Daytime Mixed Layer over the Santiago Basin: Description of Two Years of Observations with a Lidar Ceilometer

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ABSTRACT

Two years of high-resolution backscatter profiles obtained with a commercial lidar ceilometer in Santiago Basin (33.5°S, 70.6°W) are analyzed. The generally large aerosol load in the Santiago atmospheric boundary layer (ABL) facilitates the use of these backscatter profiles for the retrieval of the daytime mixed layer height (MH), especially around midday. In winter mornings, however, MH retrievals are frequently confused by upper residual aerosol layers, while in summer afternoons very low aerosol concentrations often preclude them. Based on a database formed with successful MH retrievals over Santiago, the hourly, synoptic, and seasonal variability of clear-day MHs are documented. Daytime growth rates of MH show typical values of 50 m h⁻¹ in winter and 100 m h⁻¹ in summer. MHs at 1200 LT (UTC − 4 h) present a fourfold change between the cold months (MH ≈ 200 m) and the warm months (MH ≈ 800 m). Interquartile ranges of the monthly distributions of MH are about 200 m, with little change along the seasons. This statistical description of MH data is supplemented with analysis of temperature, solar radiation, and wind data at selected stations located at the basin’s floor, at the basin’s entrance, and at an elevated location representative of a level about 400 m above the basin’s floor. Seasonal MH variability appears highly controlled by the surface energy budget, with about 30% of the top-of-the-atmosphere radiative energy being used in the warming and growth of the daytime MH. Advection of cool air from the marine boundary layer to the west of the basin also appears to be important in the basin’s ABL energy budget in some cases. Stability of the early morning temperature profile in the basin’s air mass is also a factor in the mixed layer growth. Under conditions of large bulk stability of the basin’s air mass, there exist cases of very shallow daytime mixed layers that appear to develop after nights in which the stability is highly enhanced near the surface. Results herein are a first step toward a better understanding of the dynamics of this complex terrain ABL.

1. Introduction

The depth of the mixed layer (ML) is a crucial variable in air pollution problems, since it determines the amount of vertical dilution that surface-emitted pollutants can attain. The ML height (MH) over a particular location has a large variability at seasonal, synoptic, and hourly scales, controlled mainly by the surface energy budget and by the lower-tropospheric stability, the low-level winds, subsidence, and horizontal advectons (Stull 1988). Despite its importance, long and continuous MH datasets are scarce because of the cost and difficulty of the measurements. Several algorithms have been used to derive MH from radiosonde profiles (Seibert et al. 2000), but these data have poor hourly and spatial resolution if the synoptic radiosonde network is used as data source, or have only a short-term scope if special intensive observation periods are used. Recently, various remote sensing techniques of the lower troposphere have been applied to derive mixing heights, opening the path to continuous and long-term monitoring of this variable (Emeis et al. 2008). In this work we use MHs derived from a commercial lidar ceilometer operating since 2007 in Santiago, Chile. Retrieval of MH with this technique is based on the shape of the backscatter profile observed during clear conditions, taking advantage of the fact that the relatively high concentration of aerosols in the atmospheric boundary layer (ABL) produces a well-defined layer of larger values in the backscatter profile (see section 2).
Because of the importance of aerosols for the description of the ML by means of a lidar ceilometer, the Santiago Basin can be considered a rather idealized environment for the application of this technique. The city of Santiago, Chile (33.6°S, 70.7°W), has a serious air pollution problem, which during the Southern Hemisphere cold season (April–August) manifests itself in high concentrations of particulate matter, as measured for example by the concentration of particles with size less than 10 and 2.5 μm (PM10 and PM2.5, respectively; Koutrakis et al. 2005). Together with natural and anthropogenic emission of pollutants, geographic and meteorological factors contribute to explain the magnitude of the air pollution problem of Santiago (Romero et al. 1999; Jorquera 2002). The city is located in a basin confined by the Andes Mountains to the east, the Coastal Range to the west, and with hilly chains partially blocking its northern and southern limits (Fig. 1a). Thus, the basin consists of a 30 km × 90 km floor area, confined by mountainous slopes from 500 to 3000 m high. The synoptic and mesoscale meteorological factors of the air pollution problem of Santiago have been investigated by Rutllant and Garreaud (1995) and Garreaud et al. (2002). The climate in the region is dominated by the subtropical anticyclone of the southeastern Pacific (Fig. 1b) and its attendant subsidence inversion in the lower troposphere. This large-scale environment provides a large prevalence of clear days with limited synoptic variability, which at this latitude is strong mainly in winter months, when midlatitude frontal systems pass through the region. Subsynoptic disturbances, identified as coastal lows, propagating southward occasionally strengthen the subsidence inversion and the stability of the lower troposphere, resulting in wintertime clear days with...
elevated PM10 concentrations (Garreaud et al. 2002). Unfortunately, Santiago lacks continuous monitoring of meteorological variables in the vertical, so that documentation of the effects of synoptic and mesoscale factors on the development of the ABL over the basin has been heretofore very much restricted. Nonetheless, Rutllant and Garreaud (2004) have described with some detail the wintertime diurnal cycle of the winds above the basin. Below 900 m AGL, there is a marked diurnal cycle in the winds, with maximum intensities in the afternoon and SW wind direction. During nighttime, surface winds are light with an average easterly direction, while more intense easterly flow is found around 600 m AGL. These characteristics are consistent with a mountain–valley breeze system.

Under the topographic and climatological conditions described above, the daytime boundary layer over Santiago may be characterized as a convective boundary layer (CBL) influenced by complex terrain effects, particularly those associated with local circulations in the basin, and those related to plain-to-basin (or coast-to-basin in this case) circulations. Whiteman (1990) reviewed observations and conceptual models applicable to this type of ABL. Right from the beginning, the stability of the near-surface air layers in which this CBL develops may be enhanced by topographical effects, for example, cold air drainage and accumulation at the basin’s floor (Kondo 1995), stagnation of cold air in the basin and warming aloft (Whiteman et al. 2001; Wolyn and McKee 1989), or reduction of nighttime turbulent sensible fluxes and concurrent intensification of surface radiative cooling (Garratt 1992). Later, during the morning breakup of the nocturnal inversion, the CBL growth over the basin’s floor may be slowed down by enhanced subsidence compensating for the up-slope flows that develop at the basin’s sidewalls (Whiteman et al. 1996). During daytime, a larger thermal amplitude develops in the basin, as compared to nearby plains, due to a comparatively reduced air volume receiving a possibly larger sensible flux at the surface (de Wekker et al. 1998). In turn, this increased warming of the basin’s air mass induces the formation of a thermal low pressure surface and a secondary maximum between 400 and 500 m above the surface showing higher concentrations below 200 m above the surface.

A Vaisala CL31 ceilometer has operated since March 2007 over the roof (15 m above ground level) of the Department of Geophysics Building in downtown Santiago, Chile (point C in Fig. 1a). The analysis period considered here extends from March 2007 up to April 2009, during which time the ceilometer had been in operation continuously, except for brief power outages, occasional malfunctions, maintenance checks, and three weeks in October 2008 in which the instrument was moved to a location outside of Santiago Basin. A basic technical description of the CL31 ceilometer is found in Munkel et al. (2007). The setup of the instrument in Santiago provides reports of backscatter profiles every 4 to 8 s with a vertical resolution of 20 m.

During clear days, the instantaneous CL31 backscatter profiles near the surface often provide a magnificent depiction of the evolution of the boundary layer over Santiago. An example of this is presented in Fig. 2a for 30 August 2007. Colors in this figure are scaled to backscatter values of the instantaneous CL31 measurements, and they will be interpreted henceforth as directly related to aerosol concentrations. Between 0000 and 0900 LT1 in Fig. 2a, aerosols are not well mixed in the vertical, showing higher concentrations below 200 m above the surface and a secondary maximum between 400 and 600 m. The aerosol layer above 200 m can be interpreted as a residual boundary layer of the previous day.

Section 2 of this paper describes the ceilometer data, the algorithm applied to derive MHs, and the ancillary meteorological data used in the analysis. Section 3 describes the variability of MHs in Santiago, and explores the energy and stability factors that control this variability. The statistical description of MH in Santiago presented herein has applications beyond the direct documentation of its climatology. On one hand, it provides a reference dataset against which to validate models of this complex-terrain ABL. Assessment of the relative importance on the ABL dynamics of mechanisms like the surface energy balance, the advection of the nearby marine boundary layer, or the passage of coastal lows, can be obtained through numerical or analytical modeling, only after model results are validated against actual observations of MH, like those presented here. Second, combination of MH data with aerosol concentrations (PM10 and PM2.5) routinely measured at several points over the Santiago metropolitan area shall be of interest to the validation of satellite-derived aerosol optical depths (Martin 2008; Gupta et al. 2006).

1 Throughout the paper LT will refer to UTC – 4 h, which is official wintertime time in Chile.
Its top height decreases from about 800 m at 0000 LT to about 400 m at 0900 LT. This lowering of the nocturnal residual layer top is evident in many of the collected data and may be the effect of horizontal advection or nocturnal subsidence in the basin. The latter could be of synoptic or subsynoptic origin (Garreaud et al. 2002), or result as a compensatory sinking motion associated to the near-surface drainage flows leaving the basin through its lower exits (Whiteman et al. 1996).

After 0900 LT the growth of a mixed layer with large backscatter intensities is clearly observed. This mixed layer can be clearly traced in this day from about 100 m at 0900 LT up to about 600 m at 1400 LT. After 1400 LT, the backscatter values decrease and it becomes harder to recognize the mixed layer top. After 1800 LT the aerosol layers decouple again. A near-surface aerosol layer is evident up to 200 m and a new residual layer is seen aloft, with maximum backscatter intensities at about 800 m above the surface.

b. Mixing height retrieval

Like other similar works (Emeis et al. 2008; Eresmaa et al. 2006), we derive mixing layer heights (MHs) from the CL31 backscatter intensity profiles, \( B(z) \), which is obtained directly from the CL31 in units of \((100 \text{ 000 srad km})^{-1}\), and which will be hereinafter referred to simply as backscatter units (bu). For deriving MH from \( B(z) \), we apply the idealized backscatter method described by Munkel et al. (2007), with a slight modification as described next. Backscatter profiles are first averaged over hourly periods provided there are more than 400 profiles available in the hour. Next, we eliminate the lowest five data levels. As Fig. 2a shows, the first backscatter values are most of the time very large, probably because of local clutter returns. To the remaining mean backscatter profile up to 2000 m above the surface, we fit the following analytical profile

\[
B(z) = (a + bz)f + (c + dz)(1 - f),
\]

where \( f \) is a weighting function defined by

\[
f = \frac{1 + \text{erf}[(z - h)/s]}{2},
\]

where \( \text{erf} \) denotes the error function and \( h \) is the parameter associated to the mixed layer depth. With (1) and (2) the backscatter profile tends to the linear form \( B(z) = a + bz \) for \( z \gg h \), while for \( z \ll h \) it tends to \( B(z) = c + dz \). The steepness of the transition between

![Fig. 2. (a) Instantaneous backscatter field measured by ceilometer CL31 for 30 Aug 2007. Temporal resolution is 4 s and vertical resolution is 20 m. Color scale is in backscatter units (see text). (b) As in (a), but for hourly averaged backscatter profiles. Red circles show MH retrieved by the algorithm described in the text. Retrievals at 1000 and 1100 LT were filtered out of the final analysis.](image-url)
these two linear profiles is controlled in (2) by the transition depth parameter $s$. In Munkel et al. (2007), the idealized limiting profiles were constants ($b = d = 0$). However, in our cases the observed backscatter profiles tend to show a positive (negative) slope above (below) the mixed layer top, which is the reason for adding the parameters $b$ and $d$ to the fit, and for further restricting the fit to the conditions

\begin{align*}
  b & \geq 0, \\
  d & \leq 0, \\
  s & \geq 0, \\
  h & \geq 0, \quad \text{and} \\
  (c + dh) - (a + bh) & \geq 0.
\end{align*}

The latter condition forces the backscatter inside the mixed layer to be larger than the backscatter aloft. The parameters in (1) and (2) are $a$, $b$, $c$, $d$, $h$, and $s$, which are found with a least squares algorithm. In particular, the parameter $h$ is considered to be the estimate of the ML height. As an example, Fig. 3a shows the hourly backscatter profile for 1400 LT 30 August 2007. It also shows the fitted profile and the limiting linear profiles that compound the idealized profile. In this particular case, the optimal fitted parameters were $a = 0.3$ bu, $b = 0.0262$ bu m$^{-1}$, $c = 204$ bu, $d = -0.072$ bu m$^{-1}$, $h = 490$ m, $s = 81$ m.

\section*{c. Filtering and filling of the data}

The algorithm just described was applied to all available hourly backscatter profiles for the period between 1000 and 1500 LT of each day. The hourly averaged
backscatter field for 30 August 2007 and the corresponding estimated MH are shown in Fig. 2b. It is apparent that the algorithm works well for well-developed mixing layers that show a large contrast in backscatter with respect to the layer above. In the morning, however, the algorithm is many times confused by the presence of aerosol layers related to the residual layer of the previous night. This happened in this particular day at hours 10 and 11. Figure 3b shows the evolution of the backscatter profile for hours 1000–1500 LT. During the first hours of the morning the secondary maximum of backscatter near 400 m above the surface hinders the correct determination of the aerosol layer near the surface.

Another condition that prevents determination of a mixed layer height by this method is the presence of clouds in the form of fog, low stratus, or the passage of synoptic disturbances producing precipitation. During these conditions, the backscatter profiles near the surface are far from the idealized shapes implicit in (1), and the fitted parameters are not related to a mixed layer height, even if the existence of a mixed layer can be presumed (e.g., in cases of inland penetration of the stratus-capped coastal boundary layer).

A third condition in which fitted mixing heights are not reliable, are clear days in which the aerosol load in the mixed layer is not sufficient to produce a good contrast in the backscatter profile. This happens especially in summer, or in the final hours of the afternoon, when vertical dilution of the aerosols makes it not possible to derive a mixing height from the backscatter profile. This problem is illustrated by Fig. 3c, which shows hourly averaged backscatter profiles for 14 January 2008. While at 1000, 1100, and 1200 LT it is possible to distinguish a mixed layer with larger backscatter intensities than above, from 1300 LT onward the profiles become extremely flat and no level of sharp change in backscatter is detected.

Because of the reasons stated above, it was necessary to filter the mixing heights retrieved by the algorithm, discarding those that were not clearly related to a mixing layer connected to the surface. This was done by visually inspecting all days with data and marking all retrievals affected by any of the problems described above; that is, confusion by a residual aerosol layer, existence of clouds or low aerosol load. In the cases in which only certain hourly retrievals were eliminated for some reason, the mixing height of contiguous hours were linearly interpolated to fill the gaps (less than 3% of MHs in the final database originate in this interpolation).

d. Validation of MH retrievals

Munkel et al. (2007) and Eresmaa et al. (2006) have validated MH retrievals from lidar ceilometers by comparing them with MHs derived from soundings, obtaining good agreement in dry and clear sky conditions. Lack of upper air measurements in Santiago prevents us of replicating such validation. On a few days of June 2009, however, the Chilean Weather Service (DMC) launched radiosondes in support of the monitoring of air pollution conditions in Santiago. The launching site is about 2.5 km northwest of the CL31 location. Although the number and times of these radiosondes is not enough to perform a full validation of the ceilometer-derived MHs, they still provide opportunities to compare ceilometer backscatter fields with the vertical thermodynamic structure of the Santiago ABL. Figure 4 shows CL31 hourly averaged backscatter fields and radiosonde profiles for 10 and 11 June 2009. These were clear days during an air pollution episode, in which DMC launched radiosondes at 1500 LT. Potential temperature and water vapor mixing ratio profiles suggest that a well-mixed convective boundary layer extends at that time up to about 350 m above the surface on both days. These estimates match well the MHs retrieved for these times using the CL31 backscatter profiles. Winds below 1200 m are in both cases very weak supporting the fact that convective turbulence is generally the main process inducing the growth of the daytime MH in Santiago.

e. Meteorological data

In section 3 we explore the relationship between ceilometer-derived MHs and indices representing the stability and the surface energy budget of the basin. The latter are calculated from data available at several automated meteorological stations. An index of the energy available at the surface to produce the heating and growth of the MH will be computed based on solar irradiances measured by a pyranometer collocated with the ceilometer. Near-surface basin temperature will be described by \( T_s(t) \), the average of hourly temperatures measured at 10 m above the surface at the four stations S1–S4 shown in Fig. 1a, which belong to the air pollution network of Santiago. Temperature and wind measured at surface station S5 in Fig. 1a will be used to construct an index of low-level temperature advection into the basin. Station S5 is located at the main entrance of the Santiago Basin, in a zone where the basin connects to the coastal marine boundary layer existing to the west.

Unfortunately, there are no continuous upper air measurements in Santiago with which to describe the stability of the basin's air mass. Hence, we shall use temperature data of station Lo Prado (point LOP in Fig. 1a) to characterize the temperature near the top of the basin's air mass. Lo Prado is located atop a 40-m communication tower over a saddle point (~1000 m above sea level) of the Coastal Range limiting the basin to the west.
Rutllant and Garreaud (1995) showed a good correlation between Lo Prado temperature and free tropospheric temperature at 900 hPa.

3. Results

a. Data representativity

In the next two subsections we shall describe the seasonal and hourly variation of the mixing height over Santiago as derived from the ceilometer observations. Because of the filtering of the data explained above, however, it is necessary to describe here first the representativity of the final database used in the analysis.

Figures 5a,b describe the representativity of the hourly mixing height distributions for the cold season period (Fig. 5a), and for the warm season period (Fig. 5b). The percentages shown in these figures correspond to the fraction of the days with data in the period for four categories of MH retrieval: successful, affected by cloud presence, affected by low aerosol loads, and affected by residual or upper layer of aerosols. In the winter period (Fig. 5a), maximum availability of data is reached in the afternoon, when mixing heights were retrieved successfully for about 60% of the days with data. Cloudiness and frontal system passages precluded appropriate retrieval of mixing heights the rest of the time. The figure also shows that during mornings in the cold season, more mixing heights were filtered out because of fog and low clouds and confusion of the algorithm by residual aerosol layers. Figure 5b describes the representativity of the hourly MH in the warm season from November to February. In this case, mixing height retrieval is more successful in the morning hours, covering up to 80% of the days with data. Availability of data decreases in the afternoon because of low aerosol load.
The previous analysis shows that retrievals at midday are more uniformly successful along the year. Accordingly, values at 1200 LT will be used in section 3b to describe the seasonal variability of the MH over Santiago. Figure 5c describes the monthly representativeness of MH at 1200 LT. On average, about 20% of the retrievals at 1200 LT are affected either by low aerosol loads (especially in summer) or by residual aerosol layers (especially in winter). Clouds hinder 1200 LT MH retrievals in about 10% of days in summer and up to 40% of days in winter.

b. Seasonal variation of MH

Figure 6a shows the time series of MH at 1200 LT for the full period considered and Fig. 6b shows the boxplots for the monthly distributions of the values. A large amplitude of the annual cycle of MH in Santiago is evident in these figures. Median values of MH present a fourfold change between the cold months of June and July (MH ~200 m) and the warm months of December–January with median values around 800 m. This variation is most probably due to the radiative forcing of the surface energy balance and the subsequent energy available to promote the convective growth of the daytime mixed layer. High-frequency variability in MH is also evident in Fig. 6. Interquartile ranges of the monthly distributions of MH are about 200 m (Fig. 6b), and do not change much along the year, so that the coefficient of variation of MHs peaks in the cold season. This points to the importance of synoptic factors affecting the growth of the ML. In terms of Santiago’s air pollution problem, this synoptic control is especially relevant in wintertime, when—together with the reduced energy availability—it can produce extremely shallow daytime mixed layers.

c. Hourly variation of MH

The hourly change of MH in Santiago is illustrated by Fig. 7, in which boxplot distributions of MH are shown for 1000–1500 LT for cold (Fig. 7a) and warm (Fig. 7b)

![Figure 5](https://example.com/figure5.png)

**Fig. 5.** Representativeness portrayal of ceilometer MH database. (a) Fraction of days with data in the months between May and August for different MH retrieval categories for hours 1000–1500 LT. Shading of bars indicate retrieval category. From darker to lighter shading they denote successful retrieval, cloud interference, low aerosol load, and interference by upper or residual aerosol layers. (b) As in (a), but from November to February. (c) As in (a), but for monthly analysis of 1200 LT MH retrievals.
periods of the year. For all hours considered MHs are generally larger in summer than in winter, which can be explained by the larger energy input rate available in summer, and also because at this latitude sunrise occurs about 2 h earlier in summer (0600 LT) than in winter (0800 LT).

As expected, the MH distributions show a steady increase in their median values along the hours. We fitted linear growing rates to each day with data and obtained the distributions of growing rates for winter and summer. The lower, middle, and upper quartiles of these distributions are 30, 50, and 80 in winter and 70, 100, and 150 in summer (all values expressed in m h$^{-1}$).

The summer hourly distributions in Fig. 7b show a leveling off of the median values after 1200 LT, but this behavior must be considered with caution. As explained in section 3a, the representativity of the afternoon MH values in the summer period decreases because of the lower load of aerosols, and therefore, the successful retrievals in these hours are probably biased in favor of lower MH values that may have had higher concentrations of aerosols. Nonetheless, the leveling off of MHs in the afternoon could also be the effect in part of low-level cold air advection associated to a regional coast-to-basin circulation similar to that described for the Mexico Basin by Whiteman et al. (2000). The energy analysis in the next section will show some support for this possibility.

d. Energy and stability controls of ML growth

The growth of the daytime ML over land occurs as the turbulent boundary layer gradually entrains the stable air layer above it (Garratt 1992). The ML growth rate, therefore, depends largely on the intensity of the ABL turbulence, and the lower tropospheric stability and subsidence. Daytime ABL turbulence over land is mainly controlled by heating of the surface by solar radiation and the ensuing of convection, although mechanical turbulence in windy conditions may also be important. We explore in this section the control that the surface energy balance and the stability of the basin air mass have in the MHs derived from the lidar ceilometer.
As the most simple framework in which we can analyze our observations of ML over the Santiago Basin, we consider the encroachment or thermodynamic model of convective boundary layer growth (Stull 1988, p. 454). In this case the growth of a CBL is estimated by a simple energy balance in which the only heat source of the CBL is the sensible heat flux at the surface. The integral of this flux from sunrise to a time \( t \), together with the initial potential temperature profile of the column, uniquely determine the CBL height at time \( t \). Although simple, this model is limited because it ignores several factors, such as the entrainment heat flux at the top of the growing CBL, subsidence, and advective effects. However, lacking detailed upper air data and extended surface meteorological information, we consider it an appropriate first-order conceptual model to apply to our problem.

Figure 8 illustrates the encroachment growth of a CBL starting from a stable potential temperature profile, \( \theta_{SR}(z) \), at sunrise (SR). As the surface sensible heat flux warms the boundary layer, and turbulent convection mixes it, the potential temperature profile turns to be approximately constant in the vertical at time \( t \), up to the level it intersects the initial stable profile. The area \( A \) in Fig. 8 represents the heat gained by the warming CBL, and in this model it is proportional to the integral of the surface sensible heat flux from sunrise to time \( t \). Next, using our available data, we define energy and stability indices related to this model.

An index of the energy available at the surface to produce the heating and growth of the MH is computed based on the pyranometer collocated with the ceilometer. Because of uncertainty in the absolute calibration of this pyranometer, however, we use its data only to compute the fraction, \( f(t) \), of solar radiation received at the site integrated from sunrise up to time \( t \), compared to a completely clear day. Since clear days are rather frequent, the pyranometer’s clear-day solar irradiances can be easily determined. We also compute, \( R(t) \), the incoming solar radiation at the top of the atmosphere integrated from sunrise up to time \( t \). This is calculated for this location and for each day based on standard astronomical formulae (Stull 1988, p. 257). Finally, the index for energy availability at the surface for each day at time \( t \) is simply computed as

\[
E(t) = f(t)R(t). \tag{8}
\]

Using top-of-the-atmosphere incoming solar radiation as the reference for computing \( E \) has the additional
advantage that our results can be compared to those for intermountain western U.S. valleys as reported by Wolyn and McKee (1989).

An index of the thermodynamic energy used in warming the ML from sunrise to time \( t \) is computed based upon the ceilometer-derived MH at time \( t \), \( H(t) \), and the near-surface thermal amplitude, \( \Delta \theta_s(t) \), in the following manner:

\[
A(t) = \frac{1}{2} \rho c_p H(t) \Delta \theta_s(t),
\]

where \( \rho \) is air density, \( c_p \) is the specific heat of air at constant pressure, and \( \Delta \theta_s(t) = \theta_s(t) - \theta_s(SR) \). Near-surface temperatures and potential temperatures will be considered equal (i.e., \( T_s = \theta_s \)).

The upper-air morning potential temperature \( \theta_u \) is estimated as

\[
\theta_u = \langle T_u \rangle_{07-13} + \frac{g}{c_p} H_u,
\]

where \( \langle T_u \rangle_{07-13} \) is the average of Lo Prado temperatures between 0700 and 1300 LT of each day, \( H_u \sim 400 \text{ m} \) is the level above the basin’s floor at which Lo Prado temperatures will be considered to be representative of, and \( g \) is the acceleration of gravity. Accordingly, an initial bulk stability index \( \Gamma_0 \) will be computed as

\[
\Gamma_0 = \frac{1}{H_u} (\theta_u - \langle \theta_s \rangle_{05-07}),
\]

where \( \langle \theta_s \rangle_{05-07} \) denotes the near-surface basin’s potential temperature at the time of sunrise, computed as the average of \( T_s(t) \) for hours 0500–0700 LT. Thus, \( \Gamma_0 \) is intended to represent the bulk early morning vertical gradient of potential temperature in the basin’s air mass. Another stability index, \( \Gamma_H \), can be computed based on the values of \( H(t) \) and \( \Delta \theta_s(t) \) as

\[
\Gamma_H(t) = \frac{\Delta \theta_s(t)}{H(t)}.
\]

2) RESULTS

In the case that MHs in Santiago are mainly controlled by convection, the scales \( E(t) \) and \( A(t) \) defined in (8) and (9) should be closely related. This is tested in Figs. 9a,b, for 1200 and 1400 LT, respectively. It is apparent that there is a close relationship between \( E \) and \( A \), especially earlier in the morning. At 1200 LT points scatter around the line \( A \sim 0.30E \), which would mean that typically about 30% of the top-of-the-atmosphere incoming solar radiation is used in warming the ABL. This fraction matches well what Wolyn and McKee (1989) report for intermountain western U.S. valleys. Later in the day, especially in the summer period, the \( A/E \) fraction decreases and there are days with much lower values. We hypothesize that advection of the cooler marine boundary layer existing to the west of Santiago is partly responsible for this decrease of the \( A/E \) fraction. As a preliminary test of this hypothesis, Fig. 9c shows a scatterplot of a morning temperature advection index and the \( A/E \) fraction at 1400 LT. The morning temperature advection into the basin has been estimated using wind and temperature data at station S5 in Fig. 1a. The index was simply computed by

\[
TA = -U_{SS} \frac{T_s - T_{SS}}{L},
\]

where \( U_{SS} \) and \( T_{SS} \) are the zonal wind and air temperature at station S5, respectively, and \( L \sim 4000 \text{ m} \) is a scale of the horizontal distance between station S5 and stations S1–S4. The TA index in Fig. 9c corresponds to the average of TA for hours 0600–1400 LT for each day. The scatter diagram in Fig. 9c indeed suggests that in days when cool advection at the entrance of the basin is more intense, the \( A/E \) fraction is smaller; that is, the net warming of the CBL represents a smaller fraction of the energy available at the surface.

In the case that the early morning potential temperature profile in the basin’s air mass is approximately linear, we expect \( \Gamma_H(t) \sim \Gamma_0 \), as defined by (11) and (12), respectively. This is tested in Fig. 9d. In general, \( \Gamma_H \) (1200 LT) and \( \Gamma_0 \) appear linearly related, except for cases with larger values of \( \Gamma_0 \) in which \( \Gamma_H \) can be up to 3 times \( \Gamma_0 \). These cases occur in wintertime days with a large thermal amplitude near the surface and very shallow ML according to the ceilometer (these are typically highly polluted days as well). Although on these days \( \Gamma_0 \) is also relatively large, the fact that \( \Gamma_H \gg \Gamma_0 \) suggests that during these days the early morning potential temperature profile in the basin’s air mass is not linear, but rather concave in the vertical, with stability and cooling concentrated near the surface. This curvature is consistent with nocturnal cooling dominated in the basin more by radiation than by turbulence (Garratt 1992).

4. Conclusions

This article has documented the variability of clear-day daytime MHs over Santiago, Chile, as derived from two-year observations by a lidar ceilometer. At 1200 LT, the algorithm used herein is unsuccessful in retrieving MHs in about 20% of the days, because of confusion by residual aerosol layers (especially in winter) or apparent
low aerosol load in the column (especially in summer). Earlier in the morning and later in the afternoon, successful retrievals are less frequent. Median values of 1200 LT MH exhibit a large (fourfold) seasonal change closely related to the change of solar radiation reaching the surface along the year. Synoptic variability is also important as described by interquartile ranges of about 200 m in the monthly distributions of MH. In this ABL, located in a region of generally weak pressure gradients, this synoptic variability is probably forced mainly by changes in the lower tropospheric stability by mechanisms like those proposed by Garreaud et al. (2002). Daytime growth rates of the MH successfully retrieved show typical values of 50 m h$^{-1}$ in winter and 100 m h$^{-1}$ in summer, with noticeable variability as well.

Values of MH and near-surface temperatures allowed the computation of energy and stability indices for the convective boundary layer growth. These indices were compared to the accumulated solar radiation flux measured at the surface, and to a bulk index of the early morning stability of the basin’s air mass. The conclusions of these comparisons are: i) warming and growth of the ML is highly related to the surface energy balance, especially in the morning hours, when about 30% of the top-of-the-atmosphere incoming solar radiation heats and mixes the ABL; ii) later in the day and especially in summer this
fraction decreases and is more variable, which could be due to advection of the cooler marine boundary layer existing to the west of the Santiago Basin; iii) in cases of large bulk stability of the basin’s air mass, there are events of very shallow MLs that develop a large surface thermal amplitude during the day, suggesting that in these cases the nocturnal stability is highly concentrated near the surface.

Results herein are a necessary first step toward a better understanding of the dynamics of this ABL. For several reasons, this ABL can be considered rather idealized for the study of convective boundary layer (CBL) development in complex terrain. The influence of the southeast Pacific subtropical anticyclone provides a very stable lower troposphere environment upon which this CBL grows. The main effect of synoptic variability at this latitude is to enhance or to weaken this stability, with only occasional passage of fully developed frontal systems that completely disrupt the local basin circulations. The topography surrounding the basin is steep, providing a confinement that strongly reduces low-level winds and advection. Finally, vegetation cover of this semiarid basin is very limited, reducing the uncertainty of surface latent heat fluxes to a minimum. Thus, observational analysis presented here constitutes an important benchmark for the validation and improvement of mesoscale models applied to simulate this and other similar complex-terrain ABLS. In particular, work is underway to test the ability of a mesoscale model to reproduce the very shallow MLs that the CL31 detects during specific wintertime events (Undurraga 2009). Other expected applications of the MH database documented here are in the management of the severe air pollution problem of Santiago City and in the validation of satellite retrievals of air quality parameters.

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