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Cross-shore transport variability in the California Current: Ekman upwelling vs. eddy dynamics

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ABSTRACT

The low-frequency dynamics of coastal upwelling and cross-shelf transport in the Central and Southern California Current System (CCS) are investigated using the Regional Ocean Modeling System (ROMS) over the period 1965–2008. An ensemble of passive tracers released in the numerical model is used to characterize the effects of linear (Ekman upwelling) and non-linear (mesoscale eddies) circulation dynamics on the statistics of advection of coastal waters. The statistics of passive tracers released in the subsurface show that the low-frequency variability of coastal upwelling and cross-shelf transport of the upwelled water mass are strongly correlated with the alongshore wind stress, and are coherent between the central and southern CCS. However, the offshore transport of tracers released at the surface is not coherent between the two regions, and is modulated by intrinsic mesoscale eddy activity, in particular cyclonic eddies. The transport of cyclonic eddies extends with depth and entrains water masses of southern origin, advected by the poleward California Undercurrent (CUC). The CUC water masses are not only entrained by eddies but also constitute a source for the central California upwelling system. The interplay between intrinsic (eddy activity) and deterministic (Ekman upwelling) dynamics in controlling the cross-shelf exchanges in the CCS may provide an improved framework to understand and interpret nutrients and ecosystem variability.

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1. Introduction

The California Current System (CCS) has been extensively studied through several long-term and regional sampling programs and satellite analyses, including physical, chemical and biological analyses. The California Current (CC) is the eastern boundary current of the subtropical North Pacific and is characterized by a broad (1000 km offshore), shallow (surface to 500 m) and relatively slow (mean 10 cm s^{-1}) equatorward flow (Batteen et al., 2003). In the subsurface on the continental slope, the California Undercurrent (CUC) is a relatively narrow (10–40 km width) and weak ($2\text{--}10 \text{ cm s}^{-1}$) poleward flow centered between 100–300 m depth (Hickey, 1979, 1998). Despite extensive sampling conducted by the California Cooperative Oceanic Fisheries Investigations (CalCOFI), the coarse spatial and temporal resolution of the sampling leave us with an incomplete understanding of the cross-shore transport dynamics of surface and subsurface water masses.

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On interannual to decadal time scales (referred to as “low-frequency variability” throughout the text), large-scale climate modes such as the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997) and the El Niño–Southern Oscillation (ENSO) are used to explain physical fluctuations in the Northeast Pacific Ocean, through local changes in surface wind stress and poleward coastally trapped Kelvin waves (Enfield, 1987). Di Lorenzo et al. (2008, 2009) also shows that the North Pacific Gyre Oscillation (NPGO) tracks the dominant interannual and decadal variations of salinity and nutrients in the Northeast Pacific. The ecosystem in the CCS, characterized by a high productivity stimulated by the upwelling of cold and nutrient-rich coastal water, also tends to respond to these dominant modes of climate variability. Indeed, ecosystem fluctuations have already been reported in previous studies and found to be related to large-scale climate variations in the North Pacific such as the PDO (Mantua et al., 1997; Chavez et al., 2003; Lavaniegos and Ohman, 2003; Chhak and Di Lorenzo, 2007), ENSO (Bograd and Lynn, 2001), secular warming (Roemmich and McGowan, 1995; McGowan et al., 2003; Lavaniegos and Ohman, 2007) or the NPGO (Di Lorenzo et al., 2008).

The variability of cross-shore transport of coastal water masses is likely to be critical in understanding ecosystem dynamics

because of the potential for offshore transport of nutrients, mass, and organisms. However, little is known about the dynamics controlling interannual and longer-term variability of cross-shelf transport in the CCS. To examine the temporal variability in cross-shelf transport, we use a long-term hindcast of a regional ocean model coupled with a set of passive tracers continuously released at the coast. In order to separate the offshore advection of surface coastal waters from the offshore advection of upwelled coastal water, tracers are released separately both in the surface layer (“surface-released tracer”) and in the subsurface (“subsurface-released tracer”). We use the passive tracer fields to construct proxies for offshore transport, coastal upwelling strength and Ekman transport efficiency. In addition, we divide our analysis domain into the central and southern CCS to examine the extent of south-north exchange through transport of water masses by the surface and subsurface flow.

This paper is organized as follows. Section 2 describes the model experiments and tracer approach used in this study. Sections 3–5 use the passive tracer statistics to focus on the mean, seasonal cycle, upwelling low-frequency variability, cross-shore transport and south-north exchange through the poleward undercurrent. Finally, Section 6 provides a summary and discussion of the Ekman vs. eddy dynamics.

2. Model and tracer experiment setup

The upwelling variability and offshore transport dynamics are investigated using a three dimensional, free-surface, hydrostatic, eddy-resolving primitive equation ocean model (the Regional Ocean Modeling System; ROMS; Shchepetkin and McWilliams, 2005). ROMS, a descendent of S-Coordinate Rutgers University Model (SCRUM), uses orthogonal curvilinear coordinates in the horizontal and terrain-following coordinates in the vertical. A complete report of the model numerics, open boundary conditions and mixed layer parameterizations can be found in Shchepetkin and McWilliams (2005), Marchesiello et al. (2001) and Large et al. (1994). Among others, Marchesiello et al. (2003) and Di Lorenzo et al. (2005) have already successfully used ROMS in the CCS eastern boundary current system.

Throughout the study, the CCS is defined as our model domain extending westward from the coast of California to 140°W and northward from 27°N to 42°N (Fig. 1b). The initial and monthly-averaged boundary conditions are obtained from an outer ROMS simulation that encompasses the entire North East Pacific (NEP);

Fig. 1b; westward from the coast to 180°W and northward from 25°N to 61°N). The outer experiment, with an average horizontal resolution of 10 km with 30 levels in the vertical, has been forced with 59 years (1950–2008) of monthly-averaged wind stress and surface heat flux forcing from the National Center for Environmental Prediction/National Center for Atmospheric Research reanalysis (NCEP/NCAR; Kalnay et al., 1996; 2.5° resolution) and uses the output of a high-resolution MOM3-based Ocean General Circulation Model (OGCM) code optimized for the Earth Simulator (OFES; Masumoto et al., 2004; Sasaki et al., 2004, 2008) for the initial and monthly-averaged boundary conditions. The bathymetry of the region is a smoothed version of ETOPO5 (5-min gridded elevation data; NGDC, 1988). A 59-year spinup run was first performed. The CCS grid uses the same horizontal and vertical resolution as the Northeast Pacific grid, i.e. an average horizontal resolution of 10 km and 30 levels in the vertical, with enhanced resolution near the surface (specified by the stretching parameters $(\theta_b, \theta_s, -h_c) = (0.4, 5, 41 \text{ m})$). The model experiment for the CCS uses the NCEP/NCAR reanalysis wind stress and uses the monthly climatologies of heat and freshwater fluxes computed from the outer experiment that uses nudging to climatological sea surface temperature (SST) and salinity (SSS). This model configuration with a coarser resolution of 20 km has been used to successfully study the mean, seasonal, interannual and decadal circulations of the California Current as well as low-frequency fluctuations of temperature, salinity and nutrients in the Northeast Pacific (Marchesiello et al., 2003; Di Lorenzo, 2003; Di Lorenzo et al., 2008, 2009; Chhak et al., 2009). By increasing the resolution to 10 km, we better resolve the eddy and upwelling dynamics in this region. Fig. 1 compares the mean Eddy Kinetic Energy (EKE) between the model output and satellite AVISO data for the period 1992–2008. In particular, the California Current System shows high values offshore both in the model and satellite data. However, these values seem to be underestimated by the model simulation, possibly because the 10 km model resolution does not capture all the energetic - submesoscale dynamics (Capet et al., 2008) and the monthly resolution of the wind forcing lacks the eddy energy forced by high-frequency winds. Although a 10 km model resolution forced by the NCEP winds does not allow a complete resolution of the submesoscale details and intensity of coastal upwelling (Pickett and Paduan, 2003; using Coupled Ocean/Atmospheric Mesoscale Prediction System data set), it has been shown that this model configuration does capture the large-scale low-frequency variations of upwelling in the CCS (Di Lorenzo et al., 2008). Based on these

Fig. 1. Mean eddy kinetic energy ($\text{m}^2 \text{s}^{-2}$) computed from sea surface height for (a) AVISO satellite data and (b) ROMS output for the Northeast Pacific (NEP) outer model grid.

previous findings a 10 km resolution seems appropriate to capture the statistics of larger-scale geostrophic eddies and offshore transport, allowing us to avoid long-term integrations of higher resolution (e.g. 5 km) ocean models. It is important and necessary to note that this study disregards higher resolution (<10 km) processes, occurring in the high-frequency domain (this study focuses on the interannual and decadal variability). In particular, we acknowledge that some transport contributions will be omitted such as submesoscale eddy transport (Capet et al., 2008).

The CCS is a highly productive region, resulting from strong coastal upwelling (Barber and Smith, 1981). Following the approach of Combes et al. (2009), who studied the upwelling and cross-shelf transport variability in the Gulf of Alaska circulation, we introduce a passive tracer advection–diffusion equation with a decay term:

$$\frac{\partial P}{\partial t} + \underline{u} \cdot \nabla P = A_H \nabla_H^2 P + \frac{\partial}{\partial Z} \left(A_V \frac{\partial P}{\partial Z} \right) - \frac{P}{\tau} + Q(x, y, z)$$

where P is the passive tracer concentration, $A_H = 20 \text{ m}^2 \text{ s}^{-1}$ is the horizontal diffusivity, A_V is the vertical diffusivity obtained by a KPP scheme (Large et al., 1994), Q a time independent source term and τ is the decay timescale set to 1 year (needed to avoid an infinite growth of passive tracer concentrations in the model domain interior). To characterize the transport and upwelling of subsurface nutrient-rich water, we choose the source term (Q) such that the passive tracer (P) is set to 1 in the coastal region, illustrated by the white rectangles on each figure (from the coast to ~60 km offshore), and in the subsurface from 150 to 250 m depth. To separate the upwelling regions north and south of Pt. Conception (34.45°N), we use two independent tracers, released in the central (“northern tracer”) and southern CCS (“southern tracer”). This configuration allows us to explore and quantify the alongshore exchanges. Since each tracer is continuously released in the subsurface, the concentration of each tracer found at the surface will be considered a measure of upwelling. To illustrate the passive tracer approach, Fig. 2a–d shows the distribution of a passive tracer after 57 years of model

integration (March 2007), which was continuously released in the southern subsurface region (white hatched box in Fig. 2a). The tracer concentration at 170 m (Fig. 2a–b) clearly shows a westward and northward advection of the southern water masses.

Using these tracer maps we diagnose the alongshore and cross-shore exchanges. The vertical section (Fig. 2d) illustrates the importance of mesoscale eddies in the transport of coastal water to the offshore region (high tracer concentration), in this case of CUC subsurface waters. Fig. 2b and c (SSH) also indicates that both cyclonic (“C”; negative SSH spatial anomaly) and anticyclonic (“A”; positive SSH spatial anomaly) eddies entrain and transport coastal water in their core, although we show that most of the tracer is advected offshore by cyclonic rather than anticyclonic eddies. Throughout the text and figures, the significance of correlations is estimated based on the probability density function (PDF) of the cross-correlation coefficients between 2 time series s_1 and s_2 . The PDF is built by computing the correlation of 2500 random pairs of time series that possess the same autocorrelation of s_1 and s_2 .

3. Mean and seasonal cycle

This study aims to quantify the low-frequency dynamics of the cross-shore and alongshore transport in the CCS. As explained in the previous section, the approach to this problem is to use a regional ocean model and follow the fate of a set of passive tracers. The source of the tracers is in the subsurface along the southern and central/northern California coast (hereafter referred to as “southern region” and “northern region”) so that the tracer’s concentration found at the surface corresponds to upwelled coastal water masses.

The first row in Fig. 3 shows the mean tracer concentration at the surface of the subsurface-released tracer in the northern (1st column) and southern (2nd column) regions. Therefore, Fig. 3 shows how much of that tracer has upwelled, averaged over the time of integration. For the northern tracer, we find that, on

Fig. 2. Subsurface-released tracer experiment (southern tracer). Hatched white box in panel (a) indicates the region where the tracer has been released. (a–b) Tracer concentration at 170 m depth for March 2007, (c) sea surface height for the same time. “A” and “C” on panels (b) and (c) highlight anticyclonic and cyclonic circulations, respectively. Arrows correspond to ocean currents at (b) 170 m depth and (c) the sea surface. (d) Vertical tracer section along the blue dashed line on panel (a). The white inshore section in (d) corresponds to the masking of the model grid.

Fig. 3. Subsurface-released tracer experiments. White boxes indicate the regions where the tracer has been released in the subsurface. Mean (a–b) and seasonal cycle anomaly (winter: d–e; summer g–h) of the northern tracer appearing at the surface (first column) and the southern tracer at the surface (second column). Note the different colorbars. (c), (f) and (i) indicate latitudinal variation in the NCEP alongshore wind stress in N m^{-2} .

average, the tracer is advected offshore after it has upwelled at the coast, consistent the annual mean upwelling-favorable alongshore wind stress (Fig. 3c; defined as the alongshore wind stress averaged from the coast to 50 km offshore). In contrast, for the southern tracer release, we observe a more uniform coastal upwelling along the entire coast indicating that, in addition to a southern upwelling, the poleward alongshore transport of southern water masses is an important exchange mechanism between the southern and central CCS. These exchange dynamics will be explored in more depth in a later section to show how the southern subsurface tracer is predominantly advected by the subsurface poleward flow (CUC) towards the central/northern upwelling region where it ultimately reaches the surface.

We now concentrate on the seasonal variability (2nd and 3rd rows in Fig. 3), defined as the anomaly from the mean (1st row in Fig. 3). During the winter season (December–January–February), negative tracer anomalies are found in the northern region, reflecting downwelling conditions, while in the southern region weak upwelling occurs. During summer (June–July–August), the opposite patterns are observed, with a stronger (departure from the mean) upwelling in the northern region (positive tracer anomalies). The intensification of upwelling during summer in the northern region can be tracked both with the tracer injected in the north (which directly upwells at the coast) and also with the tracer injected in the south (advected by the subsurface poleward flow and upwelled in the northern region during summer). This

view of the seasonal upwelling variability, given by the passive tracer approach, is consistent with the seasonal variability anomaly in the alongshore wind stress (3rd column in Fig. 3). The alongshore wind stress anomaly shows a strong seasonal variability, in particular in the northern region, with an upwelling favorable anomaly during summer (negative alongshore wind stress anomaly). The summer anomaly of the southern tracer (Fig. 3h) shows nevertheless a weak upwelling anomaly while the alongshore wind stress anomaly (Fig. 3i) is associated with a downwelling condition. That inconsistency appears from the fact that the stronger upwelling-favorable winds occur in the southern region in spring, in the southern region wind occurs in spring, bringing subsurface tracer anomaly to the surface during late spring. The tracer concentration anomaly observed during summer therefore corresponds to the remaining tracer upwelled a few months prior (in spring).

4. Interannual variability of upwelling and eddy cross-shelf transport

To explore the link between upwelling dynamics and cross-shore transport on the interannual timescale, we remove the climatological monthly means from the tracer fields. Fig. 4c shows the time series of the subsurface-released tracer averaged at the surface over the white box labeled 1 in Fig. 4a and b (above the region where the tracer is released), both for the tracer injected in

Fig. 4. Subsurface-released tracer experiments. All time series are anomalies from the seasonal cycle and normalized. Mean northern (a) and southern (b) tracers at the surface, as in Fig. 3. (c) Time series of tracer anomaly averaged over the section 1 (coast) and at the surface for the northern (blue) and southern (green) regions. Tracer anomaly averaged over section 1 is compared with (d–e) alongshore wind stress anomaly and (f–g) NPGO index. (h–i) show tracer anomaly averaged over each of transects 2–5 (panel a–b), together with correlations with the alongshore wind stress. Significance of correlations is shown in parentheses.

the southern region (green line) and northern region (blue line). These time series (Fig. 4c) illustrate the amount of subsurface water that upwells into each box. Both regions exhibit low-frequency variability, raising the question whether these two

upwelling centers are coherent and whether the response is consistent with Ekman dynamics. Fig. 4c indicates that there is in fact a strong coherence in the upwelling between those two regions with a significant correlation ($R = 0.59$) between the tracer

concentration averaged over the white box of the northern region (blue line) and southern region (green line). This result is consistent with the finding of Lavaniegos and Ohman (2007) that

approximately half of the interannual variability of mesozooplankton is coherent between these two regions. To address the second question concerning whether this upwelling is consistent with

Fig. 5. Surface-released tracer experiments. All time series are anomalies from the seasonal cycle and normalized. (a) Mean northern tracer and (b) mean southern tracer. (c) Times series of tracer averaged over section 2 for the northern (blue) and southern (green) tracers. (d–e) Concentration at the surface of the tracer released at the surface (light blue) compared with the concentration at 170 m depth of the tracer released at the subsurface (dark blue line), sea surface height (green) and alongshore wind stress (red), averaged over section 5. (* indicates a significance >99%).

Ekman dynamics, we compare the tracer time series from Fig. 4c with the time series of alongshore wind stress (averaged over the white box; Fig. 4d–e). We find that the interannual variability in the passive tracer concentration and alongshore wind stress are significantly correlated for the northern region with a significant correlation ($R = 0.45$; $R = 0.78$ using a 1-year lowpass filter on the time series) and a weaker but still significant correlation ($R = 0.31$; $R = 0.47$ using a 1-year lowpass filter on the time series) for the southern region. Previous studies have suggested that upwelling dynamics in the central and southern California are linked to large-scale climate variability, and in particular with the mode of climate variability associated with the NPGO (Di Lorenzo et al., 2008). In our model experiment, we also find significant correlation (0.40 for the northern region and 0.26 for the southern region) between the tracer time series and the NPGO index (Fig. 4f–g). However, neither the PDO mode nor the ENSO mode significantly explains the variability of tracer concentration found at the surface.

Having described how much subsurface coastal water reaches the surface (upwelling), we now examine if upwelling is a good indicator of how much tracer is advected in the cross-shore direction. To address this question, at various transects with increasing

distance from the coast to the offshore (white lines in Fig. 4a–b), we compute the time series of the amount of tracer reaching those transects (Fig. 4h–i). Both in the northern and southern regions, there is a significant correlation between the tracer concentration time series at each transect, suggesting that the upwelled coastal water is advected in the cross-shore direction.

In order to compare these results with the surface transport dynamics, a similar tracer experiment was performed, in which the same tracer was released directly in the surface layer (first 100 m) rather than the subsurface, thus removing the upwelling component. The mean spatial distribution of northern and southern surface-released tracer concentrations (Fig. 5a and b) are similar to the mean subsurface-released tracer concentrations (Fig. 4a and b). The primary difference with the means of the subsurface-released tracer (Fig. 4a and b) is that the surface tracer does not show the strong south-to-north displacement that was apparent in the subsurface tracer entrained in the poleward undercurrent (compare Fig. 4b with Fig. 5b). If we now consider the temporal variability of the northern and southern surface-released tracers at transect 2 (the anomaly over the white box labeled 1 being zero as it corresponds to the tracer released region), we find no correlation between their time series (Fig. 5c; $R = 0.05$), indicating that

Fig. 6. Surface composite maps of vorticity (c–d, s^{-1}), SSH (e–f, (m)), and surface-released tracer (g–h) anomalies during times when vorticity (a–b, s^{-1}), at the location of the white star, is greater than $3 \times 10^{-6} s^{-1}$ (above blue line). (a–b) also compare the time series of vorticity (blue line) and tracer concentration (red line). (i) and (j) show the composite maps of surface-released and subsurface northern tracers across the black dashed line on (g).

Fig. 7. Okubo–Weiss parameter vs. surface-released tracer anomaly at transect 5 at the surface, both in the (a) northern and (b) southern regions. A red (blue) circle corresponds to cyclonic (anticyclonic) circulation. Green percentages correspond to the fraction of circles inside each green box for all tracer anomalies greater than 0.1 or less than -0.1 . Snapshot of (c) tracer, (d) Okubo–Weiss parameter, and (e) vorticity anomalies for March 2007. “C” and “A” indicate cyclonic and anticyclonic circulation, respectively.

the offshore advection of coastal water parcels is no longer coherent between the southern (green line) and northern regions (blue line). This raises the question of what dynamics control surface transport. There is evidence to suggest that mesoscale eddies, principally generated by baroclinic instability of upwelling and along-shore currents (Marchesiello et al., 2003) or by the instability of Rossby waves (LaCasce and Pedlosky, 2004), play an important role in the offshore transport dynamics (Washburn et al., 1993). If we compare the concentration of surface-released tracer (Fig. 5d–e; light blue) and the negative SSHa (Fig. 5d–e; red blue), averaged over transect 5, we find a significant correlation coefficient of 0.24 (northern tracer) and 0.50 (southern tracer). Fig. 5d–e also indicate that the surface offshore transport of surface coastal water (light blue line) is significantly correlated with the subsurface offshore transport of subsurface coastal water (dark blue line).

To further illustrate both surface (light blue) and subsurface (dark blue) transport by mesoscale eddies, Fig. 6 shows a composite analysis of vorticity (Fig. 6c and d), SSH (Fig. 6e and f) and tracer concentration (Fig. 6g and h) anomalies during times when vorticity anomalies (Fig. 6a and b; at white star in Fig. 6c and d) are greater than $3 \times 10^{-6} \text{ s}^{-1}$ (red line; the standard deviation being $2.9 \times 10^{-6} \text{ s}^{-1}$ for the time series in Fig. 6a and $2.7 \times 10^{-6} \text{ s}^{-1}$ for the time series in Fig. 6b). The SSHa and tracer composite maps imply that high vorticity anomalies are mainly associated with cyclonic eddies (vectors correspond to surface current composite, where vectors are only plotted for positive vorticity or negative SSH composites). The vertical sections of tracer composite also reveal that the transport by cyclonic eddies extends deep in the vertical, transporting both surface-released and subsurface-released tracer offshore (Fig. 6i and j). Similar patterns are observed in both the northern (first column in Fig. 6) and southern (second column in Fig. 6) regions. However, as shown in Fig. 2b, particular events (March 2007) illustrate that anticyclonic eddies (“A1”, “A2” and “A3”) also transport coastal water (positive tracer concentration).

The scatter plots in Fig. 7a–b show the relationship between tracer concentration anomalies (from the surface-released tracer) and Okubo–Weiss parameter for each time and point along the transect #5. The Okubo–Weiss parameter (OW) has been widely used to detect eddies in the ocean (Sangra et al., 2009; Isern-Fontanet et al., 2006), defined by:

$$OW = \underbrace{\left[\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right]^2}_{S_n} + \underbrace{\left[\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right]^2}_{S_d} - \underbrace{\left[\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right]^2}_{\omega}$$

where S_n , S_s , ω , u and v are respectively the normal and shear components of the strain, the relative vorticity and the zonal and meridional component of the surface velocity. In previous studies (Chelton et al., 2007) a threshold value of $OW \leq -2 \times 10^{-12} \text{ s}^{-2}$ was taken to define eddies. For our analyses, the threshold value is set to -10^{-11} s^{-2} to be consistent with the square value of the vorticity threshold used in Fig. 6a and b ($3 \times 10^{-6} \text{ s}^{-1}$). The statistics nevertheless do not change significantly using a lower Okubo–Weiss threshold. For $OW < -10^{-11} \text{ s}^{-2}$ and $|Tracer \text{ anomaly}| > 0.1$, Fig. 7a and b also separate cyclonic (red circles; positive vorticity) from anticyclonic (blue circles; negative vorticity) events. It clearly shows that most of the eddies transporting coastal waters (positive tracer concentration) are cyclonic (78% in the northern region and 94% in the southern region). This is consistent with the composite analyses described previously, and also points out that some of the tracer can be carried by anticyclonic eddies (blue circles), as expected from Fig. 2b. This could also explain part of the low (but significant) correlation in Fig. 5d between tracer concentration and SSHa (anticyclonic having a positive SSHa). Note that most of the anticyclonic eddies transport negative tracer anomalies offshore (73% in the northern region and 85% in the southern region), meaning that anticyclonic eddies are less significant than cyclonic eddies in their contribution to the offshore export.

To illustrate the spatial pattern associated with the scatter plots in Fig. 7b, Fig. 7c–e compare southern tracer, Okubo–Weiss parameter and vorticity anomaly fields for the month of March 2007. This comparison highlights a horizontal transport of positive (negative) tracer anomaly by cyclonic (anticyclonic) eddies, where positive (negative) tracer anomaly is associated with a stronger (weaker or zero) transport of coastal water masses.

5. The poleward undercurrent

Although a modeling (ROMS) study conducted by Rivas and Samelson (2011) shows that the poleward undercurrent (CUC) plays a surprisingly small role as a direct source of Oregon upwelling water, Chhak and Di Lorenzo (2007; also using the ROMS ocean model) use adjoint passive tracers to track the origin of upwelled water masses in the CCS and show that while much of the upwelled waters at shallow depth come from offshore regions and from the north, at depths of around 200 m the CUC influences the properties of the upwelled waters with southern originated waters. Similar to Chhak and Di Lorenzo (2007), a fraction of our model tracer released in the subsurface southern region is subsequently found in the northern region. Fig. 8d presents the vertical section of the mean southern tracer at 35°N (white line in Fig. 8c;

no tracer has been released at that latitude). It shows a strong subsurface south–north exchange (positive value) and the signature of the CUC. An Empirical Orthogonal Function (EOF) analysis of the anomalous passive tracer concentration at the surface shows that, while the first mode of variability is mainly associated with the southern California upwelling region (not shown here; explaining 24% of the variance), the second mode of variability (Fig. 8a which explains 16% of the variance) corresponds to upwelling occurring in the central/northern region (the dominant upwelling region). The principal component (grey line in Fig. 8b; PC2), which exhibits a low-frequency variability, is nevertheless uncorrelated ($R = 0.04$) with the alongshore surface wind stress (red line in Fig. 8b) in the northern region. This suggests that the amount of tracer that upwells in the northern region is also controlled by the magnitude of the CUC transport. To be consistent with Fig. 4d, we now use the averaged tracer (blue line in Fig. 8e) over the northern region (dashed black line in Fig. 8c) to define the amount of southern tracer that is upwelled in the northern region (highly correlated with the PC2; $R = 0.76$). While the upwelling of the tracer released in the northern region is strongly modulated by the wind stress (seen previously in Fig. 4d; $R = 0.78$ for 1-year lowpass filtered time series), the upwelling of the tracer released in the southern region (Fig. 8e; blue line), which is first advected by the CUC, is now

Fig. 8. Tracer is released in the subsurface and in the southern region. Time series are anomalies and normalized. 2nd EOF (a) and Principal component (b; grey line) of that tracer. (c–d) mean tracer at the surface and across the white line in (c). (e) compares the tracer at the surface in the northern region (blue) with the alongshore wind stress (red) and with the strength of the poleward flow (green; defined as the tracer averaged over the surface delimited by the white dashed line in (b) – dashed green: defined as the meridional transport averaged over the region where Tracer > 0.3). Orange dashed line represents a linear model of poleward flow (green line) and alongshore wind stress (red line).

Fig. 9. Conceptual summary – Ekman vs. eddy transports.

weakly correlated with the surface wind stress (Fig. 8e; red line). That would also confirm that the strength of CUC plays an important role in the amount of tracer upwelled in the northern region. If we define the strength of the CUC as the tracer averaged in the vertical from the coast to the white dashed line in Fig. 8d, Fig. 8e shows that the tracer upwelled in the northern region (blue line) is indeed strongly modulated by the poleward flow ($R = 0.78$; green line) rather than the variability of the alongshore wind stress ($R = 0.32$; red line). Furthermore, a linear combination (orange line) of the strength of the poleward flow and the strength of the alongshore wind stress explains 71% ($R = 0.84$) of the variability of the upwelled water originating in the southern region (blue line). Therefore, understanding the dynamics and interannual variability of the subsurface poleward flow is essential to better characterize upwelling variability. Note that a lowpass filter was applied to the time series in Fig. 8e for better clarity. However, the same conclusion can be made from the raw data. Indeed, while for the filtered data the correlation coefficients are $R = 0.32$ (tracer vs. minus alongshore wind stress), $R = 0.78$ (tracer vs. poleward flow) and $R = 0.84$ (tracer vs. poleward flow + wind stress), the correlation coefficients become respectively $R = 0.20$, $R = 0.62$ and $R = 0.66$ for the unfiltered data. From a transport point of view, the strength of the CUC defined as the meridional transport (averaged over the region where Tracer > 0.3; green dashed line) exhibits a similar temporal variability as the CUC defined using the tracer concentration (green line) with an annual mean of 1.3 Sv.

The model conditions in March 2007, depicted in Fig. 2, serve to summarize our findings. Since no tracer has been injected north of 34°N, the vertical section of tracer at 35°N (Fig. 2d) shows the signature of the subsurface poleward flow (labeled “III”) at the coast. This water, originating in the CUC, is characterized by high nutrients, high salinity, and low oxygen (Hickey, 1998) and is either (1) upwelled at the coast (labeled “I”) or (2) advected offshore (labeled “II”). Fig. 2b–c again show that this specific coastal water (high positive tracer concentration in Fig. 2d) is transported in the subsurface within the cores of cyclonic eddies suggesting an important impact on the pelagic ecosystem via the transport of nutrients and planktonic organisms. It is therefore critical to understand the dynamics of the cross-shore transport and especially mesoscale physical processes (e.g. thermocline eddy statistics) in order to predict how the ecosystems will respond to

changes in atmospheric and ocean states both at the coast and in the open ocean.

6. Summary and conclusions

In this study, we use the Regional Ocean Modeling System (ROMS), forced by NCEP/NCAR reanalysis wind stress, to simulate the dynamics of the California Current System (CCS) and analyze the transport variability of the system from 1965 to 2008. We have shown that 10 km spatial resolution captures the mean large-scale features of the mesoscale activity in this region and also captures the key characteristics of eastern boundary systems such as the subsurface poleward flow. Nevertheless, it is noteworthy that the 10 km resolution experiment disregards submesoscale dynamics. By comparing five CCS high resolution experiments (12, 6, 3, 1.5, and 0.75 km spatial resolution), Capet et al. (2008) show a submesoscale transition occurring in the eddy variability as the horizontal grid scale is reduced to $O(1)$ km. In particular, they found that although this transition has no significant impact on the dominant mesoscale flow structures and horizontal eddy fluxes, the submesoscale vertical velocity is much stronger than the mesoscale one.

Our numerical simulations also present a method to track nutrient-rich coastal waters as they propagate toward the gyre interior. The vertical and cross-shore transports of coastal water masses have been explored using a passive tracer continuously released at the coast. We find that the cross-shore transport of nearshore waters is linked to both linear (Ekman upwelling) and non-linear (mesoscale eddies) dynamics. A conceptual summary of the model dynamics and findings from the passive tracer approach is presented in Fig. 9. The Ekman upwelling dynamics exert the primary control on the surface cross-shelf transport of water masses that have a subsurface origin along the coast. For example the fate of the offshore surface transport of subsurface nutrient rich water is linked predominantly to changes in upwelling favorable winds (Fig. 4). The eddy field exerts a strong control on the net horizontal cross-shelf surface transport (Fig. 5–7), specifically the cyclonic eddies as evident from the combined statistics of passive tracers and the Okubo-Weiss parameter (Fig. 7). The transport of eddies extends deep in the water column and entrains subsurface coastal

water masses of southern origin advected by the poleward California Undercurrent.

Both upwelling and eddy processes have already been mentioned by Plattner et al. (2005) in the context of nitrogen transport. On interannual time scales, the nitrogen concentration at the surface has been observed to be correlated with the North Pacific Gyre Oscillation (NPGO; Di Lorenzo et al., 2008). In our study, the NPGO has also been found to modulate the upwelling and offshore advection of tracer released in the subsurface. The present modeled passive tracer released in the subsurface can be considered a proxy for transport of quantities such as silicic acid, nitrate, oxygen, or passive planktonic organisms of coastal origins, all of which are important biogeochemical variables that are critical for explaining variability of the CCS ecosystem. Based on these findings, the cross-shore gradients of biological variables observed in previous studies (e.g. Bernal, 1981; McGowan et al., 1996) may be explained by the interplay between Ekman and eddy transport dynamics (Fig. 9), where Ekman dynamics control the gradient of biological variables that rely on subsurface nutrients, and eddy dynamics modulate the gradient of biological quantities such as zooplankton or fish larvae that possess life-cycles that are more sensitive to changes in upper ocean advection.

In addition to a cross-shore exchange, previous studies have also highlighted a northward gradient of physical, chemical and biological properties in the central and southern CCS (Bernal, 1981; McGowan et al., 1996; Venrick, 2002; Ware and Thomson, 2005). Our results (Fig. 8) provide evidence that the central CCS (north of the Southern California Bight) is enriched by subsurface coastal water originating in the southern CCS and transported in the core of the subsurface poleward flow and transported offshore by mesoscale eddies both at the surface and in the subsurface.

Considering carefully the separate and additive contributions of Ekman upwelling and eddy transport dynamics, including the north–south coherence, may improve our understanding of retention/loss of nutrients and organisms in the coastal upwelling region and its consequences for long-term variability of the pelagic ecosystem.

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