Submesoscale dominance of offshore tracer transport from the shoreline across the mid shelf

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ABSTRACT

Transport of shoreline released tracer from the surfzone across the shelf can be affected by a variety of physical processes from wind-driven to submesoscale, with implications for shoreline contaminant dilution and larval dispersion. Here, a high-resolution wave-current coupled model that resolves the surfzone and receives realistic oceanic and atmospheric forcing is employed to simulate dye representing shoreline released untreated wastewater in the San Diego, Tijuana region. Surfzone alongshore tracer transport is primarily driven by obliquely incident wave breaking. On the mid-shelf (MS, 20-30 m depth) dye is near-surface enhanced and alongshore patchy with length-scales from 0.4–4 km. Although alongshelf bathymetric variations are present, bathymetric steering is not the principal driver of offshore dye transport. A subtidal cross-shore dye exchange velocity is estimated at the MS to outer-shelf boundary (≈ 30 m depth), which exhibits episodic offshore pulses. Wind-induced Ekman transport is too weak to explain the exchange velocity as are barotropic tides and semidiurnal baroclinic tides. Baroclinic diurnal tides oscillate the dye cross-shore, but are not linked to net subtidal offshore transport. The exchange velocity is elevated for alongshore dye length-scales < 1 km and stronger alongshore density gradient \( \partial \rho / \partial y \) variability, indicating submesoscale processes drive offshore transport. The region has a mean \( \partial \rho / \partial y \) and northward surface alongshore flow is convergent. Stronger northward flows lead to elevated \( \partial \rho / \partial y \) variability, suggesting deformation frontogenesis. Baroclinic tides and surfzone processes may also provide vorticity and buoyancy gradient seeding for the submesoscale.
1. Introduction

Shoreline released wastewater or runoff enters the surfzone (SZ, region of depth-limited wave breaking) delivering pathogens and contaminants to coastal regions, in turn degrading water quality and threatening the health and sustainability of coastal ecosystems (e.g., Ahn et al. 2005; Steele et al. 2018). For example, in the San Diego Bight, south of the San Diego Bay entrance within the Southern California Bight (SCB), 35 mgd of untreated wastewater is released 10 km south of the US/Mexico border at Pt. Bandera MX (e.g., Orozco-Borbón et al. 2006). Shoreline tracer dilution needed for safe bathing occurs through exchange across the surfzone, inner-shelf, and farther offshore. Similarly, the coastal connectivity of intertidal invertebrates (e.g., Becker et al. 2007; Shanks et al. 2010) also requires cross-shelf exchange.

A fraction of shoreline released runoff or small river input is transported alongshore in the SZ, dependent on the non-dimensional flow rate and wave conditions (Wong et al. 2013; Rodriguez et al. 2018). Breaking of obliquely incident waves vertically mixes tracers (Feddersen 2012; Hally-Rosendahl et al. 2014) and drives SZ alongshore currents (Longuet-Higgins 1970; Feddersen et al. 1998), transporting SZ tracer alongcoast up to 10 km (Grant et al. 2005; Feddersen et al. 2016; Rodriguez et al. 2018). SZ tracers eventually are cross-shore transported (exchanged) to the inner-shelf (IS) due to transient rip currents (Hally-Rosendahl et al. 2014, 2015; Suanda and Feddersen 2015; Hally-Rosendahl and Feddersen 2016) or bathymetric inhomogeneities (Castelle and Coco 2013; Brown et al. 2015). Inner-shelf tracer dilution occurs through transport to the mid-shelf and outer-shelf. A variety of processes across a range of time scales (e.g., winds, tides, internal waves and submesoscale dynamics) can induce cross-shelf tracer transport. Thus, understanding tracer transport pathways and the associated physical processes has implications for any shoreline released tracer.

On subtidal (> 33 h) timescales, winds, waves, bathymetric variability, or regional-scale (10–100 km) alongshore pressure gradients (APG) can drive offshore tracer transport. For an alongshore uniform shelf, alongshore wind driven Ekman layers induce cross-shore tracer transport that decreases towards the shoreline for both stratified and unstratified conditions (e.g., Austin and Lentz 2002; Lentz and Fewings 2012). Cross-shore winds induce weaker cross-shore tracer transport relative to alongshore winds (e.g., Fewings et al. 2008; Horwitz and Lentz 2016; Wu et al. 2018). Bathymetric and shoreline variations such as headlands, capes, and shoals can steer mean
flow at a variety of length-scales \( \text{e.g., Gan and Allen 2002; Pringle 2002; Castelao and Barth 2006; Radermacher et al. 2017} \), inducing cross-shore tracer transport. A barotropic APG driven flow, induces a boundary Ekman transport and a compensating interior cross-shore geostrophic current \( \text{Lentz 2008; Marchesiello and Estrade 2010} \), yielding cross-shore tracer transport. The vertical mismatch between onshore wave-driven Stokes drift and offshore Eulerian current can also induce long-term net exchange \( \text{Lentz et al. 2008; Fewings and Lentz 2011} \).

Semidiurnal and diurnal barotropic (BT, surface) and baroclinic (BC, internal) tides also play an important role in cross-shelf exchange \( \text{e.g., Pineda 1994; Walter et al. 2014; Giddings et al. 2014; Zhang et al. 2015} \). A cross-shore dye transport is induced by nonzero covariance between the cross-shore velocity and dye over a tide cycle, analogous to tidal pumping inducing estuary-ocean exchange \( \text{e.g., Lerczak et al. 2006; Geyer and MacCready 2014} \). On the inner-shelf, BT tides can induce residual flow via tidal rectification \( \text{e.g., Ganju et al. 2011} \), particularly for strong BT tidal currents. Semidiurnal BC tides are ubiquitous in the SCB \( \text{e.g., Lerczak et al. 2003; Buijsman et al. 2012; Kumar et al. 2016; Sinnett et al. 2018} \). Diurnal BC tides are also very common in the SCB, often stronger than semidiurnal BC tides \( \text{Lerczak et al. 2001; Nam and Send 2013; Kumar et al. 2015} \), even though diurnal BC tides are subcritical at SCB latitudes. Within the SCB in 20 m depth, semidiurnal BC tides can drive a net onshore nutrient flux \( \text{Lucas et al. 2011} \). The semidiurnal BC tide can drive an onshore cold-water transport during summertime through tidal pumping \( \text{e.g., Walter and Phelan 2016} \), and are important in cross-shore heat export from the nearshore \( \text{Sinnett and Feddersen 2019} \). The presence of BC tides relative to no tides enhance model simulated horizontal and vertical tracer dispersion in 30–50 m water depth \( \text{Suanda et al. 2018} \) potentially via shear dispersion \( \text{Kunze and Sundermeyer 2015} \). BC tides also induce more mixing, reducing stratification \( \text{Suanda et al. 2017} \), resulting in residual cross-shelf flow. The diurnal internal tide is primarily responsible for the offshore and onshore tidal time scale dye advection for a shoreline released dye on a SCB beach \( \text{Grimes et al. 2019} \). Regardless of mechanism, the cross-shore exchange flow induced by BT or BC tides has magnitude less than the tidal velocity amplitude.

Fronts and filaments can be a primary driver for offshore tracer transport 100s of km from shore \( \text{e.g., Nagai et al. 2015} \). Fronts and filaments are associated with submesoscale flows that have \( O(0.1 – 10) \) km length-scales, and \( O(1) \) or greater Rossby number \( Ro = \frac{\zeta}{f} \) where \( \zeta \) is the vertical relative vorticity and \( f \) is the local planetary vorticity. Submesoscale variability is ubiqui-
tous in coastal drifter observations (e.g., Ohlmann et al. 2017) and high resolution coastal models (e.g., Dauhajre et al. 2017). Generation mechanisms of submesoscale variability such as mixed layer instability (e.g., Boccaletti et al. 2007), turbulent thermal wind balance (e.g., McWilliams et al. 2015), and deformation flow induced frontogenesis (e.g., Hoskins 1982), do not require a coastline or finite-depth. Classic frontogenesis within a non-divergent strain field has an along-front current in thermal wind balance with a cross-front density gradient that can be enhanced by an ageostrophic convergent secondary circulation (e.g., Hoskins 1982; McWilliams et al. 2009). However, bathymetric variations in coastal environments also provide a source of surface deformation flow (e.g., Pringle 2002) that may induce frontogenesis.

As the effects of the Pt. Bandera shoreline released untreated wastewater is largely unknown, we seek to understand the pathways for offshore transport in the San Diego Bight using a realistic model that includes the surfzone. This region has often been studied including San Diego Bay tidal outflow (e.g., Chadwick and Largier 1999), upwelling near Pt. Loma (Roughan et al. 2005), fate of episodic small river plumes (e.g., Warrick et al. 2007), and the evolution of shoreline released tracer (Hally-Rosendahl et al. 2014, 2015; Grimes et al. 2019). HF radars estimated regional currents reveal rich variability on length-scales of \( \approx 10 \) km with \( O(1) \) Rossby number (Kim 2010). However, scales \(< 5 \) km could not be resolved. Using HF radar derived currents, a Lagrangian transport model was used to estimate Pt. Bandera exposure kernels (Kim et al. 2009). However currents within 1 km of shore were unknown, drifters were surface trapped, and the objective mapping smooths the velocity field (Kim et al. 2010). Tracers and drifters in realistic shelf models have been extensively used to study Lagrangian transport and dispersion processes (e.g., Uchiyama et al. 2014; Romero et al. 2013; Giddings et al. 2014; Romero et al. 2016; Dauhajre et al. 2017; Suanda et al. 2017, 2018; Dauhajre and McWilliams 2019), with varying grid resolutions. Shoreline larval dispersion patterns for the San Diego Bight region were simulated using a realistic model with grid resolution of 600 m (Rasmussen et al. 2009), but did not include surfzone effects. Coupled wave and circulation models allow both the wave-driven surfzone and shelf to be resolved (Kumar et al. 2012), which is crucial as shoreline released tracer can be transported alongshore in the surfzone for 10s km (Feddersen et al. 2016; Rodriguez et al. 2018).

Here, shoreline-released tracer is simulated from the SZ to the outer-shelf in the San Diego Bight using a high-resolution wave-current coupled model with realistic atmospheric and oceanic forcing. The specific goal is to elucidate the dominant mechanism(s) responsible for offshore...
tracer transport in the mid to outer shelf region. The analysis focuses on a three month study period from mid-summer to fall characterized by weak wind forcing, prevalent southerly incident surface gravity waves, and active internal waves. The model configuration and analysis procedures are described in section 2. The spatio-temporal tracer variability and transport is presented in section 3. Alongshore transport mechanisms are examined in section 4. Mechanisms for cross-shore transport across a mid- to outer-shelf boundary are diagnosed in section 5. The sources of submesoscale variability are discussed in section 6, and section 7 provides a summary.

2. Method

a. Model configuration

Surfzone and shelf circulation is simulated using the Coupled Ocean-Atmosphere-Wave-Sediment-Transport (COAWST) model system (Warner et al. 2010; Kumar et al. 2012). The model consists of three one-way nested parent runs (from LV1 to LV2 and then LV3, referred to as parent runs hereafter) and one high resolution child run (LV4, see Fig 1a for the four grids) from mid summer to early winter 2015. The three parent runs employ the Regional Ocean Modeling System (ROMS, Shchepetkin and McWilliams 2005), whereas the child run incorporates surface gravity waves by coupling ROMS with the Simulating W Aves Nearshore model (SWAN, Booij et al. 1999). ROMS is a three-dimensional, hydrostatic model using a stretched terrain-following vertical coordinate (Shchepetkin and McWilliams 2005). SWAN is a third-generation, phase-averaged wave model that solves the wave action balance equation (SWAN, Booij et al. 1999). For the four grids (LV1 to LV4), the NOAA/NAM surface flux fields (wind stress, heat and precipitation) are used. Over the 5-month simulation period, the 21 days of NAM data gaps are filled with the Coupled Ocean-Atmosphere Mesoscale System (COAMPS) fields. The vertical viscosity and diffusivity are estimated using a $k - \epsilon$ scheme (Umlauf and Burchard 2003). The bottom stress is determined using a logarithmic bottom drag scheme with a bottom roughness $z_0 = 0.1$ cm, same value as that in Kumar et al. (2015). For the LV4 model, SWAN and ROMS are two-way coupled at 10 min intervals allowing current effects on waves and wave effects on currents through surfzone wave breaking and vortex force. The simulation time period from July 2015 to December 2015 experiences a seasonal transition from predominantly southerly incoming surface waves in summertime
to northerly wave conditions in wintertime.

1) PARENT RUN GRIDS AND SETUP

The model grids of the three parent runs downscale from 2 km horizontal resolution for the SCB (LV1 with $253 \times 390$ horizontal grid cells), to 600 m resolution resolving the southern portion of the SCB (LV2 with $266 \times 398$ grid cells), then to 200 m resolution for the greater San Diego region and adjacent shelf (LV3 with $251 \times 413$ grid cells) (see Fig 1a). All three domains have 40 terrain-following vertical ($\sigma$) levels with enhanced resolution near the surface and bottom (ROMS vertical coordinate parameters: $V_{\text{transform}} = 2$, $V_{\text{stretching}} = 4$, $\theta_s = 8$ and $\theta_b = 3$). Grid bathymetry is derived from the 3 arc-second National Oceanic and Atmospheric Administration/National Geophysical Data Center (NOAA/NGDC) coastal relief dataset.

The outermost LV1 domain inherits the boundary and initial conditions from the California State Estimate (CASE) solution, a regional implementation of the $z$-level, primitive equation MIT general circulation model (MITgcm)(Marshall et al. 1997). CASE solutions assimilate a variety of remote and in situ observations, including satellite altimetry data, satellite measured sea surface temperature, temperature and salinity profiles from Argo and Spray glider, expendable Bathythermograph (XBT) temperature transects, Autonomous Pinniped bathythermograph (APB) temperature profiles, and shipboard CTD profiles (Zaba et al. 2018). Daily averaged CASE solutions are first linearly interpolated from $z$- to $\sigma$- level coordinates, and then horizontally interpolated onto the LV1 model grid and open boundaries. As CASE does not include tidal processes, barotropic tidal elevation and velocities of 10 tidal constituents ($M_2$, $S_2$, $N_2$, $K_2$, $O_1$, $P_1$, $Q_1$, $K_1$, $M_4$ and $M_6$) are prescribed along the LV1 open boundaries with the amplitudes and phases from the ADCIRC tidal database (Westerink et al. 1993). This allows the generation and propagation of internal waves within the model domain (e.g., Kumar et al. 2015; Suanda et al. 2017; Kumar et al. 2019).

The LV1 solutions provide initial and boundary conditions for LV2. Subsequently, the LV2 solutions are used for LV3 and LV4 boundary conditions are inherited from LV3. A combination of Chapman and Flather radiation boundary conditions are used for the sea level and the barotropic (depth-independent) velocity (Flather 1976; Chapman 1985). For the baroclinic (depth dependent) flow and tracers, the Orlanski radiation condition is used together with nudging to constrain the interior solution to the parent results (Marchesiello et al. 2001). In LV1-LV3 domains, the nudging
time scale for outgoing baroclinic flow and tracers along open boundaries is 365 d⁻¹, and the nudging time scale for the incoming baroclinic flow and tracers is 6 h⁻¹. All solutions are saved at 1-hour intervals.

2) LV4 RUN GRID AND SETUP

The LV4 grid (with 486 × 1142 grid cells, area 15×36 km²) spans the outer to inner shelf and surfzone in the southern San Diego Bight (Fig. 1b). The San Diego Bight region includes the San Diego Bay (SDB) to the north. Outside of SDB and southward, the shoreline first curves and then straightens. The shoreline extends and includes the Tijuana River Estuary (TJRE), the US/Mexico border and Punta Bandera (PB) within Mexico. South of the curvature, the bathymetry is largely alongshore uniform, except for a broad shoal seaward of TJRE that extends offshore (Fig. 1b). The LV4 grid cross-shore resolution gradually increases from 110 m at the western open boundary to 8 m along the coastline, while in the alongshore direction it transits from 110 m at the southern and northern open boundaries to 8 m near TJRE. The stretched vertical domain has 15 levels (with \( \theta_s = 4.5 \) and \( \theta_b = 3 \)) and the NOAA 1/3 arc-second coastal digital elevation is used for bathymetry.

The LV4 SWAN component solves the wave action balance equation with 25 frequencies between 0.04 and 0.29 Hz and 42 directional bands spanning from 145° to 355° (wave direction in Nautical convention), covering all potential incidence angles. The shoreline normal direction south of 32.6N is approximately 265°. CDIP wave model frequency-directional wave spectra are used for open boundary conditions (O’Reilly et al. 2016). The wave-breaking parameter \( \gamma = 0.5 \) (ratio of wave height to water depth at which wave breaking occurs) is used, following Kumar et al. (2015). Wind-wave generation is also included using the wind product described in the previous section. Note that because SWAN is a wave-averaged model, the LV4 simulation has bathymetric rip currents but does not have transient rip currents which require a wave-resolving model (Feddersen 2014).

The LV4 ROMS component receives freshwater inputs from PB, TJRE and the Sweetwater river within SDB (see Fig. 1b). The parent grids did not receive freshwater input. At PB, freshwater, representing untreated wastewater, is released onto the beach at a constant discharge rate \( Q_r = 1.53 \ m^3 \ s^{-1} \) (i.e., 35 million gallons per day). TJRE freshwater discharge is given by the in-situ measurements at International Boundary and Water Commission (IBWC) gaging station.
and the discharge primarily occurs during rainfall events (Fig. 2a). Additional coastal runoff emanating from the Sweetwater River within SDB is also incorporated in LV4. The discharge rate from Sweetwater River is approximately estimated by multiplying the in-situ observed flow rate at a nearby river (San Diego river) by the ratio of the drainage area (Archfield and Vogel 2010). As PB and Sweetwater River inflow temperature is unknown, a 30-day low-pass filtered (removing weekly and higher frequency variation that could be strongly variable among sites) in-situ Tijuana River channel (station Oneonta Slough) temperature measurement is applied for all three sources. Passive tracer (dye) of constant concentration (equal to 1) is added to the PB freshwater discharge to represent the release of untreated wastewater. The off-diagonal radiation stress tensor term $S_{xy}$ (Longuet-Higgins 1970; Feddersen et al. 1998) is estimated as,

$$S_{xy} = -\frac{1}{16}\rho_0 g H_s^2 \frac{c_g}{c} \sin(\theta_w) \cos(\theta_w),$$

where $\theta_w$ denotes the mean wave angle relative to the shoreline normal, $H_s$ denotes the significant wave height, $\rho_0 = 1025$ kg m$^{-3}$ denotes a reference density, $c_g$ and $c$ denote the group and phase velocities. A positive $S_{xy}$ corresponds to southerly incident waves and large $S_{xy}$ can drive strong alongshore surfzone currents. Various quantities are subtidally filtered (refered to as subtidal) with the PL64 filter (Limeburner et al. 1985) with 33 h cutoff.

The LV4 coupled model was initialized on 12 July 2015 and integrated to 25 December 2015 and relevant time-series are shown in Figure 2. Modeled barotropic tidal amplitudes (Fig. 2b) and phases of $M_2$, $S_2$ and $K_1$ compare well with in-situ current meter measurements (not shown here). NAM wind data shows a frequently southward directed wind velocity with persistent low speed ($|U_w| < 5$ m/s, see Fig. 2c), and are consistent with nearby buoy winds (not shown). At SB, $H_s$ fluctuates between 0.5–1.5 m. There are multiple periods (yellow shading in Fig. 2e) when the LV4 simulation has southerly incoming waves (i.e., $S_{xy} > 0$). These wave characteristics also compare well with in-situ buoy measurements (not shown). At SB, the subtidal depth-averaged alongshore flow $V_{SB}$ ranges from -0.1 m s$^{-1}$ to 0.3 m s$^{-1}$ and is mostly northward (positive) during the study period (Fig. 2f). The modeled dye concentration and dye transport are analyzed during the analysis period from 22 July to 18 October 2015 (dashed line in Fig 2). This allows a 10-day period of model spin up. After 18 October, occurrences of southerly incident incoming waves end northward transport of PB dye.
b. Analysis methods

To facilitate dye transport analysis, the coastal region is divided into three regions or control volumes representing the nearshore (NS, seaward from shoreline with a cross-shore width of $L_{x}^{(NS)} = 500$ m), inner-shelf (IS, seaward of NS with a width of $L_{x}^{(IS)} = 1.25$ km) and mid-shelf (MS, seaward of IS with a width of $L_{x}^{(MS)} = 2.75$ km) outlined as magenta regions in Figure 1b. The region seaward of MS is referred to as outer-shelf (OS). The NS, IS, MS as a whole are denoted the control volume. Note, the designation NS includes both the SZ and the shallow portion of the inner-shelf. The alongshore length of the control volume is $L_{y} = 18$ km, and the southern boundary is located 4.5 km north of PB allowing dye to adjust upon release. The northern extent of the control volume is set to avoid the rapidly curving isobaths further north. The offshore boundaries of the three regimes approximately reach a water depth of 10 m, 15 m and 25 m, respectively (Fig. 1c). Hereafter, the cross- ($x$) and alongshore ($y$) directions are defined based on the control volume orientation. The coordinate origin is assigned at the southeast corner of the NS regime. Positive $x$ is directed shoreward and positive $y$ is northward directed.

Time-averages (over the analysis period) are denoted with $\langle \cdot \rangle$. Standard deviations are represented by $\text{std}(\cdot)$. Because dye is positive definite and does not have a Gaussian distribution, the time-averaged dye $\langle D \rangle$ is based on a logarithmic average such that $\langle D \rangle = 10^{\langle \log_{10} D \rangle}$ (e.g., Hally-Rosendahl et al. 2014). Temporal mean plus (minus) standard deviation of dye are defined as $\langle (\langle D \rangle +, \langle D \rangle -) = (10^{\langle \log_{10} D \rangle} + \text{std}(\log_{10} D), 10^{\langle \log_{10} D \rangle} - \text{std}(\log_{10} D))$.

Analysis using the control-volumes (NS, IS and MS) will include volume or boundary averaged dye concentration, and boundary cross-shore ($Q_{x}$) and alongshore ($Q_{y}$) dye transport (defined below) estimated at each time step. The volume averaged (3D volume averaged) dye concentration is estimated for each control volume (i.e., $\bar{D}^{(NS)}$ and $\bar{D}^{(MS)}$). The boundary-averaged (alongshore and vertical) dye concentration is estimated along the NS/IS and MS/OS boundaries (e.g., $\bar{D}^{(NS,IS)}$ and $\bar{D}^{(MS,OS)}$). Within a region, the cross-shore and depth-integrated alongshore dye transport ($Q_{y}^{(NS)}$, $Q_{y}^{(MS)}$) at an alongshore boundary $y_0$ is defined as

$$Q_{y}^{(r)}(y_0,t)=\int_{x_1}^{x_2} \int_{-h}^{\eta} v_{L}(x,y_0,z,t) D(x,y_0,z,t) \, dz \, dx \quad (2)$$

where $(r)$ represents NS or MS, $x_1$ and $x_2$ represent the offshore and onshore region locations (e.g., for NS, $(x_1,x_2) = (-L_{x}^{(NS)}, 0)$), and the Lagrangian alongshore velocity $v_{L} = v_{e} + v_{st}$, $v_{e}$ is the alongshore Eulerian velocity and $v_{st}$ is the alongshore Stokes drift. Using $Q_{y}$ at the southern and
northern boundaries, a net alongshore dye transport within a control volume is further estimated as

\[ \Delta Q_y^{(r)}(t) = Q_y^{(r)}(0, t) - Q_y^{(r)}(L_y, t). \]  

(3)

The along-boundary and depth-integrated cross-shore dye transport on the NS/IS \((Q_{x^{(NS,IS)}}^{x})\) and MS/OS \((Q_{x^{(MS,OS)}}^{x})\) boundaries are defined as (for example on the NS/IS boundary)

\[ Q_{x^{(NS,IS)}}^{x}(t) = \int_0^{L_y} \int_{-h}^{\eta} u_L(x^{(NS,IS)}, y, z, t) D(x^{(NS,IS)}, y, z, t) \, dz \, dy, \]  

(4)

where \(x^{(NS,IS)}\) denotes the cross-shore location of the NS/IS boundary, the cross-shore Lagrangian velocity \(u_L = u_e + u_{st}\), \(u_e\) is the cross-shore Eulerian velocity and \(u_{st}\) is the cross-shore Stokes drift.

Using \(\bar{D}\), alongshore and cross-shore dye transport, the alongshore dye transport velocities within NS \((V_{xs^{(NS)}}^{x})\) and MS \((V_{xs^{(MS)}}^{x})\), and the cross-shore dye exchange velocities along the NS/IS \((U_{ex^{(NS,IS)}}^{x})\) and MS/OS \((U_{ex^{(MS,OS)}}^{x})\) boundaries are calculated according to

\[ V_{xs^{(NS)}}^{x} = \frac{Q_y^{(r)}}{\bar{D}^{(r)} L_y^{(r)} \bar{h}^{(r)}} \]  

(5a)

\[ U_{ex^{(r1,r2)}}^{x} = \frac{Q_x^{(r1,r2)}}{\bar{D}^{(r1,r2)} L_y \bar{h}^{(r1,r2)}} \]  

(5b)

where \(\bar{h}^{(r)}\) denotes the mean water depth within NS (5.7 m) and MS (20.0 m) in (5a), and \(\bar{h}^{(r1,r2)}\) denotes the mean water depth along the NS/IS (9.6 m) and MS/OS (25.2 m) boundaries in (5b). To average over the tidal and higher frequency variability, the quantities \(\bar{D}, Q_y, Q_x, V_s,\) and \(U_{ex}\) are subtidally filtered.

3. Results: Spatio-temporal dye variability

a. Example of an offshore dye transport event

Upon Pt. Bandera shoreline release at \(D = 1\), the dye is advected at a range of spatio-temporal scales on the surfzone and shelf. Here, an 18-h realization of dye \(D\), density perturbation \(\sigma_t'\) and currents is presented every 6 h to show the spatio-temporal evolution of an offshore dye transport event (Fig. 3) at the surface (upper panels) and on a cross-shore transect (lower panels). Density perturbation \(\sigma_t'\) has the spatial mean removed and \(\sigma_t = \rho - 1000 \text{ kg m}^{-3}\). At the start of this event, southerly incident waves (see red shading in Fig. 2e) have just arrived driving northward NS dye
transport (Fig. 3a1–4). Low concentration dye \((D < 10^{-4})\) is present on the northern shelf and within SDB (Figs. 3a1–4). A persistent alongshore density pattern with relatively lighter water to the north is present and the surface currents on the shelf are primarily northward (Figs. 3b1–4).

At the first time step (7 Aug 14:00, Fig. 3a1), surface dye is advected northward and enters the overall control volume. Near the southern part of the control volume, a high concentration \((D > 10^{-4})\) and meandering dye patch extends seaward and obliquely crosses the MS/OS boundary with width (where \(D > 10^{-4}\)) of 1.5 km and peak value \(D = 3 \times 10^{-4}\). At the dye offshore leading edge, the offshore current is 0.1 m s\(^{-1}\). The surface density perturbation has additional small scale variability (Fig. 3b1). Along the cross-shore transect (green line in Fig. 3a1), dye is mostly concentrated within upper 5-m layer above the thermocline \(T = 19\degree C\) (Fig. 3c1). The cross-shore current is relatively weak (magnitude < 0.05 m s\(^{-1}\), Fig. 3d1).

Six hours later at the second time step (7 Aug 20:00, Fig. 3a2), the surface dye patch has been advected farther northward and elongated into a filament with MS/OS boundary width of 850 m at high peak concentration \(D = 4.9 \times 10^{-4}\). At the dye offshore leading edge, the offshore current is 0.07 m s\(^{-1}\). The elongation is due to an alongshore current convergence (deformation) that also enhances the alongshore density gradient within the control volume (Fig. 3b2). At the cross-shore transect, the surface layer onshore directed current strengthens (reaching 0.2 m s\(^{-1}\), Fig. 3d2), leading to near-surface onshore dye transport with result that, south of the filament, surface \(D\) is completely contained within the NS (Fig. 3a2). Subsurface, the \(T = 19\degree C\) isotherm and dye layer deepen to \(z = -14\) m within 1 km from shore mostly within the NS (see the vertical dye contour \(D = 10^{-4}\) in Fig. 3c2).

At the third time step (8 Aug 02:00, Fig. 3a3), the dye filament has been advected farther north orienting more cross-shore. Offshore of the control volume, the filament is advected offshore by the surface currents at 0.2 m s\(^{-1}\), but within the control volume, strong surface convergence occurs along a strong density gradient (Fig. 3b3), compressing the dye filament. At the MS/OS boundary, the dye filament has a width of 850 m with max \(D = 1.8 \times 10^{-4}\). South of the filament, surface dye is almost completely contained within the NS (Fig. 3a3). Farther south at the cross-shore transect, the onshore (offshore) directed current within the surface (bottom) layer is well developed (reaching 0.1 m s\(^{-1}\) at surface, Fig. 3d3). Both the \(20\degree C\) isotherm (Fig. 3c3) and 23.6 kg m\(^{-3}\) isopycnals (Fig. 3d3) are tilted downward toward the shoreline, and the near-bed dye layer is advected offshore reaching \(z = -18\) m, but remains within the IS (within 1.75 km of shore).
Six hours later (8 Aug 08:00), offshore of the control volume, the surface dye filament is strongly advected offshore (Fig. 3a4,b4) where strong surface convergence deepens the dye (not shown) and dilutes the surface dye. Within the control volume south of the strong alongshore density gradient, surface dye is advected offshore extending the $D = 10^{-4}$ onto the MS, as surface and bottom cross-shore currents reverse (Fig. 3c4,d4). Sub-surface, the isotherms and isopycnals flatten and the near bottom dye is advected back onshore.

**b. Dye and density statistics**

The spatial variability of temperature $T$, salinity $S$, density anomaly $\sigma_t$ and dye statistics is investigated (Fig. 4) for the realistic oceanic forcing and complex bathymetry. For depth $< 25$ m, the time-mean temperature $\langle T \rangle$ is elevated on the northern shelf and SDB by $\approx 0.5$ °C relative to the southern domain near PB (Fig. 4a1). This alongshore $\langle T \rangle$ signal is also seen in the parent LV3 results (not shown here), indicating that it is not due to the LV4 freshwater discharge. The largest temperature standard deviation $\text{std}(T)$ ($> 1.4$ °C) occurs in the nearshore ($h < 10$ m, Fig. 4b1), due to stronger diurnal warming and cooling in shallow water, without a clear alongshore gradient. Relatively low $\langle S \rangle$ and high $\text{std}(S)$ ($\text{std}(S) > 0.3$) occur near the three freshwater sources PB, TJRE, and within the SDB (Fig. 4a2,b2). The shelf $\langle S \rangle$ also has a north-south gradient, fresher to the north (Fig. 4a2). This results in an alongshore gradient of shelf $\langle \sigma_t \rangle$ (Fig. 4a3) with northern $\langle \sigma_t \rangle$ lower by 0.2 kg m$^{-3}$ especially for $h < 25$ m where $\langle T \rangle$ is elevated. The $\text{std}(\sigma_t)$ ($> 0.3$ kg/m$^3$) is elevated near freshwater sources and in the nearshore ($h < 10$ m), implying combined contributions from $S$ and $T$ variations. Within the overall control volume (shown as magenta line) deeper than 15 m, $\text{std}(\sigma_t)$ is largely alongshore uniform. The surface dye statistics $\langle D \rangle_-$ and $\langle D \rangle_+$ (defined in Section 2b, Fig. 4a4,b4) vary from $10^{-2}$ to $10^{-5}$ and show northward dye transport and dilution away from PB, resembling a diffusive plume, with higher concentrations nearshore, implying net offshore dye transport, particularly $\langle D \rangle_+$ (Fig. 4b4). Within the overall control volume, $\langle D \rangle_+$ is about 100× that of $\langle D \rangle_-$. The surface density gradient magnitude $|\nabla_H \rho|$ is defined as

$$|\nabla_H \rho| = \left[ \left( \frac{\partial \rho}{\partial x} \right)^2 + \left( \frac{\partial \rho}{\partial y} \right)^2 \right]^{1/2}, \quad (6)$$

and large values indicate fronts and filaments. Normalized surface relative vorticity $\text{rms}(\zeta/f) > 1$ indicates high Rossby number and breakdown of geostrophic balance. Both $\text{rms}(|\nabla_H \rho|)$ and
rms($\zeta/f$) are elevated in the NS (Fig. 4-a5,b5), attributed to forcings of surface wave breaking, bathymetric irregularities and surface heat fluxes. Both also indicate the dominance of submesoscale processes in the NS as $\text{rms}(\zeta/f) > 3$ (Fig. 4b5). Both statistics are also elevated near the SDB mouth. Within the overall control volume, the TJRE shoal has elevated $\text{rms}(\nabla H \rho)$ and $\text{rms}(\zeta/f)$. For the rest of the control volume, the distribution of both is largely alongshore uniform. At the MS/OS boundary, $\text{rms}(\nabla H \rho) = 1.3(\pm 0.2) \times 10^{-4}$ kg m$^{-4}$ and $\text{rms}(\zeta/f) = 0.8(\pm 0.1)$ indicating that this region often is in a submesoscale regime.

The vertical structure of the mean and std of density and dye are examined by first horizontally averaging over the NS and MS control volumes and then time-averaging (denoted as $\langle \sigma_t^{xy} \rangle$, Fig. 5). Over the summer and early Fall, both the NS and MS are strongly stratified (Fig. 5a) with mean buoyancy frequency $N \approx 0.02$ s$^{-1}$. The mean vertical density gradient ($\approx 0.4$ kg m$^{-4}$, Fig. 5a) is two orders of magnitude or larger than the $\text{rms}(\nabla H \rho)$ (Fig. 4-a5). Thus, even instantaneous horizontal stratification is much weaker than the vertical. The $\sigma_t^{xy}$ standard deviation is largely vertically uniform and varies $\pm 0.5$ kg m$^{-3}$ largely subtidally and coherent in the vertical (not shown). Associated with the vertical stratification, mean dye NS and MS $\langle D^{xy} \rangle$ is surface intensified and decays exponentially downward (Fig. 5b). Within the NS and MS, the vertical dye decay scale $\lambda$ is quantified by least-squares fit to $\langle \tilde{D}^{xy} \rangle = \langle D_S \rangle \exp(z/\lambda)$ (dashed lines in Fig. 5b), where $\langle D_S \rangle$ represents the region mean surface dye. The best fit $\lambda^{(MS)} = 6.6$ m is similar to and slightly larger than $\lambda^{(NS)} = 5.8$ m, indicating that vertical mixing primarily occurs in the NS with weak vertical mixing as dye is transported to the MS.

The temporal and alongshore variability of surface density perturbation $\sigma_t'$ and dye $D$ on the MS/OS boundary are examined in Fig. 6. Density perturbation $\sigma_t'$ has the along-boundary mean removed at each time-step. The $\sigma_t'$ has a persistent negative south to north gradient (Fig. 6a), consistent with Fig. 4a3. Significant variability is present with times of alongshore uniform $\sigma_t'$ (e.g., 6–10 Oct, Fig. 6a) or essentially a step function in $\sigma_t'$ (9 Aug, near 32.56N). Surface dye on the MS/OS boundary also has very strong variability (Fig. 6b). The northern end almost always has $D > 10^{-5}$, consistent with the mean dye fields (Figs. 4a4 and b4). The temporal $D$ variability is largely diurnal to subtidal. The alongboundary $D$ variability is often patchy with a few km or less length-scales, consistent with the offshore dye ejection events (Fig. 3). Dye can be present along the much of the boundary (26 July), only in a limited region (9 Aug), or with very low dye concentrations ($D < 10^{-5}$, 10 Sept).
The logarithmic scale in Fig. 6b obscures the strong along-boundary dye gradients, and the dye patchiness suggests that the MS/OS boundary dye has relatively short alongshore length-scales. A MS/OS boundary alongshore surface dye length scale ($L_{D}^{(MS, OS)}$) is defined as

$$L_{D}^{(MS, OS)}(t) = \left[ \frac{\int_{0}^{L_{y}} (D(y,t) - \overline{D}_{y}(t))^{2} dy}{\int_{0}^{L_{y}} \left( \frac{\partial D(y,t)}{\partial y} \right)^{2} dy} \right]^{1/2}$$

where surface $D$ is evaluated at the MS/OS boundary ($x^{(MS, OS)}$) and $\overline{D}_{y}(t)$ denotes MS/OS boundary average. The $L_{D}^{(MS, OS)}$ is evaluated at each time step when MS/OS boundary averaged dye is $> 10^{-6}$, and also subtidally filtered (Fig. 6c). The raw $L_{D}^{(MS, OS)}$ varies from 0.3–4 km and the subtidal $L_{D}^{(MS, OS)}$ varies from 0.4 to 3 km, confirming that the MS/OS boundary $D$ is patchy with relatively small-scale variation. This $L_{D}^{(MS, OS)}$ is also consistent with the 0.85–1.5 km dye length-scales observed in the example event (Fig. 3).

c. Control volume dye transport

Overall, seven individual south swell events (i.e., $S_{xy} > 0$, Fig. 2e) occur during the analysis period (yellow shading in Fig. 7 and Fig. 2e). During these events the subtidal NS-averaged dye $\overline{D}^{(NS)}$ has distinct peaks mostly at $> 10^{-3}$ (Fig. 7a) associated with breaking-wave driven northward alongshore dye transport $Q_{y}^{(NS)}$. Such alongshore SZ transport of 3–10 km has been previously observed and modeled for shoreline dye releases (Grant et al. 2005; Hally-Rosendahl et al. 2015; Hally-Rosendahl and Feddersen 2016; Feddersen et al. 2016). The large and positive $\Delta Q_{y}^{(NS)}$ indicates that the dye transport into the NS at the south is not balanced by outward dye transport 18 km to the north (Fig. 7b). Instead, the southern boundary $Q_{y}^{(NS)}$ is largely balanced (89%) by the negative (seaward) cross-shore dye transport to the IS $Q_{x}^{(NS,IS)}$ (Fig. 7c). The $Q_{x}^{(NS,IS)}$ varies largely on a synoptic (3–7 d) time-scale. Both the along- and cross-shore dye transports are almost completely contained (> 97%) in the upper 5-m layer, consistent with the NS mean $\overline{D}^{(xy)}$ vertical distribution (Fig. 5b).

Within the MS region, the average $\overline{D}^{(MS)}$ is about 10% of $\overline{D}^{(NS)}$ (Fig. 7a). The $\overline{D}^{(MS)}$ peaks are qualitatively lagged relative to $\overline{D}^{(NS)}$ peaks, and often the $\overline{D}^{(MS)}$ maxima to minima ratio are much smaller than for the NS. The MS/OS boundary mean $\overline{D}^{(MS,OS)}$ is slightly weaker than and lagged by $\approx 0.5$ d from $\overline{D}^{(MS)}$ (Fig. 7a). The south to north alongshore dye transport difference $\Delta Q_{y}^{(MS)}$ switches sign frequently with magnitude usually $< 1$ dye m$^{-3}$ s$^{-1}$ (Fig. 7b), indicating net
gain and loss from the north and south boundaries. The cross-shore dye transport from IS to MS 
\(Q_x^{(IS,MS)}\) and that from MS to OS \(Q_x^{(MS,OS)}\) are mainly offshore (i.e., negative, Figs. 7c) with 
\(Q_x^{(MS,OS)}\) and \(Q_x^{(IS,MS)}\) peaks corresponding also with a \(\approx 0.5\) d lag as dye is transported across the 
MS region. The \(Q_x^{(MS,OS)}\) peaks are stronger than and more intermittent than the \(Q_x^{(IS,MS)}\) (Figs. 7c).

However, the time-averaged \(Q_x^{(MS,OS)}\) is \(0.88\times\) smaller than \(Q_x^{(IS,MS)}\) indicating that the remaining 
dye is advected out on the alongshore boundaries. The time-average of \(Q_x^{(IS,MS)}\) is \(1.5\times\) that of 
the \(Q_x^{(NS,IS)}\), indicating that alongshore transport into the IS from the southern boundary is also 
important (not shown). Both \(Q_x^{(IS,MS)}\) or \(Q_x^{(MS,OS)}\) transports are mostly contained in upper 5-m 
layer (\(\approx 85\%\), dashed in Figs. 7c), also consistent with the enhanced near-surface dye (Fig. 5b). 
Thus, even though diurnal motions oscillates dye vertically in the NS and IS (Fig. 3), it does not 
result in subtidal IS/MS and MS/OS boundary cross-shore transport at \(z < -5\) m. Next, we explore 
the mechanisms that drive this alongshore and cross-shore dye transports.

4. Alongshore dye transport mechanisms with the NS and MS

The nearshore (NS) region is 500 m wide (Fig. 1b) with a typically 100 m wide surfzone for 
the incident wave heights (Fig. 2d). Surfzone alongshore currents are driven by the breaking of 
obliquely incident waves (Longuet-Higgins 1970; Feddersen et al. 1998) which is the dominant 
alongshore tracer transport mechanism (Grant et al. 2005; Clark et al. 2010; Hally-Rosendahl et al. 
2015; Feddersen et al. 2016). The subtidal alongshore dye transport velocity in the nearshore \(V_x^{(NS)}\) 
varies between \(-0.1\) m s\(^{-1}\) and \(0.2\) m s\(^{-1}\) (Fig. 8) corresponding to 9–17 km per day. Consistent 
with the dominant NS transport occuring in the surfzone, the subtidal \(V_x^{(NS)}\) is largely positive 
(northward directed) during southerly wave events (yellow shading in Fig. 8a) and is highly cor-
related with \(S_{zy}/\rho_0\) with squared correlation of \(r^2 = 0.6 (p < 0.05)\), best fit slope of \(\approx 1\), and 
near-zero intercept. This best fit slope is factor of two consistent with a simple surfzone along-
shore wave forcing and linear bottom friction balance assuming a 100 m surfzone width and linear 
drag coefficient of \(3 \times 10^{-3}\) m s\(^{-1}\) (e.g., Lentz et al. 1999). In contrast, \(V_x^{(NS)}\) was only weakly 
related \((r^2 = 0.13, p > 0.05)\) to the alongshore wind stress. This demonstrates the primary role of 
obliquely incident surface gravity waves in driving alongshore NS dye transport over long (10’s of 
km) distances. Other mechanisms such as wind driven currents in the outer NS, tidal currents, and 
shear dispersion can play a secondary role.
The mid-shelf (MS) region spans from 1.75 km to 4.5 km from the shoreline and has a subtidal alongshore dye transport velocity $V_{*}^{(MS)}$ of similar magnitude to the nearshore $V_{*}^{(NS)}$ (Fig. 8a). In contrast to the nearshore, the mid-shelf subtidal alongshore dye transport velocity $V_{*}^{(MS)}$ is strongly correlated to the subtidal depth-averaged SB alongshelf velocity $V_{SB}$ with $r^2 = 0.82$ ($p < 0.05$) albeit at half the magnitude of $V_{SB}$ (Fig. 8). As SB is located just offshore of the MS/OS boundary, this is attributed to a reduced current (i.e., current shear) onshore of SB. Furthermore, the $V_{SB}$ and NS alongshelf current are also likely linked as the $V_{*}^{(NS)}$ and $V_{SB}$ are also reasonably correlated ($r^2 = 0.33$, $p < 0.05$) consistent with cross-shelf coherent subtidal alongshelf currents (Kumar et al. 2015).

The analysis period is characterized by weak alongshore wind forcing (wind speed < 5 m s$^{-1}$ 95% of the time period, see Fig. 2c). Subtidal alongshore wind stress is uncorrelated ($r^2 = 0.09$, $p > 0.05$) with $V_{SB}$ and has magnitude 4× too weak (using a linear friction of $3 \times 10^{-4}$ m s$^{-1}$ Lentz and Winant 1986) to explain $V_{SB}$, suggesting that other dynamics are driving the alongshore current. Previous SCB studies (e.g., Lentz and Winant 1986; Hickey et al. 2003) have shown that the barotropic APG is a significant driver of alongshelf flow. The subtidal barotropic APG is estimated between two selected sites at depth $h = 15$ m within LV4 domain (S1 and S2 in Fig. 1b). The resulting barotropic APG largely varies between $\pm 2 \times 10^{-6}$ m s$^{-2}$, is mostly northward directed as the alongshore flow and largely varies on fortnightly time-scales (Fig. 8b). The barotropic APG is reasonably correlated with $V_{SB}$ ($r^2 = 0.49$, $p < 0.05$) and has the correct magnitude for a frictionally balanced flow (using 15 m depth and linear friction of $3 \times 10^{-4}$ m s$^{-1}$). Thus the regional alongshore current which drives alongshore dye transport within IS and MS is primarily driven by the barotropic APG.

5. Cross-shore dye transport mechanisms at the MS/OS boundary

Here, we examine separately potential mechanisms for driving the MS/OS boundary subtidal cross-shore dye transport including bathymetric steering, wind-driven Ekman transport, BT and BC tides and submesoscale flows.

a. Bathymetric steering

The LV4 domain has a curved shoreline to the north and a shoal at the TJRE mouth. Both
may steer the northward alongshore current seaward and thus favor offshore dye transport. If the dye transport vectors were aligned with the mean flow or the major axis orientation (MAO) of the depth-averaged velocity variance ellipse, then bathymetric steering is likely the dominant offshore transport mechanism. Here a time-mean bulk local dye transport velocity is estimated at each location according to

\[ U_s = \frac{\int_{-h}^{0} D \cdot u_L dz}{\int_{-h}^{0} D dz} \]  

(8)

where \( u_L = (u_L, v_L) \) is the depth dependent Lagrangian velocity. Note, the \( U_s \) direction is that of the time-mean dye transport. The depth-averaged velocity time-mean and variance ellipses are also estimated at each location. The time-mean velocity \( \langle U \rangle \) and the MAO is mostly northward directed along isobath with decreasing magnitude northward (Fig. 9). Near the TJRE mouth, no indication of offshore steering is seen in \( \langle U \rangle \) and the MAO. At the southern end of the MS/OS boundary \( U_s \) is strongly offshore directed, at an angle of 60° to \( \langle U \rangle \) and the MAO. Farther northward on the MS/OS boundary, this angle decreases to 25° at the 32.56N (TJRE mouth) and decreases further to 0° at the northern end of the MS/OS boundary. This indicates that the offshore dye transport across the entire length of the MS/OS boundary is not primarily driven by bathymetric steering, although it may contribute to the relatively weak dye transport at the northern edge of the MS/OS boundary. We also note that the statistics \( \text{std}(\sigma_t) \), \( \text{rms}(|\nabla H\rho|) \), and \( \text{rms}(\zeta/f) \) are all largely uniform along the MS/OS boundary (Fig. 4), further indicating the shoal at the TJRE mouth and large scale shoreline curvature are not driving enhanced MS/OS boundary variability.

b. Wind-driven Ekman transport

Wind-driven Ekman transport is another possible mechanism for offshore dye transport at the MS/OS boundary. Here, wind-driven Ekman surface velocity at the MS/OS boundary is estimated following Ekman (1905) and compared with the dye exchange velocity \( U_{ex}^{(\text{MS,OS})} \) (Section 2b). If the MS/OS offshore dye transport were due to Ekman transport, the \( U_{ex} \) and surface Ekman velocities would be expected to be of similar magnitude and correlated.

Classically for a uniform shelf with no alongshelf pressure gradients, the shelf velocity has Ekman component (balancing friction) and geostrophic component balancing the cross-shelf pressure gradient required to make the depth-averaged cross-shore velocity zero (Ekman 1905). With a no slip seafloor boundary condition and constant eddy viscosity \( A_v \), Ekman’s analytic solution for the
surface velocity \((U_{ek}, V_{ek})\) is (following Estrade et al. 2008)

\[
U_{ek} + iV_{ek} = (1 - i) \frac{\tau_x + i\tau_y}{\rho_0 \sqrt{2f A_v}} \frac{\sinh(mh)}{\cosh(mh)} - \frac{iV_g}{\cosh(mh)}
\]  

(9)

where \(i = \sqrt{-1}\), \((\tau_x, \tau_y)\) are the (subtidal) wind stress components (estimated from SB), \(m = (1 + i)(f/2A_v)^{1/2}\), and \(V_g\) is the alongshore geostrophic component obtained from the condition that the depth integrated cross-shore transport is zero. Here, a constant \(A_v = 4 \times 10^{-3} \text{ m}^2 \text{s}^{-1}\) is used, based on the depth- and time-averaged rms modeled eddy viscosity at SB, which is also consistent with Suanda et al. (2017).

The MS/OS dye exchange velocity \(U_{ex}^{(MS,OS)}\) is primarily negative (seaward directed) varying between 0 and \(-0.1 \text{ m s}^{-1}\) with time-mean of \(-0.03 \text{ m s}^{-1}\) (Fig. 10a). With the predominant weak southward wind (Fig. 2c), estimated surface \(U_{ek}\) is primarily negative (seaward directed, see Fig. 9b). The \(U_{ek}\) magnitude is around a quarter of \(U_{ex}^{(MS,OS)}\) and their correlation is essentially zero \((r^2 = 0.02, p > 0.05)\). This indicates that the cross-shore Ekman transport can not explain the \(U_{ex}^{(MS,OS)}\). This is further consistent with the relatively short \(L_D^{(MS,OS)}\) (Fig. 6c) and the much larger wind stress length-scales (and thus \(U_{ek}\)).

c. Cross-shore transport due to barotropic and baroclinic tides

A third mechanisms for the cross-shore dye transport across the MS/OS boundary could be barotropic (BT) and baroclinic (BC) tides. In the example offshore dye transport event (Fig. 3) the diurnal BC tide advects surface dye onshore and offshore and vertically within the NS and IS. This structure is similar to the the observed evolution of shoreline released Rhodamine WT dye near 32.59N during the analysis period (Grimes et al. 2019). If BT and BC tidal exchange mechanisms (see Section 1) were the main driver of the cross-shore exchange, then the subtidal exchange velocity \(U_{ex}\) would be less than and correlated with the slowing varying tidal amplitude. Here, in both the semi-diurnal (SD) and diurnal (DU) bands, an alongshore mean (and std) MS/OS tidal surface velocity amplitudes are estimated for BT \((\hat{U}_{SD}, \hat{U}_{DU})\) and BC \((\hat{u}_1^{(SD)}, \hat{u}_1^{(DU)})\) tides in the Appendix. These tidal amplitudes are then compared to the MS/OS boundary exchange velocity \(U_{ex}^{(MS,OS)}\).

Barotropic tidal amplitudes in relation to \(U_{ex}^{(MS,OS)}\) are examined first (Fig. 10c). The alongshore averaged \(\hat{U}_{SD}\) fluctuates fortnightly between 0.01–0.02 m s\(^{-1}\) with very small std along the boundary, indicating that the single BT velocity amplitude is representative. However, \(\hat{U}_{SD}\) is much
weaker than and uncorrelated \((r^2 = 0.04, p > 0.05)\) with \(U_{ex}^{(MS,OS)}\) (Fig. 10a,c). The alongshore mean \(\bar{U}_{DU}\) is generally \(< 0.01 \text{ m s}^{-1}\) and only weakly correlated \((r^2 = 0.23, p < 0.05)\) with \(U_{ex}^{(MS,OS)}\). Thus, the SD and DU BT tides cannot be the main driver of the cross-shore dye transport on the MS/OS boundary.

The surface DU BC tidal amplitude \(\hat{u}^{(1)}_{DU}\) is generally larger than the surface SD amplitude \(\hat{u}^{(1)}_{SD}\) (Fig. 10d), similar to previous observations (e.g., Kim et al. 2011; Johnston and Rudnick 2015) and modeling (e.g., Kumar et al. 2015) in the SCB. The alongshore averaged \(\hat{u}^{(1)}_{SD}\) is much smaller than \(\hat{u}^{(1)}_{DU}\) (reaches \(0.15 \text{ m s}^{-1}\)) to \(U_{ex}^{(MS,OS)}\) (Fig. 10a,d), but is uncorrelated \((r^2 = 0.02, p > 0.05)\). Again, the magnitude and poor correlation indicate that SD and DU BC tides are not the main driver of the cross-shore dye transport on the MS/OS boundary.

**d. Cross-shore Transport due to submesoscale processes**

The simulation snapshots (Fig. 3) show offshore propagating cross-shore elongated dye structures with a width of \(0.8 \text{ km to } 1.5 \text{ km}\). In general, the MS/OS boundary dye is patchy with subtidal dye alongshore length-scales \(L_{D}^{(MS,OS)}\) varying from 0.5 km to 3 km (Fig. 6b,c), consistent with submesoscale processes (e.g., Dauhajre et al. 2017). Alongshore surface density gradients are also consistently present at a variety of scales (Fig. 6a). Here, we examine the role of submesoscale processes in driving cross-shore transport through the relationships between the dye exchange velocity \(U_{ex}\), the surface alongshore dye length-scale \(L_{D}\) and the along MS/OS boundary rms surface alongshore density gradient \(\text{rms}(\partial \rho/\partial y)^{(MS,OS)}\).

The subtidal exchange velocity \(U_{ex}^{(MS,OS)}\) is more negative for smaller \(L_{D}^{(MS,OS)}\) (Fig. 11a) with binned-mean squared correlation of \(r^2 = 0.70 (p < 0.05)\). On average \(U_{ex}^{(MS,OS)}\) is \(2 \times\) larger for \(L_{D}^{(MS,OS)} < 0.9 \text{ km}\) than for \(L_{D}^{(MS,OS)} > 1.2 \text{ km}\). The stronger offshore \(U_{ex}^{(MS,OS)}\) linked to small \(L_{D}^{(MS,OS)}\) suggests that the offshore dye transport is largely due to submesoscale processes. The \(U_{ex}^{(MS,OS)}\) is consistently more negative for larger \(\text{rms}(\partial \rho/\partial y)^{(MS,OS)}\), with binned mean \(r^2 = 0.81, (p < 0.05)\) (Fig. 11b), consistent with enhanced transport perpendicular to strong alongshore density gradients. A cross-boundary rms along-front velocity \(U_t\) is given from a scaled thermal wind relationship as (e.g., McWilliams 2016),

\[
U_t = \left( \frac{gh}{\rho_0 f} \right) \text{rms} \left( \frac{\partial \rho}{\partial y} \right),
\]
with the mean MS/OS boundary depth \( h = 25.2 \) m. The subtidal \( \text{rms}(\partial \rho / \partial y)_{\text{MS,OS}} \) largely varies between 0 and \( 10^{-4} \) kg m\(^{-4}\), corresponding to \( U_t \) between 0–0.3 m s\(^{-1}\) (top axis in Fig. 11b). The best-fit slope between \( U_{\text{ex}}^{\text{MS,OS}} \) and \( U_t \) is 0.2, thus \( U_t \) is large enough to be a driver of exchange. By analogy, this slope of 0.2 is consistent with the ratio of velocity std to friction velocity ratio (\( \approx 0.4 \)) in turbulent wall-boundary layers (e.g., Mellor and Yamada 1982). This strongly indicates that when strong density gradient are oriented alongshore, offshore tracer transport occurs along-front (i.e., cross-shore) at these submesoscale length-scales (Fig. 11a).

**6. Discussion**

**a. Relative role of temperature and salinity in the alongshore density gradient**

These rms alongshore density gradients on the MS/OS boundary have temperature and salinity contributions, \( \alpha_T \text{rms}(\partial T / \partial y) \) and \( \alpha_S \text{rms}(\partial S / \partial y) \), where \( \alpha_T = 1.7 \times 10^{-4} \circ C^{-1} \) and \( \alpha_S = 7.6 \times 10^{-4} \) are the thermal expansion and saline contraction coefficients, respectively. Although the Pt. Bandera (1.5 m\(^3\) s\(^{-1}\)) and TJRE (Fig. 2a) freshwater input rates are typically small, they may play a role in these MS/OS density gradients as may the large-scale north-south time-averaged salinity gradient (Fig. 4a2). On average, the temperature gradient contribution is 2.4× that of salinity gradients. Occasionally (10% of the time, coincident with TJRE freshwater input, Fig. 2a) the salinity and temperature contributions are roughly equal. Thus, salinity plays a lesser, but not insignificant, role in the density fronts associated with the offshore dye transport and \( U_{\text{ex}}^{\text{MS,OS}} \). Given that many shorelines receive freshwater outflow, alongshore density gradients and associated submesoscale flows may also be important at other locations.

**b. Submesoscale mechanism for alongshore density gradient generation**

Several mechanisms have been posited for generation of submesoscale variability in regions far from coastlines such as mixed layer instability (e.g., Boccaletti et al. 2007), turbulent thermal wind balance (e.g., McWilliams et al. 2015), and deformation flow induced frontogenesis (e.g., Hoskins 1982). Although the submesoscale density gradient generation mechanism is not diagnosed in detail here, we examine the role of an alongshore surface deformation flow in enhancing the existing
mean alongshore density gradient potentially leading to the enhanced \( \text{rms}(\partial \rho / \partial y)^{\text{MS,OS}} \).

Density and alongshore current statistics time-averaged and cross-shore averaged within the MS are examined as MS material is transported across the MS/OS boundary. The time-averaged and MS cross-shore averaged surface density, \( \langle \sigma_t \rangle \) has a largely linear background alongshore density gradient with denser water to the south (Fig. 12a), consistent with the mean density distribution (Fig. 4a3). The average alongshore density gradient \( \partial \langle \sigma_t \rangle / \partial y = 8 \times 10^{-6} \text{ kg m}^{-4} \) is a factor 5-10× weaker than the \( \text{rms}(\partial \rho / \partial y)^{\text{MS,OS}} \) (Fig. 11b). During the analysis period, the subtidal depth-averaged alongshore current at SB is northward most of the time \( (V_{SB} > 0, \text{Fig. 2f}) \). The MS cross-shore averaged surface alongshore velocity is time-averaged when \( V_{SB} > 0 \) yielding \( \langle \tau_x \rangle^{(>0)}(y) \), which has a consistent negative alongshore gradient (Fig. 12b), indicating convergence. A linear best-fit yields a \( \partial \langle \tau_x \rangle^{(>0)} / \partial y = 7.5 \times 10^{-6} \text{ s}^{-1} \), an order of magnitude larger than the rms normal strain rate of mesoscale eddies (e.g., Chaigneau et al. 2008). If this northward convergent flow was responsible for strengthening \( \text{rms}(\partial \rho / \partial y)^{\text{MS,OS}} \), then a stronger \( \text{rms}(\partial \rho / \partial y)^{\text{MS,OS}} \) with increasing \( V_{SB} \) would be expected. Indeed, the subtidal \( \text{rms}(\partial \rho / \partial y)^{\text{MS,OS}} \) is consistently enhanced for stronger northward \( V_{SB} \) (Fig. 13) with \( r^2 = 0.5 (p < 0.05) \) and binned-mean \( r^2 = 0.84 (p < 0.05) \). This strongly indicates a linkage between the two processes and supports the concept that an alongshore convergent flow is promoting generation of rms alongshore density gradients.

c. Role of other mechanisms in offshore dye transport

Although submesoscale mechanisms were found to be the principal driver of offshore tracer transport on the MS/OS boundary, the region is complex and other mechanisms may play a secondary role in enhancing transport. For example, DU BC tides oscillated dye onshore and offshore, but did not principally drive subtidal offshore dye transport through a tidal exchange mechanism. However, in 20–50 m water depth, BC tides enhanced 3D and 2D horizontal dispersion relative to simulations without BC tides on the Central CA coast (Suanda et al. 2018). Here, BC tides could be similarly enhancing offshore transport relative to a no BC tide simulation (not performed). The stratification and circulation on subtidal time-scales can be modified by BT (e.g., Ganju et al. 2011) and BC (Suanda et al. 2017) tides. The TJRE shoal and large-scale bathymetry (coastline curvature, SDB entrance, and Pt. Loma) could also have a secondary effect on the MS/OS offshore dye
transport. For example, enhanced $\text{rms}(\zeta/f)$ (well above 1) and $|\nabla H\rho|$ variability at both the SDB entrance and the TJRE shoal (Fig. 4a5,b5) suggests submesoscale dominance, and local vorticity and buoyancy gradient generation. The surfzone also has strongly elevated vorticity and $|\nabla H\rho|$ variability, relative to the MS, even without the elevated vorticity (Johnson and PattiaratচCI 2006; Suanda and Feddersen 2015) and buoyancy gradients (Kumar and Feddersen 2017a,b) induced by transient rip currents. For example, the NS-region averaged $\text{rms}(\partial\rho/\partial x) = 7.6 \times 10^{-4} \text{ kg m}^{-1}$ and $\text{rms}(\partial\rho/\partial y) = 1.2 \times 10^{-3} \text{ kg m}^{-1}$, much larger than the MS/OS boundary $\text{rms}(\partial\rho/\partial y)_{(\text{MS,OS})}$ (Fig. 11b). As the NS and MS regions are material transport linked (as evidenced by dye), offshore transport of NS or TJRE shoal vorticity and $|\nabla H\rho|$ may seed submesoscale variability on the MS, although of course neither is a conserved passive tracer. Although relatively weak, the fresh water sources at TJRE, Pt. Bandera, and within SDB (Section 6a) could be an additional density gradient source. Lastly, the barotropic APG is mostly northward directed (Fig. 8b) which would induce an onshore near-surface geostrophic flow (e.g., Lentz 2008). Thus, this mechanism is largely opposite sign to the offshore dye transport (and $U_{ex}$), and cannot be a driver of near-surface offshore tracer transport. However, a southward directed APG could drive the near-surface offshore tracer transport.

d. Orientation of density gradients

Here, the simulated $\text{rms}$ alongshore density gradients are similar but 2/3 weaker than the $\text{rms}$ cross-shore density gradients from NS to MS regions. These density gradients are calculated via a local coordinate system that is largely oriented with the bathymetry (Fig. 1b). Both $\text{rms}(\partial\rho/\partial y)$ and $\text{rms}(\partial\rho/\partial x)$ are largest in the NS-region and decrease offshore. In each region, $\partial\rho/\partial y$ and $\partial\rho/\partial x$ also are uncorrelated ($r^2 < 0.05$), indicating no strong directional preference for $\nabla H\rho$. In contrast, simulations of the complex region of the Santa Monica Bay and Palos Verdes Shelf found density gradients were strongly oriented cross-shore (oriented with the local bathymetry gradient) in 10–100 m depth (Dauhajre et al. 2017), which may inhibit offshore tracer transport.

Although our and the Dauhajre et al. (2017) simulations have many similarities, they also have significant differences potentially explaining the density gradient (front) orientation difference. First is grid resolution. Whereas the Dauhajre et al. (2017) grid resolution was $(\Delta x, \Delta y) = (75, 75) \text{ m}$, here, the mean horizontal grid resolutions is significantly finer, $(9,24) \text{ m}$ and $(30,24) \text{ m}$
in the NS and MS, respectively. Inner-shelf released particle retention statistics (related to offshore transport) and dispersion depend strongly on grid resolution, even between 100 m and 36 m resolution as submesoscale processes are better resolved (Dauhajre and McWilliams 2019). Second, Dauhajre et al. (2017) did not have a surfzone, which could provide vorticity and density gradient seeding to the IS and MS (Section 6c). Third is the LV4 fresh water sources (Section 6a).

Lastly, here we use a two-equation $k-\epsilon$ model to estimate TKE and dissipation rate which yield vertical diffusivities (e.g., Burchard et al. 2008). This model can represent surface and bottom boundary layer turbulence (e.g., Umlauf and Burchard 2003) as well as the cross-shore dissipating BC tides (Kumar et al. 2016, 2019). In contrast, Dauhajre et al. (2017) obtain diffusivities from the KPP parameterization (Large et al. 1994), a bulk mixing parameterization (see Burchard et al. 2008) designed for large-scale ocean circulation models. With an extension for bottom boundary layers (Durski et al. 2004), idealized models show weak differences between KPP and $k-\epsilon$ for constant wind upwelling and downwelling shelf conditions (Wijesekera et al. 2003). However, models using KPP and $k-\epsilon$ (or another 2-equation model) have not been compared for realistic conditions that includes BT and BC tides, surfzones, and submesoscale shelf processes.

7. Summary

Transport of shoreline released tracer from the surfzone and across the shelf has implications for shoreline contaminant dilution and larval dispersion. Tracer transport can be affected by a variety of physical processes from surfzone currents, wind-driven circulation, baroclinic tides, and submesoscale flows. Here, we investigate the processes transporting a shoreline released dye representing untreated wastewater in the San Diego, Tijuana region that has a curving shoreline, an estuary, a bay, and a headland. A high resolution wave-current coupled model is used that resolves the surfzone and receives realistic air-sea forcing, tides, waves, and offshore boundary conditions inherited from a larger-scale data assimilated model. Model dye is released at the shoreline to represent shoreline untreated wastewater disposal 10 km south of the US/Mexico border and analyzed from summer to mid fall with largely southerly incident waves, strong shelf stratification, and weak wind forcing. Analysis regions are defined as the nearshore (NS, 0–10 m depth) that includes the surfzone, the inner-shelf (IS, 10–20 m depth), and the mid-shelf (20–30 m depth).

On average, dye mass is largely contained in the upper 5-m of the water column from the NS to
the MS. NS alongshore tracer transport is primarily driven by obliquely incident wave breaking. On
the MS, alongshore density gradients are persistent and dye is alongshore patchy with length-scales
from 0.4–4 km. Although alongshelf bathymetric variations are present, the mean dye transport
velocity is not aligned with the mean flow or major axis of the velocity variance ellipse. Thus,
bathymetric steering is not the principal drifter of offshore dye transport. Using the cross-shore dye
transport and the spatial-mean dye at the mid- to outer-shelf boundary, a subtidal cross-shore dye
exchange velocity is estimated, which exhibits episodic offshore directed pulses. Wind-induced
Ekman transport is too weak to explain the exchange velocity, as are the barotropic tides and
semidiurnal baroclinic tides. Baroclinic diurnal tides are relatively strong and cross-shore oscillate
the dye. However, they are not linked to offshore dye transport.

The dye exchange velocity is elevated for smaller (< 1 km) alongshore dye length-scales and
stronger root-mean-square (rms) alongshore density gradients $\partial \rho / \partial y$, indicating submesoscale pro-
cesses are important in driving offshore dye transport. During periods of northward flows, the
surface alongshore current is convergent with a relatively strong mean deformation rate. Stronger
northward flows are linked to elevated rms $\partial \rho / \partial y$, which may be due to deformation frontogenesis,
leading to offshore dye transport. Other mechanisms such a BC tides or surfzone stirring may have
a secondary impact on cross-shore tracer transport by seeding density gradients and vorticity.

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Here, barotropic and surface baroclinic tidal velocities estimation method on the MS/OS boundary is described. The cross-shore velocity is first decomposed into barotropic (BT, depth-averaged) and baroclinic (BC, deviation from BT velocity) components. These two components are then band-pass filtered to obtain the semidiurnal (SD, 16\(^{-1}\) to 10\(^{-1}\) cph) and diurnal (DU, 33\(^{-1}\) to 16\(^{-1}\) cph) band components. Here, the SD-band analysis is described.

First, we calculate the bulk amplitude (envelope) of the BT SD tidal velocity \(U_{SD}(t,y)\) that is assumed narrow-banded with form

\[
U_{SD}(t,y) = \hat{U}_{SD}(\epsilon t, y) \cos(2\pi \omega_{SD} t + \theta_{SD}),
\]

where \(\omega_{SD}\) is the SD frequency, \(\hat{U}_{SD}\) is the SD velocity amplitude that varies on longer time-scale (denoted as \(\epsilon t\)), and \(\theta_{SD}\) is a phase that will only vary slightly over the 18-km long MS/OS boundary. As the BT tidal velocity is narrow-banded, the amplitude \(\hat{U}_{SD}\) is estimated via Hilbert transform at each alongshore location. The alongshore mean and std are presented in Fig. 10c.

BC velocities vary with depth and have much shorter length-scales than the BT tidal velocities, leading to additional analysis. At each location \(y_i\) along the MS/OS boundary, the SD baroclinic cross-shore current \(u_{SD}(y_i, z, t)\) is decomposed into a vertical EOF such that

\[
u_{SD}(y_i, z, t) = \sum_{n=1}^{N} I_{SD}^{(n)}(y_i, t) \Phi_{SD}^{(n)}(y_i, z),
\]

where \(I_{SD}^{(n)}(t)\) and \(\Phi_{SD}^{(n)}(z)\) are the EOF amplitude and vertical structure at \(y_i\), and \(N = 15\) is the total number of vertical levels. At all MS/OS boundary locations, the first \((n = 1)\) EOF accounts for > 75\% of the SD-band variance (> 90\% of the DU-band variance). For both bands, the first vertical EOF \((\Phi_{SD}^{(1)}(z)\) and \(\Phi_{DU}^{(1)}(z)\)) is consistent with a first-mode baroclinic motions with mid-water column sign change and is nearly alongshore uniform on the MS/OS boundary (Fig. A1a). This indicates that the surface BC tidal cross-shore velocities can be reconstructed with a single EOF at all alongshore locations, i.e., for the SD-band surface velocity

\[
u_{SD}^{(1)}(y, \eta, t) = I_{SD}^{(1)}(y, t) \Phi_{SD}^{(1)}(y, \eta).
\]

The alongshore coherent variability of reconstructed surface SD cross-shore velocity \(u_{SD}^{(1)}(y, \eta, t)\)
(and also DU \( u^{(1)}_{DU}(y, \eta, t) \)) is further examined with a Hilbert EOF (CEOF) (e.g., Horel 1984; Merrifield and Guza 1990). A complex time series is generated according to

\[
\tilde{u}^{(1)}_{SD}(y, t) = u^{(1)}_{SD}(y, t) + i\tilde{u}^{(1)}_{SD}(y, t)
\]

where \( \tilde{u}_{SD} \) is the Hilbert transform of \( u_{SD} \) and \( i = \sqrt{-1} \). The variability of \( u^{(1)}_{SD}(y, t) \) is then CEOF decomposed into

\[
u^{(1)}_{SD}(y, t) = \sum_{n=1}^{M} B^{(n)}_{SD}(t) H^{(n)}_{SD}(y)
\]

where \( M \) is the total number of MS/OS boundary grid points, \( H^{(n)}_{SD}(y) \) is the complex eigenvector, and \( B^{(n)}_{SD}(t) \) is the complex amplitude.

The alongshore first CEOF of the SD and DU band explains 67% and 94% of the alongshore variability, respectively. The magnitude of the SD \( |H^{(1)}_{SD}(y)| \) is maximum near the center of the OS/MS boundary, and is reduced about 50% at the northern and southern ends (Fig. A1b). The DU \( |H^{(1)}_{SD}(y)| \) is largely alongshore uniform, except for a weak decrease at the northern end of the boundary (Fig. A1b). The SD along boundary phase is estimated as,

\[
\theta^{(1)}_{SD}(y) = \text{atan} \left( \frac{\mathcal{I}(H^{(1)}_{SD}(y))}{\mathcal{R}(H^{(1)}_{SD}(y))} \right)
\]

where \( \mathcal{I} \) and \( \mathcal{R} \) are the imaginary and real operators, respectively. For the SD band, the phase \( \theta^{(1)}_{SD}(y) \) varies quasi-linearly by \( \pi/3 \) (60°) along the 18-km boundary (Fig. A1c), suggesting southward propagation of the SD BC tide. For the DU band, the phase is near-zero for the southern half of the boundary, and varies \( \pi/4 \) (45°) over the northern half (Fig. A1c), also indicating a southward propagation from the northern end to about 32.54N, consistent with the southward DU-band propagation observed in this region in 12–15 m depth (Grimes et al. 2019). Overall, the phase variations indicate that the alongshore scale of the SD and DU BC tide is substantially larger than the 18 km length of the MS/OS boundary.

Spectra show that the reconstructed \( u^{(1)}_{SD} \) and \( u^{(1)}_{DU} \) are narrow banded (not shown here). The amplitude of the first CEOF reconstructed SD and DU surface velocities are estimated as for the BT tidal velocity at each \( y \) location resulting in surface \( \hat{u}^{(1)}_{SD}(y, \epsilon t) \) and \( \hat{u}^{(1)}_{DU}(y, \epsilon t) \). The alongshore mean and std are shown in Fig. 10d and described in Section 5c.
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**Figure Captions**

**Fig. 1.** (a) Bathymetry (color shading) of LV1 grid with outlines of the LV2 (yellow), LV3 (black) and LV4 (red) grids; (b) LV4 grid bathymetry and the delineation of the nearshore (NS), inner-shelf (IS) and mid-shelf (MS) control volume (magenta lines); (c) NS to outer-shelf (OS) bathymetry that is alongshore averaged following the NS/IS/OS control volume and to 5 km offshore. In panel b, the the red dots denote freshwater sources Punta Bundara (PB), Tijuana River estuary (TJRE) and Sweetwater River within San Diego Bay (SDB). The yellow dots denote the South Bay Ocean outfall (SB) mooring site in 30-m depth and two selected sites (S1, S2) for current dynamics analysis (see text for details) The US/Mexico border is also labeled.

**Fig. 2.** Time series of (a) freshwater discharge rate, $Q_r$, at TJRE, (b) sea surface elevation $\eta$ at SB, (c) wind velocity vectors $U_w$ at SB, (d) significant wave height $H_s$ at SB, (e) subtidally averaged off-diagonal radiation stress tensor $S_{xy}/\rho_0$ (positive is southerly incident waves) at SB, and (f) subtidal depth-averaged alongshore current at SB $V_{SB}$. In panel e, the yellow shading indicates times of southerly incident waves. In all panels, the two vertical blue dashed lines delineate the analysis period and the vertical red shading highlights the time period shown in Figure 3.

**Fig. 3.** Snapshots of an 18-h offshore dye transport event at (left to right) four times at 6 h interval: Surface distribution of (a) dye concentration $D$ (color shading), (b) density perturbation $\sigma'_t$ (color shading) and surface current velocity (vectors). Cross-shore and vertical profile of (c) $D$ (color shading) and isotherms (line contour), (d) density anomaly $\sigma_t$ (color shading and black contour) and cross-shore current velocity (vectors) along a chosen transect. The green line in panels a and b shows the transect location. The thick black contour in panels a and b and the cyan contour in panel c correspond to $D = 10^{-4}$.

**Fig. 4.** Horizontal distribution of (left to right) temporal mean (top) and standard deviation (bottom) of surface $T (^\circ C)$ (a1 and b1), $S$ (a2 and b2) and density anomaly $\sigma_t$ (kg m$^{-3}$) (a3 and b3), surface dye concentrations $\langle D \rangle_-$ and $\langle D \rangle_+$ (a4 and b4), root-mean-square (rms) of horizontal surface density gradient $\nabla_H \rho$ (kg/m$^4$) (a5) and normalized rms surface relative vorticity $\text{rms}(\zeta)/f$ (b5). Black contours denote the isobaths $h = [10 \ 25 \ 45] m$. The magenta line outlines the control volume. The red dot denotes the PB source.

**Fig. 5.** Vertical profile of time mean and standard deviation (errorbar) of (a) density anomaly $\bar{\sigma}_{ti}^{xy}$ and (b) dye concentration $\bar{D}_{xy}$ spatially averaged within NS (black) and MS (thick gray).
visual clarity, in both panels the results for MS are offset by -0.6 m in y axis. The dashed lines in panel b show the nonlinear least square fitting results.

FIG. 6. Hovmöller diagram (time and alongshore distance) of surface (a) density anomaly perturbation (i.e., after removing the alongshore mean density at each time step) and (b) dye along the MS/OS boundary. In panels a and b, the black contour denotes $D = 10^{-4}$. (c) Time series of alongshore dye length scale $L_D$ (7) along MS/OS boundary both hourly (black) and subtidally-filtered (blue). Gaps are when MS/OS boundary-averaged $D < 10^{-6}$. The small magenta rectangle corresponds to the time period shown in Figure 3.

FIG. 7. Time series of (a) dye concentration volumetrically averaged within NS ($\bar{D}^{(NS)}$), MS ($\bar{D}^{(MS)}$) and spatially averaged along the MS/OS boundary ($\bar{D}^{(MS,OS)}$), (b) cross-shore and vertically integrated alongshore dye transport north-south difference for the NS ($\Delta Q_y^{(NS)}$) and the MS ($\Delta Q_y^{(MS)}$). (c) alongshore and vertically integrated cross-shore dye transport along NS/IS boundary ($Q_x^{(NS,IS)}$), IS/MS boundary ($Q_x^{(IS,MS)}$), and MS/OS boundary ($Q_x^{(MS,OS)}$). The cross-shore transports are offset in the vertical. The upper 5-m layer cross-shore transport (denoted with subscript “top”) are indicated with dashed lines. The yellow shading represents times of positive off-diagonal radiation stress $S_{xy}/\rho_0$ at SB (Fig. 2e).

FIG. 8. Time series of subtidal (a) effective alongshore dye transport velocity within NS ($V_{NS}^*$), MS ($V_{MS}^*$), and depth-averaged alongshore current velocity at SB ($V_{SB}$); (b) time series of subtidal barotropic alongshore pressure gradient (normalized by density) between sites S1 and S2 (see Fig. 1b for locations). The yellow shading in panel a represents periods of positive $S_{xy}$.

FIG. 9. Horizontal distribution of bulk dye transport velocity $U_\ast$(red vectors), depth-averaged velocity time-mean ($\langle U \rangle$) and variance ellipse (gray). The magenta box outlines the control volume. Color shading represents the bathymetry.

FIG. 10. Time series of (a) cross-shore dye exchange velocity on the MS/OS ($U_{ex}^{(MS,OS)}$) boundaries, (b) estimated surface Ekman transport velocity $U_{ek}$, (c) barotropic cross-shore velocity amplitude (A1) in the semidiurnal ($\hat{U}_{SD}^{(1)}$) and diurnal ($\hat{U}_{DU}^{(1)}$) band, and (d) surface baroclinic cross-shore velocity amplitude in the semidiurnal ($\hat{U}_{SD}^{(1)}$) and diurnal ($\hat{U}_{DU}^{(1)}$) band. In panels c and d, the shading represents the alongshore std.

FIG. 11. Scatterplot (gray) with binned-means (red) of MS/OS boundary exchange velocity $U_{ex}^{(MS,OS)}$ versus (a) along-boundary dye length-scale $L_D^{(MS,OS)}$ (7), (b) the along-boundary rms alongshore density gradient $\text{rms}(\partial \rho / \partial y)^{(MS,OS)}$. All values subtidally filtered, gray points are shown
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FIG. 12. (a) The temporal mean (line) and std (shading) of surface MS cross-shore averaged density anomaly ($\langle \sigma_t^x \rangle$) (b) the temporal mean and std of surface alongshore subtidal current velocity for when $V_{SB} > 0$ ($\langle \vec{v}^x \rangle^{(>0)}$).

FIG. 13. The MS/OS boundary rms alongshore density gradient $\text{rms}(\partial \rho / \partial y)^{(MS,OS)}$ versus the depth-averaged subtidal alongshore velocity at SB ($V_{SB}$). All values are subtidally filtered, gray points are shown only every 8 h, and the binned-means have 200 h points each. The subtidal (and binned-mean) squared correlations are $r^2 = 0.5 (r^2 = 0.84)$. The horizontal dashed line is the MS background density gradient.

FIG. A1. (a) Vertical profile of the first EOF of the SD and DU band cross-shore current velocity. The alongshore variation on the MS/OS boundary of the (b) amplitude ($|H_{SD}^{(1)}|, |H_{DU}^{(1)}|$) and (c) phase ($\theta_{SD}^{(1)}, \theta_{SD}^{(1)}$) of the first cEOF of $\hat{u}_{SD}^{(1)}$ and $\hat{u}_{DU}^{(1)}$, respectively.
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