Mesoscale Wind Regimes in Chile at 30°S

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ABSTRACT

In November of 1999, four permanent surface stations were installed in the vicinity of the surface ozone monitoring station on the summit of the Cerro Tololo (2200 m MSL) in Chile at 30°S. These stations were used to study the atmospheric flow conditions, which are important for the interpretation of the ozone measurements at Cerro Tololo. In addition, radiosonde ascents were performed in March of 2000 near the coast and about 60 km inland. Different wind regimes were distinguished. Above 4 km MSL, large-scale westerly winds prevailed, while northerly winds were observed in a band along the coastline between 2- and 4-km-MSL height. The upper boundary of the northerly wind regime corresponded to the mean height of the Andes mountain range. This wind regime resulted from the westerly winds being blocked and forced to flow in parallel to the Andes (when Froude number is less than 1). The phenomenon was also confirmed by model simulations. Seasonally varying, thermally induced valley winds and a sea breeze developed below the northerly wind regime. In summer, the valley winds reached the Cerro Tololo. Diurnal variation of the top boundary of the valley winds also influenced the lower boundary of the northerly wind regime, which was less than 2 km MSL during the night and greater than 2 km MSL during the day. Thus, this observational and modeling study has shown that in summer the baseline ozone monitoring site at Cerro Tololo can be contaminated by polluted air that is transported from the plains by the thermally induced wind systems.

1. Introduction

As a part of the Global Atmospheric Watch (GAW) program, a monitoring station of radiation and surface ozone was installed by the World Meteorological Organization on the premises of the Cerro Tololo Inter-American Observatory, which is located near La Serena, Chile. Moreover, it is the objective of the “QHAWAY-RA” (Quechua for “air survey”) program to make this site a fully equipped GAW station (Gallardo et al. 2000). The GAW measurements describe long-term changes in the atmospheric conditions. This goal requires measuring sites at which the anthropogenic impact of sources in the area is either avoided or is identified such that the data can be stratified accordingly. For this reason, the locations of these stations are sometimes chosen on summits of high mountains such as Cerro Tololo. Cerro Tololo [2200 m above mean sea level (MSL)] is situated about 50 km east of the Chilean coast at 30°S, where the cities of La Serena and Coquimbo are located. The topography of the area can be described as very complex. The valleys around these mountains are deep, down to 500 m MSL, and the Andes mountain range is only 30 km east of Cerro Tololo with heights of up to 6 km MSL.

Regions with such complex terrain are frequently characterized by thermally driven wind systems, such as the sea breeze, mountain winds, valley winds, and slope winds (Atkinson 1981; Whiteman 1990), as well as by spatial variations of the boundary layer height (Dayan et al. 1988; Kalthoff et al. 1998; Koftmann et al. 1998). Often, the conditions for the evolution of more than one system exist at one site. So, merging may lead to large-scale wind systems (Kurita et al. 1990). The changes from day regimes to night regimes depend on such factors as sunrise and sunset, the orientation of the

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axis of a valley, and the size of the valley. All of this makes the wind field at ground level over complex terrain a very complex structure. These thermally induced wind systems are very important in connection with transport processes between flat areas and mountaintops (e.g., Bischoff-Gauß et al. 1998; Kalthoff et al. 2000; Fiedler et al. 2000) and may influence trace-gas concentration measurements on mountain summits.

Preliminary analysis of the Cerro Tololo data (Gallardo et al. 2000) indicated that there could be an influence of anthropogenic sources on the ozone measurements, especially during summer. When interpreting ozone data from Cerro Tololo, it is therefore important to understand the time and space variations in the regional wind field, which are caused by the sea breeze and valley wind systems between the coast and mountain ranges. So far, however, no data have been made available with regard to the regional wind systems between the coast and mountain ranges. Radiosonde data are only available from the routine radiosonde station in Quintero in the south at 33°S. Rutlant et al. (1998) investigated the boundary layer in the area of Antofagasta at 23°S. This project described here is therefore aimed at understanding the regional wind systems both by long-term characterization using observations and modeling and by a short-term intensive field campaign under late-summer conditions.

The following section provides some general background information about the topography and climatic conditions of the area. Section 3 gives an overview of the long-term measurement activities and describes the seasonal conditions of the wind field at ground level and the diurnal evolution of the boundary layer during the special observation periods (SOP) in the area around Cerro Tololo. Section 4 presents model calculations using the three-dimensional Karlsruhe Atmospheric Mesoscale Model (KAMM; Adrian and Fiedler 1991; Bischoff-Gauß et al. 1998) that help to understand some of the wind field characteristics observed.

2. Topographic characteristics and climatic conditions

The area investigated in Chile at 30°S includes the bay area of La Serena/Coquimbo in the west and the Andes in the east (Fig. 1). The Elqui Valley is one of the coastal breaks that penetrate eastward. The Andes in the east, with a mean height of about 4 km, represent...
an effective barrier to airflow. The Cerro Tololo is one of the major summits south of the Elqui Valley and is situated about 50 km east of the Pacific Ocean (Fig. 2). South of the Cerro Tololo another valley, the Limari Valley, penetrates eastward (Fig. 2).

The whole region is located in the southern part of the arid north of Chile, with the Copiapó River at 27°S marking the southern boundary of the extremely dry arid desert (Miller 1976). In the coastal region of La Serena, the mean annual precipitation is 114.5 mm yr\(^{-1}\) (Squeo et al. 1999). In Vicuña, about 60 km eastward in the Elqui Valley at an elevation of 606 m MSL, the mean annual precipitation amounts to 94 mm yr\(^{-1}\) [R. Alfaro, Instituto de Investigaciones Agropecuarias (INIA), 2000, personal communication; Fig. 2]. Where-as the rest of the year is very dry, 85%–90% of this annual rainfall is concentrated in a few winter months [May–June–July–August (MJJA)] (Weischet 1970; Miller 1976; Müller 1987). Outside of the Elqui Valley, the soil usually is very dry. The vegetation is sparse and is composed mainly of shrubs, herbs, and cacti (Miller 1976; Squeo et al. 1994; Olivares and Squeo 1999). Along the river banks, the ground is irrigated and cultivated, with fruit and wine plantations present. About 12 km west of Vicuña, construction of the Puclaro water reservoir was finished in 1999 and has been continuously filled with water since then.

The region of the Chilean coast at 30°S is influenced by the southeast Pacific high pressure area. Its center is situated at about 35°S, 90°W in January and at 25°S, 90°W in July. This situation results in a light southerly flow about 10 km offshore (Miller 1976). Along the shoreline, a sea-breeze system leads to westerly surface winds during daytime (Weischet 1996). A thin stratus layer below the subsidence inversion is frequently observed over the Pacific Ocean. It is caused by the cold oceanic Humboldt Current, which moves toward the equator along the Chilean coast. This stratus layer persists for most of the year. The base of the stratus cloud normally lies between 500 and 800 m MSL, and it is less than 250 m thick (Miller 1976; Weischet 1996). Breaks in the coastal range, such as the Elqui Valley, permit the fog to penetrate eastward. The penetration of fog into the valley and the duration can be determined roughly from the diurnal cycle of global radiation data as obtained for different sites in the Elqui Valley. The coastal fog does not affect the Cerro Tololo, so that the sky is cloud free for most of the year. The combination of low ocean surface temperatures and high land surface temperature—the latter resulting from dry soil, sparse
vegetation, and clear-sky conditions—is ideal for the development of thermally induced wind systems.

3. Surface and upper-level winds

The measurements focused on the Elqui Valley and two of its side valleys between La Serena, at the coast, and Vicuña. In addition to the ground-based stations operated by the Chilean Weather Service (La Serena Airport, Cerro Tololo), four continuously operating stations were installed in November of 1999. The positions of all the stations are shown in Fig. 2, and the parameters measured are summarized in Table 1. The stations at Pelícana and Puclaro are positioned close to the center of the valley. The Puclaro station is situated about 500 m in front of the 83-m-high dam of the new Puclaro water reservoir. The stations at Arrayán and San Carlos are located on the western slope of the side valleys, because the center of the valley may be flooded during El Niño events. The sampling rate of the parameters measured at the stations of Pelícana, San Carlos, Arrayán, and Puclaro is 5 s. The data are aggregated to 10-min-mean values. The parameters at the Chilean Weather Service station of La Serena are measured hourly. At Cerro Tololo, 15-min-mean values are obtained. During special observation periods in March 2000, radiosonde systems were operated at two sites, an Atmospheric Instrumentation Research, Inc., (AIR) global positioning system (GPS; Call 1997) at La Serena airport and an AIR radiotheodolite (Call et al. 1987) at Vicuña. Radiosondes were launched approximately every 3 h.

a. Characteristics of the near-surface wind field

Because there is little intraseasonal climate variability in this area, the diurnal cycles in the Elqui Valley reveal a high persistence from day to day. Thus,
mean diurnal cycles of the different meteorological components, $\bar{X}(t)$, for different seasons are calculated for the period from December of 1999 to November of 2000 according to

$$\bar{X}(t) = \frac{1}{N} \sum_{i=1}^{N} X_i(t),$$

where $N$ is the number of days and $X_i$ is the value on
Fig. 4. Mean diurnal cycle of the wind direction $w_d$ and wind speed $v$ for the (left) summer (Dec 1999–Feb 2000) and (right) winter (Jun–Aug 2000) at La Serena, Pelicana, and Puclaro. The measurement height of the sensors is given in Table 1.
the $i$th day at time $t$ of the day. During summer (December–January–February (DJF)), the mean sunshine duration is from about 0600 to 1900 LT (LT = UTC − 4 h) with mean maximum global radiation values up to $\overline{G} = 1100$ W m$^{-2}$ (standard deviation $\sigma_g = \pm 50$ W m$^{-2}$) at 1230 LT (Fig. 3). In winter (JJA), the sunshine duration is from 0800 to 1800 LT, and the mean maximum global radiation at 1300 LT is 600 W m$^{-2}$ ($\sigma_g =$...
Fig. 6. Relative frequency of wind directions separated into wind speed classes for summer (Dec 1999–Feb 2000) and winter (Jun–Aug 2000) in the Elqui Valley at La Serena, Pelícana, and Puclaro.

The mean daily temperature variation in the Elqui Valley is from 14°C to 29°C in summer and is from 9°C to 19°C in winter (Fig. 3).

As demonstrated by Fuenzalida (1996), the daily cycle of air pressure on the Chilean coast at 30°S is influenced by atmospheric tides (Haurwitz 1965) consisting of a diurnal and a semidiurnal component. This influence is evident from the pressure variation recorded at San Carlos (Fig. 3), where the effect of differential heating of sea and land surfaces is also evident [as noted by Kottmeier et al. (2000)]. A decrease of air pressure can be observed that is about 3.2 hPa in summer between 0830 and 1730 LT and about 2.4 hPa in winter between 1015 and 1600 LT. As compared with San Carlos, the mean global radiation at Pelícana in summer is smaller by up to 150 W m⁻², especially in the morning (Fig. 3). Also at Pelícana, the standard deviation is much higher in the morning [σ_G(1000 LT) = ±220 W m⁻²] and drops to about σ_G = ±65 W m⁻² at 1300 LT, which is similar to the value observed at San Carlos at that time. Reduced global radiation can be observed in winter until the late afternoon. The reduced radiation and the much higher standard deviation in the lower part of the Elqui Valley are caused by the marine stratus, which penetrates into the valley and sometimes reaches the area around Pelícana. The narrowing of the Elqui Valley east of Pelícana (Fig. 2) creates a natural barrier to further inland penetration of the stratus cloud, as is evident from field observations.

1) MEAN DAILY CYCLES OF THE WIND

The mean diurnal cycles of wind speed and wind direction in the Elqui Valley are shown in Fig. 4 for both the summer and winter months. At La Serena, Pelícana, and Puclaro, the wind speed at night is low in summer,
and only at Pelícana can a northeasterly mountain wind be observed in the early morning hours between 0300 and 0600 LT. The almost nonexistent nocturnal mountain wind regime is probably due to the long sunshine hours (Fig. 3). At La Serena, stable westerly winds occur at about 0800 LT (Fig. 4), associated with a slight increase of the mean specific humidity from 9 g kg\(^{-1}\) at 0600 LT to about 9.6 g kg\(^{-1}\) at 0900 LT (Fig. 3). This time probably indicates the onset of the sea breeze on the coast. At Pelícana, valley winds are established at about 0900 LT, and mean wind speeds of up to 9.5 m s\(^{-1}\) (\(\sigma_v = \pm 1.6\) m s\(^{-1}\)) appear at 10-m height at 1500 LT. These high wind speed values may result from merging of the sea breeze and valley winds.

In winter, the easterly mountain winds at La Serena, Pelícana, and Puclaro are much stronger. The change from the nocturnal wind regime to the daytime westerly winds can be observed between 1000 and 1100 LT at La Serena and Pelícana and between 1000 and 1200 LT at Puclaro, that is, 2.5–3 h after sunrise. The duration of the valley winds is reduced to about 10 h.

At Arrayán, wind direction in summer changes several times during the day (Fig. 5). At night, weak westerly winds prevail. At about 0800 LT, the wind turns to northerly valley winds with moderate wind speeds. During this time, the potential temperature and specific humidity increase continuously. However, from 1100 to 1800 LT the wind comes from the northwest, while the potential temperature and specific humidity are nearly constant in time (Fig. 3). The change from northerly to northwesterly winds appears to be associated with the passage of the sea-breeze front, although the two wind regimes would not be superpositioned in a simple manner. In winter, southerly valley winds appear at night. At about 1100 LT, the wind direction changes to a northerly mountain wind, that is, the sea breeze does not penetrate the area of the Arrayán Valley in winter.
at 0800 LT with maximum mean winds of about 7 m s$^{-1}$ ($\sigma_v = \pm 1.2$ m s$^{-1}$) at 10-m height (Fig. 5). However, the mountain wind from the south only occurs in winter. The onset of the valley wind in winter is between 0900 and 1000 LT, which is about 1 h earlier than in the broader Elqui Valley.

On Cerro Tololo, the diurnal cycle of the wind direction in summer indicates that three different wind regimes predominate during the day (Fig. 5). Weak northerly winds of 2 m s$^{-1}$ ($\sigma_v = \pm 1.0$ m s$^{-1}$) occur during the night. After sunrise, the wind direction gradually turns from north to south through east until midday. These easterly winds are accompanied by a temperature increase during the morning and appear to be associated with the valley wind of the San Carlos Valley. The mean diurnal temperature amplitude at Cerro Tololo is 6°C; the mean specific humidity is about 3.5 g kg$^{-1}$ with hardly any diurnal variation (Fig. 3). After 1300 LT, southwesterly winds of up to 5 m s$^{-1}$ ($\sigma_v = \pm 1.5$ m s$^{-1}$) predominate for about 7 h. This wind direction agrees with the orientation of the Cachacos Valley, a side valley of the Limari Valley, which ends at Cerro Tololo (Fig. 2) and indicates the arrival of valley winds. No significant mean temperature and humidity changes are associated with the onset of the southwesterly winds (Fig. 3). Additional carbon monoxide (CO) and black carbon measurements performed on Cerro Tololo in December of 2001 show that the easterly winds in the morning and the southwesterly winds at noon are accompanied by higher CO and black carbon concentrations. Because both species primarily are emitted near the surface, their presence at Cerro Tololo suggests the arrival of surface air from the valleys, although the physical processes involved do not ensure that the air follows the terrain. Because the San Carlos Valley wind system and the Cachacos Valley wind system are much weaker in winter, they do not penetrate to Cerro Tololo. Therefore, only the large-scale northerly flow can be observed at Cerro Tololo during the winter months.

2) Relative frequency of wind direction

Because of the large interseasonal variation of the solar radiation (Fig. 3), the wind direction roses at the different stations indicate large differences between the seasons. In general, the wind directions in the Elqui Valley are bidirectional (Fig. 6), that is, the wind blows parallel to the axis of the valley (Fig. 2). However, at La Serena, Pelícano, and Puclaro, the prevailing wind direction in summer is from the west, that is, hardly any easterly mountain winds occur, whereas in winter the contributions of westerly and easterly winds are nearly the same. Similar findings were obtained for the two side valleys of Arrayán and San Carlos, which are both oriented approximately north–south (Fig. 2). In San Carlos, north-easterly valley winds prevail in summer, whereas in winter the contributions of the northeasterly valley wind and the southerly mountain wind are nearly the same. At Arrayán, however, two predominant wind directions can be separated in summer—northerly valley winds and winds from the northwest (Fig. 7). As shown in the previous section, the northwesterly winds indicate the superposition of the sea breeze on the valley winds. In winter, the contribution of northwesterly winds is much lower and the contribution of southerly mountain winds is higher.

At Cerro Tololo, the predominant wind directions in summer are southwest and north; in winter the predominant wind direction is north. As proved in the previous
section, the wind direction turns to southwest when the valley wind from the Cachacos Valley arrives at Cerro Tololo. The reasons of the northerly flow will be discussed in sections 3b and 4.

b. Vertical structure of the wind regime

The vertical structure and diurnal cycle of the wind regimes on the coast at La Serena and inland at Vicuña were analyzed by means of radiosondes launched during the SOP on 28 and 29 March 2000. Three layers with characteristic wind systems can be distinguished (Fig. 8). First, westerly winds (southwesterly winds on 28 March, northwesterly on 29 March) prevailed above 3.5 km MSL on the coast and above 4 km MSL at Vicuña. This is the large-scale geostrophic flow, as is obvious from the 500-hPa pressure field (Fig. 9). Second, northerly winds occurred on the coast between 1.5 and 3.5 km MSL. At Vicuña, the upper level of the northerly flow is at about 4 km MSL; that is, it corresponds to the mean height of the Andes in that area. The lower limit of the northerly flow varies with time. During night, the limit of the northerly flow is at about 2 km MSL; during daytime the lower limit is at about 2.5–2.9 km MSL. Last, a sea-breeze system on the coast and a valley-wind system can be observed inland, both producing westerly winds during the day. On the coast, the sea breeze covers a layer of up to about 1.2 km MSL on 28 March and up to 800 m MSL on 29 March. At Vicuña, the valley wind reached up to 2.9 km MSL on 28 March and 2.6 km MSL on 29 March during the daytime, that is, an upward displacement of the lower boundary of the northerly flow can be observed. Because the top of Cerro Tololo is in the transition zone between the two wind regimes, the northerly flow can be observed on the mountaintop during the night, where as southwesterly valley winds prevail during the day (Fig. 5).

Two potential temperature and specific humidity profiles have been selected for 28 and 29 March 2000 to represent the day and nighttime conditions at La Serena and Vicuña (Fig. 10). At the coast, a daytime mixed layer can be seen, which reaches up to about 600 m MSL on 28 March and 1 km MSL on 29 March. This is in agreement with values reported by other investigations (Miller 1976; Weischet 1996). The inversion strength of the capping inversion layer is about $\Delta \Theta / \Delta z \approx 0.07$ K m$^{-1}$. Above the inversion, the potential temperature gradient is $\Delta \Theta / \Delta z \approx 0.0045$ K m$^{-1}$. Analysis of the depth of the sea-breeze winds and the temperature profile shows that the sea-breeze winds on 28 March reach beyond the top of the capping inversion layer.

At Vicuña, a strong surface inversion ($\Delta \Theta / \Delta z \approx 0.05$ K m$^{-1}$) with a depth of 400 m developed during the night. However, no typical well-mixed layer was established during daytime. Only a shallow unstable layer of about 100 m could be observed, and the atmosphere above remained stably stratified throughout the day with a potential temperature gradient of $\Delta \Theta / \Delta z \approx 0.0045$ K m$^{-1}$. The specific humidity decreased with height from about 8 g kg$^{-1}$ at the surface to about 3 g kg$^{-1}$ at 3 km MSL. The layer in which significant diurnal temperature changes can be observed (2.9 km MSL on 28 March and 2.6 km MSL on 29 March) coincided with the westerly valley winds.

4. Characteristics of the northerly flow

The observations show that in summer the lower boundary of the northerly flow varies around the top
of Cerro Tololo, and therefore different air masses influence the ozone measurements made at that site (Gal-\ lardo et al. 2000). To obtain an insight into the causes of the northerly flow, the Froude number Fr has been calculated from the observations. The Froude number is the ratio between the inertial force and the buoyancy force,

$$ Fr = \frac{\pi U}{NH} \quad (2) $$

(Stull 1988). Here, $N$ is the Brunt–Väisälä frequency, defined as

$$ N^2 = \frac{g}{\Theta \Delta z} \left( \frac{\Delta \Theta}{\Delta z} \right). \quad (3) $$

where $g$ is the acceleration of gravity. A Froude number of less than 1 indicates that the air will tend to be blocked by the mountain, whereas for a Froude number greater than 1, the air will tend to flow over the mountain. Taking into account the observed temperature gradient of $\Delta \Theta/\Delta z \approx 0.0045$ K m$^{-1}$, characteristic wind speeds of about $U = 5$ m s$^{-1}$, and a mean height of the Andes of $H = 4$ km, $Fr \approx 0.35$ results. So, it can be

Fig. 10. Profiles of $\Theta$ and $q$ at (left) La Serena and (right) Vicuña at different times on (top) 28 Mar and (bottom) 29 Mar 2000.
assumed that the westerly winds are blocked by the Andes and mountain parallel barrier winds will occur.

Blocking and mountain parallel winds are not very uncommon. They have been observed in many parts of the world, for example, in New Zealand, Iceland, Taiwan, the Alps, and the Andes (Parish 1982; Smith 1982; Barry 1992; Chen and Smith 1987; Li and Chen 1998; Garreaud and Wallace 1998; McCauley and Sturman 1999; White- man 2000). Mountain ridges may block the airflow very far upwind (Meroney 1990). Schwerdtfeger (1979) estimated a width of at least 100 km for the barrier winds for the Antarctic Peninsula. The maximum width upstream of the barrier is controlled by the Rossby radius, or radius of deformation, \( L = N H / f \) (Pierrehumbert and Wyman 1985; Chen and Smith 1987; Li and Chen 1998). Taking into account the values from above and Coriolis parameter \( f = -7.3 \times 10^{-5} \text{ s}^{-1} \) gives a value of \( L \approx 660 \text{ km} \) for the area under consideration. An explanation is given by Smith (1982, 1990). Whenever a large-scale flow is directed toward a mountain barrier, the air is forced to rise. Rising air and the accompanying adiabatic cooling aloft result in a higher pressure on the windward side. Considering the forces acting on a parcel of air approaching a mountain means that when “the air parcel nears the windward side of high pressure anomaly, it decelerates and the geostrophic balance it enjoyed initially is disturbed. The slower moving parcel feels a decreased Coriolis force and it thus begins to turn right (in the Southern Hemisphere) under the influence of the background pressure gradient” (Smith 1982). Because the mountain acts to force the motion, the barrier winds are confined mainly to the levels below the mountain crest (e.g., Parish 1982), as is also valid for the barrier flow described in section 3b.

Another reason for the change of the geostrophic wind with height can result from the thermal wind component (McCauley and Sturman 1999). However, as is obvious from the profiles of the potential temperature in Fig. 10, no significant spatial temperature differences and no temporal temperature variations can be observed in the layer between 2 and 4 km. As a consequence, this reason for the generation of the northerly flow can be excluded.

### Model calculations

Model calculations were performed with the three-dimensional KAMM to obtain insight into the mechanism that leads to the northerly flow. The entire model consists of two components, namely, the atmospheric model (Adrian and Fiedler 1991) and the soil vegetation model (Schädler et al. 1990).

The KAMM three-dimensional nonhydrostatic numerical simulation model solves the momentum, heat, and humidity equations. The inelastic form is applied to filter out the sound waves. The equations are transformed to a coordinate system which follows the terrain. The model is conceived for the meso-\( \gamma \) and micro-\( \alpha \) scale. It is driven by a large-scale or synoptic basic state that is given geostrophically and hydrostatically. Parameterization of the turbulent fluxes takes place according to the eddy diffusivity concept, in analogy to molecular exchange. It is assumed that the stress tensor is proportional to the deformation of the velocity field with the turbulent diffusion coefficient being applied as a proportionality factor. In the case of stable thermal stratification, the diffusion coefficient is determined by means of functions according to Businger et al. (1971). They are dependent on the local Richardson number. In the case of convective conditions, a nonlocal closure of Degrazia (1988) is applied. In this case, the turbulent diffusion coefficient is determined as a function of the boundary layer height, for which an additional prognostic equation has to be solved. The scheme of Orlanski (1976) is adopted for the lateral boundaries.

The soil vegetation model integrated in KAMM consists of two parts. In the first part, the soil model serves to determine prognostic equations for soil temperature and soil water content in a sufficiently thick soil layer. The second part, the vegetation model, corresponds to the one-layer model of Deardorff (1978). This means that the vegetation is not resolved in the vertical direction but is considered to be a homogeneous layer or so-

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**Fig. 11.** Profiles of \( \Theta \) (solid line) and \( q \) (dashed line) as used for the model initialization at 0000 LT.
Fig. 12. Horizontal wind fields for 1200 LT at the surface and at 1, 3, and 5 km MSL.
called big leaf between the ground surface and the lower atmosphere.

As input data, the model requires large-scale meteorological data that drive the processes in the simulation area: the terrain altitude as a function of the geographical location, data on soil, and vegetation types. The large-scale flow is described by the components of the geostrophic wind and the large-scale field of potential temperature. The model’s output are space-dependent and time-dependent distributions of wind (horizontal and vertical components), potential temperature, and specific humidity.

A horizontal grid spacing of $\Delta x = 4824$ m and $\Delta y = 4620$ m was used. With $101 \times 121$ grid points in the horizontal directions, a simulation area of $482.4 \times 544.4$ km$^2$ (from $32.5^\circ$S to $27.5^\circ$S, and from $72.5^\circ$W to $67.5^\circ$W) was obtained. In the vertical direction, a much denser grid size was used, with a vertical separation of about 25 m close to the ground and about 600 m at the upper boundary of the model area at 15-km altitude. The model run was initialized at 0000 LT with a large-scale westerly geostrophic flow of 5 m s$^{-1}$ for the whole model domain and a horizontally homogeneous temperature and humidity profile as obtained from the radiosonde ascents in La Serena during the special observation period in March of 2000 (Fig. 10). The initial conditions assumed were $\Theta = 288$ K from 0 to 750 m MSL, $\Delta \Theta/\Delta z = 0.06$ K m$^{-1}$ from 750 m to 1 km MSL, $\Delta \Theta/\Delta z = 0.005$ K m$^{-1}$ from 1 to 3.5 km MSL, and $\Delta \Theta/\Delta z = 0.002$ K m$^{-1}$ from 3.5 to 15 km MSL (Fig. 11). The sea surface temperature of the Pacific Ocean was held constant at 288 K, and a cooling rate of 0.2 K h$^{-1}$ was applied to the stratus layer between 500 and 800 m MSL over the Pacific Ocean to simulate radiative cooling. The soil temperature was initialized with the surface air temperature. Loamy sand was used as soil type, the volumetric soil moisture was set to 0.03 m$^3$ m$^{-3}$, and no vegetation was used in the model domain. Two model runs were performed: 1) using real topography and 2) using an artificial topography derived by reducing the real topography by a factor of 4. This reduction of the topography was chosen for the Froude number to become greater than 1, causing airflow over, rather than around, the mountains.

Figure 12 shows the wind fields at the different levels at 1200 LT, as calculated with the real topography data. The surface wind field reveals that the sea breeze is established and valley winds can be observed in the different valleys, such as the Elqui and the Limarí. At 1 km MSL, however, northerly winds can be found offshore while westerly winds prevail onshore. At 3 km MSL, the area of the northerly flow includes about 100 km of the ocean and about 60 km of the coastal plains toward the Andes. The remaining westerly winds represent the upper branches of the valley flows streaming over the passes of the Andes. The 5-km-MSL level already is above the top of the northerly flow regime. Hence, large-scale westerly winds can be found along the whole Chilean coast and are disturbed over the high mountains. Figure 13 shows the horizontal wind field in a vertical section through the model domain at 30°S. The area of the northerly flow agrees well with the observations at La Serena and Vicuña (Fig. 14). The top of the northerly flow regime increases slightly toward the Andes, whereas the top of the valley wind system increases much more steeply.

The results from the model calculations performed with the artificial topography are shown in Fig. 15. The near-surface wind field exhibits northwesterly winds offshore. At 1200 LT, the sea-breeze front penetrates...
about 50 km inland. Sea breeze and valley winds are much weaker than for the real topography (Fig. 12). At 1 km MSL, a northwesterly flow can be detected in most parts of the model domain while in the mountains the upper branches of the valley wind can be observed. At 2.5 km MSL, the large-scale geostrophic winds can be found. Only some disturbances can be found on the lee side of the Andes.

5. Conclusions

The area between the Chilean coast and the Andes at 30°S is characterized by several wind regimes. Above about 4-km height, the large-scale westerly flow prevails. Below this level, the Andes act as a barrier, forcing the large-scale westerly winds to flow parallel to the Andes, because the Froude number is less than 1. As a consequence, a permanent northerly flow with a vertical depth of about 2 km is observed. The lower boundary of the northerly flow varies in time and space because it depends on the temporal evolution of the height of the thermally induced wind systems. In summer during the night, the summit of Cerro Tololo is affected by the northerly flow; during the day it is influenced by the valley wind. During winter, the valley wind system does not penetrate to the summit of the Cerro Tololo. Hence, northerly flow prevails during the whole day.

The northerly flow was reproduced by model calculations using the KAMM three-dimensional model. Reduction of the model topography by a factor of 4, which results in a Froude number of greater than 1, leads to the reduction of the northerly flow in both horizontal width and vertical extension.

The results can now be used for the interpretation of the pronounced diurnal variation of ozone in summer at Cerro Tololo (Gallardo et al. 2000). It probably is caused by the different transport paths of air during the course of the day. Furthermore, model calculations with the nested version of KAMM will be performed to calculate air pollutant transport by thermally induced wind systems from the main emission sources in the bay area (the cities of La Serena and Coquimbo) and smaller urban areas in the vicinity of Cerro Tololo (Vicuña and Andacollo) and to investigate the influence of the transport on the observations made at Cerro Tololo.

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Fig. 15. Horizontal wind fields for a topography reduced by a factor of 4 for 1200 LT at the surface and at 0.5, 1, and 2.5 km MSL.
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