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Atmospheric Velocity Spectral Width Measurements Using the Statistical Distribution of Pulsed CO₂ Lidar Return Signal Intensities

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ABSTRACT

A pulsed CO₂ lidar with coherent detection has been used to measure the correlation time of backscatter from an ensemble of atmospheric aerosol particles which are illuminated by the pulsed radiation. The correlation time of the backscatter of the return signal, which is directly related to the velocity spectral width, can be used to study the velocity structure constant of atmospheric turbulence and wind shear. Various techniques for correlation time measurement are discussed, and several measurement results are presented for the technique using the information contained in the statistical distribution of a set of lidar return signal intensities.

1. Introduction

Extensive studies of atmospheric wind shear and wind velocity fluctuations have been performed using various instrumental techniques to deduce the kinematic structure of the lower troposphere in quiet and stormy weather. The two major spectral broadening mechanisms of Doppler lidar signals, due to the atmosphere itself, are wind shear and wind velocity fluctuations associated with turbulent motions. The shear term can be calculated from the range dependence of the first moment of the Doppler lidar spectrum. It can also be estimated from balloon radiosounding data, in which case one could then deduce the turbulence-related term after subtracting the contribution of the shear effect to the spectral width, providing instrumental effects are small (e.g., high laser frequency stability, suitably long transmitted pulse length). Although Doppler lidar techniques are restricted to clear air weather conditions, wind field measurements by airborne Doppler CO₂ lidars have recently been demonstrated (Bilbro 1986; Woodfield and Vaughan 1983), offering new prospects for the use of mobile Doppler lidars in mesoscale investigations of the planetary boundary layer (PBL).

In this study, we discuss coherent lidar measurements of wind shear and velocity fluctuations. Although similar measurements can obviously be obtained using different wind profilers, e.g., VHF or microwave radars, or acoustic sounders (Strauch 1985; Balser and Netteville 1981), the coherent lidar has certain complementary properties. In the limit of the Kol-

mogorov theory, all the statistical properties of atmospheric turbulence can be related to one parameter, ϵ , the dissipation rate of the turbulent kinetic energy (Pasquill 1974). To relate ϵ to the Doppler spectrum width (after removal of the shear contribution), one must assume that the size of the largest eddies within the inertial range is larger than or approximately equal to the size of the lidar resolution volume (Doviak and Zrnic 1984). In many cases, this is an adequate assumption. The lidar probe volume has a shape similar to that of a pencil, with the largest dimension related to the pulse length. For a pulsed CO₂ Doppler lidar the pulse length typically ranges from 1 to 3 μ s ($\Delta R = 150$ to 450 m). New technological developments will soon lead to the possibility of equivalent measurements with a Nd:YAG coherent lidar, with pulse lengths almost an order of magnitude shorter, implying a range resolution $\Delta R = 15$ –45 m.

Feasibility studies of Doppler lidar performance for global wind measurements have assumed laser pulse lengths of 5–7 μ s and a matched filter bandwidth (Hufaker 1985; Osmundson and Martin 1985); however, the optimum time window to calculate the power spectrum may be limited to smaller values by atmospheric effects. Turbulence and wind shear place a limitation on the temporal coherence of Doppler lidar returns in the visible and infrared wavelengths. These atmospheric effects should be assessed for various atmospheric conditions and at different altitude levels in the troposphere.

This paper presents atmospheric range-gated measurements of velocity spectrum width, using a ground based pulsed CO₂ lidar, for various atmospheric conditions, making use of supporting standard meteorological data obtained from balloon radiosounding. In a previous paper (Ancellet and Menzies 1987) we in-

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licated how measurements of the probability distribution function (PDF) of atmospheric backscatter lidar signal intensity leads to the velocity spectrum width, using the aerosol particles as tracers of atmospheric motion in the resolution volume. We discussed the measurements in terms of the temporal coherence of the Doppler lidar return signal. In this paper we present range-gated measurements and a more detailed analysis of the limiting factors associated with this measurement technique.

2. The measurement technique

a. Measurement of the coherence time

The received lidar signal arises from the backscatter of a distant volume element which contains aerosol particles moving with mean velocity equal to the bulk wind velocity. The PDF of an atmospheric backscatter coherent lidar signal intensity generally resembles an exponential density function, which is characteristic of a fully developed speckle pattern, but it changes to a K distribution when significant atmospheric turbulence is present (Gudimetla and Holmes 1982) or a modified Gamma distribution when time averaging (using a low pass filter at the receiver) is over periods longer than the coherence time of the signal (Parry 1975). The PDF is calculated from a set of several laser shots (>100). Using the first and the second moments of the statistical distribution of the received intensity at a particular time (range), one calculates the ratio of the mean to the square root of the variance, referred to as the inverse relative root variance (IRRV), which is related to the carrier-to-noise ratio (CNR), the number of independent spatial modes detected at the receiver (M_A), and the number of independent temporal coherence samples (M_T) contained in the integration time (range bin) of the receiver (Shapiro et al. 1981):

$$\text{IRRV} = \left(\frac{\text{CNR}/2}{1 + \text{CNR}/(2M_A M_T) + 1/(2\text{CNR})} \right)^{1/2}. \quad (1)$$

The CNR is the ratio of the mean signal power to the mean noise contribution. For $\text{CNR} \gg 5$, and $M_A \approx 1$ (i.e., only one spatial mode is detected). Eq. (1) becomes

$$\text{IRRV} = \sqrt{M_T}(1 + 2M_T/\text{CNR})^{-1/2}. \quad (2)$$

These two conditions are generally applicable for measurement in the lower troposphere using a typical pulsed CO_2 coherent lidar. Therefore the knowledge of the PDF leads to the number of independent temporal samples M_T .

Then one may express the number of coherent samples in the lidar range resolution element as

$$M_T = 1 + 2/(B\tau_c), \quad (3)$$

where B is the smallest full bandwidth of the receiver electronics and τ_c is the coherence time of the return signal. This equation is a direct result of the expression

of measured current fluctuations in heterodyne detection, assuming that the photocurrent is a gaussian random variable and that the admittance of the (postdetection) low-pass filter has a Lorentzian shaped band-pass (e.g., an RC filter). The validity of the relation expressed in (3) has been proven by Fukuda et al. (1984), who recorded IRRV measurements in the boundary layer using a pulse duration ($\tau \leq 200$ ns) shorter than the expected atmospheric correlation time in the boundary layer and obtained a coherence time equal to the pulse length, using (3). We observed the same behavior when operating our CO_2 lidar transmitter with short pulse durations. (These measurements are useful in proving the validity of (3), since they correspond to IRRV measurements in the case of a known coherence time, τ_c , of the detected signal.)

When using a long laser pulse duration, i.e., larger than the atmospheric aerosol correlation time, ($\geq 3 \mu\text{s}$), the coherence time of the backscatter signal is a function of velocity fluctuations or wind shear in the resolution element. Then τ_c is given by $\tau_c = 1/(2k\sigma_v)$ (Churnside and Yura 1983) and (3) becomes

$$M_T = 1 + 4k(\sigma_t + \sigma_s)/B, \quad (4)$$

where k is the wavenumber; σ_t and σ_s are, respectively, the Doppler spectral width contributions due to the wind velocity fluctuations associated with turbulence and due to the wind shear, and $\sigma_v = \sigma_t + \sigma_s$.

b. Instrumental and atmospheric decorrelation processes

To derive (4) from (3), we neglected decorrelation processes that are not related to wind velocity fluctuations or wind shear within the resolution volume. Although they might be small, they have to be identified and assessed to determine the measurement accuracy. The transmitted pulse length is not the only parameter to influence the coherence time of the backscatter signal. It might also be limited by the existence of large gradients in the aerosol distribution within the resolution volume. One has to be concerned with this limitation when measurements at the top of the boundary layer are needed; this can be studied by transferring the aerosol backscatter variation in the resolution volume to an equivalent laser pulse envelope variation. Using a laser pulse modeled as a two-step function, which can satisfactorily represent a typical TEA CO_2 laser pulse (e.g., a 200-ns narrow gain switched spike followed by a 2.8- μs tail containing 70% of the total energy), we calculated the combined effect of an aerosol distribution decay at the top of the boundary layer and of the $1/R^2$ dependence of the lidar signal on the laser pulse envelope. We assumed that the lidar is probing a resolution element located at a 1-km range from the lidar, viewing it at an angle of 30° with respect to the horizontal. Both a ground based lidar ($+30^\circ$ elevation) and an airborne lidar (-30° elevation) were considered

because the aerosol distribution decay with increasing altitude will weight the results towards shorter ranges in the first case and towards longer ranges in the second case. In both cases, the $1/R^2$ dependence will weight the results towards shorter ranges. Based on aerosol distributions presented in the next section, an aerosol vertical decay rate of 35 db/km in the resolution volume was assumed. The initial transmitted pulse and the two modified pulse envelopes are shown in Fig. 1. For a ground-based lidar, the relative importance of the narrow gain switched spike has been largely reduced. This leads to a small improvement in the coherence of the return signal. For an airborne lidar, the pulse envelope remains essentially identical to the initial pulse as a result of the two counteracting effects of aerosol distribution decay and the $1/R^2$ dependence of the return (assuming R approximately equal to 1 km). [For this particular modeled laser pulse shape, the effect of the pulse envelope itself on the coherence time of the return signal was calculated in a previous paper (Ancellet and Menzies 1987) and found to be $2 \mu\text{s}$, based on the calculated width of the signal autocovariance function.]

When turbulence is present along the path between the lidar and the volume element, the K distribution has been proposed by Gudimetla and Holmes (1982) as a reasonable approximation for the PDF of the lidar return intensity. They showed that, as the turbulence increases, the normalized variance of intensity rises above unity and reaches a peak value of approximately 1.25 at a Rytov variance, σ_x^2 , of one tenth (for a wavelength of $0.48 \mu\text{m}$). An expression for the IRRV was also proposed by Shapiro et al. (1981), for the propagation of a speckle field through a turbulent atmosphere. Assuming, for simplicity, M_A and M_T equal to unity and a nearly diffraction limited transmitted beam, this expression becomes

$$\text{IRRV} = \left(\frac{\text{CNR}^*/2}{1 + \text{CNR}^*/2A + 1/(2 \text{CNR}^*)} \right)^{1/2} \quad (5)$$

where $\text{CNR}^* = \text{CNR} \exp(4\sigma_x^2)$ and $A = \exp - (16\sigma_x^2)$. Considering daytime measurements at $10\text{-}\mu\text{m}$ wavelength along a slant path (few degrees above the horizontal) at a range of 5 km, one expects a Rytov variance of 0.01. For high CNR values, this leads to $\text{IRRV} = 0.92$. This corresponds to an equivalent velocity variance error of $3 \times 10^{-2} \text{ m s}^{-1}$ for a 500-kHz receiver bandwidth. Thus the effect of turbulence along the propagation path is negligible compared to other sources of error.

c. Comparisons with other techniques

A more direct way to measure the velocity spectrum width of the signal is to compute the signal spectrum itself for each shot. Obtaining an estimated width of the spectrum for each shot is a definite advantage compared to the IRRV technique, which requires the ac-

FIG. 1. Modification of the influence of the transmitted pulse shape (modeled as a double-step function) on the backscatter contribution function, due to an aerosol gradient in the range gate at 1 km (as described in the text), for lidar elevation angles $\theta = \pm 30^\circ$.

quisition of several shots (~ 100) to obtain the PDF of the signal. This limitation might even be a serious problem when the advecting atmosphere is far from being stationary in a statistical sense. However, from experimental measurements of the aerosol structure function in the boundary layer, Post (1978) concluded that the spatial structure of atmospheric aerosols was in approximate steady state motion for observation times less than 30 min. From our experimental observations of aerosol backscatter profiles resulting from 4-min averaging times (and with a 150-m spatial resolution), we also concluded that the mean aerosol backscatter value in each resolution volume usually varies by less than 10%–15% during a 20-min period. (See section 3.b.)

The major disadvantage of a direct measurement of the spectrum lies in the difficulty of removing the effect of the spectral processing itself. The two most common ways to estimate moments of the signal spectrum are based on the use of a Fast Fourier Transform algorithm or a poly-pulse-pair processor where the signal is converted from an analog signal to a digital sample and then subjected to autocovariance or spectral processing to produce the first three moments (Hardesty et al. 1983). The finite time window associated with the discrete nature of the spectrum will produce a velocity spectrum width; e.g., for a sampling frequency of 50 MHz and a time window of $2.5 \mu\text{s}$ the width can be as large as 6 m s^{-1} when the mean frequency falls between two adjacent discrete frequency bands, the bias of the estimated mean remaining less than 0.5 m s^{-1} (Ancellet and Menzies 1987). This effect could be corrected pro-

vided that the mean frequency is accurately known; unfortunately this is not usually the case. Because of the relatively small bias of the estimated mean frequency, one could alternately use the statistical fluctuations of the mean frequency to compute σ_v by recording a set of several spectra. However, as is true for the IRRV technique, there may be a limitation due to the length of the observation time; the motions of the aerosol inhomogeneities with large spatial scales (>500 m) would contribute to the broadening of the spectrum, reducing the ability to relate velocity spectral width to purely turbulent motions. This method is also sensitive to interpulse frequency jitter, in contrast to the IRRV method.

3. Experimental data

In this section, we will present the field measurements that were made at JPL, Pasadena, CA, using a pulsed coherent CO₂ lidar system operating at the 10P(20) laser line (10.59 μm) which provides the strongest return for boundary layer study. The experimental setup and data processing scheme were described in earlier publications (Ancellet and Menzies 1987; Menzies et al. 1984). Before recording data, we verified the intrapulse frequency stability of the TEA CO₂ laser by adjusting the cavity alignment so that consistent and reliable transverse mode selection could be obtained. [Occasionally, poor transverse mode discrimination has been proven to be the main factor limiting the laser intrapulse frequency stability (Ancellet et al. 1987).] This precaution is taken to be sure that the velocity spectral resolution is limited only by the transform-limited laser pulse duration. For small velocity spectrum width (<1 m s⁻¹), an attempt to remove the nonnegligible influence of the laser pulse (0.3 m s⁻¹) has been made for the data presented in this section by deconvolution. Both spectra in this case were modeled as Gaussian, admittedly a coarse approximation for the pulse spectrum. This assumption affects mea-

surements of velocity spectral width less than 0.5 m s⁻¹. This could be considered as a detection limit.

a. Planetary boundary layer (PBL)

First we conducted several measurements of the IRRV at a given range of 1.5 km along a nearly horizontal path. (Results are summarized in Table 1.) Measurements in the PBL along a horizontal path are not dependent on vertical wind shear and are, therefore, representative only of turbulence level encountered in the PBL. The lidar was located on a hillside at an altitude of 450 m above sea level, and a range of 1.5 km along a horizontal path corresponded to an altitude of 80 m above the ground. We recorded sets of successive aerosol return signals under various atmospheric conditions. On two occasions, we also used different bandwidths and recorded two or three successive sets of data in order to verify experimentally the sensitivity of the method. The differences between those data (which are representative of the same turbulence conditions) are very small (~ 0.1 m s⁻¹) when converted to velocity spectral width. The values obtained for the various days are all about 0.5 m s⁻¹ except on 22 January 1986, where very low values were recorded. This day was characterized by very low visibility (<2 km), associated with stable conditions which contributed to a significant haze over the Los Angeles Basin, and by a cloudy sky, which reduced the thermal convection usually observed during the daytime. The concurrence of those two factors seems to have reduced the turbulence to a level such that very little spreading of the velocity spectrum was observed. It is interesting to compute the dissipation rate of the turbulent energy using a value of 0.5 m s⁻¹ for σ_v . Assuming a Kolmogorov spectrum (Tatarski 1971), the power spectral density $S(k)$ is related to ϵ at the wavenumber k :

$$S(k) = A\epsilon^{2/3}k^{-5/3} \quad (6)$$

where $A \approx 0.5$ (MacCready 1964; Gossard et al. 1984). By integrating the spectrum for the wavenumbers larger

TABLE 1. Observations along a horizontal path in the PBL.

Date	Time (LST)	IRRV	B (kHz)	σ_v (m s ⁻¹)	Range (km)	Meteorological observations
17 Oct 85	1530	1.92	500	0.51	1.35	PBL visibility = 6 km
	1630	2.38	200	0.41	1.35	Partly cloudy sky, $T = 20.8^\circ\text{C}$
21 Oct 85	1030	1.85	500	0.46	1.35	PBL visibility = 4 km, cloudy sky; rain at 12:00; $T = 16.4^\circ\text{C}$
22 Oct 85	1515	2.94	100	0.31	1.2	PBL visibility = 5 km, hazy,
	1545	1.79	500	0.42	1.2	clear sky above
	1615	2.78	100	0.27	1.2	
22 Jan 86	1600	1.05	500	<0.1	1.5	PBL visibility = 2 km, foggy, cloudy sky. $T = 18.1^\circ\text{C}$
24 Jan 86	1000	1.74	500	0.39	1.4	PBL visibility = 4 km, clear sky $T = 21.8^\circ\text{C}$

than $1/L$ where L is the length of the resolution volume, one obtains a σ_v that in turn can be used to derive ϵ

$$\epsilon = \frac{\sigma_v^3}{L} \cdot (3A)^{-3/2}. \quad (7)$$

We obtained a value of $9.3 \times 10^{-4} \text{ m}^2 \text{ s}^{-3}$, which corresponds to a light to moderate turbulence regime according to the turbulence intensity scale defined by MacCready. Comparison with C_n^2 is rather difficult since C_n^2 is not only related to ϵ but also to N , which is the analog of ϵ for the temperature field fluctuations:

$$C_n^2 = \left(\frac{(8 \times 10^{-5})p}{T^2} \right)^2 C_T^2 \quad (8)$$

$$C_T^2 = \frac{BN}{\epsilon^{1/3}} \quad (9)$$

where B is a numerical constant of the order of unity, and p and T are the pressure and temperature fields. However, as pointed out by Tatarski, a relation exists between ϵ and N in the case of free convection, since both can be expressed in terms of the temperature flux. From these one obtains an equation for C_n^2 as a function of ϵ (independent of altitude in this case) and the altitude z :

$$C_n^2 = 0.7 \left(\frac{(8 \times 10^{-5})p}{gT} \right)^2 \left(\frac{\epsilon}{z} \right)^{4/3}. \quad (10)$$

Using a value of $9.3 \times 10^{-4} \text{ m}^2 \text{ s}^{-3}$ for ϵ and an altitude of two meters above ground, we obtain a C_n^2 value of $2.0 \times 10^{-13} \text{ m}^{-2/3}$, assuming the measurements are representative of free convection conditions. This value is typical of what we expect for daytime conditions with clear sky. It is, therefore, encouraging that our estimation of turbulence levels based on velocity spectrum width measurements compared well with other estimates of the turbulence in the PBL during the daytime.

b. Interface PBL/free troposphere

1) THE DATA COLLECTION

After the investigations of the velocity spectrum width in the boundary layer, we focused our attention on the top of the PBL, because this is a zone of transition

where we generally should observe variations in the turbulence or wind shear regime when scanning through that region. Using data obtained along a short path, we probed the top of the PBL with a vertical resolution of $\Delta z \approx 200 \text{ m}$. The measurement dates and related information are in Table 2. For all cases except 22 July 1986, we also recorded a profile of the aerosol backscatter coefficient, β , along the line of sight in order to calculate the height of the PBL or to characterize the possible existence of successive aerosol layers rather than a well-defined and uniform PBL. (The uncertainty in the absolute accuracy of the β values presented in this paper is approximately $\pm 6 \text{ dB}$). On two days we also compared our measurements with balloon radiosondes which were launched 200 m south of the lidar location. They provided vertical profiles of temperature, relative humidity, and pressure. At the same time the vertical profiles of velocity and direction of the horizontal wind were also obtained by tracking the balloon with a theodolite. On 17 January 1986, we made two measurements of σ_v in a single range gate but with two different elevation angles (measured from a horizontal plane). The two measurements were separated in time by 30 min and corresponded to two different altitudes. As shown by the two aerosol backscatter profiles (Fig. 2) recorded in 15-min intervals near 4:00 pm and 5:00 pm, the PBL was growing on that day with a vertical speed of 10 cm s^{-1} ; therefore, the first measurement was characteristic of the atmosphere 300 m above the PBL, and the second one of the top of the PBL. The contrast between the two σ_v values obtained is a good indication of the level of dynamic activity that takes place at the top of the PBL in terms of turbulence and wind shear.

For the last three dates reported in Table 2, we recorded σ_v measurements in successive range-gates to obtain a profile of σ_v in an approximately 1-km thick atmospheric layer with 100-m altitude resolution. The three profiles are displayed in Fig. 3.

2) LARGE SCALE AEROSOL MOTION

On 6 February 1987, we obtained the aerosol backscatter profiles using another CO_2 coherent lidar system, located next to the lidar used for σ_v measurements, and usually devoted to DIAL measurements in the PBL

TABLE 2. Profiles of σ_v at the top of the boundary layer.

Date	Time (LST)	Azimuth angle (°)	Elevation angle (°)	Aerosol backscatter profile	Meteorological radiosounding
17 Jan 86	1615	180	14	yes	no
	1645	180	26	yes	no
22 Jan 86	1400	180	40	no	no
4 Aug 86	1530	180	40	yes	yes
6 Feb 87	1345	180	30	yes*	yes

* Backscatter data were recorded with another lidar, described by Grant et al. (1987). See text for further description.

FIG. 2. Aerosol backscatter vertical profiles and corresponding velocity spectral width measurements at the indicated altitudes.

(Grant et al. 1987). This provided the advantage of simultaneous measurements of σ_v and of the aerosol backscatter profile. Higher temporal and spatial resolution could be attained with this system, which has a typical pulse repetition frequency (prf) of 25 Hz and transmits pulses with 300-ns duration. A first set of three profiles and a second set of four profiles were recorded, using for each individual profile an acquisition time of 5 min and an altitude resolution of 75 m. The first and the last profile, separated by 45 min, are displayed in Fig. 4, showing substantial differences in the free troposphere and also in the PBL, the altitude of which was about 2 km. We also plotted the time series of the aerosol backscatter measured at various altitudes (Fig. 5) in order to determine the observation time for which the atmosphere can be safely assumed as stationary. Within the PBL or at the transition between the PBL and the free troposphere, it is not un-

FIG. 3. Velocity spectral width vertical profiles.

FIG. 4. Aerosol backscatter vertical profiles measured using MAPM.

reasonable to use an observation time of 20 or 30 min but it is clearly not the case at altitudes above 2.8 km. It should be mentioned that a lidar system combining high PRF, high energy and long pulse duration could perform both σ_v and aerosol backscatter measurements.

3) DISCUSSION

On 22 July 1986, high values of σ_v ($\geq 5 \text{ m s}^{-1}$) were obtained at an altitude which seems to be the top of the PBL, based on an analysis of the mean values of the atmospheric returns recorded in the successive

FIG. 5. Temporal variations of 5-min averaged aerosol backscatter at indicated altitudes.

range gates. Because no supporting meteorological information was available on that day, we can only make conjectures about the mechanisms which produced such high values. One explanation might be the existence of a strong vertical wind shear of about $2 \times 10^{-2} \text{ s}^{-1}$ aligned with the direction of the lidar. The turbulence generated by the existence of a wind shear might also combine with the direct broadening effect of the shear to produce a spectral width as large as 5.5 m s^{-1} .

On 4 August 1986, the aerosol backscatter profiles (Fig. 6) reveal the existence of a 1.5-km thick homogeneous aerosol layer superposed on a thin haze layer close to the ground and fairly stable, as indicated by the small differences between the two profiles. The radiosonde data indicate an inversion layer between 0.9 km and 1.4 km (Fig. 7) and also a very strong wind shear of about $2 \times 10^{-2} \text{ s}^{-1}$ between 1 km and 1.5 km. However, because a westerly wind regime was predominant at that altitude, the contribution to the velocity spectrum broadening remains small except between 1.1 km and 1.3 km, where the wind direction was slightly northwest, and where the vertical shear of $1.4 \times 10^{-2} \text{ s}^{-1}$ in the southerly component of the wind (Fig. 8) contributed to a σ_v (the wind shear component) value of 1.4 m s^{-1} , comparable to that obtained for σ_v . No obvious effect of the turbulence induced by the wind shear can be deduced from this profile, and this appears consistent with Richardson number values larger than 0.2 near 1 km, which indicate that the turbulent motions associated with the wind shear were probably inhibited by the existence of the inversion layer at the same altitude.

On 6 February 1987, the velocity spectral width measurements corresponding to an observation time smaller than 30 min are quite similar to those obtained on 22 July 1986. However, near 1.5 km, the contri-

FIG. 6. Aerosol backscatter vertical profiles.

FIG. 7. Dry temperature (T) and potential temperature (θ) versus altitude, from radiosonde measurements. The Richardson number profile is indicated within the inversion layer.

bution of a shear in the north-south component of the wind accounts for a spectral broadening of only 1 m s^{-1} as a result of a moderate wind shear amplitude of $1.1 \times 10^{-2} \text{ s}^{-1}$ (Fig. 9). Consequently the high σ_v values recorded just below the top of the PBL do not seem related solely to the existence of a wind shear. Rather large turbulence conditions might have been encountered near the top of the PBL and at the bottom of the shear, considering the temperature lapse rate of -8.3 K km^{-1} existing on that day (Fig. 10), characteristic of a moderately unstable atmosphere. Nevertheless, this would produce a σ_v value of only 1.5 m s^{-1} , assuming a dissipation rate of $9 \times 10^{-2} \text{ m}^2 \text{ s}^{-3}$ and a resolution volume length of 450 m. Therefore, the maximum expected value for σ_v at 1.5 km would be 2.5 m s^{-1} , a factor of two less than that obtained. The reason of this discrepancy is not clear. One reason might be related to inadequate vertical resolution in the wind velocity measurement, which was obtained by a somewhat coarse method (i.e., visual tracking of a balloon using a theodolite). The existence of a large vertical shear of the horizontal wind in a layer of 100 m thickness or less would be difficult to detect using this technique. However, such a shear would strongly influence the velocity spectral width measured by the lidar.

FIG. 8. Wind horizontal component aligned with the lidar direction, deduced from balloon tracking.

FIG. 9. Wind horizontal component along the lidar azimuth, deduced from balloon visual tracking.

4. Conclusion

We have described an investigation of a new method for measurement of the velocity spectrum width in the lower troposphere. Instead of relying on a direct measurement of the width of the spectrum, which generally requires discrete spectral processing or discrete autocovariance processing, our method makes use of the statistical distribution of the backscatter lidar signal in order to compute the inverse relative root variance (IRRV). Although the observation time for this technique is longer than for the direct measurement, no spectral processing is needed. Therefore, we avoid the broadening of the spectrum by the processing itself, which is often difficult to determine when using coherent Doppler lidar.

The experimental data, obtained using CO₂ coherent lidar at 10- μ m wavelength, demonstrated the potential of this method for investigations of wind shear and/or turbulence in the planetary boundary layer (PBL). Turbulence strength values C_n^2 , deduced from wind velocity variance measurements in the lidar resolution

volume within the PBL, compared well with expected values for the weather conditions reported in Table 1. The lidar was also used to record profiles of the velocity spectrum width across the transition zone between the PBL and the free troposphere. For the two cases where visual balloon tracking provided wind velocities and directions, the spectrum broadening was shown to be related to the existence of wind shears near the top of the PBL. Values as large as 5 m s⁻¹ are likely to be the result of a combined effect of wind shear within the resolution volume and of turbulence generated by the existence of the shear. Investigations such as these are encouraging for further use of this method to obtain more information about the mechanisms responsible for velocity spectrum broadening. However, to be able to separate more accurately the wind shear contribution from the turbulence contribution, better estimates of the wind velocities and directions must be obtained using either the Doppler information contained in the lidar signal itself or in situ measurements.

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FIG. 10. Dry temperature (T) and potential temperature (θ) versus altitude, from radiosonde measurements.

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