Dominant role of eddies and filaments in the offshore transport of carbon and nutrients in the California Current System

Takeyoshi Nagai1, Nicolas Gruber2, Hartmut Frenzel3, Zouhair Lachkar2,4, James C. McWilliams5, and Gian-Kasper Plattner6

1Department of Ocean Sciences, Tokyo University of Marine Science and Technology, Tokyo, Japan, 2Environmental Physics, Institute of Biogeochemistry and Pollutant Dynamics, ETH Zurich, Zurich, Switzerland, 3School of Oceanography, University of Washington, Seattle, Washington, USA, 4Center for Prototype Climate Modeling, New York University Abu Dhabi, Abu Dhabi, United Arab Emirates, 5Department of Atmospheric and Oceanic Sciences, University of California, Los Angeles, California, USA, 6Climate and Environmental Physics, University of Bern, Bern, Switzerland

Abstract The coastal upwelling region of the California Current System (CalCS) is a well-known site of high productivity and lateral export of nutrients and organic matter, yet neither the magnitude nor the governing processes of this offshore transport are well quantified. Here we address this gap using a high-resolution (5 km) coupled physical-biogeochemical numerical simulation (ROMS). The results reveal (i) that the offshore transport is a very substantial component of any material budget in this region, (ii) that it reaches more than 800 km into the offshore domain, and (iii) that this transport is largely controlled by mesoscale processes, involving filaments and westward propagating eddies. The process starts in the nearshore areas, where nutrient and organic matter-rich upwelled waters pushed offshore by Ekman transport are subducted at the sharp lateral density gradients of upwelling fronts and filaments located at ~25–100 km from the coast. The filaments are very effective in transporting the subducted material further offshore until they form eddies at their tips at about 100–200 km from the shore. The cyclonic eddies tend to trap the cold, nutrient, and organic matter-rich waters of the filaments, whereas the anticyclones formed nearby encapsulate the low nutrient and low organic matter waters around the filament. After their detachment, both types of eddies propagate further in offshore direction, with a speed similar to that of the first baroclinic mode Rossby waves, providing the key mechanism for long-range transport of nitrate and organic matter from the coast deep into the offshore environment.

1. Introduction

Coastal upwelling systems along the eastern boundaries of the subtropical ocean basins, so-called Eastern Boundary Upwelling Systems (EBUS) such as the California, Canary, Humboldt, or Benguela Current Systems, belong to the biologically most productive continental margins in the world oceans [e.g., Carr, 2002; Pauly and Christensen, 1995]. This is a direct consequence of the strong nutrient supply from depth caused by the offshore Ekman transport driven by the persistent equatorward winds. Less clear is the fate of the abundant organic matter that is being produced in the nearshore environments as a result of the upwelling. It has been hypothesized for a long time that a substantial amount of this material might be transported laterally to the offshore regions, but the exact quantification of this offshore transport of organic matter as well as the elucidation of the governing processes has remained elusive [Brink and Cowles, 1991; Strub et al., 1991; Álvarez-Salgado et al., 2001; Gruber et al., 2011].

The potentially substantial lateral export of organic matter from the continental margins to the open ocean inspired several large-scale observation programs. Among them are the Shelf Edge Exchange Processes (SEEP) program carried out on the east coast of the U.S. in the 1980s [Biscaye et al., 1994], the Coastal Transition Zone (CTZ) Program, which focused on the physical and biological dynamics of the cold filaments formed along the northwestern U.S. coast [Brink and Cowles, 1991; Strub et al., 1991], the Ocean Margin Exchange program (OMEX), which investigated the offshore transport from the northwestern European shelf break and the northwest Iberian Margin in the 1990s [Wollast and Chou, 2001], and several others...
An important reason for the complex physical/biogeochemical processes associated with offshore transport is that the upwelled water is colder and denser than the surrounding water. This creates a coastal upwelling front [Smith, 1995] that is baroclinically unstable and induces frontal waves that cause complex and distinct horizontal structures of flows, sea surface temperature, and associated chlorophyll biomass. Because the upwelling fronts act as barriers for the transport of material properties across them [Brink et al., 1995; Smith, 1995], they tend to impede the Ekman-driven offshore transport of these properties deep into the offshore region. Instead, it is the turbulent transport by the ubiquitous mesoscale and finer-scale dynamics [Marchesiello et al., 2003; Capet et al., 2008a, 2008b; Stegmann and Schwing, 2007] that is likely to dominate the offshore transport of material properties.

First concrete support for this hypothesis came from the Coastal Transition Zone (CTZ) Program, where Strub et al. [1991] showed that most of the offshore transport in the California Current System (CalCS) is provided by the meanders of the California Current and the associated filaments. They also pointed out that the subduction in the offshore-flowing branch of the filament may move phytoplankton and nutrients below the euphotic zone, thereby generating a powerful conduit for offshore transport. The three-dimensional modeling studies by Haidvogel et al. [1991], Moisan and Hofmann [1996a], and Moisan et al. [1996] generally support this conclusion, but they also showed that the filaments have a distinct structure of offshore transport. The northern edge of the filament is characterized by offshore and downwelling flows, while the southern edge is characterized by onshore and the upwelling flows, so that the net offshore transport occurs at the center of the filaments. Using a Lagrangian model, Moisan and Hofmann [1996b] pointed out that the downward flux of organic matter in the northern part of the cold filament could determine the depth of the deep biomass maxima often observed in the CalCS. A number of recent field studies also suggested the significance of the subduction of upwelled water and organic matter in determining the physical structures and the lateral export of organic matter [Barth and Wheeler, 2005; Collins et al., 2003; Bograd and Mantyla, 2005].

While filaments may transport organic matter efficiently from the very nearshore to offshore distances of the order of 100 km, due to their limited reach, they cannot provide a conduit for the transport to greater offshore distances. Cyclonic and anticyclonic mesoscale eddies that spin off at the tips of the filaments [e.g., Mied et al., 1991; Moisan and Hofmann, 1996a] and then tend to propagate offshore may provide this conduit. Strub et al. [1991] demonstrated on the basis of a few observations that such eddies indeed propagate further offshore with a speed that is very similar to that of the first baroclinic mode of the Rossby waves, consistent with theoretical consideration [Killworth et al., 1997]. Satellite observations [e.g., Kelly et al., 1998] as well as numerical simulations [e.g., Marchesiello et al., 2003] showed that this westward propagation of eddies departing from the nearshore areas is a widespread phenomenon that characterizes the entire CalCS. A modeling study by Combes et al. [2013] suggests that across-shore transport of surface passive tracer is controlled by interplay between eddy activities and Ekman upwelling. Thus, these offshore propagating eddies may contribute substantially to offshore transport, but their relative roles for biogeochemical tracers are not well quantified.

These propagating Rossby waves and/or traveling rings or eddies are potentially very good transport agents, as they can trap material properties and transport them over great distances. The trapping occurs when the current velocity within an eddy exceeds the translational speed of the traveling eddy [Flierl, 1981]. Such transports can be considered as the mean eddy mass flux (analogous to Stokes drift). The trajectories of the surface drifters often show clear swirls translating with the eddy traveling [Strub et al., 1991]. Chemical observations moving with the eddies also indicate that the eddies can maintain the chemical characteristics from the source waters, while they move several hundreds of kilometers [Richardson, 1993; McDonagh et al., 1999; Reynolds et al., 2014]. Recent satellite observations [Chelton et al., 2011a] reveal their widespread and large impacts on the basin-scale volume-flux [Zhang et al., 2014] and on the surface chlorophyll distribution [Chelton et al., 2011b].
The offshore transport of organic matter and other biogeochemically important quantities, such as inorganic nutrients, has many important consequences. First, it tends to redistribute nutrients between the nearshore areas, where they are abundant, and the offshore region, where primary production tends to be strongly nutrient limited. Second, this transport may impact the air-sea flux of CO2 in a substantial manner, as it separates the location of where organic carbon is being formed and where it is being remineralized, thereby reducing the potential CO2 outgassing in the nearshore areas and increasing it in the offshore region [see, e.g., Turi et al., 2014]. Third, this offshore transport may also provide an important conduit for the dispersal of organisms and along with them genetic information [Bucklin, 1991; Bucklin et al., 2002].

Gruber et al. [2011] showed on the basis of a model study for the CalCS that the offshore transport of inorganic nutrients is a critical element determining nearshore productivity in this region. They demonstrated that the eddy-induced circulation subducts and laterally exports a substantial amount of nitrate from the nearshore euphotic zone to the offshore subsurface, thereby depleting the nearshore reservoir of the nitrate. As nitrate is the primary factor limiting productivity in the CalCS and nearly all other areas of the EBUS, this explains a reduced productivity in the presence of high mesoscale activity. Using satellite observations, they also showed that the magnitude of eddy variability in any EBUS tended to correlate inversely with the magnitude of net primary production [see also Lachkar and Gruber, 2012; Rossi et al., 2008, 2009]. While these studies pointed out the potentially significant role played by eddies in the offshore fluxes, they did not determine the relative role of the different processes in detail. They also quantified the eddy-induced transport only to a limited degree.

Here we aim to close this gap and investigate the spatial and temporal variations of offshore transport of organic matter and inorganic constituents in the CalCS with particular emphasis on the role of eddies and rings released from the filaments. To this end, we will use the same simulations used by Gruber et al. [2011], i.e., simulations with an eddy-resolving three-dimensional biogeochemical model coupled to the Regional Oceanic Modeling System (ROMS), configured for the U.S. West Coast (USWC) [Gruber et al., 2006].

2. Model

The model we employed in the present study is a coupled physical-biogeochemical model [Gruber et al., 2006] using a United States West Coast (USWC) configuration of the ETH-UCLA version of the Regional Oceanic Modeling System (ROMS) [Shchepetkin and McWilliams, 2005; Marchesiello et al., 2003]. ROMS adopted the curvilinear coordinate in the horizontal and the s coordinate in the vertical direction with finer resolution near the surface to resolve boundary layer mixing, which is parameterized by the nonlocal K-profile parameterization, KPP [Large et al., 1994]. The model grid covers the entire USWC that extends in latitude from the middle of Baja California (28°N) to the Canadian Border (48°N) and is 2100 km long and 1300 km wide. The curvilinear horizontal grid we used is oriented to have one axis aligned roughly across-shore direction and the other alongshore direction. The flux and velocity along the former axis is defined hereinafter as across-shore flux or velocity \( u \), and the latter as alongshore flux and velocity \( v \). Our eddy-resolving simulation was performed with \( \sim 5 \) km horizontal resolution for the entire USWC to resolve evolutions of eddies emitted from the central upwelling region over long time and distances.

The nitrogen-based ecological-biogeochemical model consists of seven state variables for phytoplankton, zooplankton, small detritus, large detritus, nitrate, ammonium, and a dynamic phytoplankton chlorophyll-to-carbon ratio. Primary production in the model is limited by the amount of photosynthetically active radiation (PAR), the concentrations of nitrate and ammonium, and temperature [Gruber et al., 2006]. Small detritus is fueled by the mortality of phytoplankton and zooplankton and egestion/excretion by zooplankton. Large detritus is formed by the coagulation of phytoplankton with small detritus, and the mortality and egestion/excretion by zooplankton. Phytoplankton and the two detrital pools are subject to sinking with constant velocities of 10 m d\(^{-1}\) for large detritus, 1 m d\(^{-1}\) for small detritus, and 0.5 m d\(^{-1}\) for phytoplankton. The conservation equations of the state variables are integrated in time, allowing them to be advected and diffused based on the physical flow field and the sinking velocities. The details of the ecological-biogeochemical model can be found in Gruber et al. [2006].

The simulation analyzed here is identical to that used by Gruber et al. [2011] and differs from the original description in Gruber et al. [2006] in two ways. First, the physical grid has a uniform resolution of 5 km and the vertical resolution was increased to 32 layers. Second, two parameters of the ecosystem model, namely...
the initial slope of the light-response curve for phytoplankton growth and the mortality rate of phytoplankton were doubled to improve the agreement with observationally based estimates of NPP (see Gruber et al. [2011], for further details).

The model was run in a climatologically forced mode, using the monthly mean climatologies of heat and freshwater fluxes from the Comprehensive Ocean Atmosphere Data Set (COADS) [da Silva et al., 1994] and the monthly mean winds from QuikSCAT [Chelton et al., 2006]. The coupled model was spun-up from the initial condition for 5 years, and continued to run for another 5 years. For our analyses, we use 2 day (averaged) output for 5 years from years 6 to 10.

3. Mean State and Offshore Transport

3.1. Climatologically Mean State

The long-term mean state of the CaICS is critical for putting our analyses of the eddy-driven offshore transport into context. We thus provide here first an overview of the model mean results (temporally averaged from years 6 to 10), focusing on the mean fluxes. Readers interested in the evaluation of the model with observations are referred to Gruber et al. [2006] and the supporting information section of Gruber et al. [2011] (http://www.nature.com/ngeo/journal/v4/n11/extref/ngeo1273-s1.pdf).

A vertical section of the mean density and circulation for the central California coast (averaged between Pt. Conception and C. Mendocino alongshore direction), shown in Figure 1a, reveals the typical pattern of strong upwelling near the coast as well as the strong offshore transport in the uppermost layers extending to more than 800 km from the coast. The alongshore range for the average is chosen on the basis of where the California Current flows roughly parallel to the shore.

The mean upwelling velocity in the simulation reaches \( O(10^{-4}) \) m s\(^{-1}\) and is confined to several tens of kilometers from the coast. The mean offshore flow of \( \sim 5 \) cm s\(^{-1}\) is confined to the upper 20 m and reflects largely the Ekman transport, consistent with field observations [cf. Lentz, 1992] and previous numerical studies [Marchesiello et al., 2003]. The mean nitrate off the central California coast shows an upward deflection of the nutrient-rich layer toward the coast (Figure 1c). This is primarily caused by the coastal upwelling, which tilts the isopycnals (Figure 1a) and consequently also the nutrient isolines (Figure 1c). This tilt is enhanced by strong local remineralization of vertically exported organic matter in the nearshore regions. The mean amount of total organic carbon (TOC, i.e., the sum of phytoplankton, zooplankton, large and small detritus in the simulation, converted from nitrogen to carbon using a fixed C:N ratio of 6.625) has a strong offshore gradient with the local vertical maximum progressively deepening with increased offshore distance (Figure 1d). TOC extends substantially deeper than net primary production (NPP) (Figure 1b) reflecting largely the sinking of the organic matter after it has been produced.

An additional substantial contribution to this widening and deepening of the TOC distribution stems from the net offshore TOC transport, which occurs predominantly in the upper 100 m and from the coast to more than \( \sim 600 \) km offshore (Figure 1f). In the near coastal 50 km, a weak onshore transport of the TOC can be seen at 30–100 m depth (Figure 1f). Also the mean advective vertical flux of TOC contributes to the widening of the TOC distribution, as there is a dominant net advective downward flux of TOC in the 100–500 km offshore region from \( \sim 20 \) m depth downward (Figure 1h). In the nearshore 50 km, the strong upwelling causes a dominant upward transport (Figure 1h). These vertical advective fluxes of TOC modify the vertical export driven by sinking particles substantially, leading to a complex mosaic of total vertical export (see also discussion in Plattner et al. [2005]).

The mean advective offshore and vertical transport of nitrate differs strongly from that of TOC, reflecting the inverse relationship between these two variables as the formation of TOC occurs with a corresponding uptake of nitrate, while the remineralization of organic matter forms nitrate. Clearly striking is the relatively strong offshore transport of nitrate occurring throughout the section (Figure 1e). In the first 0–200 km, this offshore transport is predominantly occurring in the top 30 m depth, while from \( \sim 300 \) km offshore, most of this transport occurs below 50–200 m. There, the offshore transport is superimposed by a substantial downward flux (Figure 1g).

The climatological mean states revealed in the vertical sections raise several questions. Perhaps most important is the question how TOC and nitrate can be transported all the way out to several hundred kilometers...
Figure 1. Vertical offshore sections of mean states off central California. The variables are averaged from Pt. Conception to C. Mendocino in alongshore direction (Figure 2) and represent temporal means from year 6 to year 10. (a) Potential density $\sigma_v$ with superimposed velocity vectors (scales for flows are given for 0.5 m $\text{d}^{-1}$ and 5 mm $\text{s}^{-1}$ for vertical and horizontal direction, respectively); (b) primary production (mmol N m$^{-3}$ s$^{-1}$); (c) nitrate (NO$_3$) (mmol N m$^{-3}$); (d) total organic carbon (TOC) (mmol C m$^{-3}$); (e) across-shore flux of NO$_3$ (mmol N m$^{-2}$ s$^{-1}$) (positive in onshore direction); (f) across-shore flux of TOC (mmol C m$^{-2}$ s$^{-1}$); (g) vertical flux of NO$_3$ (mmol N m$^{-2}$ d$^{-1}$) (positive in upward direction); (h) vertical advective flux of TOC (mmol C m$^{-2}$ d$^{-1}$).
from the shore given that the upwelling front, located at 100–200 km distance from the shore (Figure 1a), will impede the Ekman transport.

A plan view of the mean horizontal flow averaged over the top 100 m shows large alongshore variations (Figure 2a). This was hidden in the vertical section due to the alongshore averaging. The standing meanders extend over a few hundred kilometers in across-shore direction and create distinct regions of onshore and offshore flows even in the mean state (see also discussions in Marchesiello et al. [2003] and Plattner et al. [2005]). TOC averaged over the top 100 m is very high near the coast, but drops abruptly across the jet of the California Current (Figure 2a). The pattern of primary production (Figure 2b) is somewhat more contracted than that of TOC, reflecting the offshore transport that extends TOC beyond where it is being produced. In contrast, the pattern of the sinking flux at 100 m depth (Figure 2c) very closely mimics that of TOC concentration. This is expected given our use of constant sinking velocities.

The strong dropoff of TOC across the California Current may support the idea that it may act as a barrier to transport matter to the offshore region [Brink et al., 1995; Smith, 1995]. But the across-shore transport of TOC in the top 100 m (Figure 3a) shows a much more complex structure with many bands of alternating offshore and onshore fluxes. In the first 200–300 km from the shore, the offshore transport bands seem to extend nearly normal to the shoreline, but then turn nearly zonally thereafter. This distinct patterning illustrates that the offshore transport does not occur in a meridionally smooth pattern as might be inferred from the pattern of the Ekman transport, but that there are distinct standing transport pathways. The mean alongshore transport of TOC in the top 100 m is southward within the first 300 km from the coast except for a band of northward transport along the coast up to C. Mendocino (Figure 3d). The southward transport is largely due to the large-scale mean flow of the California Current, while the nearshore northward transport is due to the California Undercurrent that hugs the coastline over the upper slope at depth <300 m. The vertical advective TOC flux at 100 m depth is meridionally much more homogeneous and shows the consequence of upwelling near the coast, and downwelling over most of the offshore domain with some patchiness (Figures 3b and 3c).

The net impact on the horizontal advective TOC fluxes, illustrated here by the horizontal advective flux divergence (Figures 3e and 3f), shows that the advection effectively removes TOC from the coastal region and adds it to the offshore region, increasing the TOC concentration there. The large amount of TOC taken away from the coast is balanced by an excess of production in the nearshore region over vertical losses by sinking and vertical mixing. In contrast, in the offshore region, the excess of TOC contributed to by horizontal transport leads to an enhanced downward transport by both sinking (Figure 2c).
and advection (Figure 3b). Thus, the offshore transport is a dominant component of any local carbon budget in the CalCS.

How does this offshore and then downward transport occur? The prominent presence of alternating offshore and onshore fluxes implies that filaments and westward propagating mesoscale eddies are the dominant players for the offshore transport. In order to corroborate this hypothesis, we first conduct a Reynolds decomposition analysis, i.e., an analysis that separates the transport into one driven by the mean circulation, and one driven by the eddies and other time-fluctuating circulation features, such as filaments.

3.2. Reynolds Fluxes

In the Reynolds decomposition method [see, e.g., Mc Gillicuddy et al., 2003; Gruber et al., 2011], one equates the covariance terms of the time-fluctuating parts within the transport equations as the eddy fluxes. The origin of this term can be readily seen when considering the tracer conservation equation for the temporal mean of tracer $B$. 

Figure 3. Maps of mean advective fluxes of TOC for the top 100 m. (a) TOC flux in across-shore direction (mmol C m$^{-2}$ s$^{-1}$) (average over top 100 m), (b) TOC flux in vertical direction at 100 m depth (mmol C m$^{-2}$ d$^{-1}$), (c) TOC flux in alongshore direction (mmol C m$^{-2}$ s$^{-1}$) (average over top 100 m), and (d) horizontal flux divergence of TOC averaged in the upper 100 m (mmol m$^{-3}$ d$^{-1}$). These mean fluxes and the flux divergence are spatially averaged using a 60 km $\times$ 60 km grid. In Figure 3e, red indicates an increasing and blue a decreasing tendency. The 95% confidence intervals for the means using Gaussian fitting are given for (c) the vertical TOC flux and (f) the horizontal flux divergence.
where $S$ is the mean sink and source term of $B$, $u_j$ is the velocity tensor, $x_j$ is the spatial coordinate tensor, and where $K_m$ represents subgrid-scale eddy diffusivity. In (1), the second term on the right-hand side, i.e., the covariance between the fluctuations of the velocity ($u_j$) and that of the tracer concentration ($B$) represents the eddy transport term.

We obtained the mean fields by seasonally averaging the 2 day average output for 5 consecutive years (Spring: months 3–5; Summer: 6–8; Fall: 9–11; Winter: 12–2) and then subtracted these means from the instantaneous values to obtain the time-fluctuating components in (1). The resultant eddy components include thus the contribution of any temporally fluctuating transport with time scales between 2 days and several weeks. Through this choice of averaging, we exclude seasonal fluctuations as well as submesoscale variations of less than 2 days. The eddy fluxes of nitrate and TOC are, then, averaged for 5 years and from Pt. Conception to C. Mendocino.

The vertical section of the eddy-driven vertical flux of nitrate averaged in the alongshore direction shows that eddies act to subduct nitrate in the upper 100 m within the first 100 km from the coast (Figure 4b). They thus transport nitrate in the opposite direction as the mean flux, which is upward and driven by the coastal upwelling (Figure 4f). Further offshore, the eddy vertical flux of nitrate deepens, and although it is weaker than that in the inshore regions, it actually dominates the overall vertical flux in the top 100 m (Figure 4b).

The eddy-driven vertical flux of TOC is less confined to the shore than that of nitrate and extends out to 300 km offshore (Figure 4a), while keeping its maximum near the surface. The eddy vertical fluxes of TOC dominate the vertical fluxes of TOC throughout the section (Figure 4a), with the important exception of the first 100 km, where the upwelling transports a substantial amount of TOC upward (Figure 4e).

This pattern of eddy-induced subduction is very similar to our frontality-based results shown later, where we analyze the vertical fluxes in terms of their association with either the buoyancy gradient or the Okubo-Weiss parameter (5). We show that most of the subduction occurs in conjunction with straining fronts and filaments at distances of less than 100 km from the coast for nitrate, and that this subduction is somewhat broader for TOC. This consistency indicates that the eddy-induced downward fluxes of nitrate and TOC inferred from the Reynolds decomposition is largely a consequence of the subduction of tracers at fronts and filaments near the coast.

The vertical section of the eddy-driven horizontal flux of nitrate (Figure 4d) reveals that eddies cause a very strong offshore transport whose maximum is located just below the euphotic zone and largely follows the isopycnal layers. This eddy-induced offshore transport is much larger than that by the mean (Figure 4h), supporting our eddy-tracking-based analysis in the following section, where we show that a substantial amount of nitrate found in the upper 100 m in the offshore region is associated with westward traveling cyclonic eddies.

Eddies also induce a large offshore transport of TOC (Figure 4c), but compared to nitrate, it is mostly located within the top 100 m and also is of overall similar magnitude as the corresponding offshore mean flux (Figure 4g). In the upper 20 m, the mean flux dominates with the eddy flux contributing only about 20–25% to the total transport, while below between 20 and 100 m depth, the eddies actually dominate.

Overall, the Reynolds decomposition shows the critical role of mesoscale and finer-scale processes in transporting TOC and inorganic nitrate to the offshore. The transport pathways of nitrate and TOC differ in that TOC is largely transported offshore near the surface and carried by both the mean and the eddy components of the total transport. In contrast, nitrate is subducted by eddies in the first 100 km and then transported further offshore almost exclusively by eddies. This different behavior is solely due to the different dynamics between the two tracers with TOC having a near-surface source and a sink at all depths, whereas nitrate has only a sink in the upper ocean, and a source at all depths, leading to very different time-mean distributions (see Figures 1c and 1d).

The strong offshore transport of nitrate by eddies has been previously analyzed by Gruber et al. [2011] in the context of their investigation of the role of eddies in controlling productivity in EBUS. Gruber et al. [2011]
showed that this offshore transport leads to a substantial depletion of the nitrate inventory in the nearshore upwelling waters, and hence to an eddy-induced reduction of primary production in the nearshore areas. This loss of primary production is not compensated by an increase further offshore, as the offshore transport...

Figure 4. Vertical section of the vertical and across-shore transport of nitrate and TOC by the eddy and the mean flux components diagnosed from Reynolds decomposition for the central California Coast. (a) Vertical eddy flux of TOC (mmol C m\(^{-2}\) d\(^{-1}\)), (b) vertical eddy flux of nitrate (mmol N m\(^{-2}\) d\(^{-1}\)), (c) across-shore eddy flux of TOC (mmol C m\(^{-2}\) s\(^{-1}\)), (d) across-shore eddy flux of nitrate (mmol N m\(^{-2}\) s\(^{-1}\)), (e) mean vertical flux of TOC (mmol C m\(^{-2}\) d\(^{-1}\)), (f) mean vertical flux of nitrate (mmol N m\(^{-2}\) d\(^{-1}\)), (g) mean across-shore flux of TOC (mmol C m\(^{-2}\) s\(^{-1}\)), and (h) mean across-shore flux of nitrate (mmol N m\(^{-2}\) s\(^{-1}\)). The sum of the respective mean and eddy components is identical to total net fluxes shown in Figure 1. Mean fields for the Reynolds decomposition were obtained by seasonal (3 months) averaging the data over the 5 year analysis period. For the sections, the fluxes were averaged between Pt. Conception and C. Mendocino in alongshore direction.
transport occurs too deep in the water column, enriching nitrate at depth, too deep for phytoplankton growth [Gruber et al., 2011].

### 3.3. Eddy Stream Function

In order to isolate the circulation underlying the eddy-induced transports, we diagnose the eddy stream function. Here the eddy stream function includes contributions from filaments and mesoscale eddies. It is computed based on the Transformed Eulerian Mean theories [Andrews and McIntyre, 1978; Plumb and Ferrari, 2005], i.e.,

\[
\psi_e = \frac{\mathbf{u}' \times \nabla \mathbf{b}}{|\nabla \cdot \mathbf{b}|^2},
\]

where \(\psi_e\) is the eddy-induced stream function, which forms a divergence of the skew flux of the form, \(J(\psi_e, b)\), overbar represents along front mean, \(J\) is the Jacobian operator, \(J(a, b) = a_0 b_2 - a_2 b_0\), \(\mathbf{u}' = (u', w')\) represents the eddy fluctuating velocity vector, and \(\nabla \cdot \mathbf{b} = (\partial_x b, \partial_z b)\). Plumb and Ferrari [2005] pointed out that the vertical coordinate needs to be stretched to account for the large aspect ratio of the oceanic flow (\(\partial_z b / \partial_x b\)). We use the following form with the stretching factor, \(\gamma\), set as \(10^3\),

\[
\psi_e = \frac{(\nabla' \cdot \mathbf{b}) b_z - \gamma^2 (\nabla' \mathbf{b}) b_y}{\gamma^2 b_y^2 + b_z^2},
\]

where the overbar represents temporal (5 years) and alongshore (from Pt. Conception to C. Mendocino) averaging, assuming that influences from the alongshore gradients become small after temporal averaging. The same form has been examined in a number of numerical studies that showed eddy-induced circulation contributing to the restratification of fronts of the interior and of the surface mixed layer [Cerovečki et al., 2009; Mahadevan et al., 2010].

The diagnosed mean stream functions in Figure 5a show the well-known mean across-shore circulation as seen already in Figure 1a. It consists of a relatively broad anticlockwise circulation cell with a near-surface flow that is directed offshore (largely) due to Ekman transport, and a broad onshore flow at depth with upwelling near the coast. In contrast, the eddy-induced stream function, shown in Figure 5b and obtained from (3), has a clockwise rotation. Its structure is dominated by a shallow closed cell, leading to eddy-induced lateral transport toward the shore in the surface layer, strong downward motion near the coast and offshore transport in the shallow subsurface.

The strength of the eddy-induced circulation can overcome the Eulerian mean circulation near the surface, which suggests that restratification of the mixed layer front found in recent numerical studies [Capet et al., 2008b; Fox-Kemper and Ferrari, 2008; Mahadevan et al., 2010] is actually active in our simulation, despite our

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**Figure 5.** Vertical offshore sections of the stream functions for the central California Coast. (a) Mean across-shore circulation and (b) mean eddy-induced circulation estimated using (3). A positive and negative value of the stream function denotes clockwise and anticlockwise rotation, respectively. The solid black contours represent isopycnal layers, and cyan and magenta contours are for better clarity of the distribution of the stream functions.
model only having been run at a resolution of 5 km. Similar to Colas et al. [2013], the eddy-induced stream function exhibits an across-shore gradient, which suggests that eddy-induced vertical circulations may transport heat and other tracers vertically near the surface.

Superimposing the eddy-induced circulation (Figure 5b) with the time-mean distributions of nitrate and TOC (Figures 1c and 1d) largely explains the eddy-induced fluxes of these constituents (Figure 4), with the exception of the lateral eddy flux of TOC seen in Figure 4c. The eddy-induced fluxes suggest an offshore transport down to ~100 m depth, while Figure 5b suggests an eddy-induced onshore circulation in the near surface. This is because eddy-induced surface positive onshore flow tends to be associated with a negative anomaly of offshore TOC, resulting in offshore (negative) eddy TOC transport near the surface as in Figure 4c. Similarly, eddy-induced vertical TOC flux is downward over 500 km toward offshore region (Figure 4a), as TOC is a surface-intensified tracer. Thus, while eddy-induced circulation tends to flatten the isopycnal tilt of the upwelling front, their contributions to the total transport depends on the across-shore and vertical distributions of tracers as well.

4. Transport Caused by Filaments and Mesoscale Eddies

Results from the Reynolds decomposition analysis suggest that the offshore transport of nitrate and TOC are largely driven by eddies and filaments. However, as the decomposed fluctuating part includes variabilities due to both mesoscale eddies and filaments, it is difficult to assess the detailed roles of each phenomenon. Accordingly, we next analyze the role of filaments for the offshore transport, particularly with regard to the process of subduction, and then in a second step look at the role of the westward propagating eddies.

4.1. Filaments

Filaments are frequently observed in our simulation, representing a primary determinant of the surface distribution of the biological tracers in the coastal region. The typical offshore extent of the filaments is about 200 km, except in the filament-favorable spots, such as off C. Mendocino. There, a larger offshore extent of the filaments can be seen caused by a combination of the influence from the Mendocino Ridge and the existence of the cape [Haidvogel et al., 1991]. A snapshot for a summer situation (18 July of model year 9) (Figure 6) shows three major filaments, i.e., one off C. Blanco, one off C. Mendocino, and one off San Francisco Bay, reflecting the most important regions for the formation of filaments both in the model and in reality [Moisan et al., 1996; Marchesiello et al., 2003].

At the surface, the filaments are characterized by narrow jets of cold and nutrient-rich waters extending deep into the offshore regions (Figures 6a and 6b). Below these narrow jets one finds wider mounds of cold and nutrient-rich water, representing major upward lifts of the upper ocean thermocline. These modeled tracer distribution near the filaments is consistent with the observations by Cravo et al. [2010]. The filaments are rich in TOC (Figure 6c), but its distribution is more diffuse compared to that of nitrate and temperature. The narrowness of the surface cold filaments is maintained by strong eddy straining near the surface, which creates sharp surface fronts with upwelling and downwelling flows [McWilliams et al., 2009]. Although not fully resolved by our grid spacing of 5 km, the model appears to simulate two different patterns of vertical flows (Figure 6d). The first one near the surface, with scales of ~O (10 km), that can be attributed to secondary circulation associated with the submesoscale mixed layer front [Mahadevan and Tandon, 2006; Fox-Kemper and Ferrari, 2008], and the second one more prevalent below in the upper thermocline with larger scales >O (10 km) and likely caused by mesoscale forcing.

The downwelling and the upwelling in the upper thermocline are located mostly at the northern and the southern part of the filament, respectively (Figure 6d and Figure S1a of supporting information), consistent with Moisan and Hofmann [1996a]. This downward transport at the filaments creates subsurface maxima of TOC (Figure 6c) similar to those found in observations [Barth et al., 2002; Bograd and Mantyla, 2005]. As the filament is progressing offshore, the upwelling and downwelling pattern moves offshore as well and eventually will be isolated at the offshore tip of the filament, which, then, releases a cyclonic eddy that contains filament cold water and anticyclonic eddies nearby [Mied et al., 1991; Moisan et al., 1996].

We investigate what controls the locations of the upwelling and downwelling near the filaments by computing the divergence of the Q-vector [Hoskins et al., 1978] i.e.,
where $\nabla_h \cdot \mathbf{Q} = -\frac{\partial}{\partial x} \left( \frac{\partial}{\partial x} \frac{\partial b}{\partial x} + \frac{\partial}{\partial y} \frac{\partial b}{\partial y} \right) - \frac{\partial}{\partial y} \left( \frac{\partial}{\partial x} \frac{\partial b}{\partial x} + \frac{\partial}{\partial y} \frac{\partial b}{\partial y} \right)$.

Figure 6. Three-dimensional representation of the spatial variations of the physical and biogeochemical properties associated with filaments found in the simulation at 18 July, year 9, for (a) temperature (°C), (b) nitrate (mmol N m$^{-3}$), (c) TOC (mmol C m$^{-3}$), and (d) vertical velocity (m d$^{-1}$). Horizontal scales are given at the left-top and the left-bottom corners.
associated frontogenesis and frontolysis. There is not much seasonality for this relationship. This implies that the net downward TOC transport found in the upper 100 m (Figure 1d) is partially explained by the strong vertical flows associated with the fronts and/or the filaments.

This is examined by averaging the vertical fluxes of TOC and nitrate at 50 m depth, and of the associated physical parameters as a function of distance from the coast and the magnitude of the horizontal buoyancy gradient \( \nabla_h b \). For the physical parameters, we consider the relative vorticity at 50 m depth, \( \zeta = \nabla \cdot \mathbf{u}_r \), and the frontogenetic parameter, \( FP = \frac{Q \cdot \nabla_h b}{|\nabla_h b|} \). The frontogenetic parameter is useful to know whether the front is under frontogenesis (\( FP > 0 \)) or frontolysis (\( FP < 0 \)) as it is equivalent to twice the Lagrangian temporal derivative of the absolute buoyancy gradient, \( 2D_t |\nabla_h b| \) [Bluestein, 1993]. Given the transitory nature of the filaments, we performed this analysis using the 2 day output over the entire 5 years of analysis.

The large downward (negative) fluxes of nitrate and TOC at 50 m depth occur largely within the first 100 km from the coast and where the horizontal buoyancy gradient is large, i.e., near fronts (top left corner in Figures 7a and 7b). This subduction coincides with positive relative vorticity, i.e., cyclonic rotation (Figure 7e) as well as with a tendency for frontogenesis (Figure 7f). The dominant cyclonic vorticity at the sharpest part of the

**Figure 7.** Plots of vertical fluxes, relative vorticity, and frontogenetic parameter at 50 m depth as functions of the absolute lateral buoyancy gradient, \( |\nabla_h b| \) (ordinate) \( (\times 10^{-6} \text{ s}^{-2}) \) and distance from the coast (abscissa) (km). (a) Vertical fluxes of nitrate (mmol N m\(^{-2}\) d\(^{-1}\)), (b) vertical fluxes of TOC (mmol C m\(^{-2}\) d\(^{-1}\)), (c) \( \zeta / f \), i.e., the relative vorticity normalized by the local Coriolis parameter, and (d) the frontogenetic parameter, \( Q \cdot \nabla_h b / |\nabla_h b| = \frac{Q \cdot \nabla_h b}{|\nabla_h b|} \) \( (\times 10^{12} \text{ s}^{-2}) \). The solid contours indicate the integrated area in common logarithmic scale \( \log_{10}(\text{km}^2) \), over which the corresponding data are averaged. (c and d) Total vertical transport at 50 m depth for nitrate and TOC, respectively.
front on average is consistent with the asymmetry in the vorticity and ageostrophic straining often found on the dense side of the surface-intensified front [Capet et al., 2008b]. This suggests that the dense side of the narrow frontal regions under the frontogenesis is the dominant subduction site within the first 100 km from the coast. This results are confirmed by Lagrangian analyses where 30–40% of 100,000 passive Lagrangian floats released in the upper 20 m within 15 km from the coast subducted below 50 m depth after 4–6 months (Figure S2 of supporting information), and consistent with a recent numerical and field study by Omand et al. [2015], which showed that submesoscale subduction at fronts promotes particulate organic carbon exports. On the other hand, relatively weak upward tracer fluxes due to coastal upwelling occur within 30–50 km from the coast with weaker frontal buoyancy gradient over a larger area than the narrow subduction dominant cyclonic portions (Figures 7a and 7b). The mean upward transport, obtained by multiplying average area with fluxes, dominates over the subduction transport within 100 km from the coast (Figures 7c and 7d).

In order to determine the type of front responsible for the subduction, we repeated the above analysis, but replaced the absolute horizontal buoyancy gradient on the ordinate with the Okubo-Weiss parameter. This parameter has been utilized widely to detect oceanic mesoscale eddies as it separates any two-dimensional flow into a rotating regime when \( \text{OW} < 0 \), and into a straining regime when \( \text{OW} > 0 \) [Okubo, 1970; Weiss, 1991; McWilliams, 1984]. This parameter is defined as:

\[
\text{OW} = 4 \left( \frac{\partial u}{\partial x} \right)^2 + \frac{\partial v}{\partial x} \frac{\partial u}{\partial y}.
\]  

(5)

Figure 8 reveals that the large subduction fluxes of nitrate and TOC are found favorably in a straining flow regime (\( \text{OW} > 0 \)) with moderately large cyclonic vorticity and a frontogenetic tendency, which is typically seen on the dense side of isolated eddies and fronts/filaments under confluence. When the flow is in a rotating regime, i.e., when \( \text{OW} < 0 \), the cyclonic vorticity dominates (Figure 8e), but the vertical fluxes tend to be less than that in a straining regime, and directed both upward and downward. These results further illustrate that most of the subduction of tracers occurs within the first 200–300 km of the coast and at narrow fronts/filaments, of straining flow, and of frontogenetic tendency.

4.2. Westward Propagating Eddies

While filaments clearly dominate the offshore transport in the first 100 km from the shore and are also crucial for the subduction of organic matter and nitrate, it must be the westward propagating mesoscale eddies that move the material properties further offshore.

In order to illustrate this, we use Hovmoeller diagrams, i.e., plots of various parameters averaged from Pt. Conception to C. Mendocino alongshore direction as a function of time and offshore distance. Figure 9a clearly shows the westward propagation in Sea Surface Height (SSH) that has been observed in satellite data in the CalCS [Kelly et al., 1998] and has been discussed in a comparable modeling study by Marchesiello et al. [2003]. The low SSH anomalies depart from the coast in the summer upwelling season, travel westward, and reach an offshore distance of 600–800 km about a year later. The propagation speed is, therefore, \( \sim 2 \text{ cm s}^{-1} \), which is equivalent to the theoretical speed of the first baroclinic mode of the Rossby waves in this region [Killworth et al., 1997]. These westward signals seen in low SSH are the result of westward traveling eddies. The propagation is also clearly visible in the mean vertical velocity at 100 m depth (Figure 10a) and salinity (Figure 9c), but not clearly in temperature (Figure 9b), whose distribution is dominated by the seasonal variations, likely because the heating and the cooling through the surface mask the westward signals.

A striking observation is the dominance of downwelling seen in the westward propagations in the offshore region about 200–500 km from the coast. A possible explanation for this observation is the vortex stretching that occurs as the eddies travel westward and experience a deepening pycnocline. The vertical velocity caused by such a stretching can be roughly estimated from the speed of offshore travel and the mean across-shore slope of the isopycnal layers. Figure 1a shows that the \( \sigma_\theta \sim 25 \text{ kg m}^{-3} \) isopycnal descends about 40 m over an offshore distance of \( \sim 300 \text{ km} \) (between 200 and 500 km offshore), which a baroclinic eddy may take 6 months to travel across. Assuming an along-isopycnal adiabatic flow, the magnitude of the vertical velocity can be estimated as \( \sim 0.1 \text{ m d}^{-1} \), which is in good agreement with the model results depicted in Figure 10a.
In fact the westward deepening of the pycnocline, together with the presence of a front formed by the California Current flowing southward, would cause a northward deflection of the propagating eddies. This is because linear Rossby waves travel along potential vorticity contours. However, the simulated eddies/filaments initially travel west-southwest (normal to the coast, see Figures 3a and 12) and then strictly west, requiring them to stretch. This stretching causes not only a net-downwelling, but also an increase in the vorticity, $f$, which they need in order to conserve potential vorticity.

The inorganic nutrients (NO$_3$ + NH$_4$) averaged over the upper 100 m also clearly show the westward propagation of high concentration that starts in the summer near the coast in response to the coastal upwelling, and extends 500–800 km offshore (Figure 9d). The offshore propagation of the inorganic nutrients with the speed of the Rossby waves suggests that the nutrients in the offshore region (~200–600 km) are supplied by the mesoscale eddy fluxes. If this was the case, the nutrients available for the biological production in the offshore region (~600 km from the coast) around early spring would be partially supported by eddies having transported the nutrients from the coast since the previous summer. The TOC also shows some propagation, although not as clearly as the nonsinking constituents, such as nitrate (Figure 9e). This is not surprising since the leaking of TOC from eddies by sinking is bound to blur the westward propagating signals of TOC. Following the pattern of vertical velocity, the vertical flux of the inorganic nutrients (NO$_3$ + NH$_4$) and TOC at 100 m depth are similarly influenced by the westward propagating eddies (Figures 10b and 10d).

Figure 8. The same plots as in Figure 7 except that the ordinate was replaced by the Okubo-Weiss parameter ($\times 10^{-9} \text{s}^{-2}$), a measure of whether any two-dimensional flow is in a rotating regime (OW < 0) or in a straining regime (OW > 0).
Offshore of about 100 km, the downward fluxes of these tracers are entirely associated with the westward propagating signals. This is also the case for the lateral fluxes, which in fact are nearly everywhere directed in offshore direction (Figures 10c and 10e).

Figure 9. Hovmöller diagrams of physical-biogeochemical tracers averaged from Pt. Conception to C. Mendocino. Abscissa shows distance from the coast, and ordinate indicates model time in years. (a) Sea surface height (m); (b) sea surface temperature (°C); (c) salinity; (d) nitrate + ammonium (mmol N m⁻²); (e) TOC (mmol C m⁻²).

Figure 10. The same Hovmöller diagrams as Figure 9, but for (a) vertical velocity at 100 m depth (m d⁻¹), (b) vertical flux of TOC at 100 m depth (mmol C m⁻² d⁻¹), (c) across-shore flux of TOC averaged over the top 100 m (mmol C m⁻² s⁻¹), (d) vertical flux of nitrate + ammonium at 100 m depth (mmol N m⁻² d⁻¹), and (e) across-shore flux of nitrate + ammonium averaged over the top 100 m (mmol m⁻² s⁻¹).
Similar to Figures 7 and 8, top 150 m mean across-shore tracer transport for nitrate and TOC are cumulatively summed as a function of relative vorticity and OW (Figure 11). The results indicate that dominant offshore fluxes of both nitrate and TOC are mostly associated with cyclonic vorticity ($\zeta/f > 0$) in rotating regime ($\text{OW} < 0$) in summer season (Figures 11a and 11c), suggesting that cyclonic eddies are dominant agents to transport nitrate and TOC offshore in the upwelling season. This can be explained by background across-upwelling front contrasts in near-surface nitrate and TOC with larger amounts found on the inshore side or cyclonic side of the upwelling front, where these tracers tend to be held inside the cyclonic eddies. However, this tendency is weakened in the following fall, in which flow with anticyclonic rotation turns to transport nitrate and TOC with similar magnitude to that due to cyclonic rotating flow (Figures 11b and 11d). This is probably because during their westward traveling, tracers inside the cyclonic eddies leak, and some of this may entrained into anticyclonic eddies.

These results strongly suggest that the westward propagating eddies have the ability to transport inorganic nutrients and TOC to great offshore distances and in a net-downward direction. These results are thus consistent with the eddy fluxes of these tracers obtained by Reynolds decomposition analyses. We next solidify our interpretation that these offshore propagating signals stem from offshore propagating eddies, using the eddy tracking analysis. This permits us also to identify the relative roles of cyclonic and anticyclonic eddies.

4.3. Eddy Tracking
To investigate the role of the eddies in more detail, we identified and tracked individual eddies over the 5 year duration of our analysis period. We use closed contours of the Okubo-Weiss parameter (equation 5) at

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Figure 11. Sum of top 150 m averaged cross-shore transport that occurs in all the grid points for 5 years between Pt. Conception and C. Mendocino within 150–500 km offshore region as a function of top 150 m averaged Okubo-Weiss parameter ($x \times 10^{-2}$ s$^{-2}$) and normalized vorticity $\zeta/f$ for (a and b) across-shore transport of nitrate (kmol N s$^{-1}$) and (c and d) that of TOC (kmol C s$^{-1}$) in (a and c) summer (July, August, September) and (b and d) fall (October, November, December). The lateral fluxes are obtained by averaging the fluxes computed at every time step.

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for the eddy detection, with the OW parameter computed from 2 day averages of the mean horizontal flow over the top 100 m. The OW method tends to detect distinct eddy vortices, disregarding other mesoscale phenomena, such as smaller distorted filament-like eddies that occur, e.g., near the coast. As a result, this identification method tends to give a lower bound estimate of the total number of eddies. We decided to use the OW method over others [e.g., Nencioli et al., 2010], (i) since it has been widely used [e.g., Isern-Fontanet et al., 2003; Frenger et al., 2013; Kurian et al., 2011] resulting in a good understanding of its detection characteristics, and (ii) since it tends to isolate those eddies that are poised to trap material properties [Zhang et al., 2014].

Several studies have pointed out the importance of mode water eddies in the CalCS [e.g., Cornuelle et al., 2000; Collins et al., 2004; Molemaker et al., 2015], also known as California Undercurrent Eddy (hereinafter “cuddy”). Accordingly, closed contours of OW averaged between 100 and 200 m depth are used to detect these cuddies. Specifically, we identified cuddies on the basis of the averaged vorticity between 100 and 200 m depth within a closed contour exceeding that averaged over top 100 m. The detected eddies and cuddies were then screened on the basis of equivalent radius, extracting them with radiuses within the range between 15 and 200 km. The screened eddies and cuddies were then tracked in time by comparing the shared area between two consecutive time steps among eddies. For the final analysis, we retained only those eddies and cuddies that were able to track for at least 30 days.

The trajectories of the tracked and retained eddies in Figure 12 reveal their persistence as well as their westward propagation over long distances, in general agreement with the results of Chelton et al. [2007, 2011b]. We identified virtually no eddies in the nearshore 0–50 km. Although this is in part a consequence of our detection method (see above), we interpret this to be primarily due to the fact that most eddies are being formed at about 100–200 km distance from the coast, as this is where the filaments and the California Current emit the isolated eddies. Both cyclonic and anticyclonic eddies populate the entire CalCS without a clear spatial preference of one over the other, again in agreement with observations [Chelton et al., 2007, 2011a, 2011b]. In contrast, the number of cyclonic cuddies detected by this method is very small compared to that of anticyclonic cuddies, consistent with Kurian et al. [2011]. The westward travels are confirmed for the tracked eddies with the speeds similar to that of the first baroclinic mode Rossby waves, suggesting westward propagating Eulerian signals are attributed to the mesoscale eddies.

On average, we identified and tracked about 30–40 cyclonic and anticyclonic eddies at any given time off the Central California Coast (Figure 13a). There seems to be 15–25% more cyclonic than anticyclonic eddies in all seasons, consistent with Kurian et al. [2011]. The t test with a null-hypothesis of equal mean is rejected for the time mean with 95% confidence interval, while the same hypothesis is not rejected for monthly mean of number of eddies for February–March, June, and November (Figure 13a). A slightly higher number of cyclonic eddies compared to anticyclonic ones is possibly due to centrifugal instability for anticyclones [Molemaker et al., 2005]. However, the lifetime of cyclones and anticyclones is similar (41.4 days versus 40.2
days) between these means, suggesting that the asymmetry in numbers may relate to differences in forma-

tion rates.

Despite this 10% difference in their numbers, the amount of the nitrate inside cyclonic eddies is a factor of 2–3 greater than that inside of anticyclonic eddies, and the amount of the TOC inside cyclonic eddies is more than twice as large compared to that inside anticyclonic eddies (Figures 13b and 13c). We interpret this to be largely a consequence of the formation mechanism of the eddies, with the cyclonic eddy encapsulating the material properties of the cold, nutrient-rich filament from which it emanates, while the anticyclonic eddy tends to encapsulate the material properties of the nutrient-poorer waters surrounding the filament.

The number of cuddies at any given time is 5–15 and 3–5 for anticyclonic and cyclonic cuddies, respectively. The amount of TOC and nitrate within anticyclonic cuddies in top 100 m is 10–40% of that within anticyclonic eddies.

Figure 13. Statistics of detected and tracked (solid blue) cyclonic, (solid red) anticyclonic eddies, (dashed-blue) cyclonic cuddies, and (dashed-red) anticyclonic cuddies. (a) Number of eddies in every month with standard deviations. (b) Amount of nitrate (10 G mol) found inside the eddies and cuddies, integrated over the top 100 m; (c) same as Figure 13b except for TOC (100 G mole); (d) histogram of the lifetime of eddies and cuddies; and (e) histogram of the nonlinearity of eddies and cuddies defined as $e = u/c$, where $u$ is current velocity averaged inside the eddy and cuddy over the top 100 m and $c$ is the translational speed of the traveling eddy ($\sim 2$ cm s$^{-1}$). Shown and analyzed are all eddies and cuddies that were detected, screened, and tracked for at least 30 days over the 5 years of the analysis period.
The average current velocities within cyclonic and anticyclonic eddies over the top 50 m are 8 cm s\(^{-1}\) and 7.2 cm s\(^{-1}\), respectively. This is much larger than the traveling speed of eddies, i.e., ~2 cm s\(^{-1}\), indicating that eddies are strongly nonlinear and have a strong tendency for trapping material properties inside (Figure 13e). The estimated nonlinearity is very similar to that computed by Kurian et al. [2011] for the surface flow. In contrast, nonlinearity of anticyclonic eddies is higher than that of cyclonic eddies, and much smaller than that of anticyclonic eddies (Figure 13e), suggesting that anticyclonic eddies can contribute to upper layer lateral fluxes.

The eddy tracking enables us to divide the modeled tracer field into a part within the eddies (separately for cyclonic and anticyclonic ones) and the remaining part outside them. Figure 14 shows that while the total amount of inorganic nutrients or TOC inside the eddies is relatively small in the first 100 km (largely due to the lack of eddies there), the detected eddies carry a very substantial fraction of the total amount of inorganic nutrients and TOC beyond this first 100 km. The absolute amount is largest at about 100–150 km, i.e., shortly after most of the eddies have spawned off, but the relative fraction increases with increasing distance from the coast. In the far offshore region, i.e., 300–600 km away, about 13–24% of the total amount of inorganic nutrients and TOC can be found inside the eddies. This is remarkable given that these eddies occupy only 5.2% of the total area. By far the largest contribution stems from cyclonic eddies, with the relative contribution of the anticyclones decreasing with greater distance from shore. Beyond a few hundred kilometers, essentially most of inorganic nutrients and TOC associated with eddies are found inside cyclonic ones. The amount of tracers inside the detected cuddies shows that anticyclonic cuddies carry 10–40% of the TOC and nitrate found within anticyclonic eddies (Figure 14). In contrast, the cyclonic cuddly contribution is very little (Figure 14).

The eddy tracking results clearly show the relatively larger role of cyclones compared to anticyclones in transporting organic matter and inorganic constituents toward the offshore region from the central
upwelling region near the coast. But given that our eddy detection criterion may disregard many eddies that travel in offshore direction carrying material properties inside, our eddy-tracking method used here may underestimate the contribution of eddies to the overall offshore transport. However, the strong congruence of these findings with those from the Reynold’s decomposition discussed above suggests that this is likely not the case.

5. Summary and Conclusion

We investigated the mechanisms that drive the offshore transport of organic matter and inorganic nutrients in the California Current System using a three-dimensional coupled physical-biogeochemical model forced by climatological fields. While we were not able to evaluate how well our model is simulating the offshore transport, the relatively good agreement between the simulated physical and biological variables and a large number of observational constraints [see Gruber et al., 2006, 2011] gave us confidence that the model is likely also performing reasonably well with regard to the simulation of offshore transport. Since the model is also able to capture the eddy and filament fields of the California CS [see also Marchesiello et al., 2003], we assumed that it is reasonable to analyze the contributions of the eddies and the filaments to the total offshore transport. We use a number of complementary approaches to elucidate these contributions, including (i) a Reynolds decomposition, (ii) a frontality-based method, (iii) a Lagrangian particle tracking approach, and (iv) an eddy identification and tracking method. The synthetic perspective proposed next (see also Figure 15) is based on the combination of the results from all these approaches.

Every spring to fall, the coastal upwelling induces strong phytoplankton growth near the coast, causing a strong increase in the total amount of organic matter, while not entirely drawing down all available nitrate [Lachkar and Gruber, 2013]. Given the strong density differences between the upwelled cold water and the warm offshore waters, a coastal upwelling front is formed that differentiates also the high-nitrate, and high-organic and inorganic carbon nearshore waters from the low-nitrate and low-organic and inorganic carbon waters of the offshore region. This upwelling front acts as a strong barrier to transport any material property across it. This barrier can be overcome by frequently formed filaments that extend from the coast to ~200 km offshore and can distort the front in across-shore direction, thus inducing substantial offshore fluxes. Strong subduction and upwelling occur in association with the filaments, forced by submesoscale and mesoscale straining for both the surface mixed layer front located in the near-surface ocean, and a deeper baroclinic front. Particularly when these fronts are frontogenic and sharp, a substantial fraction of the inorganic nutrients and organic matter present near these fronts is subducted, leading to a strong downward transport in the first 100 km off the coast, removing the remaining inorganic nutrients from the sunlit layer.

The filaments tend to generate anticyclonic eddies nearby and to break into cyclonic eddies at their offshore tips, which in turn, trap the upwelled water, including their abundant organic and inorganic constituents. The isolated cyclonic and anticyclonic eddies emitted from the central California coast can travel westward for long distances and periods, over a 1000 km and over a year. The average current velocity within eddies, 7–8 cm s$^{-1}$, well exceeds the traveling speed of the eddies, 2 cm s$^{-1}$, which is given by the speed of the first baroclinic mode of Rossby waves. As this swirl velocity within eddies is much larger than the translational speed of these eddies, they tend to trap material properties well inside them, transporting them along westward/offshore direction. While for organic matter, most of this transport occurs in the upper ocean, the nitrate is transported offshore by the eddies below the euphotic layer, outside the reach of phytoplankton in the offshore region.

This sequence of filament and then eddy-induced lateral transport tends to dominate the overall offshore transport, particularly for the inorganic nutrients, but also for total organic carbon. In fact, at least 13–24% of the organic matter and inorganic nutrients found in the upper 100 m of the far offshore region, i.e., 300–600 km from the coast, have been laterally transported there by eddies. Thus, the continental margins of coastal upwelling systems play an important role in supplying nutrients and organic matter to the adjacent open ocean regions, highlighting the relatively good connectivity between the coastal and the open ocean. Our results confirm the conclusions drawn by Gruber et al. [2011] about the negative impact of eddies on biological productivity in EBUS, as a larger amount of eddies would tend to cause more lateral and
downward export of inorganic nutrients to the offshore subsurface layers, therefore depleting the nitrate reservoir in the nearshore euphoric zone.

Our main caveat is the relative coarseness of our model setup. This prevents our model from properly simulating a number of potentially critical details, especially associated with the pronounced nutrient injections near fronts, which occurs at submesoscales. This caveat is particularly pronounced since recent studies showed the importance of such submesoscale circulations in the vertical exchanges [e.g., Klein and Lapeyre, 2009]. As shown in Omand et al. [2015], submesoscale subduction enhances particulate organic carbon exports; our 5 km resolution simulation may have underestimated the eddy-induced subduction of organic carbon. In addition, wind-forced submesoscale ageostrophic secondary circulations [Thomas and Lee, 2005], which may not be resolved by our 5 km model, could enhance the vertical exchanges of carbon and nutrients further [Nagai et al., 2008; Shulman et al., 2015]. Therefore, it would be the next step to extend our analyses to simulations with finer resolution. We also expect substantial year-to-year variations in the offshore transport, partially in response to the variations in nearshore productivity [e.g., Frischknecht et al., 2015], and partially in response to variations in ocean circulation and eddy activity.

Figure 15. Schematic diagram summarizing the dynamics of offshore transport of organic carbon in the California Current System as diagnosed by our model simulation. After being formed in the upwelled waters, a substantial fraction of the organic matter is transported offshore, first by filaments that protrude across the coastal upwelling front, which tends to act as a barrier for offshore transport. Much of the organic matter is subducted at the fronts associated with the filaments. From there, the organic matter is readily transported further offshore by westward propagating eddies, which form at an offshore distance of about 100–200 km. Most of the initial offshore transport in summer occurs by cyclonic eddies, largely reflecting their generation at the tip of the filaments. The organic carbon in the cyclonic eddies is then stirred and entrained also into anticyclonic eddies, which travel west as well. The eddy fluxes dominate the transport in the region 200–800 km from the shore.
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Please contact the corresponding author at tnagai@kaiyodai.ac.jp to obtain the data of the numerical model described in this paper.

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