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High-resolution modeling of nearshore circulation in an upwelling ecosystem: resolving sub-mesoscale variability

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Abstract

While numerical modeling has proven successful at reproducing the physical, chemical, and biological processes associated with circulation over regional and mesoscales, challenges remain for modeling the coastal zone at sufficiently high resolution to resolve coastal flows and their interactions with larger-scale processes. Computational costs associated with grid resolution and limited observations for model validation are important concerns. This study evaluates a climatological hydrodynamic model off central Chile (28-38° S), which has three nested domains to reproduce both the regional and nearshore circulation patterns. To determine whether submesoscale processes propagate up to larger scales, results from the nested High-Resolution Model (HRM2, 32-34°S and 0.87 km resolution) and its parent (HRM0, 28-36°S and 7.4 km) were compared with those from a Low-resolution Model (LRM0) configured with the exact resolution as the HRM0. While both HRM0 and LRM0 reproduced well the climatological Sea Surface Temperature (SST) and Salinity (SSS), as well as the general spatial patterns of kinetic energy, the annual transport of the Peru-Chile Undercurrent (PCUC) increased in the nested HRM0 (-0.7 to -1.3 Sv) compared to the LRM0 (-0.58 to -0.95 Sv). Moreover, the Coastal Chilean Current (CCC) became up 2.16 faster in HRM2 than in LRM0, with a sharp shoreward decline in the maximum speed (down to

11 cm/s) nearshore, defining in this manner a coastal boundary layer (CBL), which was not reproduced by the LRM0. Also, the upwelling process in the HRMO was intensified by topographic features and was spatially consistent with ageostrophic circulation patterns. As expected, the most remarkable differences between model resolutions emerged nearshore, especially within the CBL, highlighting the importance of topographic effects on coastal circulation. Thus, although the low-resolution model can reproduce patterns of circulation in the coastal ocean fairly well, resolving submesoscale variability as in the HRMO imposes moderate to large changes in transport, current velocities, vorticity and topographic realism of the upwelling process. The extent to which variation in these processes affects biochemical processes, particle transports, and or larval dispersal must still be investigated.

15 *Keywords:* High-resolution modeling, nesting, coastal boundary layer

16 **Introduction**

17 Characterizing circulation patterns over the continental inner-shelf represents an enor-
18 mous challenge because of the complexity in the underlying physical processes that occur
19 over multiple spatial and temporal scales, largely determined by the interaction between
20 winds, fluids and the rigid coastal topography (Largier, 2003, Kirincich & Barth, 2009,
21 Lentz & Fewings, 2012, Brink, 2016). The challenge to provide useful representations of
22 the coastal circulation becomes even more difficult due to the low frequency of observations
23 that can be gathered over spatial domains and periods of time large and long enough to
24 encompass most of the processes in which we are interested, and for which we would like
25 to build observational and forecasting models. Coastal circulation processes modulate the
26 biological productivity that allows the existence of most pelagic and benthic communities
27 and ecosystems (Connolly et al., 2010, Pitcher et al., 2014, Largier, 2020), the connectivity
28 of populations via dispersal of propagules, which determines persistence of populations and
29 defines local biodiversity (Cowen & Sponaugle, 2009, White et al., 2019), and, in general,

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30 the carrying capacity and buffering of all human activities on the coastal ocean. Because
31 of these inherently limited observational capabilities of the coastal marine environment, hy-
32 drodynamic three-dimensional modeling continues to be instrumental in making advances
33 in our understanding of coastal circulation in different regions of the world, providing a tool
34 to characterize this complex environment that otherwise would be impossible to explore
35 through observational models alone (Penven et al., 2001, Mitarai et al., 2008, McWilliams,
36 2016). Multiple modeling approaches, suitable for the coastal ocean, have been developed.
37 They differ in their numerical methods, horizontal and vertical mesh discretization, the
38 manner in which they resolve topography, and the boundary conditions that can be imple-
39 mented. Additionally, computational advances have made possible the increase in spatial
40 resolution of these models, some of which are able to reproduce important circulation aspects
41 governed by both geostrophic and ageostrophic processes (McWilliams, 2003, Capet et al.,
42 2008, Soufflet et al., 2016). As with all models, a trade-off exists between model scope or
43 generality, resolution, and realism (eg Levins 1968). The comparatively recent development
44 of computationally-efficient nested numerical models that can resolve complex coastal circu-
45 lation at multiple scales, attempts to strike a practical balance in this three-prong trade-off,
46 by combining high resolution of processes within section of a more general model. Here
47 we develop a nested model for coastal circulation of a section of the Humboldt Upwelling
48 Ecosystem (HUE), assess its consistency with observations, and demonstrate that the in-
49 crease in resolution markedly improves the representation of the near-shore circulation with
50 respect to a coarser scale model.

51 Mesoscale oceanographic processes have remained the focus of hydrodynamic studies in
52 coastal regions of the world, due to both the importance of these processes in hydrographic
53 variability in coastal ecosystems, and to the practical constraints upon the resolution needed
54 to simulate the submesoscale. In the HUE, such models have provided a deeper understand-
55 ing of the upwelling circulation, topographic intensification, variability of Ekman transport,
56 mesoscale eddy activity, and the effects of riverine inputs, among others (Leth et al., 2004,
57 Dewitte et al., 2008, Aiken et al., 2008 Aguirre et al., 2012, Vergara et al., 2016). The
58 level of realism achieved in these mesoscale models has been sufficient to explore, for in-

59 stance, the broad consequences for propagule dispersal (Aiken et al., 2007, Parada et al.,
60 2012, Medel et al., 2018, Ospina-Alvarez et al., 2018) or to project effects of climate change
61 (Aiken et al., 2011, Oerder et al., 2015, Lima et al 2021). But the resolution used in these
62 models - with scales on the order of multiple kilometers - limits their applicability to resolve
63 near-shore processes, where strong gradients commonly occur over these same scales within
64 the Coastal Boundary Layer (CBL). A significant body of observational evidence challenges
65 the assumption that the unresolved processes have a negligible impact on the mesoscale (e.g.
66 Nickols et al., 2012 and 2013). The processes that define the dynamics and spatio-temporal
67 scales of variability in coastal environments commonly occur over scales of kilometers and
68 less - often called the submesoscale (Nickols et al., 2013, McWilliams, 2016). As such, these
69 processes can emerge and be correctly represented only by sufficiently increasing model res-
70 olution (Liu & Levine, 2016, Sun et al., 2020). Certainly, the presence of these submesoscale
71 structures and the increase in vertical speed associated with different energy transfer mech-
72 anisms (Capet et al., 2008a, Mahadevan, 2016) have exclusively been captured by increasing
73 the resolution into observations and models. These structures might affect the cross- and
74 alongshore transport, as well as, the final fate of particles and pollutants in the coastal
75 zone (Harrison et al., 2013, Dauhajre et al., 2017). Indeed, the presence of a nearshore
76 CBL, where an exponential decay of currents speed and a raised eddy diffusivity occurs,
77 can greatly influence the retention and dispersion capacities of nearshore matter (Nickols
78 et al., 2012, Nickols et al., 2015, Morgan, 2014). But beyond these local observations that
79 point to increased importance of submesoscale processes, it is still unclear to what extent
80 they influence the mesoscale circulation. A case in point that has received much atten-
81 tion is the determinant of dispersal and transport between coastal sub-populations of the
82 propagules produced on the shore by adults of largely sedentary species. It has been argued
83 that coastal, small-scale processes may have primacy in determining patterns of dispersal,
84 connectivity and eventual onshore recruitment (Morgan et al., 2009, Nickols et al., 2015,
85 Harrison et al., 2013), over the mesoscale variations typically represented in larval transport
86 simulations, which could potentially misrepresent these important aspects of population
87 and community persistence and productivity (Dauhajre et al., Bani et al., 2021). Therefore,

88 to realistically simulate dispersal and population connectivity, processes we may require a
89 significant increase in spatial resolution, so that finer coastal topographic complexity and
90 circulation pattern are better captured in the models, but at the same time being able to
91 resolve a domain large enough to capture mesoscale variability. In other words, to assess the
92 importance of resolving the sub mesoscale, numerical models representing the inner shelf
93 circulation should reproduce the mesoscale circulation while still capturing the CBL, whose
94 potential interaction has not commonly been considered in shelf circulation studies along
95 HUE or other upwelling ecosystems.

96 The eastern margin of the South Pacific Ocean is characterized by a narrow continental
97 shelf where the circulation is mainly forced through seasonal wind variability, modified by
98 topography and the coastline morphology (Thiel et al., 2007, Aiken et al., 2008 Strub et al.,
99 2013). Prevailing equatorward winds drive offshore Ekman transport and the coastal up-
100 welling of subsurface waters, that in turn feeds the formation of energetic mesoscale eddies,
101 meanders and coastal jets (Mesias et al., 2003, Rutllant et al., 2004, Hormazabal et al., 2004,
102 Capet et al., 2004, Chaigneau & Pizarro, 2005). As these processes largely control the rates
103 of continental shelf fertilization and organic matter exportation, understanding their dynam-
104 ics allows us to explore effects on ecological systems (Letelier et al., 2009, Correa-Ramirez
105 et al., 2007, Gruber et al., 2011, Correa-Ramirez et al., 2012). Here we investigate the
106 importance of high spatio-temporal resolution for correctly reproducing features at subme-
107 soscale, such as the CBL, and whether this influences the mesoscale circulation as simulated
108 in lower resolution models, through analysis of a nested model with variable spatial resolu-
109 tion along the coast off central Chile. Our study area is centered within one of the ocean
110 regions with highest biological productivity in the world, supporting a rich and diverse arti-
111 sanal fishing activity employing several thousand people (Thiel et al., 2007, Castilla, 2010,).
112 The outer and middle shelf circulation has been relatively well characterized through obser-
113 vations (Strub et al., 1995, Shaffer et al., 2004, Hormazabal et al., 2004) and simulations
114 (Leth & Shaffer, 2001, Aguirre et al., 2014). Alongshore, the coastal circulation is strongly
115 controlled by the seasonal equatorward wind stress, which modulates the variability of the
116 coastal upwelling (Sobarzo et al., 2007, Aguirre et al., 2012, Bravo et al., 2016, Astudillo

117 et al., 2019), the equatorward Chilean Coastal Current (CCC) and the poleward Peru-Chile
118 Undercurrent (PCUC) (Strub et al., 1995, Vergara et al., 2016, Pizarro-Koch et al., 2019).
119 The CCC interaction with the coastline geometry and the bottom topography have been
120 an issue to understand some mechanisms that affect the upwelling intensification and eddy
121 generation (Leth & Middleton, 2004, Aiken et al., 2008). As is the case generally for up-
122 welling ecosystems, the response of the HCU to long-term climate change remains uncertain
123 (Sydeman et al. 2014, Schneider et al., 2017, Aguirre et al., 2018, 2021 Weidberg et al.,
124 2020, Gibson et al., 2007, Chavez & Messié, 2009), which makes it particularly important
125 and urgent to improve our understanding of inner shelf dynamics.

126 **Methods**

127 *0.1. Hydrodynamic model*

128 The nearshore circulation over the inner-shelf of central Chile, resolving both coarse
129 and high resolution scales, was simulated using the UCLA Regional Ocean Modelling Sys-
130 tems (ROMS, Shchepetkin & McWilliams, 2005) coupled to the AGRIF package for nest-
131 ing grids (Adaptive Grid Refinement in Fortran, Debreu et al., 2012) currently known as
132 <https://www.croco-ocean.org/> CROCO (Coastal and Regional Ocean Community model).

133 The model solves the primitive equations under the Boussinesq and hydrostatic approx-
134 imations, using an explicit 3D ocean free surface scheme, sigma-type vertical coordinates,
135 and a horizontal orthogonal curvilinear system. Two simulations were run over the same
136 domain, covering the region 28°-38°S and 70°-82°W, but with different configurations. The
137 first simulation used a single grid with 1/12° horizontal resolution (7.74 km) and 40 vertical
138 levels in σ -coordinates, and is referred to as the Low-Resolution Model (LRM0). The sec-
139 ond simulation used the same coarse grid and vertical levels in σ -coordinates, and was used
140 as the parent (HRM0) of two nested regions. The nested grids were configured with 40 σ -
141 coordinates levels and horizontal resolutions of 2.6 km (HRM1) and 0.87 km (HRM2), whose
142 extensions are shown in the Table1 and Figure 4b. The solutions for each of these nested
143 grids are referred to, respectively, as the parent High-Resolution Model (HRM0), the inter-
144 mediate High-Resolution Model (HRM1), and the child High-Resolution Model (HRM2).

145 Communication between the domains in the nested configuration was established through
 146 the 2-way nesting method implemented in the AGRIF package (Adaptive Grid Refinement
 147 in Fortran, Debreu et al., 2012). This methodology allows solving the interaction between
 148 mesoscale and submesoscale processes through a viscosity and diffusion increment as the
 149 grid resolution increases.

150

151 0.1.1. Nesting Configuration

152 The Global Relief Model (doi: 10.7289 / V5C8276METOPO1) with a resolution of 1 arc-
 153 minute (~ 1.85 km) provided the topography for LRM0 and HRM0, while the General Bathy-
 154 metric Chart of the Oceans (https://www.gebco.net/data_and_products/gridded_bathymetry_data/gebco30_sediment_resolution) (~ 0.92 km), was used for the two nested HRM. The bathymetry was in-
 155 terpolated between each nested grid through a bilinear method, and a smoothing algorithm
 156 applied on the logarithm of bathymetry to limit pressure gradient errors (Penven et al.,
 157 2005). The coordinates between grids were interpolated through a fourth-order polynomial,
 158 while the bathymetric connections were carried-out along a three cells border at the same
 159 level and then connected by linear interpolation. The fine grids were extrapolated on the
 160 HRM0 mask to increase the coastal resolution utilizing a Taylor expansion to minimize the
 161 error.
 162

Table 1: Nested domains into the hydrodynamic model. The table describes the name model abbreviation, the domains coordinates, resolutions and the runtime for each simulation.

Model Name	Domain	Resolution	Runtime
Abbreviation		(km)	(yr)
LRM0	28-38°S - 70-82°W	7.74	1 to 20
HRM0	28-38°S - 70-82°W	7.74	1 to 20
HRM1	28.4-35°S - 71.7-75°W	2.63	10 to 20
HRM2	32.4-34.4°S - 71.2-72.2°W	0.87	13 to 20

163 *0.1.2. Initial Conditions and Forcings*

The general nearshore circulation and its seasonal variability were reproduced by forcing the model with hydrographic and wind conditions derived from multi-annual climatologies. The LRM0 and HRM0 were forced with a monthly wind stress climatology from <http://agate.coas.oregonstate.edu/scow/SCOW> (Scatterometer Climatology of Ocean Winds, Risien & Chelton 2008) obtained between 2000-2008 at a resolution of 50 km. The monthly boundary conditions used a temperature and salinity climatology obtained between 1958-2008 from [http://apdrc.soest.hawaii.edu/datadoc/soda2.1.6.phpSODAv2.1.6\(SimpleOceanDataAsAtmosphereDataset\)](http://apdrc.soest.hawaii.edu/datadoc/soda2.1.6.phpSODAv2.1.6(SimpleOceanDataAsAtmosphereDataset)).

164 The LRM0 and the nested HRM0 simulations were run for 20 years. The HRM1 bound-
165 ary conditions were obtained from the tenth year of the HRM0 and then simulated coupled
166 to HRM0 until the 20th year. Finally, the HRM2 started with the hydrological conditions
167 from the 13th year of the nested HRM0+HRM1 (Table 1). All simulations ran in parallel
168 using 120 processors from the high-speed computing cluster "Leftraru" from the Chilean
169 National Laboratory for High-Performance Computing (NLHPC).

170 *0.1.3. Validation*

171 For validating the ability of the model to reproduce the magnitude of the main oceano-
172 graphic variables, an annual climatology was calculated between the fourth and tenth years of
173 LRM0. The seasonal averages corresponded to austral summer (December, January, Febru-
174 ary), autumn (March, April, May), winter (June, July, August), and spring (September,
175 October, November). Model realism was evaluated through comparison of climatological
176 values to satellite products, oceanographic measurements, and reanalysis (AVISO, CARS,
177 and HYCOM, respectively). The climatological eddy kinetic energy (EKE) and its spatial
178 structure were obtained from satellite altimetry data from AVISO (2000-2014) and com-
179 pared with the LRM0 and HRM0. The EKE was calculated from the surface geostrophic
180 anomaly and the sea surface height following the method of Hormazábal and collaborators
181 (2004), where the geostrophic velocities were computed as:

$$U_g = \frac{-g \delta\eta}{f \delta y} \quad (1)$$

$$V_g = \frac{g \delta\eta}{f \delta x} \quad (2)$$

182

183 where U_g and V_g are the anomaly of zonal and meridional geostrophic velocity, respec-
 184 tively, g is the gravity, f is the Coriolis parameter, and η is the sea surface height anomaly.
 185 Thus, the average of the EKE during the period analyzed is:

$$EKE = \frac{1}{2}[\langle(U_g - \langle U_g \rangle)^2\rangle + \langle(V_g - \langle V_g \rangle)^2\rangle] \quad (3)$$

186 where $\langle \rangle$ is the time averaged within each database. Additionally, we calculated the
 187 spatial correlation between the LRM0 sea level anomalies and the observations.

The seasonal averages of temperature and salinity from LRM0 were compared with the climatological observations of the CSIRO Atlas of Regional Seas (www.cmar.csiro.au/cars CARS2009, (www.cmar.csiro.au/cars); (Ridgway et al, 2002). Meanwhile, the transport (Sv) for the main alongshore currents, as the CCC and PCUC, were compared between the LRM0, HRM0 and the HYbrid Coordinate Ocean Model reanalysis (<http://tds.hycom.org/thredds/dodsC/expt91.1-lex.europa.eu>[HYCOM; (Chassignet et al. 2007) between the years 2000 to 2010 from the experim

188 *Resolution dependence of the nearshore circulation*

189 The effect of the increasing the resolution of the nearshore was assessed by calculating
 190 geostrophic and ageostrophic EKE on the HRM2, and comparing it against the LRM0. The
 191 ageostrophic EKE was estimated by the difference between the geostrophic EKE and the
 192 EKE determined directly by both components of the simulated currents velocities. Addi-
 193 tionally, the difference between sea surface temperature and the components of the currents
 194 velocity were determined by downscaling the HRM2 to the LRM0 resolution.

195 Differences due to model resolution in the vertical profile and transport of the PCUC
 196 and CCC were determined by analyzing crossshore transects at sites identified in previ-
 197 ous observational studies as areas of intensification or weakening of upwelling (e.g. Tapia
 198 et al., 2009), Papudo (PA), Punta Curaumilla (PC), Las Cruces (LC), Punta Toro (PT)

199 and Topocalma (TO). The seasonal variability of the CCC, and the existence of a CBL,
200 was evaluated through analysis of the alongshore current transects across the continental
201 shelf (from the coast to the 200 m isobath) at 51 locations between 32.4°S and 34.4°S. The
202 former was identified by the maximum in along-shore velocity, and the latter by the loca-
203 tion of the most rapid decrease in along-shore velocity below the average seasonal speed.
204 The nearshore circulation inside the CBL for the HRM2 and LRM0 were analyzed during
205 a short favorable upwelling period (16-27 December, nineteenth year), through the current
206 frequency distribution normalized by the total number of observations and the bin width.
207 Then, the frequency distribution described the relative vorticity normalized by the Coriolis
208 parameter (ζ/f) and the vertical and horizontal velocities in 2, 10, 20, 30, 40 and 60 m. An
209 investigation of the mechanisms involved in the upwelling or downwelling processes over the
210 inner-shelf in front of the same sites described above was used to determine the causes for
211 the resolution-related differences observed in the near-shore circulation.

212 **Results**

213 The large-scale circulation was examined through three cross-shore transects showing
214 the alongshore flows obtained from the LRM0 (Fig. 1a-c) and HRM0 (Fig. 1d-f) at 30°, 33°
215 and 36°S from the coastline to 450 km offshore. Both coarser grids reproduced a dominant
216 surface nearshore flow which is consistent with the Chilean Coastal Current (CCC) (Shaffer
217 et al., 1995). This current was characterized by an equatorward flow that reached in the
218 LRM0 a maximum annual speed of 8.1 ± 2.1 cm/s across 36°S, and an lower layer depth
219 ranged at 94 to 164 meters, becoming progressively shallower and narrow (64 to 46 km)
220 downstream (Table 2). On the other hand, the nested HRM0 presented a slower maximum
221 current speed across 33°S (6.7 ± 1.6 cm/s) and similarly to the LRM0, the CCC became
222 shallower and narrow downstream, where the lower layer ranged in 73 to 173 meters and layer
223 width varying between 41 to 63 km (Table 2). The CCC transport decreased equatorward
224 without important difference in the annual average between the coarser grids (0.48 to 0.19
225 Sv and 0.4 to 0.17 Sv for the LRM0 and HRM0, respectively). Most transport seemed to
226 be concentrated during winter-spring period, where the average transport reached 0.68 Sv

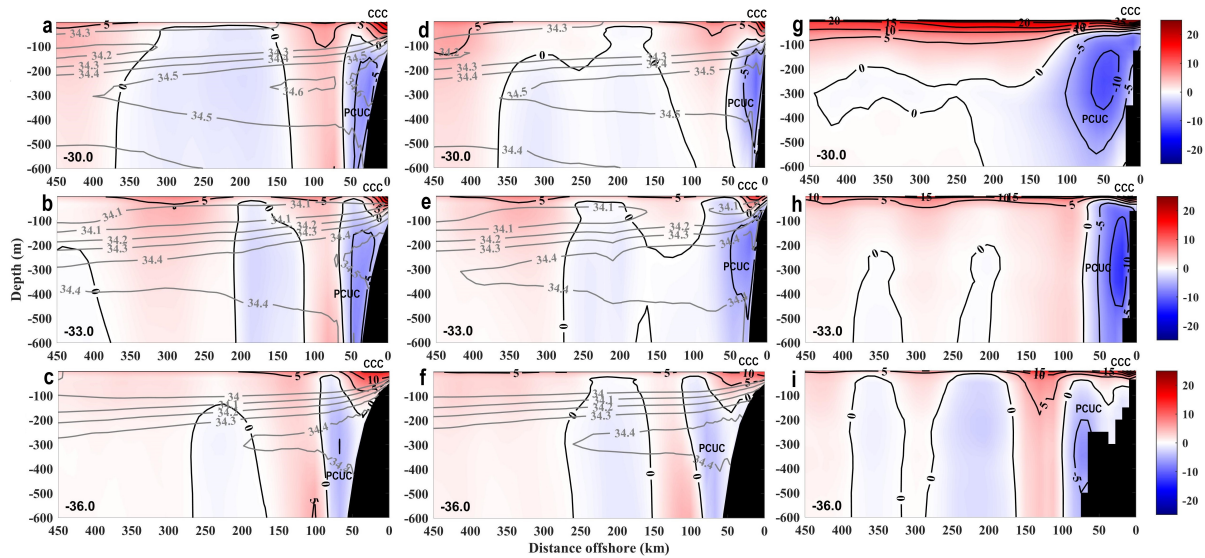


Figure 1: Offshore transects across 30°, 33° y 36°S contrasting the simulated alongshore current (cm/s) from LRM0 (a, b y c), HRM0 (d, e y f) and HYCOM (g, h y i), respectively. The model reproduce the main currents in the Humboldt system, such as the CCC and the PCUC. The contours describe the current speed (black line) and the salinity (gray line)

227 in the LRM0 and 0.66 Sv in the HRM0 around 36°S. Meanwhile, the maximum transport
 228 in 30°S took place in spring with 0.25 Sv in the LRM0 and 0.34 Sv in the HRM0 (Table
 229 2). Comparing our simulations with the transport estimated for HYCOM (Fig. 1g-i), the
 230 reanalysis resulted quite similar at 36°S (ca 8% weaker), but transported 0.9 and -0.3 fold the
 231 observed in HRM0 at 30 and 33°S, respectively (See Table 2). When the resolution increased
 232 through the HRM2, the CCC showed a cross-shore narrowing and deepening equatorwards,
 233 accelerating currents upstream of 33.5°S (i.e., Topocalma site) (Fig. 2). These trends were
 234 stronger in the HRM2 than the LRM0. The maximum alongshore average speed was used
 235 to determine the CCC position, which reached up to 40 cm/s in front of PA and 32 cm/s
 236 in front of TO and PC, and was located around 11, 14, and 8 km offshore, respectively (Fig.
 237 15b).

238 Below the CCC, both coarser grids reproduced a poleward flow adjacent to the slope
 239 and covering most of the shelf bottom layer (Fig. 1a-f), which matched well observations of
 240 the Peru-Chile UnderCurrent (PCUC) described for the region (Strub). Unlike the observed

241 with the CCC, the annual transport of PCUC was stronger in the HRM0 than the LRM0
 242 (ca 0.4 fold).

243 The maximum poleward flow in the PCUC was located at 33°S in both coarser grids
 244 reaching -4 and -4.5 cm/s for LRM0 and HRM0, respectively, although there was a decreasing
 245 speed upstream from -2.6 to -3.6 cm/s and -2.4 to -4.3 cm/s for the same simulations (Table
 246 3). The transport showed the same trend, where LRM0 with maximum values at 33°S and

Table 2: Seasonal difference between the LRM0, the nested HRM0 and HYCOM along the Coastal Chilean Current (CCC). The table compares the Svedrup transport ($1\text{Sv} = 10^8 \text{ cm/s}$), the meridional current velocity (cm/s), the zonal Layer Width (LW, km) and the vertical Lower Layer (LL, meters) of the CCC. The CCC transport estimated in the HYCOM reanalysis were as follow: 30°S=0.37 Sv; 33°S = 0.09 Sv and 36°S = 0.33 Sv.

	LRM0				HRM0			
	Sv	cm/s	LW (km)	LL (m)	Sv	cm/s	LW (km)	LL (m)
30°S								
fall	0.19	5.6	51	94	0.17	4.7	78	88
winter	0.14	6.1	36	105	0.12	6.2	30	112
spring	0.25	7.4	49	106	0.34	2.2	30	73
summer	0.17	5.2	48	89	0.05	3.9	26	71
average	0.19±0.1	6.1±0.9	46±6	98±8	0.17±0.1	4.3±1.7	41±24	86±19
33°S								
fall	0.1	4.4	45	83	0.06	5.5	37	56
winter	0.2	5.2	48	126	0.3	8.7	84	110
spring	0.06	5.1	34	72	0.06	5.4	30	58
summer	0.2	8.3	44	95	0.1	7.2	47	69
average	0.14±0.1	5.7±1.8	42±6	94±23	0.13±0.1	6.7±1.6	49±24	73±25
36°S								
fall	0.3	6	63	97	0.15	3.8	65	74
winter	0.66	7.7	58	317	0.89	6.4	74	400
spring	0.7	11	63	160	0.42	8	56	135
summer	0.24	7.7	70	83	0.12	4.5	58	84
average	0.48±0.2	8.1±2.2	64±5	164±107	0.4±0.4	5.6±1.9	63±8	173±154

247 varying from -0.58 Sv to -0.95 Sv. Similarly, the nested HRM0 resulted in a transport of
 248 -0.91 Sv and -1.32 Sv between 30°S and 33°S, with the weaker transport of -0.7 Sv at 36°S.
 249 The spring-summer season concentrated most of the transport in both coarser grids.

250 The LRM0 and nested HRM0 alongshore subsurface velocities were generally weaker
 251 than those in the HYCOM reanalysis, with the largest departure observed in the northern
 252 transect (30°S), where the PCUC transport was larger by a factor of 4.9 and 3.8, respectively,
 253 especially by the offshore extension of the PCUC in HYCOM (Fig. 1g-i). While the PCUC

Table 3: Seasonal difference between the LRM0 and the nested HRM0 along the Peru-Chile Undercurrent (PCUC). The table compares the Svedrup transport ($1\text{Sv} = 10^8 \text{ cm/s}$), the meridional current velocity (cm/s), the zonal Layer Width (LW, km) and the vertical Upper Layer (UL, m) of the PCUC. The transport estimated in the HYCOM reanalysis were as follow: 30°S = -4.42 Sv; 33°S = -1.65 Sv and 36°S = -0.53 Sv.

	LRM0				HRM0			
	Sv	cm/s	LW (km)	UL (m)	Sv	cm/s	LW (km)	UL (m)
30°S								
fall	-0.75	-3.9	57	78	-0.81	-4.3	53	76
winter	-0.63	-3.7	55	30	-0.7	-3.3	60	94
spring	-0.78	-3.4	64	46	-1.2	-5	70	64
summer	-0.8	-3.6	61	55	-0.86	-4.7	44	51
average	-0.74±0.1	-3.6±0.2	59±4	52±20	0.91±0.3	-4.3±0.7	57±11	71±18
33°S								
fall	-0.8	-3.6	61	27	-1.45	-4.8	92	24
winter	-0.5	-3.3	53	82	-0.24	-1.6	57	22
spring	-1.08	-4.5	62	34	-2	-6.1	80	20
summer	-1.4	-4.6	94	26	-1.6	-5.4	90	20
average	-0.95±0.4	-4±0.7	68±18	42±27	-1.32±0.7	-4.5±2	80±16	22±2
36°S								
fall	-0.73	-3	98	30	-0.99	-2.9	118	27
winter	-0.39	-2.2	87	75	-0.15	-1.5	84	184
spring	-0.59	-3.1	85	84	-0.41	-2.1	94	104
summer	-0.6	-2.2	103	32	-1.24	-2.9	137	31
average	-0.58±0.1	-2.6±0.5	93±8	55±28	-0.7±0.5	-2.4±0.7	108±24	87±74

254 width was extended within the first 100 km from shore in both coarser grids, the offshore
255 extension in HYCOM reached nearly 200 km offshore (Fig. 1g). Another notable difference
256 among both coarser grids was the higher salinity (34.6 psu) reached by the LRM0 and which
257 is directly associated with the PCUC features (1a-f). Contrasting the resolutions, the HRM2
258 showed that, vertically, the PCUC extends between 100 to 400 meters depth, with speeds
259 around 5 up 18 cm/s in front of Punta Curaumilla (Fig. 2c). Over the slope, the PCUC
260 was more intense, which contrasts with the LRM0 where the flow is less intense (up 8 cm/s)
261 and farther from the bottom. In the HRM2, the core poleward flow detaches from the
262 slope in front of Las Cruces (Fig. 2e), where the Maipo River submarine canyon is located,
263 decreasing its velocity compared with the downstream maximum speed reached in front of
264 Punta Curaumilla, a pattern not reproduced in the LRM0 (Fig. 2d and f).

265 The most remarkable crossshore differences associated with the resolution was the nearshore
266 forming of CBL. Here, the HRM2 showed decreasing velocities from the middle to the inner-
267 shelf (ca. 20 km) to the shore, with a further decrease below the seasonal average at about
268 5 km offshore, used here to define the position of the CBL (Fig. 3). While the cross-shore
269 velocity trends were very similar between seasons in the inner-shelf, there were seasonal
270 differences offshore of about 18 km, where the cross-shore mean velocities were stronger
271 in winter than in summer (Fig. 3a). The HRM2 also showed that the CCC equatorward
272 (alongshore) flow increased rapidly from the shore to reach a maximum close to the inferred
273 CBL position, ca. 5-7 km offshore, before decreasing again offshore until approximately 20
274 Km offshore, when it reached more spatially constant values towards the offshore ocean.
275 Summer alongshore flows were slightly higher in summer than in winter within the shelf
276 (Fig. 3c).

277 Thus, the cross-shore and alongshore current gradients confirm that the resolution of
278 the HRM2 is sufficient to allow the formation of a CBL in the inner portion of the shelf.
279 By contrast, a CBL does not form in the LRM0 (Fig. 3b, d). Although a slight decrease
280 in alongshore velocity was seen in the LRM0 within the closest 11.5 km to the coast (16.1
281 to 14.5 cm/s), the coastal speeds were still higher than the summer average. To validate
282 the performance of both coarser grids, the annual eddy kinetic energy (EKE) was compared

283 between the LRM0, HRM0 and satellite observations from AVISO in the same domain as
 284 LRM0 and below the AVISO resolution. Similarly, EKE changes related with the resolution
 285 increment were determined contrasting the LRM0 and HRM2 (Table 4). The geostrophy
 286 EKE was calculated from the sea-level anomaly and using the equation (3). The based-
 287 satellite geostrophy EKE fluctuated between 22 and 115 cm^2/s^2 (Fig. 4a) within the 73°
 288 and 78°W meridional band. The spatial distribution showed two high energy centers, the

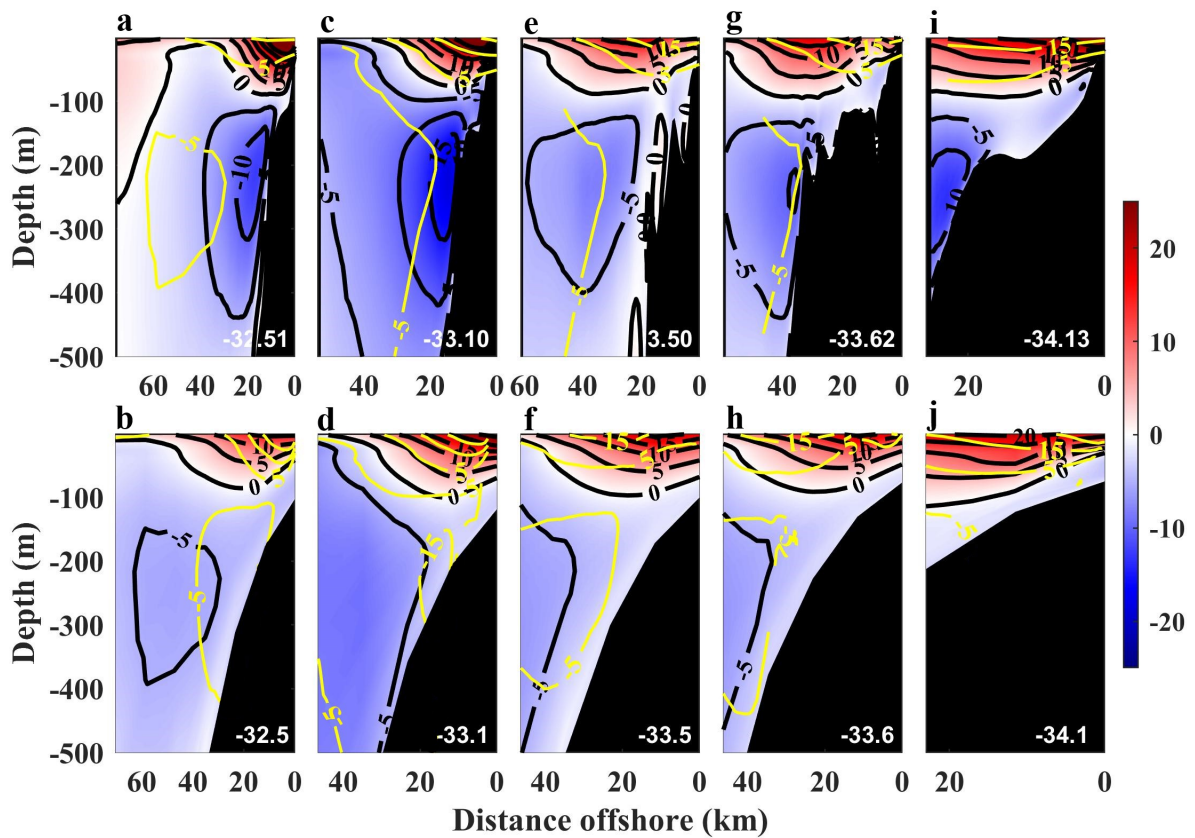


Figure 2: Alongshore currents obtained from the HRM2 (upper panels) with the LRM0 (bottom panels). The transects were located crossshore the sites of Papudo (PA, a-b), Punta Curaumilla (PC, c-d), Las Cruces (LC, e-f), Punta Toro (PT, g-h) and Topocalma (TO, i-j). The CCC (shade red) and PCUC (shade blue) were represented. In the upper (lower) panel, the black contour line represent the velocity in the HRM2 (LRM0), which have been contrasted against the yellow contour representing the current velocity in LRM0 (HRM2). Each models was compared below its own resolution.

289 main one sharper positioned between 28° and roughly 31°S, and a weaker one farther south
290 between 33°-37°S. Overall, the LRM0 exhibited higher geostrophy EKE than is present in the
291 satellite product (Fig. 4c), which ranged from 1.4 up to 266 cm^2/s^2 within the same meridional
292 band, but distributed between 28° and 34°S. The contrast between the satellite data and the
293 HRM0 was more apparent because of the stronger geostrophy EKE in the model, which
294 reached up to 422 cm^2/s^2 , concentrated mostly between 28° and 34°S (Fig. 4d)). Although,
295 there was a broad difference between the ranges of geostrophy EKE, the average values
296 were closer with 59.2, 86.7 and 107 cm^2/s^2 for AVISO, LRM0 and HRM0, respectively. The
297 difference between the coarser grids reveals much sharper turbulent activity within the HRM0
298 than the LRM0 (Fig. 4b).

299 In the HRM0, the EKE appears increasing mainly inside the northern region where the
300 nesting domains were conducted. Outside the nested domain, the southern region, and
301 specifically near Punta Lavapie (37.2°S), the EKE was at least 100 cm^2/s^2 stronger in the
302 LRM0 (Fig. 4b). The sharp EKE in the HRM0 may be related to an enhanced current speed
303 or turbulent activity associated to the nesting domains inside this coarse grid. Hence, the
304 total EKE was estimated through components of the current to solve the ageostrophy EKE
305 and determine a positive correlation with the geostrophy EKE related with the nested grid.

306 However, the ageostrophy EKE were fairly similar between both grids, with 12.9 ± 13.5
307 and 10.5 ± 15.4 cm^2/s^2 for the LRM0 and HRM0, respectively, representing the 13% and
308 9% of the total EKE. Inside the HRM2 domain (32°-34.5°S), the annual geostrophy EKE
309 increased to 186 ± 41 cm^2/s^2 in the LRM0 and 368 ± 158 cm^2/s^2 in the HRM2 (Fig. 5a),
310 where the main difference seems to form a filament towards northwest off Punta Curaumilla
311 (Fig. 5a). Despite that more energy is produced with more resolution, the HRM2 held
312 a low ageostrophy EKE ($3 \pm 6\%$ of total EKE) restricted to specific sites over the shelf,
313 closer to the headlands and recognized as upwelling sites (i.e. Papudo, Punta Curaumilla
314 and Punta de Toro, Fig. 5b). Meanwhile the LRM0 enhanced to $28 \pm 10\%$ the ageostrophy
315 EKE (Table 4). The results suggest that an important amount of energy may be transferred
316 from the shelf to the open ocean and where the higher resolution was able to capture the
317 location of important ageostrophy circulation (i.e. Ekman transport) that contributed up

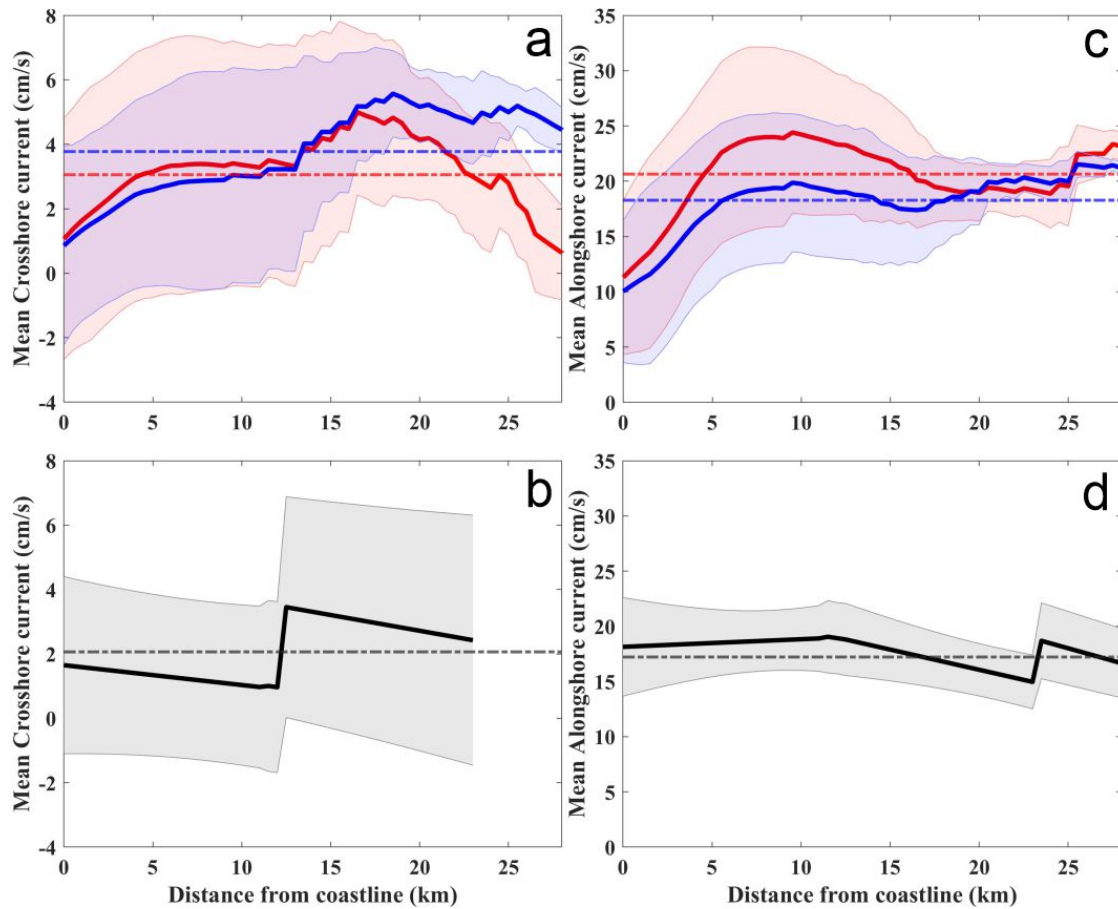


Figure 3: Coastal Boundary Layer (CBL) represented through the integrated crossshore (a, b) and alongshore (c, d) current in the HRM2 (upper panels) and LRM0 (bottom panels). The figure describe a decline in the current velocity below the seasonal average (dotted line). The blue line correspond to the winter season and the red line to the summer in the HRM2. The LRM0 only shows the summer season.

318 38% of the nearshore total EKE.

319 Another validation source compared the seasonal climatological structure of sea surface
 320 salinity (SSS) and temperature (SST) off central Chile (Fig. 10) between the LRM0 and
 321 CARS. Spatial correlation coefficients for SSS ranged from 0.94 up 0.97, while the SST
 322 varied between 0.89 and 0.94 during spring and summer, respectively. The SSS presented a
 323 standard deviation for spring and autumn close to those found in CARS observations (0.22
 324 psu, Fig. 10a and c), while SST showed overall spatial standard deviations greater than
 325 observations, but closely similar during winter and autumn (1.2 °C, Fig. 10b and d).

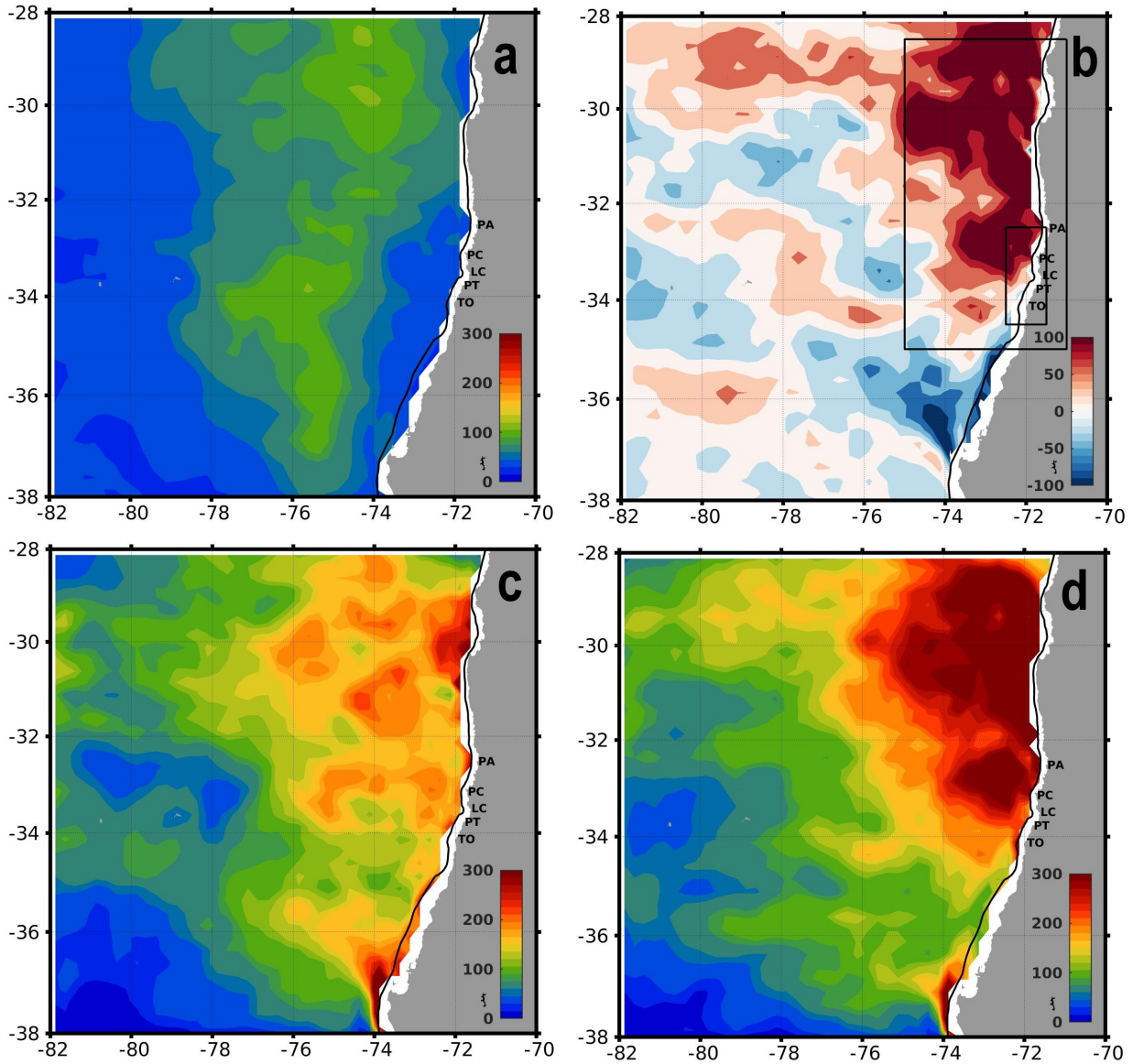


Figure 4: Annual Eddy Kinetic Energy (EKE) calculated from satellite sea-level anomaly from AVISO (2000-2014)(a) and simulated sea-level height anomaly in LRM0 (c) and HRM0 (d). Each coarse grid was previously downscaled to the AVISO resolution. The EKE difference between the nested HRM0 and LRM0 have depicted in panel b, as well as the nested domains inside HRM0 (HRM1 and HRM2), were enclosed inside the rectangles.

326 Moreover, the centered root-mean-square (RMS) showed that SST and SSS remained
 327 close to observations, with differences ranging between ± 0.41 to ± 0.55 $^{\circ}\text{C}$ and ± 0.05 to
 328 ± 0.09 psu, respectively. The LRM0 also exhibited good spatial coherence with observed

329 seasonal sea level anomaly (SLA), but with more defined sea-level structures, which rendered
 330 only a moderate spatial correlation with observations (Fig. 11a and b).

331 In contrast to the comparatively large differences in EKE between the LRM0 and HRM2,
 332 the other oceanographic variables (i.e. u , v , w , speed, temperature, density and the nor-
 333 malized relative vorticity) were seasonally described in the Table 6 and showed only small
 334 differences. Here, the water column was divided in integrated surface (0-20 m) and bottom
 335 (150-200 m) layers according with the main alongshore currents (i.e. CCC and PCUC).
 336 In the surface layer, the horizontal currents were faster in the LRM0 during the summer

Table 4: Distribution of EKE in the coarser grids (LRM0 and HRM0) contrasted with satellite observations from AVISO and the contribution of geostrophy and ageostrophy circulation to the total EKE in the LRM0 and HRM2.

	AVISO	LRM0	HRM0	HRM2
	Coarser domain and AVISO resolution			
geos-EKE				
mean	59.2	86.7	107	
SD	22.5	56.5	81.8	
range	22-115	1.4-266	1.3-422	
ageos-EKE				
mean		12.9	10.5	
SD		13.5	15.4	
range		13-94	0.3-118	
	finest domain and LRM0 resolution			
geos-EKE				
mean		186		368
SD		43.7		158
range		80-303		44-1300
ageos-EKE				
mean		78.4		11.1
SD		51.3		21.8
range		0-308		0-109

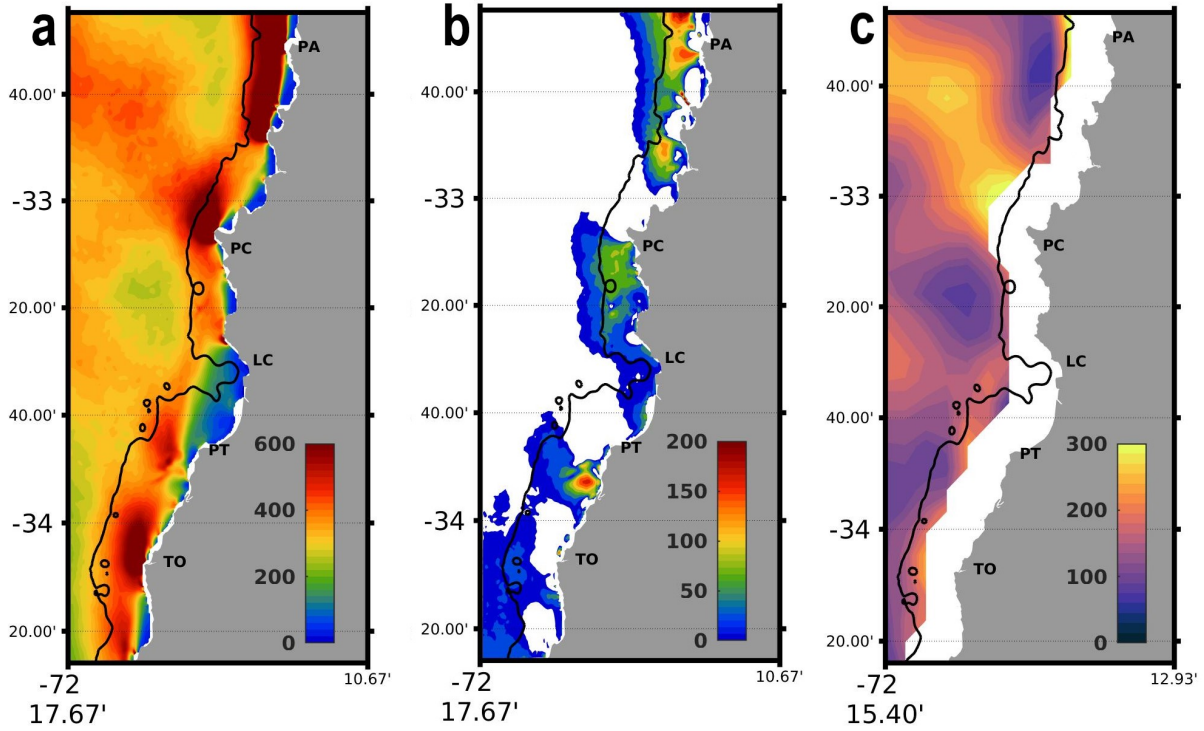


Figure 5: Annual Eddy Kinetic Energy (EKE) from the HRM2 decomposed in total EKE (a) and ageostrophic EKE (b). The panel c shows the difference between the HRM2-LRM0 associated to the total EKE. The HRM2 was downscaled to the LRM0 resolution.

337 $(12.2 \pm 4.3 \text{ cm/s})$ and slower during the winter $(10.8 \pm 5.4 \text{ cm/s})$ than the HRM2 (Fig.
 338 6a-f). However, the maximum speed that reached up 24.5 cm/s in summer and up 23.9
 339 cm/s in winter were located nearshore in the HRM2, as well as the slower speed closer to
 340 the bays (Table 6e-f). The alongshore and crossshore components of the horizontal current
 341 described the predominance of equatorward advective circulation and the seasonal Ekman
 342 transport, which are directly related with the vertical flow. Vertically, the surface alongshore
 343 current velocity was faster shoreward and the direction was predominantly equatorward in
 344 both season and model resolution (Fig 12g-l). The HRM2 showed a higher equatorward flow
 345 nearshore than the LRM0 in both seasons (up 24.1 and up 23.5 cm/s for summer and winter,
 346 respectively) (Fig 12i-l), even though in average the LRM0 was faster during the summer
 347 $(12.1 \pm 5.2 \text{ cm/s})$ but slightly lower in winter. These differences ranged in -10.1 and 8.3
 348 cm/s in summer and -15.7 and 8.7 cm/s in winter for the HRM2 respect to the LRM0. In

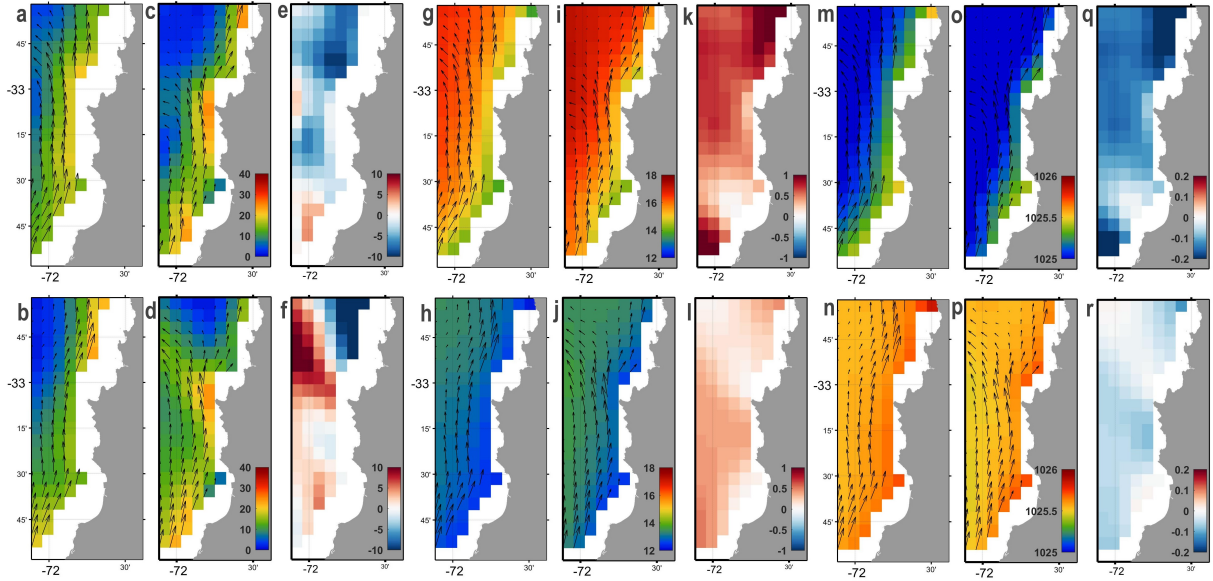


Figure 6: Seasonal average of the speed current (cm/s), SST ($^{\circ}\text{C}$) and density (kg/m^3) in the LRM0 (a-b, g-h and m-n, respectively) and HRM2 (c-d, i-j and o-p, respectively). For difference between HRM2 and LRM0 (e-f, k-l and q-r, respectively), the positive (negative) values indicate a higher (lower) magnitude in the HRM2 than LRM0. The upper panel represents the summer season and the lower panel the winter, both integrated vertically (0-20 meters). The HRM2 was downscaled to the LRM0 resolution. The arrows describe the flow speed direction

349 the bottom layer, also the HRM2 was faster (-7.2 ± 2.4 cm/s) than the LRM0 (-5.6 ± 1.1
350 cm/s), but more close to the slope (Fig 12e), with a dominant poleward flow in summer but
351 less strong in winter (Fig 12c-d). During summer and winter, the crossshore velocity in the
352 surface layer showed a clear offshore transport in the HRM2 located permanently in Punta
353 Curaumilla, which is correlated with the alongshore flow (Fig 13i-l). However, the LRM0
354 showed in average more offshore flow than HRM2 in summer (Table 6) and less in winter,
355 but the higher offshore velocities were presented in the HRM2 in both season (up -7 cm/s
356 and up -9.4 cm/s, respectively). In the bottom layer, the expected shoreward flow related to
357 the upwelling season was lower in LRM0 (-1.1 ± 0.6 cm/s) than in HRM2 (-1.1 ± 2.1 cm/s),
358 but in winter the maximum shoreward flow decreased in the HRM2 (3.3 to 1.9 cm/s), as
359 should be expected for winter season, while the LRM0 increased from 0 to 2.1 cm/s since
360 summer to winter (Fig 13a-f). The vertical flow in both coarser grids was dominated by the

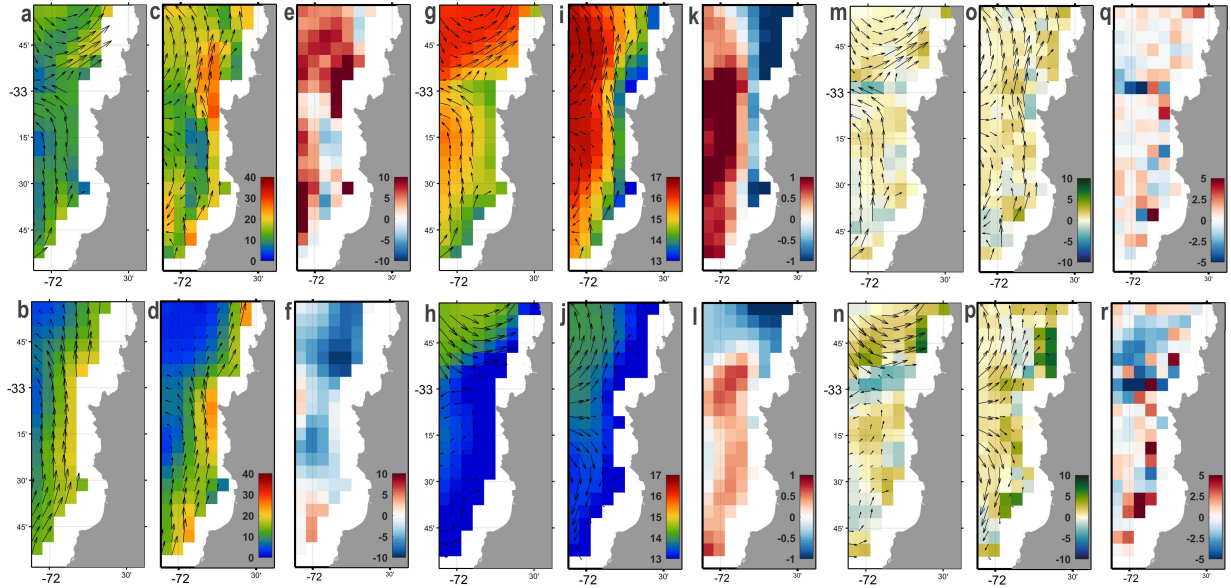


Figure 7: Mean circulation in the surface (0 - 20 meters; upper panels) and bottom layers (30 - 70 meters; lower panels), during a favorable upwelling period (16 to 27 December in nineteenth year). In order, the speed current (cm/s), SST ($^{\circ}$ C) and vertical current (m/d) in the LRM0 (a-b, g-h and m-n, respectively) and HRM2 (c-d, i-j, and o-p, respectively). For difference between HRM2 and LRM0 (e-f, k-l and q-r, respectively), the positive (negative) values indicate a higher (lower) magnitude in the HRM2 than LRM0. The upper panel represents the summer season and the lower panel the winter. The HRM2 was downscaled to the LRM0 resolution. The arrows describe the flow speed direction.

361 upward flow in both season, with maximum values along nearshore and close to the slope
 362 (Fig 14a-l). During the summer, the upward flow in the surface layer was slightly higher
 363 in the HRM2 (0.4 ± 0.5 m/d) than LRM0 (0.3 ± 0.47 m/d), but the upwelling on front
 364 Punta Curaumilla was stronger in LRM0 than in the HRM2. In winter, the upward flow
 365 decreased in both simulations and showed the same average, but HRM2 keep more variance
 366 (0.2 ± 0.3 m/d, Table 6). The seasonal currents components determined almost similar
 367 velocities in both resolution, but with more variance in the HRM2. This small difference
 368 must reflect more upwelling nearshore in higher resolution. The summer and winter temper-
 369 atures were higher in the HRM2 (16.2 and 13.3 $^{\circ}$ C) than the LRM0 (15.6 and 13 $^{\circ}$ C), but in
 370 the known upwelling sites (Punta Curaumilla and Punta de Toro), the temperatures were
 371 almost similar in both resolutions and seasons (Fig. 6k-l). Moreover, the thermal gradient

372 in the HRM2 was stronger at least in summer (Fig. 6i). Also, the more dense nearshore
 373 waters in both models specially during the summer support the upwelling occurrence, where
 374 within the LRM0, the processes seems more intense ((Fig. 6)m, o and q). However, in fact
 375 the surface colder and salty intermediate waters in the LRM0 (see Figure 1a) affected the
 376 coastal density causing a artifact. On other hand, in the bottom layer the density was the
 377 same in both resolutions without seasonal differences (Table 6). Locally, Punta Curaumilla
 378 and Punta Toro sites matched the ageostrophic EKE (Fig. 5b) and suggesting that the finer
 379 resolution intensified the Ekman transport, probably caused by the complex topography.
 380 Other sites of upwelling intensification identified in observational studies, as well as sites of
 381 weak or no upwelling, were also resolved in the HRM2.

382 Thereby, the difference between resolutions must be huge when a short upwelling period

Table 5: Mean circulation in the LRM0 and the HRM2 during a favorable upwelling period on front Punta Curaumilla (see Figure 9). The HRM2 was downscaled to the LRM0 resolution. The results were calculated averaging 6 pixels on front Punta Curaumilla and integrating the upper layer (0-20m) and the lower layer (30-70m). These results were contrasted with the same variable obtained within the CBL but along the whole coast and water column (2, 10, 20, 30, 40, 50, 70) and using the own resolution for each model (see Figure 8).

	V (cm/s)	U (cm/s)	W (cm/s)	ζ / f	T (°C)
	Upper Layer				
LRM0	6.5±1.1	-1.9±5.1	-0.6±0.5	0.06±0.02	14.6±0.3
HRM2	22.6±9.5	2.2±3	0.15±1.7	0.05±0.2	14.6±0.7
	Lower Layer				
LRM0	2±1	-3.2±3	-1.7±0.9	0.004±0.02	12.7±0.2
HRM2	9.2±6.6	2±2.6	-0.2±2.7	0.07±0.07	12.9±0.4
	Whole column				
LRM0	1.7±8.8	2±4.2	2.4±3.8	-0.1±0.1	
HRM2	8.54±13.3	1.8±6.1	2.2±24.1	0.1±0.7	

383 is analyzed. The Figure 16 and Figure 16 depicted a variability during a favorable upwelling
384 period (13 days) in a integrated surface (0-20 m) and bottom layer (30-70 m). Such as in
385 the seasonal analysis, the maximum speed appeared close to Punta Curaumilla and Punta
386 de Toro (Fig. 16c), where the nearshore flow, related to the CCC, also reached the bottom
387 (Fig. 16d) in the HRM2. During this period, the alongshore velocity was quite stronger
388 in HRM2 (11.3 ± 8.5 cm/s) than in LRM0 (6 ± 5.4 cm/s) in the surface layer and the
389 temperature, which was quite similar in both resolution ($15.4 \pm 1^\circ\text{C}$ and $15.2 \pm 0.6^\circ\text{C}$,
390 respectively) and bottom layer ($13.3 \pm 0.6^\circ\text{C}$ and $13.3 \pm 0.7^\circ\text{C}$, respectively), raised colder
391 nearshore in the finer resolution, forming a stronger horizontal thermal gradient offshore
392 (Fig. 16i and k). The vertical current associated with the normalized relative vorticity were
393 spatially difficult to described in the comparative coarse resolution because apparently these
394 act more locally. Thus, the anticyclonic (negative) vorticity must be directly related to the
395 upward flow (upwelling) and the cyclonic vorticity with the downwelling (Fig. 16m-p). In
396 average, the HRM2 showed higher upward velocity (0.4 ± 1.7 m/d) than the LRM0 (0.16
397 ± 1 m/d) and maximum upward velocities up 3.7 and 3 m/d, respectively.

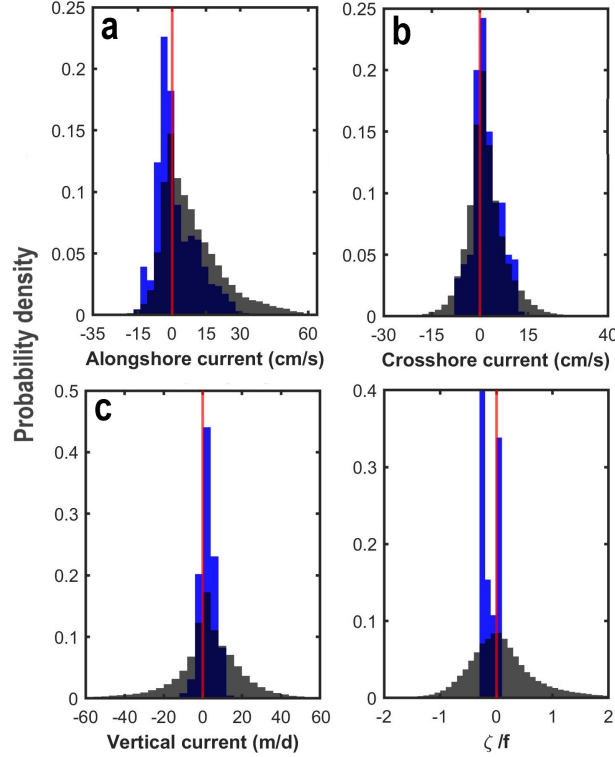


Figure 8: Relative probability (samples in a bin/total samples) of alongshore (a), crossshore (b) and vertical current (c), density (d) and relative vorticity normalized (e) during an favorable upwelling period within the CBL. The figure describes the variables for the LRM0 (blue bars) and HRM2 (gray bars). Each histogram integrated the variables through time (13 days) and depths (2, 10, 20, 30, 40, 50 and 70 m) within the CBL.

398 For the same period and regarding the CBL, the component of the current and the
 399 vorticity presented more variance in the HRM2 but the same average as the LRM0 when
 400 were compared using their own resolution (Fig. 8 and Table 5). The probability density of
 401 alongshore current was asymmetric in both resolution, but with most equatorward advective
 402 flow in the HRM2 varying between -30.6 to 65.3 cm/s, meanwhile, the LRM0 ranged between
 403 -15 to 31.7 cm/s (Fig 8a). Although, the variation showed extreme velocities, the averages
 404 were within expectations for CBL and lower to the values find outside the inner-shelf (8.5
 405 ± 13.3 cm/s in the HRM2 and 1.67 ± 8.8 cm/s in the LRM0). The crossshore velocity (Fig.
 406 8b) ranged from -25.5 to 37.2 cm/s and from -7.5 to 12.4 cm/s for the HRM2 and LRM0,
 407 respectively, but again the average were quite similar (1.8 ± 6.1 cm/s in the HRM2 and $2 \pm$

408 4.2 cm/s in the LRM0). However, the vertical flow in the HRM2 was up 44 fold faster than
 409 in the LRM0, with a range between -512 and 430 m/d in HRM2 compared with the -11 to
 410 13 m/d in the LRM0 (Fig 8c). The standard deviation was 6 fold higher in the HRM2 with
 411 2.2 ± 24 m/d contrasting with the 2.4 ± 3.8 m/d found in the LRM0. Moreover, the relative
 412 vorticity (Fig 8d) reached values that represent submesoscale processes in the HRM2 (-5.3
 413 to 5.7) and exceeding the planetary vorticity ($Ro \ll 1$), which weren't capture by the LRM0
 414 (-0.3 to 0.09).

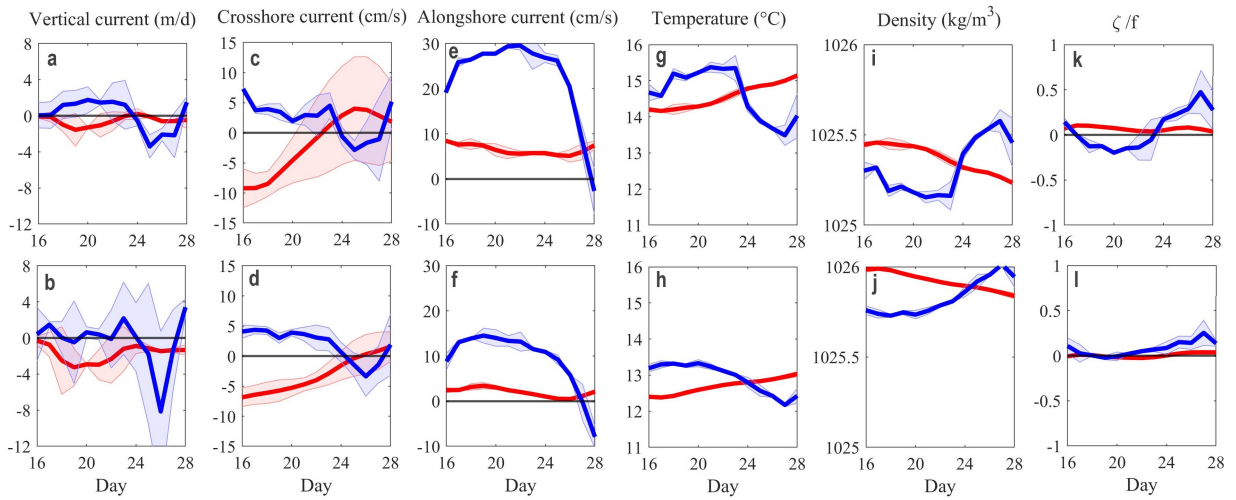


Figure 9: Time serie of mean nearshore circulation on front of Punta Curaumilla, during a favorable upwelling period between 16 to 27 December nineteenth year. The figure describe the variability of vertical (a-b), crossshore (c-d) and alongshore currents (e-f) related to changes in temperature (g-h), density (i-j) and relative vorticity normalized (k-l). The blue line represent the HRM2 and the red line the LRM0, while the upper panels to the surface layer (0 - 20 meters) and the lower panels to the bottom layer (30 - 70 meters). The standard deviation was depicted in shade color. The models were contrasted in the LRM0 resolution in an area of 15.4x23.1 km (2x3 pixels).

415 The differences in SST between models of different resolution noted above is linked to
 416 the significant differences in the ability of these models to resolve the coastal upwelling
 417 process. The evolution of the physical state in the surface and bottom layer in HRM2 and
 418 LRM0 during a single event of intense upwelling is illustrated in Figure 9. During the same
 419 period, the LRM0 and HRM2 responded differently to the identical wind stress forcing.

420 The HRM2 surface and bottom layers were accelerated to velocities of up to 30 cm/s and
421 15 cm/s respectively (Fig. 9e-f), simultaneously with an upward flow in both layers (Fig.
422 9a and b), Upon weakening of the alongshore flow 7 days into the event, an offshore flow
423 formed consistent with Ekman transport (Fig. 9a and b), the temperature diminished (Fig.
424 9g and h) and the density in both surface and bottom layers increased (Fig. 9i and j). By
425 contrast, the response of the LRM0 model to the same wind stress was weak, driving a weak
426 alongshore current and upward flow, but with greater cross-shore transport than observed in
427 the HRM2. The higher density and lower temperature during the time series, demonstrated
428 that the Ekman transport started before than in the HRM2.

429 Discussion

430 Much of the variability in both the large-scale and near-shore circulation off Chile is
431 driven by the seasonal meridional displacement of the South Pacific Anticyclone (SPA),
432 which controls the intensification of the upwelling-favorable southwesterly wind during the
433 austral spring-summer off central Chile (Bravo et al., 2016, Schneider et al., 2017). The
434 climatological wind used in all resolution models allowed captures this seasonal cycle. Both
435 coarser grids (LRM0 and HRM0) showed a good spatial correlation in SSS and SST, re-
436 producing the corresponding seasonal alongshore gradients where colder, less salty waters
437 predominate in the austral fall-winter, and warmer, salty waters dominate during the austral
438 spring-summer (Sobarzo et al., 2007). The effect of seasonality over the main meridional
439 currents was confidently reproduced by the coarser grids, increasing the transport during
440 summer and decreasing in winter for the CCC and PCUC. Also the models reproduced the
441 meridional gradient in the transport. Significantly difference between the LRM0 and the
442 HRM0 weren't observed over the CCC. However, the HRM0 had a less salty PCUC with
443 higher transport (-0.7 ± 0.5 to -1.3 ± 0.7 Sv) and speed magnitudes (-2.4 ± 0.7 to -4.5 ± 2
444 cm/s). Other simulations, for the same region, have estimated the annual transport in -0.9
445 to -0.71 Sv (Aguirre et al., 2014), in -0.8 to -0.76 Sv (Vergara et al., 2016) and between -0.88
446 to -0.71 Sv (Pizarro-Koch et al., 2019). These results are closer to our results obtained from
447 the LRM0, which early demonstrated a nesting and resolution effect able to accelerate the

448 flow in the HRM0.

449 Resolution grid differences arose more clearly when the LRM0 and HRM2 were com-
450 pared. Thus, the CCC flow transported between 0.8 to 1.6 fold more water than the LRM0,
451 meanwhile the PCUC was up 3 fold faster (0.8 to 16 cm/s) at 33°S and reached the same
452 density as the LRM0. Similar velocities have been recorded at 30°S in the PCUC core by
453 Shaffer et al. (up to 12.8 cm/s, Shaffer et al., 1999). Furthermore, other numerical simu-
454 lations have estimated comparable speeds between 7.6 to 11.8 cm/s (Leth & Shaffer, 2001)
455 and 10 cm/s (Vergara et al., 2016). In the finer resolution, the PCUC seems to interact more
456 with the slope which might induce the subsurface eddy generation due to abrupt change in
457 topography that accelerate the flow, the relative vorticity and export matter offshore (Con-
458 treras et al., 2019). However, the lower resolution bathymetry in the coarser grids avoid the
459 acceleration flow related with the vorticity causing receding from the slope and widening of
460 the stream.

461 The spatial distribution of EKE in the LRM0 resembles the Coastal Transition Zone
462 described in Hormazabal et al (2004), a region with high mesoscale eddy activity along 29°
463 to 39°S associated with strong and variable southerly wind stress. While satellite derived
464 EKE identifies two EKE cores, located around 30°S and between 33 and 37°S, the LRM0
465 matched the 30°S core (i.e. Coquimbo Bay), but also presented at least two other local
466 EKE maxima across 33°S (i.e. near Valparaiso Bay) and 37.2°S (i.e. along Punta Lavapie).
467 Southern Coquimbo Bay is located close to the Punta Lengua de Vaca, a cape known by
468 interact strongly with the equatorward flow inducing the formation of jets and eddies seaward
469 (Shaffer et al., 1999, Hormazabal et al., 2004). In the same line, Punta Lavapie is recognized
470 by the same ocean-land interaction southern Arauco Gulf and strong southerly winds stress,
471 jets and eddies formation (Mesias et al., 2001, Aguirre et al., 2012). Less windy but known as
472 an important upwelling zone is Punta Curaumilla located southern Valparaiso Bay, whose
473 topography allow the generation of northward cold filaments (Silva & Valdenegro, 2003),
474 which reproduced the HRM2 mostly in winter.

475 While increasing the horizontal resolution of the model resulted in significant improve-
476 ments in the simulation of near-shore processes, it also resulted in an over-estimation of the

477 off-shore EKE. The reason for EKE being substantially greater than estimated from satellite
478 altimetry may be related to the scatterometer winds used for forcing, which cannot resolve
479 the decrease in wind speed within the first 5km offshore (Capet et al., 2014). This will tend
480 to increase the alongshore current, Ekman transport, upwelling (Bravo et al., 2016, Astudillo
481 et al., 2019), frontal instabilities and finally EKE (Brink, 2015, Marchesiello et al., 2003).
482 To wit, the LRM0 reproduces a broader zone of cold and dense surface waters than HRM2,
483 where the upwelling is restricted to some specific sites. In the LRM0 case, the absence of
484 wind drop-off favors the upwelling by Ekman transport over Ekman pumping (Capet et al.,
485 2004) adding more EKE to the surface ocean. In the HRM2, the sources of EKE arise from
486 the Ekman transport that increased only locally as showed the nearshore ageostrophy EKE,
487 probably caused by the complex topography, the augmentation of current velocities and the
488 submesoscale circulation that drive the vertical advection through small eddy formation and
489 relative vorticity, able to transfer energy by inverse cascade mechanism to the mesoscale
490 circulation. Capet et al (2008), studying a toy model of upwelling determined that fine
491 resolution below 1.5 km may increase the submesoscale circulation. The surface flow accel-
492 eration is a known result of resolution increment (Marchesiello et al., 2003) and acts as the
493 main source of EKE, propagating the energy around the domain. Therefore, increasing the
494 grid resolution involves an increment in EKE related to flow acceleration and submesoscale
495 circulation that occurs during the year and whose magnitude is enough to be captured in
496 the seasonal mean analysis.

497 The degree to which the observed upwelling was reproduced varied between the spatial
498 and time resolution. Colder nearshore surface temperatures at Punta Curaumilla in HRM2,
499 relative to LRM0, demonstrate the increased intensification of upwelling due to increased
500 resolution, especially when the daily variability is analyzed. In fact, the warmer offshore
501 (17°C) and colder inshore (13.5°C) temperature gradient in the HRM2 during the summer
502 strengthened the region of baroclinic instability between 71.8 and 72.4° W, which is a source
503 of EKE (Brink, 2016). In the HRM2, upwelling cooled near-shore SST down up 13.5°C
504 - similar to the observations ($\sim 14^{\circ}\text{C}$) reported by Silva and Valdenegro (2003), where the
505 temperature keep low after decline the winds stress. The mean Ekman transport, calculated

506 as the vertical flow in the 356 km² varied in 533 to 627 m³/s, a value into of moderate
507 upwelling expected for a wind of \bar{u} 7 m/s. In addition to classical Ekman coastal upwelling,
508 the interaction of the CCC with the bathymetry, especially adjacent to headlands, can
509 cause it to branch upwelling (Leth et al., 2004, Leth & Middleton, 2004). Notably HRM2,
510 but weak in the LRM0, a CCC branch becomes detached from the coast in front of PC
511 during summer and winter, resembling the cold filament described by Silva and Valdenegro
512 (2003). Over the continental shelf, Leth and Shaffer (2001) simulated a equatorward surface
513 current velocity around 10-20 cm/s during the austral summer, estimating that the main
514 factor responsible of meandering and eddy formation was the bathymetry and coastline
515 variations, which may supply vorticity to the flow and sharpen the horizontal velocity shear,
516 which then causes instabilities and the subsequent growth of meanders. Thus, the strong
517 equatorward advection, represented in our model as a positive skewness dominating the
518 surface circulation over the shelf, in addition with the eddy and meander formation coupled
519 to the Ekman transport, together should be feeding the mesoescale EKE increment.

520 Despite the along-shore flow acceleration seen in the HRM2, the horizontal currents were
521 less sensitive to the change in resolution, than the vertical speed and normalized relative
522 vorticity. Marchesiello et al (2003) modeling the large-scale circulation in the California
523 Current System determined an insensitivity of the horizontal currents to the model resolution
524 through seasonal averages, indicating that the main difference occurred shoreward over the
525 EKE and sea level anomaly. Also, Capet et al (Capet et al.) concluded the same low
526 sensitivity of horizontal current to the resolution increment but regardless the first 80 km of
527 coast offshore. Similarly, Zhong and Bracco (2013), comparing two resolution models (1 and
528 5 km) forced by diurnal winds over the Gulf of Mexico, finding no difference in the horizontal
529 velocities but mainly in vorticity where the coarser grid reproduced the same structures
530 smoothed. However, the emergence of a CBL in the HRM2 allowed to capture more variance
531 in the horizontal current within the km of offshore but keeping the mean. Evenly, the
532 normalized relative vorticity showed a large variance with more positive skewness to the
533 cyclonic eddies over the shelf, meanwhile, the vertical flow showed a Gaussian distribution
534 with large variance in the HRM2. The variance was up 2, 3, 38, and 42 folds higher in the

535 HRM2 than the LRM0 over CBL into the innershelf.

536 The decline in the integrated velocity of horizontal current coincides with the cross-
537 shore observations in the Northeast Pacific shelf carried-out by Nickols et al (2012, 2012),
538 which suggest a logarithmic speed decay relative to the coastline distance even when existed
539 topographic and bathymetric difference between the sampled sites. The HRM2 reproduces
540 a CBL on average within the first 5 km offshore over the continental shelf. The lateral
541 boundary layer is caused by the transferal of lateral and bottom stress by eddy viscosity
542 (Nickols et al.). The resolution of the LRM0 is too low to resolve this process. The CBL is
543 known to impact essential ecological processes such as larval dispersal and metapopulation
544 connectivity of rocky shore communities by allowing the larval retention and accumulation
545 across the inner-shelf, facilitating the self-recruitment (Nickols et al., 2012). This higher
546 retention within the first kilometers from the coastline has been evaluated by Dauhajre
547 et al.(2017), suggesting that in models with resolution over 1 km, the retention results of
548 an underestimation of the vertical and horizontal speed currents that conform filaments and
549 fronts, which would increase the exchange within the CBL compared to grid resolutions under
550 100 m. On the other hand, Brandini et al, (2017), correlated simulations with X-frequency
551 radar inside the CBL, which showed submesoscale eddies which would be associated to the
552 interaction between the current with the bathymetrical complexity and where the vorticity
553 generation trigger a processes of current attenuation.

554 We have presented clear evidence that using fine grid of at least 1 km of resolution, to-
555 gether with a detailed bathymetry, allows to improve the simulation of small-scale processes,
556 such as submesoescale turbulence, ageostrophy circulation and the CBL. Even though the
557 finer resolution intensified of the upwelling, limiting the spatial extension seen in the coarser
558 grids. The horizontal current velocity increased most variance within the CBL, keeping
559 the average and the more relevant changes were associated to the vorticity and vertical
560 current enhanced by the topographical flow interaction. Despite the high EKE, our model
561 provides a good simulation of CBL, impossible to reproduce in lower resolution, and the
562 nearshore mixing processes, turning it in a valuable tool to applied in ecological larval dis-
563 persal studies. Thus, the application of high resolution models for open ocean questions

564 seems inadequate on the horizontal scale, unless a near-shore problem regarding the CBL
565 been addressed. The variability capture by the high resolution grids could even increase
566 when a high resolution bathymetry is adopted as initial condition, which have been sug-
567 gested in other studies (Saldías & Allen, 2020). The ecological application of high resolution
568 models is then recommended in studies applied in nearshore species.

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885 1. Taylor Diagrams

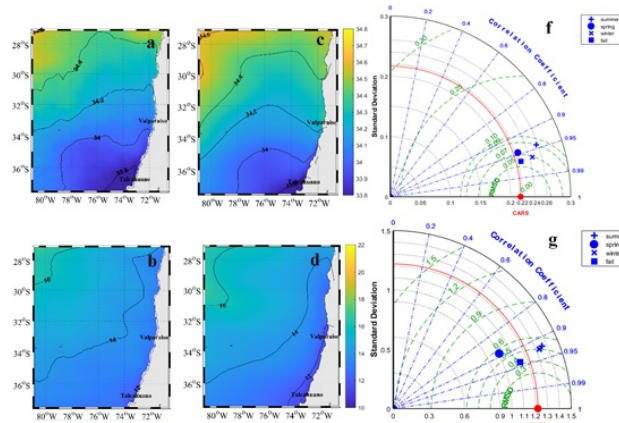


Figure 10: Sea surface temperature (SST) and salinity (SSS) during fall in the LRM0 (a and b) and observations from CARS (c and d), respectively. The Taylor diagram The Taylor diagram (c) compared the seasonal climatology of SST (f) and SSS (g) from the LRM0 with CARS observations, through the spatial correlation coefficient (dashed blue line), standard deviation (dotted black line), and the root mean square deviation (RMSD, dashed green line). The red line denote the standard deviation for the observations.

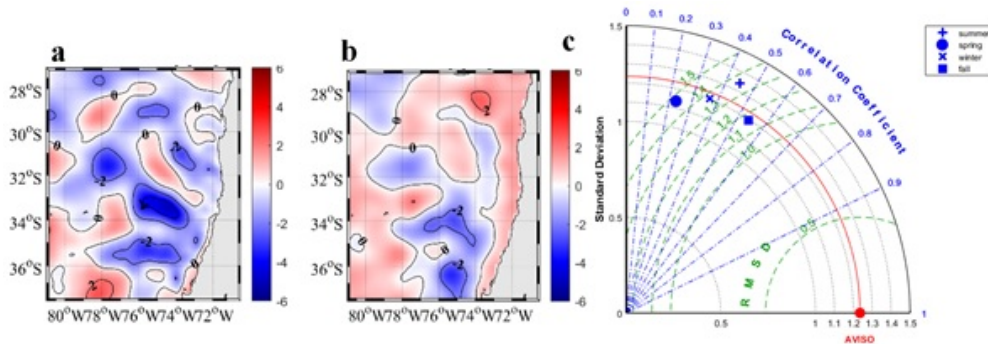


Figure 11: Annual sea-level anomaly calculated in the LRM0 (a) and satellite observation from AVISO (b). The Taylor diagram (c) compared the seasonal climatology of LRM0 with AVISO, through the spatial correlation coefficient (dashed blue line), standard deviation (dotted black line), and the root mean square deviation (RMSD, dashed green line). The red line denote the standard deviation for the observations.

886 The Taylor diagram (Fig. 11c) showed that the spatial correlation between the LRM0
 887 and observations was lowest in spring (0.23) and maximum in autumn (0.54). The stan-

888 dard deviation values indicated that the LRM0 captures an essential fraction of the spatial
 889 variability observed in SLA (± 1.23 cm), especially during winter and autumn (± 1.2 cm),
 890 while the RMSD of the LRM0 was greater (± 1.4 to ± 1.1 cm) than the satellite anomalies
 891 (Fig. 11c).

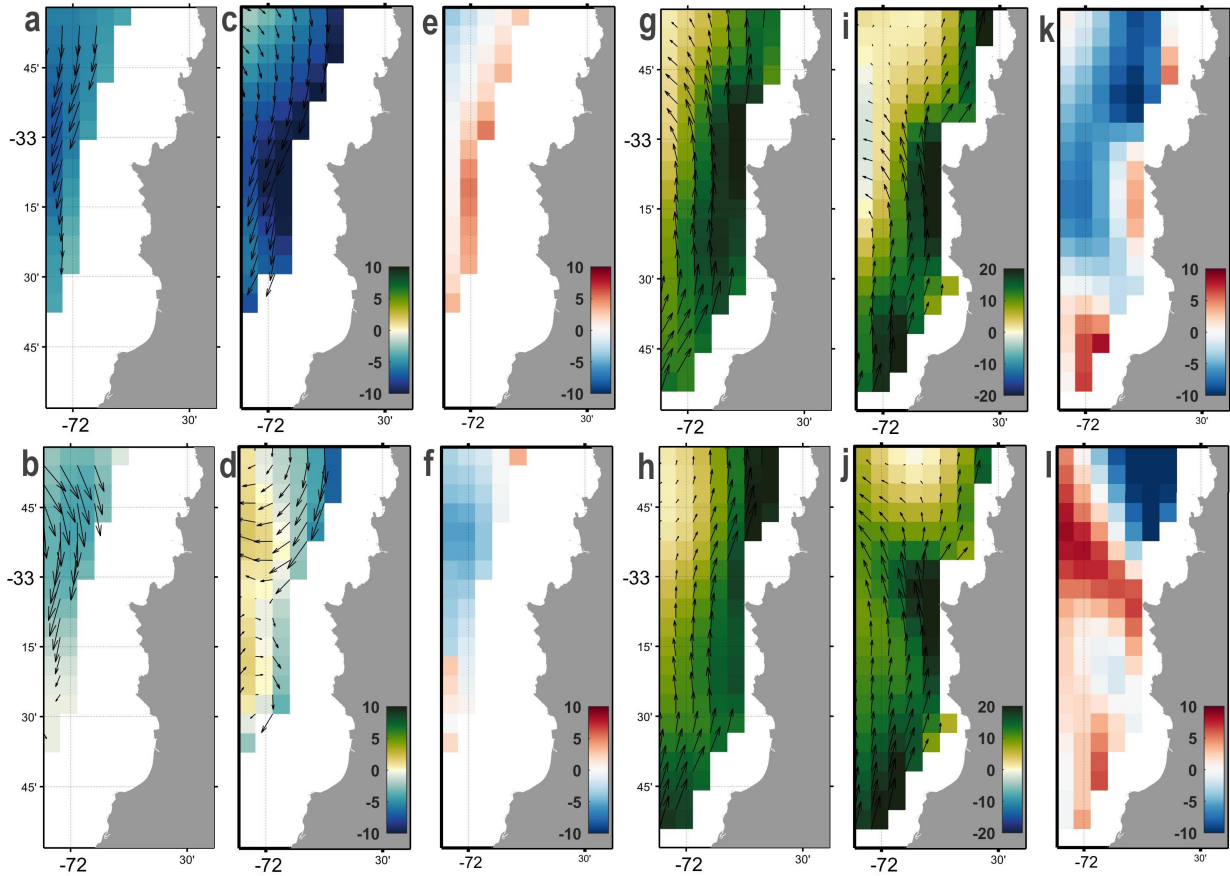


Figure 12: Seasonal average of alongshore current (cm/s) in bottom (a-f) and surface layers (g-l) integrated vertically (150-200 meters and 0-20 meters, respectively) during summer (upper panels) and winter (bottom panels). The figure compares the HRM2 (c-d and i-j) downscaled to the LRM0 (a-b and g-h) resolution. For difference between HRM2 and LRM0 (e-f and k-l), the positive (negative) values indicate a higher (lower) speed in the HRM2 than LRM0. The arrows depicts the flow speed direction.

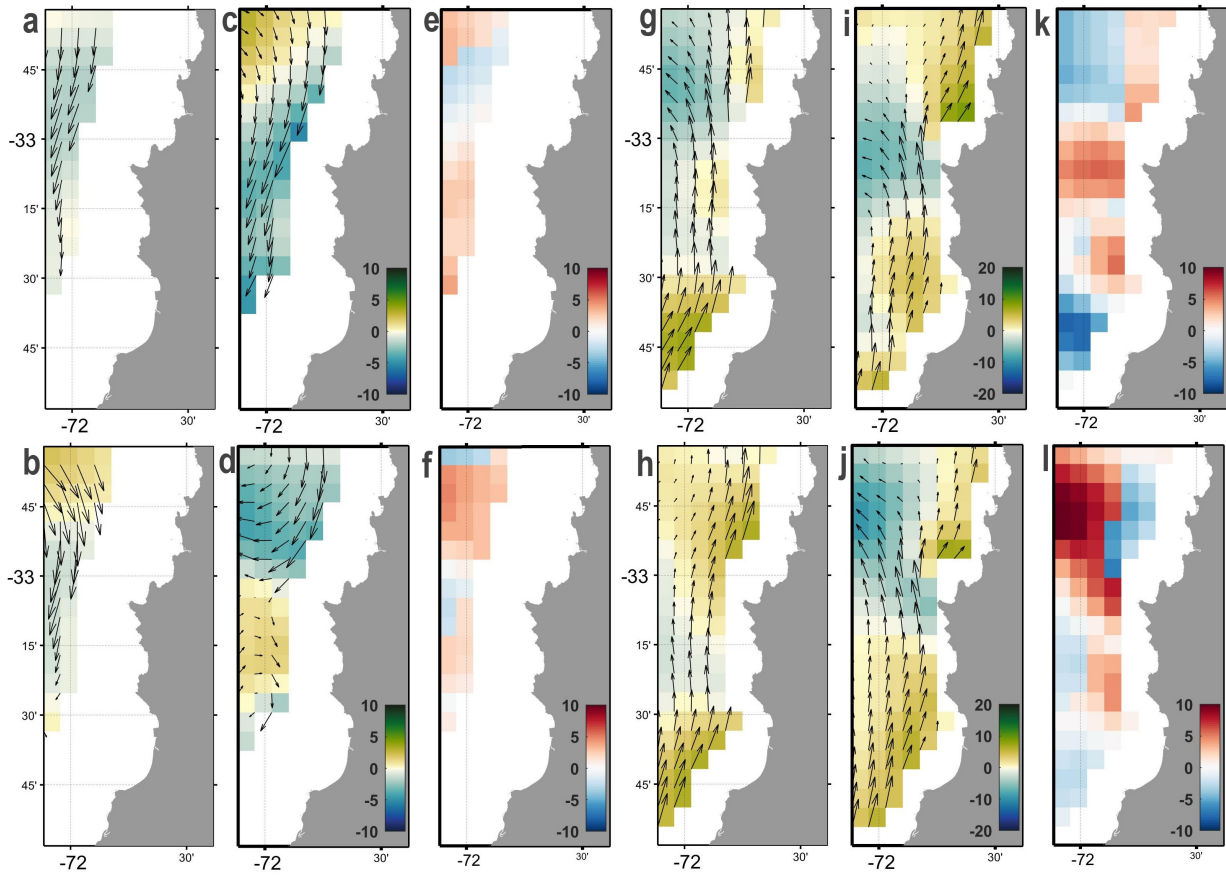


Figure 13: Seasonal crossshore current (cm/s) in bottom (a-f) and surface layers (g-l) integrated vertically (150-200 meters and 0-20 meters, respectively) during summer (upper panels) and winter (bottom panels). The figure compares the HRM2 (c-d and i-j) downscaled to the LRM0 (a-b and g-h) resolution. For difference between HRM2 and LRM0 (e-f and k-l), the positive (negative) values indicate a higher (lower) speed in the HRM2 than LRM0. The arrows depicts the flow speed direction.

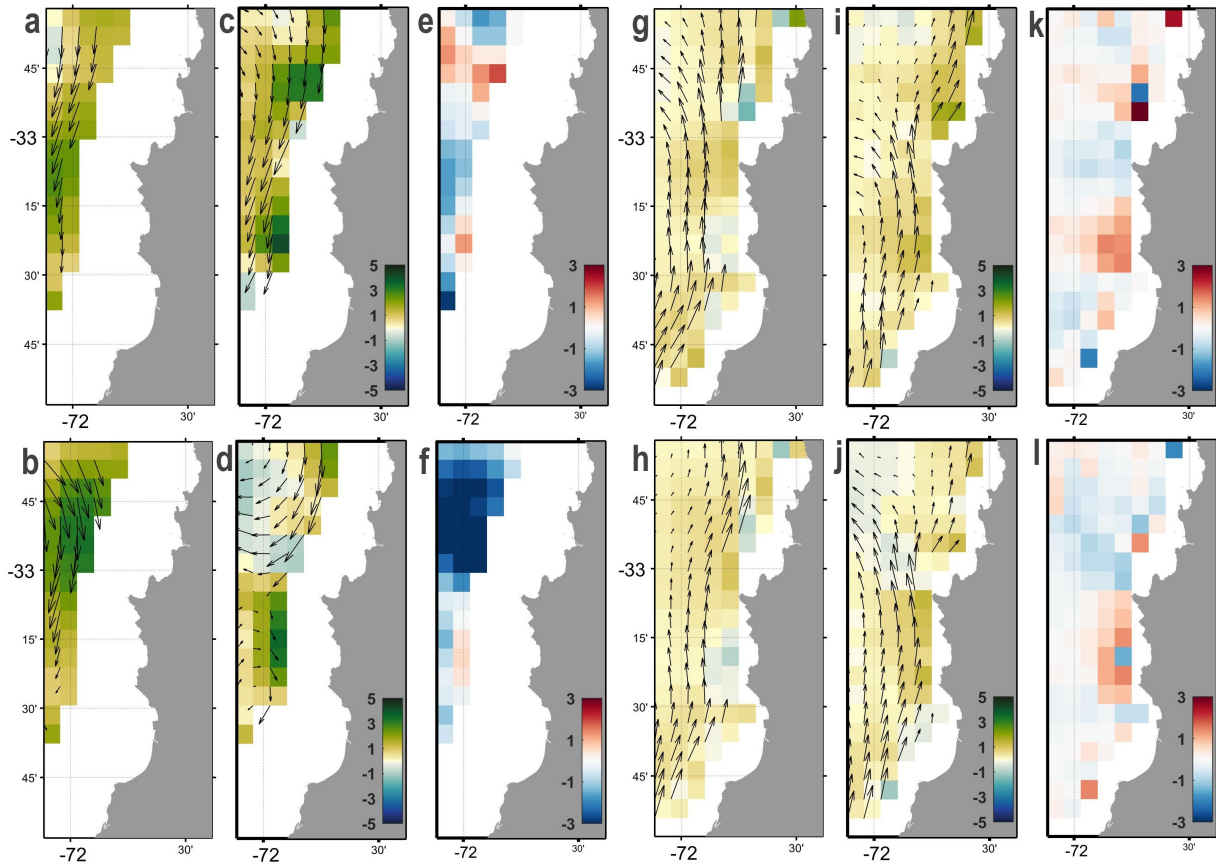


Figure 14: Seasonal average of vertical current (m/d) in bottom (a-f) and surface layers (g-l) integrated vertically (150-200 meters and 0-20 meters, respectively) during summer (upper panels) and winter (bottom panels). The figure compares the HRM2 (c-d and i-j) downscaled to the LRM0 (a-b and g-h) resolution. For difference between HRM2 and LRM0 (e-f and k-l), the positive (negative) values indicate a higher (lower) speed in the HRM2 than LRM0. The arrows depicts the flow speed direction.

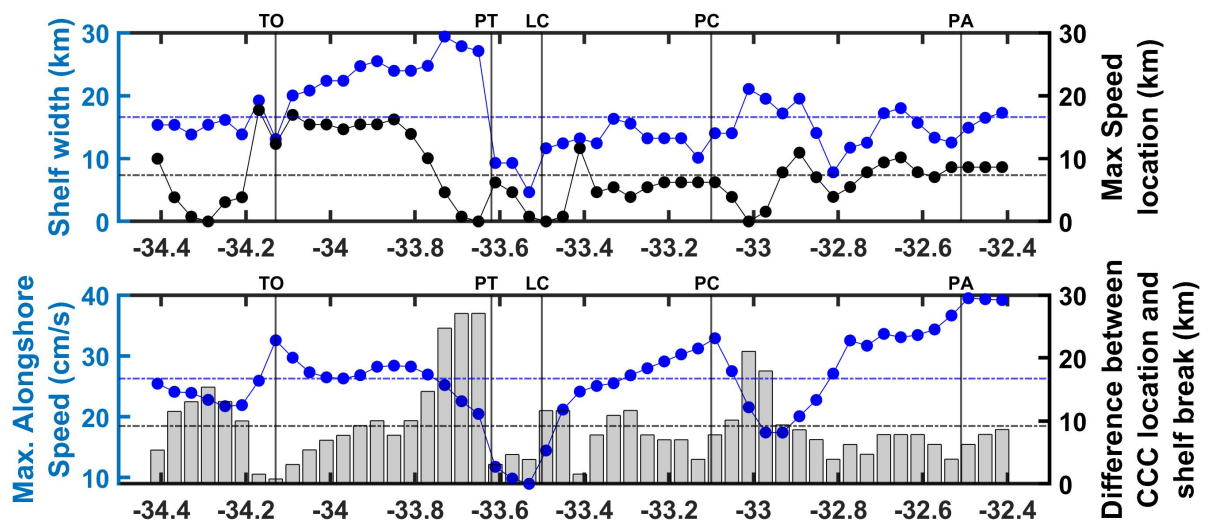


Figure 15: Upstream description of the CCC along the coastline and on front the sites TO, PT, LC, PC and PA (vertical black lines) within the HRM2. The upper panel identified the offshore location of maximum speed of CCC (black dots) and determine its distance with the shelf break (blue dots). The lower panel, The distance between the CCC and the shelf break (bars) associated with the maximum velocity of CCC. The horizontal dashed lines indicate the mean values.

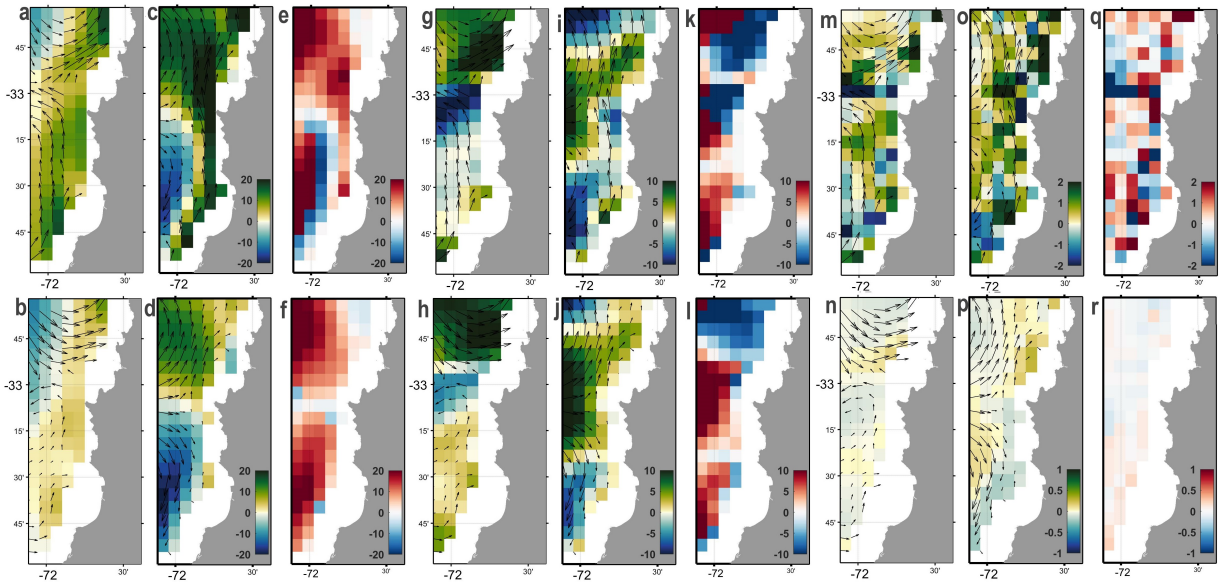


Figure 16: Mean circulation in the surface (0 - 20 meters; upper panels) and bottom layers (30 - 70 meters; lower panels), during a favorable upwelling period (16 to 27 December in nineteenth year). In order, the alongshore (cm/s) and crossshore current (cm/s) and the relative vorticity normalized (ζ/f) in the LRM0 (a-b, g-h and m-n, respectively) and HRM2 (c,d, i-j, and o-p, respectively). For difference between HRM2 and LRM0 (e-f, k-l and q-r, respectively), the positive (negative) values indicate a higher (lower) magnitude in the HRM2 than LRM0. The upper panel represent the summer season and the lower panel the winter. The HRM2 was downscaled to the LRM0 resolution. The arrows describe the flow speed direction.

Table 6: Mean circulation in the LRM0 and the HRM2 in summer, winter and annually. The HRM2 was downscaled to the LRM0 resolution. The season were integrating the upper layer (0-20m) and the lower layer (150-200m).

	V (cm/s)	U (cm/s)	W (m/d)	Speed (cm/s)	T (°C)	Density (kg/m ³)	ζ/f
Summer Integrated 0-20 m							
LRM0							
Mean	12.1	-0.6	0.3	12.2	15.6	1025.1	-0.06
SD	5.2	3.6	0.47	4.3	0.7	0.1	0.04
Range	1.2-22.2	-6.8-7.3	-1.6-2.1	3.8-20.6	13.8-16.6	1024.9-1025.5	-0.12-0.04
HRM2							
Mean	9.3	0.2	0.4	10.5	16.2	1025	-0.06
SD	7.3	3.4	0.5	6.5	0.8	0.2	0.07
Range	-2.2-24.1	-7-8.6	-0.8-1.9	1.8-24.5	14.2-17.3	1024.4-1024.8	-0.23-0.15
Summer Integrated 150-200 m							
LRM0							
Mean	-5.6	-1.1	1.4	5.7	10.6	1027.1	-0.01
SD	1.1	0.6	0.8	1.0	0.03	0	0.01
range	-7.3- -3.2	-1.9-0	-0.5-2.6	3.3-7.6	10.5-10.6	1027.1	-0.03-0.01
HRM2							
Mean	-7.3	-1.1	1.3	7.7	10.3	1027.1	0.04
SD	2.4	2.1	1.1	2.4	0.03	0	0.02
Range	-12.5-1.3	-5.2-3.3	-0.8-4.1	3.5-10.4	10.3-10.4	1027-1027.1	0.01-0.09
Winter Integrated 0-20 m							
LRM0							
Mean	11.3	1.6	0.2	10.8	13	1025.6	-0.05
SD	5.5	2.3	0.3	5.4	0.3	0.1	0.03
Range	0.9-21.8	-2.1-6.5	-1.2-1	1.8-22.7	12.2-13.4	1025.5-1025.8	-0.12-0.001
HRM2							
Mean	11.2	-0.3	0.2	12	13.3	1025.6	-0.02
SD	5.6	3.7	0.4	5.3	0.3	0.10	0.07
Range	-0.7-23.5	-9.4-7	-1.1-1.3	1.2-23.9	12.5-13.6	1025.5-1025.7	-0.15-0.18
Winter Integrated 150-200 m							
LRM0							
Mean	-2.7	0	2	3	10.5	1027.1	0.01
SD	1.2	1	0.7	1.1	0.1	0	0.01
Range	-4.2- -0.4	-1.3-2.1	0.8-3.3	0.6-4.2	10.4-10.6	1027.1	-0.02-0.03
HRM2							
Mean	-1.3	-1.3	0.6	3	10.2	1027.1	0.04
SD	2.3	1.7	1.1	1.6	0.1	0	0.03
Range	-6.9-1.9	-4.2-1.9	-1.0-3.5	0.8-7.3	10.1-10.3	1027.1	-0.03-0.09
Annual Integrated 0-20 m							
LRM0							
Mean	10.3	0.4	0.3	10	14.2	1025.4	-0.05
SD	5	2.3	0.4	4.7	0.5	0.1	0.02
Range	-0.3-18.5	-3.2-6	-1.1-1.4	1.7-19.1	12.9-14.8	1025.3-1025.7	-0.10-0.02
HRM2							
Mean	7.5	0.7	0.4	8.4	14.7	1025.3	-0.05
SD	6.1	2.7	0.3	5.6	0.5	0.1	0.06
Range	-0.7-22.6	-3.7-9.1	-0.7-1.7	1.7-23.2	13.3-15.3	1025.2-1025.6	-0.21-0.21
Annual Integrated 150-200 m							
LRM0							
Mean	-4	6 -0.6	1.8	4.8	10.5	1027.1	0
SD	1.1	0.9	0.6	1.1	0	0	0.01
Range	-5.8- -2.3	-2.1-1.4	0.4-2.8	2.5-6.2	10.4-10.6	1027.1	-0.01-0.02
HRM2							
Mean	-5.6	-0.5	1.2	6.1	10.3	1027.48	0.03
SD	2.2	2.0	0.8	2.1	0	0	0.02
Range	-10.2- -1	-4.2-3.6	-0.2-3.2	3-10.6	10.2-10.4	1027.1	0.01-0.09