

# Evaluation of high-resolution atmospheric and oceanic simulations of the California Current System

Lionel Renault<sup>\*1</sup>, James C. McWilliams<sup>\*</sup>, Faycal Kessouri<sup>\*#</sup>, Alexandre Jousse<sup>\*</sup>,  
Hartmut Frenzel<sup>†</sup>, and Ru Chen<sup>\*</sup>, Curtis Deutsch<sup>†</sup>

<sup>\*</sup> Department of Atmospheric and Oceanic Sciences,  
University of California, Los Angeles, CA 90095-1565

<sup>†</sup> University of Washington, School of Oceanography,  
Box 357940, Seattle, WA 98195-7940

<sup>#</sup> Southern California Coastal Water Research Project,  
3535 Harbor Blvd, Suite 110, Costa Mesa, CA 92626

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## Abstract

This paper is the first of two that present a 16-year hindcast solution from a coupled physical and biogeochemical model of the California Current System (CCS) along the U. S. West Coast and validate the physical solution with respect to mean, seasonal, interannual, and sub-seasonal fields and, to a lesser degree, eddy variability. Its companion paper is Deutsch et al. (2021a). The intent is to construct and demonstrate a modeling tool that will be used for mechanistic explanations, attributive causal assessments, and forecasts of future evolution for circulation and biogeochemistry, with particular attention to the increasing oceanic stratification, deoxygenation, and acidification. A well-resolved mesoscale ( $dx = 4$  km) simulation of the CCS circulation is made with the Regional Oceanic Modeling System over a hindcast period of 16 years from 1995 to 2010. The oceanic solution is forced by a high-resolution ( $dx = 6$  km) regional configuration of the Weather and Research Forecast (WRF) atmospheric model. Both of these high-resolution regional oceanic and atmospheric simulations are forced by lateral open boundary conditions taken from larger-domain, coarser-resolution parent simulations that themselves have boundary conditions from the Mercator and Climate Forecast System reanalyses, respectively. We show good agreement between the simulated atmospheric forcing of the oceanic and satellite measurements for the spatial patterns and temporal variability for the surface fluxes of momentum, heat, and freshwater. The simulated oceanic physical fields are then evaluated with satellite and *in situ* measurements. The simulation reproduces the main structure of the climatological upwelling front and cross-shore isopycnal slopes, the mean current patterns (including the California Undercurrent), and the seasonal, interannual, and subseasonal variability. It also shows agreement between the mesoscale eddy activity and the windwork energy exchange between the ocean and atmosphere modulated by influences of surface current on surface stress. Finally, the impact of using a high frequency wind forcing is assessed for the importance of synoptic wind variability to realistically represent oceanic mesoscale activity and ageostrophic inertial currents.

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<sup>1</sup>Corresponding author: [lrenault@atmos.ucla.edu](mailto:lrenault@atmos.ucla.edu)

## 38 1 Introduction

39 Subtropical eastern boundary upwelling systems like the California Current System (CCS) are  
40 among the biologically most productive coastal environments (Carr and Kearns, 2003), supporting  
41 some of the world’s major fisheries (FAO, 2009). Seasonal upwelling (mainly during spring and  
42 summer) of deep nutrient-rich water maintains high rates of productivity over broad scales (*e.g.*,  
43 Chavez and Messie (2009)). Additionally, coastal currents and oceanic mesoscale variability con-  
44 tribute to cross-shore exchange of heat, salt, and biogeochemical materials between the open and  
45 coastal oceans as well as the surface layer and interior (Hickey, 1998; Capet et al., 2008a; Gruber  
46 et al., 2011; Renault et al., 2012, 2016a).

47 The seasonal upwelling introduces water with low dissolved oxygen and low pH (*i.e.*, a below-  
48 critical carbonate saturation state) into the surface waters (Chan et al., 2008; Feely et al., 2008;  
49 Gruber et al., 2012), making this region more prone to hypoxia and acidification (Feely et al.,  
50 2018; Gruber et al., 2012). Shoaling of deep, low-oxygen and low-pH waters is particularly perti-  
51 nent in the CCS because the eastern Pacific Ocean contains the world’s largest mid-depth Oxygen  
52 Minimum Zone (OMZ). This OMZ has been expanding (*e.g.*, Stramma et al. (2010)), which makes  
53 the coast more susceptible to hypoxic intrusions onto its narrow continental shelf (*e.g.*, Penning-  
54 ton et al. (2006)). The entire CCS is subject to large-scale climate changes (*e.g.*, Stramma et al.  
55 (2012); Bopp et al. (2015); Garcia-Reyes et al. (2015)) that include deoxygenation caused in part  
56 by increased density stratification through anomalous greenhouse heating and the acidification due  
57 to anthropogenic CO<sub>2</sub> invasion. These global influences may be exacerbated in coastal regions by  
58 pollutants deposited from the atmosphere (*e.g.*, nitric and sulfuric acid), from urban wastewater  
59 effluents, and from riverine eutrophication and other contaminants. Multi-decadal declines in oxy-  
60 gen leading to hypoxia have also been observed in the coastal water off southern California and  
61 Oregon and have altered the proportions of biologically important nutrients. Dramatic responses to  
62 these perturbations have already been observed in species that form critical links in the food web  
63 (*e.g.*, pteropods for oceanic acidification (Bednarsek and Ohman, 2015; Bednarsek et al., 2017)  
64 and benthic species and anchovies for hypoxia (Grantham et al., 2004; Feely et al., 2018; Howard  
65 et al., 2020b)).

66 Regional upwelling patterns and eddies are important influences on the ecosystem. Mesoscale  
67 eddies, induced by baroclinic and barotropic instabilities of the wind-driven currents (*e.g.*, March-  
68 esiello et al. (2003)), are present everywhere in the world ocean and play a key role in many oceanic  
69 processes. Many studies have shown their crucial role in the transport of heat and freshwater (*e.g.*,  
70 Wunsch (1999); Dong et al. (2014)) and of biogeochemical materials (McGillicuddy, 2016). In the  
71 open ocean, mesoscale processes can enhance the biological production by increasing the surface  
72 concentration of limiting nutrients (McGillicuddy, 2016). In eastern boundary upwelling systems,  
73 eddies are a limiting factor that reduces the autotrophic primary production by fluxing unconsumed  
74 surface nutrients beneath the euphotic layer (“eddy quenching”) (Gruber et al., 2011; Renault et al.,  
75 2016a). As shown by Renault et al. (2016a,b); Desbiolles et al. (2016), a realistic representation  
76 of the slackening of the wind toward the coast (*i.e.*, wind drop-off) is influential for the mean and  
77 mesoscale currents, the primary production, and interior oxygen levels (Deutsch et al., 2021b).

78 Equilibrium regional oceanic circulation models have been successfully employed for more  
79 than a decade in the CCS. As detailed briefly hereafter, many of the previous modeling efforts  
80 allowed a breakthrough in the understanding and modeling of the CCS. For instance, Marchesiello  
81 et al. (2003) was one of the first realistic mesoscale resolving regional simulation of the CCS; they

82 forced a Regional Oceanic Modeling System (ROMS) simulation using climatological forcing de-  
83 rived from COADS. Veneziani et al. (2009) evaluate favorably with respect to measurements a  
84 high-resolution ROMS oceanic simulation forced by an interannual atmospheric forcing derived  
85 from COAMPS. Based on the same configuration, Neveu et al. (2016) successfully assimilate *in*  
86 *situ* and satellite data and fairly reproduce the CCS circulation and its main characteristics. Seo  
87 et al. (2016) and Renault et al. (2016d) couple a high-resolution oceanic simulation to a high reso-  
88 lution atmospheric simulation for a period of  $\approx 5\%$  years of the CCS. They show the large impact  
89 of air-sea interactions on the mesoscale activity. More recently, Fiechter et al. (2018) uses a  $dx = 3$   
90 km solution coupled with a biogeochemical model forced by the CCMP winds to assess the mod-  
91 ulation of phytoplankton variability by the wind, the oceanic circulation, and topographic effects.  
92 From these studies, some of the oceanic simulations assimilate data, some other are coupled with  
93 the atmosphere or use interannual atmospheric forcing. However, no previous simulation has been  
94 made over a long time period using high-resolution spatial and temporal atmospheric forcing that  
95 includes the effects of wind drop-off (*i.e.*, the cross-shore profile of decreasing wind speed toward  
96 the coast), current feedback on the surface stress (causing a large dampening of the mesoscale  
97 activity; Renault et al. (2016d, 2019a)), and high-frequency wind fluctuations.

98 In this paper and its biogeochemical companion (Deutsch et al., 2021a), ROMS is implemented  
99 over the CCS and is forced by the atmosphere with a regional configuration of the Weather Re-  
100 search Forecast (WRF) model for the period 1995-2010. The main objectives are to characterize  
101 and validate the behavior of the CCS circulation at different time scales with good mesoscale res-  
102 olution in both the ocean ( $dx = 4$  km) and atmosphere ( $dx = 6$  km), while also reviewing the now  
103 substantial literature on this relatively well measured regional system. We also provide an  
104 assessment of the importance of synoptic wind forcing on the mean and mesoscale currents as well  
105 as on the inertial currents. Overall, it provides a more comprehensive validation assessment than  
106 is customary, both to establish the credentials of this particular model for its intended applications  
107 (mostly biogeochemical and ecological, *e.g.*, Deutsch et al. (2021a)) and to provide an example of  
108 the state of the art for realistic regional simulations.

109 In our view, “realistic” model simulations — using forcing and bathymetry fields derived from  
110 measurements and parameterizations for the subgrid-scale effects perceived to be essential — are  
111 coming to play an increasingly central role in oceanic sciences. It is therefore important to develop  
112 a better sense in the community of just how accurate such a virtual reality is, as well as what its  
113 limitations are (*e.g.*, McWilliams, 2007). This is a necessary maturation step for this oceanic  
114 methodology, as it has long since been one for global climate science.

115 The datasets and the model components, setup, and analysis methodology are described in  
116 Sec. 2. In Sec. 3, the behavior of the atmospheric forcing is evaluated with respect to satellite mea-  
117 surements. Section 4 aims to evaluate the oceanic circulation and subsurface layer using satellite  
118 and *in situ* measurements. Finally, in Sec. 5, the oceanic mesoscale activity is evaluated and the  
119 importance of the high frequency atmospheric forcing is assessed. The results are discussed and  
120 summarized in Sec. 6.

## 2 Model Configurations, Analysis Methods, and Data

### 2.1 The Regional Oceanic Modeling System (ROMS)

The oceanic simulations are made with ROMS (Shchepetkin and McWilliams, 2005; Shchepetkin, 2015). As in Renault et al. (2016d), the primary U. S. West Coast (USW4) simulation domain extends from 144.7°W to 112.5°W and from 22.7°N to 51.1°N. Its horizontal grid is 437 x 662 points with a resolution of  $dx = 4$  km, and it has 60 terrain- and surface-following sigma levels in the vertical with stretching parameters  $h_{cline} = 250$  m, and  $\theta_b = 3.0$ , and  $\theta_s = 6$  (Shchepetkin and McWilliams, 2009).

Initial and horizontal boundary data for  $T$ ,  $S$ , surface elevation, and horizontal velocity are taken from the quarter-degree, daily-averaged Mercator Glorys2V3 product (<http://www.myocean.eu>), and applied to the outer boundary of a  $dx = 12$  km solution, which spans a larger domain and serves as a parent grid for the USW4 solution. To improve the water mass representation, in particular the density distribution, the Mercator data are corrected using the mean monthly climatology from the World Ocean Atlas (WOA) (Locarnini et al., 2013; Zweng et al., 2013) over the period 1995-2004. As we shall see in Sec. 4.2, the model does exhibit a mean bias in  $S$  (e.g., the geographical distribution on an interior density surface), and our understanding is that this is mostly inherited from WOA due to the sparsity of interior hydrographic measurements used to determine an accurate mean state around the model boundaries.<sup>2</sup>

The surface turbulent evaporation, heat, and momentum fluxes are estimated using bulk formulae (Large, 2006), and the atmospheric surface fields are derived from an uncoupled WRF simulation (Sec. 2.2), along with the precipitation and downwelling radiation; for these surface fluxes the temporal sampling interval is one hour (1H; see Sec. 5.2 for the sensitivity to this interval). As in Lemarié et al. (2012), the river-runoff forcing dataset we use is a monthly climatology from Dai et al. (2009). River runoff is included offline as surface precipitation and is spread using a Gaussian distribution over the grid cells that fall within the range from the coast to 150 km offshore; this excludes a detailed representation of river plumes.

When forced with bulk formulae, uncoupled oceanic simulations often estimate the surface stress using the absolute wind vector  $U_a$  (e.g., at 10 m height). As shown by e.g., Dewar and Flierl (1987); Duhaut and Straub (2006); Eden and Dietze (2009); Renault et al. (2016d,c); Jullien et al. (2020), such simulations overestimate the mesoscale activity because of their lack of a current feedback. The current feedback is simply the influence of the surface current on the surface stress and low-level wind. In a coupled ocean-atmosphere model, the relative velocity difference between the surface wind and current  $U_r$  is used in the bulk formula,  $\tau = \rho_a C_d |U_r| U_r$  (with  $\rho_a$  the surface air density and  $C_d$  the drag coefficient). Although the 10-m is generally much larger than the surface current (e.g., when  $U_a = 10$  m s<sup>-1</sup> and  $U_o = 1$  m s<sup>-1</sup>,  $U_r = 9$  m s<sup>-1</sup>), at the mesoscale the current feedback induces a sink of energy from the currents to the atmosphere, which causes a large dampening of the mesoscale activity (by  $\approx 40\%$  for the U. S. West Coast; Seo et al. (2016); Renault et al. (2016d)). However, in a forced oceanic model, an opposite bias arises: the mesoscale activity is underestimated because the wind response to the weakened currents through this stress

<sup>2</sup> In retrospect, a better result might have occurred had we done a density-space correction of the boundary mean  $T - S$  values, as is done for the biogeochemical properties (Deutsch et al., 2021a), but this is unlikely to overcome sampling error from data sparseness. A more elaborate procedure would be to adjust the boundary data to reduce interior bias, but this would be a form of data assimilation, which we otherwise have avoided.

160 feedback should partially re-energize the atmosphere, hence also the mesoscale currents. Renault  
161 et al. (2016d, 2020) suggest using a wind- correction approach based on the current-wind coupling  
162 coefficient  $s_w$ , estimated from a coupled simulation as the slope between the mesoscale current  
163 vorticity and the mesoscale surface stress curl. The atmospheric re-energization is then expressed  
164 as

$$U'_a = s_w U_o, \quad (1)$$

165 where  $U_o$  is the surface current,  $U'_a$  is the wind response to  $U_o$ , and  $s_w$  is a statistical regression  
166 coefficient. The surface stress, therefore, is computed using a bulk drag formula,

$$\tau = \rho_a C_D |U_r| U_r, \quad (2)$$

167 with a parameterized relative velocity,  $U_r$ :

$$U_r = U_a + U'_a - U_o = U_a - (1 - s_w)U_o, \quad (3)$$

168 where  $U_a$  is the surface wind from an uncoupled atmospheric product. For the CCS region,  $s_w =$   
169  $0.23 \pm 0.1$  (Renault et al., 2016d, 2019b). For instance, a surface current of  $1 \text{ m s}^{-1}$  is expected  
170 to induce a 10-m wind anomaly of  $0.23 \text{ m s}^{-1}$ . This simple parameterization roughly mimics the  
171 wind response to the current feedback. Although such a parameterization presents some limitations  
172 (*i.e.*, in this study  $s_w$  is constant, so does not take into account the seasonal cycle, the atmospheric  
173 boundary layer dependency, nor the global-scale geographic variation of  $s_w$ ), but it does lead to  
174 approximately the expected dampening and re-energization of the mesoscale currents.

175 The statistically equilibrated solution USW4 is integrated over the period 1995-2010 after a  
176 spin up of 1 year starting from a larger-domain ROMS parent solution with  $dx = 12 \text{ km}$ .

## 177 **2.2 The Weather Research and Forecast Model (WRF)**

178 WRF (version 3.6.1; Skamarock et al. (2008)) is implemented in a configuration with two grids,  
179 similar to Renault et al. (2016b). The WRF domains are slightly larger than the ROMS domains  
180 to avoid the effect of the WRF boundary sponge (4 grid points wide). It has horizontal resolutions  
181 of  $dx = 18 \text{ km}$  and  $6 \text{ km}$ , respectively, using only the latter over the USW4 domain. The model  
182 is initialized with the Climate Forecast System Reanalysis (CFSR with  $dx \approx 40 \text{ km}$  horizontal  
183 resolution; Saha et al. (2010)) from 1 January 1994 and integrated for 17 years with time-dependent  
184 boundary conditions interpolated from the same six-hourly reanalysis. Forty vertical levels are  
185 used, with half of them in the lowest 1.5 km, as in Renault et al. (2016b). The model configuration  
186 is set up with the same parameterizations as in Renault et al. (2016b) except that the WRF Single-  
187 Moment, 6-class microphysics scheme (Hong and Lim, 2006) is modified to take into account the  
188 spatial and seasonal variations of the droplet concentration (Jousse et al., 2016). Its Sea Surface  
189 Temperature (SST) forcing is derived from the Ostia one-day product (Stark et al., 2007) that has  
190 a spatial resolution of  $dx = 5 \text{ km}$ . The inner-nested domain (WRF6) is initialized from the outer  
191 solution (WRF18) on 1 April 1994 and integrated for 17 years. Only the period 1995-2010 is used  
192 in the model evaluations.

## 193 **2.3 Analysis Methods**

194 The numerical outputs for the solutions are daily averages, except when assessing the high fre-  
195 quency forcing importance where hourly averages outputs are saved. The winter, spring, sum-

mer, and fall seasons correspond to the months January-March, April-June, July-September, and  
 196 October-December, respectively. To assess the realism of the oceanic and atmospheric solutions,  
 197 several sub-regions are considered (Fig. 1): see the separate boxes for southern California (South),  
 198 central California (Central), and northern California plus Oregon and Washington (North). Addi-  
 199 tionally, when the the data have a spatial resolution that is high enough to consider a coastal region  
 200 (*i.e.*,  $dx < 1^\circ$ ) and do not have too large a nearshore bias zone (*e.g.*, QuikSCAT products usually  
 201 have a coastal blind zone about 30-50 km wide), both Nearshore and Offshore sub-boxes are also  
 202 considered. The mean  $\overline{(\cdot)}$  is defined with respect to the full time average (1995-2010), and it is  
 203 done separately for each season; the prime  $(\cdot)'$  denotes a deviation from the mean.

204 The oceanic geostrophic surface currents are estimated using daily-averaged sea surface height:  
 205

$$u_{og} = -\frac{g}{f} \frac{\partial h}{\partial y}, \quad (4)$$

206 and

$$v_{og} = \frac{g}{f} \frac{\partial h}{\partial x}, \quad (5)$$

207 where  $u_{og}$  and  $v_{og}$  are zonal and meridional geostrophic currents,  $g$  is gravitational acceleration,  $f$   
 208 is Coriolis frequency, and  $h$  is sea surface height.

209 Following the method described in Renault et al. (2016c), the total wind work is defined as  
 210

$$FK = \frac{1}{\rho_0} (\overline{\tau_x u_o} + \overline{\tau_y v_o}), \quad (6)$$

211 where  $u_o$  and  $v_o$  are the zonal and meridional surface currents,  $\tau_x$  and  $\tau_y$  are the zonal and merid-  
 212 ional surface stresses, and  $\rho_0$  is the mean seawater density. Substituting the decomposition of (4)  
 213 and (5) into (6), the total wind work on the geostrophic and ageostrophic flow are

$$FK_g = \frac{1}{\rho_0} (\overline{\tau_x u_{og}} + \overline{\tau_y v_{og}}), \quad (7)$$

214 and

$$FK_a = \frac{1}{\rho_0} (\overline{\tau_x u_{oa}} + \overline{\tau_y v_{oa}}), \quad (8)$$

215 where  $u_{oa}$  and  $v_{oa}$  are the zonal and meridional components of ageostrophic velocities, respectively.  
 216 The wind work terms  $FK_g$  and  $FK_a$  can be split into their mean ( $F_m K_{mg}$  and  $F_m K_{ma}$ ) and eddy  
 217 parts ( $F_e K_{eg}$  and  $F_e K_{ea}$ ):

- 218 • mean geostrophic wind work,

$$F_m K_{mg} = \frac{1}{\rho_0} (\overline{\tau_x u_{og}} + \overline{\tau_y v_{og}}); \quad (9)$$

- 219 • mean ageostrophic wind work,

$$F_m K_{ma} = \frac{1}{\rho_0} (\overline{\tau_x u_{oa}} + \overline{\tau_y v_{oa}}); \quad (10)$$

- 220 • geostrophic eddy wind work,

$$F_e K_{eg} = \frac{1}{\rho_0} (\overline{\tau'_x u'_{og}} + \overline{\tau'_y v'_{og}}); \quad (11)$$

- 221 • ageostrophic eddy wind work,

$$F_e K_{ea} = \frac{1}{\rho_0} (\overline{\tau'_x u'_{oa}} + \overline{\tau'_y v'_{oa}}). \quad (12)$$

222 As in Stern (1975), Marchesiello et al. (2003), and Renault et al. (2016d), we evaluate the  
223 following relevant eddy-mean energy conversion terms:

- 224 • barotropic (horizontal Reynolds stress) kinetic energy conversion  $K_m K_e$ ,

$$K_m K_e = - \int_z (\overline{u'_o u'_o} \frac{\partial \overline{u_o}}{\partial x} + \overline{u'_o v'_o} \frac{\partial \overline{u_o}}{\partial y} + \overline{u'_o w'} \frac{\partial \overline{u_o}}{\partial z} + \overline{v'_o u'_o} \frac{\partial \overline{v_o}}{\partial x} + \overline{v'_o v'_o} \frac{\partial \overline{v_o}}{\partial y} + \overline{v'_o w'} \frac{\partial \overline{v_o}}{\partial z}) dz \quad (13)$$

225 (where  $w$  is the vertical velocity and  $x$ ,  $y$ , and  $z$  are the zonal, meridional, and vertical  
226 coordinates, respectively) and

- 227 • eddy baroclinic potential-to-kinetic conversion  $P_e K_e$ ,

$$P_e K_e = - \int_z \frac{g}{\rho_0} \overline{\rho' w'} dz. \quad (14)$$

228  $F_m K_{mg}$  represents the transfer of energy from mean surface wind forcing to mean geostrophic  
229 kinetic energy;  $F_m K_{ma}$  represents the transfer of energy from mean surface wind forcing to mean  
230 ageostrophic kinetic energy;  $F_e K_{eg}$  represents the transfer of energy from surface wind forcing  
231 anomalies to geostrophic EKE;  $F_e K_{ea}$  represents the transfer of energy from surface wind forcing  
232 anomalies to ageostrophic EKE;  $K_m K_e$  represents the barotropic conversion from mean kinetic  
233 energy to EKE; and  $P_e K_e$  represents the baroclinic conversion from eddy available potential energy  
234 to EKE. We compute those conversion terms at each model grid point. The wind work is estimated  
235 at the free surface, while the barotropic and baroclinic conversion terms are integrated over the  
236 whole water column.

## 237 2.4 Primary Observational Datasets

238 Satellite and *in situ* measurements are used to evaluate the realism of both the atmospheric and  
239 oceanic simulations. Because of intermittent sampling with different instruments, we do not insist  
240 on exact time correspondences in computing climatological averages. To evaluate the performance  
241 of the atmospheric simulation in terms of cloud cover, we use remote sensing data retrieved from  
242 the Moderate Resolution Imaging Spectrometer level 2 data (MODIS; Platnick et al. (2003)). We  
243 use data from the Terra satellite, which is available twice daily around 10:30 am/pm local time,  
244 beginning in the year 2000. The Forcing for Coordinated Ocean-ice Reference Experiments 2  
245 (CORE; Large and Yeager (2009)) dataset is used to evaluate the surface heat and freshwater fluxes.  
246 It provides monthly surface fluxes at a spatial resolution of  $1^\circ$ . The monthly Global Precipitation

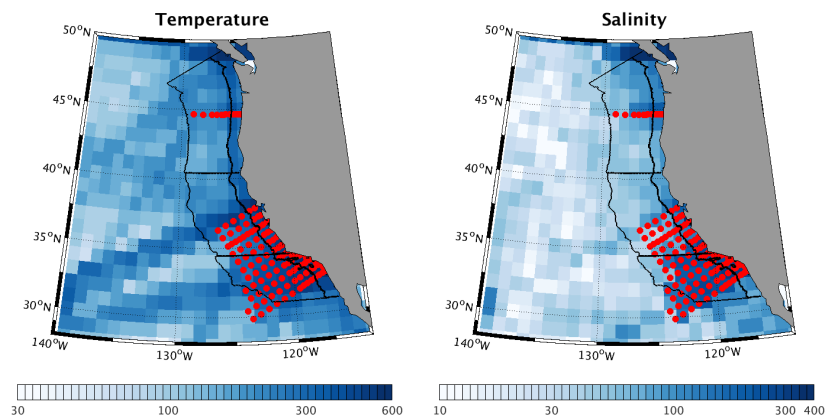


Figure 1: Data density of temperature and salinity measurements in World Ocean Database. The black lines indicate the box areas used to evaluate the simulations with respect to the measurements. Three alongshore domains are assessed: South, Central, and North. When the data have a high-enough spatial resolution to assess a coastal region (*i.e.*, more than two cross-shore valid data), Nearshore and Offshore boxes are also considered with the meridional boundary line, also in black. The red circles indicate the CalCOFI stations and the Newport hydrographic line.

247 Climatology Project (GPCP; Adler et al. (2003)) is also used to evaluate precipitation. It has a  
248 spatial resolution of  $1^\circ$ . The surface stress data is from the Scatterometer Climatology of Ocean  
249 Winds (SCOW; Risien and Chelton (2008)) product based on the QuikSCAT satellite scatterometer.  
250 It provides monthly data at a  $0.25^\circ$  resolution. To compute the wind work, the surface stress daily  
251 product processed by the Centre ERS d'Archivage et de Traitement (CERSAT; Bentamy and Fillon  
252 (2012)) is used. It provides daily surface stress at a spatial resolution of  $dx = 25$  km. SST forcing  
253 is derived from the Ostia daily product (Stark et al., 2007) that has a spatial resolution of  $dx =$   
254  $5$  km. We use data from the California Cooperative Oceanic Fisheries Investigations (CalCOFI)  
255 (*e.g.*, Bograd et al. (2003)). Since 1950 hydrographic stations have been repeatedly but irregularly  
256 sampled on a geographically fixed grid. In this study line 80 (off Pt. Conception;  $34^\circ\text{N}$ ) is used  
257 to estimate a seasonal climatology of temperature, salinity, and density, respectively, to validate  
258 the simulation from 1995 to 2010. The temperature, salinity, and density are further evaluated  
259 using the World Ocean Database 2013 (WOD13, Locarnini et al. (2013); Zweng et al. (2013)). Its  
260 fields have a resolution of  $dx = 25$  km and extend from the surface to the bottom of the ocean. The  
261 WOD13 dataset for that period includes the CalCOFI data and the Newport hydrographic line. The  
262 CSIRO (Commonwealth Scientific and Industrial Research Organization) Atlas of Regional Seas  
263 (CARS) climatology (Ridgway et al., 2002) provides an estimate of the monthly climatology of the  
264 Mixed Layer Depth (MLD) using a temperature threshold of  $\Delta\Theta = 0.2^\circ$  and  $\Delta\sigma_\theta = 0.03$   $\text{kg m}^{-3}$ .  
265 Finally, the CNES-CLS13 dataset (Rio et al., 2014) is used to evaluate the simulated mean sea  
266 surface height and to estimate the geostrophic wind work. It is a combination of GRACE satellite  
267 data, altimetry, and *in situ* measurements with a spatial resolution of  $dx = 25$  km in the analysis  
268 product. The Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVISO)  
269 dataset (Ducet et al., 2000) is used to evaluate the mesoscale activity simulated by USW4 and to



270 estimate geostrophic wind work. It provides the daily sea level anomaly at a resolution of  $dx =$   
271 25 km. Finally, the gliders (line 66.7) of Rudnick et al. (2017) are used to evaluate the geostrophic  
272 current structure.

## 273 **3 Atmospheric Fields**

### 274 **3.1 Shortwave Radiation**

275 Surface net shortwave flux is a key component of the surface energy budget in the CCS. In nu-  
276 merical models, it is strongly related to the representation of clouds and their radiative properties,  
277 which is a common difficulty for both global and regional climate models in eastern boundary up-  
278 welling regions (Nam et al., 2012; Wyant et al., 2010; Zermeño-Díaz et al., 2015). The difficulty  
279 is at least partially attributable to approximate parameterizations of processes governing the stra-  
280 tocumulus clouds that dominate these regions due to the combination of large-scale tropospheric  
281 downwelling, low humidity, and a cold oceanic SST adjacent to a generally warmer continent.  
282 WRF offers choices among various physical parameterizations; here we make our choice in accor-  
283 dance with previous work where an optimized combination of parameterizations was established  
284 in WRF for a stratocumulus region (Jousse et al., 2016). In particular, this combination (Sec. 2.2)  
285 minimize stratocumulus biases, and for the present WRF simulations similar sensitivity tests con-  
286 firm the previous choices. We also perform sensitivity tests for the surface shortwave radiation  
287 scheme (not done in Jousse et al. (2016)). Our results show a better performance of the Goddard  
288 Shortwave scheme (Chou and Suarez, 1994) in comparison to the Dudhia scheme (Dudhia, 1989).  
289 We prescribe the observed spatial variability and seasonality for the cloud droplet concentration  
290 number in the microphysics parameterization scheme WSM6 Jousse et al. (2016). This modeling  
291 strategy minimizes biases in the liquid water path over the northeast Pacific. Figure 2 demon-  
292 strates the plausible WRF results for both spatial variability and seasonal cycle in all the regions  
293 of interests. There is a general increase in incident flux moving equatorward and, in the south,  
294 shoreward, modulated by the cloud cover. These results reflect the realism of the cloud macro-  
295 physical structure (*i.e.*, total water path, TWP) in the simulation (Jousse et al., 2016). Along the  
296 central California coast, both cloud cover and mean shortwave fluxes are biased with respect to the  
297 measurements. While no doubt some of these are due to model errors, near the coast the satellite  
298 measurements have a too coarse spatial resolution to resolve the nearshore variability. There is also  
299 an underestimation of the shortwave flux over the Southern California Bight caused by the over-  
300 estimation of the cloud cover by 5-10 % (not shown). These biases are relatively small compared  
301 to global climate models (more than 30%, see *e.g.*, Fig. 2 of Richter (2015)). The behavior of  
302 the model in reproducing the interannual variability of the shortwave radiation is also revealed in  
303 Fig. 3abc. It depicts the interannual variation of the yearly mean net shortwave radiation averaged  
304 over the South, Central, and North Boxes for CORE and USW4. The simulation fairly reproduces  
305 the interannual variability with *e.g.*, a more intense shortwave radiation in 1997 (+10 W m<sup>-2</sup>) with  
306 respect to the other years.

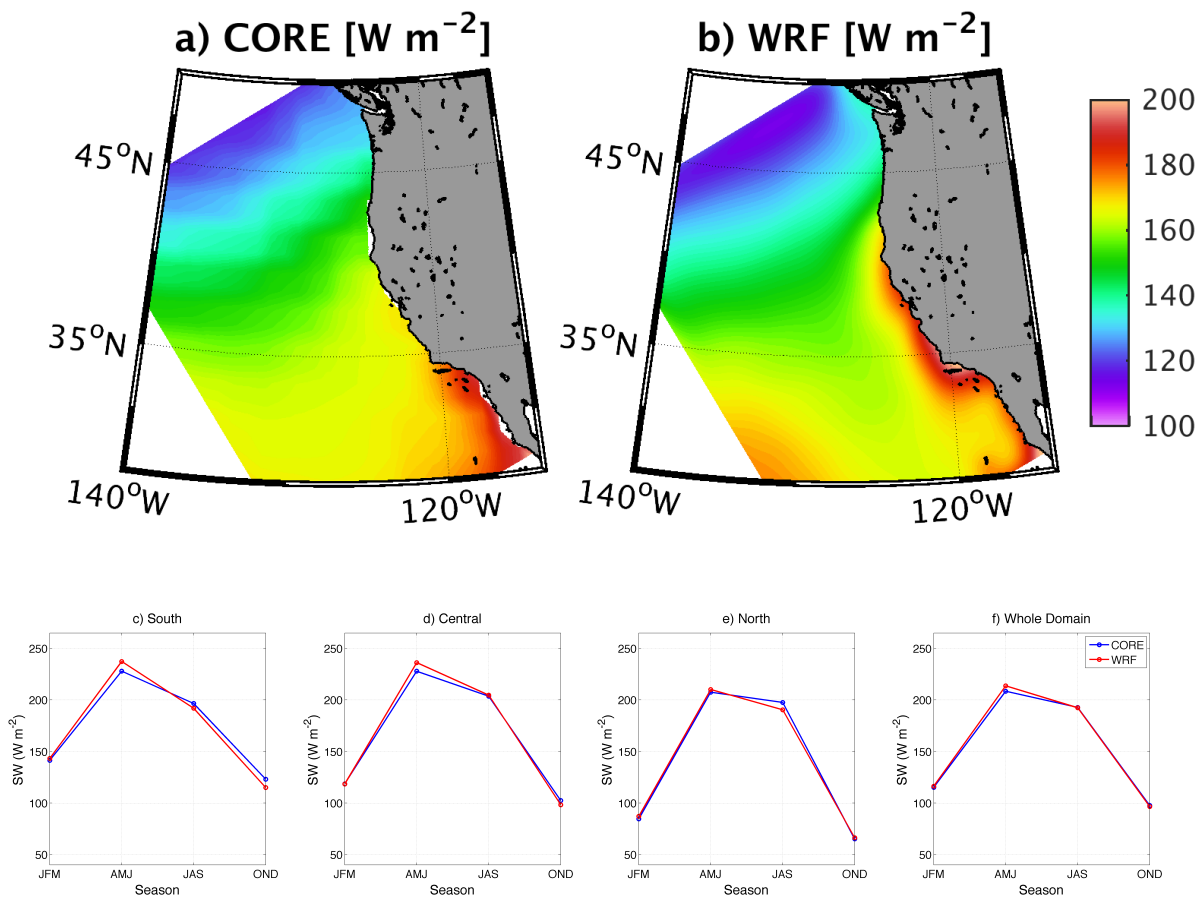


Figure 2: Mean shortwave radiation [ $\text{W m}^{-2}$ ] estimated for the period 1995-2006 from (a) CORE and (b) USW4. Panels (c), (d), (e), and (f) represent the seasonal shortwave radiation variation estimated over the same period from CORE (blue) and WRF (red), averaged over the boxes indicated in Fig. 1 or over the whole domain. The realistic representation of the cloud cover and of the liquid water path in the model allows a good representation of the shortwave radiation.

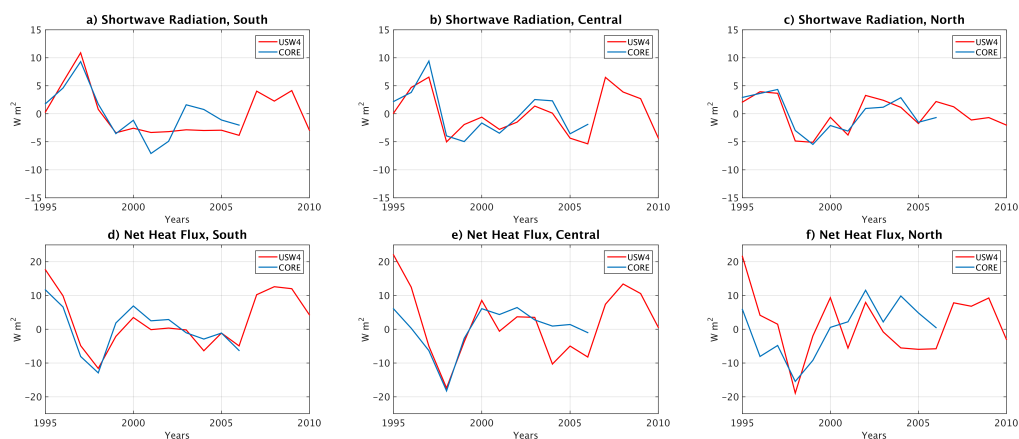


Figure 3: Interannual shortwave radiation (abc) and net radiation (def) [ $\text{W m}^{-2}$ ] over the boxes indicated in Fig. 1 from CORE and USW4.

### 307 3.2 Surface Heat Flux

308 The net surface heat flux is calculated in USW4 using the WRF solar radiation, 10 m wind, 2 m  
309 temperature and humidity fields, and SST. The resulting flux is evaluated with respect to the CORE  
310 dataset. Figure 4 shows the mean net surface heat flux and its seasonal variability in the different  
311 regions (*i.e.*, the alongshore boxes in Fig. 1). Here, we do not consider Offshore and Nearshore  
312 boxes because of the coarse spatial resolution of CORE ( $1^\circ$ ). Despite some bias compensations,  
313 there is an overall good agreement between the measurements and the simulations both in terms  
314 of spatial variability and seasonal cycle. Due to the upwelling and the cold coastal SST in the  
315 CCS, there is generally strong net heating of the ocean near the coast. The maximum of net heat  
316 flux in the Central coast box is similar in both the observations and USW4. Consistent with the  
317 overestimation of the cloud cover and the underestimation of the shortwave radiation, during the  
318 upwelling seasons (spring and summer), the largest bias is again located in the Southern California  
319 Bight, where the net surface heat flux is underestimated by  $10 \text{ W m}^{-2}$ . Along the central California  
320 coast, the cloud cover bias induces a positive bias in shortwave radiation that is mostly compen-  
321 sated for by a negative bias in longwave radiation (not shown). The turbulent (latent and sensible)  
322 heat fluxes exhibit less than a 7% error (*i.e.*, too large a latent heat flux), which is within the error  
323 range for the measurements (Large and Yeager, 2009). Finally, near the coast in a band  $\approx 30 \text{ km}$   
324 wide, the net heat flux is higher than in the measurements, which probably reflects the coarser  
325 spatial resolution of CORE. The realistic representation of the net heat flux leads to a reasonably  
326 good estimate of the spatial and the seasonal variation of the SST (Sec. 4.1); however, as shown  
327 in the companion paper (Deutsch et al., 2021a), the shortwave flux bias along the California coast  
328 can induce a positive bias in chlorophyll through photosynthesis. Figure 3def depicts the interan-  
329 nual yearly variation of the net heat fluxes over the South, Central, and North boxes for CORE  
330 and USW4. Again, the interannual variability is fairly reproduced by the simulation. Interestingly,  
331 the peak of shortwave radiation in 1997 is compensated by more intense turbulent heat fluxes (not  
332 shown).

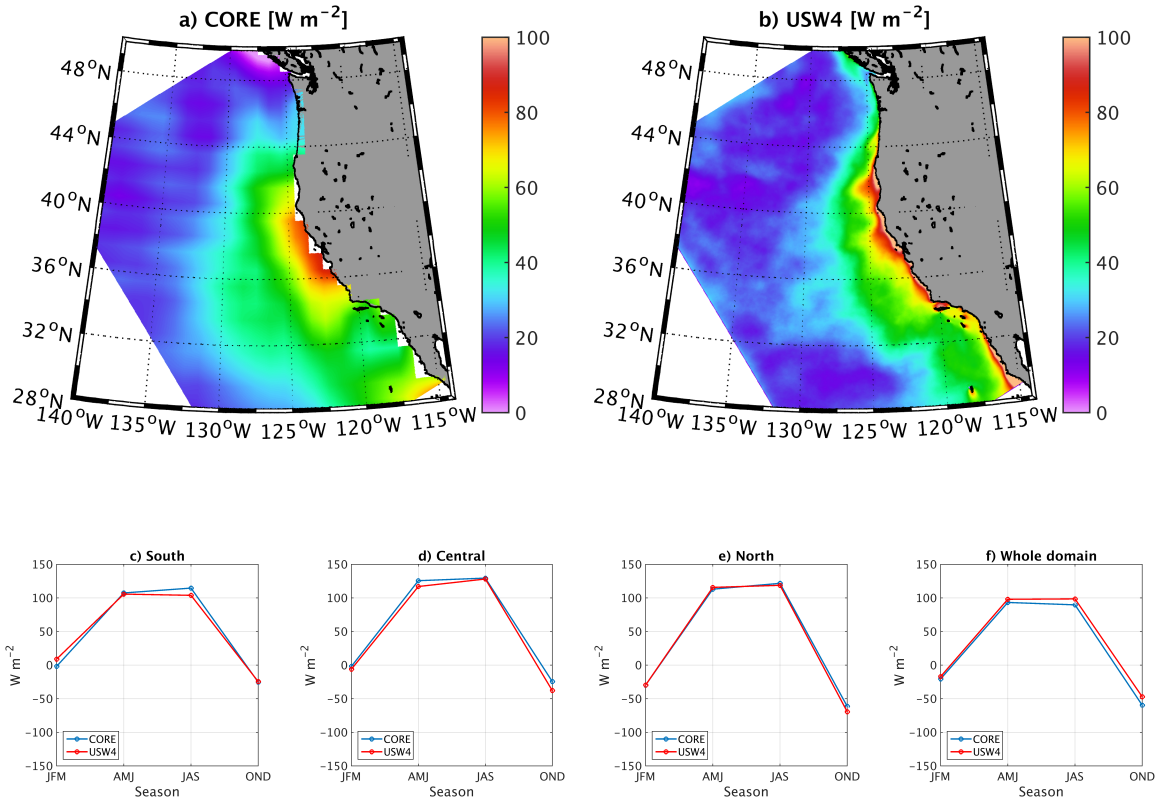


Figure 4: Mean net heat fluxes [ $\text{W m}^{-2}$ ] from (a) CORE and (b) USW4 over the period 1995-2006. Panels (c), (d), (e), and (f) represent the seasonal evolution of the net heat fluxes over the same period from CORE (blue) and USW4 (red), averaged over the boxes indicated in Fig. 1 or over the whole domain.

### 3.3 Surface Freshwater Flux

The net freshwater flux (evaporation minus precipitation) is computed by combining the precipitation from WRF and the evaporation estimated by bulk formulae from WRF's surface fields (Large, 2006). Evaporation dominates toward the south over the warmer subtropical gyre, and precipitation dominates to the north, especially close to the coast; *i.e.*, there is more precipitation in the north during the winter months due to the storm tracks. WRF generally overestimates GPCP by about 0.5 to 1 mm day<sup>-1</sup> (not shown). These differences are certainly not negligible values for the water budget. Nevertheless, they remain within the observational uncertainty range provided by the GPCP data. Moreover, due to their lack of sensitivity to drizzle, remote sensors are known to generally underestimate the precipitation produced by low clouds (Rapp et al., 2013). Because the CCS is substantially covered with low clouds, this may explain some of the discrepancies between WRF and GPCP (*n.b.*, similar results are found using CORE). In USW4, the overestimation of the precipitation is compensated by a slight excess of evaporation (consistent with the latent heat flux bias), which leads to a realistic agreement of the net freshwater flux between USW4 and CORE (Fig. 5). In both USW4 and the observations, the net freshwater flux does not have a strong inter-annual variability as shown in Fig. 6abc except over the South Box, where it reaches large values in 2009 and 2010.

### 3.4 Surface Stress

The surface stress is calculated in USW4 with the Large (2006) bulk formulae as described in Sec. 2.1; the 10 m wind, and the 2 m temperature and humidity. Renault et al. (2016b) show a good agreement between the 10 m wind and satellite and *in situ* observations with a similar model configuration. Here, we evaluate the simulated surface stress with respect to SCOW (Fig. 7). In both USW4 and SCOW, over the CCS the surface stress is mostly equatorward due to the offshore position of the atmospheric subtropical high, and this is the primary cause of offshore Ekman transport and coastal upwelling. This pattern is persistent in the South and Central boxes, with peak stresses near the coast in spring and summer. In the north, the alongshore wind stress direction reverses seasonally. The simulated surface stress is similar to that in the observations in both amplitude and direction. The seasonality and the main gradients are also realistic (Fig. 7). We do not consider separate Nearshore and Offshore boxes because, as noted by *e.g.*, Renault et al. (2009, 2016b), QuikSCAT data do not measure the stress within the first  $\approx 30$  km from the coast due to land contamination in the backscatter measurements (Chelton et al., 2004). The upwelling season in spring and summer is marked by a distinctive alongshore surface stress (up to 0.09 N m<sup>-2</sup>), and the numerous capes and mountain ranges induce so-called expansion fans (Winant et al., 1988). The main discrepancies between SCOW and WRF occur close to the coast. In the simulation there is a coastal band where the surface stress is reduced compared to its offshore value (*i.e.*, the wind drop-off). Such a slackening of the wind is mainly caused by the presence of coastal orography, coastline shape, the difference between marine and terrestrial drag coefficients, and SST; this drop-off pattern is not well captured in QuikSCAT. An indirect validation of the wind drop-off is given by the oceanic response. A too wide drop-off causes a poor representation of mean oceanic current structure and mesoscale activity (Renault et al., 2016b). Finally, the stress magnitude is slightly underestimated by USW4: the mean biases over the whole domain are 0.006 N m<sup>-2</sup> and 0.003 N m<sup>-2</sup> for the meridional and the zonal surface stress, respectively, *i.e.*, the same

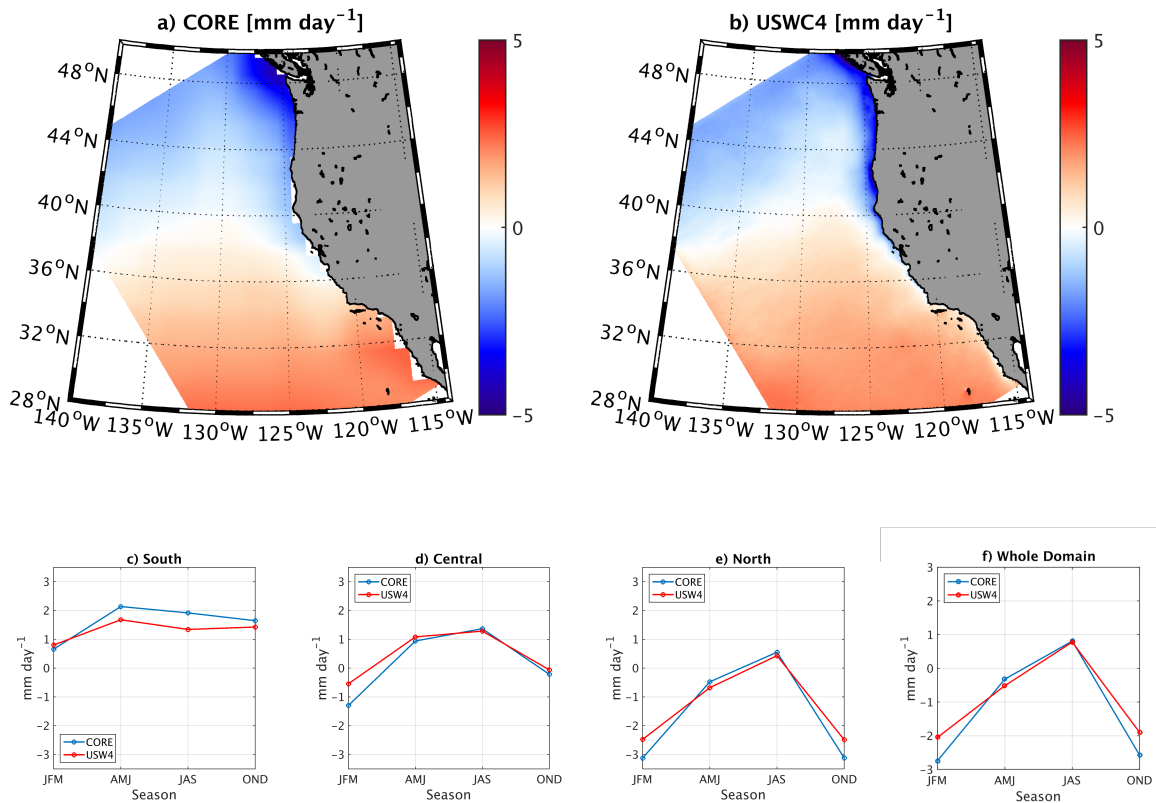


Figure 5: Net surface freshwater flux [ $\text{mm day}^{-1}$ ] from (a) CORE and (b) USW4 over the period 1995-2006. Panels (c), (d), (e), and (f) represent the seasonal evolution of the net surface freshwater flux over the same period from CORE (blue) and USW4 (red), averaged over the boxes indicated in Fig. 1 or over the whole domain. The simulation reproduces the main spatial pattern of the observed freshwater flux and its seasonal variability.

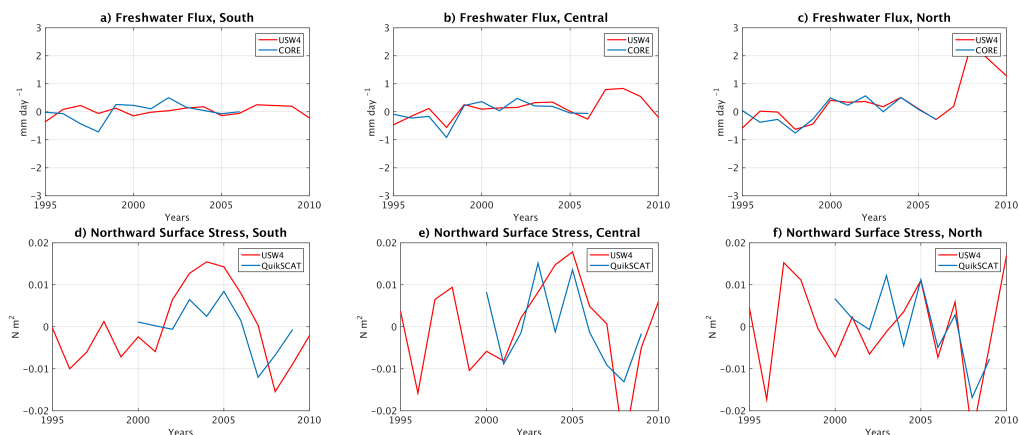


Figure 6: Same as Fig. 3 but for the net freshwater flux [ $\text{mm day}^{-1}$ ] from CORE and northward surface stress [ $\text{N m}^{-2}$ ] from QuikSCAT.

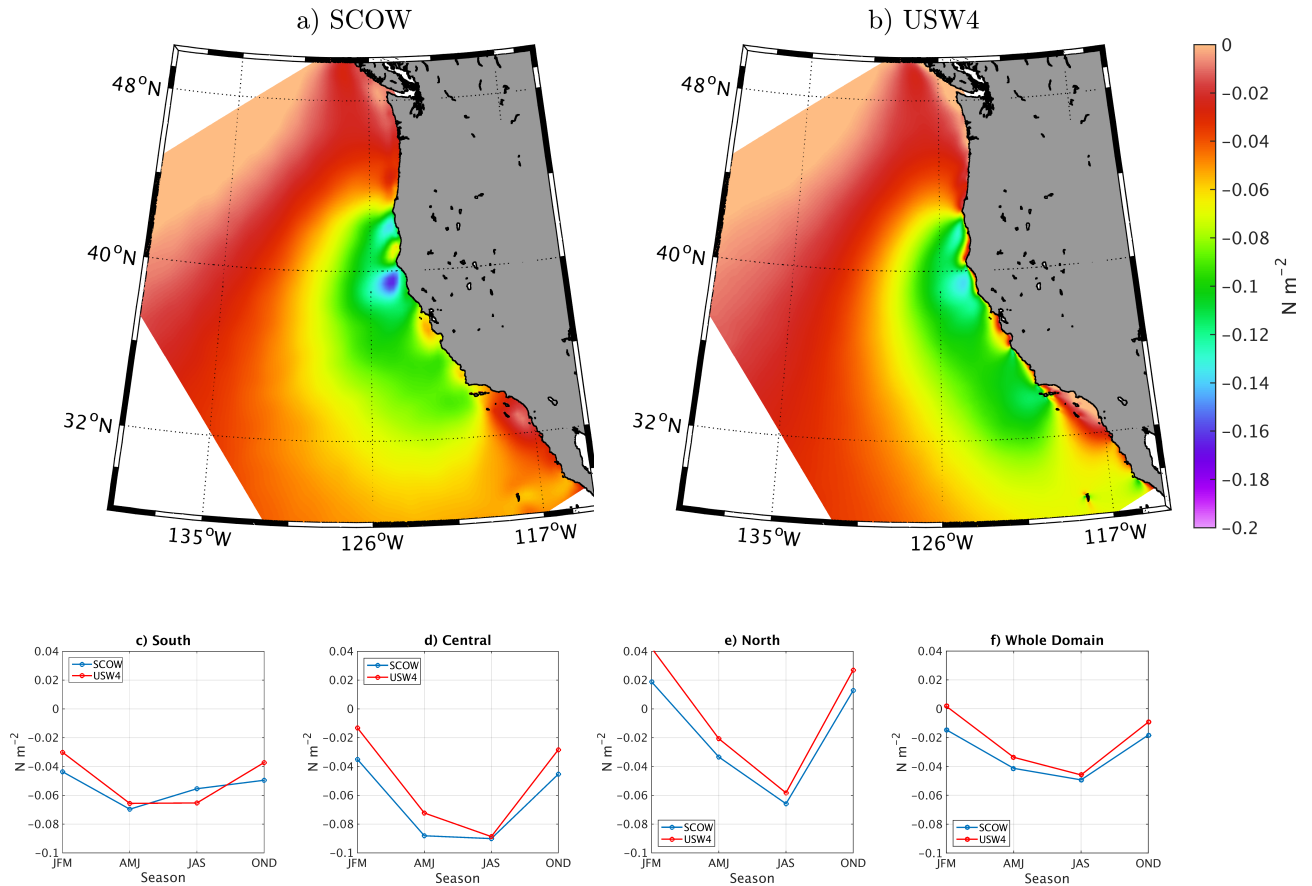


Figure 7: Mean meridional surface stress from (a) SCOW and (b) USW4 for the upwelling season (spring and summer) estimated over the period 2000-2009. Panels (c), (d), (e), and (f) represent the seasonal evolution over the same period from SCOW (blue) and USW4 (red), averaged over the boxes indicated in Fig. 1 or over the whole domain. USW4 reproduces the main surface stress spatial pattern and its seasonal evolution.

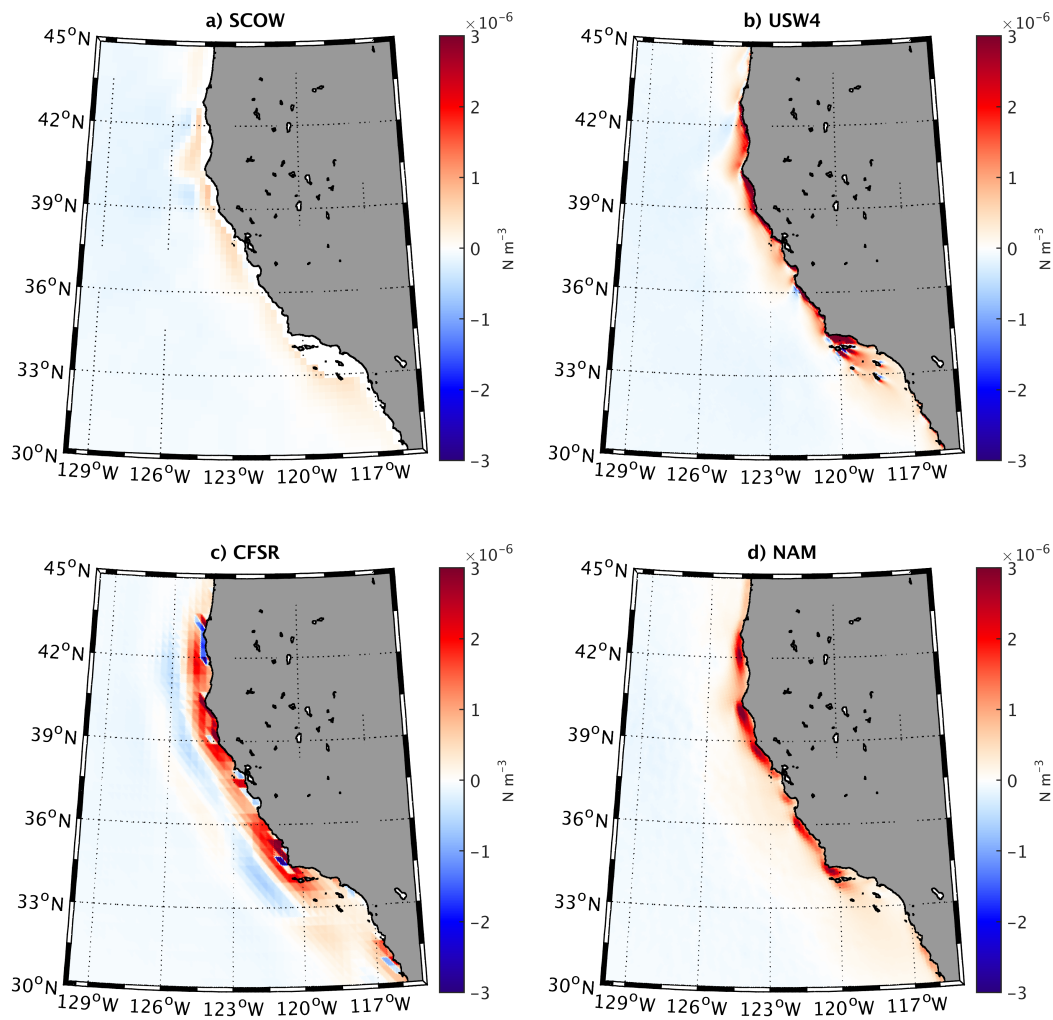


Figure 8: Mean surface stress curl during the upwelling season (Spring and Summer) from (a) QuikSCAT, (b) USW4, (c) CFSR, and (d) NAM. The USW4 atmospheric forcing allows to simulate a wind-drop off, and fine-scale stress curl such as island wind wakes. Note the nearshore blind zone (in white) of QuikSCAT in (a).



375 order of magnitude as the QuikSCAT surface stress error (Risien and Chelton, 2008). The least  
376 skillfully simulated region is again the Southern California Bight, where the surface stress is weak  
377 compared to other parts of the CCS, and it is overestimated in USW4; perhaps the complicated  
378 small island geometry is a contributing cause. Similar results are found using the QuikSCAT  
379 interannual product (Bentamy and Fillon, 2012).

380 The surface stress curl during the upwelling season from SCOW and USW4 is then compared  
381 in Fig. 8 to CFSR and NAM that are widely used to force oceanic models. As noted previously,  
382 because of the blind zone of QuikSCAT, SCOW does not represent nearshore large positive values  
383 of the surface curl caused by the wind drop-off. USW4 has large positive values of surface stress  
384 curl. Consistent with Renault et al. (2016b), the surface stress curl has spatial and seasonal vari-  
385 ability (not shown) in both its offshore extent and intensity. The offshore extent varies from around  
386 10 to 80 km from the coast and the surface stress reduction from 10 to 80 % (corresponding to the  
387 largest values of the curl). The largest curl values are situated when the mountain orography is  
388 combined with the coastline shape of a cape. Interestingly, the Santa Barbara Channel is character-  
389 ized by the presence of many fine-scale wind structures that are induced by cape effects and island  
390 mass effects. As an indirect validation, Kessouri et al. (2021a), shows that these fine-scale  
391 wind structures are responsible for intense blooms. The surface curl from CFSR does represent  
392 a drop-off, however, it is characterized by a too large cross-shore extent that is a characteristic of  
393 too coarse spatial resolution atmospheric model ( $\approx 35$  km). It also presents a pattern of positive,  
394 negative and positive values parallel to the coast, which is characteristic of the Gibbs phenomenon  
395 in spectral models (Hoskins, 1980). CFSR does not either represent any of the fine-scale wind  
396 structure along the coast. As shown by Renault et al. (2016a) and Kessouri et al. (2021a), such  
397 a wind product does not allow to represent the current structure nor the mesoscale activity and,  
398 thus, the net primary production along the CCS. NAM is a reanalysis over the United States of  
399 America from NCEP. It is a configuration of the WRF model with a spatial resolution of  $dx =$   
400 12 km. Although the representation of the wind drop-off is improved with respect to CFSR, the  
401 spatial extent of the drop-off is still too large and it does not represent all the fine-scale structures  
402 simulated by our simulations such as the island mass effects over the Santa Barbara Channel.

403 Finally, Fig. 6def shows the interannual yearly values of northward surface stress from QuikSCAT  
404 and USW4 averaged over the South, Central, and North boxes. Again, the interannual variability  
405 is reproduced by the model.

## 406 **4 Oceanic Fields**

### 407 **4.1 Surface Layer**

408 To evaluate the performance of the simulated oceanic simulation in terms of seasonal SST, we use  
409 the Ostia product as a comparison (Fig. 9), examining both Nearshore and Offshore boxes. In  
410 global coupled models, eastern boundary upwelling systems, such as the CCS, are characterized  
411 by large SST biases (up to 3°C; *e.g.*, Richter (2015)). The origin of these biases is not well under-  
412 stood, but it is likely to be caused by poor representations of the cloud cover, surface wind pattern,  
413 oceanic upwelling, and cross-shore eddy heat flux. In USW4 the mean SST large-scale patterns  
414 are qualitatively well represented compared to the SST satellite measurements, with warmer wa-  
415 ters to the west and south of the domain and colder waters to the north (Fig. 9). The favorable

416 upwelling season (*i.e.*, spring and summer) is captured by the simulation, and the upwelling signa-  
417 ture is clearly marked in a 30 km wide coastal strip (Fig. 9). The simulated SST has a weak cold  
418 bias of  $0.5^{\circ}\text{C}$  over the whole domain in all the seasons, and the coastal water is colder in the model  
419 than in the Ostia product by up to  $1^{\circ}\text{C}$  (Figs. 9 and 10). Very nearshore the bias can reach up to  
420  $2^{\circ}$ , but it is likely due to the limitation of the SST product. Ostia SST has a relatively high spatial  
421 resolution of nominally  $dx = 5$  km. However, high cloud cover over the upwelling season impedes  
422 access to high-resolution data. Therefore, the effective resolution of this product may be similar  
423 to that of the microwave satellite products ( $dx = 25$  km); this may partially explain the nearshore  
424 SST discrepancies between USW4 and the Ostia product. The largest bias is situated in the South-  
425 ern California Bight where the atmospheric forcing is also less skillful (*i.e.*, overestimation of the  
426 surface stress and underestimation of the shortwave radiation during summer). Generally, numeri-  
427 cal simulations have difficulty representing the southward extension of the cold water south of Pt.  
428 Conception (Marchesiello et al., 2003; Capet et al., 2008a; Renault et al., 2016a) (see also Fig. 10).  
429 USW4 has at least fair representation of this southward extension; *e.g.*, it has a better representa-  
430 tion of the SST and of its mean pattern than a climatological solution (or *e.g.*, the Veneziani et al.  
431 (2009) solution). In particular, in the Central Nearshore box, the climatological solution has a  
432 warm bias up to  $1.5^{\circ}\text{C}$ , whereas USW4 has a bias lower than  $0.5^{\circ}\text{C}$  there. Otherwise, the USW4  
433 SST compares well with the measurements, which is likely due to a good representation of the  
434 simulated atmospheric forcing (in particular, the cloud cover and wind drop-off) and of the surface  
435 currents.

436 Figure 11 depicts the interannual variation of the yearly mean SST in the nearshore region  
437 over the South, Central, and North boxes for Ostia and USW4. Consistent with Fig. 3, the in-  
438 terannual variability of the SST is well reproduced by USW4. In particular, the warm anomalies  
439 of  $> 0.5^{\circ}$  during 1997 and 2004 and the cold anomaly during 2008 are captured by the model.  
440 Similar results are found for the offshore region.

441 Sea Surface Salinity (SSS) is higher offshore in the subtropical gyre with its high evaporation  
442 rate and is lower in the subpolar gyre with the higher precipitation. In addition, it decreases weakly  
443 near the coast, more so in the north, mainly due to river inflow. We compare the large-scale pat-  
444 tern and seasonal cycle of SSS from USW4 to those from the WOA13 SSS large-scale pattern  
445 and seasonal cycle in Fig. 12. Due to a realistic representation of the freshwater flux by USW4  
446 (Sec. 3c), there is good agreement between the simulation and the measurements. However, off-  
447 shore in central California, the SSS is generally too low with respect to the measurements, with  
448 a maximum bias of 0.5 PSU. This is partially explained by the freshwater flux biases in Fig. 5,  
449 where the offshore flux is slightly underestimated. However, as discussed in Secs. 2.1 and 4.2, the  
450 bias is also partially inherited from the parent solution and its open boundary conditions. Near the  
451 Columbia River the USW4 SSS is conspicuously fresher than in WOD13, but the latter is probably  
452 horizontally overly smoothed.

453 In previous studies (*e.g.*, de Boyer Montégut et al. (2004)), the Mixed Layer Depth (MLD)  
454 definition can be based on different parameters such as temperature, salinity, and density. The  
455 MLD is typically defined using a threshold, for which the MLD is the depth at which potential  
456 temperature or potential density changes by a specified small value relative to its value near the  
457 surface. Here the CARS analysis definition is applied to the daily average temperature and density  
458 field in USW4. The MLD is defined using a temperature threshold of  $\Delta\Theta = 0.2^{\circ}$  and  $\Delta\sigma_{\theta} =$   
459  $0.3 \text{ kg m}^{-3}$ . The near-surface reference depth is 10 m. MLD is shallower near the coast due to an  
460 uplifted pycnocline, and it is deeper in summer in the offshore subpolar gyre due to its depressed

461 pycnocline compared to the offshore subpolar gyre with its uplifted pycnocline. The CARS MLD  
462 and the one estimated from USW4 are compared in Fig. 13. The phase and amplitude of the  
463 seasonal cycle are similar in both the model and measurements: a shallowing of the MLD during  
464 spring and summer, then a deepening from 20 m to 80 m in winter. The MLD in the model is  
465 slightly too deep compared to the climatology; this could be related to the surface forcing and to  
466 the KPP parameterization scheme (Large et al., 1994) for vertical mixing of tracers and momentum  
467 in ROMS. The vertical mixing that is too deep partially explains the cold bias in the simulated SST.

## 468 4.2 Interior $T$ and $S$

469 The CCS is stably stratified almost everywhere. It has warm temperature and fresh salinity in  
470 and above the pycnocline compared to below, and the pycnocline tilts upward toward the coast  
471 due to upwelling. Systematic large-scale hydrographic sampling of the CCS was initiated in 1949  
472 by the California Cooperative Oceanic Fisheries Investigations (CalCOFI) program. Along the  
473 zonal line 80 of the CalCOFI data (offshore from Pt. Conception, centered on 33°N), the vertical  
474 structure of the simulated temperature, salinity, and density are in general agreement with the  
475 CalCOFI climatology (Fig. 14). In both the observations and in USW4, isotherms, isohalines, and  
476 isopycnals are characterized by a positive cross-shore slope. At the surface, consistent with Fig. 9,  
477 the mean SST is well reproduced with biases lower than 0.5 °, which is likely due to a realistic  
478 representation of the net surface heat flux. As shown in Fig. 12, the mean salinity in the upper  
479 layer is too low with respect to CalCOFI (by 0.2 PSU). Finally, at depth, the mean density field is  
480 also realistic; nevertheless, there is a cold temperature bias of 1 ° (a negative density bias) that is  
481 partly compensated for by a fresh salinity bias of 0.5 PSU (a positive density bias). Similar results  
482 are found for the Newport line using the WOD13 dataset (1955-2013, Fig. 14).

483 Figure 15 shows the mean temperature, salinity, and density biases at 150 m depth. In the first  
484 500 km from the coast, the density is realistic with a very weak bias by less than 0.1 kg m<sup>-3</sup>, and  
485 nearshore, where there is more data, the bias is less than 0.05 kg m<sup>-3</sup>. However, consistent with  
486 Fig. 14, there is a bias compensation between the temperature and salinity. As suggested by Fig. 14,  
487 the temperature and salinity biases are mostly inherited from the open boundary conditions (*i.e.*,  
488 Mercator fields corrected by WOA) used in the parent-grid solution. Sensitivity tests have been  
489 made to reduce these biases. For example, more realistic results are obtained by using Mercator  
490 (compared to the Simple Oceanic Data Assimilation (SODA) by Carton and Giese (2008)) as the  
491 open boundary condition of the parent simulation and by correcting these data with WOA (not  
492 shown). In general, however, the sampling density of measurements in the offshore region is rather  
493 small, and we choose not to artificially diminish our large-scale biases by adjusting the boundary  
494 conditions within their (considerable) level of uncertainty. Deutsch et al. (2021a) discusses the  
495 importance of correctly reproducing the density field in specifying the biogeochemical boundary  
496 conditions.

497 The temperature and salinity variability in USW4 is furthermore evaluated by comparing their  
498 standard deviations (removing the long-term mean) at 150 m to the World Ocean Database 2013  
499 (WOD13, Fig. 16). In the nearshore region (first 200 km from the coast), as in the measurements,  
500 the upwelling has a signature on the water masses: there is a weaker temperature and salinity  
501 variability with respect to offshore. Offshore the  $T$  and  $S$  variability is mainly due to variations in  
502 the pycnocline depth in the subtropical gyres. In USW4 the offshore salinity variability is slightly  
503 too weak compared to the measurements; while no doubt part of this discrepancy may be due to

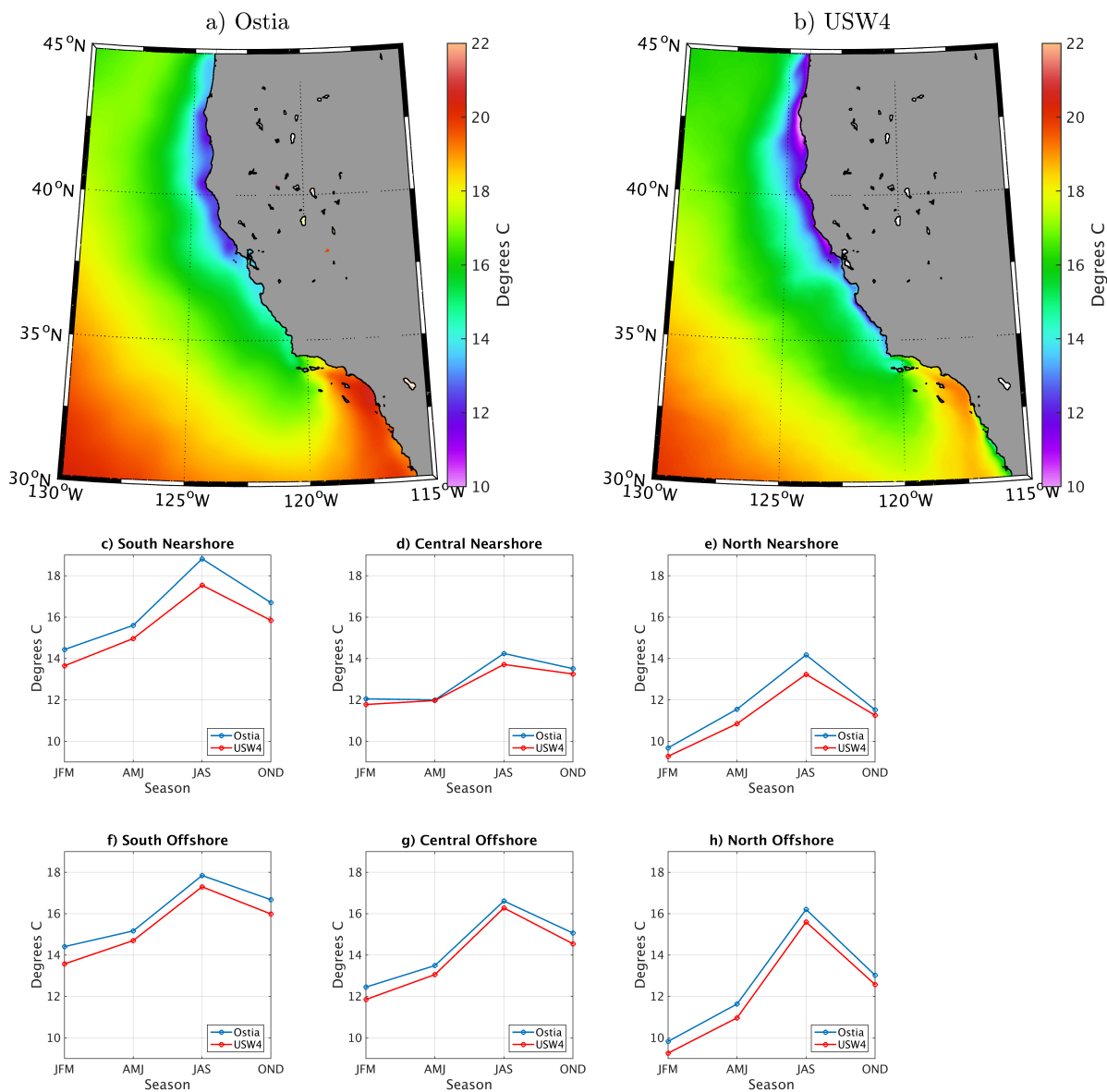


Figure 9: Mean SST [ $^{\circ}$ C] during summer from (a) Ostia and (b) USW4 (1995-2010). Panels (c) to (h) represent the seasonal evolution of the SST over the same period from Ostia (blue) and USW4 (red) and averaged over the Nearshore and Offshore boxes indicated in Fig. 1 or over the whole domain. The SST patterns are well matched in USW4 with a mean negative bias of about  $0.5^{\circ}$ C.

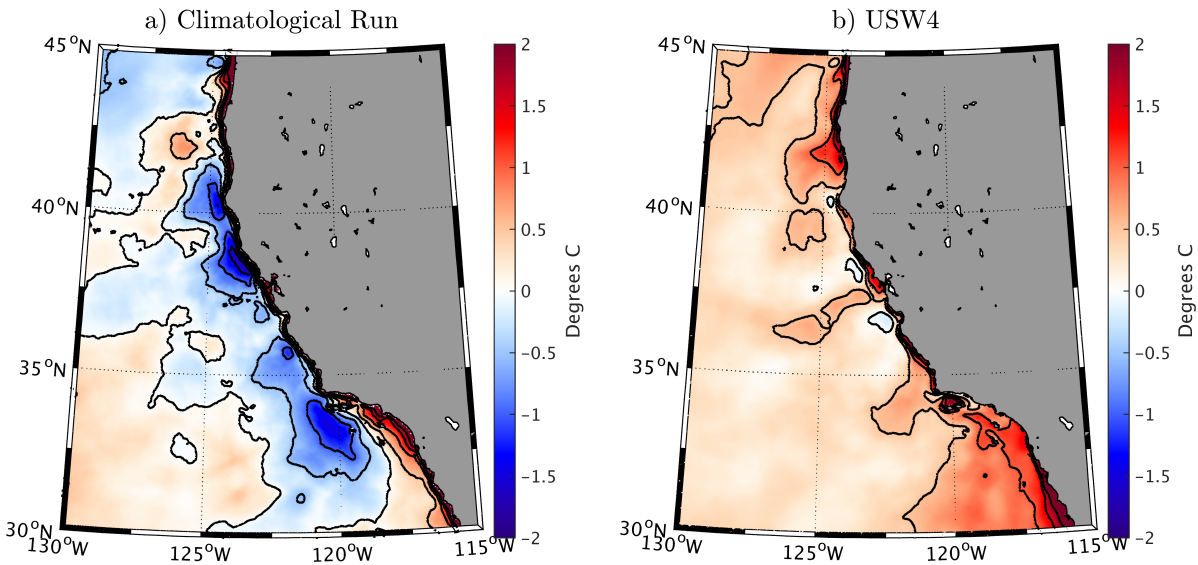


Figure 10: (a) Mean SST differences [°C] during summer between Ostia and (a) a climatological solution (*e.g.*, Capet et al. (2008b); Renault et al. (2016a)) and (b) USW4. USW4 has a cold bias ( $> 0.5$  °), in particular over the Southern California Bight; however, it is less biased than the climatological solution (up to 2°C; see text in Sec. 4.1).

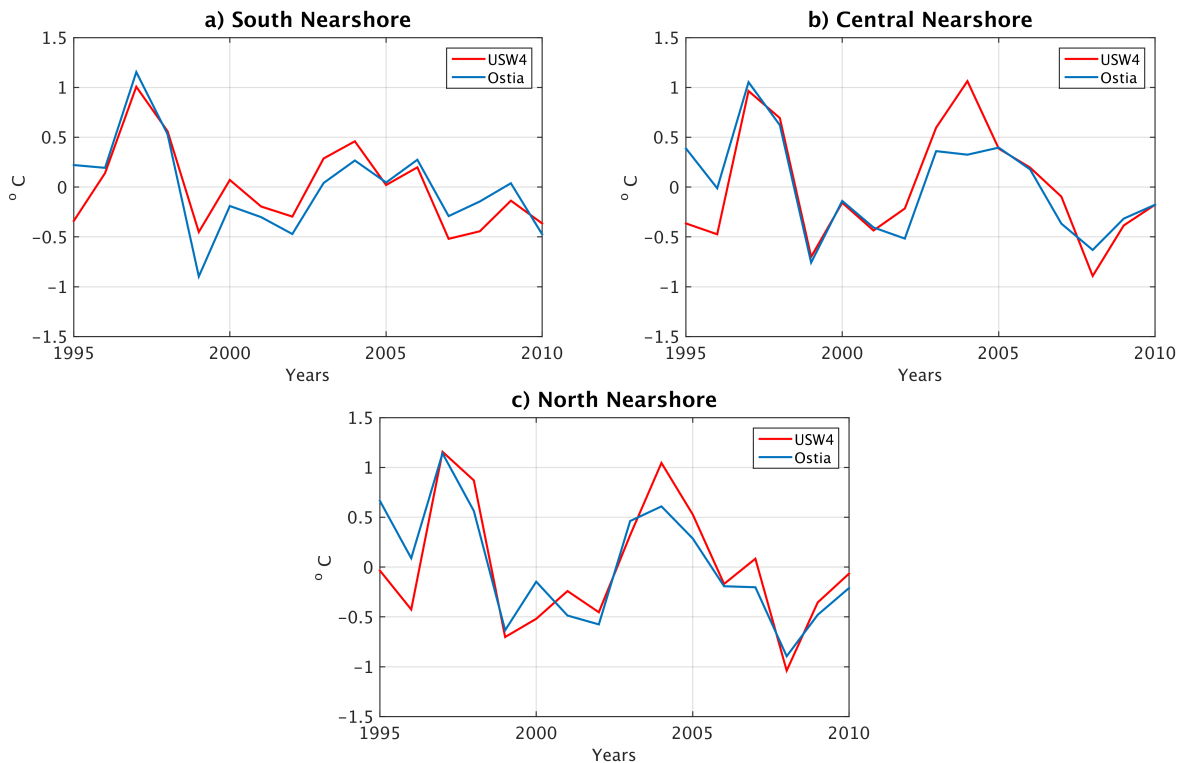


Figure 11: Interannual SST [°C] over the boxes indicated in Fig. 1 from OSTIA and USW4. Similar results are found over the offshore boxes.

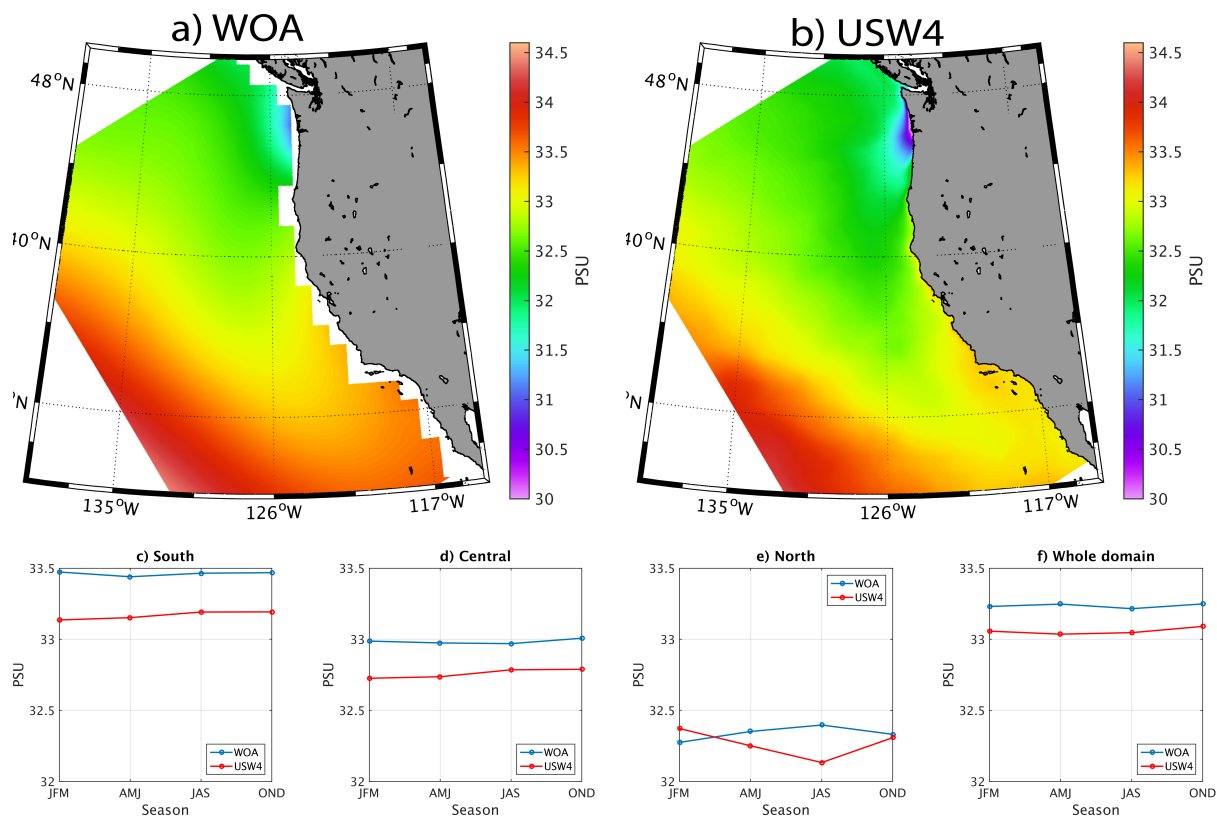


Figure 12: Mean SSS [PSU] from (a) the World Ocean Atlas and (b) USW4 (1995-2010). Due to a realistic freshwater flux, the mean SSS in USW4 is consistent with the observations despite a small negative bias (up to 0.5 PSU).

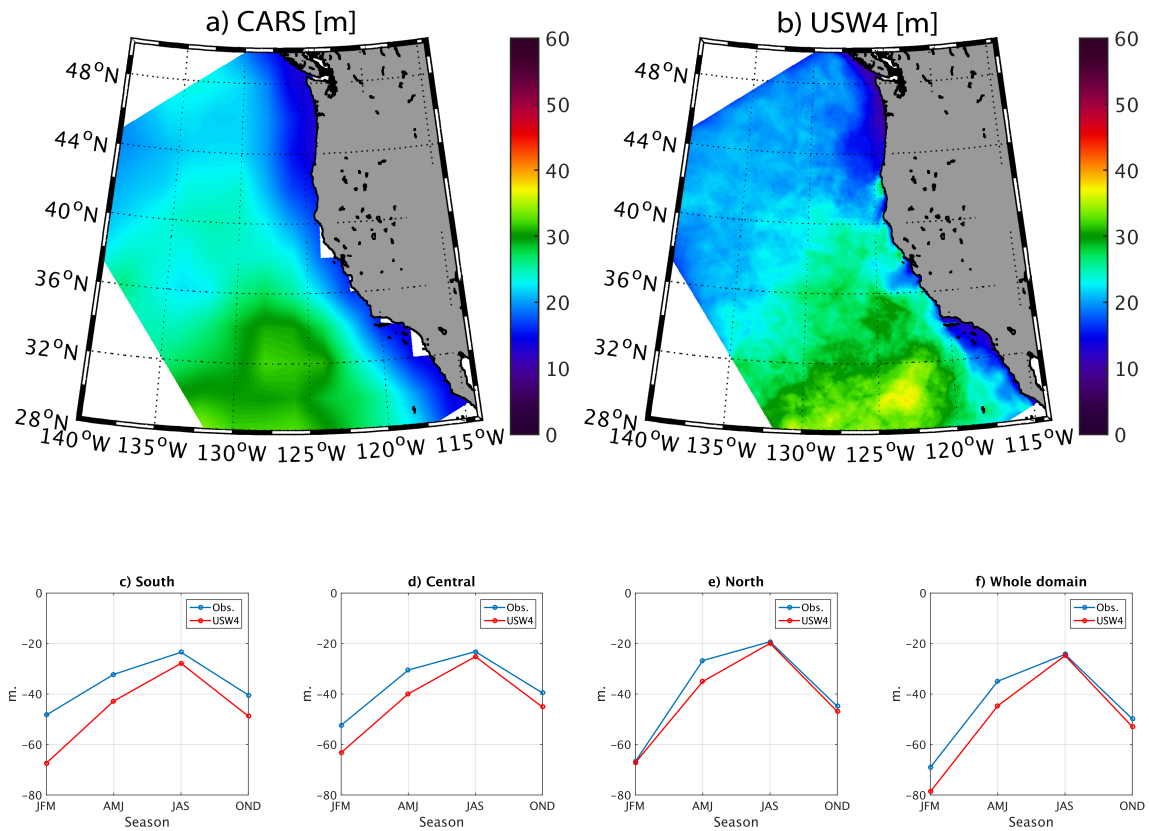


Figure 13: Mean Mixed Layer Depth (MLD) [m] estimated from (a) CARS and (b) USW4 during summer (1995-2010). Panels (c), (d), (e), and (f) represent the seasonal evolution of MLD over the same period from CARS (blue) and USW4 (red), averaged over the boxes indicated in Fig. 1 or over the whole domain. The simulation has a realistic MLD that is deeper farther offshore and during winter, but with a slight overestimation (up to 20 m in the Southern California Bight during winter) that partially explains the cold bias in SST in USW4.

504 model bias (*e.g.*, inherited from the parent solution and boundary conditions), it could also partially  
505 be explained by uncertainties in the measurements (*e.g.*, the salinity variability pattern in WOD13  
506 is somewhat noisy). However, the offshore temperature variability is well reproduced by the model.

507 Finally, Fig. 17 shows the Potential Temperature - Salinity (TS) diagram as estimated from  
508 WOA and from USW4 between 30°-49°N in the upper 1000 m near the coastline (0-100 km off-  
509 shore). The mean water masses of the CCS are realistic in USW4, although there are cold temper-  
510 ature ( $\approx 0.5^\circ$ ) and fresh salinity biases (0.2 PSU in the upper layer, maximum bias of 0.5 PSU)  
511 consistent with Fig. 14.

### 512 **4.3 Mean Vertical Velocities during the Upwelling Season**

513 Figure 18a represents the mean vertical velocities at 30 m depth (that is near the vertical peak of  
514  $w$  nearshore) during the upwelling season as simulated in USW4. Consistent with the literature,  
515 the CCS region is characterized by various upwelling cells. The largest vertical velocities reach  
516 values greater than  $0.5 \text{ m s}^{-1}$  on average are located between 42°N and 43°N, 40°N, 39°N, 38°N,  
517 36°N and in the Santa Barbara Channel, *i.e.*, near capes, complex orography, and coastline that  
518 strengthen the wind. The associated subseasonal variability is shown in Fig. 18b. It reveals a large  
519 variability reaching up to  $2 \text{ m s}^{-1}$  in the nearshore region and a non-negligible variability offshore  
520 of  $0.5 \text{ m s}^{-1}$ . Such a variability is associated with wind bursts that induce intense upwelling and  
521 larger turbulent heat fluxes (Renault et al., 2009) but also to the mesoscale activity. The interannual  
522 variability is also relatively large (Figure 18c) with values greater than  $0.5 \text{ m s}^{-1}$  nearshore, and  
523 that is mainly associated with the interannual variability of the wind (see *e.g.*, Fig. 6).

### 524 **4.4 Mean Sea Surface Height and Current**

525 The SSH (Sec. 2c) from the 16 years of USW4 is shown in Fig. 19ab, along with measurements  
526 from the  $1/4^\circ$  resolution CNES-CLS13 dataset (Rio et al. (2014), Sec. 2d). The spatial distribution  
527 and amplitude of the simulated SSH is in good agreement with the measurements. The mean  
528 Sea Surface Height (SSH) in the CCS is depressed at the coast due to the southward geostrophic  
529 current, and it further decreases poleward due to the equatorward wind stress. The main differences  
530 between the model and measurements are located along the coast. Such discrepancies can be  
531 attributed partially to the Nearshore box width (50 km) which is unresolved in the satellite data  
532 (Ducet et al., 2000; Rio et al., 2014). The negative cross-shore SSH slope is reproduced by USW4.  
533 Interestingly, the alongshore standing eddies are much less evident in USW4 and in the CNES-  
534 CLS13 dataset than in drifter measurements (Centurioni et al., 2008) or the model by Marchesiello  
535 et al. (2003), likely because of the longer time averaging used here. To better highlight the presence  
536 of standing eddies, an Empirical Orthogonal Function (EOF) analysis is applied to the 16 years of  
537 daily SSH from USW4 after removing the mean state (Sea Level Anomaly, SLA) over a nearshore  
538 region shown in Fig. 19cd. The obtained modes have therefore to be interpreted as the variation  
539 of the circulation with respect to the mean state. Here focus is mainly done on the modes that  
540 are characterized by the presence of standing eddies. The first EOF mode (not shown) explains  
541 34.8% of the variance and depicts the steric contribution. The second EOF mode (not shown)  
542 explains 22.3% of the variance, it represents the seasonal variation of the surface currents (*e.g.*,  
543 southward intensification during the upwelling season, see below). More interestingly, the third  
544 and fourth modes explain 12.2% and 6.5% of the variance, respectively. Figure 19cd depicts their



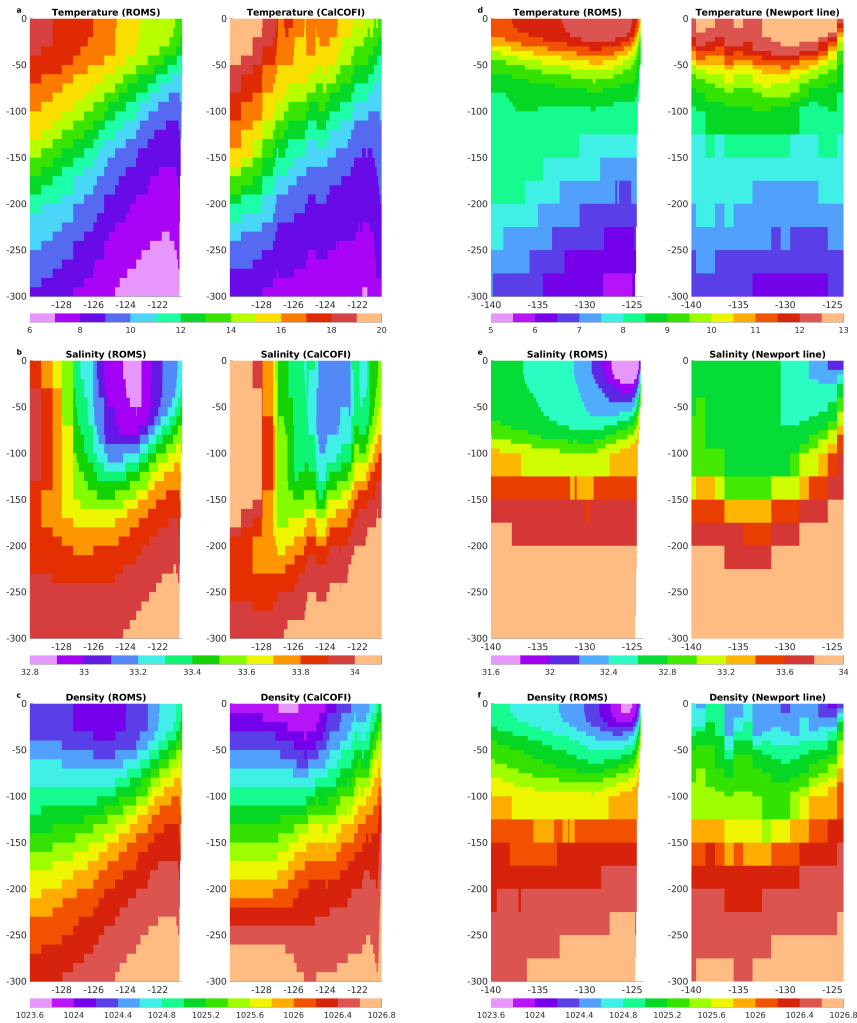


Figure 14: Mean cross-shore section along the CalCOFI lines 80 (left panels,  $\approx 33^\circ\text{N}$ ) and Newport line from WOD13 (right panels,  $\approx 45^\circ\text{N}$ ) of (a,d) temperature [ $^\circ\text{C}$ ], (b,e) salinity [PSU], and (c,f) density [ $\text{kg m}^{-3}$ ] from USW4 (1995-2010) (left column) and the measurements (period 1955-2013, right column). USW4 has approximately the right cross-shore density slope induced by the wind-driven upwelling. At the surface the salinity is too low with respect to CalCOFI. At depth the density is similar to that in the observations, but there is a cold temperature bias (a positive density bias of  $\approx 1^\circ\text{C}$ ), partially compensated for by a fresh salinity bias (a negative density bias) of  $\approx 0.2$  PSU).

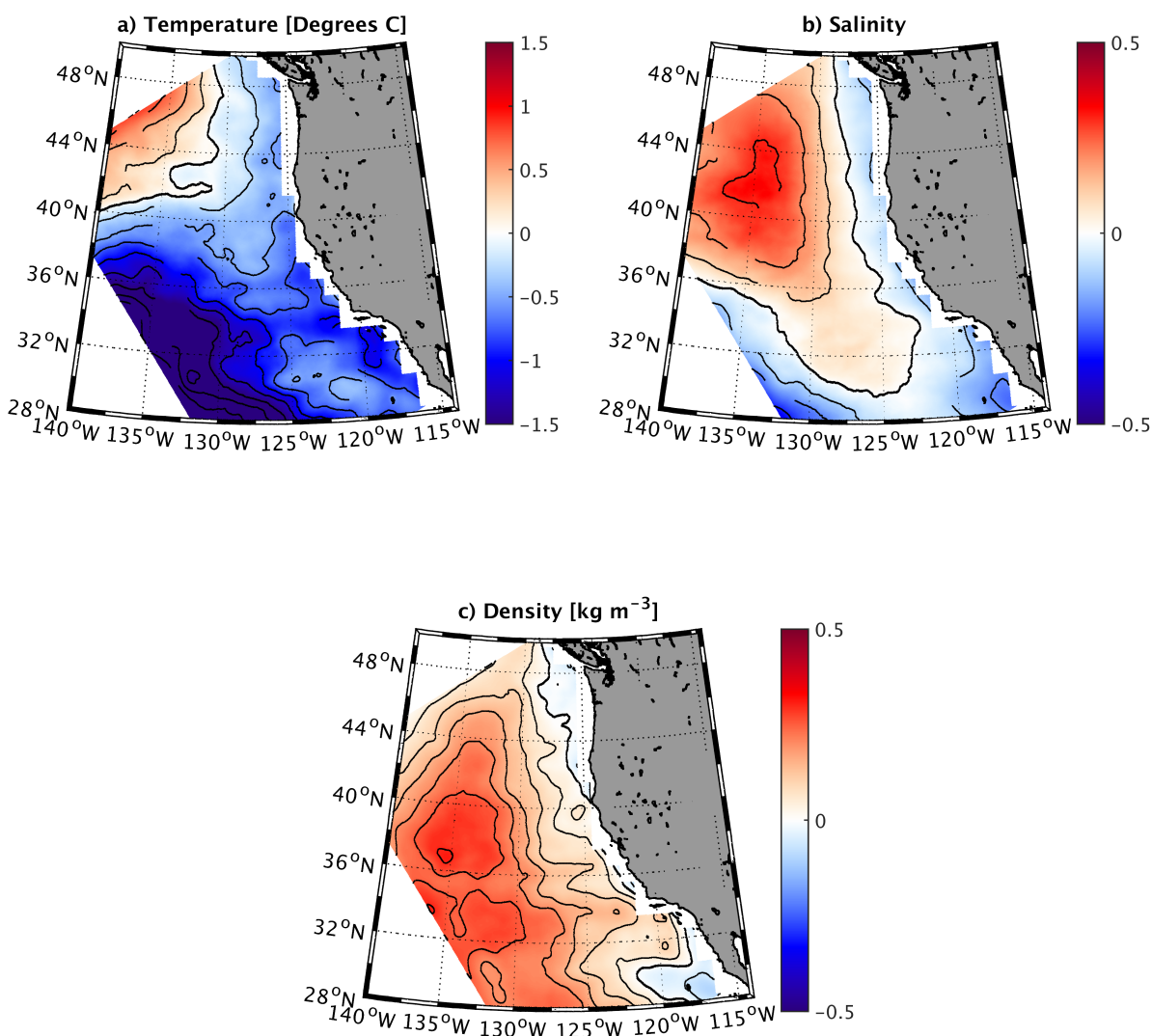


Figure 15: Mean temperature [ $^{\circ}\text{C}$ ], salinity [PSU], and density [ $\text{kg m}^{-3}$ ] differences at 150 m depth between USW4 (1995-2010) and WOA. The contour lines difference isolines, with the thick black line indicating zero difference. In the first 500 km the density at 150 m is realistic, with a very weak bias of less than  $0.1 \text{ kg m}^{-3}$ ; nearshore, where there is more data, the bias is less than  $0.05 \text{ kg m}^{-3}$ . However, consistent with Fig. 14, there is a compensation between temperature and salinity biases. Most of the salinity bias enters the domain through the northern open boundary condition.

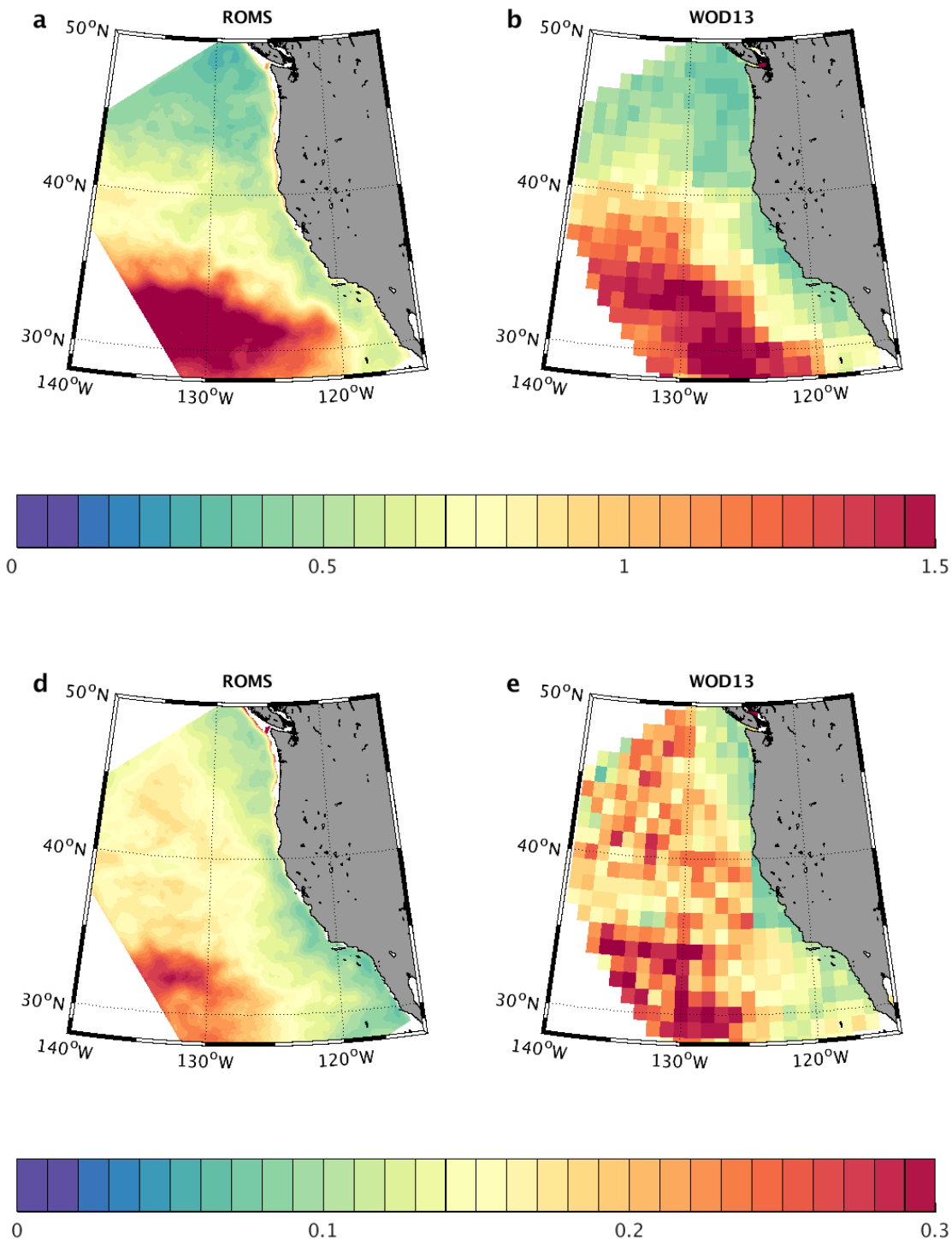


Figure 16: Top panel: Temporal standard deviation of monthly temperature [°C] at 150 m depth for (a) USW4 and (b) (WOD13). Bottom panel: same as the top panel but for the salinity [PSU]. There is a general agreement between the simulated temperature and salinity variability at 150 m depth and the measurements.

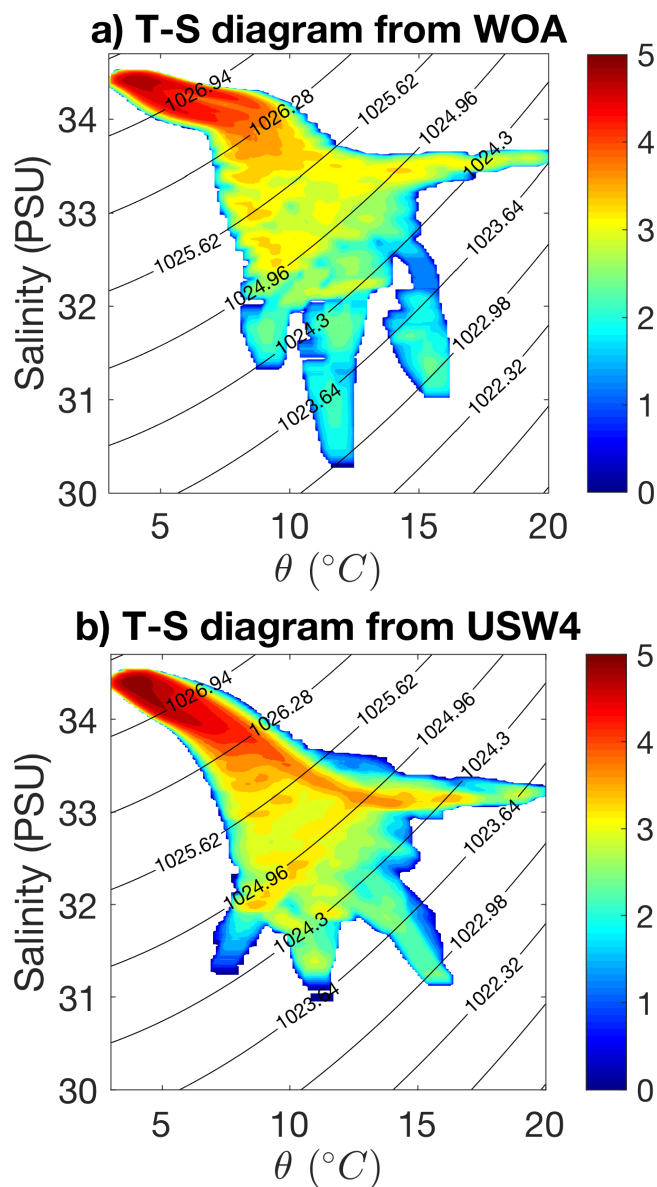


Figure 17: Potential Temperature - Salinity (TS) diagram from (a) the WOA measurements and (b) USW4, for the climatological mean between  $30^{\circ}$ - $49^{\circ}$  in the upper 1000 m near the coast (0-100 km offshore). The colorbar shows the number of data points in each ( $1^{\circ}\text{C}$ ,  $0.1\text{PSU}$ ) bin on the logarithmic scale, and black contour lines are those of density. The abscissa is potential temperature with the surface as the reference level, and the ordinate is salinity. To obtain the number of data points in each bin, we first obtain the (T,S) dataset in each season, averaged over the years 1995-2010, in the selected region regridding both measurements and USW4 over the a grid with a spatial resolution of  $dx = 4$  km in the offshore direction,  $0.05^{\circ}$  ( $\approx 5$  km) in the along-shore direction, and 20 m in the vertical direction; then, the number of data points in each bin is counted.

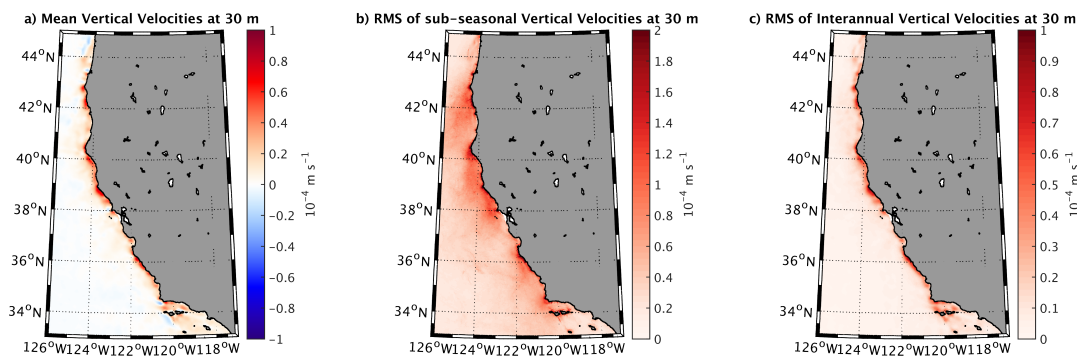


Figure 18: Vertical velocities [ $\text{m s}^{-1}$ ] during the upwelling season at 30 m depth as simulated by USW4. (a) Long-term mean, (b) subseasonal variability, (c) interannual variability.

545 spatial patterns and Figure 19ef their associated temporal variations and spectrum. Both Mode 3  
546 and 4 reveal the presence of standing eddies from 200 km offshore in particular around 39°N and  
547 36°N. Spectral analysis of the associated series reveals significant energy peaks at  $\approx 630$  and 10  
548  $\text{day}^{-1}$ . The 95s is estimated by a Markov red noise (Gilman et al., 1963). The 10  $\text{day}^{-1}$  peak likely  
549 represents wind burst that induce a modulation of the currents. Mode 3 also reveals a variation of  
550 the nearshore current that is likely responsible of the frequency peak near 25 and 90  $\text{days}^{-1}$ .

551 The CCS exhibits broad-band variability. On time and space scales larger than the mesoscale,  
552 most of the variability is partly extrinsic to the region, reflecting the larger-scale seasonal, inter-  
553 annual, and decadal climate signals that have regional manifestations along the U. S. West Coast  
554 (Chhak and Di Lorenzo, 2007; Di Lorenzo et al., 2009; Chenillat et al., 2012; Meinvielle and  
555 Johnson, 2013; Davis and Di Lorenzo, 2015). To give a sense of this variability, Fig. 20 shows  
556 the whole-coast history of the sea-surface height anomaly (SLA) and depth of a pycnocline isopy-  
557 cnal over the hindcast period (See also Fig. 22 in the biogeochemical companion (Deutsch et al.,  
558 2021a)). The along-coast coherence of SLA is striking, as is the regularity of its primarily seasonal  
559 oscillation. Its amplitude increases to the north because the amplitude of the seasonal cycle of the  
560 alongshore wind and current increase as well. A rapid poleward propagation speed  $\approx 2.5 \text{ m s}^{-1}$  is  
561 apparent on average, as has been extensively analyzed previously (Chelton, 1984; Spillane et al.,  
562 1987). The interannual variability amplitude is a modest fraction of the seasonal one, most of the  
563 time, but the 1997-98 ENSO event is particularly prominent (Kosro, 2002; Ryan and Noble, 2002;  
564 Lynn and Bograd, 2002).

565 The pattern in pycnocline depth is somewhat more complex, although the temporal correlation  
566 with SLA is evident ( $C \approx 0.8$ , more or less uniformly along the coast). The amplitudes of both  
567 quantities have an increasing trend to the north, but the depth anomaly exhibits more modulation in  
568 amplitude than the SLA with moderate drops at particular latitudes that are related to interruptions  
569 in the path of the California Undercurrent (CUC) along the coast (Chen et al., 2021).

570 The spatial pattern of the geostrophic current estimated from the observed and simulated SSH  
571 is also in good agreement (not shown). Figure 21a shows the nearshore meridional surface current  
572 during the upwelling season. The nearshore surface current is broad and generally equatorward,  
573 as observed (Swenson and Niiler, 1996). It reaches values up to  $0.2 \text{ m s}^{-1}$ . The nearshore surface  
574 current properties exhibit strong latitudinal variability (values from  $\approx 0$  to  $0.2 \text{ m s}^{-1}$ ). The maximum  
575 amplitude of the current is situated along the coast and near capes, *i.e.*, where the wind is more

576 intense. Figure 21bc represents the subseasonal and the interannual variability of the meridional  
577 surface current. The subseasonal variability of the surface meridional current during the upwelling  
578 season is large, reaching values up to  $0.3 \text{ m s}^{-1}$ , *i.e.*, larger than the seasonal averages. Such a  
579 variability is mainly associated with the mesoscale activity and with wind bursts that modulate the  
580 Ekman transport and the upwelling-associated geostrophic currents.

581 Figure 22 depicts coastal cross-shore sections of the seasonal meridional current averaged  
582 along CalCOFI glider line 66.7 (Rudnick et al. (2017) and [https://spraydata.ucsd.edu/  
583 climCUGN/](https://spraydata.ucsd.edu/climCUGN/)), *i.e.*, near San Francisco Bay, for winter, spring, summer, and fall. As reported in the  
584 measurements, the coastal CCS during the upwelling season is characterized by an equatorward  
585 surface current with a mean velocity of  $0.1 \text{ m s}^{-1}$  overlying the CUC. The CUC, one of the major  
586 components of the CCS, is a poleward flow in the upper hundreds of meters near the U. S. West  
587 Coast. It transports warm and salty equatorial equator poleward and plays a significant role in the  
588 local heat, salt and biogeochemical budgets. Quite a few studies exist characterizing the features  
589 and exploring the dynamics of the CUC (*e.g.*, McCreary et al. (1987); Lynn and Simpson (1987);  
590 Pierce et al. (2000); Gay and Chereskin (2009); Molemaker et al. (2015); Rudnick et al. (2017)).  
591 Using USW4, Chen et al. (2021) assess the CUC dynamics and show that topographic form stress  
592 is a significant northward acceleration effect for this current both in its mean and low-frequency  
593 variability.

594 The CUC structure is well represented in USW4 for the winter, summer, and spring seasons.  
595 In both USW4 and the observations, the core of the CUC is relatively shallow during winter and  
596 fall (50 m depth) and is deeper during summer (100 m depth). In summer, a surface equatorward  
597 current ( $0.05 \text{ m s}^{-1}$ ) overlies the CUC whereas in fall, the CUC outcrops the surface, reversing  
598 poleward the surface current. In Winter, the CUC still outcrops the surface but the surface current  
599 remains equatorward. These is also an indirect validation of the simulated wind drop-off, as a poor  
600 representation of the slackening of the wind toward the coast (as in CFSR) may cause occasional  
601 surfacing of the California Undercurrent (CUC) (Renault et al., 2016a) through Sverdrup dynam-  
602 ics: a positive surface stress curl produces a barotropic poleward flow that adds to the coastal  
603 baroclinic flow (McCreary and Chao, 1985; Lynn and Simpson, 1990; Marchesiello et al., 2003).  
604 The spring season is characterized by a bias in the representation of the CUC characteristics. In  
605 the observations, the CUC has intense velocities (up  $0.1 \text{ m s}^{-1}$ ) and has a core reaching a depth  
606 of 200 m (in particular in June, not shown). In USW4, the CUC remains weak (velocities of  $0.05$   
607  $\text{ m s}^{-1}$ ) and its core it not deep enough (100 m depth). Note that the definition of the seasons here  
608 differ from that of Rudnick et al. (2017). This affects the interpretation of the seasonal cycle and  
609 of this bias as the main discrepancy occurs in June (summer in Rudnick et al. (2017), spring in this  
610 study).

611 To further assess the realism of the USW4 CUC, we characterize the interannual and the subsea-  
612 sonal variabilities of the meridional geostrophic currents along line 66.7 from the gliders and from  
613 USW4 Both gliders and USW4 reveal an interannual variability associated with current anomalies  
614 reaching up to  $0.05 \text{ m s}^{-1}$  (not shown). The subseasonal variability induced anomalies larger than  
615  $0.15 \text{ m s}^{-1}$  (not shown). It is mainly associated with displacement of the CUC related to remote  
616 forcing and local wind forcing and with the presence of eddies.

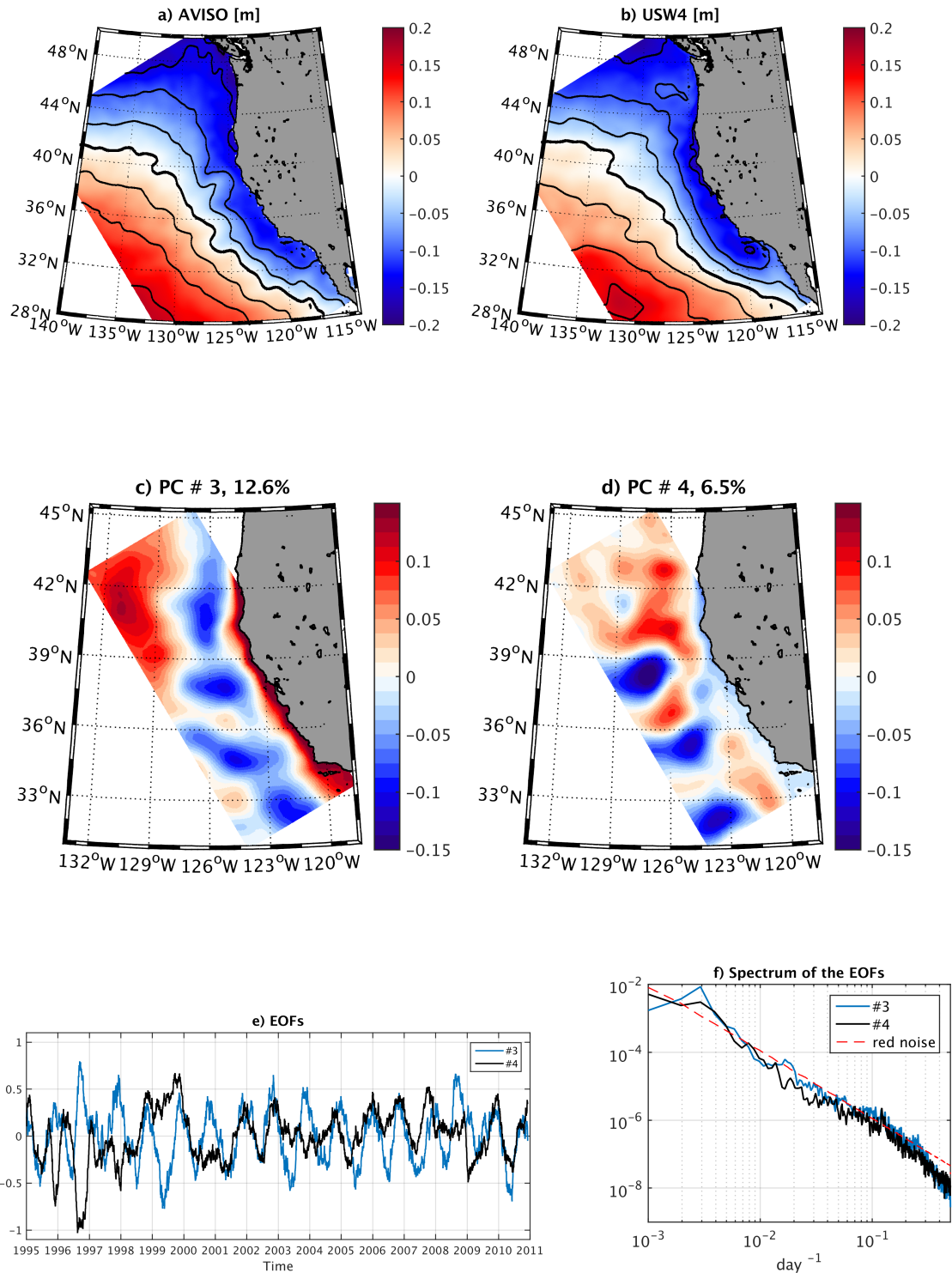


Figure 19: Mean Sea Surface Height (SSH) [m] from (a) AVISO and (b) USW4 over the period 1995-2010. Contours show 0.05 m increments of SSH, and the thick black line represents the local zero reference height contour. The simulation reproduces the mean SSH and its offshore gradient. c-d-e) represent the third and fourth mode of the Empirical Orthogonal Function (EOF) decomposition of the Sea Level Anomaly and the associated timeseries. f) Spectrum of the EOF timeseries. Standing eddies can be identified on the EOF pattern modes.

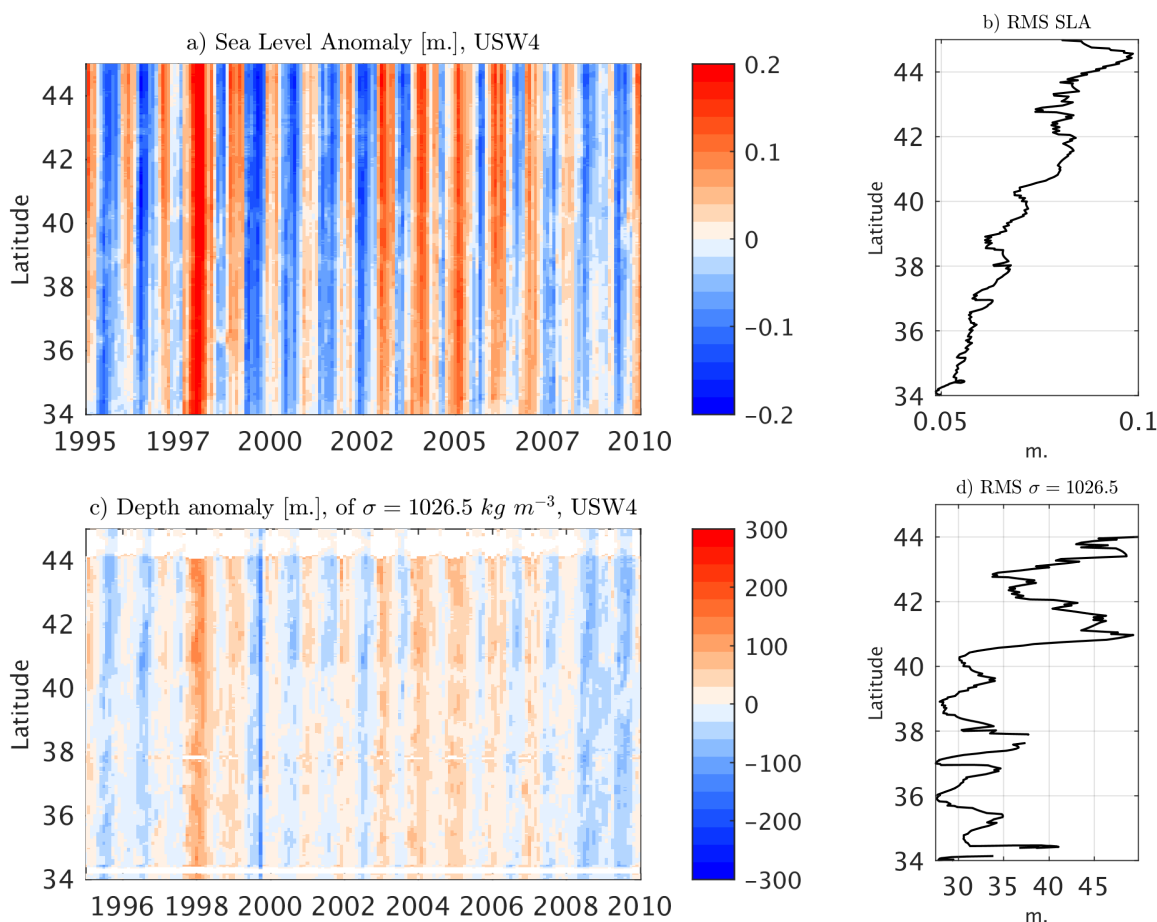


Figure 20: Hovmöller diagrams (latitude and time) for (a) SLA [m] and (c) depth of the  $\sigma_{\theta} = 25.6$  isopycnal surface [m] at a distance  $\approx 50$  km offshore. On the right of each plot is the corresponding RMS latitude profile with respect to the temporal variability.

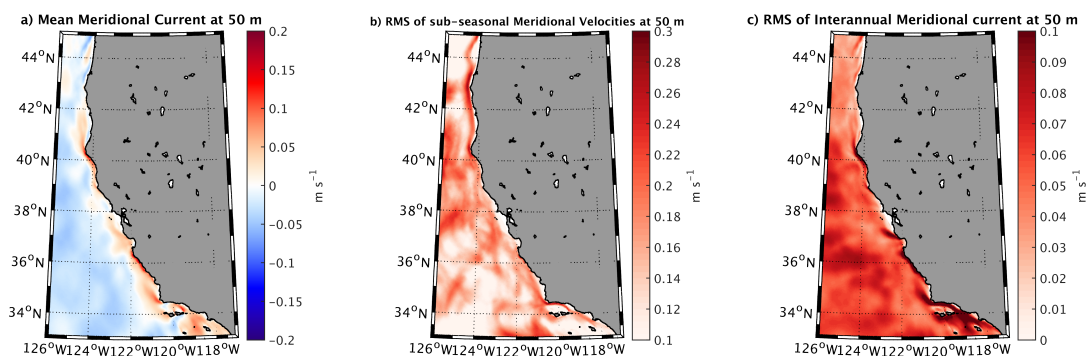


Figure 21: Meridional current at 50 m depth [ $\text{m s}^{-1}$ ] from USW4. (a) Long-term mean during the hindcast period, (b) subseasonal variability, (c) interannual variability. Note that the surface nearshore current is southward.



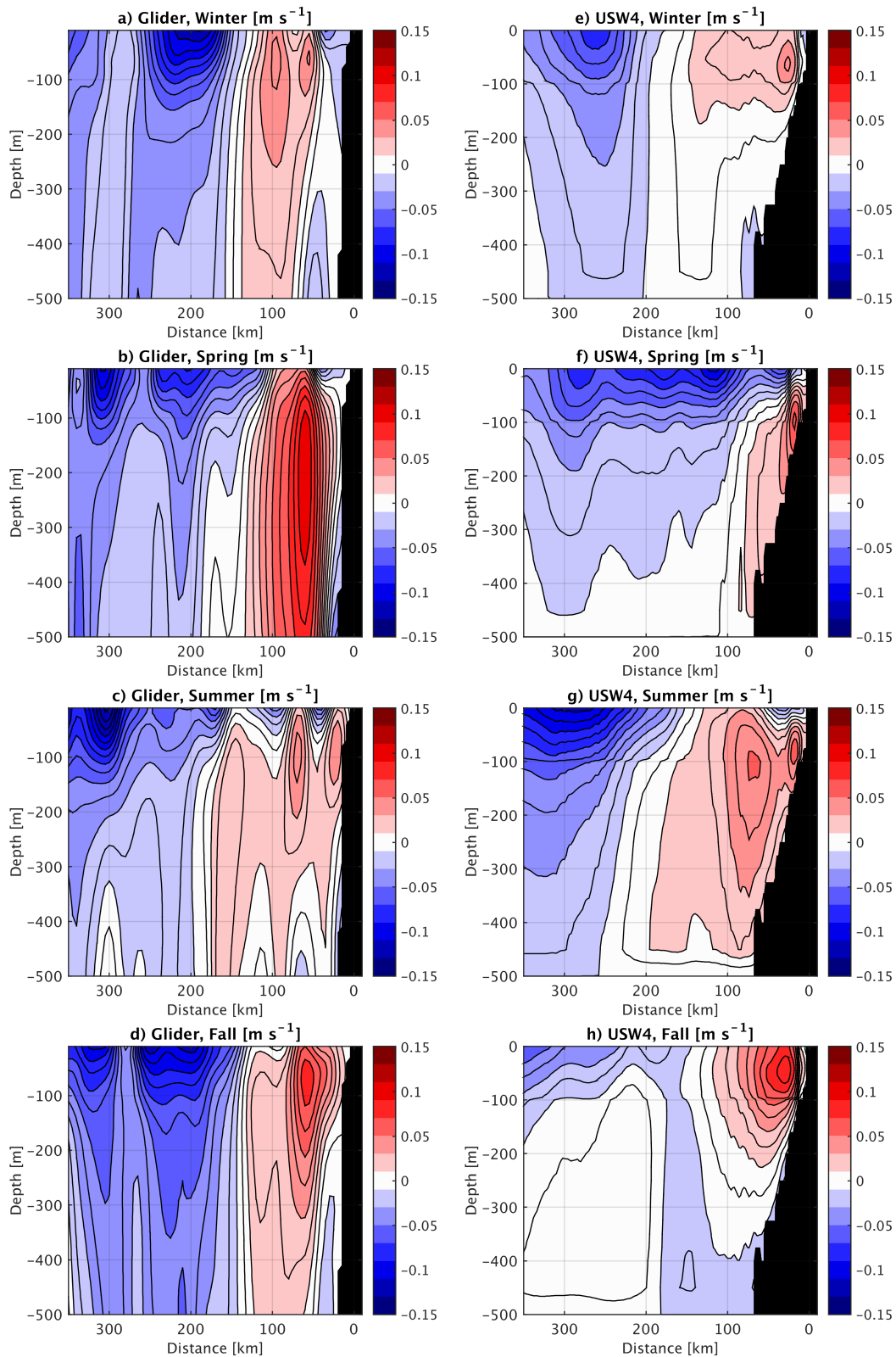


Figure 22: Seasonal means of the meridional geostrophic current [ $\text{m s}^{-1}$ ] along line 66.7 estimated from the gliders (abcd) and USW4 (e,f,g,h). Note the definition of the seasons differs from that in Rudnick et al. (2017).

## 617 5 Higher Frequency Variability

### 618 5.1 Mesoscale Activity

619 The geostrophic surface EKE is estimated over the period 1995-2010 from AVISO (Ducet et al.,  
620 2000) and from USW4 (Fig. 23ab). For the model the EKE is computed using low-pass fil-  
621 tered geostrophic velocities with a Gaussian spatial filter with a 36-km half-width and a temporal  
622 smoothing of 7 days as an approximation to AVISO's resolution (Chelton and Schlax, 2003) (More  
623 precise comparisons could be made using simulation fields processed in the same way as altimet-  
624 ric measurements.). The relevant EKE conversion rates (Sec. 2c) in USW4 are also evaluated over  
625 the period 1995-2010 (Fig. 23c-e). Consistent with Strub and James (2000); Marchesiello et al.  
626 (2003) and Renault et al. (2016d), the mean CCS circulation is unstable and generates mesoscale  
627 eddies primarily by baroclinic instability (Strub and James, 2000; Marchesiello et al., 2003) while  
628 the  $K_m K_e$  conversion is a secondary term. In both observations and USW4 The simulated EK has  
629 its largest values a couple of hundred km offshore and exhibits a wide decay zone further offshore  
630 (Fig. 23). This pattern is due to the combined influences of Ekman transport, eddy dispersion,  
631 and the eddy killing effect of the current feedback, with an overall similarity to AVISO and the  
632 literature (e.g., Capet et al. (2008a)). The overall EKE amplitude is about right in USW4, but there  
633 are some biases in the USW4 spatial pattern, mainly that the EKE is too large in the near-coastal  
634 region, and the EKE is too large in the Southern California Bight, which seems most likely due  
635 to errors in the atmospheric forcing. Part of the discrepancy may be due to model bias, e.g., the  
636 current feedback induces a dampening of the mesoscale activity, but the wind response induces a  
637 partial re-energization. In USW4, following Renault et al. (2016d), the  $s_w$  coefficient is used to  
638 mimic the wind response to the current feedback. Figure 23f depicts the temporal evolution of the  
639 surface EKE domain average from USW4 and from the coupled and uncoupled simulations used  
640 in Renault et al. (2016d). USW4 has comparable level of energy as the coupled simulation that in-  
641 cludes the wind response to current feedback (EXP3), indicating the parameterization used partly  
642 allows to re-energize the mesoscale currents.  $s_w$  is taken as spatially and temporally constant,  
643 which could induce, for example, a re-energization that is too strong in the nearshore region.

644 Figure 24a depicts a cross-section of the EKE averaged between 35°N and 40°N. It reveals that  
645 the EKE is large from 200 m depth to the surface and from the coast to  $\approx 800$  km offshore. The  
646 EKE is characterized by a peak of  $200 \text{ cm}^2 \text{ s}^{-2}$  at 200 km from the coast and at the surface and  
647 slowly decays in the offshore direction while rapidly decreasing at depth with values of less than  
648  $75 \text{ cm}^2 \text{ s}^{-2}$ .  $P_e K_e$  associated vertical structure is shown in Fig. 23b. It reveals that most of the  
649 positive values of  $P_e K_e$  occur in the first 50 m depth, from the coast to 100 km offshore. Finally,  
650 a mean cross-shore profile between 30°N and 45°N is estimated for  $P_e K_e$ ,  $K_m K_e$ , and  $F_e K_{eg}$   
651 (Fig. 24c). The geostrophic eddy wind work  $F_e K_{eg}$  profile is also estimated using a QuikSCAT  
652 product (Bentamy and Fillon, 2012) and AVISO, but only over the available QuikSCAT period  
653 (2000-2009) (Due to the QuikSCAT and AVISO coastal accuracy issues, the  $F_e K_{eg}$  value over  
654 the first 50 km off the coast is not shown.). Consistent with the measurements,  $F_e K_{eg}$  is positive  
655 in the nearshore region and then becomes negative offshore, deflecting energy from the oceanic  
656 geostrophic eddy currents to the atmosphere and thus dampening the offshore eddies. USW4  
657 deflects slightly more energy offshore than the measurements (by 10% averaged over the offshore  
658 area). This could be due to estimation errors in the measurements, but it might also be due to  
659 biases in the atmospheric and oceanic simulations, with an overestimation of the EKE reservoir

660 (more energy to be deflected) and a biased estimation of  $s_w$  when estimating the surface stress.  
661 For example,  $s_w$  has a seasonal cycle (Renault et al., 2017, 2020) and depends on the atmospheric  
662 parameterization of the marine boundary layer (Renault et al., 2016d), and these dependencies are  
663 not included in these hindcast simulations. This may explain the overestimation of the offshore  
664 EKE and the overestimation of the eddy life in EXP3 shown by Renault et al. (2016d).

665 Figure 25a shows the seasonal cycle of the EKE as estimated from AVISO and from the USW4  
666 low-pass filtered geostrophic velocities with a Gaussian spatial filter with 36-km half-width (solid  
667 line) and 28-km-half-width (dashed line). Consistent with *e.g.*, Amores et al. (2018), the EKE  
668 estimated from AVISO (and from USW4 filtered data is likely to be underestimated by a factor  
669 of 2. However, USW4 realistically simulates the seasonal evolution of the EKE, with larger EKE  
670 values in summer and fall. The EKE seasonal cycle is mainly driven the seasonal variability of  
671  $P_e K_e$  (not shown). Finally, Fig. 23f reveals a large interannual variability of the EKE, especially  
672 in the South and Central boxes.

## 673 5.2 High-Frequency Wind Forcing

674 In this section the oceanic impact of the synoptic wind variability is assessed. In a bulk formula the  
675 surface stress has a quadratic dependence on the wind. As a result, time-varying winds contribute  
676 to the time-mean surface stress. It is well known that neglecting high-frequency winds can induce  
677 large errors in the surface stress estimate (Esbensen and Kushnir, 1981; Gulev, 1994; Wu et al.,  
678 2016). Recent studies show those errors can cause large biases in kinetic energy transfer between  
679 the atmosphere and ocean (Zhai et al., 2012; Zhai, 2013). In particular, Zhai et al. (2012) using  
680 a global oceanic model showed the mean wind work increases by 70% when using a 6-hourly  
681 wind update instead of a monthly one. A few previous studies assess the role of high-frequency  
682 atmospheric wind in determining the oceanic circulation. They find that it can lead to an increase  
683 by about 50% in both the mean wind work and the EKE, as well as a strengthening of the wind-  
684 driven subtropical gyre by about 10-15% (Holdsworth and Myers, 2015; Wu et al., 2016; Condron  
685 and Renfrew, 2013).

686 To determine the importance of high-frequency wind forcing in the CCS, three additional ex-  
687 periments have been carried out for the period 1995-1999 using a 6-hourly-averaged wind forcing  
688 (6H), and a daily-averaged wind forcing (1D). The mean alongshore surface stress and the EKE  
689 averaged along a coastal band 100 km wide are illustrated in Fig. 26 and summarized in Table 1.  
690 From a 1-hourly (*i.e.*, USW4) to 6-hourly wind forcing, the surface stress is only slightly impacted  
691 by 3%. When estimating the stress using 1D, it is slightly underestimated by 6%. This is, for  
692 example, the error made by the QuikSCAT daily products for the CCS. From USW4 to 6H and 1D,  
693 the surface current is slightly reduced by 7% and 13%, respectively.

694 The error in wind work has important consequences on the mesoscale activity. As shown in  
695 Fig. 23, the main sources of EKE are the baroclinic energy conversion and the eddy wind work.  
696 By reducing the mean input of energy  $F_m K_{mg}$  and the shear of the alongshore current, the absence  
697 of the high frequency component of the wind leads to a reduction of the baroclinic conversion  
698 rate,  $P_e K_e$ , by 1%, and 4% from USW4 when the wind is temporally smoothed to 6H and 1D,  
699 respectively. The positive coastal  $F_e K_{eg}$  is also reduced by 3%, 8%, and 70% (not shown). As a  
700 result the mean alongshore EKE is reduced by 4%, 7%, and 62% in wind-smoothings of 6H and  
701 1D compared to USW4 (Table 1). A simulation forced by a daily atmospheric forcing likely under-  
702 estimates the EKE by 7%. An oceanic model could also be forced by a monthly stress (estimated

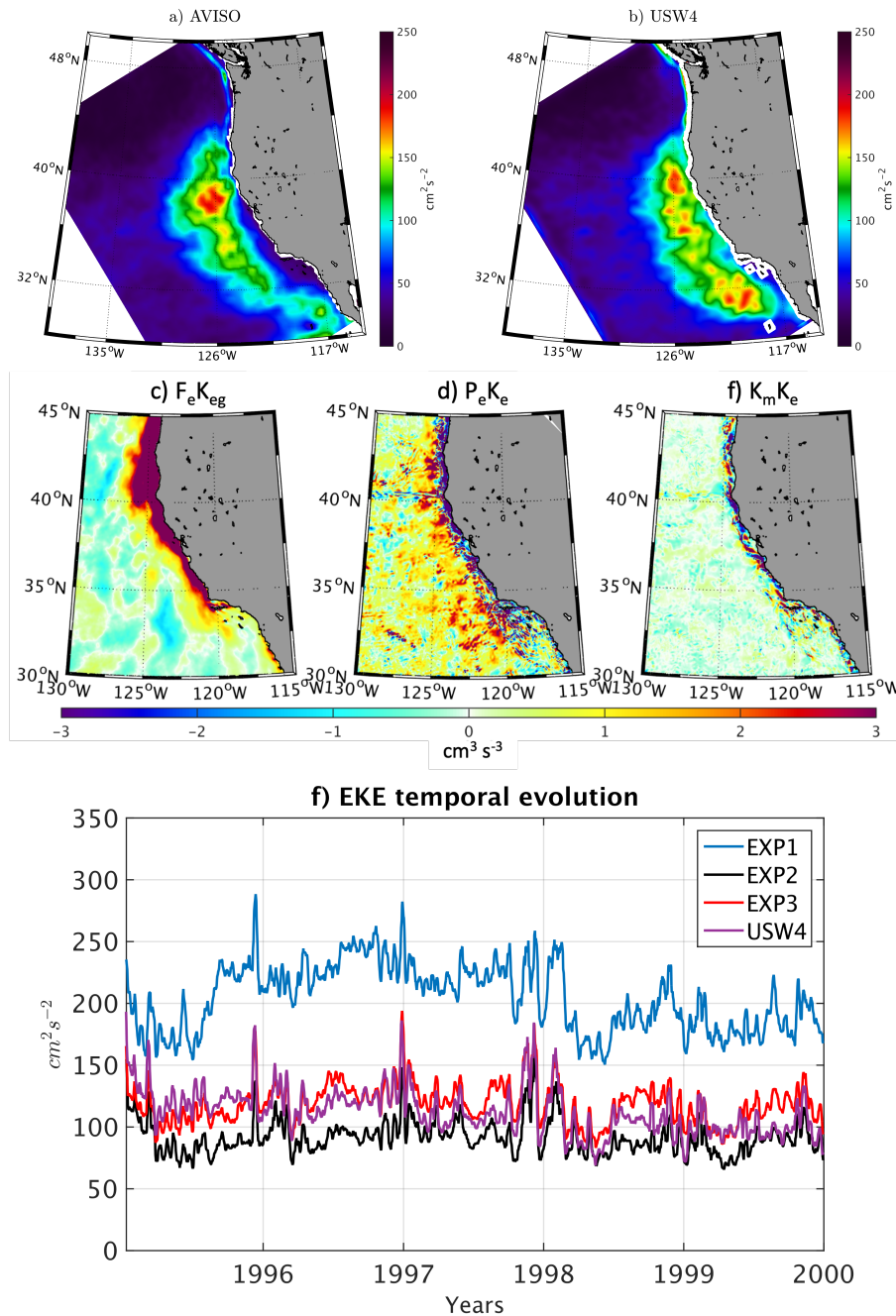


Figure 23: Mean geostrophic surface EKE [ $\text{cm}^2 \text{s}^{-2}$ ] estimated from (a) AVISO and (b) USW4. The EKE of USW4 is computed using low-pass filtered geostrophic velocities (a Gaussian spatial filter with 36-km-half-width) as an approximate match to AVISO’s spatial resolution. (c)-(e) Geostrophic eddy wind work ( $F_e K_{eg}$ ), baroclinic conversion ( $P_e K_e$ ), and barotropic conversion from the mean flow ( $K_m K_e$ ) [ $\text{cm}^3 \text{s}^{-3}$ ] in USW4 over the period 1995-2010. (f) Temporal evolution of the surface total EKE averaged over the whole domain from USW4 and the simulations from Renault et al. (2016d): EXP1 is a coupled simulation without current feedback, EXP2 is a forced simulation that uses the relative wind to the oceanic motions but without a parameterization of the wind response, EXP3 is a coupled simulation with current feedback.

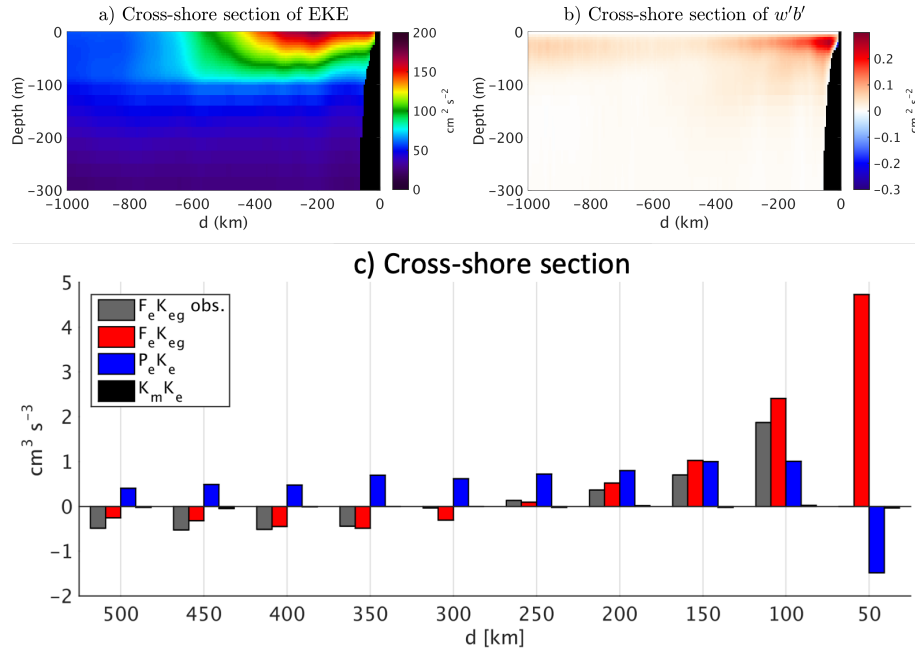


Figure 24: (a)-(b) Cross-shore sections of the mean EKE and  $P_e K_e$  in USW4 averaged between  $35^\circ\text{N}$   $40^\circ\text{N}$ . (c) Cross-shore bins of  $F_e K_{eg}$ ,  $P_e K_e$ , and  $K_m K_e$  averaged over 50 km intervals between  $30^\circ\text{N}$  and  $45^\circ\text{N}$ . The geostrophic eddy wind work ( $F_e K_{eg}$ ) is estimated from USW4 (red) and from measurements (QuikSCAT and AVISO; gray). For the first 50 km off the coast the observational estimate is not shown because of coastal contamination. The baroclinic conversion is the main energy generation term. The eddy wind work  $F_e K_{eg}$  is positive nearshore and deflects energy from the ocean to the atmosphere offshore while dampening the mesoscale activity

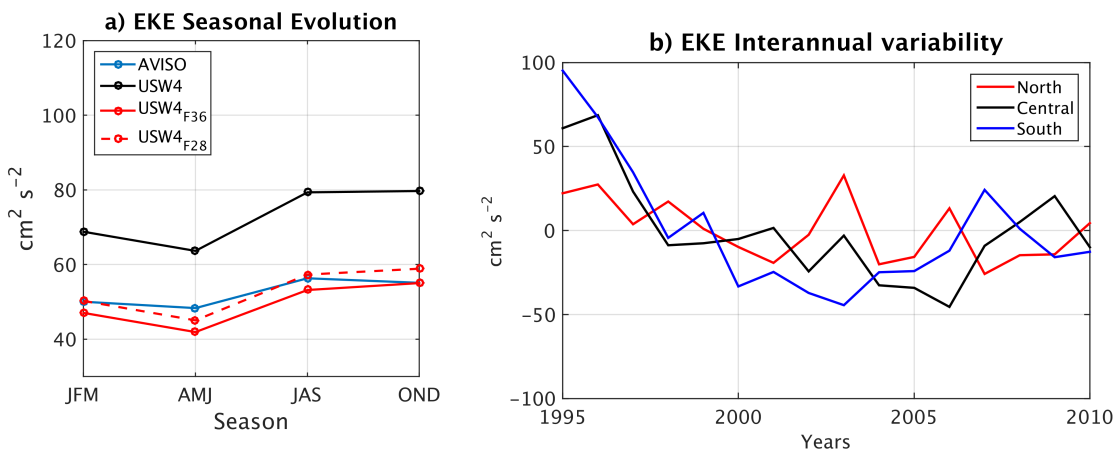


Figure 25: (a) Seasonal evolution of the mean EKE averaged over the whole domain as estimated from the measurements and from the USW4 original filtered geostrophic velocity (based on two different filter half-widths: 36 km (solid line) and 28 km (dashed line)). (b) Interannual variability of the EKE averaged over the boxes indicated on Fig. 1. The current feedback to the atmosphere dampens the eddies and thus allows the simulation to have a realistic EKE level, albeit with not quite the same spatial pattern around the Southern California Bight. Pointwise sampling errors are up to  $\approx 5 \text{ cm}^2 \text{ s}^{-2}$ , estimated using a bootstrap method (Efron and Tibshirani, 1985): the mean EKE is computed 100,000 times using random samples from the distribution, and the uncertainty is then defined as  $\pm$  the standard deviation of these values.

703 from averaging the hourly stress and thus accumulating the nonlinear effect of synoptic wind);  
 704 such a simulation would neglect the negative wind work  $F_e K_{eg}$ , leading to an overestimation of  
 705 the EKE by 60% (Fig. 23).

706 Along a coastline, the cross-shore Ekman transport is proportional to the surface stress,  $T_E =$   
 707  $\tau_{alongshore} / \rho f$ . Thus, the underestimation of the stress by neglecting the high frequency wind leads  
 708 to a similar underestimation of the transport. The mean Ekman transport is reduced by 2%, and  
 709 6% from USW4 to more smoothly varying winds with 6H and 1D averages, respectively.

710 A striking difference between USW4 and the other simulations is the level of activity in the  
 711 inertial currents (Fig. 27). By neglecting the hourly stress the inertial currents are much weaker  
 712 with 1D wind forcing. This is confirmed by the spectrum of the alongshore current: USW4 has  
 713 a large peak of energy around 18 hours that is not reproduced by 1D. This is consistent with *e.g.*,  
 714 Zhai (2017), who found that almost all the energy flux from the wind to near-inertial motions in  
 715 the mid-latitude North Pacific and Atlantic are due to a mesoscale atmospheric system with scales  
 716 less than 1000 km; a high frequency forcing is deemed to be required to represent them. Finally,  
 717 the lack of inertial currents in 6H and 1D can be seen through the eddy ageostrophic wind work  
 718 ( $F_e K_{ea}$ ) estimated over the 5 years of simulation from USW4, 6H, and 1D. The  $F_e K_{ea}$  work is  
 719 underestimated by 20% by neglecting the sub-6-hourly wind variability and underestimated by  
 720 about 70% without the sub-daily variability. Consistent with Zhai (2017) and D'Asaro (1985,  
 721 1995), the occurrence of winter storms induces larger inertial currents and  $F_e K_{ea}$  than during the  
 722 summer.

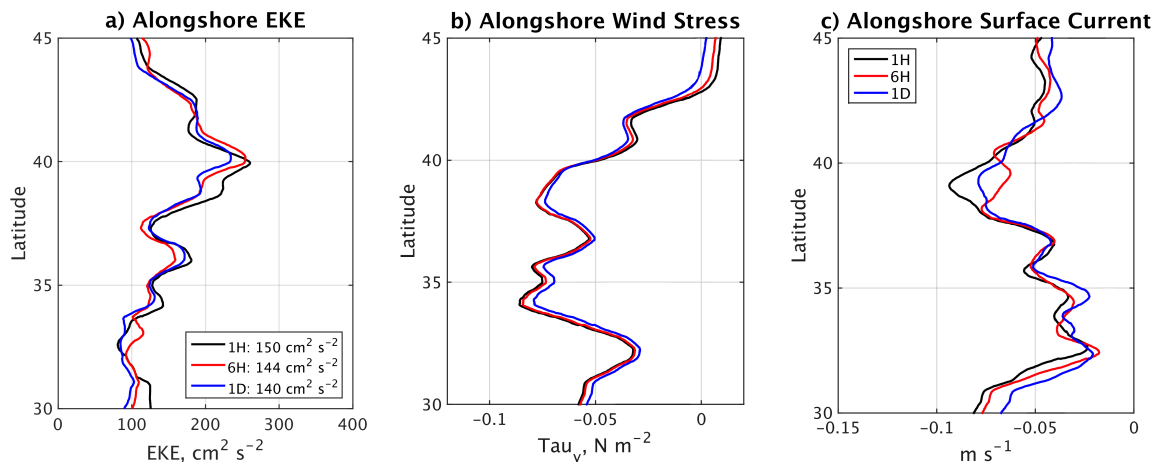


Figure 26: Influence of high-frequency wind forcing on the oceanic surface currents and the surface stress. (a) Annual mean alongshore EKE estimated over a coastal band of 100 km width in USW4 with 1 hr (1H), 6 hr (6H), and 1 day (1D) wind update intervals over the period 1995-1999. (b) Same as (a) but for the mean alongshore wind stress. (c) Same as (a) but for the mean alongshore surface current.

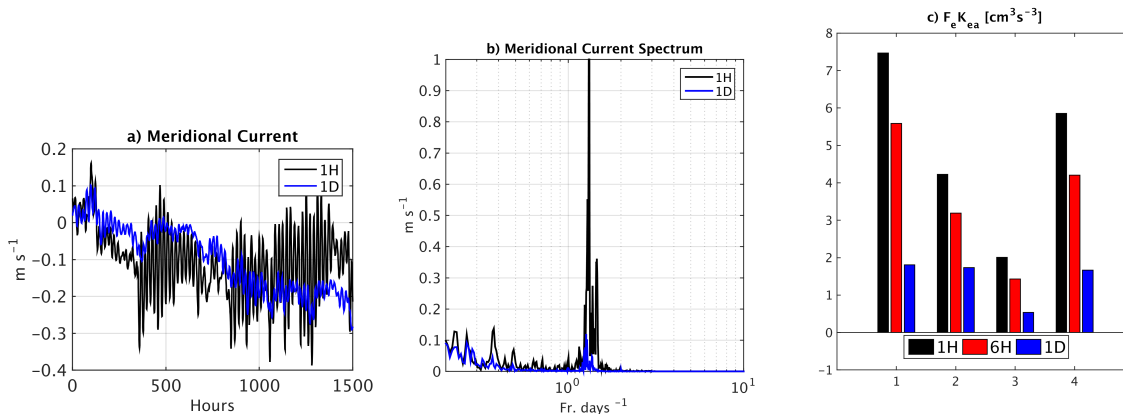


Figure 27: Surface current responses to hourly and daily wind forcing. (a) Hourly time series of the surface meridional current at ( $36^\circ\text{N}$ ,  $122^\circ\text{W}$ ) (central California coast). The black and blue lines represent the simulation forced by the hourly- (1H, as in the USW4 simulation) and the daily-updated wind (1D). (b) The temporal spectrum of the surface meridional current from the 1H (black) and 1D (blue) simulations at ( $36^\circ\text{N}$ ,  $122^\circ\text{W}$ ). (c) Mean ageostrophic eddy wind work ( $F_e K_{ea}$ ) ( $\text{cm}^3 \text{s}^{-3}$ ) averaged between  $30^\circ\text{N}$  and  $45^\circ\text{N}$  and over a 500 km cross-shore distance from the coast in the 1H, 6H, and 1D simulations. As expected, the high-frequency wind forcing enhances the ageostrophic wind work and inertial currents.

Table 1: Mean EKE, surface stress, and alongshore surface current averaged along a coastal band 100 km wide for USW4 (1H forcing) and the three additional experiments using a 6-hourly-averaged wind forcing (6H), and a daily-averaged wind forcing (1D).

	$EKE_{along}$ [ $\text{cm}^2 \text{s}^{-2}$ ]	$\tau_{along}$ [ $\text{N m}^{-1}$ ]	$V_{along}$ [ $\text{cm s}^{-1}$ ]
1H	150	-0.049	-5.3
6H	144	-0.049	-4.9
1D	140	-0.047	-4.6

## 6 Discussion

In this study regional atmospheric and oceanic model simulations are made for a 16-year hindcast period from 1995 to 2010. The simulations are evaluated against satellite and *in situ* measurements with an emphasis on the seasonal cycle and the mean and mesoscale circulations of the California Current System (CCS).

We evaluate the atmospheric forcing simulated by WRF and find, in general, a good agreement between the simulations and the measurements of the cloud cover, heat fluxes, and surface stress with modest discrepancies that are some combination of estimation errors and model biases. In particular, we show the ability of the atmospheric model to represent realistically the stratocumulus cloud deck in the northeastern Pacific. Then, by comparing the oceanic simulation to available measurements and previous modeling studies, we demonstrate the consistency of the simulations in representing the mean circulation and the seasonal and mesoscale variability of the CCS. Our results illustrate the benefits of using both oceanic and atmospheric regional simulations to simulate the seasonal variability of an eastern boundary upwelling system, at least in part because of the excessively coarse resolution in global models. Although some aspects of the interannual variability have been included in this study, more could be examined about low-frequency variability in the CCS.

The wind drop-off characteristics of a similar atmospheric simulation have been validated by Renault et al. (2016b). The simulation reported in this paper presents a good agreement with the measurements. These oceanic validations are also an indirect validation of the wind profiles simulated by WRF. An alternative simulation has been carried out using the CFSR reanalysis (Saha et al., 2010). Due to a poor representation of the wind drop-off, this simulation was characterized by an unrealistic poleward surface current and a poor representation of the mesoscale activity. The coarse resolution of CFSR (or other similar reanalysis) prevents using such a product to force this particular upwelling region and should not be used to investigate processes or trends, at least in the CCS.

Although not discussed here in any detail, the oceanic simulation is forced using various lateral open-ocean boundary forcing fields, such as Mercator or SODA. Differences in the lateral conditions can lead to significant changes in mean temperature and salinity (up to  $0.5^\circ\text{C}$  in SST and  $0.5$  PSU in  $S$ ). Probably they are the primary cause for the the salinity biases present in the USW4 simulation, which are perhaps the largest inaccuracy of the simulation. We finally chose to force the simulations using Mercator with an additional mean monthly state correction toward the measurements from the World Ocean Database over the period 1995-2004. Nevertheless, uncer-



756 tainty in open-ocean boundary conditions of the gyre-scale currents, density, and other water-mass  
757 properties do limit the possible accuracy of these quantities in a regional simulation.

758 An important contribution of this paper is our use over a long time period (1995-2010) of the  
759 parameterization of the wind and stress response to the current feedback suggested by Renault et al.  
760 (2016d) for the U. S. West Coast. Long-term comparisons with satellite measurements show real-  
761 istic simulation results for the EKE and the energy transfer between the ocean and atmosphere with  
762 this feedback — and falsely large EKE values without it — in high-resolution models. Oceanic  
763 models, if uncoupled, should take into account the current feedback and by including a parame-  
764 terization of the wind response such as Eq.(3) for a realistic kinetic energy transfer between the  
765 atmosphere and the ocean, and thus for a realistic level of mesoscale activity and mean circulation.

766 Finally, we discussed the importance of using a high-frequency wind forcing to represent the  
767 mean features of the CCS. In particular, consistent with Wu et al. (2016), we show the presence  
768 of high frequency wind prevents the use of monthly wind to force an oceanic model of the CCS.  
769 It leads to large errors in the mean stress and wind work inputs to the ocean, and, thus, to a poor  
770 representation of the mean and mesoscale currents. For the CCS we show that a 6-hourly wind  
771 forcing realistically represents the mean surface stress and the mean and mesoscale geostrophic  
772 currents. A daily wind forcing, such as QuikSCAT (commonly used to force an oceanic model),  
773 leads to an underestimation of the EKE by 7% for the CCS. However, a 1-hourly wind forcing is  
774 necessary for proper representation of the inertial currents.

775 In summary, we show the benefit of using both oceanic and atmospheric simulations for rep-  
776 resenting the mean physical state of the CCS. The atmospheric model is characterized by several  
777 biases such as too dry a lower atmosphere, too few clouds nearshore (although it realistically repre-  
778 sents the stratocumulus cloud deck in the north of the domain), too much precipitation, and slightly  
779 too low a surface stress. As a response, the oceanic model has too large a surface salinity, too cold  
780 a SST, and too deep a MLD. Some other oceanic biases are controlled by the open-boundary condi-  
781 tions, such as the too-cold 150 m depth temperature. Perfect model-measurement agreement is an  
782 impossible goal both because of sampling limitations in model and observational data and because  
783 there are too many model design options and parameterization choices to ever be fundamentally  
784 correct or precise (McWilliams, 2007). Nevertheless, the USW4 system presented here is in fairly  
785 good overall agreement with the measurements that exist, and it has no glaring failures with respect  
786 to its primary behaviors. It thus provides a reliable physical foundation for assessing biogeochem-  
787 ical cycles and climate changes in the CCS (Deutsch et al., 2021a; Howard et al., 2020a, 2021;  
788 Kessouri et al., 2021b).

789 **Codes and Simulation Data** The physical and biogeochemical codes used for our simulations  
790 are at <https://github.com/UCLA-ROMS/Code>. The simulation model output archive data can be  
791 made available by an email request to the Corresponding Author.

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799 This study has been conducted using E.U. Copernicus Marine Service Information. MODIS level  
800 2 data were downloaded from the NASA Web site (available at <http://ladsweb.nascom.nasa.gov>).  
801 GPCP data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their  
802 web site at <https://www.esrl.noaa.gov/psd/>. Gliders data can be found at <https://spraydata.ucsd.edu/climCUGN/>

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