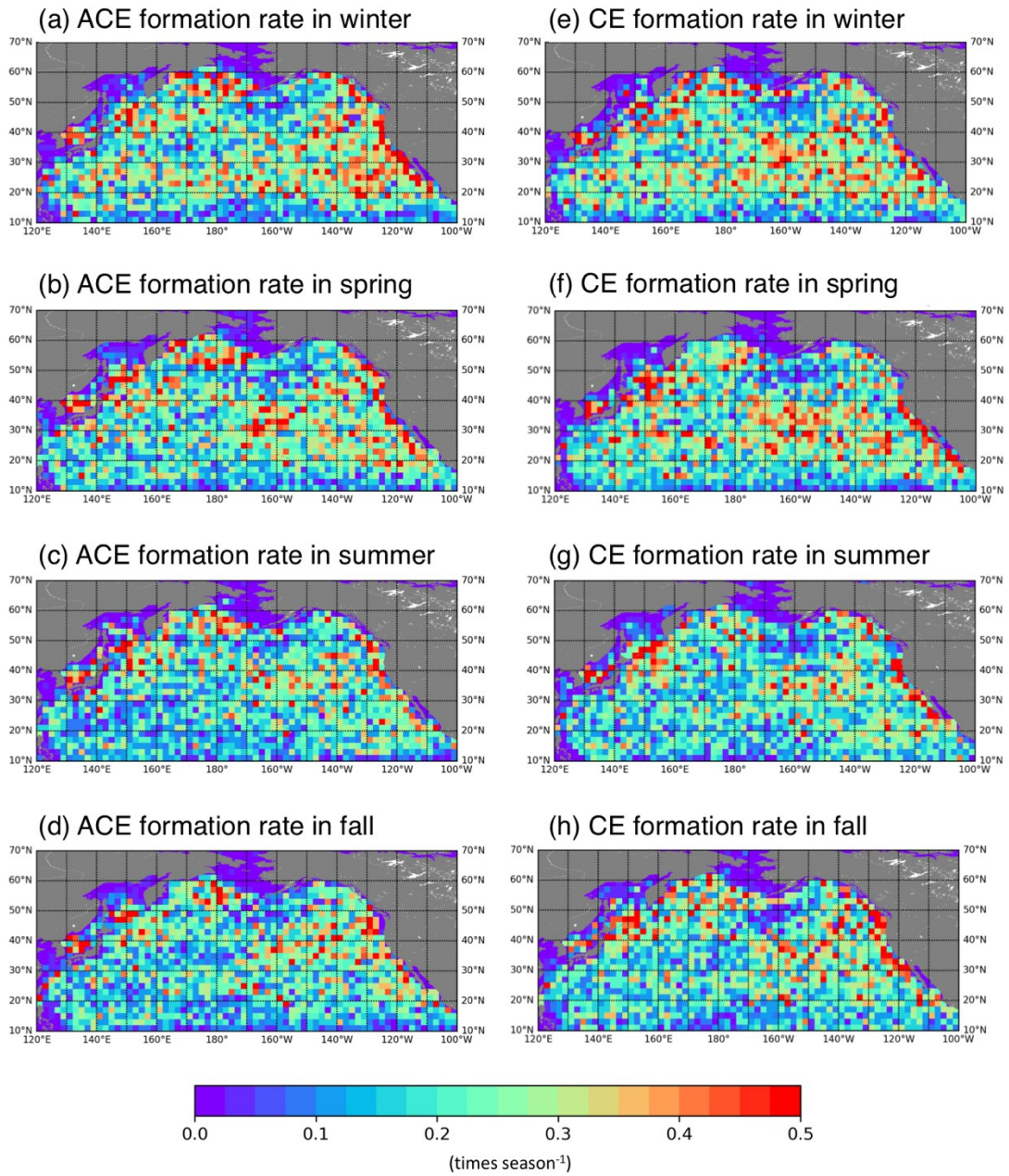


Fig. 6. Distribution of anticyclonic eddy formation rate (times season⁻¹) in (a) winter, (b) spring, (c) summer and (d) fall and cyclonic eddy formation in (e) winter, (f) spring, (g) summer and (h) fall at each 2°×2° grid from January 1, 1993 to January 18, 2018.

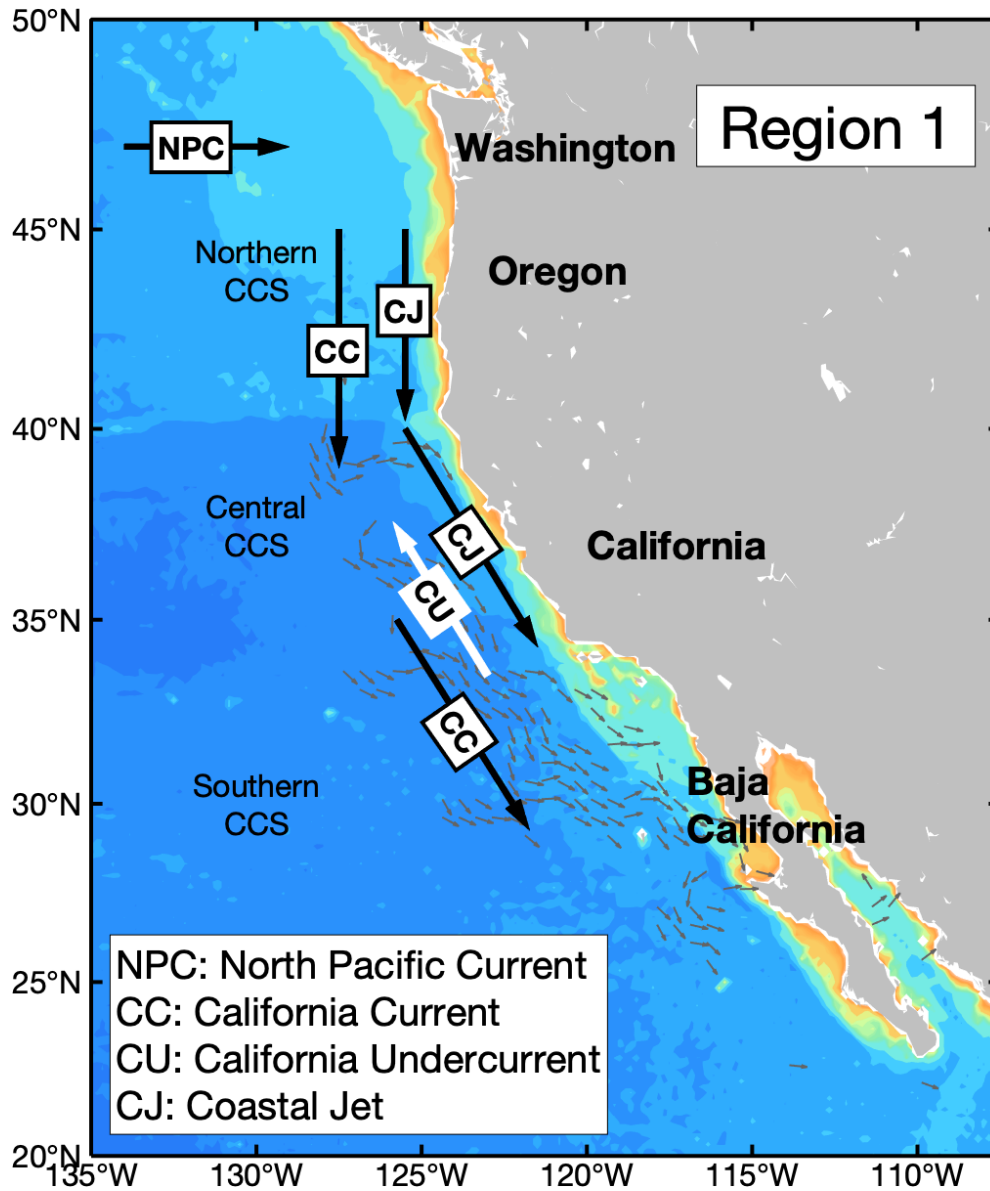


Fig. 7. Schematic representation for the currents of the CCS region (Region 1). Gray arrows indicate the direction of mean surface geostrophic velocity with speed greater than 0.05 m s⁻¹. Colors represent bottom topography (see Fig. 2). The mean geostrophic velocity data were produced by CLS (Collecte Localisation Satellites), distributed by Aviso+ with support from Cnes data center (<https://www.aviso.altimetry.fr/>), and downloaded from ftp://ftp-access.aviso.altimetry.fr/auxiliary/mdt/mdt_cnes_cls2013_global/.

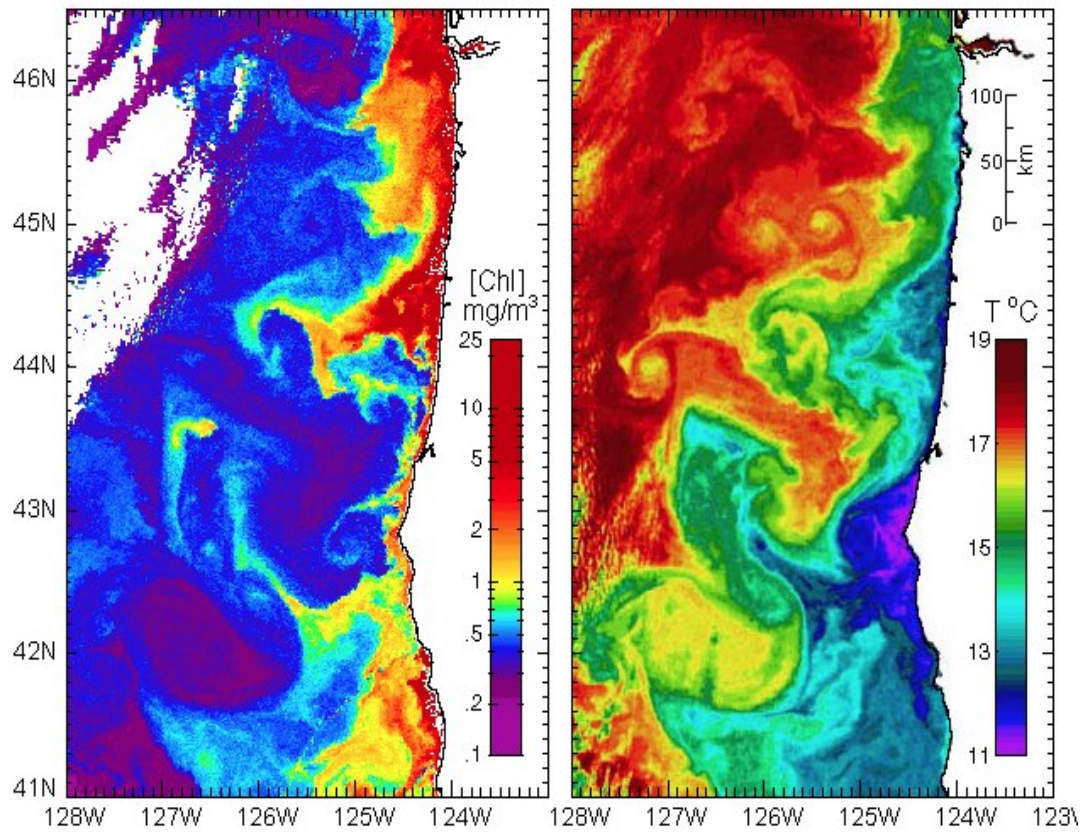


Fig. 8. Sea-surface temperature (right) and chlorophyll (left) measured by satellite along the U.S. west coast on September 26, 1998 (from Barth, 2007).

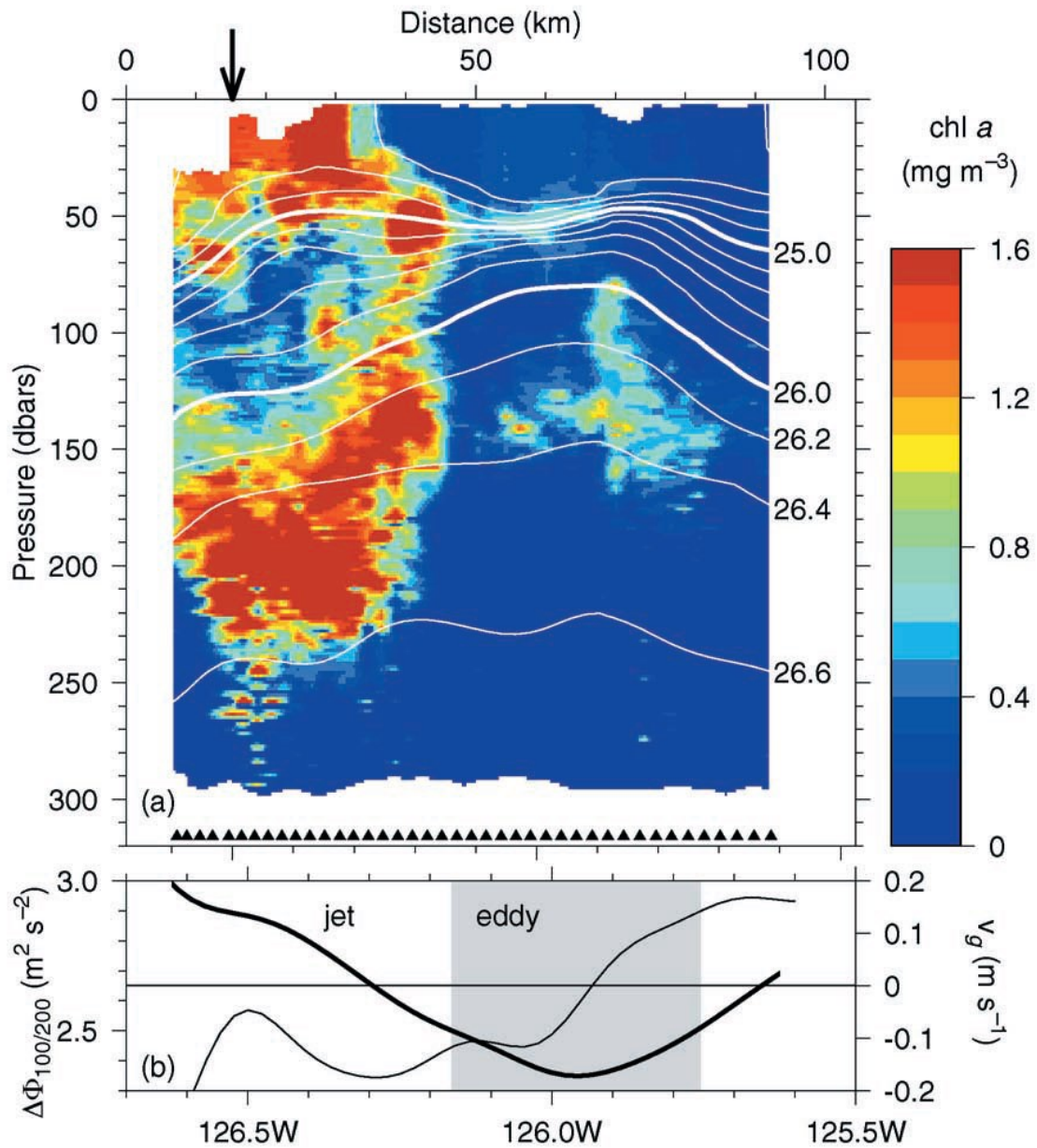


Fig. 9. (a) Vertical section of Chl a (mg m^{-3}) derived from fluorescence (color) along 37.87°N on 30 June 1993 overlaid with contours of density anomaly (kg m^{-3}). The locations of each SeaSoar up-down cycle are indicated by triangles along the bottom. (b) Geopotential anomaly at 100 m, $\Delta\Phi_{100/200}$ ($\text{m}^2 \text{s}^{-2}$) (thick curve), and 100 m north-south geostrophic velocity, v_g (m s^{-1}) (thin curve), both referenced to acoustic Doppler current profiler velocity at 200 m, along 37.87°N . From Barth et al. (2002).

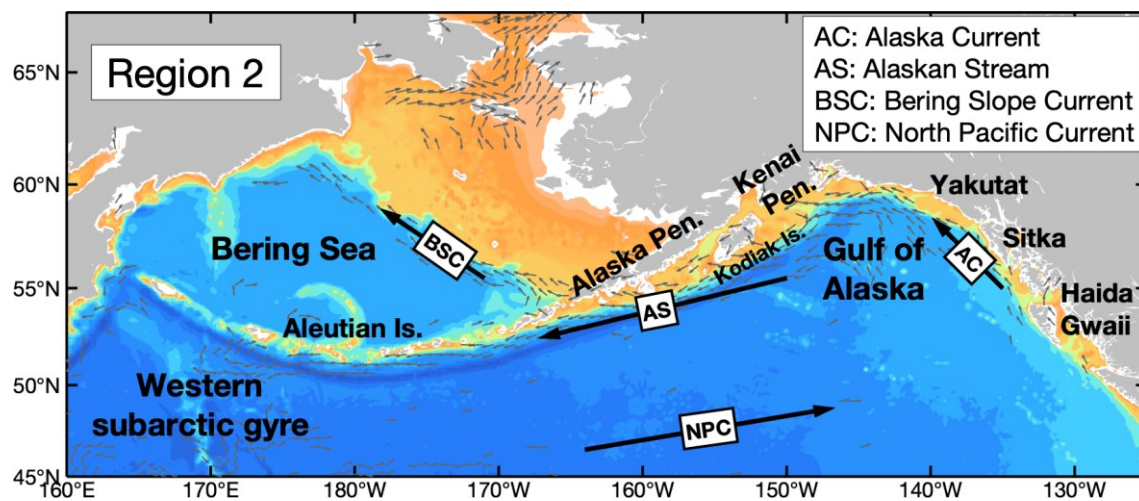
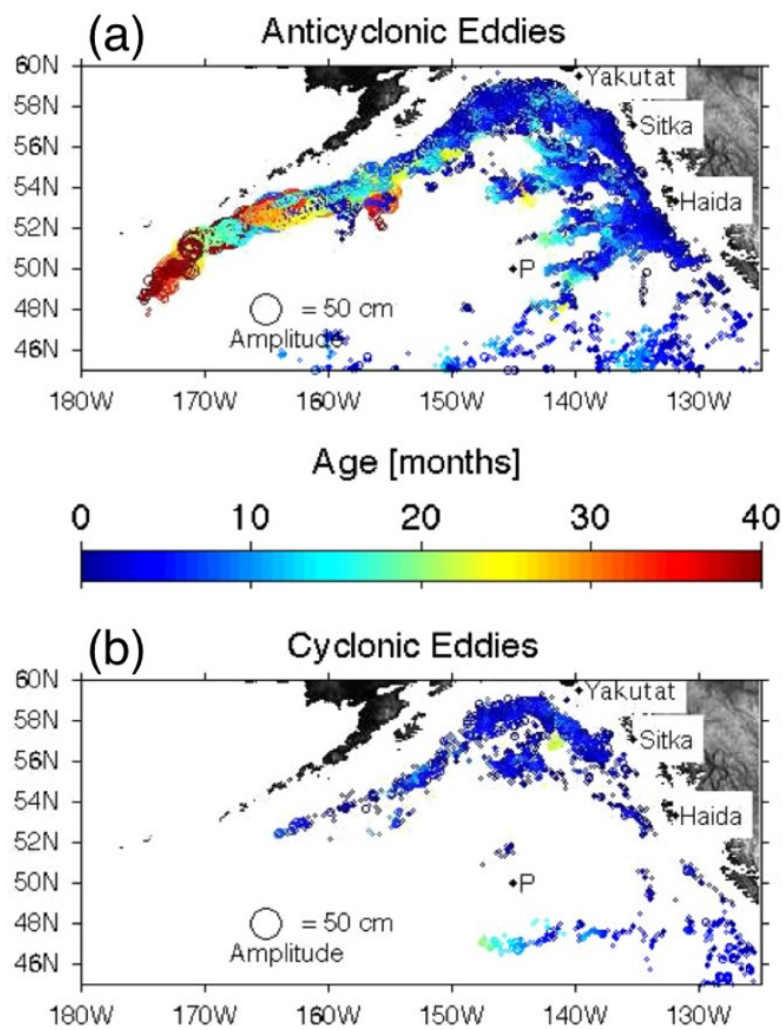


Fig. 10. The same as Fig. 7 but for the northeastern North Pacific and the Bering Sea (Region 2)

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3021 **Fig. 11.** Map of (a) anticyclonic and (b) cyclonic eddy amplitudes (circle radius), ages (color),
3022 from January 2003 to April 2012. From Lyman and Johnson (2015).

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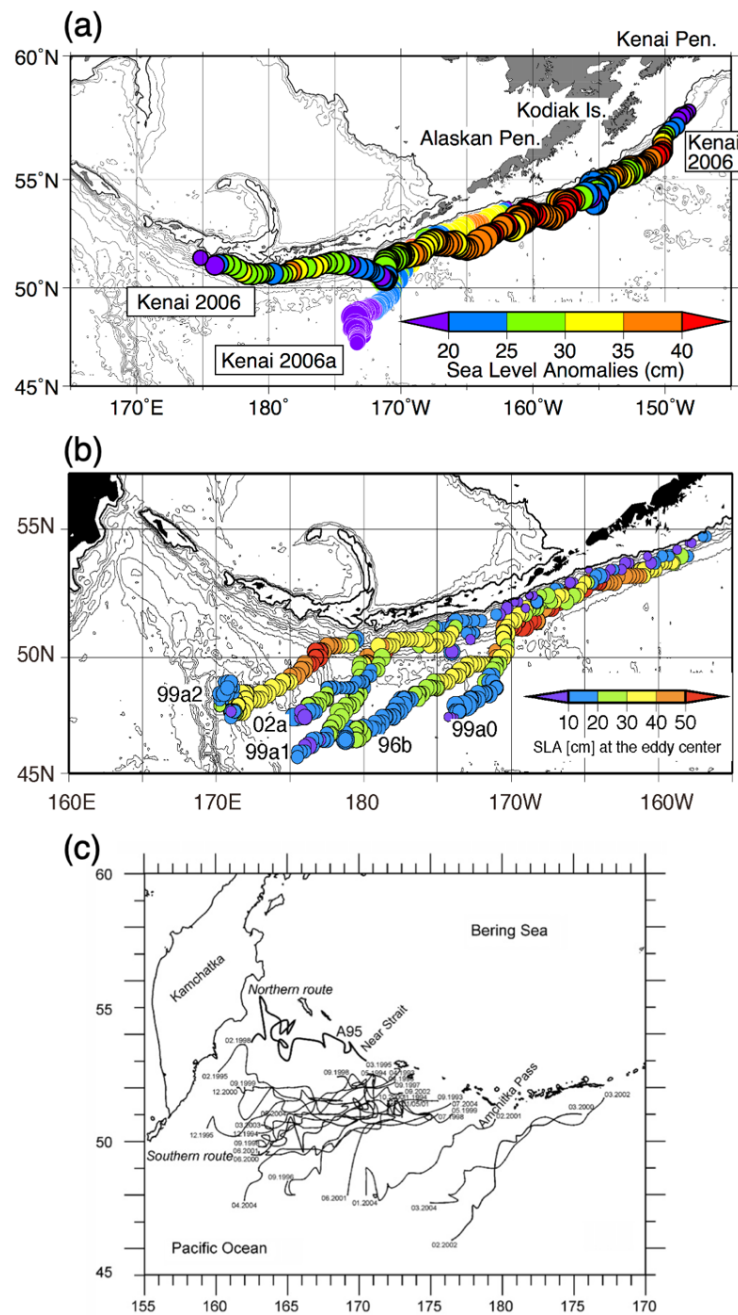


Fig. 12. (a) Trajectories of Kenai 2006 and Kenai 2006a (adapted from Ueno et al., 2012). Colors represent SLA (cm) at the eddy and the radius of each circle in the map mostly corresponds to an Okubo-Weiss radius. (b) Trajectories of long-lived Alaskan Stream eddies propagating westward along the Alaskan Stream (after Ueno et al., 2009, © American Meteorological Society). Shading represents sea level anomalies (cm) at the eddy center. (c) The track of Aleutian eddies, based on mesoscale altimetry (reprinted from Rogachev et al., 2007, with permission from Elsevier).

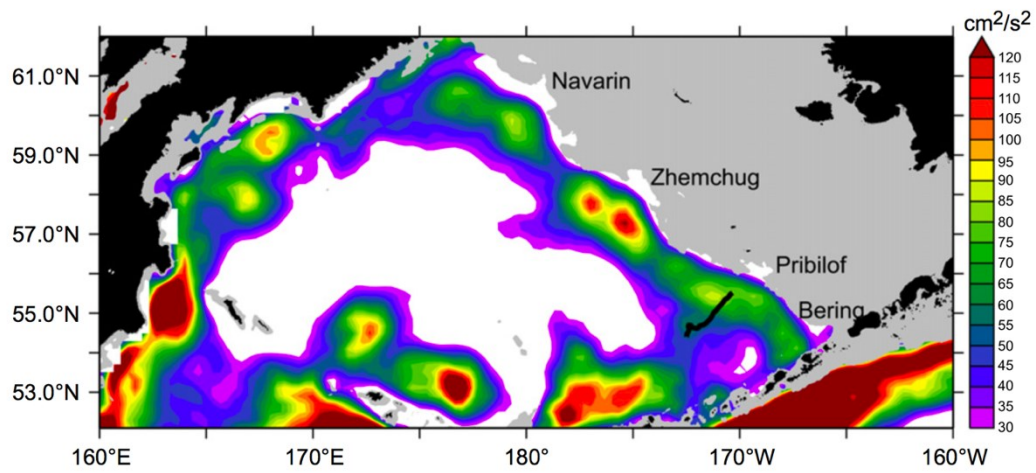


Fig. 13. (a) EKE ($\text{cm}^2 \text{s}^{-2}$) averaged over full years (1993–2009) calculated from AVISO altimetry data. Gray shading denotes shelf (< 200 m). Black line shows trajectory of 1997 Pribilof Eddy (15 June–27 August 1997) calculated from drifters. Reprinted from Ladd et al. (2012), with permission from Elsevier.

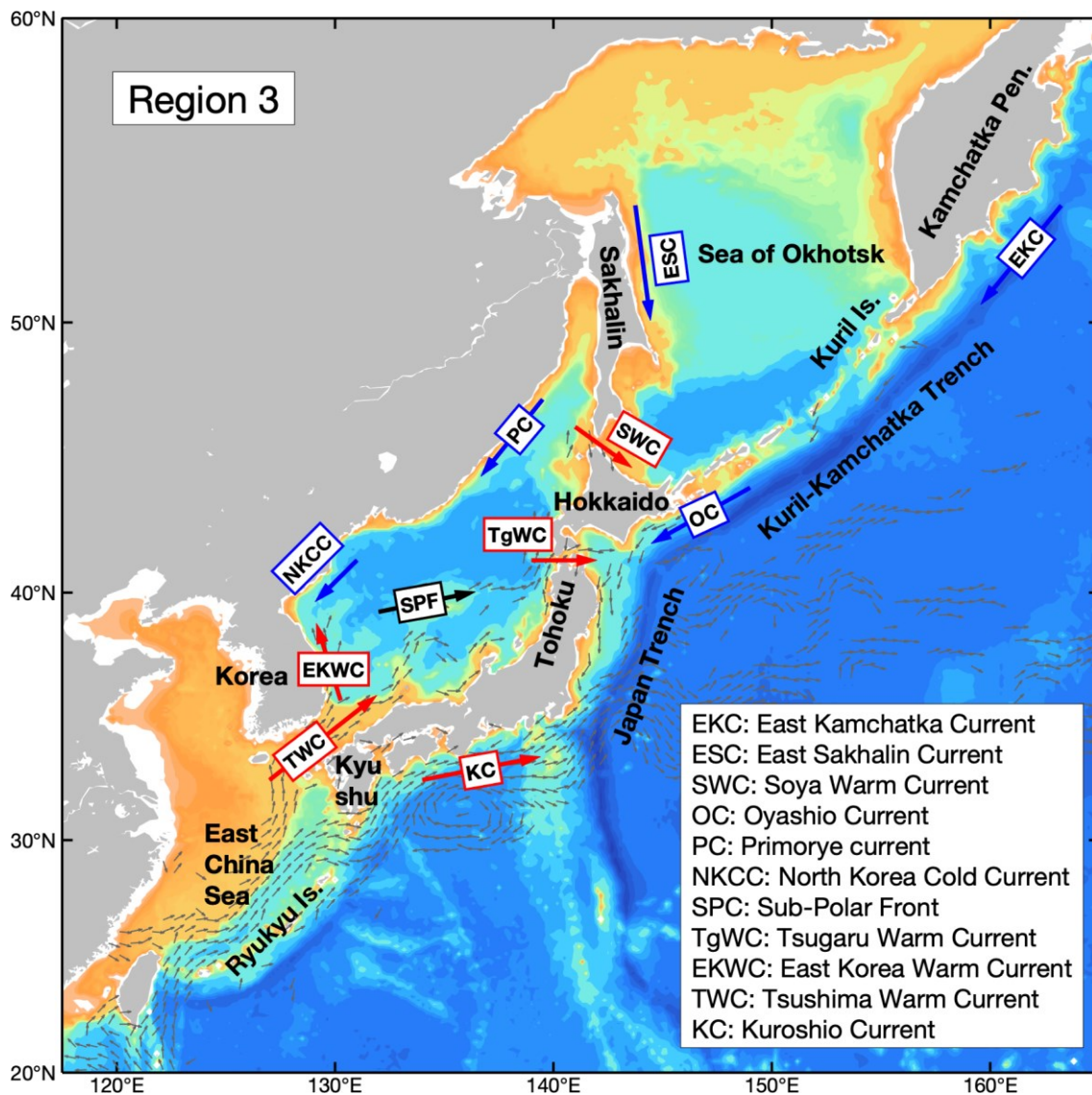


Fig. 14. As in Fig. 7 but for the western boundary of the North Pacific and marginal seas (Region 3). In this figure, gray arrows indicate the direction of mean surface geostrophic velocity whose speed is stronger than 0.1 m s^{-1} .

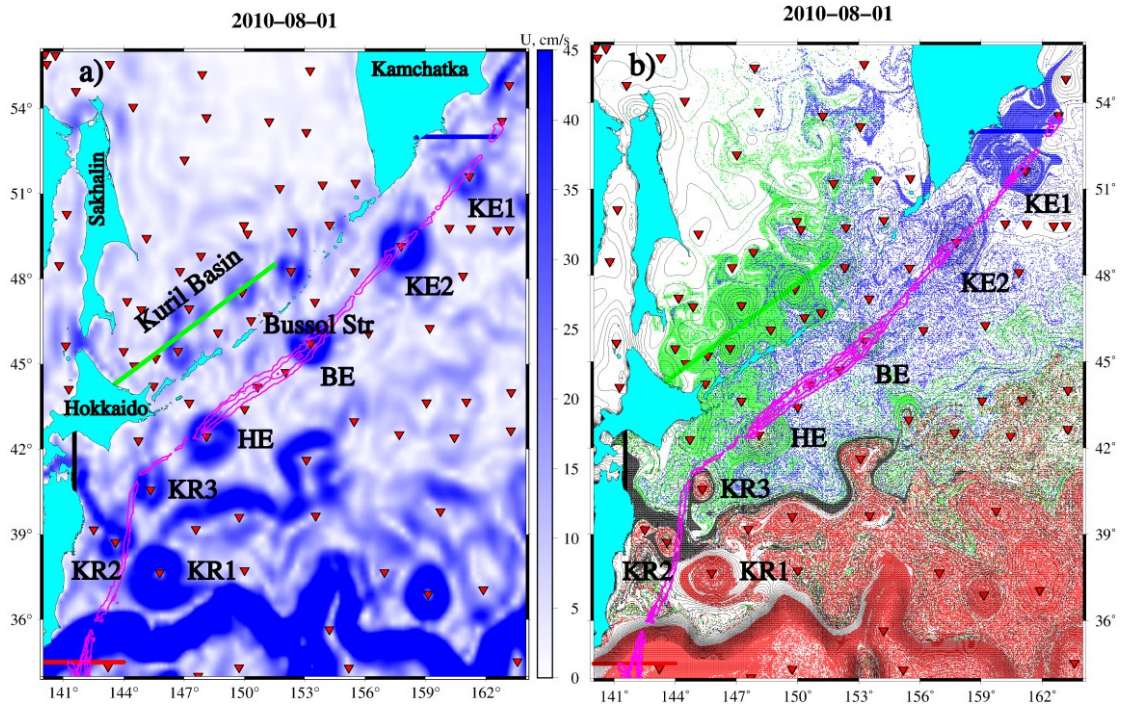


Fig. 15. (a) Snapshot current speed (cm/s) and (b) Lagrangian origin map on August 1, 2010, based on AVISO altimetry. Centers of anticyclonic features on this date are marked by triangles. Marked in (b) are Kuroshio rings as KR1-3, Hokkaido eddies as HE, Bussol' eddies as BE, Kamchatka eddies as KE1-2. (b) Colors mark fluid particles which crossed the sections of the same color shown in (a) during two years before the observation date. Isobaths from 7 to 10 km are shown by magenta contours. The origin of particles in the white areas could not be determined.

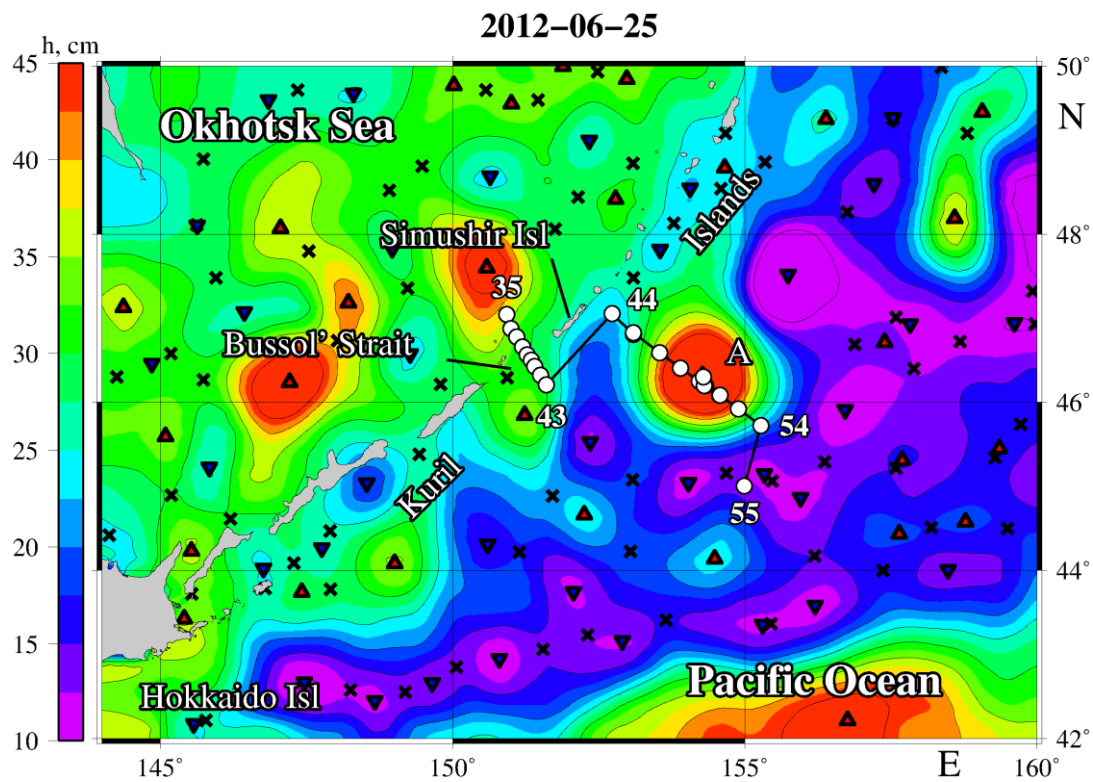
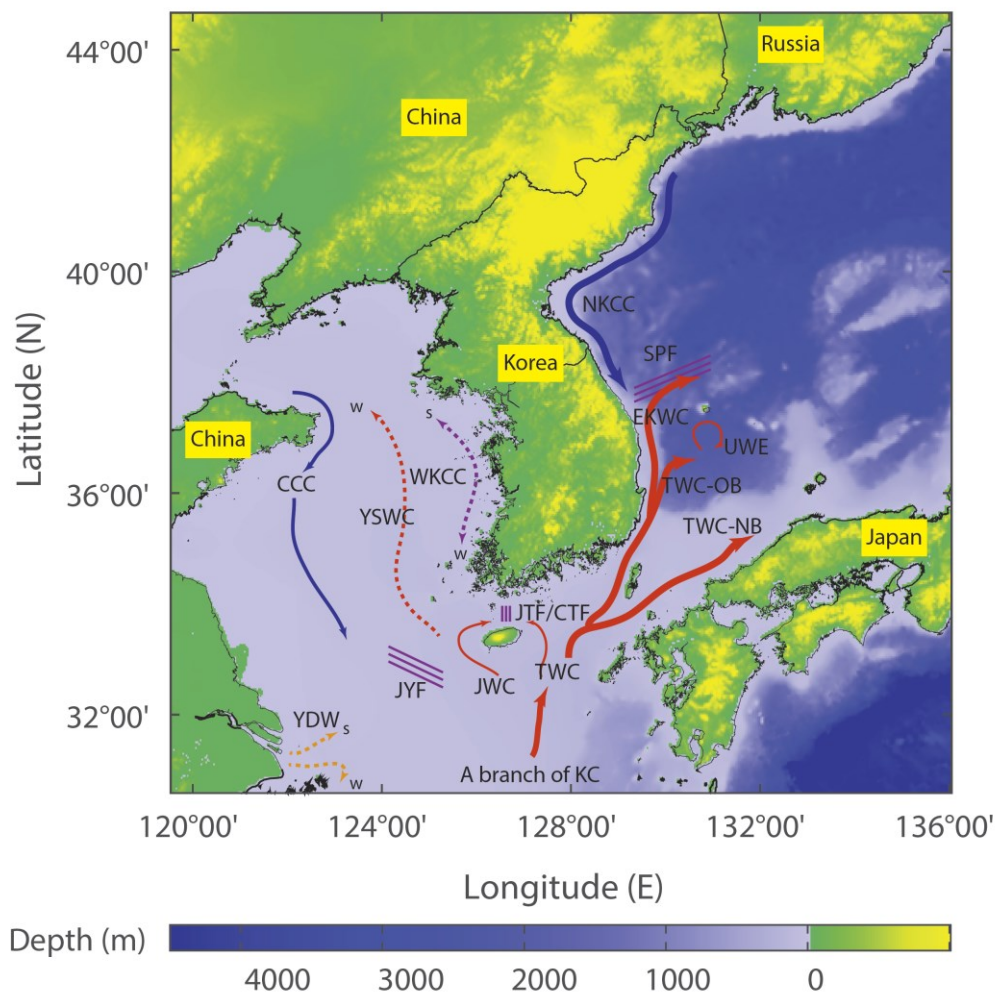


Fig. 16. SSH field showing the Bussol' eddy (A) sampled in a cruise in 2012 (white circles, Prants et al, 2016) and 'red' anticyclonic eddies with SSH > 45 cm in the Kuril Basin. Up(down)ward oriented triangles indicate the centers of anticyclonic (cyclonic) features on June 25, 2012. Reprinted from Prants et al. (2016), with permission from Elsevier.

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3061

3062 **Fig. 17.** Schematic of the regional circulation around the Korean Peninsula. Blue, red, purple,
3063 and orange colors indicate the water temperature of cold, warm, fronts, and riverine waters,
3064 respectively. Solid and dashed curves denote the persistent and seasonal currents, respectively (s
3065 and w indicate the directions of currents in summer and winter, respectively). Acronyms of
3066 primary regional currents are listed in the order of the Yellow Sea Warm Current (YSWC),
3067 West Korea Coastal Current (WKCC), Chinese Coastal Current (CCC), Jeu Warm Current
3068 (JWC), Jeu Tsushima Front (JTF)/Cheju Tsushima Front (CTF), Jeu Yangtze Front (JYF),
3069 Yangtze Diluted Water (YDW), Kuroshio Current (KC), North Korea Cold Current (NKCC),
3070 Tsushima Warm Current (TWC), Tsushima Warm Current-Nearshore Branch (TWC-NB),
3071 Tsushima Warm Current-Offshore Branch (TWC-OB), East Korea Warm Current (EKWC),
3072 Subpolar Front (SPF), and Ulleung Warm Eddy (UWE). Adapted from Lee et al. (2019).

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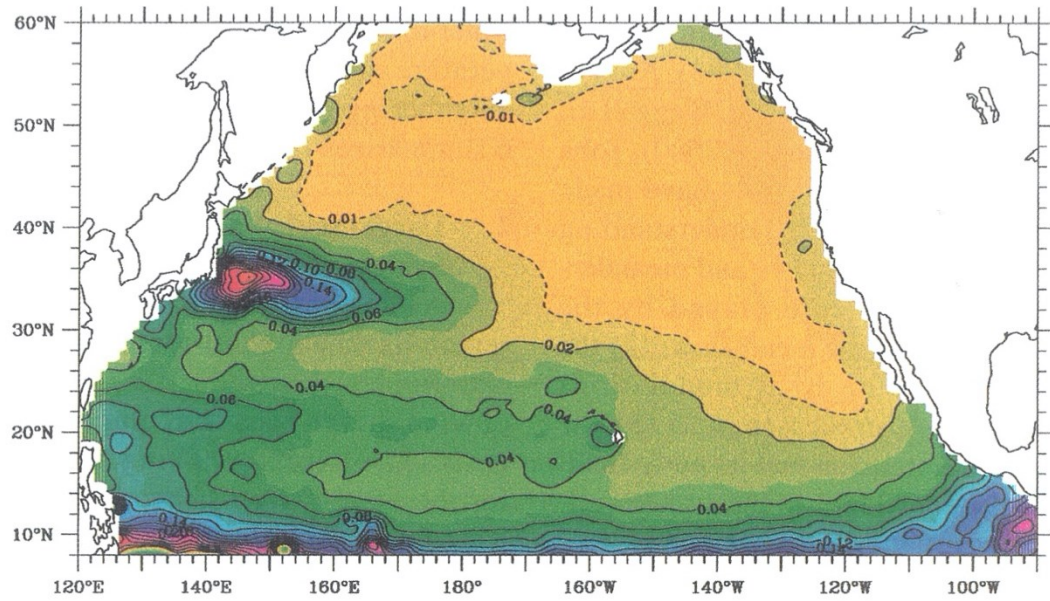


Fig. 18. EKE distributions in the North Pacific Ocean from the T/P SSH data of October 1992 to November 2000 after removing signals longer than the annual period. Contour intervals are $0.02 \text{ m}^2\text{s}^{-2}$ for solid lines; dashed lines denote the $0.01 \text{ m}^2\text{s}^{-2}$ contours. Reprinted by permission from Springer Nature: Qiu (2002).