Coastal Lows along the Subtropical West Coast of South America: Numerical Simulation of a Typical Case

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ABSTRACT

Subsynoptic, warm core low pressure areas are frequently observed along the west coast of subtropical South America during austral winter. These so-called coastal lows (CLs) tend to develop as an upper-air, midlatitude ridge is approaching the subtropical Andes and, therefore, while pressure is increasing aloft and farther to the south. These CLs have a profound impact in the coastal weather associated with a rapid transition from clear skies and stronger than average equatorward low-level flow to overcast conditions and relaxed equatorward (or even poleward) flow. Weather conditions inland mostly reflect the associated changes in the strength and height of the base of the subsidence inversion.

In this work, a mesoscale simulation of a typical CL episode is performed using a numerical weather prediction model [the fifth-generation Pennsylvania State University–NCAR Mesoscale Model (MM5)]. Comparison with observations reveals that the model simulation properly captures the large-scale pattern as well as many of the mesoscale features that characterize the CL. The model results were then used to diagnose the CL. It is found that the coastal troughing is largely due to the marked adiabatic warming of the lower troposphere (including a significant strengthening of the temperature inversion). The large-scale subsidence ahead of the incoming upper-air ridge axis is enhanced as the low-level easterly flow is constrained by the western slope of the subtropical Andes. The low-level wind off the subtropical coast is close to geostrophic balance and it is fed by air parcels that 1–2 days before had been located in the middle troposphere over the Pacific Ocean. The easterly flow is set up as the alongshore pressure gradient becomes poleward oriented because of the extratropical ridging. This gradient is further enhanced as the CL develops at subtropical latitudes. As soon as the ridge axis crosses, the low-level easterly flow vanishes and a shallow, narrow tongue of northwesterlies and stratocumulus clouds propagates poleward from northern Chile. Shortly thereafter, the trapped wind reversal merges with the incoming synoptic-scale, tropospheric deep-cyclonic circulation.

1. Introduction

The typical anticyclonic regime that prevails along the subtropical west coast of South America (25°–35°S, also referred to as central Chile; see Fig. 1) is interrupted during periods of relative low pressure, associated with the passage of midlatitude cyclones farther to the south or localized troughing, which we refer to as coastal lows (CLs). Alternance of these two phenomena gives rise to the synoptic-scale variability of the surface pressure \( p_s \) over this region, which is most marked from fall to spring. The distinctive features of a CL are that much of its development (i.e., \( \partial p_s / \partial t \leq 0 \)) occurs as pressure increases aloft, to the south of 40°S, and away from the subtropical coast, indicative of a significant disruption of the large-scale flow by the coastal range and the Andes Cordillera (Fig. 1), and a transition of the alongshore near-surface wind from strong to relaxed equatorward flow (e.g., Garreaud et al. 2002).

The regional features of South American CLs and the attending large-scale circulation were documented by Garreaud et al. (2002), using surface data from a network of automated weather stations (AWS) in central Chile, National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis fields (Kalnay et al. 1996), and satellite imagery. The results are based on a compositing analysis of 21 episodes selected on the basis of surface pressure and wind observations from an AWS at Point Lengua de Vaca (30°S, 71°W, 7 m MSL). In this station, the composite trace of \( p_s \) exhibits an asymmetric V shape with a life span of about 3 days and an amplitude of about 5 hPa. Examination of the surface pressure in the rest of the coastal stations indicates that the initial time when \( \partial p_s / \partial t < 0 \) occurs almost simultaneously along the subtropical coast (25°–35°S). Thus, the surface pressure pattern
during the onset and developing stages of a CL is characterized by an elongated trough along the subtropical coast. The attending synoptic-scale circulation is characterized by a midlevel midlatitude ridge drifting eastward from the southeast Pacific, weaker than normal midlevel westerly flow over the subtropical Andes (midlevel easterly flow was found in 30% of the cases), and a surface anticyclonic anomaly with its center at about 45°S that results in a stretching of the subtropical anticyclone into the continent.

During the developing stage of a CL there are important effects on regional weather, including (a) an enhancement of the temperature inversion that tends to depress the marine boundary layer (MBL) at and off the coast and leads to severe air pollution episodes in the major inland cities of central Chile during austral winter (e.g., Rutllant and Garreaud 1995), also increasing the risk of forest fires during summertime; (b) stronger than normal equatorward flow at the sea surface leading to enhanced upwelling of cold, nutrient-rich water along the coast (Rutllant 1993; Rutllant and Montecino 2002); and (c) a clearing of the marine stratocumulus (Scu) along the subtropical west coast, often extending more than 1000 km offshore, increasing the solar radiation reaching the ocean surface by ~40 W m⁻² (similar Scu clearing was documented by Kloesel (1992) along the coast of California).

The demise of a CL begins as soon as the midlevel ridge axis crosses the Andes and the associated surface anticyclone migrates northeastward over Argentina. It features a return of the MBL and low clouds from the coast of northern Chile (where the MBL has not been perturbed) to the south at a mean speed of $C_L = 16 \pm 5$ m s⁻¹ (Garreaud et al. 2002). Thus, at any given point along the coast, the CL culmination (the time when $\frac{\partial p}{\partial t} = 0$) is characterized by a rapid transition from stronger than normal equatorward coastal flow to relaxed equatorward flow (or even poleward flow), as well as a transition from warm, clear-sky conditions to cool, overcast conditions along the coast and farther inland associated with the recovery of the MBL. During this stage of the CL the low-level flow off the coast exhibits a cyclonic circulation, with offshore (onshore) flow in the clear (cloudy) region, as evidenced in composite time series of zonal flow and cloudiness (see Figs. 8 and 13 of Garreaud et al. 2002). The surface pressure also exhibits a weak coastal minimum within the region where the offshore flow and clear skies still prevail (Garreaud et al. 2002).

The demise of a CL seems to share some of the typical features of mobile coastal lows observed along the west and south coasts of South Africa (e.g., Gill 1982; Reason and Jury 1990) as well as coastally trapped disturbances (CTDs) along the western coast of North America (e.g., Dorman 1985; Bond et al. 1996; Nuss et al. 2000). Nevertheless, in many cases, especially during winter-

### Table 1. Physical parameterization used in the MM5 numerical simulation.

<table>
<thead>
<tr>
<th>Processes</th>
<th>Scheme</th>
<th>Reference</th>
</tr>
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<tr>
<td>Cumulus convection</td>
<td>Kain–Fritsch</td>
<td>Kain and Fritsch (1993)</td>
</tr>
<tr>
<td>Boundary layer</td>
<td>High-resolution Blackadar</td>
<td>Blackadar (1979)</td>
</tr>
<tr>
<td>Moisture</td>
<td>Simple ice</td>
<td>Dudhia (1989)</td>
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<tr>
<td>Radiation</td>
<td>Cloud–radiation</td>
<td>Benjamin and Carlson (1986)</td>
</tr>
<tr>
<td>Ground temperature</td>
<td>Force–restore</td>
<td>Blackadar (1979)</td>
</tr>
</tbody>
</table>
time, the demise of the CL is simultaneous with the approach of a midlatitude cyclone and its attending cold front, so the poleward propagation of a CTD, if any, is short lived (less than 24 h). Further similarities and differences between South American CLs and North American CTDs are described in Garreaud et al. (2002, see their section 7).

Radiosonde data and reanalyzed vertical profiles of temperature, humidity, and winds indicate that the coastal troughing is produced by the localized warming of the lower troposphere in connection with the strengthening of the climatological temperature inversion near the coast (Garreaud et al. 2002). Nevertheless, given the relatively coarse resolution of the observation’s network in this region (e.g., only one radiosonde station), several aspects of this phenomenon are not well documented, including the three-dimensional structure of the warming and the relative importance of temperature advection and diabatic heating. Furthermore, while the easterly flow appears largely forced by a surface anticyclone moving into southern South America, the coastal trough at subtropical latitudes can further steepen the alongshore pressure gradient and thus feedback on the large-scale flow. Mesoscale data are needed to assess this later effect, as well as to describe the interaction of the easterly low-level flow with the complex terrain and details of the CL demise.

Aiming at a better description and diagnosis of South American coastal lows at the meso-α scale (100–1000 km), including the aforementioned aspects and their impact on regional weather, we performed a 6-day numerical simulation of a typical coastal low that occurred in August 2001. In this study we used the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5) with finest grid spacing of 5.0 km. Whenever possible, we compared the model results with observations in order to evaluate the ability of the MM5 to simulate South American CLs. The paper is organized as follows. The model setup is described in section 2. Section 3 presents an overview of the episode as well as the model results and validation organized in synoptic-scale and mesoscale features. A discussion of the CL dynamic based on several diagnoses of the model results is presented in section 4, and conclusions are presented in section 5.

2. Model setup

To obtain a mesoscale, three-dimensional description of a coastal low, a numerical simulation of a typical episode was performed using the MM5 version 3. The MM5 is an area-limited, nonhydrostatic, numerical model formulated in a terrain-following vertical coordinate as described in detail in Dudhia (1989) and Grell et al. (1994). The parameterizations used in this simulation are listed in Table 1. The simulation features four two-way nested domains with horizontal grid spacing ($\Delta x = \Delta y$) of 135, 45, 15, and 5 km (see details in Table 2 and Fig. 1). Domain 1 (coarser mesh) covers a significant portion of subtropical South America and the southeast (SE) Pacific, and the subsequent domains zoom in on central Chile. In the vertical, all domains have 28 $\sigma$ levels, with the highest resolution in the PBL ($\Delta z \sim 70$ m in the lowest 1 km). Initial conditions for all domains and time-dependent boundary conditions for domain 1 were obtained by interpolating global NCEP surface and upper-air analysis ($1.25^\circ \times 1.25^\circ$ lat–long grids) every 12 h. Synoptic surface and radiosonde observations were not assimilated in the MM5 since they were already assimilated in constructing the global analysis. An important caveat in the initialization of CL and CTD simulations is that the MBL and the clouds at its top are often not properly represented in global analyses (e.g., Mass and Steenburgh 2000; Leidner et al. 2001).

The simulated CL took place in late austral winter and culminated around 0000 UTC 22 August 2001 (2000 local time 21 August). The episode exhibits the typical structure and evolution of a coastal low according to the compositing analysis described in Garreaud et al. (2002). Additionally, there are enough observations available for this recent case to validate the structure and evolution generated by the model. In order to capture the full evolution of the CL, the integration spans from 0000 UTC 18 August to 0000 UTC 26 August (190 h). The model results provide a high-resolution, physically consistent dataset, including meteorological fields that are difficult to measure but key in diagnosing these events (e.g., vertical velocity). To this respect, simulation of CTDs in western North America reported in the literature are encouraging: Thompson et al. (1997a,b) using the U.S. Navy Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS), Guan et al. (1998) and Jackson et al. (1999) using the

<table>
<thead>
<tr>
<th>Value</th>
<th>Domain 1</th>
<th>Domain 2</th>
<th>Domain 3</th>
<th>Domain 4</th>
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</thead>
<tbody>
<tr>
<td>Horizontal grid spacing ($\Delta x = \Delta y$)</td>
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<td>45 km</td>
<td>15 km</td>
<td>5 km</td>
</tr>
<tr>
<td>No. of grid points ($N_x \times N_y$)</td>
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<td>$54 \times 54$</td>
<td>$73 \times 73$</td>
<td>$90 \times 90$</td>
</tr>
<tr>
<td>Lat span (°S)</td>
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<td>42–24</td>
<td>37–30</td>
<td>34–32</td>
</tr>
<tr>
<td>Lon span (°W)</td>
<td>96–49</td>
<td>85–63</td>
<td>77–67</td>
<td>74–69</td>
</tr>
</tbody>
</table>

Table 2. Geometry of the model’s domains.
Regional Atmospheric Modeling System (RAMS), and Mass and Steenburgh (2000) using the MM5.

3. Model results and validation

a. Episode’s overview

The CL evolution in central Chile (33°–34°S) is encapsulated in the time series of key observed and simulated variables shown in Fig. 2. The surface pressure at Santiago (inland, 580 m MSL) remains relatively high until 1800 UTC 19 August, when it began to drop markedly for the next 36 h (Fig. 2a). After 21 August the pressure remains low for the next 2 days followed by a new negative trend. Surface pressure in other inland and coastal stations in central Chile follows a very similar evolution. In contrast, the pressure at mid- and upper levels follows an opposite evolution during the first half of the simulation, as shown by the 500-hPa geopotential height ($Z_{500}$) at Santo Domingo, a coastal radiosonde station at about the same latitude as Santiago. The increase in $Z_{500}$ culminates at 0000 UTC 22 August, nearly at the time when the surface pressure in Santiago reaches its local minimum (CL culmination hereafter). The model surface pressure and $Z_{500}$ are in good agreement with the observations, except for a slight underestimation of the CL amplitude and $\sim 9$ h lag in the CL onset.

Thirty-minute data from an AWS at Lo Prado are used to illustrate the changes in the lower troposphere during the CL (Figs. 2b–d). Station Lo Prado (33.5°S, 70.9°W, 1065 m MSL) is located near the climatological top of the subsidence inversion on a saddle point of the coastal range (wind dominated by its zonal component), and atop a 20-m-high telecommunications tower. During the developing stage of the CL there is a marked warming and drying, as well as an increase of the strength and duration of the easterly flow with respect to the mean diurnal cycle of the regional zonal wind (Rutllant and Garreaud 1995). Easterly flow, high temperatures ($10^\circ$C above the monthly mean) and very low humidities (mixing ratio less than 0.3 g kg$^{-1}$) prevail until 1800 UTC 22 August ($\sim 18$ h after the CL culmination). The subsequent cooling and moistening are less marked than the initial trends. Relative humidity reaches 100% by 24 August, at the time when the easterly flow vanished.

The simulation closely follows the observed warming and drying (albeit there are some problems in the moisture field during the first 24 h) and produces extreme values very similar to those observed.$^1$ The cooling–moistening after the culmination are, however, significantly underestimated by the model, leading to model–observed differences as large as $+8^\circ$C and $-0.7$ g kg$^{-1}$. Such a discrepancy is not found when comparing model outputs with observations from station La Dormida, $\sim 40$ km to the north of Lo Prado but at 1540 m MSL (not shown). As we show later, the problem at Lo Prado (and lower-elevation inland stations) is associated with the deficiency of the simulation in transporting low coastal clouds inland. Results from a simulation initialized at 1200 UTC 21 August exhibit the same underestimation of the inland cooling and moistening. The evolution of the simulated zonal wind is in good agreement with the observations, but with magnitude 3–5 ms$^{-1}$ smaller than observed. The stronger wind and its diurnal cycle at Lo Prado appear to be rather a local

$^1$ The simulated air temperature and mixing ratio exhibit a somewhat smaller diurnal cycle than in the observations, presumably because the model’s lowest sigma level is at 37 m MGL.
b. Synoptic-scale features

As in the compositing analysis of Garreaud et al. (2002), the midlevel circulation during the CL is characterized by the passage of a midlatitude ridge over subtropical South America with its axis oriented NW–SE, as shown by the simulated synoptic maps in Fig. 3 (domain 2, 45 km). Notice that we have not drawn the SLP field over regions where the terrain elevation exceeds 2000 m MSL (most of central Chile at this resolution) since the reduction of pressure to sea level generates a number of artificial features. By 23 August the ridge has moved to the east of the Andes and a mobile trough is off the west coast. At the surface, the southern half of the subtropical anticyclone over the SE Pacific [a climatological feature with its center at ~33°S, 90°W; e.g., Rodwell and Hoskins (2001)] is reinforced by an extratropical ridge during the first half of the CL development (Figs. 3a,b). The extratropical ridge leads to a coastal maximum of SLP at about 37°S and high surface pressure to the east of the subtropical Andes by the time of the CL culmination (Fig. 3c). Between 18 and 22 August, the SLP drops along the subtropical coast (in agreement with the few surface observations, not shown), leading to the formation and subsequent deepening of a trough, which signals the development of the CL. The characteristic high–low–high zonal pattern of SLP across the Andes during the CL development (e.g., Garreaud et al. 2002) is rapidly replaced by

(meso $\beta$) topographic channeling of the flow not resolved by the model (Rutllant and Garreaud 1995).
Subsynoptic details of the CL evolution at and off of the coast are shown in Fig. 4 by mean of the SLP field, the local rate of change of SLP (centered difference evaluated using the 1-h model outputs), and 950-hPa winds, based on the model outputs from domain 3 (15 km). During the developing phase (Fig. 4a), the troughing is confined to the west of the subtropical coast decaying westward within a cross-shore scale of ~500 km, coincident with the area in which the winds veer from SSW to SE. Half a day after the CL culmination (Fig. 4b), SLP has began to drop over most of the domain, except for a narrow coastal region to the north of 30°S. These positive tendencies are not found above 900 hPa, suggesting their association with the local return of the MBL. At this time, the SLP field exhibits a weak closed low, and the low-level flow over the ocean exhibits a cyclonic circulation. Notice that Fig. 4b shows approximately the western half of the CL since the land area has been masked. However the surface convergence toward the CL should produce northerly flow in the northeastern sector of the CL and westerly (SW) in the northwestern sector, considering only frictionally modified geostrophic flow. Ageostrophic flow components arising during the fast southward propagation of the low should enhance southerly wind components ahead of it (southwestern section of the CL) and northerly ones behind it (northwestern section of the CL).

In concert with the surface pressure drop to the west of the subtropical Andes, a marked warming begins to the north of 30°S on 18 August, but rapidly expands southward down to 40°S. As shown in the vertical cross section at 33°S in Fig. 5, the warming during the CL development encompass both the lower and middle troposphere. Low-level easterly flow also sets up almost simultaneously at subtropical latitudes on 19 August, and persists until the CL culmination, when its demise is also nearly simultaneously along the coast. By the
FIG. 6. The 48-h forward trajectories (solid line) of three air parcels released at 0000 UTC 20 Aug 2001. The position of each parcel is indicated every 12 h by a symbol, coded according to the parcel’s altitude at this time. Parcel A was released at 670 hPa, 33°S, 80°W; parcel B was released at 510 hPa, 36°S, 83°W; and parcel C was released at 520 hPa, 31°S, 77°W. For parcels B and C, the projections of their trajectories on the surface are shown by dashed lines.

Time of the CL culmination, easterly flow extends from the surface to about 700 hPa to the west of the Andes (Fig. 5). Above 600 hPa, westerly flow persists during the whole CL development, and therefore there is no flow crossing the subtropical Andes from east to west (recall that at these latitudes the Andes top is above 5000 m MSL, and the lower passes are above 4000 m MSL). The evolution of the simulated winds is in good agreement with the upper-air observations at Santo Domingo (not shown).

To further describe the three-dimensional flow during the CL development we have calculated forward trajectories for a 48-h period ending at the culmination time for a number of air parcels. Figure 6 shows the trajectories of three air parcels, all of them originating near 500 hPa over the SE Pacific (80°W). Far from the Andes, the three parcels follow an anticyclonic path in the middle troposphere and a slight descent, consistent with the presence of the ridge aloft. Parcel A (released at 670 hPa, 33°S) is not entrained into the low-level cyclonic circulation to the west of the Andes, and therefore it moves mostly eastward crossing the Andes at about 31°S. Parcel B (released at 510 hPa, 36°S) subsides below 700 hPa well off the coast, and then it follows a cyclonic path as well as a rapid descent down to 0.3 km. Parcel C (released at 520 hPa, 30°S) also subsides below 700 hPa, but closer to the western slope of the Andes than parcel B, following cyclonic path and a rapid descent right at the coast. Thus, the low-level easterly flow off the subtropical coast during the CL development is fed by air parcels that 24–48 h before had been located in the middle troposphere (500–550 hPa) over the SE Pacific Ocean. They experience a gentle descent followed by a more rapid one just upstream of the Andes below 600 hPa, at the same time that their path changes from anticyclonic to cyclonic. Air parcels originating slightly above 500 hPa also experience mid-tropospheric subsidence, but they are able to cross the subtropical Andes toward Argentina.

Stronger than normal low-level southerly (approximately alongshore) winds are observed off the coast of central Chile during the developing stage of the CL (Fig. 4a). By the time of the CL culmination, weak, shallow northerly flow begins to advance southward from the northern part of the domain (Figs. 4b and 7). Nevertheless, the poleward propagation of the northerlies at low levels rapidly merges with tropospheric-deep northerlies driven by the incoming extratropical cyclone. The signatures of the wind changes over the ocean are further illustrated in the map of 24-h wind change at the lowest \( \sigma \) level (−37 m MSL) for the period ending at 1200 UTC 21 August (Fig. 8a). The area of strong southerly local acceleration (and therefore, upwelling-favorable winds) extends about 600 km off the coast, bordered to the north by the coastal wedge of northerly local acceleration. In contrast, zonal wind changes are negli-
c. PBL changes and mesoscale features

To obtain a more detailed description of the CL-associated changes in the lower troposphere, Fig. 11 shows the evolution of potential temperature ($\theta$), specific humidity ($q$), and wind ($u$, $v$) profiles at two grid columns in subtropical latitudes. The first column is at 33°S and about 150 km off the coast. The initial westerlies are replaced by a layer of nearly uniform easterlies between 100 and 2000 m MSL that sets up at 1200 UTC 19 August and remains mostly steady until about 12 h after the CL culmination (Fig. 11a). In the lowest 100 m the easterly wind is very weak, and a transition to weak westerly (i.e., onshore) flow occurred on 21 August. After the CL culmination the westerly flow return to the whole troposphere associated with the incoming midlatitude trough. Simultaneous with the development of easterly flow there is a general warming and drying of the lower and middle troposphere (Figs. 11a,b). At the beginning of the CL, the base and strength of the subsidence inversion are close to their climatological values ($z_{base} \approx 600$ m and $\partial \theta / \partial z \approx +20^\circ K km^{-1}$); by the time of the CL culmination the inversion base has dropped to the surface and $\partial \theta / \partial z \approx +30^\circ K km^{-1}$. After the CL culmination, there is a gradual recovery of the MBL and SCu clouds, followed by a larger disruption on 25 August associated with the incoming synoptic cyclone. Figure 11b also shows the intensification of the low-level southerly wind during the CL development, which is especially marked in the lowest 1000 m, and its subsequent transition to northerly wind.

The second column is located over Santiago (Fig. 11c). The evolution of $\theta$ and $u$ are generally similar to their offshore counterparts but for two key differences. First, the low-level warming is more pronounced and persists for 24 h after the CL culmination. The aforementioned strengthening of the subsidence inversion limits the daytime growth of the inland mixed layer to $\approx 250$ m above ground during the CL development (in agreement with observations in Santiago). Observations at Santiago (not shown) reveal a marked cooling on 23

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2 The base and top of the temperature inversion were taken as the lowest and highest levels of the layer where $\partial \theta / \partial z \approx 9.8 K km^{-1}$. 

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A significant disruption of the SCu deck over the subtropical SE Pacific during the CL evolution is seen in the sequence of Geostationary Operational Environmental Satellite-8 (GOES-8) visible images shown in Fig. 9. The first image is rather typical of undisturbed conditions (Fig. 9a). On the afternoon of 19 August the SCu deck begins rapidly to break down near the coast and by 20 August the clear-sky region extends more than 1000 km off the subtropical coast. A coastal wedge of compact SCu is also evident to the north of 30°S. The wedge propagates southward at a mean speed of $\approx 13$ m s$^{-1}$ during 22 August (the propagation speed was estimated on the basis of the original visible and infrared images available every 30 min). The simulation was able to capture the rapid, widespread SCu clearing as well as the wedge of SCu along northern Chile, as illustrated by four snapshots of relative humidity and mixing ratio averaged between 1000 and 950 hPa in Fig. 10. Notice that the dry, clear-sky region moves westward into the subtropical Pacific well after the SCu deck has recovered near the coast.
August (e.g., maximum temperature drops from 27°C to 12°C) when moist MBL air and thick low coastal clouds arrived in the inland valleys. Despite an increase of the low-level flow from the coast, the model was not able to simulate the inland intrusion of the low clouds, leading to large discrepancy between the observed and simulated inland temperatures. High-resolution satellite imagery suggests that the intrusion of low clouds is favored by the existence of mesoscale gaps in the coastal range (the Maipo and Aconcagua lower valleys) that connect the coast with the inland valleys. Thus, the model’s failure to capture the onset of the coastal air and cloud intrusion might be, at least partially, related with insufficient horizontal resolution. We plan to address this important deficiency of the model in future work.

The second difference between coastal and inland conditions is the marked diurnal cycle present in the inland zonal wind, driven by the surface heating/cooling cycle over the sloping terrain (north–south-oriented ridges of ~3000 m above the valley floor are just 10–20 km to the east of Santiago). This diurnal cycle is depicted in the vertical cross section of $u$ at dawn and midafternoon on 20 August (Figs. 12a,b). Within a layer of about 1 km over the ground, the easterly flow is enhanced by the downslope flow during nighttime and early morning and offset by the upslope flow in the afternoon. These features are in qualitative agreement with piloted balloon (pibal) observations taken near Santiago during CLs (Garreaud 2002).

The low-level flow also exhibits significant horizontal variability over the complex terrain, as illustrated by the wind field on a pressure surface (900 hPa, approximately 1000 m MSL) and on a $\sigma$ surface (0.995, about 40 m above ground) in the early morning of 20 August (Fig. 13). At the 900-hPa surface, the area of strong easterly winds is rooted in the east–west Andean valleys and extends westward over the Santiago basin (Fig. 13a). Because of the strong static stability at low levels over Santiago (e.g., Fig. 11c), the downward momentum transfer is inhibited and very light winds prevail at the surface over the central valley’s flat terrain (Fig. 13b). This mesoscale variability has been well documented by local observations (Garreaud 2002). There is also strong easterly flow at the surface near the coast (Fig. 13b), presumably due to the acceleration of the low-level easterly flow over the western slope of the coastal range (300–700 m MSL).

Midlevel northerly flow is typically observed over central Chile, as a result of the topographic blocking of the free-tropospheric westerly flow upstream of the Andes Cordillera (Fig. 14a; Rutllant 1993; 1994). This barrier flow often organizes into a jetlike structure at about 800 hPa, very close to the western slope of the Andes, whose intensity depends on both the magnitude of the impinging westerlies and the midlevel static stability (e.g., Parish 1982). In the simulation the northerly flow reaches a maximum at the onset of the CL, when the westerlies are still strong and the static stability has increased in connection with the incoming warm ridge aloft (Fig. 14b). A rapid deceleration follows until the CL culmination, in concert with the weakening of the westerly flow. After the CL culmination the northerly
flow increases again, largely forced by the incoming midlatitude trough.

During the CL demise, the model simulates a poleward propagation of northerly low-level flow and SCu clouds around culmination time (Figs. 4b and 7), in general agreement with the QuikSCAT surface winds, surface observations along the coast (not shown), and satellite visible imagery. Much of this poleward advance of the near-surface northerlies occurs as the synoptically driven easterly flow aloft (e.g., 850 hPa) still prevails along the subtropical coast, suggesting a free propagation. In any case, this feature was short lived (~18 h) as a tropospheric-deep cyclone approached the coast by late 22 August. Some quantitative aspects of this phenomenon can be readily obtained from the model outputs at 45-km resolution. The simulated northerly flow and the low clouds are organized in a coastal

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3 The innermost grid (5 km) extends north to 32°S, while the propagation spans from 26° to 33°S.
Fig. 10. Low-level (1000–900-hPa average) mixing ratio (contoured every 1.5 g kg$^{-1}$) and relative humidity (shaded according to the scale at the bottom). Date and time are shown at the top of each panel. Data are from model domain 2.

4. Model diagnosis

The CL development occurs as the axis of the upper and midlevel ridge is to the west of the Andes (e.g., Fig. 3), and hence anticyclonic vorticity advection (AVA) aloft is found over the subtropical coast. In this condition, one would expect a maximum of downward motion in the middle troposphere, divergence below this level, and a surface anticyclone. Indeed, during the CL development, positive SLP increases well off of the sub-
tropical coast, farther to the south and to the east of the subtropical Andes (Fig. 4a). Over the subtropical coast, however, pressure drops below 800 hPa (maximum troughing at the surface) in connection with low-level convergence and a marked, localized warming. Of course, the enhanced warming is in turn associated with the disruption of the synoptic-scale flow by the Andes, as we will show later.

In order to quantify the role of the temperature advection and diabatic heating in the warming, the model
outputs (domain 3, 15 km) were used to evaluate the thermodynamic energy equation on several pressure surfaces:

$$\frac{\partial}{\partial t} T = -\mathbf{V} \cdot \nabla_T T + \omega S_p + Q_D, \quad (1)$$

where \(\mathbf{V}\) is the horizontal velocity on pressure surfaces; \(\omega = dp/dt\) is the downward pressure velocity; \(S_p = -T \ln \theta / \theta_p\) is the static stability parameter; \(\gamma\) and \(\gamma_s\) are the actual and dry adiabatic lapse rate, respectively; and \(Q_D\) represents the total diabatic heating.

The terms of Eq. (1) averaged between 900 and 850 hPa at 1200 UTC 20 August are shown in Fig. 15 (for sake of clarity the fields are displayed off the coast only). At this time, low-level warming encompasses the whole domain, but it is most pronounced in a band ~500 km off the coast to the north of 35°S (Figs. 15a,b). The simulated diabatic heating is relatively small (\(|Q_D| \sim 0.3°C \text{ day}^{-1}\)), as expected by the absence of clouds (little radiative heating and no latent heat release within the layer), so the warming is largely produced by advection. The major contribution is produced by the subsidence for two reasons. First, the large-scale low-level subsidence is significantly enhanced within a band of ~500 km extending westward from the Andes produced by the easterlies flowing down the sloping terrain (i.e., \(\omega = u dh_p / \partial x\)). Second, the enhanced subsidence takes place in the (very) statically stable lower troposphere (\(\partial \theta / \partial z > 10°C \text{ km}^{-1}\); cf. Fig. 11a). The dominant role of the vertical advection is held during the whole CL development. After the CL culmination, the adiabatic cooling (associated with the general upward motion ahead of the incoming extratropical trough axis and a weakly stable column) is the leading term in the cooling of the lower and middle troposphere upstream of the subtropical Andes.

The advection of warm, continental air by the offshore flow (approximately the zonal wind in this case) is considered a key ingredient in the low-level warming that precedes the occurrence of CTDs in western North America (Ralph et al. 1998). In our case, the highest temperatures are indeed found inland and at the coast so the offshore flow produces a warming as large as 5°C day\(^{-1}\) over the subtropical ocean, but it is almost completely offset by cold advection produced by the strong southerly flow.

The thermodynamic balance inland is quantitatively more complex because of the diurnally varying surface heating and the mesoscale details of the wind field over the complex topography. Equation (1) was evaluated using results from domain 4 (5 km) averaged over 4 × 4 inland grid points centered at 33.5°S, 71°W. Since the meridional flow is very weak (Fig. 14), meridional advection no longer compensates for the zonal advection. The temperature advection by the zonal wind is, at some times, as large as the vertical advection (~13°C day\(^{-1}\)), but it changes its sign during the day. Thus, when integrated during a whole day, vertical advection is still the dominant term in producing the low- and midlevel warming over the continent.

Enhanced subsidence near the coast also explains the rapid spread of the subtropical SCu clearing during the CL development. The SCu is maintained in the upper part of the MBL as long as the mixing condensation level \(z_c\) is below the inversion base \(z_i\). In undisturbed conditions (Garreaud et al. 2001), the cloud depth \((z_c - z_i)\) over the subtropical SE Pacific is about 400 m. Since \(z_c\) is determined by the surface conditions (and hence SST), it varies more slowly than the rate of change of the subsidence aloft, and therefore the SCu clearing is controlled by downward displacement of the inversion base at \(dz_i / dt = w|_{z_i} \sim 70 \text{ m h}^{-1}\) within 500 km off the coast. Thus, once the enhanced subsidence begins to act, the widespread SCu clearing has a characteristic time of ~6 h in agreement with the satellite observations.

Let us now consider the forcing of the low-level wind off the subtropical coast during the CL development.
To this effect, the total wind ($\mathbf{V}$) can be expanded in its geostrophic ($\mathbf{V}_g$) and ageostrophic components (e.g., Haltiner and Williams 1980):

$$
\mathbf{V} = \mathbf{V}_g + \frac{1}{f} \mathbf{k} \times \frac{\partial}{\partial t} \mathbf{V} + \frac{1}{f} \mathbf{k} \times (\mathbf{V}_s \cdot \nabla_s) \mathbf{V} + \frac{1}{f} \mathbf{k} \times \mathbf{F}, \tag{2}
$$

where $\mathbf{V}_s = \mathbf{V} + w_5 \mathbf{k}$ is the three-dimensional total wind and $\nabla_s$ is the three-dimensional gradient operator. The second and third terms on the rhs correspond to the isallobaric wind and advective wind, respectively. The last term represents the effect of the frictional force per unit mass. Equation (2) was numerically evaluated using the model outputs at 33.5°S, 72.5°W within the MBL and at 900 hPa.

At 900 hPa, the zonal and meridional wind components were within ±10% of their geostrophic values during most of the CL life cycle, with the exception of a brief period during the CL onset (18 August) when southwesterly isallobaric wind was as large as 1.5 m s$^{-1}$ (convergence toward the incipient low). Near the surface, the frictional component becomes significant. It points toward the lower pressure, decelerating the easterly wind and accelerating the southerly wind. As the MBL shrinks ($H \rightarrow 0$) and the meridional wind increases, the frictional component becomes strong enough, to lead to a transition of the near-surface wind from offshore (geostrophically driven) to onshore flow several hours before the CL culmination. Evaluation of equation (2) farther to the north reveals that the wind field within the coastal tongue of northerly flow is also close to geostrophic balance.

Since the coastal wind field is close to geostrophic balance, it is worthwhile to consider a rough estimate of the pressure gradient force at play. At the beginning of our simulation the alongshore SLP gradient points equatorward (as it normally does), as shown by the coastal SLP traces at 36° and 30°S in Figs. 16a,c. During 18 August the synoptic-scale ridging leads to a marked increase of the SLP to the south of 36°S while the SLP remains steady farther to the north, leading to a reversal of the SLP gradient by 0000 UTC 19 August. At this time SLP begins to drop at central Chile signaling the onset of the CL. For the next 24 h, SLP to the south of 36°S remains mostly steady while the CL develops, further enhancing the poleward-directed SLP gradient and hence the easterly, offshore low-level flow. After 0000 UTC 20 August there is a general troughing along the coast, but the poleward SLP gradient persists for the next 3 days. Thus, although the onset of the alongshore SLP gradient reversal is due to the synoptic ridging at midlatitudes, the development of the CL is key in steepening this gradient by a factor of 2 and extending its reversal well after the ridging has ceased to act. The importance of the coastal troughing is even more pronounced in setting up the offshore-directed SLP gradient that persists during the CL development and part of its demise (Figs. 16b,c) and drives the stronger than normal southerly, alongshore low-level flow.

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4 Recall that the friction within the PBL can be estimated as $|\mathbf{F}| = C_f \mathbf{V}_s^2 / (fH)$, where $H$ is the PBL depth and $C_f$ is the drag coefficient.
5. Concluding remarks

The MM5 was used to simulate a typical CL along the west coast of subtropical South America on 19–24 August 2001. The model was able to simulate the large-scale flow during the episode and the mesoscale features that characterize the CL’s developing stage, including (i) the coastal troughing and the marked warming and drying in the lower troposphere, (ii) the jetlike structure of the easterly flow to the west of the subtropical Andes, (iii) the strengthening of the alongshore equatorward near-surface winds, and (iv) the rapid clearing of the SCu clouds off the coast.

The simulated features during the CL demise are also in agreement with the observations (including a short-lived coastally trapped wind reversal propagating poleward), except for the absence of the observed intrusion of MBL air and coastal low-level clouds into the inland valleys of central Chile. This problem leads to large discrepancies between the observed and simulated temperature and humidity at lower levels. Proper simulation of the inland intrusion of MBL air and coastal SCu is a complex task (e.g., Leidner et al. 2001; Lewis et al. 1999; Kong 1999) that requires a better representation of the PBL and the land surface processes, and perhaps higher horizontal and vertical resolution to fully capture topographic mesoscale details.

The model’s results thus provide a high-resolution, physically consistent dataset useful in diagnosing the underlying dynamic of the CL. The key findings are as follows.

- The low-level alongshore pressure gradient is initially reversed from climatology by the synoptic-scale ridging that takes places at the extratropics, leading to low-level easterly flow at the subtropics. The subsequent subtropical coastal troughing further steepens the poleward-directed pressure gradient and, therefore, enhances the low-level easterly flow and adiabatic warming. A positive feedback is thus produced, extending the CL life span and increasing its amplitude. These results are in agreement with the conceptual model of CLs proposed by Garreaud et al. (2002) on the basis of a compositing analysis.

- The low-level easterly flow off the subtropical coast during the CL development is fed by air parcels that 24–48 h earlier had been located in the middle troposphere (500–550 hPa) over the SE Pacific Ocean. They experience a gentle descent followed by a more rapid one just upstream of the Andes at the same time that their path changes from anticyclonic to cyclonic.

- The pressure drop that signals the CL development is restricted to the lower troposphere over the region to the west of the subtropical Andes (27°–37°S). It maximizes inland and decays seaward over a cross-shore scale of ~500 km. The surface pressure drop is hydrostatically associated with a warming of the lower troposphere over the same region.

- The warming of the lower troposphere is in turn largely produced by enhanced low- and midlevel subsidence acting upon a highly stratified column. Offshore warm advection is mostly offset by alongshore cold advection over the coastal ocean. The large-scale subsidence is enhanced over the continent and off the coast due to the upwind-barrier effect of the western slope of the Andes Cordillera upon the synoptic-scale low-level easterly flow.

- Both the cross (zonal) and alongshore (meridional) coastal low-level winds are largely in geostrophic balance during the CL’s life span. Departures from geostrophy occurs at the CL onset due to sizeable isallobaric wind (~1.5 m s⁻¹). Surface drag deflects the near-surface winds toward the low pressure center. This later effect is particularly strong on the cross-shore surface wind, leading to a transition from easterlies to weak westerlies well before the CL culmination.

- Near the time of the CL culmination in central Chile
a wedge of weak northerly flow and SCu clouds advance poleward from the northern coast, in concert with a slight recovery of the surface pressure at the coast. The free propagation of this coastally trapped disturbance soon merges with the incoming tropospheric-deep, cyclonic circulation. We have refrained from analyzing the dynamics of this feature in detail, but we notice that the wedge’s cross-shore scale (~200 km) and propagation speed (~13 m s⁻¹) are in agreement with the characteristics of a coastally trapped Kelvin wave derived for this region.

• The model results reveal many details of the low-level flow inland, some of them supported by recently available observations. The strongest easterlies are found between 900 and 850 hPa within the narrow east–west valleys of the Andes, extending westward atop the cold air pool that forms over the central valley between the coastal and Andes ranges. The near-surface zonal flow also exhibits a marked diurnal cycle due to the superposition of the thermally driven slope flows upon the synoptically driven easterlies.

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Fig. 16. (a) SLP evolution at two points near the coast: 36°S, 72.7°W (thin line) and 30°S, 72.7°W (thick line). (b) SLP evolution at 33°S for a point at the coastline (71.5°W, thin line) and about 600 km offshore (76°W, thick line). (c) SLP differences evaluated used the grid values from (a) and (b), respectively. Data are from model domain 3.

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