The Diurnal Cycle in Circulation and Cloudiness over the Subtropical Southeast Pacific: A Modeling Study

RÉNÉ D. GARREAUD AND RICARDO MUÑOZ

Department of Geophysics, Universidad de Chile, Santiago, Chile

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ABSTRACT

The extensive and persistent deck of stratocumulus (Sc) over the subtropical southeast Pacific (SSEP) plays an important role in the regional and global climate. As in other subtropical regions, the Sc form at the top of a marine boundary layer (MBL), capped by the subsidence inversion. A distinctive feature of this subtropical deck is its pronounced dawn-to-afternoon decrease in cloud amount and liquid water path, partially associated with a regular and marked descent of the inversion base and the warming of the lower troposphere. Furthermore, coastal observations in this area reveal a diurnal cycle in air temperature encompassing up to 5 km MSL.

In this work, 15-day regional numerical simulations using the fifth-generation PSU–NCAR Mesoscale Model (MM5) in November (austral spring), May (late fall), and January (summer) 2001 were used to document the mean diurnal cycle in circulation and low-level cloudiness over the SSEP. The simulated amplitude, depth, and phase of the diurnal cycle in air temperature, wind, and cloudiness at the northern coast of Chile and over open ocean compare quite favorably with their observational counterparts.

Large-scale subsidence prevails over the SSEP on a daily average. Between 1 and 5 km, however, the vertical velocity exhibits a marked diurnal cycle, largely produced by a band of upward motion propagating from the southern coast of Peru into the SSEP during late afternoon and night. Such an “upsidence wave” was found in the three simulations. The upsidence wave produces a significant cooling, leading to a consistent diurnal cycle in air temperature in low- and midlevels over the SSEP. The impact of the vertical velocity cycle on the MBL was further studied using a 1D version of the MM5 with higher resolution. The deepening of the MBL during the upsidence period induces a more turbulent MBL and more entrainment. The warming and drying of the MBL result in a greater dissipation of the cloud layer in the afternoon, increasing the amplitude of the diurnal cycle in Sc cloud amount with respect to the cycle forced by absorption of solar radiation only.

1. Introduction

Large-scale subsidence occurs year-round over the subtropical southeast Pacific (SSEP), resulting in a quasi-permanent surface anticyclone centered roughly at 27°S, 90°W. This subtropical high drives south-south-easterly low-level winds along the west coast of South America, which in turn partially maintain a tongue of cold surface water due to coastal upwelling and equatorward alongshore advection. The cold SSTs and the adiabatically warmed air aloft leads to the formation of a cool marine boundary layer (MBL), often topped by an extensive deck of stratocumulus (Sc), and capped by a marked temperature inversion (e.g., Klein and Hartmann 1993). A distinctive feature of this subtropical Sc deck is its particularly pronounced diurnal cycle in cloud amount (Minnis and Harrison 1984; Rozendaal et al. 1995) and liquid water path (Wood et al. 2002), that is highly relevant to the quantification of the true impact of Sc on climate (Rozendaal et al. 1995; Bergman and Salby 1997).

The dawn-to-afternoon decrease in Sc cloud amount (e.g., Rozendaal et al. 1995) has been explained in terms of the afternoon decoupling of the cloud and subcloud layer within the MBL due to the radiative warming of the cloud (e.g., Turton and Nicholls 1987). The larger diurnal cycle over the SSEP, however, calls for other mechanisms in addition to solar warming. Thus, it has been speculated that diurnally varying divergence–convergence over the continent (central Andes and the Amazon) may propagate into the SSEP and contribute to the diurnal cycle of the Sc (Rozendaal et al. 1995; Gandu and Silva Dias 1998).

Ship-based, upper-air observations taken at 20°S, 85°W (Bretherton et al. 2004, see their Fig. 5) and along a transect at 27°S (Garreaud et al. 2001, see their Fig. 6) reveal that the afternoon thinning of the Sc is largely produced by a very regular descent of the inversion base, lending support to the diurnally varying divergence hypothesis. The vertical displacement of the inversion base is associated with a diurnal cycle in air temperature in the lower troposphere above the MBL.
Furthermore, Rutllant et al. (2003) report significant diurnal cycles in air temperature, wind, and moisture from 1000 m MSL up to about 5000 m MSL at Antofagasta, a coastal site at 23°S, 71°W (northern Chile).

Motivated by the observational evidence of such a significant diurnal cycle, the mean regional circulation and cloudiness over the SSEP is documented in this work using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5). Three simulations spanning 15 days were performed (November, May, and January 2001) with a finest grid spacing of 20 km. Whenever possible, we compared the model results with observations in order to evaluate the ability of the MM5 to simulate the regional circulation. The rest of the paper is organized as follows. The model setup is described in section 2. The diurnal cycle in air temperature and vertical velocity is presented in section 3, including a preliminary assessment of the seasonal changes of the diurnal cycles and a discussion of the dynamics at play. The diurnal cycle in cloudiness and the impact of the diurnally varying subsidence upon the Sc amount is analyzed in section 4, using a single-column (1D) version of the MM5. A summary of our main findings is presented in section 5.

2. Model setup

To obtain a three-dimensional description of the tropospheric circulation over the SSEP, multiday numerical simulations were performed using the PSU–NCAR MM5, version 3 (see Grell et al. 1994 for further details). The simulations feature two two-way nested domains with horizontal grid spacing (\(\Delta x = \Delta y\)) of 60 and 20 km (see details in Table 1) and 30 \(\sigma\)-levels in the vertical (\(\Delta z \sim 60\) m in the lowest 1 km). Domain 1 (coarser mesh) covers a significant portion of tropical and subtropical South America and the eastern Pacific, and domain 2 zooms in the SSEP (see for instance Fig. 2). Outputs from domain 2 were saved every 1 h. The parameterizations used in this simulation are listed in Table 2. Of particular relevance is the representation of the turbulence that controls the development and structure of the MBL and the Sc at its top. In these model runs we use a 1.5-order turbulence scheme that includes a prognostic equation for turbulence kinetic energy (TKE) and diagnosis of the mixing and dissipation length scales (Gayno 1994; Shafran et al. 2000). The turbulence model uses total water (\(Q\)) and liquid water potential temperature (\(\Theta\)) as conservative variables, and has an all-or-nothing treatment of saturation effects in the TKE buoyancy production term (Muñoz et al. 2000).

Three simulations were performed, starting on 14 November, 14 May, and 11 January 2001. Most of our analysis is concentrated on the November simulation (austral spring) when the Sc deck over the SSEP reaches its maximum extent off the coast of Peru and northern Chile (e.g., Klein and Hartmann 1993). In each simulation the model was initialized and continuously integrated for the next 15 days, using time-dependent boundary conditions interpolated to the boundary of domain 1 from the (National Centers for Environmental Prediction) NCEP–NCAR reanalysis grids (2.5° × 2.5° latitude–longitude) every 6 h. SST was held constant to its climatological value. The reanalysis data do capture the day-to-day variability in the SSEP (e.g., Garreaud et al. 2001) and our simulations can be considered as a dynamical downscaling of them.

3. Diurnal cycle in air temperature and vertical velocity

a. November 2001 simulation

The overall vertical structure and time variability of the mid- and lower-troposphere over the SSEP is synthesized in Fig. 1, by the time–height diagram of several variables at 21°S, 76°W, a grid column about 700 km off the coast of South America. The simulated MBL has a mean depth of 700 m, often topped by a cloud layer about 300 m thick, and capped by the trade inversion with a temperature increment of \(\sim 10\) K. South-southeasterly trade winds prevail in the lowest 1500 m MSL and westerly winds aloft. These results are in general agreement with ship observations interpolated to this point (see Garreaud et al. 2001 and Bretherton et al. 2004). The day-to-day variability features two tropospheric warming events that culminate on 20 and 29 November associated with easterly (offshore) flow at

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<th>Table 1. Geometry of the model’s domains (MM5-3D) for the regional numerical simulations.</th>
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<td><strong>Value</strong></td>
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<td>Number of grid points ((N_x \times N_y))</td>
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<th>Table 2. Physical parameterizations used in the MM5 (3D and 1D) for the regional numerical simulations.</th>
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<td><strong>Processes</strong></td>
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low- and midlevels, resembling coastal lows observed farther south (Garreaud and Rutllant 2003).

Most notable is the very regular diurnal cycle in potential temperature (Θ), with a pronounced maximum between 2100 and 0200 UTC [about 1600–2100 local time (LT)], similarly evident in other grid columns over the SSEP (see Fig. 1b). At 800 hPa the mean amplitude of the diurnal cycle exceeds 1.2 K from the South American coast westward to ~85°W and it is larger than the day-to-day standard deviation of potential temperature over most of the SSEP (Fig. 2). Thus, the mean diurnal cycle in Θ, and other variables can be properly documented using the values averaged over the 15 days at each hour (Θ̄h, h = 1, 2, ..., 24; the subscript h is omitted hereafter).

Vertical profiles of Θ̄ in three columns over the SSEP, generally representative of near-coastal and open ocean conditions, reveal a consistent cycle below 5 km MSL (Fig. 3). To show this cycle more clearly we subtracted the potential temperature basic state [Θ̄(z) = average over all hours and all days]. The first column (50 km off southern Peru) exhibits a maximum amplitude near 3 km MSL, with extremes occurring at 1800 (maximum) and 0200 (minimum) UTC. These extremes lead for about 5 h those of a distinctive cycle observed below 1 km (within the MBL). The second column (100 km off northern Chile) exhibits a near-simultaneous phase at all levels above 1 km, with extremes occurring at 1900 and 0300 UTC. Again, the free-tropospheric cycle leads by about 4 h the one in the MBL. The third column (about 1000 km off the coast) has a cycle similar to those near the coast, but with half their amplitude and extending down to the surface.

The simulated diurnal cycles compare favorably with available observations. First, the simulated amplitude, depth, and phase of the mean diurnal cycle in Θ̄, ρ, and v̄ off the northern coast of Chile are very similar to those measured by Rutllant et al. (2003) at Antofagasta (23°S, 71°W) during January 1997 and 1998 (see their Figs. 3a,e). The agreement is even better when comparing the observations with our January simulation (not shown) despite the difference in the year and length of the average. Second, Fig. 4 includes the simulated mean diurnal cycle of several variables at 20°S, 85°W, that can be compared with the mean diurnal cycles measured in October 2001 and reported by Bretherton et al. (2004; see their Fig. 5). Again, there is good agreement between the simulated and observed amplitude and phase of the diurnal cycle in cloud top and integrated cloud liquid water (CLW). Compared with the observations, however, the modeled MBL is significantly shallower, with cloud tops between 800 and 1000 m (as compared to observed values between 1200 and 1400 m above the surface). One-dimensional model calculations described...
Later suggest that this problem is related to a sensitivity of the turbulence model to the vertical resolution, and that with finer vertical spacing the calculated PBL depth can increase significantly (see section 4).

Returning to Fig. 3, the lower row shows the diurnal cycle in mean vertical velocity ($\overline{w}$) for the same three columns. Subsidence prevails during most of the day in the lower and middle troposphere, interrupted by a period of somewhat stronger ascending motion ($\overline{w} \sim +2$ cm s$^{-1}$). In all cases, the ascending motion defined as "upsidence" is limited below 5 km and peaks at about 3 km MSL, and tends to be flanked by periods of strong subsidence.\(^1\) For a point at 20°S, 85°W, the amplitude and phase of the simulated diurnal cycle in $\overline{w}$ at 850 hPa (not shown) is very similar to the mean vertical velocity derived from daily European Centre for Medium-Range Weather Forecasts (ECMWF) forecasts for October 2001 (Fig. 5 in Bretherton et al. 2004). Note the agreement in the timing of the ascent and cooling (compare upper and lower panels in Fig. 3). Application of the thermodynamic energy equation at various levels between 2 and 5 km confirms that the local rate of change that gives rise to the diurnal cycle in $\overline{\theta}$ is mostly due to the diurnally varying component of the vertical advection, whereas the basic state $\overline{\Theta}(z)$ results from a balance between day-mean subsidence warming and net radiative cooling (infrared cooling plus short-wave heating).

The diurnal cycle in vertical velocity in the free lower troposphere is further illustrated in Fig. 5 by the maps of $\overline{w}$ at 800 hPa. Qualitatively similar results are found between 850 and 500 hPa. Subsidence prevails over most of the SSEP during morning and afternoon (1000–1800 UTC). Along the coast of southern Peru, subsidence maximizes at 1600 UTC (local noon) evolving into strong ascending motion within the next 2–3 h. The upsidence region detaches from the continent by 2200 UTC (1700 LT) forming a crescent-shaped band about 400 km wide and 5 km deep that can be traced for at least 12 h as it moves southwestward at $\sim 1.8$ h$^{-1}$ (30 m s$^{-1}$; see also Fig. 6). Also evident in Fig. 5 is the offshore expansion of the area that experiences strong subsidence off northern Chile from 1000 to 1800 UTC, with isolines of $\overline{w} < 0$ moving westward at $\sim 1.7$ h$^{-1}$ (50 m s$^{-1}$). Below 850 hPa, the cycle in $\overline{w}$ is weaker and it has several maxima and minima. Still, upsidence prevails during nighttime over most of the SSEP. Interpolated to the inversion base ($z_i$), the cycle in $\overline{w}$ implies a vertical displacement of about $\pm 70$ m with respect to its daily mean height by the end of the upsidence (nocturnal)–subsidence (diurnal) periods, roughly 70% of the simulated amplitude in $z_i$. In section 4, however, we show that the modeled entrainment rate is not independent of the large-scale mean vertical velocity.

\(^{1}\) Near the coast there is a weaker diurnal cycle in $\overline{w}$ and $\overline{\theta}$ above 5 km, mostly out-of-phase with the cycle in the middle and lower troposphere, but its analysis is beyond the scope of this paper.
The mesoscale band of upsidence at 800 hPa along the coast of southern Peru during the afternoon requires vertically integrated convergence. At 800 hPa, for instance, a coastal band of convergence is collocated with the band of upsidence (Fig. 7b). Inspection of the horizontal divergence field reveals that convergence in this coastal area occurs at all levels below 700 hPa (including surface convergence), in contrast with large-scale divergence over open ocean and along the coast of northern Chile. The simulated band of convergence along the southern coast of Peru during the afternoon is in agreement with the seasonal mean fields of surface divergence derived from marine Comprehensive Ocean–Atmosphere Data Set (COADS) reports (Dai and Desser 1999) and from National Aeronautics and Space Administra\’s Quick Scatterometer (QuikSCAT) winds (R. Wood 2003, personal communication).

b. May and January simulations

As a first approximation to the seasonal changes in the diurnal cycle over the SSEP, we conducted a 15-day simulation for January 2001 (austral summer) and May 2001 (late autumn). During the austral summer, the Sc cloud cover off the coast of South America is minimum, whereas the area of strong deep convection over the continent reaches its southernmost position encompassing the southern Amazon basin and the Altiplano. By May, widespread deep convection has migrated to the north of the equator over Colombia and Central America.

The diurnal cycle in $\bar{\theta}$ at 800 hPa averaged between
Fig. 6. Space–time cross section of mean vertical velocity at 800 hPa for the (left) Nov, (center) May, and (right) Jan 2001 simulations. The cross section is taken along a line extending for 2000 km from the coast of southern Peru (17°S, 73°W) to 25°S, 80°W (see also Fig. 2b). The mean diurnal cycle is repeated twice and time advances upward. The vertical velocity is shown in units of 0.1 cm s⁻¹, contoured every three units, negative values in dashed lines, and the zero contour is omitted. Shading indicates upward motion.

Fig. 7. Daily mean wind and divergence at 800 hPa averaged between 1800 and 2100 UTC for the Nov 2001 simulation. (a) Wind vectors (scale at the bottom; m s⁻¹). (b) Divergence. (c) Zonal contribution to divergence (u/αx). (d) Meridional contribution to divergence (u/αy). Divergence and its two contributions are shaded according to the scale at the bottom in units of 10⁻² s⁻¹.

22°–18°S and 78°–74°W is shown for the three simulations in Fig. 8a. In all cases, the cycle features a diurnal warming and nocturnal cooling, with amplitude of 2 K in January, 1.2 K in November, and 0.7 K in May. As noted before, the cycle in  is largely dictated by the cycle in  shown for the three simulations in Fig. 8b. In both January and November the cycle in  exhibits a period of marked ascent that peaks at 0200 UTC, and a secondary maximum in  that does not exceed 0. In May, subsidence decreases during the early night but not enough to produce upsidence. Despite differences in timing and amplitude, the cycles in  largely result from the upsidence pulse that emanates from the southern Peru coast and the daytime increase of subsidence off northern Chile, and is qualitatively similar among the simulations (Fig. 6).

c. Discussion

The diurnal cycles in surface heating and deep convection over the continent are plausible sources of the diurnal cycle in vertical velocity over the SSEP. Determination of a cause–effect relationship, however, would require sensitivity studies (such as dry and no-mountain numerical simulations) that are beyond the scope of this work. Still, based on the model results from our “control” simulation (November 2001), we can speculate on the dynamics underlying the diurnal cycle in mean vertical velocity over the SSEP.

The marked diurnal cycle of deep convection over the Amazon basin and the central Andes (peaking about 1800 LT ~ 2200 UTC) during the austral summer has been suggested as a forcing of the diurnally varying
component of $w$ offshore (e.g., Rozendaal et al. 1995; Minnis and Harrison 1984). Mass adjustment to the vigorous upward flux within convective cells excites a series of gravity waves propagating outward from the area of deep convection, with the leading (fastest) mode inducing tropospheric-deep subsidence and surface divergence (Mapes 1993). Nevertheless, consistent diurnal cycles in $\bar{w}$ and $\bar{\theta}$ over the SSEP are mostly confined below 5 km MSL and peak at 3 km MSL. Furthermore, the diurnal cycles in $\bar{w}$ for May and November are similar to the one in January, despite the lack of deep convection over South America (south of the equator) from fall to spring. Thus, the effect of the diurnal cycle of continental deep convection seems to have little relevance in causing the diurnal cycle over the SSEP.

Given the hyper-aridity of the west coast of South America (in the latitude span of interest) and the prominent coastal topography (the Andes cordillera rises above 5000 m MSL within 100–300 km from the coastline), thermally driven circulations are likely to be deep and strong over this region. A west–east cross section of the mean zonal wind ($\bar{u}$) between 21° and 23°S at 1200 UTC (0800 LT; Fig. 9a) shows a weak land breeze flowing down the slope of the Andes (north–south oriented in this section) and merging with the SE trade winds further offshore. Subsidence prevails over the whole region. By 1800 UTC (1400 LT), a sea breeze develops inland also encompassing a band of about 200 km over the ocean² (Fig. 9b). The transition from offshore to onshore flow at low levels tends to increase the divergence over coastal areas, with $\partial \bar{w}/\partial x$ being the leading contributor to the afternoon divergence, and hence subsidence, at and off the coast of northern Chile (Fig. 7c). Inland, the sea breeze flows up the slope of the Andes producing strong but localized upsidence. In this context, the larger amplitude of the diurnal cycle in vertical velocity off the coast in January relative to November and May (Fig. 8), might be related with the larger dry surface heating over the Andes in austral summer.

² The P.M.–A.M. difference in zonal wind indicates the continental surface heating exerts its influence up to about 1000 km off the coast (not shown).
The previous argument could be applied to the coast of southern Peru. Nevertheless, the south-southeasterly low-level flow that prevails over the SSEP is roughly parallel to the coast of northern Chile but near normal to the coast of southern Peru (Fig. 7a). Considering the mean height of the Andes in southern Peru $H \sim 5$ km, typical upstream velocities $V \sim 5$ m s$^{-1}$, and static stability $N \sim 2 \times 10^{-2}$ s$^{-1}$, the resulting Froude number is $Fr \sim 0.05$, indicative of very strong blocking. The diurnal cycle of the mean meridional wind ($\overline{v}$) in this region is shown in Fig. 10 by a south–north cross section at $74^\circ$–$73^\circ$W. Over open ocean, $\overline{v}$ exhibits a morning-to-afternoon increase, likely the effect of the surface heating over the western slope of the Andes in northern Chile through geostrophic adjustment, while $\overline{v}$ remains small close to the coast. Off the coast of southern Peru, the afternoon low-level convergence, and hence upsurge, is largely due to $\partial\psi/\partial y$ (Fig. 7d), and we speculate that it is caused by the blocking effect of the Andes upon the strengthened meridional flow. Of course, the daytime development of a sea breeze in southern Peru also tends to increase subsidence off the coast. This sea breeze, however, is weaker and more localized than its counterpart in northern Chile, presumably due to the steeper slope of the Peruvian Andes.

The band of upsurge along the coast of southern Peru separates from the continent around 2200 UTC (1700 LT), coincident with the onset of the land breeze that subsides down the slope of the Peruvian Andes. In turn, the transition from up-to downslope flow occurs nearly simultaneously with onset of the surface cooling over the sloping terrain. Once detached from the continent, the band of upsurge propagates over the SSEP at $C \sim 30$ m s$^{-1}$ (Fig. 6). Using a height scale $H \sim 5$ km and a reduced gravity $g^* = g\Delta\theta/\theta_0 \sim g/35$ that is typical over the SSEP, the phase speed of a free gravity wave in this region is $c = (g^*H)^{1/2} \sim 35$ m s$^{-1}$. Thus, the “observed” propagation speed of the upsurge pulse agrees well with that of a gravity wave traveling against the mean flow.

4. Diurnal cycle in cloudiness

Figure 11 shows the average low-level cloud fraction field at 1400 UTC computed by the model for the No-
November 2001 simulation. The fractions refer to 1° × 1° boxes for which we counted the fractional number of grid cells that have nonzero integrated liquid water in the lowest 3000 m. These fractions were then averaged for the 15-day period. The cloud fraction field shows a band of maximum coverage that is parallel to the coast of southern Peru, where the model averaged cloud fractions are over 90%, in agreement with cloud amounts inferred from satellite images (Minnis and Harrison 1984). The averaged diurnal cycles in mean liquid water path (LWP) and cloud fraction for a 2.5° × 2.5° box centered at 20°S, 85°W is shown in Figs. 4b,c. The maximum LWP at the end of the night is similar to that reported by Wood et al. (2002) for a 2-yr average for the same area computed from satellite microwave radiometer data. The decrease of LWP during daytime is qualitatively similar to observations, although the model has a larger amplitude in the diurnal cycle. On the other hand, the amplitude of the average diurnal cycle in mean cloud cover for the same box is smaller than those reported by Minnis and Harrison (1984) for November 1978. If the threshold in integrated cloud liquid water used to define a cloudy column is increased from 0 to 5 g m⁻² this amplitude increases significantly, with minimum cloud amounts around 60% (the maximum cloud amounts are not too sensitive to this threshold) in closer agreement with Minnis and Harrison results.

The diurnal cycle in cloud amount and related parameters over the SSEP might be forced by the diurnal cycle in solar radiation, the diurnal cycle in vertical velocity, or a combination of both. In order to separate these two effects we performed additional simulations using a one-dimensional version of the MM5 model (MM5-1D). The MM5-1D includes the same physics as the MM5-3D, but is run over just one model column. The vertical velocity profile, the horizontal advection tendencies, and the geostrophic wind need to be prescribed. When forced with the hourly outputs (profiles) of the MM5-3D model run, we verified at several points that the MM5-1D was able to reproduce the 3D model results reasonably well. To assess more clearly any impact of the mean vertical velocity on the Sc cloud layer, however, it was necessary to run the MM5-1D with a larger vertical resolution in the PBL (~30 m in the first 1500 m above the surface) and the radiative computations were performed every 10 min, instead of every 30 min as in the MM5-3D model.

Two types of experiments were performed: in the first type (experiment WVAR) the hourly profiles of (large-scale) vertical velocity as extracted from the MM5-3D runs were used in the 1D model runs. In the second type (experiment WMEAN), the daily averaged vertical velocities were used at each vertical level. In both experiments the horizontal advective tendencies of temperature and water mixing ratio were also set to zero. Thus, the diurnal cycle in cloud-related parameters in experiment WMEAN can only be produced by the diurnal cycle in solar radiation, while their counterparts in experiment WVAR also include the forcing (if any) associated with the diurnal cycle in vertical velocity. Vertical sections of cloud liquid water content are presented in Fig. 12 for a point located at 21°S, 76°W, for day 16 November 2001. This point was chosen because it is located in the region where the signal of the up-sidence wave is strongest (e.g., Fig. 5).

Before addressing the differences between the two experiments, we first note in Fig. 12 that the depth of the PBL computed in these high-resolution runs is significantly larger than the PBL computed in the low-resolution 3D model run (between 200 and 300 m for this day and point). Lenderink and Holtslag (2000) showed also that the depth of a Sc layer computed by these type of turbulence models can have large sensi-

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*The 1D model runs were started at 1800 UTC of the previous day, so that the initial profiles coming from the low vertical resolution 3D model have time to adjust to the 1D vertical resolution.
tivity to vertical resolution. In cases with large-scale subsidence they described a numerical entrainment effect that tends to lock cloud top in a fixed model layer. Besides this numerical problem, the entrainment rate is sensitive to the specification of the turbulent length scales in the stable layer above the cloud, to the definition of the TKE profile, and to the formulation of the turbulent transport of TKE. All these factors could be playing a role in these model results as well.

The diurnal variation of the vertical velocity has a large impact on the depth of the PBL and the cloud layer (Fig. 12). During the first 6 h of the day, for example, when there was subsidence in the 3D model, the PBL rises about 200 m in experiment WVAR. By contrast, the results in experiment WMEAN show a relatively modest increment in the PBL depth which stays around 400 m above the surface. By the end of the day the cloud layer in run WVAR has fully dissipated, while in case WMEAN the Sc layer persists during all day, with only a thinning during the daytime hours. Thus, the alternance of upsidence and subsidence significantly enhances the amplitude of the diurnal cycle in cloud amount and LWP on top of its counterpart forced by solar radiation only. The different cloud fields in the WVAR and WMEAN experiments have an effect in the radiative fluxes reaching the surface. The 24-h average shortwave flux reaching the surface in experiment WMEAN is 17% lower than in the WVAR experiment, while the longwave flux is 7% larger. Such specific features of the diurnal cycle of the cloud layer shown in Fig. 12a will depend on the location and the phase correlation between the upsidence wave and the radiative diurnal cycle. The thinning of the cloud layer around 1000 UTC in Fig. 12a, for example, may be related to the increased subsidence that follows after the upsidence period earlier in the night (see Fig. 12c). A point located much farther from the coast—like that used in Fig. 4—does not show that feature in its averaged diurnal cycles (see Fig. 4a).

In an idealized model in which the vertical velocity does not affect the thermodynamic and turbulent structure of the PBL, the two experiments described earlier should produce similar results by the end of the day. In that case the vertical velocity has a reversible effect on the PBL. When positive, the vertical velocity deepens the PBL and increases the liquid water content at its top; when negative, it decreases the PBL height and reduces the cloud water at the top. The model results, however, suggest an irreversible effect of the vertical velocity. This irreversibility appears to be linked to an interaction between vertical velocity and the entrainment rate. The difference between the deepening rate of the PBL and the vertical velocity at the PBL top (Fig. 12c) is a measure of the entrainment rate of the turbulent layer into the stable layer above. During the first 6 h of the day the run WVAR has an entrainment rate about 50% larger than the run WMEAN.

Figure 13 shows the thermodynamic trajectories of...
vorably with their observational counterparts (also based on short records). Similarly, the simulated spatial field of cloud coverage and its diurnal cycle is in good agreement with satellite estimates.

The main discrepancy between model results and observations appears in the depth of the MBL. The model computes a boundary layer height shallower than what observations suggest. The 3D results and ongoing work using a higher vertical resolution in the 3D model suggest that in this climate regime the turbulence scheme is sensitive to the vertical grid resolution. It is possible that a deeper MBL might respond less strongly to the mean vertical motion, but it is also true that the magnitude of the mean vertical velocity acting at the top of the MBL will generally increase in the case of a deeper BL. More work is then needed to verify that the effect of the upwelling wave upon the MBL suggested by this work are still valid for a deeper and more realistic MBL.

As expected from its geographical location, large-scale, tropospheric-deep subsidence prevails over the SSEP on a daily mean basis ($\bar{\nabla}w \sim -0.5 \text{ cm s}^{-1}$). The vertical velocity field at low- and midlevels, however, exhibits a marked and very regular diurnal cycle, largely implied by an “upsidence wave” propagating from the continent into the SSEP. Strong ascending motion ($\bar{\nabla}w \sim +2 \text{ cm s}^{-1}$) first appears along the coast of southern Peru during afternoon, detaching from the continent by 2200 UTC (1700 LT). The band of upwelling is about 400 km wide and 5 km deep, and can be traced for at least 12 h as it moves southwestward at $\sim 1^\circ \text{ h}^{-1}$, resembling a free gravity wave. An increment of the subsidence off the coast of northern Chile during daytime, also contributes to the diurnal cycle in the vertical velocity over the SSEP.

Given the height scale of the upwelling wave ($H \sim 5 \text{ km}$) and its somewhat similar structure and evolution in spring, summer, and winter, we speculate that deep convection ($H \sim 15 \text{ km}$) over the adjacent continent (central Andes and Amazon basin) has little relevance in forcing the wave over the SSEP. Rather, we suggest a more prominent role of the mechanical blocking exerted by the steep, NW–SE-oriented Peruvian Andes upon the SE low-level flow over the ocean. The southerly flow along the coast of northern Chile accelerates during afternoon, in geostrophic response to the land heating, but remains near zero close to the Peruvian coast. The strong deceleration of the southerly flow leads to convergence, and hence upwinding in that coastal region.

The upwelling wave acting on a very stratified troposphere produces significant cooling through vertical advection. Thus, a consistent diurnal cycle in air temperature with a mean amplitude on the order of 2 K is found over most of the SSEP between 1 and 5 km MSL. The cycle is not really sinusoidal, with the cooling phase being the shorter/sharper. Assuming a constant entrainment rate, the cycle in $\nabla w$ also implies a vertical displacement of the inversion base of about $\pm 70 \text{ m}$ by the end of the upwelling/subsidence period. A simulation using a 1D version of the MM5 with finer vertical resolution, however, reveals that the diurnal cycle in $\nabla w$ does impact the modeled turbulence in the MBL. The significant deepening of the MBL during the upwelling period induces a more turbulent MBL and more entrainment. The warming and drying of the MBL result in a greater dissipation of the cloud layer in the afternoon, significantly enhancing the amplitude of the diurnal cycle in cloud amount and LWP with respect to the cycle forced by absorption of solar radiation only.

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