The Development of a Scanning Raman Water Vapor Lidar for Boundary Layer and Tropospheric Observations

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ABSTRACT

A scanning, ultraviolet, Raman water vapor profiling lidar designed primarily for boundary layer measurements has been built and operated by the Los Alamos National Laboratory Ground-Based Earth Observing Network team. The system provides high temporal and spatial resolution measurements of the atmosphere within and above the atmospheric boundary layer (ABL). Several examples of the types of data collected and the techniques for processing the data are presented. The typical horizontal range for the lidar is approximately 700 m when scanning, while the vertical range with photon counting can be up to 12 km with corresponding spatial resolutions of 1.5 m in the near field to 75 m in the far field. The uncertainty in the water vapor mixing ratio was found to be \( \pm 0.34 \) g kg\(^{-1}\). The development of the scanning Raman lidar is directed at questions about the behavior of the surface atmosphere interface. These questions address the nature of spatial variability and intermittent microscale convective transport in the ABL and lower troposphere.

1. Introduction

A study of the feedback mechanisms involved in global circulation models was tested during the Central Equatorial Pacific Experiment (CEPEX) in March of 1993 (Kuettner 1993). Los Alamos National Laboratory participated in this experiment by fielding a Raman water vapor profiling lidar specifically designed and built for use on the R/V Vickers. This Raman lidar was used to characterize the distribution of water vapor in the lower troposphere. The lidar was mounted on the forward part of the uppermost deck of the Vickers, approximately 8 m above the ocean surface. After this experiment, the lidar was further modified by adding a turret-type scanner that allows scanning 360° in azimuth and ±22° in elevation. In 1996, the lidar was again mounted on a ship and fielded to the central equatorial Pacific as part of the Combined Sensor Program (CSP) to quantify boundary layer water vapor variability (Post et al. 1997).

Data for CEPEX were collected during the period of 10–22 March 1993, and data from CSP were collected from 16 March to 4 April 1996. Since the system was capable of day- and nighttime operation, the limiting factor for continuous data collection was bad weather. The Raman lidar can be operated simultaneously in a current mode and a photon-counting mode. The current mode maximizes the near-field high spatial resolution for boundary layer observations, while the photon-counting mode provides long-range soundings. Current-mode operation uses direct, high-speed digitization (100 MHz) of the signal from photomultiplier tubes (PMTs) allowing for high signal-to-noise measurements in the near field of the lidar. The photon-counting mode uses a multichannel scaler to count the number of incoming photons arriving during successive time intervals (of 500 ns). This methodology provides long-range measurements, albeit with lower range resolution in the far field. Thus, in the near field, high spatial resolution (1.5 m) data are available up to an altitude of 2500 m or horizontal range of 700 m in the scanning mode, while far field, lower spatial resolution (75 m) data are available up to about 4 km during the day, and as high as 12 km at night. The standard deviation between lidar and radiosonde hygrometer data taken at concurrent time and space was found by regression to be \( \pm 0.34 \) g kg\(^{-1}\). Much of this uncertainty is due to significant differences between the lidar and radiosonde, which occur in regions of rapid change in water vapor with altitude. In these regions, the slower time resolution of the ra-
diosonde’s capacitance hygrometer “averages” the changes in atmospheric structure, as compared to higher-resolution lidar data (Cooper et al. 1996).

This paper describes the physical aspects of the lidar instrument and general data reduction used in CEPEX and CSP. Details of the design and operating characteristics are discussed as well as several of the system limitations. In addition, emphasis is placed on data analysis techniques and the types of information that may be obtained from this system. Examples of derived information includes the multidimensional images of water vapor mixing ratio and latent energy evaporative flux from the sea surface.

2. Measurement methodology
   a. Raman lidar technique

   The solar-blind Raman water vapor lidar used in these experiments is based upon the Raman technique pioneered by Melfi et al. (1969) and Cooney (1970) and extended for daytime, solar-blind operation by Renault et al. (1980), Cooney et al. (1985), and Renault and Capitini (1988). The device operates by emitting a pulsed ultraviolet laser beam into the atmosphere. Nitrogen gas and water vapor react to this light via the Raman scattering process, causing light of longer wavelengths to be scattered back to the lidar. During the day, the system operates in the solar-blind region of the spectrum using krypton fluoride as the lasing media to obtain light at 248 nm. The Raman-shifted nitrogen signal returns at 263 nm and the Raman-shifted water vapor signal returns at 273 nm. At night, when the solar background light is negligible, xenon fluoride is used in the laser to generate 351-nm light. The Raman-shifted nitrogen signal returns at 382 nm and the water vapor Raman-shifted signal returns at 403 nm. Since wavelengths longer than 300 nm are not strongly absorbed by atmospheric ozone, and Rayleigh scattering is reduced at longer wavelengths, greater range is possible at night than in the day. Simultaneous measurement of the water vapor and nitrogen returns provides a simple method for obtaining absolute measurements. Because nitrogen is the dominant atmospheric gas, dividing the Raman-shifted return signal from water vapor by that of nitrogen normalizes each pulse and corrects for first-order atmospheric transmission effects, variations in laser energy from pulse to pulse, and telescope field-of-view (FOV) overlap with the laser beam. The divided returns are then proportional to the absolute water vapor content of the air. A correction is required to account for the differential atmospheric attenuation between the nitrogen and water vapor wavelengths.

   The backscattered signals from the elastic, nitrogen, and water vapor channels are given by the following equations:

\[
P_{\text{elastic}}(r) = \frac{C_1 E [n_e(r) \sigma_{\text{scatter,aerosol}}(r) + n_g(r) \sigma_{\text{scatter,Rayleigh}}(r)] \exp \left\{ -2 \int_0^r \left[ \alpha_{\text{atten,Rayleigh}}(r') + \alpha_{\text{atten,aerosol}}(r') \right] dr' \right\}}{r^2},
\]

\[
P_{N_2 \text{Raman}}(r) = \frac{C_2 E n_{N_2}(r) \sigma_{\text{scatter,N}_2 \text{Raman}}(r) \exp \left\{ -2 \int_0^r \left[ \alpha_{\text{atten,water}}(r') + \alpha_{\text{atten,N}_2}(r') \right] dr' \right\}}{r^2},
\]

\[
P_{H_2O \text{Raman}}(r) = \frac{C_3 E n_{H_2O}(r) \sigma_{\text{scatter,H}_2O \text{Raman}}(r) \exp \left\{ -2 \int_0^r \left[ \alpha_{\text{atten,water}}(r') + \alpha_{\text{atten,H}_2O}(r') \right] dr' \right\}}{r^2},
\]

where \(P_{\text{elastic}}(r)\), \(P_{N_2 \text{Raman}}(r)\), and \(P_{H_2O \text{Raman}}(r)\), are the received signal strengths in each channel; \(r\) is the range from the lidar; \(E\) is the laser energy per pulse; \(\sigma_{\text{scatter,aerosol}}\), \(\sigma_{\text{scatter,Rayleigh}}\), \(\sigma_{\text{scatter,N}_2 \text{Raman}}\), and \(\sigma_{\text{scatter,H}_2O \text{Raman}}\) are the microscopic Rayleigh, Mie, and Raman backscatter coefficients (\(\text{m}^2 \text{sr}^{-1}\)) at 180° for the laser wavelength; \(n_e(r)\), \(n_g(r)\), \(n_{N_2}(r)\), and \(n_{H_2O}(r)\) are the number densities of the particulate and molecular scattering centers (\(\text{m}^{-3}\)) and those of nitrogen and water vapor specifically; \(\alpha_{\text{atten,Rayleigh}}\), \(\alpha_{\text{atten,aerosol}}\), \(\alpha_{\text{atten,N}_2}\), and \(\alpha_{\text{atten,H}_2O}\) are the macroscopic attenuation coefficients (\(\text{m}^{-1}\)) at the laser wavelength due to molecular and aerosol scattering, the total attenuation at the laser wavelength, and at the nitrogen Raman-shifted and water Raman-shifted wavelengths; and \(C_1\), \(C_2\), and \(C_3\) are the system coefficients that take into account the effective area of the telescope, the transmission efficiency of the optical train, and the detector quantum efficiency at each of the three wavelengths. The water vapor mixing ratio is the mass of water vapor divided by the mass of dry air in a given volume. Because nitrogen is in con-
stant proportion to dry air in the lower atmosphere, the mixing ratio can be written as

\[ q(r) = \frac{n_{\text{H}_2\text{O}}(r)M_{\text{H}_2\text{O}}}{n_{N_2}(r)M_{N_2}} \times (\% \text{ nitrogen content}), \]  

(4)

where \( M_{\text{H}_2\text{O}} \) and \( M_{N_2} \) are the molecular weights of water and nitrogen, respectively. One can use (2), (3), and (4) to determine the mixing ratio at any distance from the lidar from the following expression:

\[ q(r) = \frac{P_{\text{H}_2\text{O}}(r)}{P_{N_2}(r)} \left( \frac{C_s \sigma_{n,\text{H}_2\text{O}} M_{N_2}}{C_s \sigma_{n,N_2} M_{\text{H}_2\text{O}} (\% \text{ N content})} \right) \times \exp \left( \int_0^r [\alpha_{\text{att},\text{H}_2\text{O}}(r') - \alpha_{\text{att},N_2}(r')] \, dr' \right). \]  

(5)

Thus the water vapor mixing ratio at any range is given by the ratio of the magnitude of the signal in the water vapor channel to the magnitude of the signal in the nitrogen channel, a multiplicative constant, and a correction due to difference in attenuation between the nitrogen-shifted and water vapor-shifted wavelengths.

The multiplicative constant was determined by comparison of the lidar signal with radiosondes released from the ships. Over land, the constant can be accurately determined by aiming the lidar horizontally and comparing to calibrated water vapor point sensors. This method was not possible aboard ship. While there are known problems with radiosonde calibration (e.g., Ferrare et al. 1995; Connell and Miller 1995; Melfi et al. 1989), efforts were made to mitigate their effects on the lidar calibration. For example, each radiosonde was tested for accuracy prior to launch and the altitudes used for calibration were limited to those with high relative humidity where the calibrations for the hygrometers are reliable to a few percent.

When using a wavelength of 248 nm, the difference in the attenuation at 263 nm (from nitrogen) and 273 nm (from water vapor) is significant, especially at long ranges (Vaughn et al. 1988). This is due to the effects of molecular scattering, which varies as \( 1/\lambda^4 \) (Teillet 1990); aerosol scattering, which varies as \( 1/\lambda \) over short wavelength differences (Whiteman et al. 1992); and ozone attenuation, which varies as the ozone concentration (Trakhovsky 1985). At a wavelength of 351 nm, similar corrections are required but are smaller in magnitude and do not include an ozone correction.

Data were processed in a manner similar to that described by Whiteman et al. (1992). The differential attenuation at the various wavelengths due to ozone, molecular, and aerosol scattering was derived from several sources. The ozone concentration was obtained from ship-launched ozonesondes and averaged to provide a day average and a night average concentration with height. Cross sections for ozone attenuation as a function of wavelength were obtained from Trakhovsky (1985). The differential attenuation due to molecular scattering is calculated from molecular densities derived from the U.S. Standard Atmosphere 1976 (NOAA 1976) and corrected with the measured pressure aboard ship. The Rayleigh attenuation model was developed by Teillet (1990). Estimates of the differential attenuation due to particulate scattering is based upon the standard atmosphere compiled by Elterman (1964).

A dead time correction was applied to the photon-counting data in the manner described by Knoll (1979) for a paralysable detector. The dead time correction is required to account for those photons that arrive during the time required for the scalar to record a previous photon (about 9 ns). When recording the first photon, the scalar is effectively “dead” or incapable of recording the second photon. At long distances from the lidar this correction is negligible. Nearer the lidar (from 500 m to 1 km), when high counting rates occur, the correction can be as high as 15%. A detailed discussion of this effect and the necessary corrections can be found in Funck (1986) and Donovan et al. (1993).

Figure 1 is a comparison of lidar and radiosonde measured water vapor mixing ratio for seven different radiosonde flights that were not used as calibration references. The correlation coefficient \( r^2 \) is 0.99 with a regression derived slope of 1.00. The root-mean-square
(rms) deviation from the regression is 0.34 g kg⁻¹. However, this value can be misleading. A detailed comparison of radiosonde and lidar far-field measurements taken nearly simultaneously (the lidar scan requires approximately 10 min, while the radiosonde can take well over 30 min to achieve 10 km altitude) is illustrated in Fig. 2. The lesser range resolution of the radiosonde and the long time constant of the capacitance hygrometer lead to significant differences between the two instruments when there are large gradients (e.g., at 1000, 2000, 3200, and 4700 m in Fig. 2). Thus while it can be said qualitatively that both instruments agree on the vertical distribution of water vapor, individual measurements (such as those at 2000 and 2250 m) may differ by as much as 2 g kg⁻¹. An analysis of some of these deviations can be found in Cooper et al. (1996).

Signals from laser pulses are summed to increase the statistical significance of the measurements and reduce the effects of system noise. Longer summing times (5 to 10 min) enable the system to achieve measurements at longer distances. However, shorter summing times enable measurements to be taken that allow visualization of discrete atmospheric structures, albeit at much shorter ranges. This capability has been used over land to study the surface atmosphere interface (Eichinger et al. 1993a; Cooper et al. 1992). Figure 3 is an example of such a vertical profile in which convective plumes of moist air are rising through the mixed layer.

**b. Raman lidar uncertainty determination**

The determination of the measurement uncertainty begins with the uncertainty in the analog signal level from a PMT (Inaba and Kobayasi 1972):

\[
\frac{\delta P(r)}{P(r)} = \frac{n(pr)^{1/2}}{[n + 2(n_s + n_d)]^{1/2}},
\]

where \( n \) is the number of summed pulses, \( p \) is the number of photoelectrons per unit time, \( r \) is the sampling time interval, \( n_s \) is the number of photoelectrons due to background light, and \( n_d \) is the number of effective photoelectrons due to dark current in the PMT. The wavelengths used in this system are in the region of the solar spectrum where ozone strongly attenuates the incoming light; thus the background photon count within a 3-nm window is negligible. The operating voltage is limited so that the sum of the dark current and background light is limited to less than a microvolt in comparison to a peak signal strength of 0.5 V, thus making the effects of \( n_s \) and \( n_d \) negligible during a 500-ns time period. The digitization of the signal provides a major source of uncertainty at longer ranges. The 8-bit digitizer registers changes in signal level with a resolution of 2 mV with a peak input of 0.5 V. Since the signal falls off as \( 1/r^2 \), the digitizer resolution becomes more significant at longer ranges. Other sources of uncertainty include the photomultipliers and bases, which have been selected to provide linearity to 1%. Using a least squares fit to a calibrated reference, the calibration constant can be determined to less than 2% (the use of a radiosonde as a reference may lead to significantly higher uncertainty; Ferrare et al. 1995; Connell and Miller 1995), while the uncertainty due to correction for differential attenuation is estimated to be less than 1% (out of a total correction of 12%–15%). Combining these independent sources of uncertainty, the root-sum-squared uncertainty in the mixing ratio can be written as (Bevington and Robinson 1992):

\[
\frac{\delta q(r)}{q(r)} = \left[ \frac{1}{n_{H_2O}pT} + \frac{1}{n_{N_2}pT} + \frac{(1 \text{ mV})^2}{P_{H_2O}(r)} + \frac{(1 \text{ mV})^2}{P_{N_2}(r)} + \left( \frac{\delta C}{C} \right)^2 + \left( \frac{\delta A}{A} \right)^2 + L_{H_2O}^2 + L_{N_2}^2 + \delta T^2 \right]^{1/2},
\]

where \( P_{H_2O}(r) \) and \( P_{N_2}(r) \) are the signal levels in volts, \( C \) is the calibration constant, \( A \) is the correction for attenuation, \( L_{H_2O} \) and \( L_{N_2} \) are contribution of PMT non-linearity to the uncertainty, and \( \delta T \) is the contribution of the uncertainty due to changes in the return signal due to temperature. Because of the first four terms, the uncertainty in the mixing ratio values are a strong function of distance from the lidar. For a midrange distance, the uncertainty is approximately 3.6%. This is consistent with the comparisons of lidar and calibrated references over land surfaces along horizontal paths. As noted above, the use of radiosondes as calibration references may lead to much higher uncertainties. In this case, the uncertainty in the radiosonde values dominates the other terms in estimating the total uncertainty.
c. Atmospheric transmission by Raman lidar

The Raman lidar has the advantage of having an additional signal from molecular nitrogen in the atmosphere as well as the elastic signal. The backscatter coefficient from nitrogen is proportional to the nitrogen density with altitude (which is known or can be calculated), while the attenuation coefficient is nearly the same as that of the elastic signal. Thus, the nitrogen signal can be inverted to give the extinction (Ferrare et al. 1992; Ansmann et al. 1991; Ansmann et al. 1992; Mitev et al. 1992) as

\[
\alpha_{\text{aerosol}}(r) = \frac{d}{dr} \ln \left( \frac{N(r)}{P(r)r^2} \right) - \alpha_{\text{molecular}}(r) - \alpha_{\text{molecular}}(r) - \alpha_{\text{molecular}}(r)
\]

\[1 + \left( \frac{\lambda_{\text{laser}}}{\lambda_{\text{Raman}}} \right)^k\]

(8)

where \( r \) represents not only the range from the lidar but the altitude as well. The subscript laser refers to the attenuation at the laser wavelength and the subscript Raman to the Raman-shifted wavelength. The \( \lambda_{\text{laser}}/\lambda_{\text{Raman}} \) term corrects for the small difference in aerosol attenuation between the elastic and Raman wavelengths. The constant \( k \) is normally taken as 1.0 (e.g., Klett 1985; Whiteman et al. 1992). The \( N(r) \) is the total molecular number density at altitude \( r \) (the atmosphere is assumed to be entirely nitrogen). The attenuation coefficients for molecules are well known and can be found from Rayleigh scattering theory (Measures 1984).

3. Raman lidar instrument description

The current configuration of the instrument is described in Table 1. The lidar is mounted in two major pallets and a single auxiliary electronics rack housed in an electronics enclosure. The main pallet contains the laser, its power supply, the telescope, and the scanner in an aluminum framework (Fig. 4). The second pallet contains the laser chiller, a cryogenic gas filtration system, a transformer to convert ship power at 480 V to
TABLE 1. Lidar operational parameters.

<table>
<thead>
<tr>
<th>Day</th>
<th>Night</th>
</tr>
</thead>
<tbody>
<tr>
<td>Output wavelength</td>
<td>248 nm</td>
</tr>
<tr>
<td>Return signal–elastic</td>
<td>248 nm</td>
</tr>
<tr>
<td>N₂</td>
<td>263 nm</td>
</tr>
<tr>
<td>H₂O</td>
<td>273 nm</td>
</tr>
<tr>
<td>Energy per pulse</td>
<td>0.40 J</td>
</tr>
<tr>
<td>Max repetition rate</td>
<td>200 Hz</td>
</tr>
<tr>
<td>Beam divergence</td>
<td>0.4 mrad</td>
</tr>
<tr>
<td>Telescope</td>
<td>61 cm, f/8, Cassegrain</td>
</tr>
<tr>
<td>Detector</td>
<td>Side-on PMT, Hamamatsu</td>
</tr>
<tr>
<td>Filters</td>
<td>3 nm wide, 8%–10% transmission</td>
</tr>
<tr>
<td>Rejection</td>
<td>10⁴ at elastic wavelengths</td>
</tr>
<tr>
<td>Digitizers</td>
<td>Signatec, DASP100, 100 MHz, 8 bit</td>
</tr>
<tr>
<td>Computer</td>
<td>Industrial Computer Source, Neptune 90-MHz 80586</td>
</tr>
<tr>
<td>Photon counters</td>
<td>DSP 4101, 500-ns summation</td>
</tr>
<tr>
<td>Range resolution</td>
<td>Near field–1.5 m</td>
</tr>
<tr>
<td>Scanner angular resolution</td>
<td>0.003°</td>
</tr>
</tbody>
</table>

three-phased 208 V, and limited power conditioning to provide stable 110-V power for the computer and data collection electronics. A standard electronics rack is used to house all the electronics involved in command and control.

A Lambda Physics LPX 220 excimer laser was used as the laser source. The laser beam is aligned with turning mirrors so that it is coaxial to the optical axis of a 61-cm-diameter f/8 UV-optimized Cassegrain telescope. The laser beam is expanded by approximately five times with a combination of a 52-mm-diameter diverging lens and a 203-mm-diameter collimating lens mounted between the turning mirrors and in front of the secondary mirror of the receiver telescope, respectively. A 920 mm by 610 mm flat mirror is mounted at 45° to the receiver telescope, which is able to turn through 360°, allowing scanning in one plane. A 14-bit absolute optical encoder is mounted on the turntable with up to 0.03° resolution. The lidar is equipped with four pendulum sensors; two mounted to the frame and two mounted on the scanner. The two pendulums on the scanner are used in feedback loops with the motor controller to actively stabilize the scanner in order to correct for ship roll. The other two pendulums (roll and pitch) are automatically polled at the beginning and end of each scan along a given line of sight. These give the position of the ship with respect to the horizontal, and, when combined with the scanner encoder angle (the angle of the scanner with respect to the ship), allow an estimate to be made of the effectiveness of the stabilization system, and potential corrections applied if necessary. Since pitch of the ship (an average of 2° with excursions to 5°) was relatively small, compared to roll (an average of 5° with excursions greater than 10°), no attempt was made to correct for the pitch. A second 1000-mm flat mirror is mounted on top of the lidar on a motor-driven bracket that rotates the mirror ±22° limited by the size of the mirror) off the optical axis, which allows for elevation scans. The bracket assembly is mounted on a 26-in. ring gear orthogonal to the optical axis that allows for 360° scans in the azimuthal direction. The entire turret is mounted on a set of slides that allows the scanner to be pushed out of the optical path so that vertical profiles can be acquired. In addition, 14-bit optical encoders are mounted on the bracket and ring gear with a maximum of 0.003° resolution. Figure 5 shows the current configuration with the scanner.

a. Detector optical train

Behind the telescope, three PMTs are mounted at equal distances from the prime focus of the telescope and at right angles to the light emerging from the telescope (Fig. 6). Dielectric mirrors are used to reflect approximately 3% of the collected light to the elastic channel PMT; 9% of the light at 263 nm to the nitrogen channel PMT; and the remaining light to the water vapor channel PMT. This approximately equalizes the number of photons impacting nitrogen and water vapor PMTs. Mechanical extensions were added to the PMT bases before the filter holders in elastic and nitrogen channels to equalize the optical pathlengths. Between the mirror and each PMT are interference filters of the appropriate

![Fig. 4. A schematic of the primary Raman lidar pallet. This pallet contains the telescope, laser, and scanning mirror.](image-url)
wavelength and a quartz lens. The focal length of the lens is chosen so as to focus the telescope secondary onto the active area of each PMT. Each of the interference filters has a bandpass of 3 nm and is well blocked (from $10^{-4}$ to $10^{-6}$) at the other two wavelengths as well as rejecting visible light. The strongest blocking is at the laser output wavelengths. The danger of bleed-through from the 248-nm channel is strongest in the 263-nm (nitrogen) channel where two filters in series provide blocking approaching $10^{-10}$. The total transmission of the filters ranges from 8% to 12%.

While the rotational fine structure of the Raman backscatter signal is highly temperature dependent, the integrated total of the Raman spectrum is independent of temperature. Thus the width of the filters (3 nm) is chosen to completely cover the Raman signal, precluding any significant temperature sensitivity. Contributions to the Raman signal from the O, Q, and S branches cover a band of less than 0.95 nm in the 263-nm part of the spectrum and 0.9 nm in the 273-nm part of the spectrum. With a 3-nm-wide filter arrangement, the resulting error due to temperature effects should be less than 0.2% (Whiteman et al. 1993). The resulting signals from the three PMT channels are fed to a PC-based digitizer data collection system, which digitizes the raw signal in the near field and counts photons in the far field.

The lidar system behavior is routinely checked by using 263-nm (or 382 nm at night) filters (detecting the atmospheric nitrogen density) in front of all of the PMTs simultaneously and examining the resultant signal. The signal from each of the three detectors should be identical. Variations in these signals are used as a diagnostic and are indicative of problems with changing PMT fields of view, amplification nonlinearities, drift of the high voltage power supply, or variations in the photocathode efficiency in any of the PMTs. Normalizing these signals allows corrections to be made to the different PMTs.

b. Photomultiplier tubes

Special consideration has been given to the electrical characteristics of the PMT bases. Increased capacitance has been added across the last five dynodes to minimize voltage drops due to high photocurrents. The maintenance of constant voltage across the dynodes is the major factor in achieving approximately 1% signal linearity with less than 0.5% undershoot. This was checked through the use of light pulses of known amplitude and duration. To increase the dynamic range of the system, the current mode signal is taken from the last dynode in the chain, while the photon-counting signal is taken from the anode (see Fig. 7) at the end of the PMT amplification chain. The PMTs were chosen to have sufficient gain to produce a 0.5-V peak analog signal without further amplification. Because of the potential difficulties with noise and nonlinearity with external amplifiers, this capability was considered desirable.

The PMTs used during the day are R166UH solar-blind tubes made by Hamamatsu. The tubes used at night are R1527 bialkali tubes also made by Hamamatsu. These are opaque photocathode side-on tubes with high quantum efficiency. However, the size of the active area (1.0 cm by 1.0 cm) is only slightly larger than the size of the image of the laser beam in the near field (the region between 10 and 250 m from the lidar). This causes the FOV of each PMT to be slightly different and necessitates the use of an FOV corrector in the near field (less than 150 m), which is derived from the periodic use of 263-nm or 382-nm filters in front of all of the PMTs (Grant 1991). Figure 8 is an example of an FOV correction function. In addition, these types of PMTs, with large interstage gain, do not allow for clear resolution of noise pulses from single photon pulses.
Thus, in the photon counting mode, a discriminator setting that eliminates the noise pulses also eliminates a large number of true photon pulses, reducing the overall efficiency of the system. PMTs with lower interstage gain and an increased number of stages would be more appropriate for this use.

c. Data acquisition system

The signal from the anode of each of the PMTs is fed directly to on 8-bit, 100-MHz Signatec DASP100 digitizer (see Table 1). The signal from the last dynode of the PMT is amplified using a 1.1-GHz Comlinear amplifier, sent to a 300-MHz Phillips Scientific Model 708 discriminator, and then to a DSP Model 4101 multichannel scalar for photon counting (Table 1). The system is PC based and controlled by an Intel 80586 microprocessor. Data acquisition can be performed at rates up to a maximum of 200 Hz when taking 1 kbyte of data in each of three near-field channels and 256 bytes of data from the three far-field channels.

A Keithley DAS1601 multipurpose board in the computer is used to read the position encoder on the scanner, and through A/D converters on the board, measures the ambient temperature, humidity, pressure, and ship orientation (from the pendulums). This board also generates the pulses, which are used to trigger the laser, allowing the operator to control the on/off state of the laser as well as the firing rate. The computer also houses a motor controller board to send commands to the motor controller as appropriate.

A recent feature added to the system is a real-time data display system that displays the data as it is being taken on a second display screen. While any of the datasets currently being taken can be displayed, normally, the near-field water vapor concentration is displayed in two dimensions. This feature is useful in identifying malfunctions in the system as well as identifying the times and locations where processes of interest occur.

4. Data display formats

The utility of the lidar for atmospheric science studies in the boundary layer is derived from the ability to map the water vapor mixing ratio in multiple dimensions by access to scanning options available to the user from the two coaxial scanners. For shipboard operation, vertical soundings can be acquired with active control of the lower scanner that creates consistent high quality data regardless of ship roll. If the platform is stable, then multidimensional scans can be acquired to quantify the three-dimensional structure of the ABL. The following is a short description of the scanning options.

a. Vertical sounding

A vertical sounding file is created by setting the lower scanner to the vertical position and pushing the upper scanner out of the optical path; a user-selected number of laser pulses are fired, averaged together in the computer, and saved to disk. The typical operation acquires the current from the PMTs and is digitized at 100 MHz creating a 1.5-m range-resolved two-dimensional profile of the atmosphere analogous to what a high-resolution radiosonde would observe in the ABL (see Fig. 9). Range is a function of the number of averaged laser pulses and is inversely proportional to temporal resolution. Thus, the user requirements involved in the study of the atmosphere must make a trade-off between range and temporal resolution. If deep sounding of the troposphere is desired, then the multichannel scalars are also employed for photon counting, probing the lower troposphere during the day and up to 12 km at night.

b. Time series

A time series of range-resolved water vapor mixing ratio is the compilation of multiple vertical soundings acquired sequentially in time. The fundamental time resolution of the series is the time required to create individual profiles, which can be as short as 1/200 of a second (the maximum pulse rate of the laser is 200 Hz) or as long as the computer memory can accommodate the data (up to 45 min in the current system). A time series can be acquired in any direction the user desires from parallel to the surface to vertical by orienting the upper and/or lower scanners to any desired geometry.

c. Vertical scans

Vertical two-dimensional scans show the water vapor content in a plane perpendicular to the surface. These can be generated from the upper scanner by locking the lower scanner and maintaining constant azimuth on the upper scanner turret while varying the elevation angle of the upper mirror. The upper scanner acquires user-defined short time series data at a given elevation, after the short time series is acquired the scanner elevation is changed incrementally by a stepping motor and an additional short time series is acquired. The number of laser pulses per short time series, direction of scan, number of angle increments in a series, and incremental angle change are all user selectable. A vertical image (in polar projection) is generated by compiling multiple incrementally changed elevation angle time series, after the scan is created the mirror is repositioned to a user-defined azimuth and elevation for another scan sequence. The time required to make a vertical scan is a function of the number of laser pulses averaged into a single time series line of sight, the range of angle swept by the scanner and the incremental angular change, thus vertical scans can be acquired in as little as 15 s for modest areas. Figure 10 is an example of a vertical scan.
d. Horizontal scans

Horizontal scans parallel to the surface are created the same way as the vertical scans, except that the elevation angle on the upper scanner is held constant and an incremental change is made to the ring gear, thus affecting the azimuth. Multiple lines of sight as a function of azimuth are compiled together to form a horizontal image of the surface in question. Figure 11 is an example of a horizontal scan.

5. Data analysis

The data collected by the Raman water vapor lidar are used to evaluate the scalar and flux properties of the atmospheric boundary layer (ABL) and the troposphere. This section will briefly describe some of the analysis methods used for the CEPEX and CSP programs. More detailed papers on the structure of the ABL, evaporative fluxes, and boundary layer processes are either in press or forthcoming.

a. ABL structure

The traditional approach to characterizing the ABL is by use of one-dimensional sensors, such as radiosondes or temperature–humidity probes that measure the scalar properties (humidity, temperature, and wind) of the atmosphere over time. The size and direction of the values and gradients are used to determine the structure and behavior of the ABL, such as the depth of the surface and mixed layers, and the top of the ABL. Time-averaged lidar data are the direct analogs to the radiosonde, and are used here as an additional tools to understand the ABL properties (Fig. 9). Note that in the first several hundred meters the lidar and radiosonde are in good agreement; above this region the two sensors diverge at the inversion that divides the ABL from the lower tropospheric free atmosphere (Stull 1988). The inversion is more pronounced in the lidar data, as well as the variability. The cause for the divergence between the radiosonde and lidar can be better explained once we expand the temporal resolution of the lidar profile from a 30-min average to less than 10 s per line of sight (Fig. 3). The lidar data suggests that about every 200 s a moisture-rich microscale convective plume moves from somewhere in the surface layer to the top of the ABL at approximately 650 m. These plumes, and the regions between the plumes, cause the top of the ABL to be somewhat variable and show that the ABL is dominated by turbulent intermittent structures that occur on space and time scales too small for the radiosonde to sense. The probability that a radiosonde will move through the appropriate combination of plume and interplume region to characterize adequately the structural properties of the ABL is small (Cooper et al. 1996). The range-resolved time series shown in Fig. 3 indicates that these plumes have their origin somewhere in the surface layer, which suggests that the lower portion of the ABL might not follow the simple scaling laws commonly used in numerical tools such as similarity theory (Brutsaert 1982). Due to the limitations of the lidar during CEPEX in 1993, it was not possible to directly probe the first 50 to 100 m of the surface–atmosphere interface. With the addition of a turret-type scanner, three-dimensional images are now possible so that the origin and structure of the surface–atmosphere interface can be observed and evaluated. We returned to the central equatorial Pacific in 1996 with the goal of identifying the source of the plumes previously observed. During the CSP voyage, the Raman lidar made hundreds of vertical and horizontal scans covering the first 150 m of the marine atmospheric surface layer, such as the lidar image shown in Fig. 10. The image shows that very small plumes on the order of 10 to 20 m begin to grow from the surface–atmosphere interface and appear to circulate in a region 30 to 40 m above the ocean. Further analysis indicates that periodically some of these plumes penetrate in the mixed layer at about 120 m (Cooper et al. 1997). The features seen in Fig. 10 suggest eddy circulation patterns previously theorized by dynamical models (Williams and Hacker 1993).
horizontal scans acquired soon after the vertical scans show the spatial distribution of these plumes and their cross-section dimensions (Fig. 11). The plumes from this image are approximately 25 m in diameter, with similarly sized nearby dry regions, which are presumed to be regions of upwelling and downwelling. Ongoing analysis of the lidar data incorporates large-eddy simulation and direct numerical simulation of the ABL that will hopefully give improved insight into the relationship between the boundary layer and both the troposphere and surface.

The evolution of the scanning Raman lidar has been directed to provide the types of data with spatial and temporal resolution that address questions about the behavior of the surface atmosphere interface. These questions address the nature of spatial variability and intermittent microscale convective transport in the ABL. Future design developments of the lidar will concentrate upon the measurement of other scalars, such as temperature or gases such as CO₂.

b. Evaporative fluxes

Evaporative fluxes derived from the lidar provide a check on the values obtained from other methods aboard ship. The lidar can be used to obtain the water vapor flux using the Monin–Obukhov similarity theory (Eichinger et al. 1993b). In this theory, the relationship between the water vapor concentration at the surface and some height \(z\) is

\[
q_s - q(z) = \frac{LE}{0.4L_u \rho} \ln \left( \frac{z - d_0}{z_\infty} \right) + \psi_c \left( \frac{z - d_0}{L} \right),
\]

where \(d_0\) is the displacement height, \(L\) is the Monin–Obukhov length, \(q_s\) is the surface water vapor concentration, \(q(z)\) is the water vapor concentration at height \(z\), \(u_\ast\) is the friction velocity, \(LE\) is the evaporative flux, \(L_e\) is the latent heat of vaporization, \(\rho\) is the air density, \(z_\infty\) is the roughness length for water vapor, and \(\Psi_c\) is the diabatic correction for water vapor. If the vertical profile of the water vapor concentration in the atmosphere is measured, and a least squares fit can be made using (9) to determine the value of \(LE/(L_u \rho)\), then the flux can be determined from this value if the friction velocity is known. While this technique will only give the latent energy flux, this is the dominant mass and energy transfer mechanism from the ocean to the atmosphere. Figure 12 is a comparison of evaporative fluxes obtained through this method and from the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment specified drag coefficient method (Fairall et al. 1996) using fixed-point measurements taken on the R/V *Vickers* (Brutsaert 1982) aboard ship. Specific details on the calculations are presented in Eichinger et al. (1998, manuscript submitted to *J. Climate*).

6. Comparison to other water vapor lidars

The choice of a Raman lidar to scan and determine spatial water vapor measurements is unusual. However, examination of the choice leads one to the conclusion that it is the best option to pursue for the purpose of making high temporal and spatial observations for the purpose of studying surface–atmosphere interactions and transport in the boundary layer.

The main competitor to the use of Raman systems is the differential absorption lidar method (DIAL). In this method, two different, but closely spaced, wavelengths of laser light are used. One wavelength is absorbed by the molecule of interest, one is not. Because the two wavelengths are nearly the same, their atmospheric transmission properties are similar, except for absorption by the molecule of interest. Thus, the differences
in the returning light intensity can be used to determine the range-resolved concentrations. The concentration is proportional to the product of the molecular concentration at a given range, the distance over which the attenuation occurs (which is the range resolution), and the absorption coefficient of the line used. Many of these types of lidars have been or are in use (e.g., Higdon et al. 1994; Ehret et al. 1993; Hinkley 1976).

DIAL systems have a number of inherent advantages over Raman lidars. They are generally smaller, lighter, and use considerably less energy than Raman systems. They are especially well suited for use in aircraft because they rely on atmospheric backscatter which increases with range in a downward-looking system, and thus partially compensates for the $1/r^2$ decrease in signal with range. Because of the strong atmospheric attenuation of UV and near-UV light, DIAL systems in the near-infrared are better suited for deep atmospheric sounding. Both systems must average the return from large numbers of laser pulses to obtain data with sufficient statistical significance. However, high energy (250–400 mJ per pulse) excimer lasers with pulse rates of 200 Hz are common. In contrast, the materials commonly used in water vapor DIAL systems (flashlamp or Nd:YAG-pumped dyes, alexandrite, and Ti:sapphire lasers) have inherent limitations on the pulse rates, which limit high-power systems to 10 to 50 Hz. Thus, Raman systems have the potential for better temporal resolution. While the range resolution of DIAL systems is dependent upon the strength of the absorption feature used, existing systems have range resolutions on the order of 15 to 500 m as compared to 1.5 m for the Raman technique.

Because of the limitations on pulse rates and range resolution, it is our belief that DIAL systems will not have in the near future the spatial or temporal resolution necessary to capture atmospheric structures with sizes on the order of 5 to 100 m in size and with lifetimes of seconds to a few minutes. Our investigations show that the processes involved in surface–atmosphere interactions and transport in the boundary layer require spatial and temporal resolutions on this scale. This is not to say that a DIAL system with this capability cannot or will not be built. But in the context of adapting ex-
Fig. 11. Horizontal lidar scan from the CSP experiment at 0421 UTC (1821 local time) 26 Mar 1996 acquired 10 m from the sea surface, showing the spatial distribution of plume features.

Fig. 12. Comparison of evaporative flux measurements made aboard the R/V Vickers from the lidar and using conventional drag coefficient methods. The line drawn is a best fit with a slope of 0.91 and a correlation coefficient, $r^2$, of 0.74. The rms difference between the lidar-derived fluxes and drag coefficient fluxes is 16 W m$^{-2}$.

7. Summary

The scanning Raman water vapor lidar has been successful in acquiring data, which is being used to characterize the lower marine troposphere from aboard ships. The lidar operated nearly continuously on the two experiments during both CEPEX and CSP programs (as well as other terrestrial experiments), in conjunction with many other shipborne and airborne instruments. The internal, online diagnostic hardware and software made it possible to acquire high-resolution water vapor data in a relatively difficult environment—the open Pacific Ocean. The water vapor lidar measurements were corrected for variations in laser power output; field of view; and attenuation from ozone, molecular, and aerosol constituents. The Raman lidar collected vertical pro-
files of water vapor up to 10 km, with spatial resolution of 1.5 m in the first 2 km and 75 m in the remaining 8 km. The temporal resolution of the Raman lidar was between 30 s and 30 min. These lidar provided a wide range of information on the ABL and lower troposphere with spatial and temporal resolutions currently unavailable by any other method. They are providing information to address questions concerning climate feedback mechanisms, the spatial variability of the lower boundary layer, and the nature of transport processes near the surface. The data that the Raman lidar provided will be used to better understand the role of the lower troposphere in ocean–atmosphere interactions.

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