Characterization of the atmospheric boundary layer in a narrow tropical valley using remote-sensing and radiosonde observations and the WRF model: the Aburrá Valley case-study

Laura Herrera-Mejía1,2 | Carlos D. Hoyos1,2

1Facultad de Minas, Universidad Nacional de Colombia, Sede Medellín, Colombia
2Sistema de Alerta Temprana de Medellín y el Valle de Aburrá, Área Metropolitana del Valle de Aburrá, Colombia

Correspondence
Laura Herrera-Mejía, Facultad de Minas, Departamento de Geociencias y Medio Ambiente, Universidad Nacional de Colombia, Sede Medellín, Colombia. Email: lherreram@unal.edu.co

Abstract
The spatiotemporal evolution of the atmospheric boundary layer (ABL), in a narrow, highly complex terrain located in the Colombian Andes, is studied using radiosondes and remote-sensing equipment. Different techniques are implemented to automatically estimate the ABL height using ceilometer backscattering profiles and a combination of a radar wind profiler and microwave radiometer retrievals. The large aerosol load from anthropogenic emissions within the valley allows the use of ceilometer-based ABL height detection methods, especially under stable atmospheric conditions. However, convective atmospheres favour aerosol dispersion, increasing the uncertainty associated with the estimation of the convective boundary layer using ceilometers. In contrast, the multisensor technique is more robust, performing better in stable and unstable conditions. All ceilometer-based methods and the multisensor scheme capture the observed ABL height diurnal cycle. The main difference among ABL height retrievals occurs in the afternoon and during the night when Richardson number estimates tend to detect the top-down contraction of the residual layer, while ceilometer-based estimates detect the sudden bottom-up onset of the nocturnal stable layer. The results also show that intra-annual and annual variations of cloudiness strongly condition the ABL expansion, leading to a modulation of the ABL height diurnal cycle. The amount of aerosol particles near the surface is influenced by the evolution of the ABL, modifying the available control volume for the pollutants to interact and disperse. The evolution of ABL over the slopes and at the valley floor differs as a result of the local thickening associated with upslope winds. Weather Research and Forecasting model simulations, from a climatological point of view, skilfully simulate the observed ABL height for both the diurnal and annual cycles; the model skill is higher over the valley floor than over the slopes.

KEYWORDS
atmospheric boundary layer; complex terrain; tropical narrow valley; remote sensing; ABL automatic retrieval methods; ABL model simulations; ABL diurnal cycle; ABL annual cycle

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1 INTRODUCTION

A better knowledge of the atmospheric boundary layer (ABL) structure and dynamics is fundamental to understand the mechanisms associated with the behaviour of meteorological variables in urban areas (Baklanov et al., 2009), the evolution of air pollutant dispersion and disposal in the atmosphere (Nieuwstadt and van Dop, 1981; Kastner-Klein and Rotach, 2004), and the development of deep convection on different spatial scales (Mapes, 2000; Nesbitt and Zipser, 2003; Kalthoff et al., 2011; Rochetin et al., 2014). The triggering of local convection strongly depends on the spatiotemporal variability of near-surface processes, determining the diurnal cycle of rain showers and thunderstorms as daytime solar radiation heats the surface, triggering thermal instability (Betts et al., 2002; Nesbitt and Zipser, 2003; Liu and Liang, 2010; May et al., 2012). The convective boundary layer (CBL) plays a fundamental role by acting as an interface for exchanging momentum, water vapour, gases, and pollutants from the surface into the free atmosphere (Betts, 1973). Fair-weather cumuli are, for example, a result of the near-surface exchanges; hence, their life cycle is strongly coupled to the variability of surface fluxes in the ABL (Brown et al., 2002; Betts and Viterbo, 2005; Chandra et al., 2013). The diurnal cycle of lower-troposphere processes, which are governed by the incidence of solar radiation, also controls the state of atmospheric stability and evolution of the ABL, modulating the concentration of pollutants from mobile and fixed sources (Yu et al., 2001; Whiteman et al., 2014). The ABL height determines the control volume for pollutants to disperse (De Wekker and Kossmann, 2015; Lotteraner and Piringer, 2016): when the meteorological conditions are not favourable for the growth of the CBL, anthropic emissions build up and lead to critical air pollution episodes. In this sense, the ABL height implicitly determines the pollutant concentration (Dabberdt et al., 2004; Eresmaa et al., 2012; Wiegner et al., 2014). Consequently, in recent years, urban meteorology has become a priority for city planning and human health studies (Di Sabatino et al., 2013; Shrivastava et al., 2014).

Despite the importance of the ABL dynamics, there is not a single, reliable and widely accepted technique to retrieve mixing heights (Seibert et al., 2000). The lack of a robust tool is particularly critical for complex terrains where the challenge is not only restricted to the development of an ABL height-retrieval algorithm, but also extends to the definition of the ABL itself. De Wekker and Kossmann (2015) and Lehner and Rotach (2018) highlight current challenges in our understanding of the ABL over mountainous terrain, noting that it exhibits a multilayered structure and that the existing detection methods might be inefficient or represent different components of an intertwined dynamic. Numerous definitions and methods result in a wide range of estimates from in situ and remotely sensed measurements (Eresmaa et al., 2012; Lotteraner and Piringer, 2016). Indeed, evidence from previous studies suggests that using more than one method and different datasets is highly recommended to overcome the deficiencies of individual techniques (Emeis et al., 2008; Quan et al., 2013; Uzan et al., 2016; Wang et al., 2016; Su et al., 2017) and, in complex terrains, to capture the footprint of different processes, resulting in a multilayered structure. Among the most widely used methods are the gradient method using radiosondes (Seidel et al., 2010; Lee et al., 2014), the minimum gradient method, maximum variance using ceilometer profiles (Hayden et al., 1997; Stachlewska et al., 2012), and the bulk Richardson number method (Stull, 1988; Chandra et al., 2014; Zhang et al., 2014a).

Radiosondes are still widely regarded as the most accurate method for mixing-layer height retrievals, especially for enabling direct measurements of temperature, humidity, and pressure (Lokoshchenko, 2002; Hennemuth and Lammert, 2006; Seidel et al., 2010; Emeis et al., 2012; Sawyer and Li, 2013; Lee et al., 2014; Zhang et al., 2014b). However, their limited temporal resolution does not allow the detailed evaluation of ABL evolution throughout the entire diurnal cycle. A significant improvement in remote-sensing techniques over the past sixty years, in particular over the past two decades, has allowed researchers to use vertical profiles from active remote-sensing instruments, such as ceilometers, lidars, and radar wind profilers (RWPs), as well as passive sensors such as microwave radiometers (MWRs), for long-term ABL height retrievals (Hennemuth and Lammert, 2006; Sundström et al., 2009; Bianco et al., 2011; Stachlewska et al., 2012; Hannesdóttir, 2013; Nisperuza, 2015; Lotteraner and Piringer, 2016; Uzan et al., 2016), facilitating the study of the ABL temporal evolution.

The Aburrá Valley, a low-latitude highly complex mountainous terrain located between the Colombian west and central mountain ranges, is home to approximately four million people living in an area of 1,152 km² and has endured the onset of critical air quality episodes in recent years (Figure 1a). The evidence suggests a substantial modulation of the local air pollutant concentration associated with ABL variability. In March 2016, a critical air quality episode highlighted the relationship between local meteorological conditions, ABL variability, and pollutant concentrations (Figure 1b). Pérez Arango (2008), Correa et al. (2009), and Pérez Arango et al. (2011) used wind and virtual potential temperature vertical profiles from a short field experiment (three days in 2006 and one day in 2017) using pilot and tethered balloons to study boundary-layer winds and their interaction with the free atmosphere. Their results suggest a change in the atmospheric stability, from stable to neutral conditions, at approximately 1000 local time (LT= UTC–5 hr). Nisperuza (2015), using a 532 nm lidar, presented evidence of a marked mid-morning transition between the stable and convective...
layers. The results of these studies represent the first observational steps towards a better understanding of ABL evolution within the Aburrá Valley. Despite the advances, there is a need for a long-term continuous ABL height dataset to identify and study the meteorological and climatological conditions that modulate the ABL dynamics and lead to critical air quality episodes.

The central aim of this study is to examine the ABL height variability over the Aburrá Valley, on different time-scales, associating it with the evolving regional climate and the meteorological conditions within the valley that are important in modulating the ABL behaviour on the local scale using observational evidence and limited-area numerical simulations using the Weather Research and Forecasting (WRF) model. In this work, ABL height refers to the distance from the surface (ground level) to the top of the boundary layer. There is a growing need for skilful ABL numerical simulations and forecasts associated with evaluating air quality projections and urban planning scenarios (Baklanov et al., 2009; Shrivastava et al., 2014; Wagner et al., 2015). Moreover, correct simulation of the ABL using climate and weather models is vital for pollutant dispersion modeling, and it also conditions the overall skill of numerical weather prediction models (van der Kamp and McKendry, 2010; Di Giuseppe et al., 2012; Zhang et al., 2014a). This work includes the development and implementation of both conventional and new methods to retrieve ABL heights based on information from radiosonde campaigns, ceilometers, a RWP, and a MWR. The remote sensors operate continuously as part of the Sistema de Alerta Temprana de Medellín y el Valle de Aburrá (SIATA; https://siata.gov.co; accessed 18 June 2019). In Section 2, we present a description of the data used for the ABL assessment and the different techniques implemented to estimate ABL heights from single-sensor measurements and by combining the retrievals from multiple instruments. Section 2 also includes a description of the numerical model configuration and the experimental set-up. Section 3 presents the results of the ABL height spatiotemporal analysis, identifying the climate and meteorological conditions favouring the pollutant dispersion in the Aburrá Valley. An evaluation of the ABL height simulation and forecast skill for the Aburrá Valley is also included in Section 3. Finally, Section 4 presents the most important conclusions of the study.

2 | DATA AND METHODOLOGY

Observational and theoretical evidence suggests there are significant gradients of different meteorological variables in the lower troposphere due to the transition between the ABL and the free atmosphere, allowing the use of vertical profiles of atmospheric variables to infer the ABL height (Chandra et al., 2014). The vertical structure of the aerosol concentration, virtual potential temperature ($\theta_v$), wind speed and relative humidity are some of the most commonly used profiles for indirect ABL height estimates. The ABL retrieval methods proposed in this work use data from various ceilometers, a MWR, and a boundary-layer RWP. ABL height retrievals are compared to estimates using radiosonde data during intensive observation periods (IOPs).

2.1 | Geographical context

Figure 2 shows the geographical context of the Aburrá Valley and the location of the remote-sensing equipment used for ABL estimates. The valley is 64 km long, and it is located in the Central Andes mountain range between 6°N and 6.5°N and 75.3°W and 75.6°W. The widest section of the valley, from divide to divide, is approximately 18.2 km wide, and the width of the narrowest section is approximately 3 km. Most
FIGURE 2  Geographical context of the Aburrá Valley, located in the Department of Antioquia, Colombia. The map shows (white to blue) the main topographic features of the region and the location of the remote sensing equipment used in this study: C = Ceilometer, MWR = Microwave Radiometer, RS = Radiosonde, RWP = Radar Wind Profiler. Most sensors are located at the base of the valley, except for the southernmost ceilometer (Itaguí), which is installed at a considerably higher altitude than the valley floor on a western hill. The MWR is located at SIATA’s main operations centre (Torre SIATA) sensors are installed on the base of the valley except for the southernmost ceilometer, installed on the western hill, allowing the evaluation of the dependence of ABL structure on the topography inside the valley.

2.2  |  Data

2.2.1  |  Ceilometer data

Ceilometers provide information about the laser-pulse energy that is backscattered by clouds and other atmospheric components, including particulate matter, which is expressed as the backscattering attenuated coefficient, $\beta$ (Emeis et al., 2009; Münkel and Roininen, 2010; Kambezidis et al., 2012). Measured profiles from three Vaisala CL51 ceilometers, installed at different locations inside the valley (Figure 2), are used to estimate ABL heights. Vaisala CL51 ceilometers work at the 910 nm wavelength and emit a laser signal every 67 ns, providing backscattering attenuated coefficient measurements with a vertical resolution of 10 m and temporal resolution of 16 s. Ceilometer profiles are available continuously (with some missing data periods) since October 2014 and November 2015, at the Torre Siata and Itagui sites, respectively.

2.2.2  |  MWR data

MWRs measure the radiation emitted by atmospheric gases at submillimetre-to-centimetre wavelengths and are useful for retrieving the thermodynamic state of the atmosphere at different levels, allowing the assessment of atmospheric stability in real time. The Aburrá Valley MWR (Figure 2) is a Radiometrics MP-3000A profiler and provides continuous retrievals of temperature, relative humidity and liquid water up to a height of 10 km above the surface under nearly all weather conditions. The MWR is located at the top of the SIATA main operations centre on the valley floor, approximately 60 m from the surface, and it provides vertical profiles with variable spatial resolution: 50 m from the surface to 500, 100 m up to 2 km, and 250 m up to 10 km. Thermodynamic profiles are available with a 2 min time resolution, and the data record spans from January 2013 to date.

2.2.3  |  RWP data

Doppler radars used for vertical wind profiling rely on refractive index variations caused by changes in humidity, temperature, and pressure. The Aburrá Valley wind profiler (Figure 2), a RAPTOR VAD-BL by DeTect Inc., works at a nominal frequency of 915 MHz, reaching up to 8 km above the surface under high humidity conditions (Lau et al., 2013). The RWP is designed to measure the wind profile in different modes that differ in their vertical resolution and in the atmospheric domain. In our case, we use two overlapping modes: in the higher-resolution mode (60 m), the RWP measures the wind profile from 77 to 3,500 m, and in the lower-resolution mode (72 m) from 2,500 to 8,000 m. In this study, we use only the higher-resolution mode since the ABL was never higher than 3,500 m. The temporal resolution is 5 min. The radar was installed in January 2015.

2.2.4  |  Radiosonde data

Three 5-day radiosonde IOPs were conducted from 28 January to 2 February (IOP 1), from 24 to 28 March (IOP 2) and from 4 to 8 May (IOP 3) in 2015. During each IOP, there were eight radiosonde flights per day launched at the base of the valley from the Torre SIATA site, resulting in a total of 120 atmospheric profiles. The design of the IOPs responds to the aim of sensing the atmosphere during the three main seasons in the Andean Region of Colombia (dry, dry-to-wet transition and wet seasons). However, anomalous meteorological conditions resulted in atmospheric behaviour different from the typical climatological patterns expected over the region. During the three IOPs, the surface air temperature, surface wind speed, and surface relative humidity were remarkably homogeneous, and they were characterized by persistent cloudiness, particularly during early morning
hours, and isolated precipitation events. The average surface air temperature for the entire 5-day period was 21.8, 21.7 and 23.2°C, respectively. The average surface wind speed was 3.1, 2.9, and 3.2 m/s for IOPs 1, 2 and 3, respectively, and the average relative humidity was 61.4, 61.8 and 55.5%. The most salient difference among the three IOPs is the cumulative rainfall: 2.2, 16.4, and 6.8 mm, respectively. However, these differences do not condition the behaviour of the atmospheric stability during the IOPs. Figure 3 presents the measured vertical profiles of the lower-troposphere air temperature, virtual potential temperature, and wind speed for the three IOPs for the first 3 km showing a marked diurnal cycle. The amplitude of the diurnal cycle is particularly high up to approximately 2–2.5 km, where the direct influence of the surface seems to decay considerably. Despite the apparent regularity, the maximum height of the thermal expansion varies from day to day, unequivocally leading to ABL variability. Regarding the wind speed, it can be seen that the magnitude is considerably higher above the valley, as expected, with values inside the valley typically less than 3 m/s and approximately 8 m/s at 2 km.

2.2.5 | In situ, satellite, and reanalysis data

In addition to the atmospheric profiles obtained using ceilometers, MWR, RWP, and radiosondes, we use in situ records of fine particulate matter (PM2.5) concentration from eight BAM1020 (MetOne equipment) monitoring stations installed in the valley, 5 min precipitation records from two tipping-bucket gauges installed at Torre SIATA, 1 min radiation from a Kipp & Zonen pyranometer also at Torre SIATA, and 1 min wind speed and direction from two Vaisala all-in-one WXT520 weather stations installed at the Torre SIATA and Itagüí sites. In addition to the in situ data, we also use the 3-hourly Tropical Rainfall Measuring Mission (TRMM) 3B42 v7 product, which is in good agreement with in situ stations globally and regionally (Kummerow et al., 1998; Huffman et al., 2007; NASA, 2011; Ceccherini et al., 2015), and low-, medium-, and high-level cloud data from the ERA-Interim global reanalysis project (Dee et al., 2011). The TRMM-3B42 v7 product has a 0.25° by 0.25° spatial resolution, and a coverage spanning the entire zonal band from 50°S to 50°N. The spatial resolution of the ERA-Interim dataset is approximately 80 km.

2.3 | Determination of ABL heights using ceilometers

ABL height detection methods using ceilometer backscatter intensity (BI) profiles are based on the assumption that a significant aerosol concentration reduction takes place at the top of the ABL, in the transition between the well-mixed
near-surface troposphere and the “free” troposphere. During fair-weather days, BI profiles often provide an excellent depiction of the ABL evolution within the valley and throughout the day. Figure 4a shows an example of the ceilometer BI for 9 May 2015. A 300–400 m high stable layer is clearly observed between 0300 and 0900 LT. A secondary layer transition, with a smaller BI amplitude, is positioned between 1,000 and 1,500 m between 0000 and 0400 LT. The latter corresponds to the residual layer. At approximately 0930 LT, the stable layer tends to disappear due to the transition to a CBL. The onset of the CBL generates a reduction in BI, and consequently, the ABL is significantly more difficult to detect using aerosols as a proxy. Initiation of the CBL leads to the formation of fair-weather cumuli with a height of 2–3.5 km between 1000 and 1800 LT. This behaviour is observed until 1600–1800 LT when the mixing processes cease, and the onset of the stable layer starts to occur.

The ceilometer-based methods allow the identification of internal layers within the ABL caused by weak mixing and stratification (Emeis et al., 2008; Young, 2013; Young and Whiteman, 2015). Most profiles are complex and show multiple breakpoints due to aerosol accumulation on weakly mixed layers, the presence of clouds, high humidity conditions, and the inherent multilayer nature of ABL in complex terrains (Martucci et al., 2010; Lehner and Rotach, 2018; Serafin et al., 2018). The ceilometer-based ABL height detection methods used herein correspond to the minimum gradient method, maximum variance scheme, and a continuous wavelet transform-based technique. Considering the ABL definition by Stull (1988), we use 30-min average BI profiles for ABL detection to maximize the signal-to-noise ratio, concomitantly providing sufficient information about the ABL structure and temporal variability.

2.3.1 Minimum gradient and maximum variance methods

One of the most common tools for determining the ABL height using BI profiles is the vertical gradient method proposed by Hayden et al. (1997), which has been used by several authors (Steyn et al., 1999; Münkel et al., 2004; Emeis et al., 2012; Stachlewska et al., 2012). The minimum of the vertical gradient profile is considered to be the ABL height. Figure 4b shows the average ceilometer BI for 9 May 2015 between 0930 and 1000 LT, and as an example, Figure 4c shows the results for the ABL height detection using the minimum gradient method. On the other hand, the maximum variance method locates the sharpest changes in the BI profiles due to the layer transition by finding the maximum variance in the vertical profile. The variance profile is calculated using a 200 m sliding window. Figure 4d shows the detection using the maximum variance method.
One of the main advantages of ceilometers is the ability to operate continuously under any atmospheric conditions. However, in the presence of rain, cloud droplets, and hail, light is scattered more efficiently in comparison to aerosol-only scattering (Young, 2013). As a result, the profiles do not represent the typical behaviour in the atmosphere, and the transition between layers becomes almost impossible to detect. Therefore, clouds and precipitation have to be filtered out before determining the ABL height using ceilometers. To overcome this issue, we developed a simple algorithm to detect cloud layers in the profiles and subsequently filter them out using linear interpolation of BI between the cloud-free edges to complete the missing data where the cloud is located.

### 2.3.2 Continuous wavelet transform-based method

The vertical distribution of aerosols in the troposphere is the result of complex nonlinear interactions among several physical processes occurring on a broad set of temporal and spatial scales. This interaction and multilayer structure of the ABL in mountainous terrains (Lehner and Rotach, 2018; Serafin et al., 2018) imposes difficulty in the automatic determination of the ABL height. A spectral analysis tool such as the wavelet transform is ideal for detecting changes in the BI profile associated with the height of the ABL by isolating the appropriate spatial scales and identifying the maximum variance close to the surface. Various authors have successfully used wavelet-based strategies for automatic ABL detection (Cohn and Angevine, 2000; Morille et al., 2007; Baars et al., 2008; Granados-Munoz et al., 2012). However, most studies have used the Haar function as the mother wavelet and a fixed dilation parameter. In this paper, we use the full discretization of the continuous wavelet transform (CWT) technique following the formulation described by Torrence and Compo (1997) using Morlet and Paul wavelets.

The methodology proposed herein for ABL detection using CWT is performed by (a) constructing the wavelength–height wavelet spectrum, (b) integrating the wavelet spectrum on selected spatial scales corresponding to the vertical extent of the physical processes involved in the ABL structure, in this paper on scales ranging from 50 to 500 m, (c) detecting, bottom-to-top, the maxima in the band-passed spectrum, where the ABL height typically corresponds to the first maximum, and (d) nevertheless, checking the relative magnitude of the subsequent spectral peaks to guarantee the presence of a considerable BI reduction. If the reduction is not significant, the algorithm selects the following spectral peak, and step (d) is repeated. Figure 4f shows the ABL detection for the profile in Figure 4b using the CWT-based scheme. The Figure shows the wavelength–height spectrum, the band-passed spectrum, and the selection of the maximum corresponding to the estimated ABL height.

### 2.4 Richardson number method: ABL height determination using RWP and MWR profiles

We propose and implement a novel multisensor scheme for continuous ABL detection combining information from the RWP and the MWR. The profiles obtained from these two sensors allow the estimation of the bulk Richardson number \((Rib)\), a dimensionless parameter that is typically implemented using radiosonde data. The method integrates dynamic and thermodynamic variables to establish a critical threshold \(Rib_c\) where the turbulent flow becomes laminar in the free atmosphere (Stull, 1988; Eresmaa et al., 2006; Granados-Munoz et al., 2012; Zhang et al., 2014a; Schween et al., 2014; Chandra et al., 2014). The \(Rib\) for a specific \(z\) level, is computed as

\[
Rib(z) = \frac{g}{\theta_s} \frac{\theta(z) - \theta_s}{u(z)^2 + v(z)^2} z, \tag{1}
\]

where \(\theta_s\) is the potential temperature for a reference level near the surface, \(\theta(z)\) is the potential temperature, and \(u(z)\) and \(v(z)\) are the zonal and meridional wind components. The RWP and the MWR are located inside a radius of approximately 3 km and have different vertical resolutions. Thus, to estimate \(Rib\) preprocessing of the data is required. Figure 4e shows an example of ABL height detection using the multisensor \(Rib\)-based method for different \(Rib_c\), showing good agreement with the previously described ceilometer-based methods.

The selection of a \(Rib_c\) is not straightforward as it potentially influenced not only by the vertical resolution of the profiles but also by the measurement strategy used for obtaining the wind speed and potential temperature (Seibert et al., 2000). Figure 5a shows an example of the diurnal evolution of the Richardson number profile for 2 February 2015. The profiles show a marked transition from stable conditions in the morning (from 0000 to approximately 1030 LT) to a thermally unstable regime (1030 to 1530 LT), to stable again during the night. Vertical gradients of \(Rib\) under stable conditions are very high, with all the contours closely grouped. The Figure also shows an expansion of the low \(Rib\) area between 1000 and 1800 LT, and a clear separation of the contours during the convective stage. Under stable conditions, the differences in contour height as a function of \(Rib\) are not substantial, being less than 100 m; however, during the convective phase, these differences could exceed 1,000 m in a single day. It is interesting to note that contours corresponding to \(Rib_c = 1, 2, 3\), which are closer to the surface, are more variable than those for \(Rib_c = 4, 5\), underlining two aspects: (a) the irregularity of convective activity near the surface, and...
(a) Evolution of the Richardson number ($\text{Rib}$) profile for 2 February 2015. The purple-to-lime lines denote contours of different $\text{Rib}_c = 1, 2, 3, 4, 5$. (b) diurnal cycle of the mixed-layer height retrieved using the multisensor method with different $\text{Rib}_c$. The diurnal cycle is derived for the period January 2015 to June 2018.

(b) the maximum height of thermal convection, as suggested by the relatively smooth upper contours. Figure 5b shows the diurnal cycle of the mixed-layer height estimated using the multisensor method with different $\text{Rib}_c$ for the period from January 2015 to June 2018. Similar to Figure 5a, while the general behaviour according to different $\text{Rib}_c$ is the same, there are considerable differences in the ABL height. These differences have practical implications since the maximum ABL height for deep valleys might determine whether or not aerosols are able to disperse away from the surface and outside of the valley.

### 2.5 Determination of ABL height using radiosondes

We use two methods to estimate the ABL height using radiosonde data. The first approach uses the Holzworth method, commonly referred to as the parcel method (Holzworth, 1964). This method establishes that the ABL height is determined by the first interception between the virtual potential temperature profile ($\theta_v(z)$) and the dry adiabat ascending from the surface, corresponding to the height where the parcel is in equilibrium with its surroundings (Holzworth, 1964; Lokoshchenko, 2002; Münkel et al., 2006; Seidel et al., 2010). The second approach corresponds to the $\text{Rib}$ method described previously. In this case, potential temperature, wind speed and direction are obtained directly from the radiosondes. Zhang and Li (2019) recently presented a long-term assessment (30 years, 1988–2017) of the climatological characteristics of the ABL over Japan, showing that the bulk Richardson number method performed the best among all the methods tested. Other authors, such as Rampanelli and Zardi (2004), have proposed algorithms based on the potential temperature profiles from soundings to find the CBL using best-fit concepts. We explored the Rampanelli and Zardi (2004) methodology, but the results were not robust during all cases.

### 2.6 WRF ABL forecast assessment

The WRF model, version 3.7.1 (Skamarock et al., 2008), is the basis for the operational numerical weather forecasts issued by SIATA on a daily basis. We used the output of the daily 0000 UTC WRF 24 hr forecasts for three years, focusing on the evaluation of the representativeness of the ABL diurnal cycle and the skill assessment of the 24 hr average ABL height numerical forecasts (0–24 hr forecast horizon). The model configuration includes three nested domains with 18 km ($191 \times 191$), 6 km ($82 \times 118$) and 2 km ($136 \times 136$) grid spacing, and 40 vertical levels up to 50 hPa. The first domain (18 km), $10^\circ$S–20$^\circ$N, $60–90^\circ$W, covers the entire geography of Colombia, the Caribbean Sea, the Colombian sector of the Pacific Ocean, and Amazonia, to include the main external forcing factors of the regional atmospheric circulation and precipitation over the territory. The second domain (6 km) includes the Andean region of Colombia ($1–10^\circ$N, $72–78^\circ$W). The third and last domain (2 km) is centred around the Aburrá Valley and covers $5–7.5^\circ$N, $74.5–76.8^\circ$W. The model runs use the output from the 1200 UTC Global Forecast System (GFS) output as initial and boundary conditions. The integration time step is 90, 60, and 10 s in the 18, 6, and 2 km domains, respectively. SIATA operational forecasts include different ensemble members with different microphysics parametrizations. Table 1 summarizes the schemes and parametrizations selected for the ABL analysis in this study. We focus on the assessment of the ABL height corresponding to the highest grid spacing domain.
3 | RESULTS

This section presents the results of the different schemes used for estimating the ABL height, highlighting the capabilities of the various instruments and the robustness of these estimations under different atmospheric conditions, assessing the temporal and spatial variability of the ABL height, and studying first-order aspects of the role of regional climate in modulating boundary-layer processes.

3.1 | Assessment of the ABL Detection

Figure 6 shows examples of ABL height estimations for 4 March 2015, 13 March 2016, 25 January 2018, and 26 February 2018, presented together with the evolution of BI profiles using the Torre SIATA ceilometer. In general, all ABL height estimations satisfactorily reproduce the diurnal evolution associated with radiation forcing and thermally forced convection. All ceilometer-based ABL heights show a relatively good correspondence among them and, as expected, with the observed BI profiles. Initiation of the atmospheric instability is the main driver of the ABL structure; this process is characterized by the triggering of thermal convection near the surface and is modulated by the increasing temperature gradient between the lower atmosphere and the surface. As the surface temperature increases, air parcels start to ascend, which is usually evident after 0900–1000 LT. Figure 6 shows that the ABL reaches the maximum height at approximately 1300 LT; at this moment, the temperature gradient between the surface and the atmosphere starts to reverse, and a cooling mechanism initiates, forcing a shrinkage of the layer. Once the surface is no longer receiving the energy from solar radiation, the convective layer becomes the residual layer, as the nocturnal stable layer concurrently develops.

One of the most salient features that differentiates the ABL height estimates takes place after 1600 LT when the Richardson number method tends to detect the relatively slow top-down contraction of the residual layer, while ceilometer-based estimates appear to detect the sudden bottom-up onset of the nocturnal stable layer. The CWT-based method is the only ceilometer-based technique that shows a preference for selecting the residual layer as the height of the ABL. Detection of the nocturnal stable layer by the maximum variance and minimum gradient methods is likely caused by the combination of high anthropic aerosol emissions associated with the afternoon commute and the sudden stabilization of the near-surface atmosphere as a result of the solar radiation reduction. In other words, sufficient aerosols are emitted to quickly replenish the control volume confined by the surface and the first atmospheric capping inversion, causing a noticeable discontinuity in the BI. The CWT method, in contrast, tends to isolate the residual layer as a result of the BI reduction. Detection of the residual layer by Richardson number estimates is a direct consequence of the processes contrasted in the adimensional number, indicating that turbulence does not shut down in the entire column immediately with the onset of near-surface stability (the atmosphere is turbulent between the top of the nocturnal boundary layer and the top of the residual layer), but rather decays with the slow weakening of sensible heat flux.

While all dates in Figure 6 correspond to dry days (no precipitation), the dates in the top panels were predominantly cloudy as shown in the ceilometer profiles with a cloud ceiling of approximately 1.2 to 2.5 km. The average specific humidity in the first 2 km in the dates of both top panels, from radiometer retrievals, was approximately 20% higher than usual, ranging between 11 and 12 g/kg. Similarly, the average cloud radiative forcing in the 0600–1800 LT for these two days was approximately −150 W/m² compared to a clear day. Given the reduction in solar radiation reaching the surface, it is likely that sensible heat flux was also anomalously low. Potential temperature profiles from the MWR (not shown) suggest that both days were predominately stable, leading to pollutant accumulation near the surface. The average daily PM2.5 concentration for both days was, in fact, 61 and 81 μg/m³, which is considerably higher than the long-term average (Figure 1). Under these conditions, aerosols become excellent tracers of ABL behaviour. In addition, atmospheric humidity accentuates the size of hygroscopic aerosol particles, enhancing the near-surface ceilometer BI signal (Emeis et al., 2012; Young and Whiteman, 2015).

In contrast, the last two dates in Figure 6 correspond to fair-weather days, with clear skies, and some isolated cumuli over the valley. The cloud radiation forcing is relatively weak (less than 20 W/m²), and the specific humidity is lower than the first two dates (approximately 9 g/kg). Under the mentioned circumstances, the sensible heat exchange tends to be sufficient to break the lower-troposphere stability, establishing a convective regime after 1000 LT. After the atmosphere becomes unstable, and the vertical dispersion takes

### Table 1

| Process          | Scheme                                      |
|------------------|---------------------------------------------|
| Microphysics     | Eta (Ferrier)                               |
| Radiation        | Rapid Radiative Transfer Model (RRTMG)     |
| PBL              | Mellor–Yamada–Janjić                       |
| Land surface     | Unified Noah land-surface model             |
| Surface          | Monin–Obukhov (Janjić Eta)                 |

**TABLE 1** WRF model schemes and parametrizations (Skamarock et al. 2008 give details).
place, aerosol concentrations near the surface and within the convective layer considerably reduce: once aerosols exceed the depth of the valley, trade winds generally advect pollutants away to the east of the Aburrá Valley. The 24 hr average PM2.5 concentration was considerably lower than for the first two dates (31.6 and 38.8 μg/m³, respectively). During dates with similar weather conditions, ceilometer-based techniques tend to represent well the nocturnal boundary layer, but the reduction of tracers results in a less skilful ABL detection. In contrast, the Richardson number methodology is robust in all four cases, allowing a direct ABL height estimation with no dependence on the existence of tracers. However, the main drawback of the Richardson number technique is its dependence on the proper selection of a critical threshold.

### 3.2 ABL height estimates comparison using radiosondes

The first step in the ABL height intercomparison is to evaluate the applicability of the Holzworth and Richardson number methods using radiosonde data as reference values. ABL height estimates are intercompared considering different $R_i b_c$ to find an optimal threshold. The scatterplots in Figure 7a,b,c show the level of agreement between the Holzworth method and three different $R_i b_c$ values (0.25, 1, 3). Pearson correlation and root mean square (RMS) deviations are shown in each of the panels, and the colours of the full circles correspond to the hour (local time) of the radiosonde flights. The distribution of colours in a time-ABL height clustered fashion suggest that both methods, in all three cases, represent well the existence of a marked ABL diurnal cycle during IOPs. The correlations among both the Holzworth and Richardson number estimates are very high, with a maximum of 0.75 for $R_i b_c = 0.25$. The least RMS deviation is also obtained for $R_i b_c = 0.25$. The dispersion level among the Holzworth–Richardson ABL estimates, expressed as the RMS deviation, is significant. However, it is important to consider the fact that on some occasions, both methods represent different sublayers within the ABL, as discussed in the previous section: Richardson ABL height estimates correspond to a top-down approach, while Holzworth estimates correspond to a bottom-up approach. This phenomenon is also evident in the figure since Richardson number ABL height estimates tend to be higher than those using the parcel method, and the dispersion is higher during the convective stage and the contraction of the residual layer. Considering the overall evidence, we select both Holzworth and $R_i b_c = 0.25$ radiosonde-based methods as references for comparing remote-sensing ABL height estimates.

The scatterplots in Figure 7d,e,f show the comparison between the Holzworth estimates and the multisensor Richardson number ABL height estimates for three different $R_i b_c$ values (1, 3, 5). The figures show the same general
features regarding the ABL diurnal cycle. In this case, the optimal $Rib_c$ selected for the multisensor method is different from the one used for radiosonde-based estimates, confirming its dependence on the measurement strategy. Additionally, the highest correlation does not coincide with the lowest RMS deviation. For this reason, and to summarize the intercomparison results, we use a Taylor diagram to contrast all the ceilometer and multisensor estimates with the two radiosonde reference estimates during the IOPs (Figure 8). Taylor diagrams allow the evaluation of the degree of correspondence between two patterns in terms of their correlation, their centred RMS deviations, and the amplitude of their variations represented by the individual standard deviations (Taylor, 2001). The diagram uses normalization of the dimensional quantities considering that we use two different radiosonde references. In other words, ABL height estimates are scaled relative to the standard deviation of the corresponding reference, either Holzworth or $Rib_c = 0.25$, in each case.

The intercomparisons in Figure 8 include, in different groups organized by shape and colour palette, five sets of estimates comparing Holzworth (the reference) with radiosonde Richardson number estimates (blue circles), five sets of estimates comparing Holzworth with multisensor Richardson number estimates (pink-to-violet squares), three sets comparing Holzworth with ceilometer-based estimates (green triangles), five sets comparing radiosonde $Rib_c = 0.25$ with multisensor Richardson number estimates (purple crosses), and three sets comparing radiosonde $Rib_c = 0.25$ with ceilometer-based estimates (orange-to-brown rhombs). In general, twelve out of sixteen radiosonde versus remote-sensing estimates show a correlation greater than 0.5; however, all estimates underrepresent the variability observed in ABL heights from radiosondes. The latter result is in part explained by the finding that the vertical resolution of the radiosonde is higher than that of the ceilometer, MWR, and RWP. Overall, the methodology showing the highest correspondence with the Holzworth technique during IOPs, considering correlation and a balance between RMS deviation and variability representation, is the Richardson number multisensor method with $Rib_c = 4$. If the radiosonde
Ribc = 0.25 is used as a reference, the highest agreement is achieved with the multisensor Ribc = 4 and ceilometer maximum variance method. Considering this result, the analysis in the following subsection will focus on the multisensor Ribc = 4 and ceilometer maximum variance method.

### 3.3 ABL height long-term comparison

Radiosonde ABL height estimates are based on direct measurements of atmospheric profiles and, hence, constitute an important reference for the estimates in this study. However, the radiosonde-based estimates are also prone to artifacts associated with the Lagrangian nature of the platform: radiosonde profiles do not correspond to the same atmospheric column, and in a narrow valley, the differences between the profiles at the base and at the hills could potentially be large, introducing uncertainty in the ABL estimates. In our case (not shown), most of the radiosonde flights below 4,500 m are within the area sensed by the MWR and RWP in the same altitude range, however there is a westward biased towards the hills in the radiosonde which could potentially affect the intercomparison. Figure 9 summarizes a detailed long-term assessment of the ABL height estimates obtained with the different schemes. Since none of the methods could be selected a priori as a reference standard, Figure 9a shows the cross-correlation array in which each pixel corresponds to the zero-lag correlation between all 30-min resolution ABL height time series for all the methods, and Figure 9b presents the amplitude of each time series obtained with the method denoted in each row relative to the method in each column.

The first evident feature is that, in terms of variability, the ABL height estimates obtained with variations of the same base method such as ceilometer-variance (minimum gradient and maximum variance) methods, ceilometer CWT-based methods, and Richardson number methods are very similar among the approaches, with correlations larger than 0.68 despite the high temporal resolution used in the analysis. The CWT-based methods show the lowest correlations with the other methods, and the correlation between the minimum gradient and maximum variance methods with the Richardson number method is larger for lower values of the critical Richardson number. The latter characteristic is due to the sensitivity of the ceilometer-variance methods and the low critical Richardson number (Ribc = 1) estimates to the highly variable near-surface atmospheric changes in time; hence, the variability of the estimates is higher than those obtained using Ribc = 4. This phenomenon is also clear in Figure 9b, where
the overall variability (amplitude of the diurnal cycle and intraday variability) of the ABL height series is largest for estimates using the Richardson number method with $R_{ibc} = 1$ and lowest for CWT-based estimates.

An assessment of ABL height estimates with the different schemes suggests that, in the presence of stable atmospheres, the ABL height estimates using the three backscatter-based methods exhibit high correspondence with each other. If the BI profiles present well-defined vertical structures with noticeable transitions, the retrievals are usually very robust. By the time the convective layer starts to develop due to the onset of atmospheric instability, the concentration of aerosols near the surface and within the valley decreases. In this case, backscatter-based estimates differ from one method to the other, and it is not possible to distinguish which of the methods is more accurate in determining the ABL height. In the latter cases, the Richardson number method becomes a more robust tool for estimating the ABL height.

3.4 ABL height temporal variability

The long-term ABL height identification allows the assessment of the influence of local-scale phenomena versus regional forcing on ABL variability. In this section, we examine the most salient time-scales present in the ABL height time series, identifying the processes that lead to the existence of such variability. Spectral analysis of the long-term ABL height time series allows identification of the main temporal scales of ABL variability. The average ABL height spectrum was estimated using the fast Fourier transform (FFT), and the variability of the spectrum itself was evaluated by computing a sliding FFT with a fixed temporal window, to assess the regularity (stationarity) of the spectral peaks in terms of their amplitude, as a response to regional forcing. Figure 10a shows the average FFT spectrum and the variability of the spectra (+/– one standard deviation) for all block-moving 21.333-day (1,024 30-min data points) ABL height series, estimated using the maximum variance method. We use the mentioned ABL height series considering a length of the ceilometer record longer than that of the RWP. Each 21.333-day spectrum is also computed after tapering the signal using a Hann window. The relative variance is estimated by integrating the FFT spectrum between 20 and 24 hr and dividing the result by the series variance. The most salient ABL time-scale with a period shorter than 21.333 days, in terms of variance, corresponds to the diurnal cycle, caused by the modulation controlled by solar radiation and the onset of convective activity. The other noticeable spectral peaks in Figure 10a correspond to the harmonics of the diurnal cycle (8 and 12 hr peaks).

Figure 10a suggests the presence of substantial temporal variability of the relative amplitude of the diurnal cycle. Considering the latter, Figure 10b presents the histogram of the relative variance of the diurnal cycle, as a percentage, obtained from all the 21.333-day spectra, showing that the diurnal cycle represents, in some cases, just approximately
25% of the total variance, compared with more than 50% in other cases. In other words, the amplitude of the diurnal cycle changes on the annual and semi-annual time-scales. The practical implications of this variability of the amplitude of the diurnal cycle are linked to the knowledge that a low-amplitude ABL diurnal cycle could lead to pollutant accumulation. The spectrum of 100 m BI (not shown), a proxy for the aerosol concentration, also shows a marked diurnal cycle and a 7-day peak (and 14- and 21-day subharmonics), implying important variability due to the weekly pollutant emission cycle within the valley. After filtering out the diurnal cycle, the ABL height spectrum (Figure 10c) shows significant annual and semi-annual peaks that are potentially associated with the cloud radiative forcing resulting from the progression of the Intertropical Convergence Zone (ITCZ) over Colombia. In addition to these spectral peaks, there are other low-variance intraseasonal spectral peaks that are not explored in this work. It is important to note that the length of the ABL height series is approximately 3.5 years, precluding the exploration of interannual time-scales.

Figures 11a,d present the diurnal cycles for the multisensor Richardson number $Ri_b = 4$ and maximum variance ABL height estimates, respectively. The shaded band represents the ±0.5 standard deviation. Both time series show the remarkable diurnal cycle, with similar variability around the mean values. In the long term, the most evident difference appears to be the same as that described in Section 3.1 regarding the multisensor method preference for detecting the residual layer and BI methods for selecting the nocturnal stable layer. The consequence, in the case of the maximum variance method, is an almost symmetric diurnal cycle, while the multisensor method cycle is clearly asymmetric following the typical evolution of the near-surface radiative energy balance. On the valley floor, the thermal turbulence triggered by the incoming solar radiation triggers sensible heat flux and, as a consequence, the mixing process, causing the near-surface layer to grow. Figures 11b,e show the average monthly ABL height in the 1100–1500 LT interval (convective period). We focus on this time period given its importance for the vertical dispersion of pollutants outside the valley. The correlation among both monthly series is high (0.6), but the amplitude of the multisensor series is higher because ceilometer-based convective layer estimates tend to be shallower when aerosol concentrations are low. Both series show a large variability around the mean value, and in some cases, the monthly mean ABL height in the convective period is lower than the average depth of the valley ($\sim 1, 100$ m). Figures 11c,f show the monthly percentage of the days when the ABL height surpasses the top of the valley during the 1100–1500 LT interval, using estimates from the Richardson and maximum variance method. On average, considering both methods, the convective boundary layer in the 1100–1500 LT interval in the Aburrá Valley exceeds the top of the mountains on 45% of the days. Months with low percentages of above-valley ABL and scarce effective below-cloud scavenging are prone to trigger critical air quality episodes.

In addition to the diurnal and annual cycles, there is important ABL variability over longer time-scales. Figures 12a,b show the evolution of the monthly diurnal cycle of the ABL height at the base of the valley using the multisensor Richardson number estimates with $Ri_b = 4$ and the maximum variance method, respectively. The black contour corresponds to the depth of the Aburrá Valley. If the ABL does not reach this critical threshold, once the convective phase ceases, the aerosols will recirculate within the valley, prompting an increase in aerosol concentration during the night as the residual layer shrinks. The monthly evolution of the ABL height diurnal variability presents evidence of significant differences among the different months, suggesting a strong variability most likely as a result of external forcing
FIGURE 11  (a) Diurnal cycle of the ABL height estimates using the multisensor Richardson number method. The horizontal dashed line denotes the average depth of the valley. The vertical dashed lines are at 1100 and 1500 LT. (b) Monthly time series of average ABL height between 1100 and 1500 LT for the multisensor Richardson number method. (c) Monthly percentage of days when the ABL height exceeds the top of the valley during the 1100–1500 LT interval, using the estimates from the Richardson method. (d, e, f) are as (a, b, c) but using the maximum variance method. In (a, b, d, e), the shaded band denotes ± 0.5 standard deviation

FIGURE 12  Evolution of the monthly diurnal cycle of the ABL height at the base of the valley using (a) the multisensor Richardson number estimates with \( R_{ih} = 4 \) and (b) the maximum variance method. The black contour indicates the depth of the Aburrá Valley. Evolution of (c) the monthly diurnal cycle precipitation, and (d) BI, at the Torre SIATA site. Dashed horizontal lines indicate the months of March in 2015 to 2018 associated with the migration of the Intertropical Convergence Zone over Colombia. Figure 12c presents the evolution of the monthly diurnal cycle precipitation at the Torre SIATA site, showing an important evolution of daytime versus night-time rainfall events, with wet seasons characterized by important night-time stratiform rainfall and daytime convective precipitation events. The importance of the ABL height, together with night-time rainfall, is evident: Figure 12d shows that months with a shallow ABL height and limited night-time rainfall result in BI peaks 100 m from the surface, a proxy
for the pollutant concentration. It can be observed how, in general, during March (dashed horizontal lines), the ABL height barely reaches the critical threshold and the night-time is scarce, leading to the highest annual concentrations. Conversely, months such as July or August, when the ABL tends to be the deepest of the year, present the lowest aerosol concentration. Since 2017, strong regulations were implemented by the environmental authority to reduce emissions during the critical air pollution episodes in March, effectively reducing the pollutant concentration compared to other years.

Figure 15a presents the relationship between the mean radiation in the 0600–1800 LT period and the anomaly of the daily mean ABL height relative to the long-term mean, showing a marked and direct relationship between radiation and the ABL height, as suggested before, with low radiation associated with ineffective growth of the ABL. Precipitation (and cloudiness) is also an important modulator of the radiation and hence the ABL height, with rainy days corresponding to below-average daily ABL height. Additionally, Figure 15b presents the scatterplot between the anomaly of the daily mean ABL height and the average daily PM2.5 concentration in the valley, showing a marked conditional association between these two variables. Days with extreme PM2.5 concentrations only occur when the daily mean ABL height is below the long-term average. Correspondingly, days with a below-average ABL height and low PM2.5 concentrations correspond to rainy days, most likely favouring below-cloud scavenging.

Other studies have also shown a link between negative anomalies of ABL height and pollutant accumulation (Mamtimin and Meixner, 2011; Liu et al., 2013; Pal et al., 2014; Pal et al., 2015; Tang et al., 2016). Pal et al. (2014) specifically studied the effect of ABL height on the diurnal variability of aerosol concentration at a valley site, finding an inverse relationship between a decreasing trend in ABL height and an increasing tendency in near-surface pollutant concentrations, in particular the fine fraction (0.3–0.7 μm). However, the evidence of Pal et al. (2014) is from a short field campaign compared to the long-term analysis included here.

3.5 ABL height spatial variability

The divergence of winds on the synoptic scale, the structure of the mesoscale circulation, and the development of valley winds driven by topography and temperature gradients, considerably influence the thermal and mechanical turbulent fluxes, which in turn modulate the evolution of the ABL in complex terrains (Bianco et al., 2011; De Wekker and Kossmann, 2015). Within the valley, the height of the ABL changes considerably from one location to another, and more so if these locations are, for example, on the valley floor and on a slope. Figure 16a shows the long-term ABL height diurnal cycle at valley floor (Torre SIATA) and western hill (Itagüí) sites using the maximum variance method. While the month-to-month evolution is similar, with the same months with a low (March, October, and November) and a high ABL (January, June, and July), the maximum ABL height in the early afternoon is considerably less at the western hill site (65–70% of the ABL at the base of the valley) and, more importantly, the time of the maximum tends to be delayed by an hour. Throughout most of the day, the absolute ABL height at the western hill location is lower than the ABL at the base of the valley; however, after 1600 LT, the opposite
occurs. Considering that the maximum variance method in the afternoon tends to favour the selection of the nocturnal stable layer, in relative terms, the onset of the nocturnal stable layer at the western hill occurs later than at the valley base. The diurnal cycle of radiation in both sites (not shown) is similar; hence the observed differences are not due to insolation but rather an indirect response due to the differences in surface wind behaviour between the two sites. Figures 16b,c show the diurnal cycle of wind direction at the base of the valley and at the western hill site, respectively. At the valley floor, winds are from the north (upvalley) all day; at the western hill, winds are downslope from the west (southwest to northwest) before 0600 LT, upslope from the east from 0600 to 1600 LT, and downslope again after 1600 LT. Upslope winds in the 0600–1600 LT period thicken the boundary-layer height locally, effectively delaying the onset of the nocturnal stable layer (De Wekker and Kossmann, 2015; Serafin et al., 2018).

### 3.6  |  WRF simulations

Figures 17a–c show the spatial distribution of the diurnal cycle of the ABL height relative to the surface, computed for the 2 km domain, using all the WRF 24 hr forecasts runs described in Section 2. Each panel represents an 8 hr period. The evolution of the diurnal cycle, as represented by WRF, shows the signature of topography and solar radiation forcing, with the development of the convective layer in the 0800–1600 LT period. The dashed line in the Figure corresponds to the cross-section of the Aburrá Valley shown in Figure 17d, and it is remarkably clear that the deepest convective layer occurs over the Aburrá Valley, in the most densely populated and urbanized area. Figure 17d shows the average diurnal evolution of ABL height over the Aburrá Valley, following the terrain as represented in the model. Each coloured line corresponds to 1 hr of the diurnal cycle. The model simulations skilfully represent the general features of the observed
**FIGURE 16**  (a) Annual cycle of ABL height at the base of the valley (Torre SIATA site) and at the western hill (Itagüí site) using the ceilometer-based maximum variance method. Diurnal cycle of wind direction at (b) the base of the valley site and (c) the western hill site. The frequency of the wind direction is plotted as a 2-hourly wind rose; times listed on the right are the beginning of the 2 hr interval.

**FIGURE 17**  Spatial distribution of the ABL height relative to the surface, computed for the 2 km WRF domain, using all the WRF 24 hr forecasts runs described in Section 2. Average ABL height over the periods (a) 0000 to 0800 LT, (b) 0800 to 1600 LT and (c) 1600 to 2359 LT. The black lines in the centre of each panel indicate the political division of the Aburrá Valley (as shown in Figure 2). (d) Average diurnal evolution of ABL height following the terrain for all the 24 hr WRF forecasts runs available. Each coloured line represents one hour of the diurnal cycle.
**Figure 18** Average ABL height in the 2 km WRF domain by IGBP land use category. Category 13 corresponds to urban and built-up areas, and category 12 to croplands. Categories 1–7, with the smallest ABL, denote different types of forests (1–5) and shrublands (6–7), and categories 8–10 savannas and grasslands.

ABL height diurnal cycle within the valley. The top of the ABL is below the height of the surrounding mountains before 1000 LT and after 1700 LT. In the 1000–1700 LT period, the radiation-driven thermal expansion of the lower troposphere leads to an ABL deeper than the valley, favouring moisture and pollutant exchange with the free troposphere. Figure 17 hints at an interesting aspect regarding the simulated thickness of the ABL in the middle of the valley compared with over the slopes and over the relatively flat terrain at the east of the valley. During both stable stratification and unstable conditions, the maximum ABL height occurs in the middle of the valley, which highlights, among other factors related to the magnitude of the turbulent heat fluxes, the deeper inversion heights in the valley compared with those over the slopes (Lehner and Rotach, 2018; Serafin et al., 2018). Figure 18 shows the average ABL in the 2 km WRF domain categorized by land use. As suggested before, the deepest ABL occurs over the urbanized area, where the integrated sensible heat flux is larger than for other areas. In contrast, the forested areas exhibit the smallest ABL heights, as a result of the small Bowen Ratio typical of tropical forests.

Figures 19a,b show, similar to Figure 12a, the evolution of the monthly diurnal cycle of the ABL height at the base of the valley and at the western hill, respectively, as simulated using WRF. The continuous (dashed) black contour corresponds to the 1,100 m ABL height contour in WRF (observations–Richardson method). From a climatological point of view, the correspondence between the observed and simulated ABL height is good for both the diurnal and annual cycles (Figure 12), suggesting that the evolution of the ABL in WRF responds well to the slow-evolving regional forcing. However, it is possible to observe that the ABL expansion rate during morning hours in the model is higher than in the observations, which is a crucial feature with practical implications for air quality. This is easier to see in Figures 19c,d, showing the comparison between the long-term average ABL height diurnal cycle from WRF forecast simulations for the grid model colocated with the Torre SIATA site (valley floor) and the Itagüí site (western hill), and the Richardson number and maximum variance method estimations. Figures 19c,d also suggest that the skill of the model simulating the ABL height evolution is considerably higher for the valley floor than for the western hill site. The latter results are different from the studies in predominantly flat terrains. The work by Hu et al. (2010) explored the performance of different planetary boundary layer schemes in the WRF model in the south central United States, given their importance simulating the ABL height. Despite the relevance of the results, the terrains there are very different from those present in the Colombian Andes and more so in the urbanized Aburrá Valley. In order to improve the ABL simulations at the hill site, it would be desirable in a subsequent study to evaluate the performance of urban canopy parametrizations as in Salamanca et al. (2011) but in complex terrains.

The skill of the model representing the diurnal and annual cycle of the ABL height appears to be the result of the proper simulation of the cloud radiative forcing. Figure 20a shows evidence of the conditional relationship between daily low-level cloudiness, represented as the 600 hPa cloud fraction, the 24 hr average short-wave radiation reaching the surface, and the ABL height. In the model, high values of low-level cloud fraction are associated with negative radiative forcing and low values of ABL height, similar to the observations. In contrast, low values of low-level cloud fraction allow, in general, the efficient expansion of the ABL during the daytime. However, the low-level cloud fraction is not the only variable controlling the short-wave radiation reaching the surface. Thus low values of low-level cloud fraction is a necessary but not a sufficient condition for
high radiative forcing and ABL expansion, as suggested in Figure 20a.

An assessment of the ABL height forecast skill at the base of the Aburrá Valley is presented in Figure 20b. The Figure shows the scatterplot and the correlation between the 24 hr ABL forecasts and the estimated ABL using the Richardson number method. The linear relationship is clear and the correlation is high (0.59), suggesting a good agreement between the observed values of ABL height and the forecasts. The only evident issue is that, in some cases, the expansion of the ABL is underestimated by the model. In general, the results of the assessment skill of the model are encouraging from the applications point of view, considering the fact that the processes involved in the advanced simulation of the ABL are complex, and depend on different coupled parametrizations.

4 CONCLUSIONS AND DISCUSSION

The primary goal of this work is the study of the variability of ABL in time and space in a narrow tropical urban valley in order to explore the interaction between the complex terrain and the tropospheric processes on local and regional spatial scales. One of the current research challenges in our understanding of the ABL over mountainous terrain is related to its multilayered structure. The development of a deep CBL in the Aburrá Valley is vital for the local development of extreme precipitation events and for reducing the likelihood of critical air quality episode occurrences. The implemented schemes use atmospheric profiles from remote-sensing retrievals and measurements, from ceilometers, a radar wind profiler, and a microwave radiometer. Four ABL height detection methods are implemented, three using ceilometer BI observations and one using the obtained non-dimensional bulk Richardson number.

In all cases, the ABL height estimates capture the diurnal cycle associated with the radiation forcing and the thermally forced convection. The temperature gradient between the surface and the lower troposphere is one of the primary modulators of the ABL structure. All ABL height estimates show a high correspondence among the methods, especially under stable conditions, which favour the appearance of marked transitions in the lower atmosphere. Under stable conditions, ABL detection using ceilometer-based methods is very reliable since aerosols become excellent tracers of ABL evolution. After the atmosphere becomes unstable and vertical dispersion takes place, the aerosol concentration near the surface and within the convective layer considerably decreases, reducing the reliability of ceilometer-based methods. In contrast, the Richardson number methodology does not depend on the existence of tracers and is, in general, more reliable than ceilometer-based methods. The principal challenge using the Richardson number technique is the selection of an optimal critical threshold for its value.

Despite the dependence on the selection of an optimal threshold, a detailed intercomparison of ABL height retrievals suggests that the Richardson number multisensor method and the ceilometer-based maximum variance technique show the highest correspondence with the Holzworth technique during the radiosonde IOPs. The main difference between the ABL height retrievals occurs after 1600 LT when the Richardson number ABL height estimations tend to detect the slow top-down contraction of the residual layer, while ceilometer-based estimations appear to detect the sudden bottom-up onset of the nocturnal stable layer. After 1600 LT, sufficient anthropic aerosols quickly replenish the control volume confined between the surface and the first atmospheric capping inversion, causing a noticeable discontinuity in the BI, consequently swaying the BI methods to detect the
Figure 20  (a) Scatterplot of the daily low-level cloudiness, represented as the 600 hPa cloud fraction, and the 24 hr average ABL height, categorized (in colours) by short-wave radiation reaching the surface. (b) Scatterplot of the 24 hr ABL forecast heights and those estimated using the Richardson number method. The Pearson correlation between the two variables is 0.59

nocturnal stable layer. The Richardson number ABL height method identifies the residual layer as a direct consequence of the turbulence not shutting down immediately with the onset of near-surface stability.

In addition to the diurnal cycle, the results suggest the existence of intra-annual and annual spectral peaks triggered by ITZC modulation of cloudiness, precipitation, and surface incident radiation. The lowest average ABL heights occur during October, November, December, and March, where the ABL height does not exceed, on average, the depth of the Aburrá Valley. In contrast, July and August are the months with the thickest ABL, exceeding the critical threshold. When the climate conditions are not favourable for ABL expansion, the anthropic emissions build up and lead to critical air pollution episodes. In this sense, the ABL height implicitly determines the pollutant concentration over time and space. In other words, the amount of aerosol in the lower troposphere is influenced both by anthropogenic emissions and by the vertical dispersion conditioned by topography and the ABL dynamics, namely, the evolution of the ABL height. The meteorological conditions during March are, as a result of regional-scale forcing, the most adverse for the vertical dispersion of pollutants, thus becoming the most sensitive period of the year for reaching unhealthy pollutant concentrations within the valley. During this month, ABL is relatively shallow, and the effect of scavenging processes is limited given the reduced nocturnal rainfall. These results for the transition season (March), in contrast to the dry and wet seasons, hint at an essential question in air quality studies regarding the estimation of the relative portion of particulate matter that is removed from the local atmosphere by convectively triggered vertical dispersion and by scavenging processes. In March 2016, for example, the average monthly ABL did not reach the critical threshold, limiting vertical aerosol dispersion. The unfavourable conditions were exacerbated by a prolonged absence of recurrent precipitation as a result of El Niño conditions limiting the scavenging processes, thus leading to a “perfect storm” type condition favouring pollutant accumulation and leading to the occurrence of the most critical air quality episode in the region.

Regarding the spatial variability, the absolute ABL maximum height above the surface at the western hill is considerably less than at the valley floor, and the peak time of the maximum height tends to be delayed by approximately 1 hr. Additionally, the onset of the nocturnal stable layer at the western hill occurs later than at the base of the valley. These disparities appear to be linked to the differences in surface wind direction, with upslope winds at the western hill acting to thicken the boundary-layer height locally. In contrast, WRF model simulations skilfully represent the general features of the observed ABL height diurnal cycle within the valley, indicating that the thickness of the ABL in the middle of the valley is greater than over the slopes and higher-altitude flat terrain. The model simulations also capture the ABL annual cycle resulting from the slow evolving regional forcing associated with the migration of the ITCZ. However, the skill of the model simulating the ABL height evolution is considerably higher for the valley floor than for the western hill site, emphasizing the need to improve our current understanding and modelling capabilities for boundary layers over complex terrain. The assessment of the 24 hr forecasts of the ABL height at the base of the valley is quite promising, showing a high correlation between the simulated and estimated values. These results are remarkable considering the fact that the processes involved in the simulation of the ABL depend on different parametrizations.

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The authors declare that they have no conflict of interest.

ORCID

Laura Herrera-Mejía @ https://orcid.org/0000-0002-0501-0545
Carlos D. Hoyos @ https://orcid.org/0000-0002-9010-9784

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