Diurnal cycle of surface winds over the subtropical southeast Pacific

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Received 12 February 2008; revised 12 April 2008; accepted 30 April 2008; published 3 July 2008.

[1] The diurnal variability of surface winds over the SE Pacific is characterized with QuikSCAT data for years 2000–2006. The twice daily satellite passes occur around 0600 and 1800 LT (referred to as AM and PM, respectively). The PM-AM wind difference maximizes in a region between 20°S and 30°S, extending from the Chilean coast up to 75°W. This difference is mostly meridional (along the coast) with seasonal variation characterized by summertime weakening of the morning winds. The observational analysis is supplemented with mesoscale model results for December 2003. They show that the coastal zonal flow increases between 0600 and 1200 local time (LT) but then decreases in the next 6 hours, which explains the absence of an onshore component in the QuikSCAT PM-AM difference. Additional model diagnostics show that in this region the diurnal cycle hodographs are more simple and larger above the marine boundary layer (MBL) than at the surface. Evaluation of the momentum and temperature budgets reveals the importance of the diurnal cycle of vertical velocity coupled with the thermal structure in controlling the diurnal cycle of the low-level winds. Around 1300 LT coastal subsidence maximizes in a layer sloping from 1000 m at 36°S to 3000 m at 22°S, inducing the development of a coastal trough at the surface. During the night the passage of a high-pressure perturbation reverses the surface meridional pressure gradient force, inducing the decrease of the coastal meridional winds. These results stress the important role played by the dynamics of the low troposphere above the MBL (e.g., the heating of the Andes Cordillera to the east) in controlling the wind diurnal cycles near the surface.

Citation: Muñoz, R. C. (2008), Diurnal cycle of surface winds over the subtropical southeast Pacific, *J. Geophys. Res.*, *113*, D13107, doi:10.1029/2008JD009957.

1. Introduction

[2] The southeast (SE) Pacific region off north central Chile $(19^{\circ}-35^{\circ}S, 80^{\circ}-70^{\circ}W)$ is a region of considerable climatic interest. It encompasses a subtropical anticyclone that maintains an extensive and persistent deck of lowlevel stratocumulus (Sc) clouds which have a large impact on the planetary radiative budget [Hartmann et al., 1992]. To the east, the region is bounded by topography rising in less than 200 km from sea level to the peaks of the Andes Cordillera (heights >4000 m), leaving in between the extremely arid Atacama Desert. In previous work we have presented data analysis [Garreaud and Muñoz, 2005] and numerical modeling results [Muñoz and Garreaud, 2005] aimed at describing and understanding a low-level wind maximum which exists along the eastern border of the subtropical SE Pacific region. The present work extends the study into the diurnal cycle of low-level winds in the same region.

[3] The diurnal variability of low-level clouds in this geographical area has already been shown to be relatively

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large [e.g., *Rozendaal et al.*, 1995], and, although we do not dwell on this matter here, we may presume that diurnal variations of winds and clouds are well related dynamically and thermodynamically. In fact, for a geographical setting similar to the one considered in this work, *Koračin and Dorman* [2001] showed that off the California coast the diurnal variation of the nearshore cloudiness is highly coupled to the convergence and divergence of the low-level winds. More generally, right at the coast a band of afternoon clearing and onshore wind acceleration are wellknown features of sea breeze circulations [e.g., *Simpson*, 1994, p. 21] that may be important to explain the diurnal cycles in the SE Pacific region. However, the complex topography in this area, among other factors, can certainly complicate the way that the sea breeze takes form.

[4] Besides their expected interaction with the low-level cloudiness, the diurnal variability of low-level winds may also impact the oceanic circulation. Near the coast, for example, the nocturnal weakening of the winds induces a significant surface stress rotor important in the maintenance of coastal upwelling of deeper waters [*Rutllant et al.*, 2004]. *Lerczak et al.* [2001] show evidence of diurnal variability in subsurface currents that may be related to the diurnal component of the surface wind variability.

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[5] Satellite scatterometry has provided in recent years a wealth of information about surface winds over oceanic regions. QuikSCAT data used in this study have been available since July 1999. They consist of nominal 10 m wind speed and direction gridded with 25 km horizontal resolution. Besides their application in real-time assimilation into weather forecast models, averages of these data have proven useful in building up a climatology of oceanic winds and related fields such as vorticity and divergence [e.g., McNoldy et al., 2004]. Moreover, the fact that these data are generally available twice per day has allowed some inferences about the diurnal cycle of surface winds over the sea. Gille et al. [2003] used 3 years of QuikSCAT data to perform a global analysis of the difference between the morning and evening QuikSCAT wind fields. Their results were interpreted in the framework of land/sea breeze systems. For the SE Pacific region they found a statistically significant diurnal cycle signal with an orientation parallel to the coast, a peculiarity they suggest may be related to topographical constraints. Gille et al. [2005] refined the observational analysis of the diurnal cycle of oceanic winds by studying the period of April-October 2003 in which there were four daily scatterometer wind fields available (from two satellites in operation). On the basis of the averaged fields for these four observations per day they performed an elliptical fit to the diurnal cycle wind hodograph at a global scale. Again, the SE Pacific off the west coast of South America shows a significant amplitude of the diurnal cycle ellipse in a region extending more that 1000 km offshore (north of 30°S). The diurnal cycle ellipse is very meridional in the north central part of Chile. According to their results, the timing of the maximum meridional winds also shows a significant meridional change along the Chilean coast, occurring several hours earlier off the northern part of the country (and southern Peru) than off the central region of Chile.

[6] Prior to the scatterometer wind era, *Dai and Deser* [1999] had performed a global analysis of the diurnal and semidiurnal variability of surface winds based on land stations data and marine reports. With relatively coarse horizontal resolution ($\sim 2^{\circ}$), their database covers a long period (1976–1997) with a 3 h time resolution. For the eastern rim of the SE Pacific region their results show a maximum in the magnitude of the diurnal component of the variability of the meridional winds and wind speeds. Remarkably, their Figure 5.1 suggests that for the months of December–February the global maximum in the diurnal component of wind speed variability is found along northern Chile.

[7] In this work we center the analysis of QuikSCAT data and numerical modeling on the diurnal cycle of surface winds in the SE Pacific. The regional focus and the more extended QuikSCAT database that is now available allow a more robust and detailed analysis of the data, compared with the studies described above. Results of the data analysis are presented in section 2. Next, we use the Pennsylvania State University–National Center for Atmospheric Research (PSU-NCAR) mesoscale model (MM5) to perform monthlong simulations of the meteorological conditions over the region of interest. After validation against the available data, the model results are extensively diagnosed in section 3 in order to (1) produce a more complete description of the diurnal cycle in the region, (2) better understand its forcings, and (3) propose answers for some of the questions posed by the data analysis. Section 4 summarizes the results.

2. Data analysis and Model Validation

2.1. QuikSCAT Data Analysis

[8] The climatology of the surface wind field over the ocean is constructed with version 3a QuikSCAT data for years 2000-2006. The data are on a $0.25^{\circ} \times 0.25^{\circ}$ latitude/ longitude grid derived from the original swath data available at www.ssmi.com. There are two daily passes of the satellite occurring around 0600 and 1800 LT, which we call AM and PM wind fields, respectively.

[9] Mean features of the surface wind field obtained by averaging over the complete QuikSCAT database are shown in Figure 1. Figure 1a presents the vector average of the AM winds, showing the prevalence of southerly winds off the Chilean coast, as part of the SE Pacific anticyclonic circulation. The region of maximum winds around 30°S is the result of the frequent occurrence of a southerly lowlevel jet [Garreaud and Muñoz, 2005; Muñoz and Garreaud, 2005]. The mean PM-AM wind velocity difference fields are shown in Figure 1b. A conspicuous region of enhanced afternoon wind acceleration is evident east of 75°W and between 20°S and 30°S, with maximum magnitudes close to the coast. Two salient features of this region are (1) its relatively large offshore extent and (2) the orientation parallel to the coast shown by the PM-AM wind differences. By contrast, south of 30°S the PM-AM wind difference field has smaller magnitudes; it is much more constrained to the coast, and its orientation is more onshore.

[10] The statistical distributions of the meridional component, V, of the QuikSCAT winds for a four-pixel (100 km) coastal band are presented in Figure 2. Figure 2a shows that for the month of December and in the region between 20°S and 30°S the PM and AM meridional coastal winds are significantly different, with no overlap of their interquartile ranges. The annual variation of these distributions is illustrated in Figure 2b for the 2° latitudinal band centered at 27°S. Figure 2b suggests that the increased PM-AM difference in the summer months is due to a decrease of the AM meridional winds, the latter showing a more modest annual cycle.

[11] The fact that QuikSCAT data are available only twice per day limits the understanding that can be gained about the surface wind diurnal cycle based solely upon them. For example, the absence of a zonal component in the PM-AM vector fields in Figure 1b may just be due to the timing of the satellite passing above the region. Other related questions are the mechanisms behind the observed PM-AM meridional wind difference and the location and offshore extent of its region of maximum strength. More generally, it is of interest to examine if the observed diurnal cycle of the low-level circulation in this region fits into a standard sea breeze conceptual model or if the extreme topography and special climatic elements (Atacama Desert and SE Pacific Sc region) alter this model in a substantial way.



Figure 1. Global averages of QuikSCAT winds for years 2000-2006. (a) Vector average of AM velocity fields. (b) Vector average of PM-AM velocity difference fields. In each plot the contours show the magnitude of the corresponding vector field (labeled in m s⁻¹), and only one out of every three vectors in each direction is shown.

2.2. Model Setup and Validation

[12] We have used the PSU-NCAR MM5 [*Grell et al.*, 1994] to compute surface wind fields over the region of interest. One model domain is used with horizontal resolution of 30 km and 32 levels in the vertical (first model layer at 23 m above sea level and 12 levels below 1000 m). Turbulence in the planetary boundary layer is computed with the 1.5 order turbulent kinetic energy scheme Gayno-Seaman parameterization [*Gayno*, 1994; *Shafran et al.*, 2000]. Boundary conditions are taken from the National Centers for Environmental Prediction/NCAR Reanalysis, and sea surface temperature is kept fixed in time at the monthly mean field derived from advanced microwave scanning radiometer–Earth Observing System (AMSR-E) data obtained from www.ssmi.com (see Figure 3).

[13] The skill of the MM5 (with similar setups) to reproduce the synoptic variability and the large-scale features of the surface winds in the SE Pacific region has been shown elsewhere [*Muñoz and Garreaud*, 2005; *Muñoz et al.*, 2006]. Here we evaluate its performance in reproducing features of the diurnal cycle of the surface winds. For this purpose we concentrate on the period of 2003 in which a second scatterometer operated on the Midori-2 satellite, providing additional wind fields at about 1100 and 2300 LT. Unfortunately, this second satellite operated successfully only between April and October, a period in which, as we have shown, the diurnal signal in this region is not the strongest. Therefore we perform a validation model run for the period September–October 2003, but the full diagnosis of the diurnal cycle and the description of its forcings are outlined in section 3 based on a model run for December 2003.

[14] Figure 4 shows a comparison between satellite and model meridional winds for the average fields and for the two independent semidiurnal differences that can be computed from the available data. The region of the maximum meridional wind is well captured by the model, as well as the more stagnant region at the northeastern portion of the domain, the latter presumably related to the abrupt change in the orientation of the coastline and orography. Some of the magnitude differences of the winds are due to the fact that scatterometer winds are nominally representative of 10 m winds, while the first layer of the model results is located 23 m above the surface. In terms of the semidiurnal wind difference fields, their spatial patterns are relatively well captured by the model. In particular, the region of maximum V18 - V06 off the coast of Chile between 20°S and 25°S is well reproduced in shape and magnitude.

[15] Figure 5 compares model results against satellite scatterometer data for a point located at 24.125°S, 71.875°W. Figure 5a shows that the model reproduces reasonably well the interdaily variability of the meridional AM and PM winds at this location. To further evaluate the



Figure 2. Statistical box plots of coastal meridional winds based on QuikSCAT data for years 2000-2006. The boxes define the interquartile range of each distribution, and the horizontal line marks its median. Open boxes correspond to AM data, and shaded boxes correspond to PM data. The four data pixels closest to the coast are considered coastal, and the data have been binned in 2° latitudinal bands. (a) Latitudinal variation of coastal meridional wind distributions for the month of December. (b) Annual variation of coastal meridional wind distributions for the 2° region around 27° S.

model with respect to the diurnal variation of the meridional winds, Figure 5b shows the anomalies with respect to the daily mean based on model results and scatterometer winds for the 25 days in the period in which all four satellite passes were available at this point. The model reproduces well the tendency in the observations of having reduced meridional winds at the end of the night and in the morning and increased magnitudes and variability by late afternoon.

3. Diurnal Cycle Characterization

3.1. Assessing the QuikSCAT Timing Effect

[16] Model results for December 2003 are used here to build a more complete picture of the diurnal cycle of the low-level winds over the SE Pacific. In what follows, UTC minus 5 h will be used as local time (LT) for describing model results. Figure 6 addresses the question of why the PM-AM zonal wind difference in the QuikSCAT data is so small. We have plotted there the mean 6-hourly changes in zonal (Figures 6a–6d) and meridional (Figures 6e–6h) winds. Figure 6b shows that there is indeed an increase in the zonal winds along a coastal band between 0600 and 1200 LT, but in the next 6 hours they experience a comparable decrease in magnitude. The net result for the 0600-1800 LT period (similar to the integration period of the QuikSCAT data) is only a very modest mean variation of the zonal wind.

[17] Figures 6e-6h show the corresponding 6-hourly changes of the meridional wind. In the 0600-1800 LT period (Figures 6f and 6g) the coastal band off north central Chile shows a relatively persistent increase of this wind component, explaining the larger values of the PM-AM meridional wind difference captured by the QuikSCAT data. On the other hand, during the nocturnal phase of the diurnal cycle (Figures 6e and 6h) the meridional wind changes in the SE Pacific appear to be highly related to the surface pressure perturbation induced by the "upsidence wave" feature described by Garreaud and Muñoz [2004]. From the results above we conclude that the picture given by the QuikSCAT data of the diurnal cycle of the surface winds in this region may indeed be affected by the timing of the satellite passes and its relation with the phases of the daytime heating and the propagation of the upsidence wave perturbation during the night.

3.2. Diurnal Cycle Hodographs

[18] The average wind diurnal cycle hodographs computed by the model are shown in Figure 7 for the layer at 23 m (Figure 7a) and for that at 1404 m (Figure 7b) (the



Figure 3. Sea surface temperature (SST) (line contours in °C) used in model run for September– October 2003. Shaded contours (meters above sea level) correspond to the topography of the model.

mean wind vector has been subtracted at each point). The amplitude of the diurnal cycle hodographs at the surface is larger at the grid points closer to the Chilean coast, compared with grid points offshore. The structure of these coastal hodographs, however, can be quite complex. For example, between 20° S and 30° S the zonal wind diurnal cycle near the coast exhibits a double cycle in its diurnal variation giving the hodograph a characteristic "figure eight" form. South of 30° S the coastal hodographs are more like what may be expected for a simple coastal circulation driven by the land-sea temperature difference and the Coriolis effect.

[19] Figure 7b shows the diurnal cycle hodographs for a model layer well above the marine boundary layer (MBL). In a region between 20°S and 30°S and extending about 500 km offshore, the diurnal cycle hodographs are significantly "larger" above the MBL than at the surface. They are also much simpler, an indication that the double cycle in the zonal wind variation described above is a near-surface feature. South of 30°S the amplitude of the diurnal cycle hodographs at this level is very small.

[20] These model results suggest that north of 30° S the larger amplitudes of the wind diurnal cycles are found

above the MBL, while those near the surface present considerable complexity. In contrast, south of 30° S the diurnal cycles are larger in the MBL than above and are important in a region more constrained to the coast. Partial observational support of these results is given by *Rutllant et al.* [2003], who describe significant diurnal cycles of wind, temperature, and mixing ratio in a layer extending from 1000 to 4000 m above sea level according to rawinsonde measurements carried out at 23° S.

3.3. Diurnal Cycle Forcings

[21] Figure 8a shows the mean diurnal cycle of the meridional wind, V, at a point located at 25.93°S, 72.48°W for the region between the surface and 3000 m. Maximum meridional winds are found at about 600 m (top of the modeled MBL), reaching mean values over 14 m s⁻¹ between 1800 and 2200 LT. Figure 8b shows the time rate of change of the meridional wind. The diurnal cycle of V has larger amplitudes in the region below 2000 m, with maximum decelerations ($\partial V/\partial t < 0$, where t is time) around 0100 LT and maximum accelerations ($\partial V/\partial t > 0$) between 1300 and 1600 LT. Below 1000 m

Figure 4. Comparison of satellite scatterometer winds and MM5 results for September–October 2003. (a) Mean meridional wind at 1800 LT from QuikSCAT. (b) As in Figure 4a but for MM5 results. (c) Average difference of meridional winds between 1800 and 0600 LT satellite passes. (d) As in Figure 4c but for MM5 results. (e) Average difference of meridional winds between 2300 and 1100 LT satellite passes. (f) As in Figure 4e but for MM5 results.



Figure 4



Figure 5. Comparison of model results and QuikSCAT data for point located at 24.125°S, 71.875°W for September–October 2003. (a) Meridional winds at AM (circles) and PM (crosses) satellite passes. (b) Meridional wind diurnal anomalies from satellite scatterometer data (circles) and from MM5 results. Only 25 days with four satellite data per day are shown.



Figure 6. Averaged 6-hourly changes of (a-d) zonal and (e-h) meridional winds computed from MM5 results for December 2003. Shaded contours represent positive changes, and dashed contours represent negative changes. Contour values are in m s⁻¹ (the zero contour has been omitted).



Figure 7. Mean diurnal cycle hodographs computed from MM5 results for December 2003. The mean wind vector for each point has been subtracted from the hodographs. Crosses mark the grid point corresponding to each hodograph (one out of every six model grid points is shown). Hodographs begin at 0100 LT and end at 2400 LT, and a dot marks the 1800 LT point. In the top right corner there is a reference hodograph with a 2 m s⁻¹ radius. (a) Diurnal cycle hodographs for 23 m. (b) As in Figure 7a but for 1404 m.

the values of V increase continuously between 0700 and 1700 LT, reaching maximum values between 1700 and 1900 LT. The diagnosis of all forcings in the meridional momentum budget shows that the main factor responsible for this acceleration is the meridional pressure gradient force (PGF), whose mean time-height structure is shown in Figure 8c. Above 2000 m the mean meridional PGF at this point remains negative all day, consistent with a westerly geostrophic wind. Below 2000 m, however, the PGF is positive most of the day, maximizing at the surface at 1000–1300 LT and becoming negative between 2300 and 0500 LT. The latter feature is responsible for the nocturnal deceleration of the low-level meridional wind in this region, which, as we showed in section 3.1, is an important part of its diurnal cycle.

[22] The modeled mean diurnal cycle of surface pressure along the coast is shown in Figure 9. The coastal surface pressure has a relatively large diurnal cycle in the region north of 30°S. In northern Chile $(18^{\circ}-22^{\circ}S)$, minimum surface pressures occur between 1300 and 1600 LT, and maximum values occur at about 0100 LT. Although the timing of the minimum coastal pressure does not vary much latitudinally, the timing of the maximum surface pressure occurs about 6 h later at 30°S than at 20°S. The structure of the diurnal pressure cycle in Figure 9 shows that the pressure gradient along the coast maximizes its diurnal cycle in the transition region between 30°S and 22°S, where it even reverses sign during the nocturnal phase.

[23] Now we address the factors explaining the diurnal cycle of surface pressure in the model results. Figure 10a shows a time-height cross section of pressure anomalies computed by subtracting at each level the pressure averaged over the full modeling period. The diurnal pressure cycle has a very clear pattern with maximum magnitudes at the surface, maximum pressures around 0700 LT, and minimum values around 1600 LT. In turn, this diurnal pressure



Figure 8. Mean modeled diurnal cycles at 25.93°S and 72.48°W for December 2003. (a) Meridional wind (m s⁻¹). (b) Meridional wind speed change (m s⁻¹ h⁻¹). (c) Meridional pressure gradient (PG) acceleration (m s⁻¹ h⁻¹). Zero contours are solid, and negative contours are dashed. DVDT is rate of change of meridional wind speed.

cycle is largely explained by the diurnal variation in the temperature profile. Indeed, Figure 10b shows the diurnal cycle of temperature anomalies, which has a more complex structure and presents large magnitudes at the subsidence inversion level. These temperature anomalies have been hydrostatically converted into pressure perturbations (using a reference density of 1 kg m⁻³) and then integrated from 4000 m down to each level to produce the pattern shown in Figure 10c. The strong correspondence between Figures 10a and 10c suggests that most of the diurnal pressure cycle is thermodynamically driven by the diurnal temperature cycle vertically integrated over the lower troposphere.

[24] Next, we focus in Figure 11 on the diurnal cycle of temperature and its forcings. Figure 11a shows the timeheight structure of the diurnal temperature cycle at the same point analyzed before. Maximum temperatures at about 1000 m mark the top of the persistent subsidence inversion in the region. The structure of the temperature time rate of change is shown in Figure 11b, and the diurnal structure of its vertical advection forcing is presented in Figure 11c. The similarity between Figures 11b and 11c implies that vertical advection is a main forcing in the temperature budget of this region.

3.4. Diurnal Cycle of Vertical Velocity

[25] Section 3.3 has shown that the vertical velocity field (coupled with the thermal structure) is key in modulating the diurnal cycle of the lower tropospheric pressure field and the concomitant diurnal cycle of the low-level winds. We end this section, therefore, describing the structure of the diurnal cycle of vertical velocity, W, according to the model. Figure 12a shows the average time-height structure of W at the same point as that of Figures 8, 10, and 11. As expected, this field is dominated by subsidence, although with relatively large diurnal and vertical variation. Subsidence maximizes in the entire column around 1200– 1300 LT. This enhanced subsidence is present all along the Chilean coast, as seen in Figure 12b which shows a smoothed meridional-vertical cross section of the coastal Wfield averaged at 1300 LT. The maximum afternoon subsi-



Figure 9. Mean modeled diurnal cycle of surface pressure along the coast for December 2003. Contour labels are in hPa. Values are smoothed with a 5×5 grid point average.

dence occurs in a conspicuous layer sloping from ~1000 m at 36°S to ~3000 m at 22°S. The height, vertical extent, and slope of this afternoon subsidence layer suggest that the heating of the Andes Mountains to the east of the coast plays a role in forcing this diurnal circulation. A second feature of the *W* field in Figure 12a is a nucleus of positive values occurring around 0200 LT between 2000 and 4000 m. This feature does not occur simultaneously along all the coast (not shown), but it is a mark of the "upsidence wave" type of signal described by *Garreaud and Muñoz* [2004].

[26] The structure of the diurnal cycle of W below 1500 m in Figure 12a is more complex than above. Larger values of W are found at 0300 and 1800 LT, and minimum values are found around 2200 and 1200 LT, suggesting the existence of a double cycle in the 24 h period. This different character in the W diurnal cycles closer to the surface rather than farther above is more evident in the Hovmüller diagrams presented in Figures 13a and 13b. While the 2000–4000 m layer shows a one-cycle period in the W diurnal cycle, the 0–1000 m layer shows in a 300 km coastal band a clear two-cycle diurnal evolution. We relate this complex diurnal cycle of the low-level W to the complex low-level hodographs described in section 3.2. In fact, Figure 13c shows Hovmüller diagrams for the diurnal cycle of the zonal wind anomalies at the same

latitude as Figures 13a and 13b. In a coastal band similar to that in Figure 13b the surface zonal wind also shows a twocycle diurnal variation, a feature that gave the hodographs in Figure 7a their characteristic figure eight shape. Finally, Figure 13d shows the corresponding Hovmüller diagrams for the diurnal cycle of the MBL height anomalies. Minimum heights are reached between 1200 and 1400 LT, and maximum heights occur around 0600 LT. Again, in the 300 km band closer to the coast the mean diurnal cycles in MBL height are more complex than offshore. These results suggest that close to the coast the dynamical interaction of the MBL wind field with the coastal topography may produce convergence/divergence patterns that significantly alter the diurnal cycles of MBL height, W, and zonal winds. A more detailed study of this interaction, however, requires the application of simplified analytical models or numerical models with higher resolution, which falls beyond the scope of this work.

4. Summary and Conclusions

[27] The purpose of this work has been to provide an improved characterization and a better understanding of the diurnal cycle of the low-level winds in the subtropical SE Pacific region. Analysis of QuikSCAT data for years 2000–2006 has shown that the surface wind difference between hours 1800 and 0600 LT maximizes in a region extending



Figure 10. Mean modeled diurnal cycles at 25.93° S and 72.48° W for December 2003. (a) Pressure anomaly (hPa). (b) Temperature anomaly (K). (c) Hydrostatic pressure anomaly computed from temperature anomaly integrated down from 4000 m (hPa). Zero contours are solid, and negative contours are dashed. DP is pressure anomaly; DT is temperature anomaly.

between 75°W and the coast and between 20°S and 30°S. This PM-AM difference resides mainly in the meridional (alongshore) component of the wind. We have also shown that the PM-AM wind difference is larger in the warm period of the year, with the seasonal variation being stronger in the AM coastal winds than in the PM winds.

[28] The QuikSCAT data analysis has been supplemented with MM5 runs simulating conditions for the month of December 2003. Model results have a similar spatial structure for the region of maximum PM-AM wind field, compared with the QuikSCAT results. They show also that the diurnal cycle wind hodograph is larger above the MBL than near the surface.

[29] The observations and model results described here suggest that the diurnal cycle of the low-level winds in this region is not a simple sea breeze type of circulation driven mainly by the sea surface–land temperature contrast. It appears to be also influenced by the lower troposphere circulation, specifically, the diurnal cycle of vertical velocity in the 2000–4000 m layer. The daytime phase of this

diurnal cycle is characterized by enhanced afternoon subsidence all along the coast of north central Chile. The nighttime phase, on the other hand, is influenced by the passage of the vertical velocity perturbation described by *Garreaud and Muñoz* [2004]. Because of the large thermal stratification in this region, this diurnal variability of the W field aloft has a large imprint in the model surface diurnal pressure cycle. In the afternoon, a coastal trough develops with minimum pressures in northern Chile, while during the night the pressure gradient force along the coast reverses sign.

[30] The diurnal cycle of the MBL variables in a 300 km coastal band is more complex than offshore, with the zonal wind and the vertical velocity in the lowest 1000 m exhibiting a double cycle in their mean diurnal variation. We interpret this behavior as the response of the shallow MBL constrained to the east by the topography and forced by the W diurnal cycle aloft. In this interpretation, the region where the upper W diurnal cycle forcings and the



Figure 11. Mean modeled diurnal cycles at 25.93°S and 72.48°W for December 2003. (a) Temperature (°C). (b) Temperature hourly change (K h^{-1}). (c) Temperature tendency due to vertical advection (K h^{-1}). Zero contours are solid, and negative contours are dashed.



Figure 12. (a) Mean modeled diurnal cycle of vertical velocity (*W*) at point located at 25.93°S and 72.48°W. (b) Mean meridional-vertical cross section of coastal vertical velocity at 1300 LT smoothed with a 5×5 grid point average. Zero contours are solid, and negative contours are dashed. Contour units are cm s⁻¹. Averaging period is December 2003.



Figure 13. (a) Longitude-time cross section of vertical velocity (cm s⁻¹) averaged between 2000 and 4000 m at 26°S. (b) Longitude-time cross section of vertical velocity (cm s⁻¹) averaged between 0 and 1000 m at 26°S. (c) Longitude-time cross section of zonal velocity anomalies (m s⁻¹) at 23 m at 26°S. (d) Longitude-time cross section of boundary layer height anomalies (m) at 26°S. Averaging period is December 2003. Zero contours are solid, and negative contours are dashed. The diurnal cycles have been repeated in each plot for clarity.

MBL response add constructively is where the largest surface wind diurnal cycles are to be found.

[31] The exact nature of the interaction of the lowtropospheric flow with the Andes topography that produces the nocturnal upsidence wave and the afternoon coastal enhanced subsidence has not been addressed in this work. Sensitivity model runs with changes in the Andes topography could be used in the future to shed some light on this problem. Meanwhile the results herein stress the importance of adequate monitoring and modeling of the conditions in the low troposphere above the MBL if one is to reproduce appropriately the dynamics of the MBL in this region. This conclusion is particularly relevant in view of the upcoming Variability of the American Monsoon Systems (VAMOS) Ocean-Cloud-Atmosphere-Land Study Regional Experiment (VOCALS-REx) to be carried out by the end of 2008, aimed at better understanding and modeling the dynamics of this important stratocumulus capped MBL (R. Wood et al., 2006, VOCALS-SouthEast Pacific Regional Experiment (REx) Scientific Program Overview, available at http://www.eol. ucar.edu/projects/vocals/science_planning/VOCALS_ SPO_rev_aug06.pdf).

[32] Acknowledgments. The author thanks R. Garreaud, J. Rutllant, B. Barrett, and two anonymous reviewers for numerous suggestions that improved the paper. This research was supported by CONICYT (Chile) under grants ACT-19 and Fondef D05110038. QuikSCAT and SeaWinds data are produced by Remote Sensing Systems and sponsored by the NASA Ocean Vector Winds Science Team. AMSR-E data are produced by Remote Sensing Systems and sponsored by the NASA Earth Science REASON DISCOVER Project and the AMSR-E Science team. Data are available at www.remss.com.

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