Wind-driven ocean synoptic variability off south-central Chile

Catalina Aguirre\textsuperscript{a}, René D. Garreaud\textsuperscript{a} & José A. Rutllant\textsuperscript{a,b}

\textsuperscript{a} Departamento de Geofísica, Universidad de Chile. Blanco Encalada 2002, Santiago, Chile.

\textsuperscript{b} Centro de Estudios Avanzados en Zonas Áridas (CEAZA), Universidad de la Serena. Raúl Bitrán s/n, Colina El Pino, La Serena, Chile.
Abstract

The effect of the synoptic wind variability upon the ocean off south-central Chile is investigated using an ocean model. We focus our analysis in austral summer, when the regional wind experience significant variability superimposed in a mean, upwelling favorable flow. To evaluate the nature and magnitude of these effects, we performed two identical simulations except for the wind forcing: the *synoptic* run used daily QuikSCAT wind stress and the *climatological* run used long-term monthly mean QuikSCAT wind-stress. No major differences between simulations were found in the mean currents and surface geostrophic EKE, and both agree well with those observed in Southeast Pacific. Nevertheless, significant more ageostrophic EKE is found in the *synoptic* simulation, impacting on the total surface EKE and diffusivities, which are twice as large as their *climatological* counterparts. Summer mean SSTs are also similar in both simulations in agreement with observations, but SST variability within the first 50 km offshore is larger in the *synoptic* run, suggesting a rather linear response of the surface ocean to cycles of southerly wind strengthening and relaxation. A high-resolution simulation shows that at the beginning of a southerly wind event vertical mixing acts first eroding stratification, allowing the onset of vertical velocities leading to a cooling at mid levels. During northerly wind conditions, the major warming process is the horizontal advection by the surface onshore flow in the upper 30 m. Thus, the magnitude of the coastal warming seems controlled by the duration of the wind reversal rather than by its intensity.
1. Introduction

Upwelling of cold, nutrient rich waters along much of the Chilean coast induces high primary productivity and maintains one of the major fisheries of the world [e.g., Rutillant and Montecino, 2002; Food and Agriculture Organization, 2004]. Coastal upwelling is in turn forced by alongshore, southerly (equatorward flow) winds, being particularly strong around major capes at 30°S (Punta Lengua de Vaca), 33°S (Punta Curaumilla) and 37°S (Punta Lavapié) because of topographic / bathymetric effects [Figueroa and Moffat, 2000].

To the north of 33°S upwelling-favorable winds prevail year round whilst off south-central Chile (35-40°S) monthly mean surface winds alternate between moderate northerlies in winter to strong southerlies in summer (DJF), following the migration of the south Pacific anticyclone [Saavedra and Foppiano, 1992]. South of 35°S synoptic variability of the winds is very pronounced in winter –because of the passage of extratropical atmospheric disturbances- but it is still present in summer [e.g., Sobarzo et al., 2010] due to the quasi-weekly occurrence of southerly coastal jet events alternated with periods of weak southerlies or even northerly flow [Garreaud and Muñoz, 2005]. The jet events are characterized by an elongated maximum of surface southerly winds (over 15 m s\(^{-1}\)) generally rooted in Punta Lavapié and extending northward several hundred of km [Muñoz and Garreaud, 2005]. Consistent with the mean, upwelling-favorable winds and the occurrence of strong coastal jets, the region around Punta Lavapié exhibits the highest primary productivity along the Chilean coast during summer [e.g., Montecino et al., 2006].

Although the basic features of coastal upwelling have been described, including their regional distribution and seasonal variability [e.g., Figueroa and Moffat, 2000; Leth and
Shaffer, 2001; Blanco et al., 2001; Sobarzo et al., 2007], there is less information on the surface ocean response to atmospheric jet events / northerly wind reversals, partly because of the scarcity of in situ data. Using satellite observations Renault et al., [2009] studied the effect of the atmospheric low-level jet on sea surface temperature (SST) off central Chile (30°S), but the blind zone close to the coast was an important limitation to capture the upwelling response to the wind forcing which is particularly important within the first 20 km off the coast [e.g., Perlin et al., 2007; Aiken et al., 2008]. Modeling efforts have also been conducted to understand upwelling and ocean circulation off central Chile [Batteen et al., 1995; Leth and Shaffer, 2001; Mesias et al., 2001; Mesias et al., 2003; Leth and Middleton, 2004, Aiken et al., 2008]. Most of these studies have used climatological (long-term-mean) wind stress data to force an ocean model, perhaps loosing an important source of variability to the south of 35°S. Only Mesias et al., [2001] and Aiken et al., [2008] used daily wind stress data, but the former utilized a coarse horizontal resolution (2.5° × 2.5° lat-lon grid) and the latter applied a spatially homogeneous wind stress field recorded at a coastal station. The lack of spatial structure of the near-shore winds in both studies may strongly influence the patterns of the circulation and SST [Capet et al., 2004].

In the present work, we forced the Regional Ocean Model (ROMS) [Shchepetkin and McWilliams, 2005] off Chile with daily QuikSCAT-derived wind stress [Piolle and Bentamy, 2002], allowing us to analyze the effects of the day-to-day changes in wind upon ocean circulation. Although the model domain extend along most of the Chilean coast (25°-45°S) and the runs span 8 full years, we focus our study off south-central Chile during austral summer, when there is significant synoptic variability in the wind superimposed on a mean, upwelling-favorable flow. To gauge the nature and magnitude of these effects, we
have done a twin simulation, but using the long-term-mean monthly QuikSCAT wind-
stress, and compared the results of both simulations with observations whenever possible.
To further clarify the physics of some results in the wind-variable simulation relative to the
constant-wind case, we have used our synoptic simulation to describe in detail the three-
dimensional structure of the coastal upwelling (downwelling) response to a strong coastal
jet event (wind reversal period) that occurred in December 2000 (January 2001) off Punta
Lavapié.

The rest of the paper is organized as follows: section 2 describes the model, its
configuration and the experiments performed; section 3 focuses on comparisons between
simulations and validation using available observations. A case study of upwelling
(southerly jet event) and downwelling (wind reversal period) is presented in section 4.
Finally, section 5 summarizes the main results and conclusions of this study.

2. Model configuration and experimental setup

The model used in this study is the Regional Oceanic Modeling System (ROMS), a split-
explicit, free surface, topographically-following-coordinates oceanic model that solves the
primitive-equations assuming hydrostatic and incompressible conditions [Shchepetkin and
McWilliams, 2005]. Radiation conditions are used in open boundaries in order to treat the
inward and outward information fluxes separately. When information fluxes are outward
the boundary is passive and when they are inward the boundary is active, nudging toward
its lateral boundary conditions [Marchesiello et al., 2001]. Near the open boundaries a
sponge layer is used, which is a region of increased horizontal viscosity. The vertical mixing is parameterized using a K-Profile Parameterization (KPP), a non-local closure scheme based on the boundary layer formulation by Large et al., [1994].

To compare the effect of the synoptic-scale wind stress variability relative to a smooth seasonal variation, we performed two simulations (termed as climatological and synoptic) that only differ in their surface wind forcing. The model domain spans from 25° to 45° S and from 70° to 80° W, where the northern, western and southern boundaries are open (Figure 1). The horizontal resolution is 1/10°, which corresponds to ~7.9-10.1 km, with a 248 × 101 grid points. The bathymetry was extracted from the 2’ resolution ETOPO database [Smith and Sandwell, 1997]. In spite of the fact that a pressure gradient scheme prevents large errors in the computation of pressure gradients [Shchepetkin and McWilliams, 2003], the bathymetry was smoothed in order to maintain a “slope parameter” $r = \nabla h / 2h < 0.25$ [Beckmann and Haidvogel, 1993]. The vertical grid has 32 vertical levels and stretching parameters of $\theta_s=7$, $\theta_b=0$ and $hc=10$ m [Song and Haidvogel, 1994] allowing a good representation of the surface coastal ocean. The time step was 13.3 min and outputs were averaged over each day.

The climatological run (CliSim) used long-term monthly means wind stress as surface boundary condition, calculated from QuikSCAT data between 2000 and 2007. This simulation was integrated for 10 years of 360 days each, repeating the mean annual cycle of the surface stress interpolated linearly to each time step. Statistics are analyzed for the last 8 years (the first 2 years are considered model spin-up). The synoptic run (SynSim) had an initial two year integration using daily QuikSCAT for the year 2000 (spin up period),
followed by an integration forced with daily QuikSCAT data between 2000 and 2007 (thus, we also computed model statistics using 8 years).

In both simulations, climatological monthly means of air temperature, humidity, precipitation and wind from the Comprehensive Ocean-Atmosphere Data Set (COADS) [Da Silva et al., 1994] were used to calculate surface heat and freshwater fluxes. Thus, our synoptic run only includes the effects of the variable wind on ocean dynamics. Likewise, in both runs the initial condition and lateral boundary conditions were climatological monthly means obtained from the World Ocean Atlas 2005 [Locarnini et al., 2006; Antonov et al., 2006].

A second set of climatological (forced with long-term monthly mean winds) and synoptic simulations (forced by daily Quikscat winds) were performed with higher resolution (~4.4 km) to compare effects on surface dispersion and to study a well-defined jet event (wind reversal period) occurred in December 2000 (January 2001). We explicitly refer to these simulations as high-resolution when used in our analysis. In these simulations the domain is smaller (33° - 40° S and 71.3° - 77.5° W, Figure 1) and the vertical grid has 20 vertical levels. As in the previous simulations lateral boundary conditions are obtained from WOA, and the atmospheric data used to calculate surface fluxes are obtained from COADS. These runs are independent from the previous simulations and both have 2 years of spin up, and December 2000 and January 2001 are analyzed from the synoptic simulation. The time step was 6.6 min and the output was averaged hourly.
3. Inter-simulation comparison and validation

3.1 Mean currents

Figure 2 shows the summer mean surface ocean meridional velocity obtained from CliSim and SynSim, complemented with depth-longitude cross sections at 30.3° and 36.5°S. Overall, both fields are similar and reproduce the major features of the circulation in the southeast Pacific as identified by Strub et al., [1998]. The Chile-Perú surface Current (CPC, also known as the Humboldt Current) is the eastern branch of the subtropical South Pacific gyre. This flow has been described using altimetry data and exhibits higher equatorward intensity during summer [Fuenzalida et al., 2008]. In our simulations the CPC intensifies during summer with a mean value of ~15 cm s\(^{-1}\). This speed is somewhat higher than those characterizing satellite-derived surface currents in this season that reach values larger than 10 cm s\(^{-1}\) in the jet core. However, it is important to remind that satellite-derived currents only account for the geostrophic component of the currents. The Chile Coastal Current (CCC) flows equatorward as an ocean coastal jet (speeds larger than 25 cm s\(^{-1}\)) during summer and it is related to upwelling dynamics [e.g., Aiken et al., 2008]. Both CliSim and SynSim clearly capture the mean CCC with a cross-shore scale of about 30 km off south-central Chile and flow speed in the 15-25 m s\(^{-1}\) range. The simulated ocean jet exhibits a coastal separation near Punta Lavapié, an important feature that has recently observed by hydrographic data [Letelier et al., 2009] confirming previous simulations using the Princeton Ocean Model [Mesias et al., 2001; Mesias et al., 2003; Leth and Middleton, 2004].
The Perú-Chile Countercurrent (PCCC) is a surface poleward flow located west of the CCC, about 100-300 km offshore [Strub et al., 1995]. This current has been described using satellite-derived sea-surface height anomalies data, showing a maximum during austral spring and a minimum in fall. In our simulations a mean poleward surface current consistent with the observed location of the PCCC is found during summer at around 27° S with speeds of about 5 cm s\(^{-1}\). Farther south (~30° S) another patch of weak poleward surface flow is observed. Finally, the Perú-Chile Undercurrent (PCU) is a well-defined poleward flow located over the continental shelf and continental slope [Shaffer et al., 1997; Shaffer et al., 1999; Pizarro et al., 2002]. The vertical sections of our simulations reproduce the PCU with summer means of >10 cm s\(^{-1}\) at 30.3° S (Figure 2c,d) and ~5 cm s\(^{-1}\) at 36.5° S (Figure 2e,f). The mean value obtained at 30.3° S (12 cm s\(^{-1}\)) agrees well with the annual mean value of 13 cm s\(^{-1}\) obtained for a six year period of current measurements near the PCU core over the slope at 30° S [Shaffer et al., 1999].

Further verification of the model performance in simulating the mean currents is provided in Figure 3 by the summer mean vertical profiles of zonal (u) and meridional (v) components along with their observational counterparts obtained from two Acoustic Doppler Current Profilers (ADCP). The first ADCP is located ~13 km from the shore on the continental slope at 30.3°S (COSMOS). Its summer mean current was obtained considering only 2004, 2005 and 2006. The second ADCP is ~20 km from the shore on the continental shelf at 36.5°S W (S18). In this case the summer mean considered only January and February 2009. As found before, mean profiles from CliSim and SynSim are very similar (except for the v-component in S18), their shapes agree well with the observations and they capture the reversal of the currents at different depths. The agreement is better in
S18, within our target area, while farther north (COSMOS) the model profiles show an overestimation of the northward and eastward flow components of the surface current (Figure 3a,b).

3.2 Eddy Kinetic Energy (EKE)

Given the complexity of EKE generation, surface EKE comparison between a model and data is a stringent test of the upper-oceanic circulation [Capet et al., 2008]. To test our simulations, we used the altimetry-derived gridded product (1/3°) of the current anomalies of the Archiving, Validation and Interpretation of Satellite Oceanographic Data (AVISO) spanning 2000 to 2007 to obtain the observed geostrophic EKE. The model outputs allow calculation of both total EKE (time average of the sea surface velocities anomalies) and geostrophic EKE (obtained from the sea-surface height anomalies).

In the offshore region, the spatial pattern of geostrophic EKE is similar in both simulations and in agreement with satellite data (Figure 4a,b,c). Higher summertime values are found in a band rooted along the coast between 38-30°S. This region corresponds to the energetic coastal transition zone off Chile described by Hormazabal et al., [2004] using 7.5 years of AVISO data. The model and observed geostrophic EKEs show higher values in a band extending offshore in a SE-NW direction from Punta Lavapié, consistent with the jet separation. The offshore values of the simulated geostrophic EKE are, however, 40-60% larger than the observations, an overestimation also found in other upwelling systems [Veitch et al., 2010; Capet et al., 2008]. This could be attributed to an overestimation of eddy generation at the CPC via baroclinic instability [Leth and Middleton, 2004] and/or to
nearshore eddy generation propagating westward [Hormazabal et al., 2004], as a major sources of geostrophic EKE offshore. In particular, the observed EKE minima close to the coast, could be attributed to the nearshore eddy scale (<50 km), not resolved in satellite data [Ducet et al., 2000]. Furthermore, to the south of 37°S the CliSim and SynSim exhibit high geostrophic EKE (≈ 150 cm² s⁻²) in a coastal strip of about 70 km wide (coincident with the CCC) that is not present in the observations probably because AVISO is not reliable within the first ~50 km offshore. We interpret the agreement in geostrophic EKE between CliSim and SynSim as the result of intrinsic variability of the ocean arising from instabilities in the currents (especially CPC and CCC) regardless of the presence (or absence) of high-frequency variability in the wind forcing. Analysis of the energy conversion in a numerical study made by Leth and Shaffer [2001] showed that baroclinic instability is indeed a major mechanism for the generation of meanders and eddies (quasi-geostrophically balanced) off central Chile.

In contrast to the geostrophic EKE, the ageostrophic kinetic energy (obtained as the difference between the full and geostrophic EKE) is significantly different between both simulations (Figure 4d,e). The synoptic simulation exhibits a coastal band, collocated with the CCC, of high ageostrophic EKE (>120 cm² s⁻²) from 35 to 45°S accounting for more than a half of the total surface EKE. A coastal band of high ageostrophic EKE is also found in the climatological simulation, but smaller and weaker than in SynSim (only 20-30% of the total EKE). A similar pattern was found in the California Current System by Marchesiello et al., [2003]. Thus, variable winds significantly enhance the level of eddy kinetic energy in the coastal zone off south-central Chile over a background EKE arising from the intrinsic current’s instability.
3.3 Lagrangian properties of the surface flow

We showed that forcing ROMS with synoptic-scale variable winds makes little difference in the resulting mean currents and geostrophic EKE compared with a smooth (climatological) forcing, but it does significantly increase the ageostrophic EKE along the coast. To further quantify this effect, here we study the properties of the surface flow by tracking particles released at the surface ocean using a Lagrangian drifter-tracking code, developed by X. Capet\(^1\). The offline version of the code advects numerical floats from stored ROMS outputs of the velocity field. For more details of the algorithm the reader is referred to Carr et al., [2008]. For this analysis we used the ROMS outputs from the high-resolution simulations during December.

The drifter-tracking module was run with a time-step of 400 s and drifters positions were recorded hourly during one month. Drifters were released simultaneously at the surface within a strip of \(\sim 250 \text{ km} \) from the shore, separated by 5 km in the cross-shore direction and 10 km in the along-shore direction. As an example, Figure 5 shows the trajectories of some virtual floats 23 days after release. The drifters in the synoptic simulation were clearly more dispersed and reached farther offshore than their counterparts in the climatological simulation.

In a Lagrangian frame, the velocity vector of each drifter \((\mathbf{U}(t))\) can be written as \(\mathbf{U} + \mathbf{u}'(t)\), where \(\mathbf{U}\) is the mean flow, representative of the large spatial scale and calculated by averaging the velocity component over each drifter trajectory, and \(\mathbf{u}'\) is the time varying

\(^{1}\text{Roms Offline Floats (Roff) available at http://www.atmos.ucla.edu/~capet/Myresearch/my_research_floats.html}\)
perturbation [e.g., *Kundu and Cohen*, 2000]. For the zonal \((u')\) and meridional \((v')\) components of the eddy velocity we calculated the integral time scale \(T=(T_u, T_v)\) as

\[
T_{u,v} = \frac{1}{R_{u,v}(0)} \int_0^\Gamma R_{u,v}(\tau) d\tau
\]

where \(R_{u,v}(\tau)\) is the Lagrangian autocovariance function and \(\Gamma\) is the lag at first zero crossing [e.g., *Chaigneau and Pizarro*, 2005]. The autocovariance and integral times were computed for each drifter spending the complete month inside the domain (1203 and 1083 drifters in the climatological and synoptic runs, respectively). The integral length scale \(L=(L_u, L_v)\) and the horizontal diffusivity was obtained using \(K=(K_u, K_v)\) are given by [e.g., *Chaigneau and Pizarro*, 2005].

\[
L = \sqrt{\overline{u'^2}} \cdot T
\]

\[
K = \overline{u'^2} \cdot T
\]

The results of these statistics are presented in Table 1. On the basis of a \(t\)-test, with a significance level of 5\%, we obtained that all mean values are statistically different between the climatological and synoptic simulations. Most importantly, the variances and diffusivities in the synoptic simulation are twice as large as their climatological counterparts. The \(K\)-values in our synoptic run are also similar to the diffusivities obtained by *Chaigneau and Pizarro* [2005] tracking actual drifters on the CPC region (24\(^\circ\)-34\(^\circ\)S; 70\(^\circ\)-82\(^\circ\)W; see Table 1). These results highlight the relevance of using variable (synoptic) winds to force the ocean models if one aims at describing the turbulent component of the
upper-ocean circulation, like in biological modeling where dispersal of chemical and larvae is a key issue.

3.4 Sea surface temperature

The long-term mean of the summer SST observed from the Advanced Very High Resolution Radiometer (AVHRR, Pathfinder V5, years 1985-2001) and those simulated (8-years) in the climatological and synoptic runs are presented in Figure 6. Both runs are able to reproduce a realistic pattern of the summer mean field of SST, with warmer waters offshore and colder waters at the coastal region, superimposed on a general north-south decrease in SST. The simulated summer upwelling front (i.e., the offshore limit of the coastal high SST gradient) is about 100 km from the coast in agreement with the satellite observation [Letelier et al., 2009].

Close to the coast of south-central Chile the ClimSim and SynSim SST summer means are 2-3°C colder than satellite observations (the differences reduce to ~1°C during other seasons, not shown). A colder bias along the coast has also been reported in climatological simulations of the Benguela upwelling system [Penven et al., 2001; Veitch et al., 2010] and the Perú upwelling system [Penven et al., 2005], and attributed to the low spatial resolution of the wind forcing. In our case, the wind stress over the first ~50 km off the coast was simply extrapolated from nearest QuikSCAT ocean grid point (0.5° grid spacing), typically resulting in an overestimation of the wind speed, and hence Ekman transport, in the near shore coastal strip [Capet et al., 2004]. Assumption of alternative drop-off of near coastal
wind is difficult to justify in our case because of the lack of observations in the coastal strip
[Garreaud et al., 2010].

Simulated SSTs are also 1-2°C colder than in-situ measurements as summarized in Table 2 and illustrated in Figure 7 by a few daily time series at selected stations. In spite of the cold bias, the model SST (synoptic run) follows quite well the observed SST, with multi-year correlation coefficients around 0.4-0.5 (Table 2) but as high as 0.7 in individual years. This is a remarkable result considering that (a) coastal stations are typically located in small embayments so their records are influenced by local-scale processes (e.g., upwelling shadows) hardly resolved by the model, (b) the lack of a realistic (daily) air-sea fluxes, highlighting the prominent role of wind-driven upwelling cycles in coastal SST variability, and (c) any possible errors propagating from the model’s lateral boundary conditions.

Somewhat expected, the synoptic run exhibits significantly higher summer standard deviation of SST (σSST) than the climatological run (Figure 8). The SynSim values of σSST are close to the observations (Table 2). The major differences in SST variability are found in a coastal strip off south-central Chile, coincident with the region of maximum wind variability [Renault et al., 2009], where σSST in SynSim is about two times larger than its climatological counterpart. The cross-shore scale of the σSST maximum is about 50 km and it indicates the region of strong coupling between the wind and SST via upwelling dynamics [e.g., Perlin et al., 2007]. Farther offshore, σSST is similar between both runs and smaller than observed (Table 2), suggesting that SST day-to-day variability over the open ocean is more controlled by air-sea fluxes (whose variability is not included in our simulations). The weak SST variability in the climatological simulation is related to
the generation of meanders and filaments along the coast that, as discussed before, arises by
the intrinsic instability of the ocean currents [Marchesiello et al., 2003].

3.5 Summary of CliSim and SynSim differences

It is interesting to note the similarities in the spatial pattern of the mean SST and currents
between both simulations in contrast with the much higher level of coastal SST variability
and ageostrophic EKE in the synoptic run, indicative of rather linear responses of the
surface ocean to cycles of southerly wind strengthening (upwelling favorable) and
weakening (downwelling favorable in the case of northerly winds). Indeed, Figure 9 shows
the scatter plot between simulated (SynSim) SST and meridional wind ($v_{\text{wind}}$, daily values,
one-day lag) off Punta Lavapié, from where a negative linear relationship is evident for
moderate to strong southerly winds ($v_{\text{wind}} > 5 \text{ m/s}$). In Figure 9 we also included the time
evolution (in the SST-$v_{\text{wnd}}$ phase space) of a two consecutive upwelling / downwelling
cycles in December 2000 (analyzed in detail next). When the meridional wind relaxes
($v_{\text{wnd}} < 5 \text{ m/s}$) or reverses ($v_{\text{wnd}} < 0$) SST increases, sometimes abruptly, but there is
considerable more dispersion in this range. In any case, synoptic alternance of strong/weak
southerlies will produce large SST variability without inducing a significant cold (or warm)
bias relative to a constant wind case. Notably, the highest SST values don’t occur during
days of strong northerlies, presumably because the enhanced vertical mixing in the upper
ocean that scales with the wind speed regardless of its direction. The scatter plots between
$v_{\text{wnd}}$ and surface ocean currents ($u_{\text{sfc}}$, $v_{\text{sfc}}$) off Lavapié are similar to the SST case (not
shown), with an approximately linear relationship in the moderate and high range of meridional wind and more dispersion for the low or reversed range.

Given these approximately linear relationships, the spatial response of the surface ocean to the wind variability can be described by regressing the SST and surface current fields upon the meridional wind off Punta Lavapié (one-point correlation maps in Figure 10). The correlation of the local wind with the meridional wind elsewhere shows the characteristic pattern of the atmospheric coastal jet events during summertime, rooted off Lavapié and extending offshore in a NW direction [Garreaud and Muñoz, 2005; Renault et al., 2009].

The correlation between SST and $v_{wnd}$ is larger than -0.5 in a coastal strip about 50 km wide under the area of maximum wind anomalies, and rapidly decays offshore. This coastally confined response of SST to variable winds is consistent with our previous analysis of the $\sigma$SST field. The correlation between the meridional wind and the zonal component of the surface velocity shows maximum negative correlations values of about -0.6 at zero lag, indicating that the surface Ekman flow varies in close phase with the wind.

In contrast with the coastally confined SST response, the highest $v_{wnd} - u_{sfc}$ correlations under the strong southerlies extend from coast for at least 500 km offshore. The correlation between $v_{wnd}$ and surface meridional current is also coastally confined and implies acceleration (deceleration) of the Chilean coastal current (CCC) during jet events (wind reversal periods).
4. Case study

To obtain more details on the processes coupling the wind field and the surface ocean, we considered the austral summer 2000/2001. Daily time series of simulated SST (from SynSim) and meridional wind (from QuikSCAT data) averaged 50 km alongshore and about 20 km off Punta Lavapié are shown in Figure 11. Also included in this figure is the vertical velocity at 40 m depth, which clearly show a southerly-upwelling and northerly-downwelling pattern, consistent with the coastal location of the column. This period exhibits well defined synoptic variability of the surface wind, including a strong southerly coastal jet (17-20 December) and an extended period of wind reversal (northerly flow, 2-7 January). Each of these wind events produced transient SST excursions of about $\pm 2^\circ$C relative to the full period average ($\sim 12^\circ$C, see also Figure 6). Surface cooling (warming) also occurs during other days with moderate southerly (weak northerly) winds, although with variable intensity. In particular, the wind reversal on 24 December was as strong as the event in January, but the sea surface only recovered from a cold condition (caused by the previous southerly jet) to average conditions, suggesting that surface warming is more dependant on the duration of the low wind anomalies rather than on wind intensity. That explains, at least partially, the high dispersion in the $v$-SST plot when $vwind < 5$ m/s (Figure 9).

4.1 Coastal jet / upwelling period

To portrait the spatial extent of the surface changes between the jet event (17-20 December 2000) and the previous calm days (11-14 December) Figure 12 shows the wind and SST
differences between both periods. The southerly wind increased from 36°S to 42°S reaching values as large as 15 m s\(^{-1}\) on the 18\(^{th}\) and 19\(^{th}\) off Lavapié; farther north the coastal winds decreased but they remained blowing from the south. Consistently, SST decreased about 3°C under the jet core and increased about 2°C farther north. The major changes are observed within the first 30 km from the coast and no changes were observed beyond 100 km offshore. The simulated pattern of surface cooling/warming between the jet-minus-calm periods is in good agreement with the satellite derived SST fields (Figure 12c) in spite of a minor mismatch in the dates used for constructing the differences.

Longitude-depth sections averaged 50 km alongshore off Lavapié of the jet-minus-calm conditions for ocean \(T\), \(u\), and \(v\), are shown in Figure 13. The coastal cooling during the jet encompasses the upper 40 m of the ocean. The stronger southerlies also accelerate the upper-ocean currents near the coast. The zonal ocean velocity increases more than 10 cm s\(^{-1}\) in the upper 30-m layer with the flow increasing offshore. That leads to upper-ocean divergence in the near-coastal region ultimately compensated by upward vertical velocities. The zonal flow acceleration is consistent with the theoretical Ekman flow [e.g., Cushman-Roisin, 1994]

\[ U_{EK} = \frac{\tau_y}{\rho f H} \quad (4) \]

where \(\tau_y\) is the meridional wind stress, \(\rho\) is sea water density (~1025 kg m\(^{-3}\)), \(f\) is the Coriolis parameter and \(H\) is the Ekman layer depth. Considering \(H \sim 30\) m, the Ekman zonal velocity increased from about 2 cm s\(^{-1}\) during the calm period (\(\tau_y \sim 0.05\) Pa) to about 18 cm s\(^{-1}\) during the jet (\(\tau_y \sim 0.5\) Pa) in agreement with the model results. The equatorward coastal flow (CCC) also increased its value in more than 5 cm s\(^{-1}\) in the upper 20 m layer.
In order to diagnose the origin of the cooling during the jet event, a heat budget was performed using hourly outputs from the model and neglecting the contributions of the horizontal mixing because of their low values. Therefore, our heat budget is reduced to

$$\frac{\partial T}{\partial t} = -\left( u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} \right) + \frac{\partial}{\partial z} \left( k \frac{\partial T}{\partial z} \right) + Q $$

(5)

where $T$ is temperature, $u$, $v$ and $w$ are the zonal, meridional and vertical velocities respectively, $k$ is the vertical diffusion coefficient and $Q = Q_{SW} - Q_{LW} + Q_{SEN} + Q_{LAT}$ is the net air-sea heat flux.

Of particular relevance, Figures 14c,d show the divergence of the heat flux due to advection and vertical mixing, respectively. It is important to note that the zonal and vertical advectons, contributing to the heat flux divergence, are almost in balance, each one being one order of magnitude larger than the total heat flux divergence. During the calm conditions before the jet event (11-15 December) the vertical profile of temperature (Figure 14b) shows a strong thermocline at about 10-m depth. When the southerly wind starts to increase (early on 17 December) vertical mixing leads to a rapid cooling in the upper 20-m layer and a slight warming immediately below. The weaker stratification permits the development of strong vertical velocities leading to a mid level cooling due to advection divergence. One day later (when the $vwnd$ reaches a maximum) the cooling increased in the upper-layer (led by vertical mixing) and extended down to 50 m (led by net advection). On 19 December, still under strong wind, the upper 50 m column became nearly isothermal at about 10.5°C and both vertical mixing and net advection cease their cooling effect. The upper-ocean remains cool and weakly stratified for several days until the surface meridional
wind reverses \( v_{\text{wnd}} < 0 \) on 23 December producing a small downwelling event that raised the temperature up to pre-jet, near-average values.

Evaluating the contribution of the surface fluxes to the ocean temperature change is problematic because the model used monthly climatological values of air temperature, humidity, solar radiation and wind speed to compute those fluxes. Indeed, the surface fluxes used in the model changed less than 17\% between the calm period and the jet event, with little (if any) impact on the simulated ocean cooling, but one may wonder about their actual contribution. Of particular relevance is the surface evaporative cooling \( Q_{\text{LAT}} \) since it significantly increases during a jet event; the other surface fluxes are either small (net longwave radiation, sensible heat flux) or have little change between calm and windy days (shortwave radiation). Here we use a bulk formula

\[
Q_{\text{LAT}} = \rho_a L C_E \omega_s (q_s - q_a) \quad (6)
\]

where \( \omega_s \) and \( q_a \) are the wind speed and specific humidity near the interface ocean-atmosphere (COADS data); \( q_s \) is the saturation specific humidity, \( \rho_a \) is the air density and \( L \) is the latent heat of evaporation. With \( \omega_s \sim 5 \text{ m s}^{-1} \), \( q_s - q_a \text{(SST=13°C)} \sim 1.3 \cdot 10^{-3} \text{ g/Kg} \) results in \( Q_{\text{LAT,calm}} \sim 28.6 \text{ W/m}^2 \). During the jet event \( \omega_s \sim 15 \text{ m/s} \), \( q_s - q_a \text{(SST=11°C)} \sim 1.2 \cdot 10^{-3} \text{ g/Kg} \) and \( Q_{\text{LAT,jet}} \sim 75.5 \text{ W/m}^2 \). Considering a 15-m deep mixed layer (H), the cooling due to enhanced evaporation results \( \partial T / \partial t = (Q_{\text{LAT,jet}} - Q_{\text{LAT,calm}}) / (\rho_{\text{sw}} C_{\text{sw}} H) \sim 0.1^\circ/\text{day} \), a number one order of magnitude lower than the cooling produced by vertical mixing and net advection in the ocean.
4.2 Wind reversal / downwelling period

In this subsection we repeat the previous analysis for the period 2-7 January 2001, when weak northerly winds ($v_{\text{wnd}} \sim -5$ m/s) prevailed off south-central Chile. SST near Lavapié had dropped down to 11°C in the previous days under weak/moderate southerly winds followed by a marked increase up to 15.5°C (well above the seasonal average, Figure 11), during the wind reversal period. The spatial extent of the warming is also confined to the coastal region (~ 50 km) under the northerly flow with a small cooling farther north (Figure 12). The response of the ocean temperature and velocities in the vertical cross-sections to the wind reversal is also, in first approximation, the inverse to the observed in the jet event (compare the left and right panels in Figure 13), and includes convergent, onshore flow in the upper layer and a deceleration of the coastal ocean jet. Nevertheless, the column that experiences warming is deeper (+2°C down to 60 m) and narrower than in the cooling case.

The intense warming at the surface began on 2 January (soon after the surface wind became northerly) and rapidly progressed downward (Figure 14). The pattern of temperature change due to vertical mixing is the same as in the jet event (cooling aloft and warming below), so the warming of the upper 40 m is largely controlled by heat flux divergence due to advection. Of particular relevance is the horizontal warm advection, product of the onshore flow in the upper 30 m, acting during most of the wind reversal period. The surface warming re-forms the thermocline after 4 January, acting against the downward vertical velocities even when the surface northerly winds still prevail.
4.3 Trajectory analysis

To further illustrate the differences in ocean circulation between a jet event and the wind reversal period, we calculated the trajectories of the water parcels that arrived to the surface at a coastal point (38.1°S, 73.50°W; 10 km off Lavapié) using the backward capability of the Lagrangian drifter-tracking algorithm described before. Figure 15 shows selected 2-day backward trajectories (marked every 24 hrs) for selected days. Under weak, moderate southerly winds (15-16 December) the trajectories are very shallow and come directly from the south, consistent with the horizontal transport in the CCC at about 10 cm s⁻¹. Water parcels arriving to Lavapié at the beginning of the jet event (18 December) still followed a near-surface trajectory but they come from farther south as the CCC was accelerated by the wind. One day later (19 December) the trajectories originated not far to the south of Lavapié but at about 25 m deep, experiencing most of the lifting in the previous 36 hrs. On 20 December the trajectories return to the surface but they originate well to the south given the acceleration that experienced the CCC under strong southerlies. In contrast with the mostly coastal, along-shore trajectories during the jet event, the trajectories during wind reversal period are mostly zonal (cross-shore) and shallow. On 5 January, for instance, the water parcels reaching Lavapié were about 20 km off the coast 2 days before, to the west of the coastal-upwelling synoptic front.

5. Conclusions

In this work we have investigated the effects of the synoptic variability of the winds on the ocean off south-central Chile (35-45°S) using the regional ocean model (ROMS). We
performed two continuous 8-year long simulations (plus two years of spin up) forced by
QuikSCAT “observed” daily winds (synoptic run, SynSim, 2000-2007) and by the long-
term monthly mean winds (climatological run, CliSim). The simulations are identical
otherwise, including lateral boundary conditions from the World Ocean Atlas and COADS
climatological values to calculate surface fluxes. Our study focuses in the austral summer
(December to February) when south-central Chile experience southerly, upwelling
favorable mean wind but with substantial day-to-day variability including strong southerly
jet events and wind reversal (northerly flow) periods.

The simulated summer mean fields of upper-ocean currents and SST are very similar
between both runs and they reproduce the major observed features of the circulation in the
southeast Pacific, such as the Chile-Perú Current, the Perú-Chile Countercurrent and
Undercurrent, the Coastal Chilean Current, and the coastal band of upwelled cold waters.
Furthermore, the surface geostrophic EKE is similar between SynSim and CliSim,
consistent with the idea that this is intrinsically generated by instability of the major
currents. The model geostrophic EKE also agrees well with the spatial pattern observed by
satellite-data, but with higher values. In contrast, the ageostrophic EKE is significantly
higher (by a factor of 2) in SynSim compared with CliSim in a coastal band about 50 km
wide under the region of strong winds and collocated with the CCC. This has a direct
impact in the simulated diffusivities, which are twice as large in the synoptic simulation
than in the climatological run, highlighting the relevance of using variable (synoptic) winds
to force the ocean models if one aims at describing the turbulent component of the upper-
ocean circulation.
Discrepancies between both simulations are also found in coastal SST variability; within the first 50 km offshore. The standard deviation of the daily SST (calculated for each season) in synoptic run is about twice larger than its climatological counterpart. Comparison of the simulated SST in SynSim with in-situ observations suggests the prominent role of wind-driven upwelling cycles in coastal SST synoptic variability. The cross-shore scale (~50 km) of the differences between total EKE and SST variability between CliSim and SynSim is consistent with the region of coastal upwelling response to variable winds observed/simulated elsewhere [e.g., Perlin et al., 2007; Aiken et al., 2008]. Furthermore, the similarities in the spatial pattern of the mean SST, currents and geostrophic EKE between both simulations in contrast with the much higher level of coastal SST variability and ageostrophic EKE in the synoptic run is indicative of rather linear responses of the surface ocean to cycles of southerly wind strengthening (upwelling favorable) and relaxation (downwelling favorable in the case of northerly winds). We verified that feature by examining the scatter plot between meridional wind (simulated) coastal SST and surface currents.

Finally, we used a high-resolution synoptic simulation to describe the mechanism behind SST changes during two periods of strong southerlies and wind reversal. At the beginning of a strong southerly wind event vertical mixing leads to a rapid cooling in the upper 20-m layer and a slight warming immediately below. The weaker stratification permits the development of strong vertical velocities leading to a rapid deepening of the cooling due to net advection divergence. Eventually the upper 50 m column becomes nearly isothermal and both vertical mixing and net advection cease their cooling effect even under strong southerly winds. During the jet event, the Ekman flow varies in close phase with the wind
and there is a clear intensification of the Coastal Current off southern Chile. In contrast, when northerly winds are present the only process acting to warm the upper-ocean is the horizontal advection of warm waters product of the onshore flow in the upper 30 m. Thus, the magnitude of the coastal warming seems controlled by the duration of the wind reversal rather than by the wind speed.

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Figure Captions

Figure 1. Climatology of the austral summer (DJF) meridional wind (colors, in m s^{-1}) and
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(inner) rectangle indicates the model domain of the regular (high-resolution) ROMS
simulations used in this study.

Figure 2. Austral summer mean (DJF) of the surface meridional current (colors, in cm s^{-1})
and surface velocity (vectors are shown if the surface current speed is higher than 10 cm s^{-1})
for the (a) climatological and (b) synoptic simulations. Middle panels show vertical
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climatological and (c) synoptic simulation. Lower panels show vertical sections of the
austral summer mean of meridional currents at 36.5°S for the (e) climatological and (f)
synoptic simulation. Labels indicate the location of the major current systems (see text for
details).

Figure 3. Upper panels: austral summer mean of the simulated and in-situ measurements
(ADCP) of the zonal (u) and meridional (v) current profiles at 30.3°S – 71.7°W
(COSMOS). The observed mean was calculated using data from 2004 to 2006. Lower
panels: As upper panels but for a point at 36.5°S – 73.1°W (Station 18). The observed mean
was calculated using data from January and February 2009.

Figure 4. Austral summer mean of the surface eddy kinetic energy (EKE, cm² s⁻²).
Geostrophic EKE in (a) climatological simulation, (b) synoptic simulation and (c) satellite
derived from AVISO. Ageostrophic EKE in (d) climatological simulation and (e) synoptic
simulation.

Figure 5. Trajectories of surface drifters 23 days after being released simultaneously close
to the coast off south-central Chile using outputs from the high resolution climatological (a)
and synoptic (b) simulations.

Figure 6. Austral summer mean (DJF) of the sea surface temperature (°C). (a)
Observations from the Advanced Very High Resolution Radiometer (AVHRR, Pathfinder
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White lines in panels (b) and (c) indicate location of the mean coastal front.
Figure 7. Daily time series of the simulated SST (grey) in the synoptic run and \textit{in-situ} observations (black) at different locations during summer of 2006 and 2007. The correlation coefficient is indicated for each season within the box.

Figure 8. Austral summer standard deviation of the daily sea surface temperature (colors in °C) for the (a) Climatological and (b) Synoptic simulations.

Figure 9. Scatter plot of the daily meridional wind (QuikSCAT) versus the simulated SST (synoptic run) during austral summer (DJF from 2000 to 2007) off Punta Lavapié (37.4°S and 73.7°W). The dark circles are selected days during December 2000 illustrating the cooling during a jet event (18-20 Dec) and the subsequent warming as the wind relaxes. Also shown are the histograms for each variable including its mean (black line) and median (dotted line) values.

Figure 10. One-point correlation map between meridional wind ($v_{wnd}$ from QuikSCAT) at 34.7°S - 74.6°W and SST at one day lag (colors), meridional wind elsewhere at zero lag (contours), and the zonal ($u_{sfc}$) and meridional ($v_{sfc}$) component of surface currents at zero lag (vectors).

Figure 11. Coastal time series of the (a) meridional wind (QuikSCAT data), (b) SST and (c) vertical velocity ($w$) at 40 m depth for December 2000 and January 2001. SST and $w$ from the high-resolution synoptic simulation. All variables were averaged on line 50 km alongshore and 20 km off Punta Lavapié.
Figure 12. Upper panels: differences between the jet event (17-20 December 2000) and the previous calm days (11-14 December) in (a) meridional wind (QuikSCAT), (b) simulated SST and (c) observed SST (AVHRR). Lower panels as above but for the differences between the wind reversal period (2-7 December 2000) and previous days under weak southerly winds (31 December 2000 to 2 January 2000).

Figure 13. Left panels: differences between the jet event (17-20 December 2000) and the previous calm days (11-14 December) in a vertical section off Punta Lavapié of ocean (a) temperature, (c) zonal current and (e) meridional current. The differences are shown in colors; solid lines are the seasonal mean of each variable. Results from the high resolution synoptic run. Right panels: as left panels, but for the differences between the wind reversal period (2-7 December 2000) and previous days under weak southerly winds (31 December 2000 to 2 January 2000).

Figure 14. (a) Time series of the coastal meridional wind (QuikSCAT) off Lavapié for part of December 2000 and January 2001. (b) Time-depth section ocean temperature (contours, °C) and local rate of change of temperature (colors, °C s⁻¹). (c) Divergence of the heat flux due advection (°C s⁻¹). (d) Divergence of heat flux due to vertical mixing (°C s⁻¹). Results from the high resolution synoptic run in a coastal column about 20 km off Punta Lavapié and 50 km long in the along-shore direction.

Figure 15. Two-day backward trajectories of the drifters reaching a coastal surface location (10 cm depth at 38.1°S and 73.5°W) off Punta Lavapié. Numbers indicates the day the drifter reaches this point. Circles are plot every 24-hr and colored according to the drifter
depth at that time. Trajectories calculated using outputs from the high resolution synoptic run.

Table 1. Lagrangian time ($T_u, T_v$) and length ($L_u, L_v$) scales, velocity variances ($\overline{u^2}, \overline{v^2}$) and diffusivities ($K_u, K_v$) for the zonal and meridional components of the surface flow in both simulations (see more details in the text).

<table>
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<tr>
<th>Simulation</th>
<th>$T_u$, days</th>
<th>$T_v$, days</th>
<th>$L_u$, km</th>
<th>$L_v$, km</th>
<th>$\overline{u^2}$, cm$^2$ s$^{-2}$</th>
<th>$\overline{v^2}$, cm$^2$ s$^{-2}$</th>
<th>$K_u, 10^7$ cm$^2$ s$^{-1}$</th>
<th>$K_v, 10^7$ cm$^2$ s$^{-1}$</th>
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</thead>
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<tr>
<td>CliSim</td>
<td>2.3±0.06</td>
<td>2.7±0.06</td>
<td>10.6±0.3</td>
<td>13.0±0.4</td>
<td>40</td>
<td>40</td>
<td>0.65±0.02</td>
<td>0.86±0.04</td>
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<tr>
<td>SynSim</td>
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<td>2.3±0.06</td>
<td>13.2±0.3</td>
<td>17.3±0.6</td>
<td>81</td>
<td>77</td>
<td>1.20±0.04</td>
<td>1.60±0.07</td>
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</table>

The diffusivities obtained by Chaigneau and Pizarro [2005] using drifter buoy trajectories in the CPC region (24°-34°S;70°-82°W) were $K_u=2.2±0.5 \cdot 10^7$ cm$^2$ s$^{-1}$ and $K_v=1.8±0.4 \cdot 10^7$ cm$^2$ s$^{-1}$.
Table 2. Annual and summer (DJF) mean and standard deviation of *in-situ* and simulated SST at different locations along the coast and offshore. The correlation coefficient between simulated (synoptic run) and observed daily SST is shown for each station considering the period 2000-2007.

<table>
<thead>
<tr>
<th>Location</th>
<th>Data</th>
<th>Annual mean (°C)</th>
<th>Annual std (°C)</th>
<th>Summer mean (°C)</th>
<th>Summer std (°C)</th>
<th>Annual R</th>
<th>Summer R</th>
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<td>0.58</td>
<td>0.47±0.2</td>
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<tr>
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<td><em>in-situ</em></td>
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<td>2.0</td>
<td>14.6</td>
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<td>2.1</td>
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<td>1.2</td>
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Figure 11. Coastal time series of the (a) meridional wind (QuikSCAT data), (b) SST and (c) vertical velocity (w) at 40 m depth for December 2000 and January 2001. SST and w from the high-resolution synoptic simulation. All variables were averaged on line 50 km alongshore and 20 km off Punta Lavapié.
Figure 12. Upper panels: differences between the jet event (17-20 December 2000) and the previous calm days (11-14 December) in (a) meridional wind (QuikSCAT), (b) simulated SST and (c) observed SST (AVHRR). Lower panels as above but for the differences between the wind reversal period (2-7 December 2000) and previous days under weak southerly winds (31 December 2000 to 2 January 2000).
Figure 13. Left panels: differences between the jet event (17-20 December 2000) and the previous calm days (11-14 December) in a vertical section off Punta Lavapié of ocean (a) temperature, (c) zonal current and (e) meridional current. The differences are shown in colors; solid lines are the seasonal mean of each variable. Results from the high resolution synoptic run. Right panels: as left panels, but for the differences between the wind reversal period (2-7 December 2000) and previous days under weak southerly winds (31 December 2000 to 2 January 2000).
Figure 14. (a) Time series of the coastal meridional wind (QuikSCAT) off Lavapié for part of December 2000 and January 2001. (b) Time-depth section ocean temperature (contours, °C) and local rate of change of temperature (colors, °C s⁻¹). (c) Divergence of the heat flux due advection (°C s⁻¹). (d) Divergence of heat flux due to vertical mixing (°C s⁻¹). Results from the high resolution synoptic run in a coastal column about 20 km off Punta Lavapié and 50 km long in the along-shore direction.
Figure 15. Two-day backward trajectories of the drifters reaching a coastal surface location (10 cm depth at 38.1°S and 73.5°W) off Punta Lavapié. Numbers indicates the day the drifter reaches this point. Circles are plot every 24-hr and colored according to the drifter depth at that time. Trajectories calculated using outputs from the high resolution synoptic run.