Observatory for Atmosphere Space Interaction Studies (OASIS)

Editors:

Stephen Hall – hall37@illinois.edu

Gary Swenson – swenson1@illinois.edu
Table of Contents

Expected Altitude Coverage and Measurement Accuracies of the Large Aperture Lidar Observatory ..................................................1
Rayleigh Lidars ...........................................................................................................39
The State-of-the-art Na lidars for MLT dynamics studies ...............................47
Fe Doppler and Boltzmann LIDAR for Exploring Atmosphere and Space ...........................................................................................................59
He Lidar for LALO ........................................................................................................67
Influx of cosmic dust and how it affects the structure and evolution of the Earth’s atmosphere ...........................................................................................................73
Incoherent Scatter Radar ..............................................................................................84
Meteor Radar ................................................................................................................89
Coherent Scatter Radars ...............................................................................................92
In-Situ Observations .....................................................................................................96
Global Navigation Satellite Systems ...........................................................................102
Correlative and in-situ measurements complementing the Large Aperture Lidar ...........................................................................................................108
Exospheric Studies ........................................................................................................114
Optical/lidar Workshop Report ...................................................................................119
Signal-to-Noise Ratio Calculations for Imaging Bi-Static Rayleigh Lidar ...........................................................................................................126
Expected Altitude Coverage and Measurement Accuracies of the Large Atmospheric Lidar Observatory

Chester S. Gardner
Department of Electrical and Computer Engineering
University of Illinois at Urbana-Champaign
Urbana, Illinois

Abstract
The Large Atmospheric Lidar Observatory (LALO) is envisioned to include a transportable 100 m² telescope array, along with Rayleigh, Na, Fe and He(2³S) Doppler lidars and other correlative instrumentation. Collectively, these four lidars cover the altitudes ranges of 30-150 km and 300-1000 km. At night the signal-to-noise power ratios (SNRs) would be about 1200 times larger and the measurement errors would be about 35 times smaller than the more sophisticated middle and upper atmosphere lidars in operation today. During the day the SNRs would be about 400 times larger and the measurement errors would be about 20 times smaller. At night the Rayleigh, Na and Fe lidars are capable of measuring horizontal and vertical wind and temperature profiles with accuracies better than 6 m/s, 1.5 m/s and 7 K, respectively, between 30 and 150 km at a vertical resolution of 500 m and a temporal resolution of 2.5 min. At this resolution, the signal-to-noise ratios are so large that photon noise is negligible for measurements of gravity wave vertical heat and constituent fluxes between 30 and 150 km and for measurements of gravity wave momentum fluxes between 30 and 135 km. Also at this resolution, photon noise is negligible for measurements of the eddy heat flux and eddy thermal diffusivity profiles (k_H) between 30 and 60 km and between 80 and 105 km. In these altitude ranges the k_H profile can be measured with an accuracy of 4 m²/s or better. At the lower resolution of 2.5 km and 2 h, the k_H profile can be measured with an accuracy of better than 20 m²/s between 30 and 130 km. Above 80 km where both eddy constituent and heat transport are measured, the LALO Na and Fe lidars can also measure the turbulent Prandtl number (Pr) at high spatial and temporal resolution. The He(2³S) lidar is capable of measuring horizontal and vertical wind and temperature profiles with accuracies better than 35 m/s, 8 m/s and 7.5 K, respectively, between 300 and 1000 km at resolutions of 25 km and 5 min.

1. Introduction
The development and refinement of sophisticated remote sensing technologies during the past five decades have contributed enormously to our knowledge of the atmosphere, especially the upper atmosphere above 30 km altitude. Major radar facilities, such as AMISR, Arecibo, EISCAT, Sondrestrom, Jicamarca, Millstone Hill and the MU Radar, have permitted researchers to study directly the ionized atmosphere with very high accuracy and resolution while enabling inferences of neutral gas properties and dynamics. At the time these facilities were commissioned, each represented a major step forward in observational capabilities. Today these radars continue to play central roles in many ionospheric studies.

Lidar technology has enjoyed a similar renaissance since the invention of the laser 50 years ago. The first lidars were built in the 1930s and 1940s using mechanically modulated searchlights. Today, modern laser-based systems are used to probe composition and structure of the neutral atmosphere from the troposphere into the lower thermosphere. The last two decades has been a period of substantial growth in lidar capabilities and applications, principally because of advances in critical areas of laser technology. Perhaps the most important of these has been the development of high-power, ultra-stable narrowband lasers, which are now being used in Doppler lidars for middle and upper atmosphere applications. Furthermore, robust tunable fiber
lasers are also being used in laser guide star applications for ground-based astronomical imaging and for sensing helium in the Earth’s thermosphere [Carlson et al., 2009].

While the recent advances in lidar technology have been impressive, the accuracy, resolution, sensitivity and range of many systems are still limited by signal levels. Experiments conducted during the past fifteen years using the 3-meter class telescopes at the Starfire Optical Range, NM and Haleakala, Maui (Maui/MALT Program) demonstrated clearly the substantial scientific advantages of employing large steerable lidar telescopes in combination with correlative radars, passive optical instruments and rocket probes to study the middle and upper atmosphere. In this paper we examine the anticipated measurement performance of the Large Atmospheric Lidar Observatory (LALO). The centerpiece of the observatory would be a 10-meter class telescope that would serve as the receiving system for several very large Rayleigh, Na, Fe and He(2^3S) lidar systems. By 10-meter class we mean a telescope with an effective aperture area comparable to a 10 meter diameter mirror (~100 m^2). It is envisioned that the facility might consist of a large fully steerable telescope (3-4 meters in diameter) in combination with a large array of smaller fix-pointed telescopes (e.g. 10x10 array of 1 meter telescopes) yielding a total collecting area of approximately 100 square meters. The observatory would also include an appropriate complement of other important instruments such as radars, imagers, spectrometers and perhaps in situ measurement capabilities using balloon and rocket probes.

In the following, we compute the expected signal-to-noise ratios (SNR) and analyze the measurement accuracies of modern Rayleigh, Na, Fe and He(2^3S) Doppler lidars when they are coupled with the 100 m^2 LALO telescope array. The calculations are summarized in the following sections, while the detailed error formulas and computations are described in the appendices.

2. Observatory Configuration

LALO is envisioned to be a versatile 5-beam, transportable system utilizing one beam fixed pointed at zenith and four beams that can be pointed at several zenith and azimuth angles. Nominally these four beams would be pointed θ degrees off-zenith (θ=30°, 60° and 0°) with azimuth angles of 0 degrees (north), 90 degrees (east), 180 degrees (south) and 270 degrees (west). The telescope array is designed so that sub-arrays of telescopes can be pointed in the same directions as the laser beams. The system is flexible so that for each beam direction, the laser power level and the aperture area of the telescope sub-array can be selected to optimize the observations for the scientific issue being investigated.

The line-of-sight (LOS) wind velocities observed by the lidar are

\[ V_z = w \]
\[ V_N = v \sin \theta + w \cos \theta \]
\[ V_S = -v \sin \theta + w \cos \theta \]
\[ V_E = u \sin \theta + w \cos \theta \]
\[ V_W = -u \sin \theta + w \cos \theta \]

where \( V_z, V_N, V_S, V_E \) and \( V_W \) are the LOS velocities observed by the zenith, north, south, east and west beams, respectively and \( w, u \) and \( v \) are the vertical, zonal and meridional wind components, respectively. The three geophysical wind components are derived from the LOS winds as follows.

\[ w = V_z \]
\[ u = \frac{V_E - w}{\tan \theta} = -\frac{V_W}{\sin \theta} + \frac{w}{\tan \theta} \]
\[ v = \frac{V_N - w}{\tan \theta} = -\frac{V_S}{\sin \theta} + \frac{w}{\tan \theta} \]
Because the vertical wind is dominated by gravity wave perturbations and the off-zenith beams probe different regions of the atmosphere, the vertical wind component observed on the off-zenith and zenith beams are mutually uncorrelated. Therefore, the horizontal wind measurements will be contaminated by the small vertical winds that contribute to the observed radial winds. This problem is not unique to LALO. It arises anytime upper atmosphere winds are measured from a single site. The velocity errors are given by

\[
\begin{align*}
\text{Var}(\Delta w) &= \text{Var}(\Delta V_z) \\
\text{Var}(\Delta u) &= \frac{\text{Var}(\Delta V_E)}{\sin^2 \theta} + \frac{\text{Var}(\Delta V_N)}{\sin^2 \theta} = \frac{\text{Var}(\Delta V_w)}{\tan^2 \theta} + \frac{\text{Var}(w)}{\tan^2 \theta}, \\
\text{Var}(\Delta v) &= \frac{\text{Var}(\Delta V_E)}{\tan^2 \theta} + \frac{\text{Var}(\Delta V_N)}{\tan^2 \theta} = \frac{\text{Var}(\Delta V_w)}{\tan^2 \theta} + \frac{\text{Var}(w)}{\tan^2 \theta},
\end{align*}
\]

where \(\Delta V_z, \Delta V_N, \Delta V_E\) and \(\Delta V_w\) are the lidar measurement errors associated with photon noise and other effects and \(\text{Var}(\cdot)\) denotes the variance of the parameter. Photon noise errors decrease with increasing signal level, which is proportional to the power-aperture product of the lidar beam. The zonal and meridional wind errors increase with increasing zenith angle because the range to a given altitude increases and the signal level decreases. However, the vertical wind contamination decreases with increasing zenith angle. For a large facility like LALO, vertical wind contamination is typically the major source of error for horizontal wind measurements.

LALO is designed to be operated at several fixed zenith and azimuth angles depending on the scientific objectives, viz. \(\theta=0\) degrees (full telescope and a single laser beam pointed at zenith, optimum for vertical wind, temperature and eddy flux observations), \(\theta=6\) degrees (2-, 4- or 5-beam configuration, optimum for momentum flux observations) and \(\theta=30\) degrees (1-, 2-, 4-, or 5-beam configuration, optimum for horizontal wind observations).

### 3. Winds, Temperatures and Densities

The atmospheric parameters can be written as the sum of the longer-term climatology (e.g. weekly or monthly means), medium time-scale perturbations associated with planetary waves, tides and gravity waves (i.e. periods ranging from several minutes to several days) and the short time-scale fluctuations caused by turbulence (i.e. periods ranging from seconds to several minutes).

\[
T(z,t) = T^0 + T_{GW}^r + T_{\text{Turb}}^r \\
w(z,t) = w^0 + w_{GW}^r + w_{\text{Turb}}^r \\
u(z,t) = u^0 + u_{GW}^r + u_{\text{Turb}}^r \\
v(z,t) = v^0 + v_{GW}^r + v_{\text{Turb}}^r \\
\rho_c(z,t) = \bar{\rho}_c + \rho_{CGW}^r + \rho_{\text{Turb}}^r
\]

The overbar denotes the long-term sample mean, the primes denote the fluctuations, subscript \(GW\) denotes the large-scale wave fluctuations and the subscript \(\text{Turb}\) denotes the small-scale turbulence fluctuations.

We focus on the more difficult gravity wave and turbulence observations, which require the highest resolutions. The shortest gravity wave vertical wavelengths and periods are about 1 km and 5 min. So for gravity wave observations, we assume the lidar resolutions are nominally \(\Delta z=500\) m and \(\Delta t=2.5\) min. For turbulence, the inner scale \(l_i\) varies from a few meters to perhaps \(100-200\) m [Lübken et al., 1993; Lübken, 1997], while the shortest temporal scales are on the order of \(2l_i / \sqrt{u_z^2 + v_z^2}\). They vary from a few seconds to a few tens of seconds, depending on the horizontal wind velocity. We are most interested in measuring the eddy heat and constituent fluxes and the associated eddy diffusivities. Thus, the resolution needs to be small enough to

3
capture the majority of the energy in the turbulence fluctuations. For turbulence observations, we assume the lidar resolutions are \(\Delta z=50\,\text{m}\) and \(\Delta t=10\,\text{s}\) (see Section 4).

The key specifications of the lasers for the Na, Fe, He and Rayleigh Doppler lidars are listed in Table 1. These specifications were provided by the authors of the lidar descriptions included in the Engineering and Technical Supplement. They reflect what has been or could be readily achieved with current technology. In each case we assume diode laser pumping of the laser rods or fibers and frequency multiplication or mixing to produce the required output wavelengths.

### Table 1. Key Laser and Receiver Sub-System Parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Rayleigh</th>
<th>Na</th>
<th>Fe</th>
<th>He(2^2S)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\lambda_{\text{Laser}}) - Optical Wavelength</td>
<td>532 nm</td>
<td>589 nm</td>
<td>372 nm</td>
<td>1083 nm</td>
</tr>
<tr>
<td>(P_{\text{Laser}}) - Average Power</td>
<td>50 W</td>
<td>10 W</td>
<td>10 W</td>
<td>20 W</td>
</tr>
<tr>
<td>(\sigma_{\text{Laser}}) - RMS Linewidth</td>
<td>50 MHz</td>
<td>50 MHz</td>
<td>15 MHz</td>
<td>10 MHz</td>
</tr>
<tr>
<td>(\Delta \sigma_{\text{Laser}}) - Uncorrected Linewidth Jitter</td>
<td>(\leq 500,\text{kHz})</td>
<td>(\leq 500,\text{kHz})</td>
<td>(\leq 500,\text{kHz})</td>
<td>(\leq 500,\text{kHz})</td>
</tr>
<tr>
<td>(\Delta f_{\text{Laser}}) - Laser Frequency Locking Jitter</td>
<td>NA</td>
<td>(\leq 100,\text{kHz})</td>
<td>(\leq 100,\text{kHz})</td>
<td>(\leq 100,\text{kHz})</td>
</tr>
<tr>
<td>(\Delta f_{\text{Chirp}}) - Uncorrected Chirp Jitter</td>
<td>(\leq 100,\text{kHz})</td>
<td>(\leq 100,\text{kHz})</td>
<td>(\leq 100,\text{kHz})</td>
<td>(\leq 100,\text{kHz})</td>
</tr>
<tr>
<td>(\Delta f_{\text{Filter}}) - Filter Frequency Locking Jitter</td>
<td>(\leq 500,\text{kHz})</td>
<td>NA</td>
<td>NA</td>
<td>NA</td>
</tr>
<tr>
<td>(R_{\text{Det}}) - Detector Dark Count Rate</td>
<td>(~100/s)</td>
<td>NA</td>
<td>NA</td>
<td>NA</td>
</tr>
<tr>
<td>(R_{\text{Laser}}) - Pulse Repetition Rate</td>
<td>750 pps</td>
<td>100 kHz</td>
<td>500 kHz</td>
<td>10 W</td>
</tr>
<tr>
<td>(\theta_{\text{Laser}}) - Beam Divergence (single TEM(_{00}) mode)</td>
<td>(\theta_{\text{Laser}} \leq 0.5\text{mrad})</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Measurement Accuracies and Fundamental Limits at High SNR:** For the optimized resonance fluorescence (RF) Doppler lidars (see Appendix A), measurements are made at three frequencies, viz. at the peak of the fluorescence line \(f_c\) and in the wings of the line at \(f_w = f_c \pm \delta f\). The errors include contributions from photon noise, laser tuning and linewidth errors and signal calibration errors. For nighttime observations, the total temperature, wind and constituent density measurement errors are

\[
\Delta T_{\text{RF}}^{\text{rms}} = \left[ \frac{3.78(1+\beta^{-1})}{\bar{N}_C} + \frac{2\Delta f_{\text{Chirp}}^2}{f_0^2} + 18.9 \frac{\sigma_{\text{Laser}}^2}{f_0^2} \frac{\Delta \sigma_{\text{Laser}}^2}{f_0^2} \right]^{1/2} T(z)
\]

\[
\Delta W_{\text{RF}}^{\text{rms}} = \left[ \frac{0.577(1+\beta^{-1})}{\bar{N}_C} + \frac{\Delta f_{\text{Laser}}^2}{f_0^2} + \frac{\Delta f_{\text{Chirp}}^2}{2f_0^2} + 3.72 \frac{\sigma_{\text{Laser}}^2}{f_0^2} \frac{\Delta \sigma_{\text{Laser}}^2}{f_0^2} \right]^{1/2} \lambda_C f_0
\]

\[
(\Delta \rho_C)^{\text{RF}}_{\text{rms}} = \left[ \frac{3.13(1+\beta^{-1})}{\bar{N}_C} + 10^{-4} \left(1 - 2\Delta t / T_i\right) \right]^{1/2} \rho_C(z)
\]

\[
\text{SNR} = \frac{\bar{N}_C^2}{(\bar{N}_C + \bar{N}_B + \bar{N}_D)} = \frac{\bar{N}_C}{(1+\beta^{-1})}
\]

\[
\beta = \frac{\bar{N}_C}{(\bar{N}_B + \bar{N}_D)}
\]

where \(\bar{N}_C\) is the expected fluorescence signal photon count when the laser is tuned to \(f_c\), \(\bar{N}_B\) is the expected background noise count, \(\bar{N}_D\) is the expected detector dark count and \(T_i\) is the inertial period. \(\beta\) is the ratio of the signal count to the sum of the background noise and dark
counts. The corresponding total errors for an optimized Rayleigh Doppler lidar that employs two stabilized edge filters in the receiver are (see Appendix B)

\[
\Delta T_{rms}^{Hod} = \left[ \frac{2.33(1+\beta^{-1})}{N_{Mol}} \right]^{1/2} T(z)
\]

\[
\Delta w_{rms}^{Ray} = \left[ \frac{1.17(1+\beta^{-1})}{N_{Mol}} + \frac{\Delta f_{\text{filter}}^2}{8 f_{\delta}^2} + \frac{\Delta f_{\text{Chirp}}^2}{4 f_{\delta}^2} \right]^{1/2} \lambda_{\text{Laser}} f_{\delta}
\]

\[
(\Delta \rho_{\text{Atm/o}})_{rms}^{Ray} = \left[ \frac{2.33(1+\beta^{-1})}{N_{Mol}} + 10^{-4}(1-2\Delta t/T_i) \right]^{1/2} \rho_{\text{Atm/o}}(z). \tag{6}
\]

\[
SNR = \frac{\bar{N}_{\text{Mol}}}{(\bar{N}_{\text{Mol}} + \bar{N}_B + \bar{N}_D)} = \frac{\bar{N}_{\text{Mol}}}{(1+\beta^{-1})}
\]

\[
\beta = \frac{\bar{N}_{\text{Mol}}}{(\bar{N}_B + \bar{N}_D)}
\]

\(\bar{N}_{\text{Mol}}\) is the expected molecular backscattered signal count. The errors associated with uncertainties in the laser tuning and line width, receiver filter tuning and signal calibration, represent the fundamental limits to measurement accuracy, which can be achieved at very high signal levels (large \(\bar{N}_c\) and \(\bar{N}_{\text{Mol}}\)). The fundamental limits (see Table 2) are about ±0.05 K, ±10 cm/s and ±1% (density). These fundamental errors are correlated throughout the measurement profiles while the photon noise errors in different range bins are uncorrelated.

**Table 2. Fundamental Limits of Measurement Accuracies @ High SNR**

<table>
<thead>
<tr>
<th>Lidar</th>
<th>(\Delta w)</th>
<th>(\Delta T/T)</th>
<th>(\Delta \rho/\rho)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Night</td>
<td>Day</td>
<td>Night</td>
</tr>
<tr>
<td>Na</td>
<td>7.9 cm/s</td>
<td>7.4 cm/s</td>
<td>2.2x10^{-4}</td>
</tr>
<tr>
<td>Fe</td>
<td>4.6 cm/s</td>
<td>1.7x10^{-4}</td>
<td>2.4x10^{-4}</td>
</tr>
<tr>
<td>He(2^S)</td>
<td>13 cm/s</td>
<td>5.5x10^{-5}</td>
<td>7.8x10^{-5}</td>
</tr>
<tr>
<td>Rayleigh</td>
<td>9.8 cm/s</td>
<td>0</td>
<td></td>
</tr>
</tbody>
</table>

Notice that at nighttime \(1<<\beta\) and the SNR is equal to the mean signal count which is proportional to power-aperture product \(P_{\text{Laser}} A_{\text{Tele}}\). Because of the very large aperture area of LALO (100 m^2) and higher laser power levels (5-7 times larger), the nighttime SNR for LALO would be about 1200 times larger, and the measurement errors about 35 times smaller, than existing lidar systems that employ much smaller telescopes (0.5-0.8 m^2) and lower laser power levels (5-7 times smaller).

**Detector Noise:** Since we would like to understand the observational limits for the optimum LALO configuration, we focus on nighttime observations during new moon. We assume the background noise is negligible, but we do include the effects of detector dark counts, which limit the performance at the highest altitudes where the signal is weak.

\[
\bar{N}_D = \frac{2}{c} R_{\text{Det}} R_{\text{Laser}} \Delta t \Delta z
\]

\[
\beta = \frac{\bar{N}_c}{\bar{N}_D} \text{ or } \frac{\bar{N}_{\text{Mol}}}{\bar{N}_D}
\]

**Beam Swinging Versus Multiple Beams:** Depending on the application, it may be necessary to point the lidar beam in several different directions. For example, to measure horizontal winds the beam must be pointed in at least two directions in order to measure both the zonal and meridional wind components. To measure momentum fluxes, four different beam directions are required.
(see Section 4). Measurement performance can be significantly different depending upon whether a single beam, which is pointed sequentially in multiple directions (beam swinging), or multiple beams are employed.

Let $P_{\text{Laser}} A_{\text{Tele}}$ denote the maximum power-aperture product of the lidar corresponding to a single beam. Let $k$ denote the number of different beam directions required to make a particular measurement. For beam swinging, the single beam must be time shared among the $k$ different beam directions so the effective power-aperture product for each beam direction is $P_{\text{Laser}} A_{\text{Tele}} / k$. If multiple beams are employed, then the aperture area for each beam is $A_{\text{Tele}} / k$ and the average laser power, which must be divided among the beams either by time sharing or splitting the laser power, is $P_{\text{Laser}} / k$ for each beam. Consequently, the effective power-aperture product per beam is $k$ times smaller for the multiple beam scenario.

$$\frac{(P_{\text{Laser}} A_{\text{Tele}})_{\text{eff}} / \text{Beam}}{k} = \begin{cases} P_{\text{Laser}} A_{\text{Tele}} / k & \text{Beam Swinging Configuration} \\ P_{\text{Laser}} A_{\text{Tele}} / k^2 & \text{Multiple Beam Configuration} \end{cases} \quad (8)$$

$k = \text{number of beams}$

Although the beam swinging approach yields larger signal levels and more precise measurements, it is more expensive to implement since the whole telescope array and laser beam must be capable of swinging precisely to all the beam directions and doing so quickly and repetitively for long periods of time. For gravity wave studies, all beam directions would need to be sampled within at least $T_B/2 \approx 2.5$ min and the process repeated for as long as several hours to several tens of hours. Gravity wave momentum flux observations require the largest number of beams (4 or 5 if zenith observations are also required). However, as we will see later, for LALO the accuracy of the momentum flux observations is dominated by statistical noise, not photon noise, and so the multiple beam configuration would not compromise the accuracy of the flux measurements. Horizontal wind measurements require at least two beams. The SNR for the 2-beam configuration would be half that for the beam swinging configuration and the measurement errors would be $\sqrt{2}$ times larger. This is a small performance penalty compared to the much higher cost of implementing beam swinging. Since LALO will be designed to be transportable, by constructing the telescope array in modules, we assume that most observations will be made with multiple beams.

**Wind and Temperature Observations 30-1000 km:** First we consider the performance of LALO when configured to make horizontal wind, temperature, and density observations using Rayleigh, Na, Fe and He(2$^2$S) Doppler lidar techniques. In this case, two beams are used to probe the atmosphere at 30$^\circ$ off-zenith to the north and east. The laser power is distributed equally between the two beams and the telescope array is also divided equally into two sub-arrays (50 m² each), pointed in the directions of the two laser beams. As noted in Section 2, the horizontal winds will be contaminated by the vertical wind (see equations (2) and (3)) even for a 30$^\circ$ zenith angle. However for simplicity, we ignore the vertical wind contamination in the error calculations. To observe vertical winds and temperature, a single laser beam and the whole telescope array (100 m²) are pointed at zenith.

Detailed SNR and error calculations for horizontal and vertical winds and temperatures are provided in Appendix C. The results are summarized below in Tables 3 and 4. In the middle atmosphere and lower thermosphere (30-150 km) for resolutions of 500 m and 2.5 min, the horizontal and vertical wind accuracies are better than about 5 m/s and the temperature accuracy is better than about 5 K. In particular, highly accurate Na and Fe lidar observations of temperatures and winds can be made in the thermosphere between 100 and 150 km at the resolution sufficient to observe the full gravity wave spectrum. Although the Na and Fe densities above 100 km are tenuous (see Appendix C and references therein), the large 100 m² aperture provides adequate signal levels even when the densities fall below 10 cm$^{-3}$. Between 80 and 100 km where the Na and Fe densities are greatest, the wind and temperature accuracies are better than a few 0.1 m/s and 0.1 K.

The upper thermosphere between 300 and 1000 km is devoid of gravity waves because severe damping at lower altitudes caused by molecular diffusion eliminates them. Consequently,
scientifically useful observations of this region can be made at much coarser resolutions. For resolutions of 25 km and 5 min, the He(2^3S) Doppler lidar is capable of observing the horizontal and vertical winds with accuracies of a few tens of m/s or better and temperatures with accuracies better than 10 K.

Table 3. LALO Wind and Temperature Measurement Accuracy 30-150 km

<table>
<thead>
<tr>
<th>z(km)</th>
<th>Δu_{rms}(m/s)</th>
<th>Δw_{rms}(m/s)</th>
<th>ΔT_{rms}(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rayleigh Na</td>
<td>Fe</td>
<td>Rayleigh Na</td>
</tr>
<tr>
<td>150</td>
<td>16.50</td>
<td>5.75</td>
<td>3.8</td>
</tr>
<tr>
<td>145</td>
<td>12.40</td>
<td>4.50</td>
<td>2.9</td>
</tr>
<tr>
<td>140</td>
<td>8.40</td>
<td>3.80</td>
<td>1.9</td>
</tr>
<tr>
<td>135</td>
<td>5.70</td>
<td>3.25</td>
<td>1.3</td>
</tr>
<tr>
<td>130</td>
<td>3.90</td>
<td>2.50</td>
<td>0.91</td>
</tr>
<tr>
<td>125</td>
<td>2.85</td>
<td>2.05</td>
<td>0.66</td>
</tr>
<tr>
<td>120</td>
<td>1.95</td>
<td>1.65</td>
<td>0.45</td>
</tr>
<tr>
<td>115</td>
<td>1.30</td>
<td>1.35</td>
<td>0.30</td>
</tr>
<tr>
<td>110</td>
<td>0.87</td>
<td>1.10</td>
<td>0.21</td>
</tr>
<tr>
<td>105</td>
<td>0.51</td>
<td>0.74</td>
<td>0.14</td>
</tr>
<tr>
<td>100</td>
<td>0.24</td>
<td>0.44</td>
<td>0.090</td>
</tr>
<tr>
<td>95</td>
<td>0.18</td>
<td>0.23</td>
<td>0.081</td>
</tr>
<tr>
<td>90</td>
<td>0.17</td>
<td>0.15</td>
<td>0.081</td>
</tr>
<tr>
<td>85</td>
<td>0.19</td>
<td>0.16</td>
<td>0.082</td>
</tr>
<tr>
<td>80</td>
<td>0.32</td>
<td>0.27</td>
<td>0.10</td>
</tr>
<tr>
<td>75</td>
<td>1.45</td>
<td>1.80</td>
<td>0.34</td>
</tr>
<tr>
<td>70</td>
<td>1.6</td>
<td></td>
<td>0.38</td>
</tr>
<tr>
<td>60</td>
<td>0.73</td>
<td></td>
<td>0.19</td>
</tr>
<tr>
<td>50</td>
<td>0.38</td>
<td></td>
<td>0.12</td>
</tr>
<tr>
<td>40</td>
<td>0.24</td>
<td></td>
<td>0.10</td>
</tr>
<tr>
<td>30</td>
<td>0.20</td>
<td></td>
<td>0.099</td>
</tr>
</tbody>
</table>
Table 4. LALO Wind and Temperature Measurement Accuracy 300-1000 km

          \( \Delta z = 25 \text{ km} - \Delta t = 5 \text{ min} \)

<table>
<thead>
<tr>
<th>z(km)</th>
<th>( \Delta u_{\text{rms}} ) (m/s)</th>
<th>( \Delta w_{\text{rms}} ) (m/s)</th>
<th>( \Delta T_{\text{rms}} ) (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>34</td>
<td>8.0</td>
<td>7.4</td>
</tr>
<tr>
<td>950</td>
<td>30</td>
<td>6.7</td>
<td>6.3</td>
</tr>
<tr>
<td>900</td>
<td>25</td>
<td>5.9</td>
<td>5.4</td>
</tr>
<tr>
<td>850</td>
<td>22</td>
<td>5.1</td>
<td>4.7</td>
</tr>
<tr>
<td>800</td>
<td>19</td>
<td>4.4</td>
<td>4.0</td>
</tr>
<tr>
<td>750</td>
<td>16</td>
<td>3.8</td>
<td>3.5</td>
</tr>
<tr>
<td>700</td>
<td>14</td>
<td>3.3</td>
<td>3.0</td>
</tr>
<tr>
<td>650</td>
<td>12</td>
<td>2.9</td>
<td>2.7</td>
</tr>
<tr>
<td>600</td>
<td>11</td>
<td>2.5</td>
<td>2.3</td>
</tr>
<tr>
<td>550</td>
<td>10</td>
<td>2.3</td>
<td>2.2</td>
</tr>
<tr>
<td>500</td>
<td>9.6</td>
<td>2.2</td>
<td>2.0</td>
</tr>
<tr>
<td>450</td>
<td>9.8</td>
<td>2.3</td>
<td>2.1</td>
</tr>
<tr>
<td>400</td>
<td>12</td>
<td>2.7</td>
<td>2.5</td>
</tr>
<tr>
<td>350</td>
<td>17</td>
<td>4.0</td>
<td>3.7</td>
</tr>
<tr>
<td>300</td>
<td>35</td>
<td>8.1</td>
<td>7.3</td>
</tr>
</tbody>
</table>
4. Gravity Wave and Eddy Fluxes

Gravity Wave Vertical Fluxes: The gravity wave vertical flux ($<VF_{GW}>$) is defined as the expected value of the product of the vertical wind fluctuations ($w'_{GW}$) and, either the temperature ($T'_{GW}$), constituent density ($\rho'_{GW}$) or horizontal wind fluctuations ($u'_{GW}$ or $v'_{GW}$) caused by gravity wave action (excluding turbulence)

$$<VF_{GW}>=<w'_{GW}X'_{GW}>$$

where the angle brackets denote ensemble averaging and

$$X'_{GW} = \begin{bmatrix} T'_{GW} & \text{gravity wave heat flux} \\ \rho'_{GW} & \text{gravity wave constituent flux} \\ u'_{GW} & \text{zonal momentum flux} \\ v'_{GW} & \text{meridional momentum flux} \end{bmatrix}$$

Because the smallest gravity wave vertical wavelengths and periods at mesopause heights are 1 km and 5 min, the resolutions of the density, wind and temperature data should be on the order of 500 m and 2.5 min to ensure that all of the gravity wave energy is captured in the flux estimates.

To reduce the variance of the flux estimate, the instantaneous flux measurements are averaged over time or altitude or both

$$\overline{VF_{GW}} = \frac{1}{L\tau} \int_{-\tau/2}^{\tau/2} dt \int_{z_0-L/2}^{z_0+L/2} dz \left[ (w'_{GW} + \Delta w)(X'_{GW} + \Delta X) - (\Delta w\Delta X)_{est} \right]$$

$$<\Delta w\Delta X> = <w'_{GW}X'_{GW} >$$

$$<\overline{VF_{GW}}> = \frac{1}{L\tau} \int_{-\tau/2}^{\tau/2} dt \int_{z_0-L/2}^{z_0+L/2} dz <(w'_{GW} + \Delta w)(X'_{GW} + \Delta X) - (\Delta w\Delta X)_{est}> = <w'_{GW}X'_{GW}>$$

(11)

where $L$ and $\tau$ are, respectively, the averaging height range and period (overbar denotes sample average) and $(\Delta w\Delta X)_{est}$ is an unbiased estimate of the measurement noise covariance.

$\overline{VF_{GW}}$ is an unbiased estimator of $<w'_{GW}X'_{GW}>$, and its variance is [Gardner and Yang, 1998]

$$Var(\overline{VF_{GW}}) = \left\{ 1 + \frac{\Delta z\Delta t}{\Delta z_{av}\Delta t_{av}} \left[ \frac{Var(\Delta w)}{Var(w'_{GW})} + \frac{Var(\Delta X)}{Var(X'_{GW})} \right] \right\} \frac{\Delta z_{av}\Delta t_{av}}{L\tau} Var(w'_{GW})Var(X'_{GW})$$

$$\Delta z_{av} = 1 km$$

$$\Delta t_{av} = 10 \text{ min}$$

(12)

$\Delta z_{av}$ and $\Delta t_{av}$ are respectively, the vertical correlation length and correlation time of the gravity wave vertical fluxes.

Momentum fluxes are measured using the dual-beam technique pioneered by Vincent and Reid 1983). Thorsen et al. [2000] have explored the measurement uncertainties associated with this technique. If the lidar employs two sets of dual co-planar beams, one set pointing $\theta$
degrees off-zenith due east and west, and the other pointing due north and south, then the variance of the zonal momentum flux is given by [Thorsen et al. 2000].

\[
\text{Var}(w'_{GW}u'_{GW}) = \frac{\Delta z \Delta t}{L \tau} \left[ \text{Var}(w'_{GW}) \text{Var}(u'_{GW}) + \frac{\text{Var}(w'_{GW})}{2} \frac{\text{Var}(u'_{GW})}{\tan^2 \theta} + \frac{\text{Var}(u'_{GW})}{4} \tan^2 \theta \right]
\]

Because the momentum flux measurements are made using small zenith angles, the LOS wind errors caused by photon noise (i.e. terms in (13) depending on \( \text{Var}(\Delta V_{EW}) \)) are approximately equal to the vertical wind error \( \text{Var}(\Delta w) \). For large SNRs, the photon noise component is small and the variance is dominated by statistical fluctuations represented by the first bracketed terms on the right-hand-side of (13). The momentum flux variance is minimum for the zenith angle satisfying

\[
\theta_{\text{min}} = \tan^{-1} \left( \frac{\text{Var}(w'_{GW})}{\text{Var}(u'_{GW})} \right),
\]

and is given by

\[
\text{Var}(w'_{GW}u'_{GW})_{\theta_{\text{min}}} = \left\{ 1 + \frac{\Delta z \Delta t}{\Delta z \Delta t_{\text{FW}}} \frac{\text{Var}(\Delta w)}{\text{Var}(w'_{GW})} \left[ 1 + \frac{\text{Var}(\Delta w)}{4 \text{Var}(w'_{GW})} \right] \right\} \frac{\Delta z \Delta t_{\text{FW}}}{L \tau} \text{Var}(w'_{GW}) \text{Var}(u'_{GW})
\]

(15)

The variance formula for the meridional momentum flux is similar. For typical values of the vertical and horizontal wind variances measured at the Starfire Optical Range, NM (viz. 3.5 \( \text{m}^2/\text{s}^2 \) and 350 \( \text{m}^2/\text{s}^2 \), respectively), the zenith angle that minimizes the momentum flux variance is about 6 degrees.

Photon noise is negligible and statistical noise dominates all the flux variances whenever the following conditions hold

\[
\begin{align*}
\frac{\Delta z \Delta t}{\Delta z \Delta t_{\text{FW}}} \frac{\text{Var}(\Delta w)}{\text{Var}(w'_{GW})} & \approx 1 \frac{\text{Var}(\Delta w)}{8 \text{Var}(w'_{GW})} \\
\frac{\Delta z \Delta t}{\Delta z \Delta t_{\text{FW}}} \frac{\text{Var}(\Delta T)}{\text{Var}(T'_{GW})} & \approx 1 \frac{\text{Var}(\Delta T)}{8 \text{Var}(T'_{GW})} \\
\frac{\Delta z \Delta t}{\Delta z \Delta t_{\text{FW}}} \frac{\text{Var}(\Delta \rho_c)}{\text{Var}(\rho'_{GW})} & \approx 1 \frac{\text{Var}(\Delta \rho_c)}{8 \text{Var}(\rho'_{GW})}
\end{align*}
\]

(16)

At mesopause heights, statistical noise dominates whenever

\[
\begin{align*}
5 \times 10^3 & \ll \text{SNR}_{\text{Na}} \\
2 \times 10^3 & \ll \text{SNR}_{\text{Fe}} \\
9 \times 10^3 & \ll \text{SNR}_{\text{Ray}}
\end{align*}
\]

(17)
For heat and constituent fluxes, which are measured using a single zenith-pointing beam, these conditions are satisfied for the Rayleigh lidar between 30 and 90 km and for the Na and Fe lidars between 75 and 150 km (see SNR values tabulated in Appendix C Tables C5-C7). For momentum fluxes, which are measured using four beams pointed 6 degrees off zenith, the conditions are satisfied for the Rayleigh lidar between 30 and 70 km and for the Na and Fe lidars between 75 and about 135 km. Note, the SNRs for momentum flux measurements are equal to 1/16\(^{th}\) of the single-beam values listed in Appendix C Tables C1-C3 (to account for the smaller power-aperture product for each of the four beams).

In summary, heat and constituent fluxes can be measured by LALO with accuracies limited only by statistical noise between 30 and 150 km. Momentum fluxes can be measured with accuracies limited only by statistical noise between 30 and about 135 km. In other words, for the LALO gravity wave flux measurements, the SNRs are so large that the effects of photon noise are negligible within these height ranges. Furthermore, because the gravity wave and eddy fluxes are computed from the wind, temperature and density perturbations, not their absolute values, the flux measurements are not affected by the fundamental error limits associated with laser frequency and linewidth errors or receiver filter tuning jitter.

**Eddy Fluxes:** The vertical eddy heat and constituent fluxes are defined, respectively, as the expected values of the product of the vertical wind and temperature fluctuations \(<w_{Turb}'T_{Turb}'\>) and the product of the vertical wind and constituent density fluctuations \(<w_{Turb}'\rho'_{CTurb}\>) caused by turbulence. The eddy fluxes are computed by first processing the measured vertical wind, temperature and constituent density data to derive the turbulence fluctuations \((w_{Turb}', T_{Turb}'\) and \(\rho'_{CTurb}\)). Like the gravity wave fluxes, the eddy fluxes are computed by averaging the instantaneous flux estimates over both altitude and time.

The variances of the eddy heat, constituent (i.e. Na and Fe) and momentum fluxes are

\[
\text{Var}(w_{Turb}'T_{Turb}') = \frac{\Delta z_{EF}\Delta t_{EF}}{L\tau} \text{Var}(w_{Turb}') \text{Var}(T_{Turb}') + \frac{\Delta z_t}{L\tau} \left[ \text{Var}(w_{Turb}') \text{Var}(\Delta T') + \text{Var}(\Delta w) \text{Var}(T_{Turb}') + \text{Var}(\Delta w) \text{Var}(\Delta T') \right]
\]

\[
\text{Var}(w_{Turb}'\rho'_{CTurb}) = \frac{\Delta z_{EF}\Delta t_{EF}}{L\tau} \text{Var}(w_{Turb}') \text{Var}(\rho'_{CTurb}) + \frac{\Delta z_t}{L\tau} \left[ \text{Var}(w_{Turb}') \text{Var}(\Delta \rho_{CTurb}) + \text{Var}(\Delta w) \text{Var}(\rho'_{CTurb}) + \text{Var}(\Delta w) \text{Var}(\Delta \rho_{CTurb}) \right]
\]

\[
\text{Var}(w_{Turb}'u'_{Turb}) = \frac{4}{3} \frac{\Delta z_{EF}\Delta t_{EF}}{L\tau} \text{Var}^2(w_{Turb}') + \frac{16}{9} \frac{\Delta z_t}{L\tau} \left[ \sqrt{3} \text{Var}(w_{Turb}') \text{Var}(\Delta w) + \text{Var}^2(\Delta w) \right]
\]

\[
\Delta z_{EF} = \frac{l_0}{10} \sim 50m = \Delta z
\]

\[
\Delta t_{EF} = \frac{l_0}{5\sqrt{\bar{u}^2 + \bar{v}^2}} \sim 10s = \Delta t
\]

where \(l_0\) is the outer scale of turbulence (\(~500-1500\) m), \(\Delta z_{EF}\) is the eddy flux correlation length and \(\Delta t_{EF}\) is the correlation time. We assume that the turbulence is largely isotropic so that the variances of the vertical and horizontal winds \((u\) and \(v\)) are approximately equal. At mesopause heights the variances of the turbulence fluctuations are [Lübken et al.,1993; Lübken,1997]
\[ \text{Var}(\omega_{\text{Turb}}) = \text{Var}(u_{\text{Turb}}) \sim 1 - 4m^2/s^2 \]
\[ \text{Var}(T'_{\text{Turb}}) \sim 0.5 - 1K^2 \]
\[ \text{Var}(\rho'_{\text{Turb}}) \sim 0.1 - 1\%^2 \]

Because of the high resolution required to observe the turbulence fluctuations (\(\Delta z=50\) m and \(\Delta t=10\) s), both photon noise and statistical noise affect the accuracies of the eddy flux measurements. Photon noise is negligible when all the following conditions are satisfied

\[
\frac{\Delta z \Delta t}{\Delta z_{EF} \Delta t_{EF}} \frac{\text{Var}(\Delta w)}{\text{Var}(w'_{\text{Turb}})} = \frac{\text{Var}(\Delta w)}{\text{Var}(w'_{\text{Turb}})} << 1
\]
\[
\frac{\Delta z \Delta t}{\Delta z_{EF} \Delta t_{EF}} \frac{\text{Var}(\Delta T)}{\text{Var}(T'_{\text{Turb}})} = \frac{\text{Var}(\Delta T)}{\text{Var}(T'_{\text{Turb}})} << 1
\]
\[
\frac{\Delta z \Delta t}{\Delta z_{EF} \Delta t_{EF}} \frac{\text{Var}(\Delta w) \text{Var}(\Delta T)}{\text{Var}(w'_{\text{Turb}}) \text{Var}(T'_{\text{Turb}})} = \frac{\text{Var}(\Delta w) \text{Var}(\Delta T)}{\text{Var}(w'_{\text{Turb}}) \text{Var}(T'_{\text{Turb}})} << 1
\]

This requires a SNR (computed with \(\Delta z=50\) m and \(\Delta t=10\) s) of about \(10^6\) or greater. This condition is satisfied for the Rayleigh lidar between 30 and 50 km altitude, for the Na lidar between 80 and 100 km and for the Fe lidar between 80 and 95 km. In these height ranges the eddy flux (and eddy diffusivity) measurements are dominated by statistical noise, not photon noise.

It is most convenient to characterize the eddy fluxes in terms of the eddy transport velocities, which are obtained by normalizing the fluxes by the mean temperature and mean constituent density.

\[
\text{Heat Transport Velocity} = \frac{w'_{\text{Turb}} T'_{\text{Turb}}}{\overline{T}}
\]
\[
\text{Constituent Transport Velocity} = \frac{w'_{\text{Turb}} \rho'_{\text{Turb}}}{\overline{\rho_C}}
\]

The eddy coefficient for constituent diffusion is related to the eddy constituent flux as follows [Colegrove et al., 1966]

\[
\frac{w'_{\text{Turb}} \rho'_{\text{Turb}}}{\overline{\rho_C}} = -k_{zz} \overline{\rho_\text{Atmos}} \frac{\partial}{\partial z} \left( \frac{\overline{\rho_C}}{\overline{\rho_\text{Atmos}}} \right) = -k_{zz} \overline{\rho_C} \left( \frac{1}{\overline{H}} + \frac{1}{\overline{T}} \frac{\partial \overline{H}}{\partial z} + \frac{1}{\overline{\rho_C}} \frac{\partial \overline{\rho_C}}{\partial z} \right). \tag{22}
\]

The term in brackets on the right-hand-side of (22) is the inverse of the concentration scale height of the constituent. The eddy diffusivity can be computed by multiplying the measured eddy flux by the concentration scale height and dividing by the mean density. Near the mesopause, the eddy transport velocities are typically a few cm/s and the eddy diffusivities range from a few tens of m\(^2\)/s to perhaps 200 m\(^2\)/s [Gardner and Liu, 2010; Stroebel et al., 1985; Lübken et al., 1993 and 1997]. Thus, to be scientifically useful, the transport velocities and diffusivities must be measured with accuracies of several mm/s and m\(^2\)/s, respectively.

To derive the eddy diffusivity from the constituent transport velocity, the scale factor in (22) must be derived at the same resolution as the eddy fluxes (viz. L and \(\tau\)), which may include gravity wave fluctuations. The scale factor falls to zero at the peaks of the Na and Fe concentration profiles so in these regions the eddy diffusivity cannot be derived. This is expected since there can be no eddy mixing in regions where the concentration profile is constant with altitude. Fortunately, the nominal peaks of Na and Fe layers are separated by several km so that in principle the diffusivity profile could be measured without any altitude gaps by making simultaneous Na and Fe observations. However, the scale factor can also fall to zero on the
bottom- and topsides of both layers in the presence of strong gravity wave perturbations, which distort the profile shapes. Furthermore, it has been shown that these wave-induced Na and Fe density fluctuations are highly correlated [Chen and Yi, 2011; Huang et al., 2013]. These small correlated regions of no eddy transport will change randomly with time as the gravity wave field propagates through the Na and Fe layers. Therefore, we used the root mean-square value of the scale factor to evaluate the average error in the eddy diffusivity measurement (see Appendix D).

\[ \Delta k_{zz} = \frac{\Delta w'_{Turb} \rho'_{CTurb}}{\left( \frac{\rho_c}{H} + \frac{\rho_c}{T} \frac{\partial T}{\partial z} + \frac{\partial \rho_c}{\partial z} \right)_{rms}} \]  

(23)

While the fundamental vertical wind, temperature and density must be measured with high resolution (\(k_H=50 \text{ m and } \Delta t=10 \text{ s}\)) to capture the turbulence fluctuations, the instantaneous eddy flux profiles are averaged in height (L) and time (t) to reduce errors. Turbulence is generated by breaking gravity waves and so it is desirable to resolve the eddy fluxes and diffusivities with resolutions comparable to the shortest wavelength, shortest period gravity waves. For LALO, we assume that the transport velocity and diffusivity profiles are derived with a vertical resolution of \(L=500 \text{ m}\) and a temporal resolution of \(\tau=2.5 \text{ min}\). We also consider the accuracies of the measured diffusivities derived at somewhat lower resolutions of \(L=2.5 \text{ km}\) and \(\tau=2 \text{ h}\), which is useful for studying turbopause structure and dynamics above 100 km altitude where the Na and Fe SNRs are much smaller and the concentration scale heights are much larger.

The heat and momentum fluxes are related to the eddy coefficients for thermal \((k_{tt})\) and momentum diffusion \((k_{mm})\) as follows

\[
\begin{align*}
\overline{w'_{Turb} \theta'_{Turb}} &= -k_h \frac{\partial \overline{\theta}}{\partial z} = -k_h \overline{\theta} \left( \frac{\Gamma_{ad} + \partial \overline{T}}{T} / \partial z \right) \\
k_h &= \frac{-\overline{w'_{Turb} \theta'_{Turb}} / \overline{\theta}}{\left( \frac{\Gamma_{ad}}{T} + \frac{1}{T} \frac{\partial T}{\partial z} \right)} = \frac{-\overline{w'_{Turb} T'_{Turb}} / \overline{T}}{\left( \frac{\Gamma_{ad}}{T} + \frac{1}{T} \frac{\partial T}{\partial z} \right)} = \frac{-\overline{w'_{Turb} T'_{Turb}} / \overline{T}}{\left( \frac{\Gamma_{ad}}{T} + \frac{1}{T} \frac{\partial T}{\partial z} \right)} \\
\overline{w'_{Turb} u'_{Turb}} &= -k_m \frac{\partial \overline{u}}{\partial z} \\
\overline{w'_{Turb} u'_{Turb}} &= Var\left(\overline{w'_{Turb} u'_{Turb}}\right) / \left( \partial \overline{u} / \partial z \right)^2
\end{align*}
\]

where \(\theta\) is the potential temperature and \(u\) is the horizontal wind in the dominant direction of the momentum flux. Because we are interested in measuring the eddy heat flux and thermal diffusivity at high resolution, we evaluate the error by taking into account gravity wave perturbations.

\[
\Delta k_h = \frac{\Delta w'_{Turb} T'_{Turb}}{\left( \frac{\Gamma_{ad} + \partial T}{\partial z} \right)_{rms}} = \frac{\Delta w'_{Turb} T'_{Turb}}{\sqrt{\left( \frac{\Gamma_{ad} + \partial \overline{T}}{\partial z} \right)^2 + Var\left( \partial T / \partial z \right)}} \]

(25)

The errors for the measured eddy heat transport velocity (eddy heat flux), \(k_h\) and \(k_{zz}\) are tabulated in Tables 5 and 6 between 30 and 130 km. It was assumed that LALO employed a single beam pointed at zenith to make the eddy flux and diffusivity measurements. At the highest resolution (500 m and 2.5 min), the heat transport velocity can be measured with accuracies of 1
cm/s or less between 30 and 115 km altitude. The eddy thermal diffusivity can be measured with accuracies of about 4 m$^2$/s or less between 30 and 60 km and between 80 and 105 km. At the lower 2.5 km and 2 h resolution, the eddy thermal diffusivity can be measured with accuracies of 20 m$^2$/s or less from 30 to 130 km. Similar accuracies can be achieved for $k_{zz}$ above 80 km. Note, that while the Rayleigh lidar can measure the heat flux it cannot measure constituent flux because the molecular backscatter signal arises from all the atmospheric constituents. Consequently, it is not possible to measure the $k_{zz}$ below about 80 km with the Rayleigh lidar alone. However, if the Rayleigh lidar is paired with a sufficiently powerful ozone lidar or a Raman lidar, which measures specific constituents that are not uniformly mixed in the atmosphere, eddy flux measurements below 80 km would be possible.

The ratio

$$Pr = \frac{k_M}{k_H}$$

is called the turbulent Prandtl number. If $Pr$ is small (<1) then the net effect of turbulence on the thermal budget is cooling associated with downward heat transport. Conversely if $Pr$ is large (>>1), the net effect is heating [Strobel et al., 1985; Liu, 2000; Liu, 2009]. Given the accuracies of the diffusivity measurements tabulated in Tables 5 and 6, LALO will be capable of providing accurate measurements of the turbulent Prandtl number throughout the mesopause region and lower thermosphere at high spatial and temporal resolution. Furthermore, it should be possible to extend the Prandtl number measurements below 80 km through the addition of powerful ozone and Raman lidars.
Table 5. Eddy Heat Transport and Thermal Diffusivity ($k_H$) Measurements 30-130 km
Single Zenith Beam – Rayleigh, Na and Fe Lidars

<table>
<thead>
<tr>
<th>z(km)</th>
<th>$\frac{\Delta w'<em>T \tau'</em>{Turb}}{\bar{T}} (cm / s)$</th>
<th>$\Delta k_H (m^2 / s)$</th>
<th>$\Delta k_H (m^2 / s)$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>L = 500 m $\tau = 2.5$ min</td>
<td>L = 500 m $\tau = 2.5$ min</td>
<td>L = 2.5 km $\tau = 2$ h</td>
</tr>
<tr>
<td>Rayleigh</td>
<td>Na</td>
<td>Fe</td>
<td>Rayleigh</td>
</tr>
<tr>
<td>130</td>
<td>5.1</td>
<td>3.3</td>
<td>485</td>
</tr>
<tr>
<td>120</td>
<td>1.3</td>
<td>1.6</td>
<td>102</td>
</tr>
<tr>
<td>115</td>
<td>0.59</td>
<td>1.0</td>
<td>39</td>
</tr>
<tr>
<td>110</td>
<td>0.30</td>
<td>0.71</td>
<td>14</td>
</tr>
<tr>
<td>105</td>
<td>0.13</td>
<td>0.36</td>
<td>4.1</td>
</tr>
<tr>
<td>100</td>
<td>0.057</td>
<td>0.16</td>
<td>1.4</td>
</tr>
<tr>
<td>97.5</td>
<td>0.049</td>
<td>0.094</td>
<td>1.1</td>
</tr>
<tr>
<td>95</td>
<td>0.047</td>
<td>0.077</td>
<td>1.0</td>
</tr>
<tr>
<td>92.5</td>
<td>0.046</td>
<td>0.062</td>
<td>0.97</td>
</tr>
<tr>
<td>90</td>
<td>0.046</td>
<td>0.056</td>
<td>0.93</td>
</tr>
<tr>
<td>87.5</td>
<td>0.047</td>
<td>0.056</td>
<td>0.92</td>
</tr>
<tr>
<td>85</td>
<td>0.048</td>
<td>0.057</td>
<td>0.90</td>
</tr>
<tr>
<td>80</td>
<td>3.0</td>
<td>0.073</td>
<td>0.088</td>
</tr>
<tr>
<td>75</td>
<td>1.0</td>
<td>0.75</td>
<td>1.8</td>
</tr>
<tr>
<td>70</td>
<td>0.56</td>
<td>11</td>
<td>0.71</td>
</tr>
<tr>
<td>60</td>
<td>0.14</td>
<td>3.1</td>
<td>0.20</td>
</tr>
<tr>
<td>50</td>
<td>0.065</td>
<td>1.9</td>
<td>0.12</td>
</tr>
<tr>
<td>40</td>
<td>0.048</td>
<td>1.6</td>
<td>0.10</td>
</tr>
<tr>
<td>30</td>
<td>0.045</td>
<td>1.3</td>
<td>0.084</td>
</tr>
</tbody>
</table>
Table 6. Eddy Diffusivity (k_{zz}) Measurements 30-130 km
Single Zenith Beam – Na and Fe Lidars

<table>
<thead>
<tr>
<th>z(km)</th>
<th>Δk_{zz}(m^2/s) L = 500 m τ = 2.5 min</th>
<th>Δk_{zz}(m^2/s) L = 2.5 km τ = 2 h</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Na</td>
<td>Fe</td>
</tr>
<tr>
<td>130</td>
<td>863</td>
<td>248</td>
</tr>
<tr>
<td>125</td>
<td>461</td>
<td>174</td>
</tr>
<tr>
<td>120</td>
<td>214</td>
<td>118</td>
</tr>
<tr>
<td>115</td>
<td>99</td>
<td>78</td>
</tr>
<tr>
<td>110</td>
<td>28</td>
<td>49</td>
</tr>
<tr>
<td>105</td>
<td>3.8</td>
<td>22</td>
</tr>
<tr>
<td>100</td>
<td>2.3</td>
<td>8.5</td>
</tr>
<tr>
<td>97.5</td>
<td>3.6</td>
<td>5.2</td>
</tr>
<tr>
<td>95</td>
<td>10</td>
<td>4.5</td>
</tr>
<tr>
<td>92.5</td>
<td>3.0</td>
<td>12</td>
</tr>
<tr>
<td>90</td>
<td>1.6</td>
<td>4.7</td>
</tr>
<tr>
<td>87.5</td>
<td>1.2</td>
<td>2.2</td>
</tr>
<tr>
<td>85</td>
<td>0.86</td>
<td>1.5</td>
</tr>
<tr>
<td>80</td>
<td>0.88</td>
<td>1.3</td>
</tr>
<tr>
<td>75</td>
<td>6.8</td>
<td>20</td>
</tr>
</tbody>
</table>

5. Daytime Observations

LALO would also yield superb measurements during the daytime. However, background noise from scattered sunlight can significantly reduce the lidar SNR and measurement accuracy during daytime. The expected background photon count per range bin is given by

\[ \bar{N}_B(\lambda) = \eta_{Telescope} \frac{S_{Sky}(\lambda)A_{Telescope} \Delta \lambda \Omega_{FOV}}{hc/\lambda} \frac{2 \Delta z \Delta t R_{Laser}}{c} \]

\[ \eta_{Telescope} = \text{optical efficiency of the telescope including detector QE} \]

\[ S_{Sky}(\lambda) = \text{sky spectral radiance (W/m}^2/\text{nm/sr)} \]

\[ A_{Telescope} = \text{telescope aperture area (m}^2) \]

\[ \Delta \lambda = \text{telescope optical bandwidth (nm)} \]

\[ \Omega_{FOV} = \text{telescope solid angle field-of-view (sr)} \]

\[ R_{Laser} = \text{laser pulse rate (s}^{-1}) \]

\[ h = \text{Planck's constant (6.63x10}^{-34}\text{J/s)} \]

\[ c = \text{velocity of light (3x10}^8\text{m/s)} \]

Narrowband optical filters such as Fabry-Perot etalons or Faraday filters [e.g. Chen et al., 1996] and narrow fields-of-view are employed to reduce background noise from the bright daytime sky. The sky’s spectral radiance depends on many factors including the elevation angle of the Sun, the lidar pointing direction relative to the Sun, the altitude of the lidar and the laser wavelength. Absorption by atomic constituents in the outer atmosphere of the Sun (Fraunhofer lines) and by
molecular constituents in the Earth’s atmosphere have a significant influence on the sky’s spectral radiance.

Estimates of \( S_{\text{Sky}} \) at the Rayleigh (532 nm), Na (589 nm) and Fe (392 nm) wavelengths are given by Gardner [2004]. The impact of daytime background radiation on the performance of the LALO lidars can be determined by evaluating (27) and then substituting the result into the daytime versions of equations (5) and (6) (see Tables A1 and B1).

\[
SNR_{\text{Day}} = \frac{SNR_{\text{Night}}}{(1 + \beta^{-1})}. \tag{28}
\]

The major impact is to reduce the SNRs by the factor \((1 + \beta^{-1})^{-1} = (1 + N_B/N_C)^{-1}\) or \((1 + N_B/N_{\text{Mol}})^{-1}\) or equivalently increase the rms photon noise errors by the factor \(\sqrt{1 + \beta^{-1}}\). For a given constituent, \(\beta^{-1}\) and the daytime SNR are proportional to

\[
\beta^{-1} = \frac{N_B}{N_C} \propto \frac{z^2 S_{\text{Sky}}(\lambda) \Delta \lambda \Omega_{\text{FOV}} R_{\text{Laser}}}{P_{\text{Laser}} \bar{p}_C(z)}
\]

\[
SNR_{\text{Day}} = \frac{N_C}{N_C + N_B} \propto \frac{P_{\text{Laser}} A_{\text{ele}} \Delta z \Delta t \bar{p}_C(z) / z^2}{\beta} \propto \frac{P_{\text{Laser}}^2 A_{\text{ele}} \Delta z \Delta t}{R_{\text{Laser}} z^4 S_{\text{Sky}}(\lambda) \Delta \lambda \Omega_{\text{FOV}}} \tag{29}
\]

Notice that \(\beta^{-1}\) is largest where the signal levels are smallest. Thus performance is eroded during daytime at the highest altitudes where the signals are weakest or where the atmospheric, Na, Fe and He densities are smallest. However, \(\beta^{-1}\) can be minimized by employing narrowband optical filters, small fields-of-view and large laser pulse energies \((P_{\text{Laser}} / R_{\text{Laser}})\).

Typically, coarser vertical and temporal resolutions would be employed during daytime to reduce the impact of background noise and increase the SNR, thereby improving the measurement accuracies. Because of the very large aperture area of LALO (100 m\(^2\)) and higher laser power levels (5-7 times larger), the daytime SNR for LALO would be about 400 times larger, and the measurement errors about 20 times smaller, than current lidar systems that employ much smaller telescopes (0.5-0.8 m\(^2\)) and lower laser power levels (5-7 times smaller) at much lower pulse repetition rates (50 vs 750 pps). Even so, daytime errors for LALO could be 10 or more times larger than the values listed in Tables 3-5.
Appendix A - Resonance Fluorescence Doppler Lidar

Doppler resonance fluorescence (RF) lidars probe the fluorescence spectra of the observed constituents at three different frequencies. The laser is frequency-locked to the peak of the fluorescence line \( f_{\text{Laser}} = f_C \) and an acousto-optic modulator is used to shift the frequency several hundred MHz into the wings of the line at \( f = f_{\text{Laser}} \pm f_\delta \) [She and Yu, 1994]. Measurements are made by tuning the laser sequentially to each of the three frequencies \( f_C - f_\delta, f_C, \) and \( f_C + f_\delta \). The backscattered signals are then used to derive the constituent density, temperature and radial velocity. The frequency shift \( f_\delta \) and the dwell times at each frequency are chosen to minimize the errors. The optimum frequency shift and dwell times depend on the ratio of the fluorescence signal to the background noise and they are different for each parameter being measured [Gardner, 2004].

Numerous authors have analyzed the performance characteristics of Doppler RF lidars [e.g. Bills et al., 1991a,b; She et al., 1992; She and Yu, 1994; Papen et al., 1995; Gardner, 2004; Chu and Papen, 2005]. To derive accurate estimates of winds, temperatures and densities, the data processing algorithms must account for constituent isotopes, multiple hyperfine lines, the laser spectral shape and in certain situations, saturation of the fluorescence line [e.g. Welsh and Gardner, 1989; von der Gathen, 1991].

However, in most cases measurement precision can be adequately assessed by modeling the laser-excited fluorescence line or equivalently, the effective backscatter cross-section, as a symmetric Gaussian profile,

\[
\sigma_{\text{eff}}^{C}(f, V_R, T, \sigma_{\text{Laser}}) = \frac{\varepsilon^2 f_{\text{osc}} / 4 \varepsilon_0 m_e c}{\sqrt{2\pi}} \exp\left[-\frac{(f - f_C - V_R/\lambda_C)^2}{2(\sigma_C^2 + \sigma_{\text{Laser}}^2)}\right] \frac{\sqrt{\gamma_t T}}{\sqrt{\sigma_C^2 + \sigma_{\text{Laser}}^2}} \tag{A-1}
\]

where \( e \) is the electron charge, \( f_{\text{osc}} \) is the oscillator strength, \( \varepsilon_0 \) is the permittivity of free-space, \( m_e \) is the electron mass, \( c \) is the velocity of light, \( \sigma_{\text{Laser}} \) is the rms laser linewidth and \( \sigma_C \) is the rms width of the thermally broadened fluorescence line.

\[
\sigma_C = \sqrt{\frac{k_B T}{\lambda_C^2 m_e}} = \sqrt{\gamma_t T} \tag{A-2}
\]

\( k_B = 1.38 \times 10^{-23} \) J/K is Boltzmann’s constant and \( m_C \) is the atomic mass of the constituent atom. In resonance fluorescence scattering, the species absorbs the incident photons emitted by the laser, which are then spontaneously reradiated. If the bulk motion of the atoms is away from the laser beam (i.e. the radial velocity of the atoms \( V_R \) is positive), then to these atoms the laser frequency will appear to be Doppler shifted to a lower frequency by the amount \( V_R/\lambda_C \). Equivalently, the absorption cross-section of the species will appear to have been Doppler shifted to a higher frequency.

The backscattered photons are collected by a broadband receiving telescope whose bandwidth is much larger then the fluorescence linewidth. The expected backscattered photon count as a function of altitude and laser frequency is given by the lidar equation [Bills et al., 1991a; Chu and Papen, 2005].

\[
\bar{N}_C(z,t,f) = \frac{\eta T_{\text{Atmos}}^2 P_{\text{Laser}} A_{\text{Focal}} \alpha_{\text{ext}}(f,z) \sigma_{\text{eff}}^{C}(f,V_R,T,\sigma_{\text{Laser}}) \rho_C(z,t) \Delta z \Delta t \cos \theta}{4 \pi z^2 h_f^2} \tag{A-3}
\]

where
\[ z = \text{altitude (m)} \]
\[ t = \text{time (LT)} \]
\[ \theta = \text{zenith angle} \]
\[ \eta = \text{lidar system efficiency including detector quantum efficiency} \]
\[ T^{2}_{\text{Atmos}} = \text{2-way atmospheric transmittance} \]
\[ P_{\text{Laser}} = \text{laser power (W)} \]
\[ A_{\text{tel}} = \text{telescope aperture area (m}^2) \]
\[ \alpha_{\text{ex}}(f, z) = \text{extinction caused by absorption and scattering} \]
\[ \sigma_{\text{eff}}^{C}(f, V_R, T, \sigma_{\text{Laser}}) = \text{effective constituent backscatter cross-section} \]
\[ \rho_{C}(z, t) = \text{constituent density profile (m}^3) \]
\[ \Delta z = \text{vertical resolution (m)} \]
\[ \Delta t = \text{profile integration period (i.e. time laser is tuned to f) (s)} \]
\[ h = \text{Planck’s constant=6.63x10^{-34} J/s} \]

For a pure Gaussian absorption cross-section, the temperature, winds and constituent density are derived from the measured photon counts as follows,

\[ T(z) = \frac{f^2}{\gamma_C} R_T - \frac{\sigma^2_{\text{Laser}}}{\gamma_C} \]
\[ R_T = 1/\ln \left[ \frac{(1-\chi)^2}{4\chi^2} \frac{N^2_C(f_c)}{N_C(f_c-f_\delta)N_C(f_c+f_\delta)} \right] = \frac{\sigma^2_{\text{Laser}}}{f_\delta} \]
\[ V_T(z) = \frac{\lambda_C f_\delta}{2} R_T R_V \]
\[ R_V = \ln \left[ \frac{N_C(f_c-f_\delta)}{N_C(f_c+f_\delta)} \right] = \frac{2f_\delta V_R(z)}{\lambda_C \left( \sigma^2_C + \sigma^2_{\text{Laser}} \right)} \]
\[ \rho_C(z) = \frac{\xi R^2_\rho}{\chi + (1-\chi)\exp(-1/2 R_T) \cosh(R_V/2)} \]
\[ R^2_\rho = N_C(f_c-f_\delta) + N_C(f_c) + N_C(f_c+f_\delta) = \left[ \chi + (1-\chi)\exp(-1/2 R_T) \cosh(R_V/2) \right] R_C(f_c) \]  
(A-4)

where \( \chi \) is fraction of time (dwell time) the lidar is tuned to the peak of the fluorescence line \((f_{\text{Laser}}=f_C)\) and \((1-\chi)/2\) is the fraction of time the lidar is tuned to each of the wing frequencies \((f_{\text{Laser}}=f_C\pm f_\delta)\). For pure Gaussian backscatter cross-sections, the temperature, wind and constituent density depend linearly on the respective lidar signal metrics, \( R_T, R_V \) and \( R_\rho \). In practice, rather than using the formulas in (A4) to derive \( T, V_T \) and \( \rho_C \), numerical models, which account for departures from purely Gaussian cross-sections caused by multiple hyperfine lines and isotopes, are used to compute lookup tables which map the observed signal metrics \( R_T, R_V \) and \( R_\rho \) into the correct temperature, wind and density values.

**Signal Photon Noise:** For zenith pointing, the rms photon noise errors are given by [Gardner, 2004]
\[ \Delta T_{\text{rms}} = 2 \left[ \frac{1}{\chi} + e^\alpha (1 + \beta e^{-\alpha/2}) \right]^{1/2} \cdot \frac{T(z)}{\sqrt{\text{SNR}(z,f_c)}} \]
\[ \Delta w_{\text{rms}} = \frac{e^\alpha (1 + \beta e^{-\alpha/2})}{\alpha(1 - \chi)(1 + \beta)} \cdot \lambda_c \sqrt{\frac{\sigma_C^2 + \sigma_{\text{Laser}}^2}{\text{SNR}(z,f_c)}} \]
\[ (\Delta \rho_C)_{\text{rms}} = \frac{1}{\alpha} \left[ \frac{(\alpha - 1)^2}{\chi} + e^\alpha (1 + \beta e^{-\alpha/2}) \right]^{1/2} \cdot \frac{\rho_c(z)}{\sqrt{\text{SNR}(z,f_c)}} \]

where

\[ \alpha = f_\delta^2 / (\sigma_C^2 + \sigma_{\text{Laser}}^2) \]
\[ \beta = \bar{N}_C(z,f_c) / (\bar{N}_B + \bar{N}_D) \]  

and the effective signal-to-noise power ratio is

\[ \text{SNR}(z,f_c) = \frac{\bar{N}_C^2(z,f_c)}{\bar{N}_C(z,f_c) + \bar{N}_B + \bar{N}_D}. \]

\( \bar{N}_C(z,f_c) \) is the expected signal count (overbar denotes sample mean) from the range bin corresponding to altitude \( z \) when the laser is tuned to the peak of the fluorescence line during the whole temporal resolution cell (i.e. when \( \chi = 1 \)). \( \bar{N}_B \) is the expected background count and \( \bar{N}_D \) is the expected detector dark count.

For observations below the turbopause, the eddy heat flux is the most difficult parameter to measure because the observations must be made at high temporal and spatial resolution. The eddy flux error is typically dominated by photon noise in the wind and temperature measurements (see Section 4). Therefore, we choose the dwell time (\( \chi \)) and frequency shift (\( f_\delta \)) to minimize the product of the wind and temperature errors. The optimum parameters and the resulting photon noise errors are listed in Table A1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Nighttime (1&lt;&lt;\beta)</th>
<th>Daytime (\beta&lt;&lt;1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \alpha = f_\delta^2 / (\sigma_C^2 + \sigma_{\text{Laser}}^2) )</td>
<td>3.7272</td>
<td>1.8636</td>
</tr>
<tr>
<td>( \chi = \text{Dwell Time @ } f_c )</td>
<td>19.5%</td>
<td>19.5%</td>
</tr>
<tr>
<td>(1-( \chi ))/2 = Dwell Time @ ( f_c \pm f_\delta )</td>
<td>40.25%</td>
<td>40.25%</td>
</tr>
<tr>
<td>( f_\delta = \sqrt{\alpha \cdot (\sigma_C^2 + \sigma_{\text{Laser}}^2)} )</td>
<td>1.931</td>
<td>( \sqrt{\sigma_C^2 + \sigma_{\text{Laser}}^2} )</td>
</tr>
<tr>
<td>( \Delta T_{\text{rms}} )</td>
<td>( 1.945T(z) / \sqrt{\text{SNR}(z,f_c)} )</td>
<td>( 3.890T(z) / \sqrt{\text{SNR}(z,f_c)} )</td>
</tr>
<tr>
<td>( \Delta w_{\text{rms}} )</td>
<td>( 1.466\lambda_c \sqrt{\sigma_C^2 + \sigma_{\text{Laser}}^2} / \sqrt{\text{SNR}(z,f_c)} )</td>
<td>( 2.073\lambda_c \sqrt{\sigma_C^2 + \sigma_{\text{Laser}}^2} / \sqrt{\text{SNR}(z,f_c)} )</td>
</tr>
<tr>
<td>( (\Delta \rho_C)_{\text{rms}} )</td>
<td>( 1.823\rho_c(z) / \sqrt{\text{SNR}(z,f_c)} )</td>
<td>( 1.846\rho_c(z) / \sqrt{\text{SNR}(z,f_c)} )</td>
</tr>
</tbody>
</table>
The key system parameters and photon noise errors for Na, Fe and He(2^3S) RF Doppler lidars are listed in Tables A2-A4. The signal levels were computed by assuming each system uses state-of-the-art optical components and detectors. The overall system efficiencies were assumed to be between 6% and 12%, which includes atmospheric transmittance, optical efficiency of the telescopes and filters and the quantum efficiency of the detectors. The results were benchmarked against the best performing systems in current operation to confirm the calculations.

Table A2. Parameters and Measurement Errors for Na RF Doppler Lidars

@ T=185 K, w=0 m/s, \( \lambda_{\text{Na}} = 589 \text{ nm} \), \( \sigma_{\text{Na}} = 439 \text{ MHz} \), \( \sigma_{\text{Laser}} = 50 \text{ MHz} \)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Nighttime</th>
<th>Daytime</th>
</tr>
</thead>
<tbody>
<tr>
<td>( f_\delta )</td>
<td>853 MHz</td>
<td>603 MHz</td>
</tr>
<tr>
<td>( \Delta T_{\text{rms}} )</td>
<td>( \frac{360K}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
<td>( \frac{720K}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
</tr>
<tr>
<td>( \Delta w_{\text{rms}} )</td>
<td>( \frac{382m/s}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
<td>( \frac{540m/s}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
</tr>
<tr>
<td>( (\Delta \rho_{\text{Na}})_{\text{rms}} )</td>
<td>( \frac{1.823 \rho_{\text{Na}}(z)}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
<td>( \frac{1.846 \rho_{\text{Na}}(z)}{\sqrt{\text{SNR}(z,f_\text{Na})}} )</td>
</tr>
<tr>
<td>( T_e \text{T}_{\text{Amos}} )</td>
<td>( \eta_{\text{Te}}T_e^2 )</td>
<td>~6%</td>
</tr>
<tr>
<td>( \bar{N}<em>{\text{Na}}(z,f</em>\text{Na}) )</td>
<td>( \sim 2 \cdot 10^{-3} \left( \frac{100km}{z} \right)^2 P_{\text{Laser}}A_{\text{Te}} \rho_{\text{Na}}(z)\Delta z\Delta t \cos \theta )</td>
<td></td>
</tr>
<tr>
<td>( \sigma_{\text{Na}}^f(f_{\text{Na}}, V_R = 0, T, \sigma_{\text{Laser}} = 0) )</td>
<td></td>
<td>14.87 \cdot 10^{-12} \text{ cm}^2</td>
</tr>
<tr>
<td>( \gamma_{\text{Na}} )</td>
<td></td>
<td>1.042 \cdot 10^3 (\text{MHz})^2 / \text{K}</td>
</tr>
</tbody>
</table>

Units for parameters in signal level formula: \( z(\text{km}) \), \( P_{\text{Laser}}(\text{W}) \), \( A_{\text{Te}}(\text{m}^2) \), \( \rho_{\text{Na}}(\text{cm}^{-3}) \), \( \Delta z(\text{m}) \), \( \Delta t(\text{s}) \)
Table A3. Parameters and Measurement Errors for Fe RF Doppler Lidars
@ T=185 K, w=0 m/s, $\lambda_{Fe}=372$ nm, $\sigma_{Fe}=446$ MHz, $\sigma_{Laser}=15$ MHz

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Nighttime</th>
<th>Daytime</th>
</tr>
</thead>
<tbody>
<tr>
<td>$f_\delta$</td>
<td>862 MHz</td>
<td>609 MHz</td>
</tr>
<tr>
<td>$\Delta T_{rms}$</td>
<td>$360 K \sqrt{\text{SNR}(z,f_{Fe})}$</td>
<td>$720 K \sqrt{\text{SNR}(z,f_{Fe})}$</td>
</tr>
<tr>
<td>$\Delta w_{rms}$</td>
<td>$243 m / s \sqrt{\text{SNR}(z,f_{Fe})}$</td>
<td>$344 m / s \sqrt{\text{SNR}(z,f_{Fe})}$</td>
</tr>
<tr>
<td>$(\Delta \rho_{Fe})_{rms}$</td>
<td>$1.823 \rho_{Fe}(z) \sqrt{\text{SNR}(z,f_{Fe})}$</td>
<td>$1.846 \rho_{Fe}(z) \sqrt{\text{SNR}(z,f_{Fe})}$</td>
</tr>
<tr>
<td>$\eta_{Tele} T_{Atmos}^2$</td>
<td>~7%</td>
<td></td>
</tr>
<tr>
<td>$\bar{N}<em>{Fe}(z,f</em>{Fe})$</td>
<td>$\sim 10^{-4} \left( \frac{100 km}{z} \right)^2 P_{Laser} A_{Tele} \rho_{Fe}(z) \Delta z \Delta t \cos \theta$</td>
<td></td>
</tr>
<tr>
<td>$\sigma_{He}^{eff}(f_{He}, V_R = 0, T, \sigma_{Laser} = 0)$</td>
<td>$0.944 \times 10^{-12}$ cm$^2$</td>
<td></td>
</tr>
<tr>
<td>$\gamma_{Fe}$</td>
<td>$1.075 \times 10^3 (MHz)^2 / K$</td>
<td></td>
</tr>
</tbody>
</table>

Units for parameters in signal level formula; z(km), $P_{Laser}(W)$, $A_{Tele}(m^2)$, $\rho_{Fe}(cm^{-3})$, $\Delta z(m)$, $\Delta t(s)$

Table A4. Parameters and Measurement Errors for He(2$^3$S) RF Doppler Lidars
@ T=1000 K, w=0 m/s, $\lambda_{He}=1083$ nm, $\sigma_{He}=1.330$ GHz, $\sigma_{Laser}<10$ MHz

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Nighttime</th>
<th>Daytime</th>
</tr>
</thead>
<tbody>
<tr>
<td>$f_\delta$</td>
<td>2.569 GHz</td>
<td>1.816 GHz</td>
</tr>
<tr>
<td>$\Delta T_{rms}$</td>
<td>$1945 K \sqrt{\text{SNR}(z,f_{He})}$</td>
<td>$1365 K \sqrt{\text{SNR}(z,f_{He})}$</td>
</tr>
<tr>
<td>$\Delta w_{rms}$</td>
<td>$2112 m / s \sqrt{\text{SNR}(z,f_{He})}$</td>
<td>$2987 m / s \sqrt{\text{SNR}(z,f_{He})}$</td>
</tr>
<tr>
<td>$(\Delta \rho_{He})_{rms}$</td>
<td>$1.823 \rho_{He}(z) \sqrt{\text{SNR}(z,f_{He})}$</td>
<td>$1.846 \rho_{He}(z) \sqrt{\text{SNR}(z,f_{He})}$</td>
</tr>
<tr>
<td>$\eta_{Tele} T_{Atmos}^2$</td>
<td>~12%</td>
<td></td>
</tr>
<tr>
<td>$\bar{N}<em>{He}(z,f</em>{He})$</td>
<td>$\sim 6 \times 10^{-11} \left( \frac{500 km}{z} \right)^2 P_{Laser} A_{Tele} \rho_{He}(z) \Delta z \Delta t \cos \theta$</td>
<td></td>
</tr>
<tr>
<td>$\sigma_{He}^{eff}(f_{He}, V_R = 0, T, \sigma_{Laser} = 0)$</td>
<td>$2.794 \times 10^{-12}$ cm$^2$</td>
<td></td>
</tr>
<tr>
<td>$\gamma_{He}$</td>
<td>$1.770 \times 10^3 (MHz)^2 / K$</td>
<td></td>
</tr>
</tbody>
</table>

Units for parameters in signal level formula; z(km), $P_{Laser}(W)$, $A_{Tele}(m^2)$, $\rho_{He}(cm^{-3})$, $\Delta z(m)$, $\Delta t(s)$
Laser Frequency and Linewidth Errors: The laser local oscillator is frequency-locked to a hyperfine line of the fluorescence spectrum by employing Doppler-free, saturation-absorption spectroscopy techniques [e.g. She and Yu, 1995]. Acousto-optic (AO) modulators, which are driven by crystal controlled radio frequency oscillators, are then used to shift the laser frequency into the wings of the spectrum [She and Yu, 1994]. The frequency-shifted local oscillator beam is either amplified or used to seed a ring laser to achieve the required power levels and pulse format. Frequency locking can be achieved with an accuracy of several tens of kHz or less, while the AO frequency shift can be controlled with accuracy of a few Hz or less. However, pulsed lasers typically experience frequency chirping that can approach several tens of MHz [e.g. Gangopadhyay et al., 1994 and Melikechi et al., 1994], which affects both the center frequency and the linewidth. Fortunately, heterodyne detection techniques can be used to measure the frequency chirp as a function of time on a pulse-to-pulse basis [Fee et al., 1992; Chu et al., 2010]. This information can then be processed to determine the laser pulse spectrum, center frequency and linewidth.

Because of the architecture of modern Doppler RF lidars that employ AO modulators [see e.g. Chu and Papen, 2005], the laser frequency jitter, associated with locking the laser to a hyperfine line (Δf Laser), is the same for all three frequencies, while errors in the offset frequency f o, associated with the AO modulator, are negligible. However, the laser frequency chirp error (Δf Chirp) and the associated linewidth error (Δσ Laser) are different (uncorrelated) for each frequency. For RF lidars, laser frequency jitter primarily affects the wind measurement because both the temperature and density metrics (R T and R ρ) in equation (7) are insensitive to small changes in f Laser. Chirp affects all the measurements. The measurement errors caused by laser frequency jitter and chirp are

\[
\Delta T_{\text{rms}} = \frac{\sqrt{2}}{f_\delta} \left\{ \Delta f_{\text{Chirp}}^2 + \left[ 2 + (\alpha - 1)^2 \right] \frac{\sigma_{\text{Laser}}^2}{f_\delta^2} \Delta \sigma_{\text{Laser}}^2 \right\}^{1/2} T(z)
\]

\[
\Delta w_{\text{rms}} = \frac{\lambda_c}{\sqrt{2}} \left[ 2 \Delta f^2_{\text{Laser}} + \Delta f_{\text{Chirp}}^2 + (\alpha - 1)^2 \frac{\sigma_{\text{Laser}}^2}{f_\delta^2} \Delta \sigma_{\text{Laser}}^2 \right]^{1/2}
\]

\[
(\Delta \rho_C)_{\text{rms}} = \frac{(1 - \chi)}{(1 - \chi) + \chi e^{\alpha/2} \sqrt{2} f_\delta} \left\{ \Delta f_{\text{Chirp}}^2 + \left[ 2 \chi^2 e^{\alpha} \frac{\sigma_{\text{Laser}}^2}{f_\delta^2} \Delta \sigma_{\text{Laser}}^2 \right] \frac{\sigma_{\text{Laser}}^2}{f_\delta^2} \Delta \sigma_{\text{Laser}}^2 \right\}^{1/2} \rho_C(z)
\]

We assume that state-of-the-art techniques are employed to monitor the laser pulse spectrum and to lock the CW local oscillator laser to the hyperfines lines of the species being measured. For this analysis we make the conservative assumption that the residual errors in the laser frequency and linewidth are

\[
\Delta f_{\text{Laser}} = 100 \text{kHz}
\]

\[
\Delta f_{\text{Chirp}} = 100 \text{kHz}
\]

\[
\Delta \sigma_{\text{Laser}} = 500 \text{kHz}
\]

They represent the residual errors in the laser spectrum after averaging the laser pulses for several tens of seconds.

Signal Calibration Errors: The detected signals are typically normalized by the Rayleigh signal near 30 km altitude to eliminate fluctuations associated with variations in laser power and the transmittance of the lower atmosphere. The temperature and winds are derived from the ratios of the normalized photon counts (R T and R ρ) in equation (A-4) and so the signal processing does not require knowledge of the absolute signal levels. However, the constituent density is derived from the sum of the photon counts (R ρ) and so it is necessary to know the absolute values of the signals in order to determine the absolute constituent density. This is accomplished by first calibrating the Rayleigh signal at the normalizing altitude (~30 km), which requires knowledge of the atmospheric density, or equivalently the pressure and temperature, at the time the
observations are being made. It also requires precise knowledge of lidar response function versus altitude. The response function depends on the telescope field-of-view and the geometry of the laser beam and telescope. We assume the response function is determined with an accuracy of better than 0.1% so that the calibration error is dominated by errors in the assumed atmospheric density at the normalization altitude.

Typically, models are used to determine the atmospheric density but that approach does not account for wave perturbations and weather variations in the lower atmosphere that could introduce errors of up to 10%. By launching high altitude balloons during the observations to measure the atmospheric density, the error can be reduced. This approach would eliminate errors associated with weather variations and long-period waves but could not account for the random perturbations associated with gravity waves. The residual error in atmospheric density caused by gravity wave perturbations is about 1%, which means the error in the derived constituent density is also about 1%. The gravity wave temporal spectrum varies approximately as $-\omega^{-2}$ between the inertial $(2\pi/T_i)$ and buoyancy frequencies $2\pi/T_B$. Thus, if the balloon observations are averaged over the period $T_B/2 < \Delta t < T_i/2$, the density error caused by the residual gravity fluctuations is

$$\frac{(\Delta \rho_C)_{\text{max}}}{\rho_C} = 1\% \cdot \left( \frac{T_i - 2\Delta t}{T_i - T_B} \right)^{1/2} = 1\% \cdot \sqrt{1 - \frac{2\Delta t}{T_i}}$$

$$T_B = 5\text{ min}$$
$$T_i = 12\text{ h} / |\sin \phi|$$
$$\phi = \text{latitude}$$

Unfortunately, long averaging times, comparable to a substantial fraction of the inertial period, are required to appreciably reduce the residual density error.

For a large facility like LALO, which operates over a very large altitude range, the dynamic range of the signals can be extremely large. For example the Rayleigh signal from 150 km altitude is about $5 \times 10^9$ times smaller than the signal at 30 km. None of the detectors available today are linear over such a large dynamic range. Detector saturation is a common problem for large lidars that can be addressed by carefully calibrating the detector and then insuring that during operation the signal levels do not exceed the calibrated limits. For this analysis we assume that detector nonlinearity and saturation issues have been properly addressed.
**Appendix B - Rayleigh Doppler Lidar**

Several techniques have been employed to make Doppler wind measurements with Rayleigh lidars. Fabry-Perot spectrometers coupled with detector arrays can measure the spectrum of the molecular backscattered signal to deduce the Doppler shift and thermal broadening. Alternatively, narrowband optical filters such as iodine cells [e.g. Baumgarten, 2010] or Fabry-Perot etalons [Tang et al., 2012] can be used to analyze the backscattered signal. The performances of well-engineered systems are often similar and can sometimes approach within a few dB of the theoretical limit for the hypothetical optimum system design (see below, (B-9)).

For this analysis we assume that the Rayleigh Doppler lidar receiver uses up to three identical filters to probe the spectrum of the molecular backscattered signal at the peak frequency $f_{Laser}$ and in the wings of the spectrum at $f_{Laser} \pm f_\delta$. The filters are all frequency locked to the laser. In the absence of aerosol scattering, the spectrum of the molecular backscattered signal is Gaussian and we assume the passbands of the three filters can be modeled as Gaussian functions of frequency with identical widths but of course different center frequencies. The expected signal count as a function of the center frequency of the filter (f) is given by

$$
\bar{N}_{Rec}(z,f) = \frac{\chi(f)N_{Mole}(z,f_{Laser})\sigma_{Rec}}{\sqrt{\sigma_{Mole}^2 + \sigma_{Laser}^2 + \sigma_{Rec}^2}} \exp\left[-\frac{(f-f_{Laser} + 2V_R/\lambda_{Laser})^2}{2(\sigma_{Mole}^2 + \sigma_{Laser}^2 + \sigma_{Rec}^2)}\right]
$$

where $\chi(f)$ is the fraction of the total signal directed to the filter, $\sigma_{Rec}$ is the rms filter bandwidth and $\bar{N}_{Mole}(z,f_{Laser})$ is the equivalent signal count that would be detected if the telescope employed a single broadband filter ($\chi=1$ and $\sigma_{Mole}<<\sigma_{Rec}$) tuned to the laser frequency ($f=f_{Laser}$). $\sigma_{Mole}$ is the rms spectral width of the molecular backscattered signal.

$$
\sigma_{Mole} = \sqrt{\frac{4k_BT}{\lambda_{Laser}^2 m_{Amos}}} = \sqrt{Y_{Mole}T}
$$

$m_{Amos} = 4.811x10^{-23} g$ = mean mass of the atmosphere

The Rayleigh lidar equation is given by (A-3) with the effective RF backscatter cross-section and constituent density replaced by the molecular backscatter cross-section and atmospheric density.

$$
\bar{N}_{Mole}(z,t,f_{Laser}) = \frac{\eta T_{Amos}^2 P_{Laser} A_{Tele} \alpha_{ext}(f_{Laser},z)\sigma_{Ray}(\lambda_{Laser})\rho(z,t)\Delta z \Delta t \cos \theta}{4\pi z^2 hf_{Laser}}
$$

$$
\sigma_{Ray}(\lambda_{Laser})\rho(z) = \frac{3.692 \cdot 10^{-31} P(z)}{T(z)}
$$

$\sigma_{Ray}$ is the Rayleigh (molecular) backscatter cross-section ($m^2$), $\rho$ is the atmospheric density ($m^3$), $P$ is the atmospheric pressure (mb), $T$ is the temperature (K) and $\lambda_{Laser}$ is the laser wavelength (m).

The signal outputs from the three filters are processed in the same way as the 3-frequency RF measurements to derive the radial wind, kinetic temperature and relative atmospheric density. In addition, the output counts of the three filters can be summed and used to derive the hydrostatic temperature from the relative atmospheric density profile.

**Photon Noise:** The measurement errors are given by
that operates at 532 nm. The optimized system parameters and measurement frequencies are measured. The receiver filter bandwidths wing frequency filters (i.e. the fraction of signal directed to each of the filters tuned to \( f_{\text{Laser}} \pm f_\delta \). The effective signal-to-noise power ratio is

\[
\Delta T_{\text{rms}} \Delta T_{\text{Hyd}} = \frac{\Delta \rho_{\text{rms}} T(z)}{\rho(z)} = \left( \frac{\sigma_{\text{Mole}}^2 + \sigma_{\text{Laser}}^2 + \sigma_{\text{Rec}}^2}{\sigma_{\text{Rec}}} \right)^{1/2} \cdot \frac{1}{2} \cdot \left[ \frac{e^\alpha (1 + \beta e^{-\alpha/2})}{\alpha(1 - \chi)(1 + \beta)} \right]^{1/2} \cdot \lambda_{\text{Laser}} \frac{\sqrt{\sigma_{\text{Mole}}^2 + \sigma_{\text{Laser}}^2 + \sigma_{\text{Rec}}^2}}{\sqrt{\text{SNR}(z)}} \cdot \frac{\rho(z)}{\sqrt{\text{SNR}(z)}} \]

where

\[
\alpha = f_\delta^2 / (\sigma_{\text{Mole}}^2 + \sigma_{\text{Laser}}^2 + \sigma_{\text{Rec}}^2) \]

\[
\beta = \frac{N_{\text{Mole}}(z, f_{\text{Laser}})}{N_{\text{Mole}}(z, f_{\text{Laser}}) + N_B + N_D} \]

\( \chi \) is the fraction of the total received signal that is directed to the filter tuned to \( f_{\text{Laser}} \) and \( 1 - \chi/2 \) is the fraction of signal directed to each of the filters tuned to \( f_{\text{Laser}} \pm f_\delta \). The expected background count \( N_B \) is the expected detector dark count. Notice that when three different filters are employed both the kinetic and the hydrostatic temperatures may be derived from the signal.

We focus on the special 2-filter case where the signal is split equally and directed the two wing frequency filters (i.e. \( \chi=0 \)) so that winds, relative atmospheric density and only the hydrostatic temperature are measured. The receiver filter bandwidths \( \sigma_{\text{Rec}} \) and off-set frequencies \( \pm f_\delta \) are chosen to minimize the product of the vertical wind and temperature errors \( \Delta w_{\text{rms}} \Delta T_{\text{Hyd}} \). We assume the Rayleigh lidar employs a pulsed frequency-doubled Nd:YAG laser that operates at 532 nm. The optimized system parameters and measurement errors are summarized in Table B1.
Table B1. Values of $\alpha$, $\chi$, $\sigma_{Rec}$ and $f_\delta$ that Minimize the Product $\Delta w_{rms}\Delta T_{rms}^{Hyd}$ and the Resulting Measurement Errors for Rayleigh Lidars

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Nighttime (1&lt;&lt;\beta)</th>
<th>Daytime (\beta&lt;&lt;1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha = f_\delta^2 / (\sigma_{Mole}^2 + \sigma_{Laser}^2 + \sigma_{Rec}^2)$</td>
<td>1</td>
<td>1/2</td>
</tr>
<tr>
<td>$\chi$</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$(1-\chi)/2$</td>
<td>1/2</td>
<td>1/2</td>
</tr>
<tr>
<td>$\sigma_{Rec} = \sqrt{\sigma_{Mole}^2 + \sigma_{Laser}^2}$</td>
<td>867 MHz</td>
<td>867 MHz</td>
</tr>
<tr>
<td>$f_\delta = \sqrt{\alpha \sigma_{Mole}^2 + \sigma_{Laser}^2 + \sigma_{Rec}^2}$</td>
<td>1.227 GHz</td>
<td>867 MHz</td>
</tr>
<tr>
<td>$\Delta T_{Kin}^{rms}$</td>
<td>NA</td>
<td>NA</td>
</tr>
<tr>
<td>$\Delta T_{Hyd}^{rms}$</td>
<td>$\frac{1.527T(z)}{\sqrt{\text{SNR}(z)}} = \frac{282K}{\sqrt{\text{SNR}(z)}}$</td>
<td>$\frac{1.527T(z)}{\sqrt{\text{SNR}(z)}} = \frac{282K}{\sqrt{\text{SNR}(z)}}$</td>
</tr>
<tr>
<td>$\Delta w_{rms}$</td>
<td>$1.080\lambda_{Laser}\sqrt{\frac{\sigma_{Mole}^2 + \sigma_{Laser}^2}{\text{SNR}(z)}}$</td>
<td>$1.527\lambda_{Laser}\sqrt{\frac{\sigma_{Mole}^2 + \sigma_{Laser}^2}{\text{SNR}(z)}}$</td>
</tr>
<tr>
<td>$\Delta \rho_{rms}$</td>
<td>$1.527 \rho(z) / \sqrt{\text{SNR}(z)}$</td>
<td>$1.527 \rho(z) / \sqrt{\text{SNR}(z)}$</td>
</tr>
<tr>
<td>$\text{SNR}(z)$</td>
<td>$\tilde{N}<em>{Mole}(z,f</em>{Laser})$</td>
<td>$\tilde{N}<em>{Mole}^2(z,f</em>{Laser}) / \tilde{N}_B$</td>
</tr>
<tr>
<td>$\eta_{Tel} T_{Amos}^2$</td>
<td>~9%</td>
<td></td>
</tr>
<tr>
<td>$\tilde{N}<em>{Mole}(z,f</em>{Laser})$</td>
<td>~$10 \left( \frac{100 km}{z} \right)^2 P(z) T(z) P_{Laser} A_{Tel} \Delta z \Delta t \cos \theta$</td>
<td></td>
</tr>
<tr>
<td>$\sigma_{Ray}(532 nm) \rho(z)$</td>
<td>$5.458 \times 10^{-6} \frac{P(z)}{T(z)}$</td>
<td></td>
</tr>
<tr>
<td>$\gamma_{Mole}$</td>
<td>$4.054 \times 10^3 (MHz)^2 / K$</td>
<td></td>
</tr>
</tbody>
</table>

Units for parameters in signal level formula; $z$(km), $P$(mb), $T$(K), $P_{Laser}$(W), $A_{Tel}$(m$^2$), $\Delta z$(m), $\Delta t$(s)

**Laser and Receiver Frequency Errors:** The receiver filters are temperature and pressure stabilized Fabry-Perot etalons that are locked to a CW reference laser. The reference laser frequency is nominally the same as the pulsed laser that is used to probe the atmosphere. The frequency jitter of the filter tuning ($\Delta f_{Filter}$) is different for each filter. The reference laser is also used to monitor the frequency of pulsed laser spectrum by applying the same heterodyne techniques that are employed to measure the pulse spectrum of RF lidars [e.g. Fee et al., 1992 and Chu et al., 2010]. For Rayleigh Doppler lidars where each filter channel processes a selected portion of the same backscattered signal, the effects of the uncorrected laser frequency and line width errors caused by chirp ($\Delta f_{Chirp}$ and $\Delta \sigma_{Laser}$) are identical for each of the three channels.

The temperature, wind and atmospheric density errors caused by filter tuning errors and laser tuning and linewidth errors are...
\[ \Delta T_{\text{Kin}}^{\text{rms}} = \frac{\sqrt{2}}{f_s} \left( \Delta f_{\text{Filter}}^2 + 2 \alpha^2 \frac{\sigma_{\text{Laser}}^2}{f_s^2} \Delta \sigma_{\text{Laser}}^2 \right)^{1/2} T(z) \]

\[ \Delta T_{\text{Hyd}}^{\text{rms}} = 0 \]

\[ \Delta w_{\text{rms}} = \frac{\lambda_{\text{Laser}}}{2} \sqrt{\Delta f_{\text{Chirp}}^2 + \Delta f_{\text{Filter}}^2 / 2} \]

\[ \Delta \rho_{\text{rms}} = \frac{(1 - \chi)}{(1 - \chi) + \chi e^{\alpha^2/2} f_s} \left\{ \frac{\Delta f_{\text{Chirp}}}{2} + \left[ \alpha - 1 - \frac{\chi e^{\alpha^2/2}}{(1 - \chi)} \right] \frac{\sigma_{\text{Laser}}^2}{f_s} \Delta \sigma_{\text{Laser}}^2 \right\}^{1/2} \rho(z) \]

Because these errors affect the whole signal profile detected by each of the filter channels, they do not affect the relative atmospheric density profile nor do they affect the hydrostatic temperature, which is derived from the relative density. We assume that state-of-the-art techniques are employed to lock the receiver filters to the reference laser and to monitor the laser pulse spectrum. For this analysis we make the conservative assumption that the filter tuning errors and the residual errors in the laser frequency and linewidth are

\[ \Delta f_{\text{Filter}} = 500 \text{kHz} \]

\[ \Delta f_{\text{Chirp}} = 100 \text{kHz} \]

\[ \Delta \sigma_{\text{Laser}} = 500 \text{kHz} \]

These values represent the residual errors in the laser spectrum and filter tuning after averaging for several tens of seconds.

This 2-filter Doppler Rayleigh lidar is a suboptimum design but does perform within a few dB of the ideal receiver with optimum signal processing. The ideal receiver, which cannot be realized with existing technology, would measure precisely the frequency of each detected photon. The temperature and wind errors for the ideal receiver with a signal level \( \bar{N}_{\text{mol}} \) and assuming \( \bar{N}_b = \bar{N}_D = 0 \), are

**Ideal Receiver with Optimum Signal Processing and \( \bar{N}_b = \bar{N}_D = 0 \)**

\[ \Delta T_{\text{Kin}}^{\text{rms}} = \frac{\sqrt{2}}{\sqrt{\bar{N}_{\text{mol}}(z)}} \frac{T(z)}{\sqrt{\text{SNR}(z)}} \]

\[ \Delta T_{\text{Hyd}}^{\text{rms}} = \frac{T(z)}{\sqrt{\bar{N}_{\text{mol}}(z)}} = \frac{T(z)}{\sqrt{\text{SNR}(z)}} \]

\[ \Delta w_{\text{rms}} = \frac{\lambda_{\text{Laser}}}{2} \sqrt{\sigma_{\text{mol}}^2 + \sigma_{\text{Laser}}^2} \frac{1}{\sqrt{\bar{N}_{\text{mol}}(z)}} \]

To achieve the same hydrostatic temperature accuracy as the ideal receiver, the 2-filter receiver requires a factor of \((1.527)^2 = 2.332 \text{ (3.7 dB)}\) more signal. To achieve the same vertical wind accuracy, the 2-filter receiver requires a factor of \((2 \cdot 1.080)^2 = 4.666 \text{ (6.7 dB)}\) more signal. Other suboptimal approaches for measuring temperature and winds may perform better than the 2-filter receiver, but the improvement is not expected to be significant given that the 2-filter receiver performance is so close to that of the ideal receiver. Hence, the 2-filter system analyzed here provides a realistic assessment of the performance capabilities of a well-engineered Doppler Rayleigh lidar, regardless of how the backscattered signal is detected and processed by the receiver.
Appendix C - Wind and Temperature Accuracy Calculations

We assume horizontal winds are measured by pointing two beams 30 degrees off zenith to the north and east. The SNRs and the wind and temperature errors of the Rayleigh, Na, Fe and He(2^3S) lidars were computed using the formulas listed in Tables A2, A3, A4 and B1. The nominal SNR of the Rayleigh lidar versus altitude between 30 and 150 km altitude and the accuracy of the derived horizontal wind and temperature measurements are tabulated in Table C1. For altitudes greater than about 100 km, the Rayleigh lidar SNR is not sufficient to observe the shorter wavelength and shorter period waves so we increase Δz and Δt to improve the SNR and the precision of the derived parameters. The pressure and temperature values were obtained from the 1976 U.S. Standard Atmosphere. Because the Rayleigh signal levels are so weak above 100 km, even for LALO, we expect the Rayleigh lidar will be used primarily for observations below 80 km. Note that above 100 km analysis of the molecular backscattered signal must account for gravitational separation of the atmospheric constituents.

Table C1. Horizontal Wind and Temperature Measurements 30-150 km

<table>
<thead>
<tr>
<th>z(km)</th>
<th>T(K)</th>
<th>P(mb)</th>
<th>SNR_{Ray}</th>
<th>Δu_{rms}(m/s)</th>
<th>ΔT_{rms}(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Δz = 10 km - Δt = 10 h</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>150</td>
<td>634</td>
<td>4.5x10^6</td>
<td>785</td>
<td>36</td>
<td>35</td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>7.2x10^6</td>
<td>3200</td>
<td>18</td>
<td>15</td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>1.3x10^6</td>
<td>1.7x10^4</td>
<td>7.6</td>
<td>5.5</td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>2.5x10^6</td>
<td>9.6x10^3</td>
<td>3.2</td>
<td>1.8</td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>7.1x10^5</td>
<td>8.0x10^3</td>
<td>1.1</td>
<td>0.41</td>
</tr>
</tbody>
</table>

| Δz = 500 m - Δt = 2.5 min |
| 100   | 195  | 3.2x10^4 | 1300    | 28           | 8.3         |
| 90    | 187  | 1.8x10^3 | 9600    | 10           | 2.9         |
| 80    | 199  | 0.011    | 7.0x10^4 | 3.8          | 1.1         |
| 70    | 220  | 0.052    | 3.9x10^7 | 1.6          | 0.54        |
| 60    | 247  | 0.22     | 2.0x10^9 | 0.73         | 0.27        |
| 50    | 271  | 0.80     | 9.6x10^9 | 0.38         | 0.13        |
| 40    | 250  | 2.9      | 5.9x10^7 | 0.24         | 0.05        |
| 30    | 227  | 12       | 4.8x10^8 | 0.20         | 0.02        |

Tables C2 and C3 include the nominal SNRs of the Na and Fe Doppler lidars, respectively, between 75 and 150 km and the accuracies of the derived horizontal wind and temperature measurements. The Na and Fe densities were obtained from various mid- and high-latitude observations reported in the literature [see references in Gardner, 2004] including observations of the tenuous topside metal layers [Höffner and Friedman, 2004 and 2005; Raizada et al., 2004] and sporadic layers well into the mesosphere [Chu et al., 2011; Lübken et al., 2011; Wang et al., 2012; Tsuda et al., 2012; Huang et al., 2013; Friedman et al., 2013]. While the main Na and Fe layers are persistent between about 80 and 105 km, the abundances vary seasonally. The values listed represent the nominal seasonal means at mid-latitudes. Above 105 km, the tenuous metal densities vary substantially and in fact can sometimes be much smaller or larger than the values listed. The tabulated values represent the nominal mean densities for the sporadic, but relatively common, topside thermospheric metal layers reported in the above references.
Table C2. Horizontal Wind and Temperature Measurements 75-150 km
2-Beam Na Doppler Lidar - $\theta=30^\circ$ - $P_{\text{Laser}}A_{\text{Tele}} = 250 \text{ Wm}^2 / \text{Beam}$

$\Delta z = 500 \text{ m} - \Delta t = 2.5 \text{ min}$

<table>
<thead>
<tr>
<th>$z$ (km)</th>
<th>$T$ (K)</th>
<th>$\rho_{\text{Na}}$ (cm$^{-3}$)$^1$</th>
<th>SNR$_{\text{Na}}$</th>
<th>$\Delta u_{\text{rms}}$ (m/s)</th>
<th>$\Delta T_{\text{rms}}$ (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>634</td>
<td>0.15</td>
<td>2125</td>
<td>16.50</td>
<td>27</td>
</tr>
<tr>
<td>145</td>
<td>599</td>
<td>0.25</td>
<td>3825</td>
<td>12.40</td>
<td>19</td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>0.5</td>
<td>8250</td>
<td>8.40</td>
<td>12</td>
</tr>
<tr>
<td>135</td>
<td>517</td>
<td>1</td>
<td>1.8x10$^4$</td>
<td>5.70</td>
<td>7.5</td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>2.0</td>
<td>3.8x10$^4$</td>
<td>3.90</td>
<td>4.7</td>
</tr>
<tr>
<td>125</td>
<td>417</td>
<td>3.5</td>
<td>7.3x10$^4$</td>
<td>2.85</td>
<td>3.0</td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>7</td>
<td>1.6x10$^4$</td>
<td>1.95</td>
<td>1.8</td>
</tr>
<tr>
<td>115</td>
<td>300</td>
<td>15</td>
<td>3.7x10$^4$</td>
<td>1.30</td>
<td>0.96</td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>30</td>
<td>8.0x10$^4$</td>
<td>0.87</td>
<td>0.52</td>
</tr>
<tr>
<td>105</td>
<td>209</td>
<td>85</td>
<td>2.5x10$^4$</td>
<td>0.51</td>
<td>0.26</td>
</tr>
<tr>
<td>100</td>
<td>195</td>
<td>550</td>
<td>1.8x10$^4$</td>
<td>0.24</td>
<td>0.010</td>
</tr>
<tr>
<td>95</td>
<td>188</td>
<td>2500</td>
<td>9.0x10$^4$</td>
<td>0.18</td>
<td>0.057</td>
</tr>
<tr>
<td>90</td>
<td>187</td>
<td>3300</td>
<td>1.3x10$^4$</td>
<td>0.17</td>
<td>0.061</td>
</tr>
<tr>
<td>85</td>
<td>189</td>
<td>1300</td>
<td>5.8x10$^4$</td>
<td>0.19</td>
<td>0.064</td>
</tr>
<tr>
<td>80</td>
<td>199</td>
<td>150</td>
<td>7.6x10$^4$</td>
<td>0.32</td>
<td>0.15</td>
</tr>
<tr>
<td>75</td>
<td>208</td>
<td>5</td>
<td>2.9x10$^4$</td>
<td>1.45</td>
<td>0.75</td>
</tr>
</tbody>
</table>

$^1$Na scale height = 7.5 km from 110-150 km

Table C3. Horizontal Wind and Temperature Measurements 75-150 km
2-Beam Fe Doppler Lidar - $\theta=30^\circ$ - $P_{\text{Laser}}A_{\text{Tele}} = 250 \text{ Wm}^2 / \text{Beam}$

$\Delta z = 500 \text{ m} - \Delta t = 2.5 \text{ min}$

<table>
<thead>
<tr>
<th>$z$ (km)</th>
<th>$T$ (K)</th>
<th>$\rho_{\text{Fe}}$ (cm$^{-3}$)$^1$</th>
<th>SNR$_{\text{Fe}}$</th>
<th>$\Delta u_{\text{rms}}$ (m/s)</th>
<th>$\Delta T_{\text{rms}}$ (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>634</td>
<td>10</td>
<td>7200</td>
<td>5.75</td>
<td>15</td>
</tr>
<tr>
<td>145</td>
<td>599</td>
<td>15</td>
<td>1.2x10$^4$</td>
<td>4.50</td>
<td>11</td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>20</td>
<td>1.7x10$^4$</td>
<td>3.80</td>
<td>8.5</td>
</tr>
<tr>
<td>135</td>
<td>517</td>
<td>25</td>
<td>2.2x10$^4$</td>
<td>3.25</td>
<td>6.7</td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>40</td>
<td>3.8x10$^4$</td>
<td>2.50</td>
<td>4.7</td>
</tr>
<tr>
<td>125</td>
<td>417</td>
<td>55</td>
<td>5.7x10$^4$</td>
<td>2.05</td>
<td>3.4</td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>75</td>
<td>8.5x10$^4$</td>
<td>1.65</td>
<td>2.4</td>
</tr>
<tr>
<td>115</td>
<td>300</td>
<td>105</td>
<td>1.3x10$^5$</td>
<td>1.35</td>
<td>1.6</td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>145</td>
<td>1.9x10$^5$</td>
<td>1.10</td>
<td>1.1</td>
</tr>
<tr>
<td>105</td>
<td>209</td>
<td>300</td>
<td>4.4x10$^5$</td>
<td>0.74</td>
<td>0.61</td>
</tr>
<tr>
<td>100</td>
<td>195</td>
<td>800</td>
<td>1.3x10$^6$</td>
<td>0.44</td>
<td>0.33</td>
</tr>
<tr>
<td>95</td>
<td>188</td>
<td>3000</td>
<td>5.4x10$^6$</td>
<td>0.23</td>
<td>0.16</td>
</tr>
<tr>
<td>90</td>
<td>187</td>
<td>8250</td>
<td>1.7x10$^6$</td>
<td>0.15</td>
<td>0.095</td>
</tr>
<tr>
<td>85</td>
<td>189</td>
<td>6750</td>
<td>1.5x10$^6$</td>
<td>0.16</td>
<td>0.10</td>
</tr>
<tr>
<td>80</td>
<td>199</td>
<td>1500</td>
<td>3.8x10$^6$</td>
<td>0.27</td>
<td>0.20</td>
</tr>
<tr>
<td>75</td>
<td>208</td>
<td>25</td>
<td>7.2x10$^6$</td>
<td>1.80</td>
<td>1.5</td>
</tr>
</tbody>
</table>

$^1$Fe scale height = 15 km from 110-150 km
Listed in Table C4 are the nominal SNR of the He(2$^3$S) Doppler lidar between 300 and 1000 km and the accuracies of the derived horizontal wind and temperature measurements. Dr. Lara Waldrop and colleagues provided the metastable He densities which they derived from model calculations [Waldrop et al., 2005; Bishop and Link, 1999]. They represent the typical values expected at mid-latitude during winter.

Table C4. Horizontal Wind and Temperature Measurements 300-1000 km
2-Beam He(2$^3$S) Doppler Lidar - $\theta$=30$^\circ$ - $P_{\text{Laser, Tele}}$ = 500 Wm$^2$ / Beam

<table>
<thead>
<tr>
<th>z(km)</th>
<th>T(K)</th>
<th>$\rho_{\text{He}}$(m$^{-3}$)</th>
<th>SNR$_{\text{Fe}}$</th>
<th>$\Delta u_{\text{rms}}$(m/s)</th>
<th>$\Delta T_{\text{rms}}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>1000</td>
<td>3.1x10$^4$</td>
<td>1.5x10$^5$</td>
<td>34</td>
<td>16</td>
</tr>
<tr>
<td>950</td>
<td>1000</td>
<td>3.8x10$^4$</td>
<td>2.0x10$^5$</td>
<td>30</td>
<td>14</td>
</tr>
<tr>
<td>900</td>
<td>1000</td>
<td>4.6x10$^4$</td>
<td>2.7x10$^5$</td>
<td>25</td>
<td>12</td>
</tr>
<tr>
<td>850</td>
<td>1000</td>
<td>5.6x10$^4$</td>
<td>3.8x10$^5$</td>
<td>22</td>
<td>10</td>
</tr>
<tr>
<td>800</td>
<td>1000</td>
<td>6.6x10$^4$</td>
<td>5.0x10$^5$</td>
<td>19</td>
<td>8.7</td>
</tr>
<tr>
<td>750</td>
<td>1000</td>
<td>7.8x10$^4$</td>
<td>6.7x10$^5$</td>
<td>16</td>
<td>7.5</td>
</tr>
<tr>
<td>700</td>
<td>1000</td>
<td>9.1x10$^4$</td>
<td>9.0x10$^5$</td>
<td>14</td>
<td>6.5</td>
</tr>
<tr>
<td>650</td>
<td>1000</td>
<td>1.0x10$^5$</td>
<td>1.2x10$^5$</td>
<td>12</td>
<td>5.7</td>
</tr>
<tr>
<td>600</td>
<td>1000</td>
<td>1.1x10$^5$</td>
<td>1.5x10$^5$</td>
<td>11</td>
<td>5.0</td>
</tr>
<tr>
<td>550</td>
<td>1000</td>
<td>1.1x10$^5$</td>
<td>1.8x10$^5$</td>
<td>10</td>
<td>4.6</td>
</tr>
<tr>
<td>500</td>
<td>999</td>
<td>1.0x10$^5$</td>
<td>1.9x10$^5$</td>
<td>9.6</td>
<td>4.4</td>
</tr>
<tr>
<td>450</td>
<td>998</td>
<td>7.8x10$^4$</td>
<td>1.9x10$^5$</td>
<td>9.8</td>
<td>4.5</td>
</tr>
<tr>
<td>400</td>
<td>996</td>
<td>4.3x10$^5$</td>
<td>1.3x10$^5$</td>
<td>12</td>
<td>5.3</td>
</tr>
<tr>
<td>350</td>
<td>990</td>
<td>1.5x10$^5$</td>
<td>6.0x10$^5$</td>
<td>17</td>
<td>7.9</td>
</tr>
<tr>
<td>300</td>
<td>976</td>
<td>2.7x10$^5$</td>
<td>1.4x10$^5$</td>
<td>35</td>
<td>16</td>
</tr>
</tbody>
</table>

We assume vertical winds are measured using a single beam pointed at zenith. Tabulated in Table C5 is the nominal SNR of the Rayleigh lidar versus altitude between 30 and 150 km altitude and the accuracy of the derived vertical wind and temperature measurements. For altitudes greater than about 100 km, the Rayleigh SNR will not be sufficient to observe the shorter wavelength and shorter period waves. Consequently, $\Delta z$ and $\Delta t$ have been increased to improve the SNR and the precision of the derived parameters. Similar data for the Na, Fe and He(2$^3$S) lidars are tabulated in Tables C6-C8.
Table C5. Vertical Wind and Temperature Measurements 30-150 km
Single Zenith Beam Rayleigh Doppler Lidar - $\theta=0^\circ$ - $P_{L}A_{T}=5000 \text{ Wm}^2$

<table>
<thead>
<tr>
<th>z(km)</th>
<th>T(K)</th>
<th>P(mb)</th>
<th>SNR$_{Ray}$</th>
<th>$\Delta w_{rms}$(m/s)</th>
<th>$\Delta T_{rms}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>634</td>
<td>4.5x10$^{-6}$</td>
<td>1.4x10$^5$</td>
<td>4.3</td>
<td>8.3</td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>7.2x10$^{-6}$</td>
<td>4.7x10$^5$</td>
<td>2.3</td>
<td>4.0</td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>1.3x10$^{-6}$</td>
<td>1.8x10$^5$</td>
<td>1.2</td>
<td>1.7</td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>2.5x10$^{-6}$</td>
<td>7.2x10$^5$</td>
<td>0.60</td>
<td>0.65</td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>7.1x10$^{-6}$</td>
<td>4.2x10$^5$</td>
<td>0.26</td>
<td>0.18</td>
</tr>
</tbody>
</table>

$\Delta z = 10 \text{ km} - \Delta t = 10 \text{ h}$

<table>
<thead>
<tr>
<th>z(km)</th>
<th>T(K)</th>
<th>$P_{Na}$(cm$^{-3}$)$^a$</th>
<th>SNR$_{Na}$</th>
<th>$\Delta w_{rms}$(m/s)</th>
<th>$\Delta T_{rms}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>100</td>
<td>195</td>
<td>3.2x10$^{-4}$</td>
<td>6125</td>
<td>6.4</td>
<td>3.8</td>
</tr>
<tr>
<td>90</td>
<td>187</td>
<td>1.8x10$^{-5}$</td>
<td>4.5x10$^4$</td>
<td>2.4</td>
<td>1.4</td>
</tr>
<tr>
<td>80</td>
<td>199</td>
<td>0.011</td>
<td>3.2x10$^4$</td>
<td>0.88</td>
<td>0.53</td>
</tr>
<tr>
<td>70</td>
<td>220</td>
<td>0.052</td>
<td>1.8x10$^5$</td>
<td>0.38</td>
<td>0.25</td>
</tr>
<tr>
<td>60</td>
<td>247</td>
<td>0.22</td>
<td>9.3x10$^6$</td>
<td>0.19</td>
<td>0.12</td>
</tr>
<tr>
<td>50</td>
<td>271</td>
<td>0.80</td>
<td>4.4x10$^5$</td>
<td>0.12</td>
<td>0.062</td>
</tr>
<tr>
<td>40</td>
<td>250</td>
<td>2.9</td>
<td>2.7x10$^6$</td>
<td>0.10</td>
<td>0.023</td>
</tr>
<tr>
<td>30</td>
<td>227</td>
<td>12</td>
<td>2.2x10$^5$</td>
<td>0.099</td>
<td>0.007</td>
</tr>
</tbody>
</table>

$\Delta z = 500 \text{ m} - \Delta t = 2.5 \text{ min}$

Table C6. Vertical Wind and Temperature Measurements 75-150 km
Single Zenith Beam Na Doppler Lidar - $\theta=0^\circ$ - $P_{L}A_{T}=1000 \text{ Wm}^2$ / Beam

<table>
<thead>
<tr>
<th>z(km)</th>
<th>T(K)</th>
<th>$P_{Na}$(cm$^{-3}$)$^a$</th>
<th>SNR$_{Na}$</th>
<th>$\Delta w_{rms}$(m/s)</th>
<th>$\Delta T_{rms}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>634</td>
<td>0.15</td>
<td>1.0x10$^5$</td>
<td>3.8</td>
<td>12</td>
</tr>
<tr>
<td>145</td>
<td>599</td>
<td>0.25</td>
<td>1.8x10$^5$</td>
<td>2.9</td>
<td>8.7</td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>0.5</td>
<td>3.8x10$^5$</td>
<td>1.9</td>
<td>5.6</td>
</tr>
<tr>
<td>135</td>
<td>517</td>
<td>1</td>
<td>8.2x10$^5$</td>
<td>1.3</td>
<td>3.5</td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>2.0</td>
<td>1.8x10$^5$</td>
<td>0.91</td>
<td>2.2</td>
</tr>
<tr>
<td>125</td>
<td>417</td>
<td>3.5</td>
<td>3.4x10$^5$</td>
<td>0.66</td>
<td>1.4</td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>7</td>
<td>7.3x10$^5$</td>
<td>0.45</td>
<td>0.82</td>
</tr>
<tr>
<td>115</td>
<td>300</td>
<td>15</td>
<td>1.7x10$^5$</td>
<td>0.30</td>
<td>0.45</td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>30</td>
<td>3.7x10$^5$</td>
<td>0.21</td>
<td>0.25</td>
</tr>
<tr>
<td>105</td>
<td>209</td>
<td>85</td>
<td>1.2x10$^5$</td>
<td>0.14</td>
<td>0.13</td>
</tr>
<tr>
<td>100</td>
<td>195</td>
<td>550</td>
<td>8.3x10$^5$</td>
<td>0.090</td>
<td>0.060</td>
</tr>
<tr>
<td>95</td>
<td>188</td>
<td>2500</td>
<td>4.2x10$^5$</td>
<td>0.081</td>
<td>0.045</td>
</tr>
<tr>
<td>90</td>
<td>187</td>
<td>3300</td>
<td>6.1x10$^5$</td>
<td>0.081</td>
<td>0.044</td>
</tr>
<tr>
<td>85</td>
<td>189</td>
<td>1300</td>
<td>2.7x10$^5$</td>
<td>0.082</td>
<td>0.047</td>
</tr>
<tr>
<td>80</td>
<td>199</td>
<td>150</td>
<td>3.5x10$^5$</td>
<td>0.10</td>
<td>0.079</td>
</tr>
<tr>
<td>75</td>
<td>208</td>
<td>5</td>
<td>1.3x10$^5$</td>
<td>0.34</td>
<td>0.35</td>
</tr>
</tbody>
</table>

$^a$Na scale height = 7.5 km from 110-150 km
Table C7. Vertical Wind and Temperature Measurements 75-150 km
Single Zenith Beam Fe Doppler Lidar - $\theta = 0^\circ$ - $P_{Laser}A_{Tele} = 1000$ Wm$^2$/Beam

$\Delta z = 500$ m - $\Delta t = 2.5$ min

<table>
<thead>
<tr>
<th>$z$(km)</th>
<th>$T$(K)</th>
<th>$\rho_{Fe}$(cm$^{-3}$)</th>
<th>SNR$_{Fe}$</th>
<th>$\Delta w_{rms}$(m/s)</th>
<th>$\Delta T_{rms}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>634</td>
<td>$10 \times 3.3 \times 10^4$</td>
<td>1.3</td>
<td>6.8</td>
<td></td>
</tr>
<tr>
<td>145</td>
<td>599</td>
<td>$15 \times 5.3 \times 10^4$</td>
<td>1.1</td>
<td>5.0</td>
<td></td>
</tr>
<tr>
<td>140</td>
<td>560</td>
<td>$20 \times 7.6 \times 10^4$</td>
<td>0.88</td>
<td>3.9</td>
<td></td>
</tr>
<tr>
<td>135</td>
<td>517</td>
<td>$25 \times 1.0 \times 10^5$</td>
<td>0.76</td>
<td>3.1</td>
<td></td>
</tr>
<tr>
<td>130</td>
<td>469</td>
<td>$40 \times 1.8 \times 10^5$</td>
<td>0.59</td>
<td>2.2</td>
<td></td>
</tr>
<tr>
<td>125</td>
<td>417</td>
<td>$55 \times 2.6 \times 10^5$</td>
<td>0.48</td>
<td>1.6</td>
<td></td>
</tr>
<tr>
<td>120</td>
<td>360</td>
<td>$75 \times 3.9 \times 10^5$</td>
<td>0.39</td>
<td>1.1</td>
<td></td>
</tr>
<tr>
<td>115</td>
<td>300</td>
<td>$105 \times 6.0 \times 10^5$</td>
<td>0.32</td>
<td>0.76</td>
<td></td>
</tr>
<tr>
<td>110</td>
<td>240</td>
<td>$145 \times 9.0 \times 10^5$</td>
<td>0.26</td>
<td>0.49</td>
<td></td>
</tr>
<tr>
<td>105</td>
<td>209</td>
<td>$300 \times 2.0 \times 10^5$</td>
<td>0.18</td>
<td>0.29</td>
<td></td>
</tr>
<tr>
<td>100</td>
<td>195</td>
<td>$800 \times 6.0 \times 10^5$</td>
<td>0.11</td>
<td>0.16</td>
<td></td>
</tr>
<tr>
<td>95</td>
<td>188</td>
<td>$3000 \times 2.5 \times 10^5$</td>
<td>0.067</td>
<td>0.080</td>
<td></td>
</tr>
<tr>
<td>90</td>
<td>187</td>
<td>$8250 \times 7.6 \times 10^5$</td>
<td>0.054</td>
<td>0.052</td>
<td></td>
</tr>
<tr>
<td>85</td>
<td>189</td>
<td>$6750 \times 7.0 \times 10^5$</td>
<td>0.054</td>
<td>0.054</td>
<td></td>
</tr>
<tr>
<td>80</td>
<td>199</td>
<td>$1500 \times 1.8 \times 10^5$</td>
<td>0.074</td>
<td>0.098</td>
<td></td>
</tr>
<tr>
<td>75</td>
<td>208</td>
<td>$25 \times 3.3 \times 10^5$</td>
<td>0.42</td>
<td>0.70</td>
<td></td>
</tr>
</tbody>
</table>

Fe scale height = 15 km from 110-150 km

Table C8. Vertical Wind and Temperature Measurements 300-1000 km
Single Zenith Beam He(2$^3$S) Doppler Lidar - $\theta = 0^\circ$ - $P_{Laser}A_{Tele} = 2000$ Wm$^2$

$\Delta z = 25$ km - $\Delta t = 5$ min

<table>
<thead>
<tr>
<th>$z$(km)</th>
<th>$T$(K)</th>
<th>$\rho_{He}$(m$^{-3}$)</th>
<th>SNR$_{Fe}$</th>
<th>$\Delta w_{rms}$(m/s)</th>
<th>$\Delta T_{rms}$(K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>1000</td>
<td>$3.1 \times 10^5$</td>
<td>7.0$\times 10^5$</td>
<td>8.0</td>
<td>7.4</td>
</tr>
<tr>
<td>950</td>
<td>1000</td>
<td>$3.8 \times 10^5$</td>
<td>9.5$\times 10^5$</td>
<td>6.7</td>
<td>6.3</td>
</tr>
<tr>
<td>900</td>
<td>1000</td>
<td>$4.6 \times 10^5$</td>
<td>1.3$\times 10^5$</td>
<td>5.9</td>
<td>5.4</td>
</tr>
<tr>
<td>850</td>
<td>1000</td>
<td>$5.6 \times 10^5$</td>
<td>1.7$\times 10^5$</td>
<td>5.1</td>
<td>4.7</td>
</tr>
<tr>
<td>800</td>
<td>1000</td>
<td>$6.6 \times 10^5$</td>
<td>2.3$\times 10^5$</td>
<td>4.4</td>
<td>4.0</td>
</tr>
<tr>
<td>750</td>
<td>1000</td>
<td>$7.8 \times 10^5$</td>
<td>3.1$\times 10^5$</td>
<td>3.8</td>
<td>3.5</td>
</tr>
<tr>
<td>700</td>
<td>1000</td>
<td>$9.1 \times 10^5$</td>
<td>4.2$\times 10^5$</td>
<td>3.3</td>
<td>3.0</td>
</tr>
<tr>
<td>650</td>
<td>1000</td>
<td>$1.0 \times 10^6$</td>
<td>5.3$\times 10^5$</td>
<td>2.9</td>
<td>2.7</td>
</tr>
<tr>
<td>600</td>
<td>1000</td>
<td>$1.1 \times 10^6$</td>
<td>6.9$\times 10^5$</td>
<td>2.5</td>
<td>2.3</td>
</tr>
<tr>
<td>550</td>
<td>1000</td>
<td>$1.1 \times 10^6$</td>
<td>8.2$\times 10^5$</td>
<td>2.3</td>
<td>2.2</td>
</tr>
<tr>
<td>500</td>
<td>999</td>
<td>$1.0 \times 10^6$</td>
<td>9.0$\times 10^5$</td>
<td>2.2</td>
<td>2.0</td>
</tr>
<tr>
<td>450</td>
<td>998</td>
<td>$7.8 \times 10^5$</td>
<td>8.7$\times 10^5$</td>
<td>2.3</td>
<td>2.1</td>
</tr>
<tr>
<td>400</td>
<td>996</td>
<td>$4.3 \times 10^5$</td>
<td>6.0$\times 10^5$</td>
<td>2.7</td>
<td>2.5</td>
</tr>
<tr>
<td>350</td>
<td>990</td>
<td>$1.5 \times 10^5$</td>
<td>2.8$\times 10^5$</td>
<td>4.0</td>
<td>3.7</td>
</tr>
<tr>
<td>300</td>
<td>976</td>
<td>$2.7 \times 10^5$</td>
<td>6.7$\times 10^5$</td>
<td>8.1</td>
<td>7.3</td>
</tr>
</tbody>
</table>
Appendix D – Eddy Diffusivity Accuracy Calculations

The eddy diffusivity \( k_{zz} \) is a fundamental atmospheric parameter that is used to assess vertical constituent transport associated with turbulent mixing. The eddy transport of all species at a particular point and time is governed by the same eddy diffusion coefficient. The measured constituent flux and diffusivity are related as follows [Colegrove et al., 1966]

\[
\frac{\partial}{\partial z} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right) = -k_{zz} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right)
\]

\[\text{(D-1)}\]

where the sample means on both sides of the equation are taken over identical time periods and altitude intervals \((L \text{ and } \tau)\). In most cases the eddy fluxes will be affected by gravity waves propagating through the region, because breaking waves generate turbulence, and because non-breaking waves distort the constituent profile and the magnitude of the eddy mixing. Because of the very high SNRs for the LADO lidars, the eddy fluxes can be derived at the high spatial and temporal resolutions sufficient to observe these gravity wave effects. In this case (D-1) may be written in the form

\[
\frac{\partial}{\partial z} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right) = -k_{zz} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right)
\]

\[\text{(D-2)}\]

To account for these gravity wave perturbations when calculating the error in the derived eddy diffusivity, we use the rms value of the right-hand side of (D-2).

\[
\Delta k_{zz} = \frac{\Delta \bar{p}_C}{\bar{H}} \frac{\Delta \bar{p}_C}{\bar{T}} \frac{\Delta \bar{p}_C}{\partial z} \frac{\Delta \bar{p}_C}{\partial z} \frac{\Delta \bar{p}_C}{\partial z}
\]

\[\Delta k_{zz} = \frac{\Delta \bar{p}_C}{\bar{H}} \frac{\Delta \bar{p}_C}{\bar{T}} \frac{\Delta \bar{p}_C}{\partial z} \frac{\Delta \bar{p}_C}{\partial z} \frac{\Delta \bar{p}_C}{\partial z} \]

\[\text{(D-3)}\]

The derivation of the mean-square value of the scale factor in (D-3) is involved but straightforward. The result, calculated by using the zeroth- and first-order perturbation terms, is

\[
\frac{\partial}{\partial z} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right) = -k_{zz} \left( \frac{\bar{p}_C}{\bar{H}} + \frac{\bar{p}_C}{\bar{T}} \frac{\partial \bar{T}}{\partial z} + \frac{\partial \bar{p}_C}{\partial z} \right)
\]

\[\text{(D-4)}\]

where \( \bar{H} = R \bar{T} / g = 30.2 \bar{T} = 5.6 km \) is the pressure scale height of the atmosphere, \( \Gamma_{ad} = g / C_p = 9.5 K / km \) is the adiabatic lapse rate, \( g = 9.5 m / s^2 \) is the acceleration of gravity and \( R = 287 m^2 / K / s^2 \) is the universal gas constant. At mesopause heights the lapse rate variance is about \( Var(\bar{T}_{GW} / \partial z) = 30 K^2 / km^2 \) and the temperature variance is about \( Var(\bar{T}_{GW}) = 60 K^2 \) [Gardner and Liu, 2007]. The mean-square scale factor was evaluated using these parameters and the Na and Fe density profiles tabulated in Table C2 and C3. The results were used in (D-3) along with (18) to calculate the errors for the derived eddy diffusivity profile that are tabulated in Table 6.
The heat flux is related to the eddy coefficient for thermal diffusion \((k_H)\) as follows

\[
\overline{w'_T \theta'_T} = -k_H \overline{\theta \frac{\partial T}{\partial z}} = -k_H \frac{(\Gamma_{ad} + \frac{\partial \overline{T}}{\partial z})}{\overline{T}}
\]

where \(\theta\) is the potential temperature. Because we are interested in measuring the eddy heat flux and thermal diffusivity at high resolution, like we did for \(k_{zz}\), we evaluate the error by taking into account gravity wave perturbations.

\[
\Delta k_H = \frac{\Delta w'_T \theta'_T}{\Delta T}\n\]

The mean-square scale factor was evaluated using the temperature profiles tabulated in Table C2 and C3. The results were used in (D-6) along with (18) to calculate the errors for the derived eddy thermal diffusivity profiles that are tabulated in Table 5.

### References


layers and fast gravity waves in the thermosphere (110-155 km) at McMurdo (77.8°S, 166.7°E), Antarctica, Geophys. Res. Lett., 38(23), L23807, doi:10.1029/2011GL050016.

Chu, X., and W. Huang (2010), Fe Doppler-free spectroscopy and optical heterodyne detection for accurate frequency control of Fe-resonance Doppler lidar, Proc. 25th ILRC, St. Petersburg, Russia.


Tsuda, T. T., X. Chu, T. Nakamura, M. K. Ejiri, and T. Kawahara (2012), Sodium layer in the thermosphere (110-130 km) observed at Syowa Station (69.0°S, 39.6°E) in Antarctica, American Geophysical Union Fall Meeting, San Francisco, California, 3–7 December 2012.


Rayleigh Lidars

Richard L. Collins
Geophysical Institute and Department of Atmospheric Sciences
University of Alaska Fairbanks
Fairbanks, Alaska

Gerd Baumgarten
Leibnitz Institute for Atmospheric Physics
University of Rostock
Kühlungsborn, Germany

1. Introduction

Rayleigh lidars employ nonresonant Rayleigh scatter to determine density, temperature, and wind profiles in the atmosphere (see papers in collection by Grant et al. (1997)). While Rayleigh scatter is as often described as elastic scatter, in air composed of molecules it is composed of three distinct components: Cabannes scatter, rotational Raman scatter, and vibrational Raman scatter (Young 1981; Miles et al., 2001; She 2001), of which the Cabannes scattering is the elastic component. For nitrogen molecules (N2) at a pressure of 1 atmosphere and a temperature of 300 K, the Cabannes scatter has a half-linewidth of 0.03 cm⁻¹, the rotational Raman scatter consists of a series of lines separated by 12 cm⁻¹, and the vibrational Raman scattering is offset by 2331 cm⁻¹. Thus for laser excitation at 532 nm, the Cabannes scatter has a half linewidth of 0.85 pm, the rotational Raman lines are separated by 340 pm, and the vibrational Raman line is at 607 nm. The intensity of the rotational Raman scatter is less than 1% of the Cabannes scatter, and the intensity of the vibrational Raman scatter is less than 0.1% of the Cabannes scatter. Analyses of Rayleigh scattering from a molecular gas such as air shows that Raman scatter is an integral part of Rayleigh scatter (Young 1981; Miles et al., 2001; She 2001). The principles, implementation, and performance of Rayleigh and Raman lidar systems through the early 2000s are described in two recent reviews (Fuji and Fukuchi, 2005; Weitkamp, 2005).

In this chapter we report the observational capabilities of currently operational Rayleigh lidar systems and the critical laser characteristics, discuss possible improvements in lasers over the next ten years, and finally discuss observational capabilities that would be achievable with a 100 m².

2. Observational Capabilities of Current Rayleigh Lidar Systems

Current Rayleigh lidar systems are divided into two classes: hydrostatic Rayleigh lidar systems that measure density and temperature profiles, and Doppler Rayleigh lidar systems that measure wind.

Hydrostatic Rayleigh Lidar Systems: The Rayleigh lidar operating at Poker Flat Research Range, Chatanika, is a representative contemporary hydrostatic system (Thurairajah et al., 2010; Collins et al., 2011). The lidar signal is proportional to the density of the atmosphere where no aerosols are present. Under the assumption of hydrostatic equilibrium the temperature profile can be determined from the relative density profile. The lidar systems employs a flashlamp-pumped Nd:YAG laser and a 62 cm telescope, and yields temperature and density measurements in the 40-80 km altitude region (Table 1). The measured density and temperature profiles at timescales of minutes and hours (Figure 1 and 2) yield measurements of waves and tides in the middle atmosphere that support studies of the wave-driven circulation. Hydrostatic Rayleigh lidar systems are robust systems that have yielded stable long-term measurements of the temperature structure of the middle atmosphere and revealed significant cooling.
trends in the mesosphere that have been attributed to long-term trends in ozone depletion (Ramaswamy 2001).

Table 1: Hydrostatic Rayleigh Lidar Systems (Thurairajah et al., 2010)

<table>
<thead>
<tr>
<th>Transmitter</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Laser</td>
<td>Nd:YAG</td>
</tr>
<tr>
<td>Wavelength</td>
<td>532 nm</td>
</tr>
<tr>
<td>Repetition Rate</td>
<td>20 Hz</td>
</tr>
<tr>
<td>Peak Power</td>
<td>~40 MW</td>
</tr>
<tr>
<td>Average Power</td>
<td>8 W</td>
</tr>
<tr>
<td>Energy</td>
<td>400 mJ/pulse</td>
</tr>
<tr>
<td>Pulse Length</td>
<td>10 ns</td>
</tr>
<tr>
<td>Tuneability</td>
<td>NO</td>
</tr>
<tr>
<td>Beam Expander</td>
<td>x 10</td>
</tr>
<tr>
<td>Divergence</td>
<td>0.2 mrad</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Receiver</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Aperture</td>
<td>0.60 m</td>
</tr>
<tr>
<td>Detector</td>
<td>PMT</td>
</tr>
<tr>
<td>Bandwidth</td>
<td>0.3 nm</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Measurement Resolution and Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Altitude Range</td>
</tr>
<tr>
<td>Altitude Resolution</td>
</tr>
<tr>
<td>Time Resolution</td>
</tr>
<tr>
<td>Accuracy</td>
</tr>
<tr>
<td>Temperature</td>
</tr>
<tr>
<td>Density</td>
</tr>
</tbody>
</table>
Figure 1. Temperature profiles measured by Rayleigh lidar at Chatanika, Alaska, on the night of 17-18 February 2009. The nightly average temperature measured over the interval 1947-0712 LST (0447-1612 UT, thick dashed line) and the temperature profile measured over the two hour interval 0104-0304 LST (1004-1204 UT, thin solid line) are plotted. The one-sigma temperature errors due to the statistical fluctuations in the lidar signal (dashed lines) and the SPARC February temperature profile (dashed line and solid squares) are also plotted. The linear fit to the topside of the MIL is indicated. From Collins et al., 2011.

Figure 2. Relative density perturbations measured by Rayleigh lidar at Chatanika, Alaska on 10-11 January 2005. The perturbations are spatially band-limited between wavelengths 2 km and 30 km and temporally band-limited between time periods of 30 minutes and 4 hours. The positive values are colored red (0-1%, 1-2%, >2%) and the negative values blue (0 - -1%, -1 - -2%, < -2%). The white contour marks the zero line. From Thurairajah et al., 2012.
Doppler Rayleigh Lidar Systems: The Rayleigh lidars at the Arctic Lidar Observatory for Middle Atmosphere Research (ALOMAR), Andoya, and Heifei Normal University, Heifei, are contemporary Doppler wind lidars reporting wind measurements in the middle atmosphere. Both lidar systems use seed-injected flashlamp-pumped Nd:YAG laser systems. The ALOMAR system uses a stabilized iodine cell to lock the transmitter laser and as a filter to detect Doppler shifts in the backscattered light and determine the line-of-sight wind. The Heifei system employs Fabry-Perot interferometers to lock the transmitter laser and as a filter to detect Doppler shifts in the backscattered light and determine the line-of-sight wind. Both systems yield measurement of the wind profile over periods of hours (Table 2). The wind measurements have been validated with a combination of radiosonde measurements, reanalysis, and radar measurements (Figures 2 and 3). The ALOMAR and Heifei systems have similar power, but the ALOMAR system has a telescope area 16 times larger and accordingly measures winds to higher altitudes.

Figure 3. Temperature, vertical wind, and meridional wind measured by ALOMAR RMR lidar on 17 January 2009 between 17:00 and 19:00 UT. ECMWF data from 18:00 UT (solid) and 12:00, 24:00 UT (dashed) are shown. Simultaneous observations by the collocated Meteor radar are shown. The gray area indicates altitudes with aerosol contribution as measured by the lidar. From Baumgarten 2010.

3. Critical Laser Capabilities

Current Rayleigh lidar systems combine traditional “giant pulse” lasers (> 400 mJ/pulse at > 20 pps) with large telescopes (> 0.5-1.8 m) to yield measurements in the stratosphere and mesosphere. The reliability of current flashlamp-pumped Nd:YAG lasers is a critical component in the robustness of these systems, where the lasers operate in a turnkey fashion with routine maintenance. However, these Nd:YAG lasers are a mature technology with current highly reliable commercial systems operating at pulse energies of 1 J at 30 pps.

4. Laser Characteristics over the Next Decade

While the reliability of flashlamp-pumped Nd:YAG lasers may be expected to improve over the next decade, the fundamental performance of these systems will not necessarily improve in terms of laser pulse energy and pulse repetition rate. However, significant progress is expected in diode-pumped Nd:YAG lasers, with 100’s of W at 10’s of kHz. In 2009 Coherent Inc. of Santa Clara California
announced the “Mamba Green” high-power industrial laser that produced 325 W of power at 532 nm at 10 kHz with operating lifetimes of greater than 25,000 h (Coherent, 2009). Diode-pumped lasers of this class are operating at average powers 10 times that of current flashlamp-pumped lasers. For lidar applications with measurements up to 150 km the maximum pulse repetition rate is 1 kHz. Development of diode-pumped lasers with average powers ~ 100 W with pulse repetition rates of 500 Hz to 1 kHz appears reasonable in the next decade.

5. Operating with 100 m² Receiver Telescopes

A receiver telescope with 100 m² collecting area would improve over the aperture of the Rayleigh lidar systems at Chatanika, Andoya, and Heifei by factors of 330, 40, and 630 respectively. This would extend the scope of these measurements in three distinct ways by allowing measurements at (1) current accuracy at higher resolution (where the increase in resolution would be factors of 18, 6, and 25 respectively), (2) higher accuracy at current resolution, and (3) higher altitudes.

Increasing the resolution of current measurements is critical for accurate measurement of gravity waves and instabilities. For the hydrostatic Rayleigh lidar system this would allow temperature and density measurements at a resolution of 10 min up to 80 km. For Doppler Rayleigh lidar systems this would allow wind measurements at resolutions of 20 min up to 80 km and 3 min up to 40 km. Increasing the accuracy of current measurements is critical for accurate measurements of fluxes where nonlinear propagation of measurement errors can swamp the measurement of the physical flux products. For the hydrostatic Rayleigh lidar system this would allow temperature measurements with errors of 0.5 K and density measurements with errors of 0.2% at 80 km. For the Doppler Rayleigh lidar systems this would allow wind measurements with errors of 2.5 m/s at 80 km and 0.02 m/s up to 40 km respectively.
Table 2. Wind Rayleigh Lidar Systems (Baumgarten 2010, Tang et al. 2012)

<table>
<thead>
<tr>
<th>Transmitter</th>
<th>Nd:YAG seeded with I2-frequency stabilized cw laser</th>
<th>Injection-seeded Nd:YAG</th>
</tr>
</thead>
<tbody>
<tr>
<td>Laser</td>
<td>532 nm</td>
<td>355 nm</td>
</tr>
<tr>
<td>Repetition Rate</td>
<td>30 Hz</td>
<td>30 Hz</td>
</tr>
<tr>
<td>Peak Power</td>
<td>~60 MW</td>
<td></td>
</tr>
<tr>
<td>Average Power</td>
<td>14 W</td>
<td>12 W</td>
</tr>
<tr>
<td>Energy</td>
<td>467 mJ/pulse</td>
<td>400 mJ/pulse</td>
</tr>
<tr>
<td>Pulse Length</td>
<td>8 ns</td>
<td>3-7 ns</td>
</tr>
<tr>
<td>Pulse Width</td>
<td>20 - 70 MHz</td>
<td>200 MHz</td>
</tr>
<tr>
<td>Tuneability</td>
<td>NO</td>
<td>NO</td>
</tr>
<tr>
<td>Seeding Power</td>
<td>50 mW</td>
<td>-</td>
</tr>
<tr>
<td>Beam Expander</td>
<td>x 20</td>
<td>x 10</td>
</tr>
<tr>
<td>Divergence</td>
<td>0.18 mrad</td>
<td>0.1 mrad</td>
</tr>
</tbody>
</table>

| Receiver   | 1.8 m                                           | 0.45 m                  |
| Detector   | APD                                             | PMT                     |
| Bandwidth  | 130 pm at night, 4 pm in day                    | 150 pm                  |

Wind Measurement Resolution

| Altitude Range | 30 - 80 km | 10 - 40 km |
| Altitude Resolution | 2 km       | 500 m     |
| Time Resolution   | 2 h        | 1 h       |
| Accuracy           | 0.6 m/s at 49 km, 20 m/s at 80 km | 0.5 m/s between 5 and 20 km |

Increasing the upper altitude of the lidar measurements requires that the lidar signal overcome the loss in Rayleigh signal with altitude due to the exponential decrease in density of the atmosphere with height. In the upper mesosphere the scale height $h$ is $\sim 8$ km. Thus an increase in telescope aperture by a factor of $2.7^2$ or 7.4 will increase the altitude range of the measurements by 8 km while maintaining their accuracy and precision. For the Chatanika hydrostatic Rayleigh lidar system the altitude range would increase by 2 scale heights or $\sim 16$ km. For the Andoya Doppler Rayleigh lidar system the altitude range would increase by 1 scale height or $\sim 8$ km. For the Heifei Doppler Rayleigh lidar system the altitude range would increase by 3 scale heights or $\sim 24$ km.
What is significant about these improvements in lidar performance is that they extend the scope of the Rayleigh lidar systems across one of the defining boundaries in the atmosphere; the turbopause. The turbopause is the transition where molecular diffusion dominates over eddy diffusion in the atmosphere. Rayleigh lidar systems employing a Large Aperture Telescope could directly observe the composition and dynamics of the turbopause region. One of the key signatures of this transition is the change in scale height of the atmosphere in the transition from a well-mixed gas with a single scale height to a gas where each species gravitationally settles independently and so each species has an independent scale height. With a 100 m$^2$ collecting area, Raman measurements that are currently limited to the stratosphere become viable in the lower thermosphere. Assuming a vibrational Raman scattering cross section 1000 times less than the Rayleigh cross section, Raman measurements of N$_2$ could be made with the hydrostatic Rayleigh lidar system with an accuracy of 2% at 80 km and a resolution of 2 km and 12 h. Measurements with an accuracy of 5% could be made at 100 km. This would provide a nightly averaged measurement of the N$_2$ profile in the upper mesosphere and lower thermosphere. Increase in laser power from the current 8 W of the Chatani-based Rayleigh lidar system would serve to extend the scope of these measurements even further.

![Figure 4](image_url)

**Figure 4.** Profiles of wind speed and direction measured by using a Rayleigh lidar system compared with data from a collocated pilot balloon on 2 August 2010 from 22:00 PM to 23:00 PM. From Tang et al., 2012.

**References**


The State-of-the-art Na lidars for MLT dynamics studies

Chiao-Yao She
Department of Physics
Colorado State University
Fort Collins, CO

Ralph Burnham
Fibertek, Inc.
Herndon, VA

Abstract
Na Doppler lidar as an essential instrument for atmospheric dynamics studies in the mesosphere and lower thermosphere (MLT) has been proven beyond doubt. Going forward, there are two challenges for Na lidar technology and science: first, to develop and implement a higher power transmitter that is robust enough to realize automation potential and to facilitate high quality and large quantity of data acquisition, and second, to develop and implement receivers with aperture at the level of 10 - 100 m² to permit observations with high temporal and vertical resolution, so that turbulence dynamics and eddy diffusion can be observed directly, along with temperatures and winds at various time scales. This paper is an attempt to lay out the strategies for achieving these objectives.

1. Introduction
Due to its large resonance scattering cross-section, Na resonance lidar has been recognized as a gold standard [Gardner, 2004] for investigating atmospheric dynamics in the mesopause region, often termed as the mesosphere and lower thermosphere (MLT). At this juncture, Na lidar systems on hand have already proven to be a reliable tool for observing temperature and winds as well as gravity wave (GW) dynamics as well as GW-tidal interactions. That the latter studies are still rare is due mainly to the fact that the current transmitter is not yet robust enough to allow full automation for multiple day (either nighttime only or 24-hour continuous) operations with minimum human intervention. To remedy this defect, it is necessary to employ all-solid-state technology that not only can realize the potential of automation but also enjoys broad based industrial support. We therefore discuss the state-of-the-art Q-switched Nd:YAG lasers and the method of using them to generate stable, high power, coherent radiation at 589 nm in Section 3. In addition, to understand gravity wave dynamics fully, lidar data should be capable of evaluating the associated turbulence and eddy diffusion coefficients. It is now well known that while 0.5 km and 2.5 min resolution is sufficient for investigating gravity wave dynamics (in addition to slower perturbations), a much higher resolution, about 20 m and 2 sec, is necessary if the full turbulence spectrum is to be probed [Gardner and Liu, 2013]. In other words, for comparable photon uncertainties in temperature, wind and Na density, it requires about 5x10^6 times more signal for turbulence measurements. We therefore discuss and propose deployment geometry with large aperture receivers, along with a conservative estimate of their measurement uncertainties in Section 4. A recent Na-density-only lidar using a 6-m telescope has demonstrated the science potential in high resolution with direct observations of detailed fine structures and turbulence billows in the mesospheric Na layer [Pfrommer et al., 2009], which awaits quantification of background atmospheric state variables in such short temporal and spatial scales.

2. A brief overview of Doppler Na lidar measurement techniques
Since the translational motion of naturally occurring atoms in the mesopause region is in thermal equilibrium with ambient atmosphere, atmospheric temperature and winds may be determined from a ground-based resonance scattering lidar by monitoring, respectively, the Doppler-broadening and Doppler-shift of the return signal. This principle was first demonstrated by Gibson et al. [1979] by measuring the mean nocturnal temperature of the Na layer. Fricke and von Zhan [1985], who used a similar but improved flash-lamp pumped pulsed dye laser to scan over the NaD$_2$ spectrum, were able to resolve temperatures at different altitudes within the Na layer. Measurements in North America, with the collaboration between the research groups of She at Colorado State University (CSU) and Gardner at University of Illinois (UIUC), used a continuous-wave (CW) single-frequency tunable dye laser seeded pulsed dye amplifier as lidar transmitter. With this system, mesopause temperatures were first measured [She et al., 1990] by locking the seed laser to two absolute frequencies (at the peak, $\omega_a$, and cross-over, $\omega_c$, resonances) of the NaD$_2$ Doppler-free spectrum [She and Yu, 1995]. The same laser system was expended to simultaneous temperature and line-of-sight (LOS) wind measurement, initially tuned to a third frequency [Bill et al., 1991] and later with a three-frequency technique [She and Yu, 1994] by locking the laser at $\omega_a$ and up- and down-shifting the frequency by a fixed amount (630 MHz for example) to $\omega_a$ and $\omega_c$ using a dual acousto-optic modulator. Though human intervention is required in a long campaign, this system has been in operation for more than two decades and continues to be the workhorse for the three Na lidar facilities in the NSF sponsored Consortium of Resonance and Rayleigh Lidars (CRRL).

3. A proposed state-of-the-art Na laser transmitter

To go forward we are looking for all-solid-state technology that not only delivers more power (~10 times) but also has the potential for full automation and enjoys a broad-based industrial support for basic components. Our proposed lidar source, based on Q-switched Nd:YAG is a solid-state laser with robust performance for diverse high-power applications. The same Nd:YAG crystal can also be made to lase at 1319 nm. Combining these two frequencies via sum-frequency generation (SFG) in a nonlinear crystal like LiB$_3$O$_5$ (LBO), one can produce light tunable through the 589 nm Na resonance line. The SFG process has been carried out with CW or pulsed YAG lasers. The narrowband CW systems [Vance et al., 1998] can be used to seed the pulsed dye amplifier for Na lidar application, though the dye based system requires high-maintenance and can be troublesome at a remote observatory. The pulsed system, traditionally flash-lamp pumped though all-solid-state, typically does not have long maintenance-free operating life or the precision locking mechanism required for demanding LOS wind measurements [Kawahara et al., 2002] unless Doppler-free techniques are introduced into the seeding mechanism [She et al., 2007]. Both systems have low average power of ~1 W.

Due to the interest in producing laser guide-stars from the atmospheric Na layer for astronomical telescope applications, high power CW lasers at 589 nm have undergone rapid development with varied methods investigated. One of the most successful has been frequency doubled fiber lasers [Taylor et al., 2009]. This approach has produced systems yielding as much as 50 W, with 20 W systems commercially available. Unfortunately, a pseudorandom modulation code is required to range resolve long-range lidar measurements with a CW system. The associated penalty [Norman and Gardner, 1988] that arises due to continuous reception of background light could be as high as 20 times in power requirement[She et al., 2011]. While flash-lamp pumped Nd:YAG lasers produce high pulsed power with lower efficiency and at low repetition rate, the more recent diode pumping technique can produce Q-switched Nd:YAG lasers with higher efficiency and higher repetition rate. Transverse diode pumping of Nd:YAG slabs can produce pulses at ~100 Hz rate with excellent beam quality, but more recently the end pumping method has produced pulsed Nd:YAG laser with higher efficiency at repetition rates up to 10s of kHz. During the past decade, Fibertek, Inc., has gained extensive experience in custom-building such lasers for advanced lidar applications, thanks to the vision of NASA research laboratories. This work has resulted in numerous fielded airborne and space-based lidar systems. The proposed state-of-the-art Na lidar transmitter will be a sum-frequency generated solid-state system that utilizes Fibertek’s Nd:YAG lasers. Using proven design codes, Fibertek has investigated the trade-off between repetition rate, efficiency, power, and beam
quality of these lasers. The favored choice is an end-pumped modular system with repetition rate of 750 Hz, which can cover a range of 200 km unambiguously. This system has a plug efficiency of 1.5% including the nonlinear conversion efficiency of 33%; two such laser systems will be considered for the proposed Na lidar applications; one has 5 W output, and the other 10 W at 589 nm. The performance parameters for both 100 Hz and 750 Hz Na system are compared in Table 1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Slab Osc</th>
<th>Osc/Amp</th>
<th>End Pumped Osc/Amp</th>
<th>Osc/Amp</th>
</tr>
</thead>
<tbody>
<tr>
<td>PRF</td>
<td>100 Hz</td>
<td></td>
<td>750 Hz</td>
<td></td>
</tr>
<tr>
<td>Energy (1064 nm)</td>
<td>30 mJ</td>
<td>60 mJ</td>
<td>13.5 mJ</td>
<td>27 mJ</td>
</tr>
<tr>
<td>Energy (1319 nm)</td>
<td>15 mJ</td>
<td>30 mJ</td>
<td>6.5 mJ</td>
<td>13 mJ</td>
</tr>
<tr>
<td>Beam Quality (M^2)</td>
<td>1.5</td>
<td></td>
<td>1.2</td>
<td></td>
</tr>
<tr>
<td>Pulse Width</td>
<td>10-20 ns</td>
<td></td>
<td>10-20 ns</td>
<td></td>
</tr>
<tr>
<td>Volume</td>
<td>2 cubic feet</td>
<td></td>
<td>2 cubic feet</td>
<td></td>
</tr>
<tr>
<td>Mass</td>
<td>50 kg</td>
<td>75 kg</td>
<td>20 kg</td>
<td>40 kg</td>
</tr>
<tr>
<td>Pump Laser Efficiency</td>
<td>4% (1064), 1.5% (1319)</td>
<td></td>
<td>10% (1064), 4% (1319)</td>
<td></td>
</tr>
<tr>
<td>Power Input</td>
<td>250 W</td>
<td>500 W</td>
<td>350 W</td>
<td>700 W</td>
</tr>
<tr>
<td>Power Out (589 nm)</td>
<td>1.5 W</td>
<td>3.0 W</td>
<td>5 W</td>
<td>10 W</td>
</tr>
<tr>
<td>Efficiency (589 nm)</td>
<td>0.6%</td>
<td></td>
<td>1.5%</td>
<td></td>
</tr>
</tbody>
</table>

As the schematic in Figure 1 shows, the proposed Na transmitter system starts with two ring laser resonators at 1064 and 1319 nm. Each resonator YAG crystal is end-pumped by a pair of diodes at 808 or 880 nm. The master frequency source that seeds each resonator is a continuous-wave (CW) temperature-stabilized DFB (distributed feedback) laser. Including jitter, the linewidth of the seed is <10 kHz, and its frequency is set by stabilized temperature and a current tuning. In view of the stability of these seeds, the 1319 nm seed will be set at a fixed wavelength. The 1064 nm seed will be tuned by PDH or FM lock to keep the sum frequency at the peak, _a_ of the NaD2 hyperfine transition, using the proven Doppler-free spectroscopy technique developed at CSU [She and Yu, 1995]. To provide the CW light source for Doppler-free spectroscopy, a small portion of the seeds will be used to sum-generate a light beam at 589 nm (~1 µW). Most of the seeds (1319 nm at fixed frequency, and 1064 nm modulated by AOM to provide _a_ + and _a_ - cyclically or on command) will be used to seed the Q-switched pulsed resonators, respectively. The pulsed oscillator output will be amplified to provide ~30 W total from the
YAG systems that will then be mixed in a nonlinear crystal to generate 10 W output at 589 nm. The output characteristics of the pulsed YAG are ~15 ns in pulse width with Fourier transform limited frequency width of 33 MHz. Though there exists pulse height jitter of a few percent, the jitter in the center frequency of the pulse spectrum is estimated to be within ~1 MHz. The expected pulse width, frequency width, center frequency variation, and pulse height jitter of the 589 nm output will then be, respectively, 10 ns, 50 MHz, less than 1 MHz, and 2%. Since the shortest time duration for lidar measurements is 2 s, these variations and jitters will be much reduced when they are averaged over more than 1500 pulses. Conceptually, the proposed Fibertek system is similar to a recent diode pumped Japanese system, both based on the proposal for an all-solid-state flash-lamp pumped system [She et al., 2007]. While all the components of the Japanese system [Tsuda et al., 2011], which produced 2 W of power, are laid out on an optical table, the proposed Fibertek system is modular and compact with most optical elements hermetically sealed. The excellent Fibertek engineering and experience with airborne laser instrumentation and reliable thermal and electrical control will result in modular laser systems with enhanced frequency and pulse stability along with high-power capability of 10 W.

![Figure 1: The optical schematic of proposed Fibertek 10 W Na system. Approximately 2% electrical efficiency, 200 W optical pump, 500 W electrical input.](image)

4. Measurement accuracy estimation and proposed deployment strategies

We begin by presenting the received signal and background of the CSU Na lidar system in 2002, when it became possible for 24-hour continuous operation [She et al., 2003]. This system deployed two beams pointing to the north and east, both at 30° from zenith. The laser power is 0.5 W per beam at 50 Hz repetition rate, and the lidar signal and sky background is received by telescopes of 35 cm diameter within 0.8 mrad field of view, and fiber coupled to photomultipliers (PMTs) with 20% efficiency. The power-aperture product (PA) of the system is 0.05 Wm² per beam. Table 2 shows the parameters of the north beam, which will be used as reference data, from which the measurement uncertainty of the proposed lidar systems will be estimated. Four observational scenarios are listed: 2310_night and 2205_night, representing winter and summer nights, respectively, when sky background is negligibly small, and 2310_noon and 2205_noon, representing winter and summer noon, respectively, when the sky
background is maximal, with a Faraday filter [Chen et al., 1996] incorporated in the receiver to reduce sky background under sunlit conditions. The table shows, in Columns 3 and 4, the signal $S$ and background $B$ in a vertical range of 130 m (or slant range of 150 m) at the edge of the Na layer (98 or 99 km) integrated over a time of 40 s. Since the signal count is maximum at the centroid of the Na layer, the photon signal-to-noise decreases from the center to the edge of the layer, so we consider only an altitude range (as shown for each scenario) of ~15 km, and present the largest measurement uncertainty (at the edges) for temperature $T$, LOS wind $V$, and Na density $D$, respectively, in columns 5, 6, and 7, at a vertical resolution of 0.5 km and 1 hr. We note that the measurement uncertainties of the last two rows, even with the use of Faraday filter, are still considerable, as noon is the worst case scenario. In this situation, a UV lidar would work better. For a conservative estimate, we chose the data of the lower signal north beam as the reference. For 40 s integration, the photon count of the Na profile from the north beam on 2310_night is 98261 (from the east beam it is 142240), corresponding to 49 cts per pulse. When this is converted to a Na lidar system at University of Colorado using 30 Hz laser and 81 cm dia. telescope, the Na layer cts/pulse is 877 cts. This number can be 50% to 100% lower than that recently achieved [Private communication, X. Z. Chu], indicating again that our estimate based on the reference system (Table 2) is a conservative one.

<table>
<thead>
<tr>
<th>Observation DOY_time</th>
<th>Altitude range</th>
<th>Edge S (cts)</th>
<th>Edge B (cts)</th>
<th>$\Delta T$ (K)</th>
<th>$\Delta V$ (m/s)</th>
<th>$\Delta D$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2310_night</td>
<td>85 - 99 km</td>
<td>268</td>
<td>1.3</td>
<td>1.36</td>
<td>1.29</td>
<td>0.33</td>
</tr>
<tr>
<td>2205_night</td>
<td>86 - 98 km</td>
<td>107</td>
<td>0.51</td>
<td>2.3</td>
<td>1.99</td>
<td>0.52</td>
</tr>
<tr>
<td>2310_noon</td>
<td>85- 99 km</td>
<td>14.4</td>
<td>40</td>
<td>26.6</td>
<td>21.88</td>
<td>2.75</td>
</tr>
<tr>
<td>2205_noon</td>
<td>86-98 km</td>
<td>6</td>
<td>63</td>
<td>64.3</td>
<td>68.9</td>
<td>7.44</td>
</tr>
</tbody>
</table>

Resolution: 130 m-40 s (0.5 km and 1 hr) for S and B ($\Delta T$, $\Delta V$, $\Delta D$).

**Photon noise uncertainty of a dual-beam and a vertical-beam configuration:** In order to measure zonal momentum flux of gravity waves, the CSU lidar was upgraded in 2006 by installing 40% efficient PMTs, larger telescopes (PA = 0.2 Wm$^{-2}$) and a dual-beam geometry in the east-west plane, as shown in Figure 2. It is a straightforward calculation to deduce the photon noise uncertainties for temperature, zonal and vertical wind, $\Delta T$, $\Delta u$, $\Delta w$, of this dual-beam configuration from the uncertainties, $\Delta T$ and $\Delta V$, of one beam pointing at 30° off zenith. To relate these quantities, we note that an altitude range $\Delta z$ is related to a slant range, $\Delta r$, at an angle $\theta$ from zenith, by $\Delta z = \Delta r / \cos \theta$. For the same altitude range $\Delta z$, the relationships between the received signal and background for a beam pointing at an angle $\theta$ and those received at 30° are $S_\theta = S_{30^\circ} / \cos \theta$; $B_\theta = B_{30^\circ} / \cos \theta$.
The relationship of photon noise measurement uncertainties from a beam pointing at 30° (as a reference), to those from one beam pointing at $\delta T_{30}$, $\delta V_{30}$, and $\delta D_{30}$ are then given in (1), to those from one vertical beam system, $\delta T_{v}$, $\delta V_{v}$, and $\delta D_{v}$ given in (2); and to those from a dual-beam configuration, $\Delta T$, $\Delta u$, $\Delta w$, and $\Delta D$ given in (3).

$$\frac{\delta T_{30}}{\delta T_{v}} = \frac{\delta V_{30}}{\delta V_{v}} = \frac{\delta D_{30}}{\delta D_{v}} = \sqrt{\frac{\cos \theta}{\cos 30^\circ}} \quad (1)$$

For one vertical beam,

$$\frac{\delta T_{v}}{\delta T_{30}} = \frac{\delta V_{v}}{\delta V_{30}} = \frac{\delta D_{v}}{\delta D_{30}} = \sqrt{\frac{\cos 0^\circ}{\cos 30^\circ}} = 1.075 \quad (2)$$

For the dual-beam configuration,

$$\frac{\Delta T}{\delta T_{30}} = \frac{\Delta D}{\delta D_{30}} = \frac{1}{\sqrt{2}} ; \quad \frac{\Delta u}{\delta V_{30}} = \frac{1}{\sqrt{2} \sin \theta} ; \quad \frac{\Delta w}{\delta V_{30}} = \frac{1}{\sqrt{2} \cos \theta} \quad (3)$$

In figure 3, the relative to the uncertainty of one beam pointing at 30° off zenith, the uncertainties of temperature, vertical wind and Na density from a vertical beam, Eq.(2), is indicated by a big cross (Left Scale), signifying $\Delta T_{30} / \Delta T_{v} = \delta T_{30} / \delta T_{v} = \delta D_{30} / \delta D_{v} = 1.075$. The relative uncertainties for $\Delta T$ and $\Delta D$ (Right Scale) in blue, for $\Delta u$ (Left Scale) in black, and for $\Delta w$ (Right Scale) in red, are also plotted in Figure 3, respectively. Note that the uncertainty in zonal wind $\Delta u$ increases as angle $\theta$ decreases, while it is only increased (decreased) by a small amount for temperature, $\Delta T$ (vertical wind, $\Delta w$). Though it will do little to improve measurement accuracy of temperature and vertical wind, $\Delta T$, and $\Delta w$, an additional vertical beam is often added as will be explained.

Because of the interest in the dual-beam geometry for momentum flux measurements, whose uncertainty is dominated by statistical fluctuations, the choice of angle $\theta$ requires further consideration with a compromise between zonal wind and momentum flux uncertainties in mind. Since the momentum flux is the statistical average of the product of vertical wind and horizontal wind perturbations, $MF = \overline{w' \cdot u'}$, the variance of MF depends on the interplay between the variances of $w'$ and $u'$, var($w'$) and var($u'$), as shown [Gardner and Liu, 2007] below:
\[
\text{var}(w'u') = C \left[ \text{var}(u') \frac{\tan^2 \theta}{4} + \text{var}(w') \frac{\cot^2 \theta}{4} + \frac{1}{2} \text{var}(w') \text{var}(u') \right]
\]  

where \( C \) is a constant depending only on resolution and measurement details; it is independent of \( \theta \). Therefore, the functional dependent of \( \text{var}(w'u') \) on \( \theta \) depends on the ratio \( \frac{\text{var}(u')}{\text{var}(w')} \). Since this ratio is \( \sim 100 \) [Gardner and Liu, 2007], the minimum of \( \text{var}(w'u') \) occurred at \( \theta = 5.7^\circ \); the dependence of \( \text{var}(w'u') \) on \( \theta \) for this case is also plotted in Figure 3 in dashed black (Left Scale). Here, as indicated in Figure 3, we see at \( \theta = 10^\circ, 15^\circ, \) and \( 20^\circ \), \( \Delta u = 4.34, 2.89, \) and \( 2.31 \) times \( \Delta V \) and \( \text{var}_{\text{opt}}(w'u') = 1.35, 2.31, 3.81 \) times optimum (minimum) value of \( \text{var}(w'u')_{\text{min}} \), respectively. We note that in the literature \( 10^\circ, 13.5^\circ \) and \( 20^\circ \) were used in SOR measurements [Gardner and Liu, 2007], in model calculation [Thorsen et al., 2000], and in measurements at CSU [Acott et al., 2010] and ALOMAR. Therefore, depending on the signal level, the angle of \( 6^\circ \) (or \( 5.7^\circ \)), \( 10^\circ \), \( 13.5^\circ \) or \( 20^\circ \) may be selected, respectively, for \( PA = 20, 10, 2 \) and \( 0.5 \text{ Wm}^2 \) per beam. Since at these angles the separations between the dual beams at 100 km are, 21, 35, 54, and 73 km, respectively, the determination of \( u', w' \) and \( T' \) (and associated fluxes) at high resolution from a dual-beam geometry may not include the contributions of short wavelength GWs. For this reason, a vertical beam should be added. Then, the difference in heat and constituent fluxes between those determined by the vertical beam and by the dual-beam geometry may be compared to assess the importance of short wave contributions.

**Measurement uncertainties estimated for two proposed systems:** As discussed, the state-of-the-art pulsed laser systems at 589 nm operate at 750 Hz with an average power of 5 W or 10 W. Here, we propose and estimate the performance of two systems for potential Facility deployment. First, we propose a moderate 5-beam system with single vertical and two sets of dual beams (pointing at \( 15^\circ \) off zenith), each with \( PA = 2.0 \text{ Wm}^2 \), which may be implemented with 1 W power and four 80 cm dia. telescopes. This moderate system (a 5 W laser and twenty 80 cm telescopes) may be constructed for mobile deployment. Second, we propose a large 5-beam system with a single vertical beam and two sets of dual beams (pointing at \( 10^\circ \) off zenith), each with \( PA = 10.0 \text{ Wm}^2 \). To beef up the turbulence measurements, the vertical beam may be replaced by one with \( PA = 500 \text{ Wm}^2 \). The per beam value (at \( 30^\circ \) off zenith) of \( S_{30^\circ}, B_{30^\circ}, \delta T_{30^\circ}, \delta V_{30^\circ}, \) and \( \delta D_{30^\circ} \) for these proposed systems may be scaled from the information in Table 2 by first considering a beam with the same PA of 0.05 \text{ Wm}^2, converting from 20% to 40% PMT and from 50 Hz to 750 Hz repetition rate; this will double the signal \( S \) and increase the background \( B \) by \( 2 \times 15 = 30 \) times. Then, for a system with higher PA, we further increase \( S \) in proportion to PA, and keep the background \( B \) the same (with the same receiver field-of-view).
With these new values of $S$ and $B$, the measurement uncertainties $\Delta T$, $\Delta u$, $\Delta w$, $\Delta D$ by a pair of dual beams and $\Delta T$, $\Delta w$, $\Delta D$ by the vertical beam may be calculated from Eqs. (2) and (3). These are given in Table 3 for the moderate system proposed and in Table 4 for the large system proposed, both with resolution of 0.5 km and 2.5 min for night (1 hr for day with Faraday Filter), and with units for the uncertainties same as in Table 2.

### Table 3. Performance of the moderate system

<table>
<thead>
<tr>
<th>Observation</th>
<th>Dual-beam at 15°, with 2.0 Wm² per beam</th>
<th>Vertical beam with 2.0 Wm²</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\Delta T$</td>
<td>$\Delta u$</td>
</tr>
<tr>
<td>2310_night</td>
<td>0.56</td>
<td>2.04</td>
</tr>
<tr>
<td>2205_night</td>
<td>0.94</td>
<td>3.15</td>
</tr>
<tr>
<td>2310_noon</td>
<td>1.63</td>
<td>5.20</td>
</tr>
<tr>
<td>2205_noon</td>
<td>3.52</td>
<td>14.6</td>
</tr>
</tbody>
</table>
Table 4. Performance of the large system

<table>
<thead>
<tr>
<th>Observation DOY_time</th>
<th>Dual beam at 10⁰, 10.0 Wm² per beam</th>
<th>Vertical beam at 10.0 Wm²</th>
<th>Vertical beam at 500 Wm²</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ΔT</td>
<td>Δu</td>
<td>Δw</td>
</tr>
<tr>
<td>2310_night</td>
<td>0.25</td>
<td>1.37</td>
<td>0.24</td>
</tr>
<tr>
<td>2205_night</td>
<td>0.42</td>
<td>2.11</td>
<td>0.37</td>
</tr>
<tr>
<td>2310_noon</td>
<td>0.57</td>
<td>2.69</td>
<td>0.47</td>
</tr>
<tr>
<td>2205_noon</td>
<td>0.96</td>
<td>5.89</td>
<td>1.04</td>
</tr>
</tbody>
</table>

Since under sunlit conditions, the sky background at 589 nm is too high for successful measurements to resolve GWs (at 0.5 km, 2.5min) or turbulence (at 20 m, 2 s), we should only estimate nighttime uncertainties for the resolution of 20 m and 2 s. The results are shown in Table 5. Though the uncertainties at 2 s integration could be substantial, a moderate integration can beat down the photon induced uncertainties [Gardner and Liu, 2013]. This is particularly so for the 500 Wm² beam when the uncertainty of a single Na density measurement with 2 s integration is below 1%. The performances of vertical beams with 2.0 Wm² and 10.0 Wm² are not shown in Table 5, but they can be calculated easily.
Table 5. Nighttime performance of proposed systems with 20 m and 2 s resolution.

<table>
<thead>
<tr>
<th>Observation DOY_time</th>
<th>Altitude range</th>
<th>Dual beam at 15°, with 2.0 Wm² per beam</th>
<th>Dual beam at 10°, with 10.0 Wm² per beam</th>
<th>Vertical beam with 500 Wm²</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>ΔT  Δu  Δw  ΔD</td>
<td>ΔT  Δu  Δw  ΔD</td>
<td>ΔT  Δu  Δw  ΔD</td>
</tr>
<tr>
<td>2310_night</td>
<td>85-99 km</td>
<td>26.4 96.7 25.7 5.8</td>
<td>11.9 64.9 11.5 2.7</td>
<td>2.40 2.27 0.53</td>
</tr>
<tr>
<td>2205_night</td>
<td>86-98 km</td>
<td>44.6 149 39.9 9.2</td>
<td>20.1 100 17.7 4.1</td>
<td>4.05 3.51 0.84</td>
</tr>
</tbody>
</table>

We point out that the above uncertainty estimates are based on photon noise errors. Though the photon noise dominates the measurement uncertainty, the frequency of the pulsed output is broadened with an asymmetric shape compared to the CW seed. This frequency chirp gives rise to a shift between the center-of-mass of the pulse spectrum and the frequency of the CW seed. Its effects on temperature and wind need to be determined and accounted for. In principle, this chirp-caused frequency shift may be measured by a heterodyne beating technique [White et al., 2004] on a pulse-to-pulse basis, though it is much more difficult to do for a short pulse of ~10 ns. Fortunately, since a lidar measurement is done in seconds if not in minutes, this frequency shift (or wind bias) can be monitored by averaging over many pulses with a spectral filter technique [Yuan et al., 2010]. Hopefully, the state-of-the-art Na laser system is stable and repeatable; the broadening and chirp-caused frequency shift can then be determined in the laboratory and their effects accounted for in the data analysis. Also, as pointed out, the use of a 750 Hz laser (compared to a 50 Hz laser) increases the sky background by 15 times, making daytime reception more difficult. At the same time, the higher rep rate will reduce the peak pulse power, enabling the use of a system with 10 times more average power, free from the Na layer saturation problem.

5. Conclusions

With the state-of-the-art diode-pumped, high-efficiency Nd:YAG lasers readily available, a single frequency tunable, Q-switched pulsed laser at 589 nm and 750 Hz may be custom-made at an average power of 5 W or 10 W. Based on these robust and potentially automatable laser transmitters, we have conceptually proposed two Na lidar systems for Facility deployment: a moderate 5-beam system employing a vertical beam plus two sets of dual beams pointing at 15° from zenith with PA = 2 Wm² per beam, and a large 5-beam system employing two sets of dual beams pointing at 10° from zenith with PA = 10 Wm² per beam, plus a vertical beam with either PA = 10 Wm² or PA = 500 Wm². With the signal and background from an existing reference system, we conservatively estimated the measurement uncertainties of temperature, wind and Na density for the two proposed systems and found them capable of measuring temperature, wind components and Na density with resolutions for probing GWs (0.5 km and 2.5 min) as well as for probing turbulence (20 m, 2s), though a longer time integration is needed with the moderate system.

Acknowledgments

The lead author thanks Chet Gardner for his continued leadership and interest in the research of metal resonance lidar technology and science, as well as the information he provided for this paper on the requirement for turbulence/diffusion measurements.
References


She, C.-Y. (2007), A proposed all-solid-state transportable narrowband sodium lidar for mesopause region temperature and horizontal wind measurements, Canadian Journal of Physics, 85, 111 – 118.

She, C.-Y. et al. (2011), Mesopause-region temperature and wind measurements with pseudorandom modulation continuous-wave (PMCW) lidar at 589 nm, Appl. Opt., 50, 2916-2926.


Fe Doppler and Boltzmann LIDAR for Exploring Atmosphere and Space

Xinzhao Chu and Wentao Huang
Cooperative Institute for Research in Environmental Sciences & Department of Aerospace Engineering Sciences
University of Colorado at Boulder
Boulder, Colorado

Jonathan S. Friedman
Arecibo Observatory, SRI International & Puerto Rico Photonics Institute, Universidad Metropolitana
Arecibo, Puerto Rico

1. Introduction

The scientific motivation to explore the neutral properties of the Earth’s atmosphere and space is compelling. LIDAR (Light Detection And Ranging) is certainly one of the most advanced observational technologies for such exploration, especially when pushing science to the next levels. A new discovery of neutral iron (Fe) layers reaching 170 km in the thermosphere is just one of the advances in atmosphere and space sciences brought by cutting-edge lidar technologies and observations. With this extraordinary observation made by an Fe Boltzmann lidar in Antarctica, we were able to measure the neutral atmosphere temperatures from 30 to 170 km for the first time in the world [Chu et al., CEDAR, 2013a]. It brings us one step closer to realizing a dream of making whole atmosphere (0–200 km) lidar observations. Certainly, such discovery challenges our understanding of electrodynamics, neutral dynamics, chemistry, composition and energetics in the Earth’s upper atmosphere and space environment, therefore prompting new research growth.

There is growing recognition that meteorological sources of wave energy from the lower atmosphere are responsible for producing significant variability in the upper atmosphere. Furthermore, energetic particles and fields originating from the magnetosphere regularly alter the state of the ionosphere. These influences converge through the tight coupling between the ionosphere plasma and neutral thermosphere gas to produce emergent behavior in the space-atmosphere interaction region (SAIR). Unfortunately measurements of the neutral thermosphere are woefully incomplete and in critical need to advance our understanding of and ability to predict the SAIR. To fully explore neutral-ion coupling in the critical region between 100 and 200 km requires measurements of the neutral atmosphere to complement radar observations of the plasma. Lidar measurements of neutral thermospheric winds, temperatures and species can enable these explorations, an objective of highest priority for the upper atmosphere science community [CEDAR: The New Dimension, 2011].

To push the atmosphere and space science to the next levels, there are several new demands on the lidar observational capabilities. The first is to push the measurement range well above 100 km, to 200 km for example, for advancing the SAIR science. This is not only for detection of species, but also for measurements of neutral temperature and wind. Traditionally lidar was labeled for temperature and wind measurements below 100 km but almost nothing above 100 km where the space-atmosphere interactions become more and more critical. The second demand is to significantly increase the photon bucket for achieving unprecedented high precision and resolution in temperature and wind measurements. This is essential for study of fast dynamical processes in the atmosphere, such as turbulence and eddy flux. The third demand is to accurately characterize and control the spectrum of lidar pulses to sub-MHz accuracy and precision, not only on its central frequency, but also on its linewidth and line shape. This is the key for achieving temperature and wind measurements with 0.1 K and 1 cm/s accuracy, which is crucial for investigations that demand accuracy of absolute values, like vertical winds.
associated with general circulation and long-term temperature monitoring. The last but not the least demand is to push for the whole atmosphere lidar profiling of temperature and wind from ground to 200 km. This is for tracing various atmospheric waves from the lower atmosphere to the upper atmosphere.

There could be several different ways to meet these observational demands as described in other sections. In this section, we outline a very promising approach – the pulsed-alexandrite-ring-laser-based Fe Doppler lidar that is very likely to meet all the demands listed above when deployed with the large-aperture lidar observatory (LALO) receivers. Not only has it strong technological heritage from the field-proven Fe Boltzmann temperature lidar, Na Doppler lidar, scanning Fe Doppler lidar, and even the coherent Doppler lidar, but also it integrates many new high-precision spectroscopy, high-power and solid-state laser, high-sensitive and high-speed photonics, computer and data acquisition technologies. Such an Fe Doppler lidar can serve as a pivot to stimulate new lidar technology development and bring lidars to the next generations.

2. Neutral and Ionic Iron (Fe) Atoms in the Earth’s Atmosphere

Neutral Fe layers in the thermosphere reaching over 155 km were first discovered with the Fe Boltzmann lidar observations in May 2011 at McMurdo (77.8°S, 166.7°E), Antarctica [Chu et al., 2011]. As illustrated in Figure 1, these layers appear as vertically converged layers starting in the upper E region and then descend in height with time following the downward phase progression of gravity waves. Taking these Fe atoms as excellent tracer, the lidar data can be used to derive the neutral atmosphere temperatures from 30 to 150 km as shown in Figure 1c. Such layers appear to be relatively common during the last three years of lidar observations at McMurdo. They have been observed in all the months of March through August when the lidar signal-to-noise ratio (SNR) under low solar elevation is much higher than that during polar summer months. The morphology of the thermospheric Fe layers is also diverse. Besides gravity wave driven cases like in Figure 1, some Fe layers appear to be tide driven and some layers appear to be converged by gravity-wave wind shear but further modulated by longer-period waves [Chu et al., AGU, 2012]. A case on 20 March 2012 is shown in Figure 2 that the thermospheric Fe layers stay continuously over 10 hours with much higher density than the case in Figure 1. The derived temperatures are much closer to the MSIS00 model than in Figure 1.

![Figure 1](image1.png)

*Figure 1. Fe lidar measurements on 28 May 2011 at McMurdo: (a) Thermospheric Fe density contour, (b) MLT temperature contour, and (d) temperature from 30 to 150 km [Chu et al., GRL, 2011].*
Figure 2. (a) Fe and Rayleigh temperatures above and below 70 km on 20 March 2012 at McMurdo. (b) Mean lidar temperature between 9 and 17 UT, compared to MSIS [Chu et al., AGU, 2012].

Recently, the altitude record of neutral Fe layers has been elevated to 170 km by the lidar observations at McMurdo on 1 June 2013 (Figure 3a and 3b). Such Fe atom tracer enables the derivation of neutral atmosphere temperature to nearly 170 km. Figure 3c illustrates the lidar-measured temperatures from 30 to nearly 170 km using the combination of Fe Boltzmann technique [Gelbwachs, 1994; Chu et al., 2002] and Rayleigh integration technique [Hauchecorne and Chanin, 1980]. It is worth to point out that for the 1-h integration data the temperature error bars above 145 km are quite large due to the limited capabilities of the two-decade-old Fe Boltzmann lidar deployed at McMurdo Station. This Fe lidar was equipped with only 40-cm in diameter telescopes, and its transmitter consists of broadband and relatively low power (~2 W) lasers, both factors resulting in low signal levels. The situation can be significantly improved by using a modern Fe Doppler lidar with the proposed LALO receivers to not only reduce the temperature errors by 50+ times, but also enable the measurements of neutral winds in the thermosphere. Details will be discussed later in the technical sections.
These thermospheric metal layers are not confined to McMurdo Station but most likely a global phenomenon. Strong supporting evidence comes from the converged Fe layers regularly reaching over 140 km in the thermosphere observed at Davis (69°S, 78°E), Antarctica [Lübken et al., 2011] by a scanning Fe Doppler lidar operating at 386 nm [Höffner and Lautenbach, 2009; Lautenbach and Höffner, 2004]. In addition, Fe atoms have been observed up to 130 km in the so-called “layer topside” at Külingsborn and Arecibo that were first seen as the exponential decay of the topside of the main metal layers [Höffner and Friedman, 2004, 2005]. Careful inspection shows that some of the cases exhibit converged layer signatures. Later lidar observations at Wuhan also exhibit the exponential decay of the main layers [Yi et al., ??] as well as the sporadic Fe and Na layers above 105 km and up to 120 km [Ma and Yi, 2010]. Following the publication of Chu et al. [2011], lidar colleagues worldwide have searched through their database and found many more thermospheric events in K layers up to 150 km at Arecibo Observatory () [Friedman et al., 2013] and in Na layers up to 130 km at Syowa () [Tsuda et al., 2012], Beijing [Wang et al., 2012], Lijiang () [Xue et al., 2013], and Wuhan () [Gong et al., 2003]. These observations cover a very large geographic latitudinal range, supporting the global phenomenon hypothesis. A comparison between Fe and Na layers simultaneously observed by two lidars at Table Mountain in Boulder, Colorado [Huang et al., 2013] is shown in Figure 4. This is a fairly typical situation, that is, the Fe layers in the lower thermosphere are much more dense and occur more often than the Na layers. While the column abundance ratio of Fe to Na is about 3 for the main layers from 75 to 102.5 km, the ratio above 102.5 km can easily reach 10 or even higher during 12 nights of simultaneous observations at Boulder.
The largest altitude extension so far and most frequent occurrence of Fe layers in the thermosphere, when compared to other metal layers such as Na and K, are most likely due to the facts that Fe atoms are the most abundant meteoric metal species in the Earth’s upper atmosphere, and Fe$^+$ ions are a major species in sporadic E layers. Numerous lidar observations have shown that the Fe column abundance from 75 to 105 km is about 3 times larger than Na and several hundreds of times larger than K, Ca and Ca$^+$. Rocket measurements have found that while Fe$^+$ and Mg$^+$ are the major species in converged ion layers, Na$^+$ and K$^+$ occupy very minor percentages. As hypothesized in Chu et al. (2011), the thermospheric Fe layers observed at McMurdo in Figure 1 are most likely formed through the neutralization of vertically converged Fe$^+$ ion layers that descend in height following the gravity wave downward phase progression. Friedman et al. (2013) put forward a similar hypothesis that the observed thermospheric K layers at Arecibo were formed through neutralization of tidal-converged K$^+$ ion layers. Recent numerical modeling of Fe and Fe$^+$ layers reproduces the McMurdo lidar observations in Figure 1 and thus supports the above theory. Because Fe$^+$ ions are much more abundant than Na$^+$ and K$^+$ ions in the E and F regions, it is reasonable to expect the neutral Fe layers in the thermosphere to be more dense and occur more frequently than the neutral Na and K layers, consistent with the findings by limited simultaneous and common volume lidar observations of multiple metal species. This fact makes the neutral Fe layers a favorable tracer for extending the lidar measurement range to the thermosphere (say at least 200 km in altitude) in order to advance the SAIR science. Again, it is worth to point out that the current lidar detection sensitivity largely limits the metal species occurring height, frequency and geographic coverage. When the lidar SNR is significantly improved using modern lidars with LALO receivers, it is reasonable to expect much higher detection altitudes and more frequent occurrence worldwide.

### 3. Principles of Fe Doppler Lidar

The Fe Doppler lidar in discussion utilizes the Doppler effects of the strong 372-nm Fe absorption line originating from the ground state $^5\text{D}_4$ to an excited state $^5\text{F}_5$. Temperature and wind in the Fe distribution altitudes are derived from the inferred Doppler line width broadening and line frequency shift, respectively. Because of its UV operating wavelength, the Fe Doppler lidar possesses strong Rayleigh, aerosol and Raman scatterings, which are utilized to measure temperature and wind in the atmosphere lower than the Fe layer heights, and to detect aerosols in all altitude ranges.
**Fe-Resonance Doppler Lidar:** Iron (Fe) is the most abundance metal species in the mesosphere and thermosphere, providing an excellent tracer for temperature and wind profiling in these regions. The Fe atom is a transition metal and has 26 electrons in orbitals. The electronic configuration of Fe atom in ground state is $1s^22s^22p^63s^23p^63d^64s^2$. Fe atoms have four naturally stable isotopes $^{56}$Fe, $^{54}$Fe, $^{57}$Fe, and $^{58}$Fe with natural abundance of 91.75%, 5.85%, 2.12%, and 0.28%, respectively. An energy level diagram of the main Fe species $^{56}$Fe is illustrated in Figure 5, where the lowest energy level $3d^64s^2\tilde{a}^5D_4$ is the ground state. Due to its zero nuclear spin, $^{56}$Fe spectrum has no hyperfine structure therefore exhibiting much simpler energy level structures than those of alkali metals (e.g., Na and K).

![Energy level diagram of atomic $^{56}$Fe isotope](image)

*Figure 5. Energy level diagram of atomic $^{56}$Fe isotope that is closely related to the Fe lidar discussion.*

Several absorption lines of Fe are of interest to lidar applications. The energy separation between $a^5D_3$ and the ground state $a^5D_4$ is 415.932 cm$^{-1}$. In thermodynamic equilibrium, the ratio of the populations in the states of $a^5D_3$ and $a^5D_4$ obeys the Boltzmann distribution law and is a sensitive function of temperature. By measuring this population ratio, the ambient temperature can be derived as first proposed by Gelbwachs [1994]. The first realization of such an Fe Boltzmann temperature lidar was achieved by Chu et al. [2002] using two injection-seeded and frequency-doubled pulsed alexandrite lasers to probe two absorption lines at 372 and 374 nm and then take the ratio of resonance fluorescence at corresponding wavelengths. This Fe Boltzmann lidar is now being deployed at McMurdo that has pushed the Fe layer detection range to 170 km. From Figure 5 it is obvious that other absorption line pairs can also be used to detect the Boltzmann distribution, e.g., a pair of 386 and 389 nm lines shown in Figure 5, but they are less favorable because of the much lower effective cross-sections when compared to the pair of 372 and 374 nm. Boltzmann technique is insensitive to wind velocities, therefore incapable of wind measurements.
Table 1. Atomic Parameters for Fe Absorption Lines

<table>
<thead>
<tr>
<th>Wavelength in vacuum (nm)</th>
<th>Energy levels</th>
<th>A_{ki} (10^8 s^{-1})</th>
<th>g_i</th>
<th>g_k</th>
<th>Oscillator Strength f_{ik}</th>
<th>Branching Ratio B_{Fki}</th>
<th>Effective Scattering Cross Section at 200 K (x10^{-12} cm^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>248.40209</td>
<td>a^2D_4 - x^2F_5^o</td>
<td>4.81</td>
<td>9</td>
<td>11</td>
<td>0.5438</td>
<td>8.289</td>
<td></td>
</tr>
<tr>
<td>372.09928</td>
<td>a^2D_4 - z^2F_5^o</td>
<td>0.162</td>
<td>9</td>
<td>11</td>
<td>0.0414</td>
<td>0.9959</td>
<td>0.940</td>
</tr>
<tr>
<td>373.81941</td>
<td>a^2D_3 - z^2D_4^o</td>
<td>0.141</td>
<td>7</td>
<td>9</td>
<td>0.0382</td>
<td>0.9079</td>
<td>0.797</td>
</tr>
<tr>
<td>386.10058</td>
<td>a^2D_4 - z^2D_4^o</td>
<td>0.0969</td>
<td>9</td>
<td>9</td>
<td>0.0217</td>
<td>0.7863</td>
<td>0.403</td>
</tr>
</tbody>
</table>

Although Fe Boltzmann lidar has accomplished astonishing science achievements in the last 15 years, the lidar for future science exploration must be capable of both wind and temperature measurements. Therefore, Fe Doppler lidar is a natural choice as one of the next-generation lidars. Three different absorption lines originating from the same ground state a^2D_4 have central wavelengths of 248, 372, and 386 nm, respectively. Their atomic parameters are compared in Table 1. Although the 248 nm line has ~9–20 times higher effective cross section than the other lines, it falls in the short UV wavelength so cannot be used from the ground, although it would be a good choice for spaceborne Fe Doppler lidar. Because of the different Einstein coefficient A_{ki}, degeneracy factors, and wavelengths, the oscillator strength of the 372-nm line is 1.9 times of the 386-nm line. If further considering the different branching ratios, the effective scattering cross section of the 372-nm line is 2.33 times higher than that of the 386-nm line. This much higher effective cross section leads to our choice of the 372-nm lidar as the operating wavelength of the next-generation Fe Doppler lidar.

Fe atoms have four naturally stable isotopes ^{56}Fe, ^{54}Fe, ^{57}Fe, and ^{58}Fe with natural abundance of 91.75%, 5.85%, 2.12%, and 0.28%, respectively. The isotope line shifts relative to the ^{56}Fe line and the relative strengths are illustrated in Figure 6 and documented in [2].
Figure 6. The absorption cross-section of the Fe 372-nm line with Doppler effects for three temperatures at $V_R = 0$ m/s (a) and for three radial velocities at $T = 200$ K (b).

The 372-nm line is Doppler broadened by the Maxwellian thermal velocity distribution (~1 GHz at 200 K) and Doppler shifted by the radial wind velocity $V_R$ (~2.7 MHz per m/s).

Rayleigh/Aerosol/Raman Lidar: In the region free of aerosol and fluorescence, pure Rayleigh scattering is used to derive the relative atmosphere density and then temperature. The lidar employs short UV wavelength, high pulse energy, and daytime capability, enabling very sensitive detection of aerosols and clouds like polar mesospheric and stratospheric clouds (PMC and PSC). In the presence of PMC, the 3-frequency technique is insufficient to derive temperature and wind, since the Mie scattering from the clouds contaminates the Fe resonance fluorescence. We will employ a frequency scanning technique as an alternative operation mode to address this issue.

In principle, the MRI lidar can be used as a Rayleigh Doppler lidar to measure neutral wind from 0-70 km if spectrum analysis filters such as Fabry-Perot etalons are added to the lidar receiver.
He Lidar for LALO

Tony Mangognia and Gary Swenson
Department of Electrical and Computer Engineering
University of Illinois at Urbana-Champaign
Urbana, Illinois

He lidar has been a method of investigating the Earth’s thermosphere ever since it was originally simulated by Gerrard et al., 1997. The method is to resonate with the He \((2^3S)\) metastable state which has a significant population between 250 and 750 km for resonant remote sensing. Ground based spectroscopists have investigated the resonant emission due to solar scattering in the upper atmosphere (e.g. Tinsley, 1968). The He \((2^3S)\) is energetically \(\sim 19\text{eV} \) above ground state, and is produced in the Earth’s upper atmosphere by solar EUV photons as well as photoelectrons, and has a lifetime of several 10’s of hours. Consequently, the He metastable has a persistence in the upper atmosphere at low and mid latitudes, and particularly in the winter when the winter He bulge persists in the upper atmosphere.

Waldrop et al. [2005] performed a chemical analysis study and deduced the state was quenched significantly below 300 km (Figure 1). Knowledge of the production/loss enables further calculations of the populations of the metastable for two sites at low latitude, both likely candidates for fielding such a lidar for thermospheric investigations. Figure 2 is a plot of metastable populations vs. altitude for both Arecibo, PR and Huancayo, Peru.

![Figure 1. Metastable He\((2^3S)\) loss rates as a function of altitude for the dominant loss processes. From [Waldrop et al., 2005].](image)
Figure 2. Altitude profiles of metastable He density during morning twilight (SZA ~ 100°) at Arecibo Observatory (left panel) and at Huancayo Geophysical Observatory (right panel) under various seasonal and solar activity conditions.

Note that the peak population is near 400 km and slowly drops in density with altitude above that altitude of peak population for the SZA~100°.

The considerations to implement lidar observations in this region are unique and unprecedented. Narrow-band Doppler lidar capabilities would be able to investigate the He (2^3S) line shape and shift for measurements of wind, temperature, and non-Boltzmann distributions of line shapes where source distributions of the observed volume may be different. Considerations are also to transport the system to the Auroral regions where secondary electrons are produced in auroral arcs producing large quantities of the metastable. Those distributions of the metastable which are time are immersed in ion flows due to electric fields along arcs would also be observed.

Laser technologies are evolving with solid state technologies to be able to tune to wavelengths not heretofore feasible for lidar remote sensing. The laser described here is a MOPA (Master Oscillator Power Amplifier). In this case, the seeder source for this laser is a DBR (Diode Bragg Diffraction) laser diode. The diode is fabricated so that the Bragg diffraction is tuned to transmit at roughly 1083 nm, with a few mW of power. Gain is provided by multi-stages of power diode pumped Yb fiber amplifiers. The seeder is 'locked' to the He (2^3S) peak resonance. This is accomplished by generating a gas cell of He, exciting a significant population to the He (2^3S) with an RF discharge, and then tuning a small beam of the source emission through the cell and locking on a diode sense of the resonance peak [Carlson et al. 2008]. The center wavelength of the DBR is then controlled by both the temperature and the current applied to the DBR. The laser lock controller developed by Carlson et al. works reliably. The first laser produced ~10 W cw with a bandwidth of ~150 MHz, and the latest configuration is producing ~40 W [Mangognia et al., 2011].
Figure 3. A master oscillator, power amplifier configuration with an overall gain of 36 db, 10 W single mode output, with 150 MHz bandwidth.

Figure 4. A scan and lock to the 1083.032 nm resonance.

The Doppler broadened line shape for the thermosphere He metastable is shown in Figure 5. The methods used in Doppler resonance measurements of Na has routinely performed 3-frequency, and that methodology would be similarly implemented to perform He Doppler measurements.

Figure 5. A simulation of the thermospheric He metastable line width for expected thermospheric temperatures.
Figure 6. The technology for 3-frequency shifting in order to deduce the population Doppler width and shift is accomplished by Acoustic Modulators, a proven technology.

Testing for He (2³S) currently uses bistatic methods, in which the laser is separated from the transmitter by some distance, \( d \), and the receiving telescope images the beam, which has range information by the position of the beam element in the image (see Figure 7). This method works reasonably well for zenith-oriented beams where vertical winds and temperatures might be sampled, but it is more challenging to construct the off-axis sampling.

Figure 7. A bistatic configuration used with a CW laser beam.

Pulsed fiber lasers have power limits due to the limitations of the fiber to be able to handle high energy densities associated with pulsing; however, those technologies are being worked on by a number of fiber groups including our own at the University of Illinois, and likely by the time we field a LALO facility, this type of laser will be available with a pulsed capability. The laser may also be chopped to produce a pulse. A 100 W beam can be chopped with a 20% duty cycle, or 20 W average power. A pulse of 100
km would be 0.66 ms with a period of 3.30 ms. There is little or no return between 30 and 250 km, with a major interest in sampling at 300-800. With this type of chopped beam, a traditional pulse-gated return lidar can be implemented for He.

A further discussion of bistatic lidar is described. The method is valuable for testing and large altitude integrals. A larger baseline (1.5 km) will be used in Chile with the SOAR telescope (see Figure 9).

Figure 8. A projection of pixels onto the U of I telescope for a 100 m baseline between the transmitter and receiver.

Figure 9. The He metastable configuration for SOAR in Chile.

He lidar adds unique capabilities to measure the vertical wind in the thermosphere. The capabilities are extended to a reference of a few m/s, or temperatures of a few K with statistical samples of ~5000 counts. For long exposures, there is an inherent advantage with pulse-gated return lidar. The comparison is made to the time spent on a range volume for a time-gated return, versus staring at that volume from the ground for the duration of that integrated volume. The time-gated return for a 50 Hz, 1 W with a 1 km range gate results in an integration time of signal, background and sensor dark count, of .33 ms for 1 sec of pulses or 1 W. The background and dc integration time for a bistatic configuration, for a CW laser that is 1 W average power for 1 s, is 1 s, or 3000 times longer than the traditional pulsed system.
The current technology for 1083 He cw Doppler laser transmitters is stable in locking, to a few m/s. The laser transmitter has been built and laboratory demonstrated.

As in all lidar, the Power*Aperture product defines the system, and the laser is available, as is the technology for the aperture. The LALO will truly enable quality He lidar measurements to explore, for the first time, the thermosphere and exosphere with lidar.

It should be asked how else these measurements can be made. Clearly, a space-based limb interferometer can provide some useful information, but likely not vertical velocity. Upper thermosphere and exosphere measurements have not been accomplished with the temporal, spectral, and spatial resolution necessary to address the aeronomic problems facing the community.

References
Mangognia, T., and G. Swenson (2011), Resonance Fluorescence He LIDAR, Poster session presented at: Joint CEDAR-GEM Workshop, The 26th Annual Summer CEDAR Workshop, 2011 Jun 26 – Jul 01; Santa Fe, NM.
Influx of cosmic dust and how it affects the structure and evolution of the Earth’s atmosphere

John Plane
School of Chemistry
University of Leeds
Leeds, United Kingdom

The main sources of dust in the solar system are the sublimation of comets as they approach the sun, and collisions between asteroids in a belt between the orbits of Mars and Jupiter (1). There are also minor inputs from the Kuiper belt and the Oort cloud. Dust particles from long-decayed cometary trails (in particular from comets Encke and 55P/Tempel-Tuttle) dominate the continuous sporadic input. Fresh dust trails produced by comets which crossed the earth’s orbit recently (within the last 100 years or so) are the origin of meteor showers (e.g. the Perseids). The sporadic background provides a much greater mass flux on average than meteor showers.

In this section we will address the apparently straightforward question: what is the magnitude of the cosmic dust input to the earth’s atmosphere? And then consider the corollary: what impacts does this have on the atmosphere? Table 1 shows that estimates of the Interplanetary Dust Particle (IDP) input into the atmosphere vary from ~2 to 270 t d$^{-1}$ (tonnes per day)! Zodiacal cloud observations and spaceborne dust detection (dark blue shading in Table 1) indicate a daily input of 100 – 300 t d$^{-1}$, which is mostly in agreement with the accumulation rates of cosmic elements in polar ice cores and deep-sea sediments (grey shading). In contrast, measurements in the middle atmosphere (light blue shading) – by radar, lidar, high-flying aircraft and satellite remote sensing – indicate that the daily input is only 2 - 50 tonnes (2). It is important to note that these middle atmosphere measurements probe the products of meteoric ablation, which is the process of rapid evaporation of the constituent elements in a meteoroid once it has melted (temperature > 1800 K).

There are two reasons why this enormous discrepancy matters. First, if the upper range of estimates is correct, then vertical transport in the middle atmosphere may be considerably faster than is generally thought to be the case, or the degree of ablation of the incoming material – which creates the atoms, ions and aerosols observed in the middle atmosphere – may be significantly overestimated. On the other hand, if the lower range is correct, then our understanding of dust evolution in the solar system, and transport mechanisms from the middle atmosphere to the earth’s surface, will need substantial revision. Second, cosmic dust particles enter the atmosphere at high speeds (11 - 72 km s$^{-1}$), and their ablation injects metals into the atmosphere which are involved in a diverse range of impacts, including the formation of layers of metal atoms and ions (3), nucleation of noctilucent clouds (4), effects on stratospheric aerosols and O$_3$ chemistry (5), and fertilization of the ocean with bio-available Fe (6). These impacts of meteoric ablation obviously depend on the magnitude of the IDP input.

There is also good evidence that the IDP input has varied substantially during the earth’s history, and there have been attempts to link these variations to fluctuations in climate. For example, using $^3$He accumulation in Cretaceous limestone, the IDP flux was shown to have varied episodically by a factor of 4 between 73 and 100 My BP (7). Ir measurements indicate that the rate of micrometeorite deposition in Antarctica has varied by a factor of 25 in the last 120 ky (8). In fact, there is empirical evidence of a mass imbalance in the zodiacal cloud, which appears to be over-massive in relation to current sources by 1 to 2 orders of magnitude (9, 10). This can be explained by the injection of dust from a massive comet which entered into a short-period near-Earth orbit, when the IDP flux might have been large enough to affect terrestrial climate by changing the optical depth of the atmosphere (11). A survey with the Spitzer Space Telescope has shown that these short-period Jupiter family comets are very likely to generate substantial debris trails (12). It has recently been postulated that ice age conditions lasting for a millennium around 12.9 ky BP were caused by debris from a large short-period comet, whose remnants now exist as the Taurid Complex (13). An increase in the IDP flux has also been suggested as a cause of the “snowball”
Earth glaciations during the Neo-proterozoic era (~1000-540 My BP) (14), probably not by direct negative radiative forcing but through indirect forcing following the nucleation of ice clouds (15).

Table 1. Estimates of the global IDP input rate to the Earth’s atmosphere (deep blue = extra-terrestrial estimate; light blue = middle atmosphere estimate; grey = ice core/deep-sea estimate). Taken from (2).

<table>
<thead>
<tr>
<th>Technique</th>
<th>IDP input t d⁻¹</th>
<th>Reference</th>
<th>Potential problem of technique</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zodiacal dust cloud observations and modelling</td>
<td>270</td>
<td>Nesvorný et al. (16)</td>
<td>Needs to be constrained by terrestrial meteor radars</td>
</tr>
<tr>
<td>Long Duration Exposure Facility</td>
<td>110 ± 55</td>
<td>Love &amp; Brownlee (21)</td>
<td>Sensitive to IDP velocity distribution</td>
</tr>
<tr>
<td>High performance radars</td>
<td>5 ± 2</td>
<td>Mathews et al. (18)</td>
<td>Possible velocity bias / selective mass range</td>
</tr>
<tr>
<td>Conventional meteor radars</td>
<td>44</td>
<td>Hughes (17)</td>
<td>Extrapolation, selective mass/velocity range</td>
</tr>
<tr>
<td>Na layer modelling</td>
<td>20±10</td>
<td>Plane (49)</td>
<td>Sensitive to vertical eddy diffusion transport</td>
</tr>
<tr>
<td>Fe layer modelling</td>
<td>2</td>
<td>WACCM (see text)</td>
<td>Depends on vertical transport</td>
</tr>
<tr>
<td>Fe/Mg in stratos sulphate layer</td>
<td>22 – 104</td>
<td>Cziczo et al. (5)</td>
<td>Data has limited geographic extent</td>
</tr>
<tr>
<td>Optical extinction measurements</td>
<td>10 – 40</td>
<td>Hervig et al. (39)</td>
<td>Particle refractive indices uncertain</td>
</tr>
<tr>
<td>Fe in Antarctic ice core</td>
<td>15 ± 5</td>
<td>Lanci et al. (50)</td>
<td>Very little wet deposition by snow</td>
</tr>
<tr>
<td>Fe in Greenland ice core</td>
<td>175 ± 68</td>
<td>Lanci &amp; Kent (51)</td>
<td>Uncertain atmospheric transport/deposition</td>
</tr>
<tr>
<td>Ir and Pt in Greenland ice core</td>
<td>214 ± 82</td>
<td>Gabrielli et al. (45)</td>
<td>Uncertain atmospheric transport/deposition</td>
</tr>
<tr>
<td>Cosmic spherules in Antarctic ice</td>
<td>7 2</td>
<td>Taylor et al. (34)</td>
<td>Flux of unablated material only</td>
</tr>
<tr>
<td>Os in deep-sea sediments</td>
<td>101 ± 36</td>
<td>Peuker-Ehrenbrink (48)</td>
<td>Focusing by ocean currents</td>
</tr>
<tr>
<td>Ir in deep-sea sediments</td>
<td>240</td>
<td>Wasson &amp; Kyte (47)</td>
<td>Focusing by ocean currents</td>
</tr>
</tbody>
</table>
1. IDPs in the Solar System and Meteoric Ablation

Zodiacal light is the very faint diffuse glow caused by sunlight scattering off IDPs concentrated close to the ecliptic (i.e., the plane containing the sun and orbits of the planets). A recent dust cloud model (16), constrained by observations of the zodiacal cloud made by the Infrared Astronomical Satellite (IRAS), predicts that ~90% of the dust in the inner solar system comes from Jupiter family comets with a period of typically 20 years. These IDPs should enter the terrestrial atmosphere from a near-prograde orbit with a mean speed of ~ 14 km s\(^{-1}\), producing a global mass input around 270 t d\(^{-1}\), the highest estimate in Table 1.

The input flux of meteoroids into the atmosphere is so uncertain because no single technique can observe particles over the mass range from about 10\(^{-12}\) to 1 g which make up the bulk of the incoming material (1). Figure 1 shows that the particle mass can vary by 30 orders of magnitude, although the largest contribution of mass entering the atmosphere on a daily basis comes from particles around 10 \(\mu\)g. Assuming a meteoroid density of ~2.8 g cm\(^{-3}\), these particles will have a diameter of ~ 200 \(\mu\)m. There is a population of huge impactors with masses greater than 10\(^{10}\) g which make a significant contribution, but only on a geological timescale! Any single measurement technique will only sample a subset of this size distribution. For instance, optical camera networks which observe visible meteors detect particles larger than about 1 mg in mass, or 1 mm in radius.

![Figure 1. Mass influx (per decade of mass) plotted against particle mass (2).](image)

Meteor radars measure particles with masses between about 10\(^{-9}\) and 10\(^{-3}\) g, and therefore cover the most important mass range (Figure 1). During meteoric ablation the evaporating metal atoms ionize through hyperthermal collisions with air molecules, creating a trail of electrons behind the meteoroid which the radar detects. Meteor radar data was used to produce a much-quoted estimate of 44 t d\(^{-1}\) for the global input, although this involved artificially increasing the size distribution to match visual meteor observations (17). In the past two decades, high-powered large aperture (HPLA) radars, such as at Arecibo and the EISCAT radars in the Arctic, have been able to detect by incoherent scatter the meteor head echo (i.e. the ball of plasma around the ablating particle as it descends through the atmosphere). This enables measurements of the direction, velocity, deceleration and (indirectly) mass to be made (18). Both types of radar suffer from sampling biases. Conventional meteor radars do not efficiently detect meteors at higher altitudes (> 100 km), because of the rapid diffusion of the ionized trails. Since faster meteors generally occur at higher altitudes, distributions measured by meteor radars are biased towards lower speeds. In the case of HPLA radars, the magnitude of the head echo depends on the meteoroid mass and velocity, so each instrument is sensitive to a particular mass range, and the velocity distribution of the smallest particles measured by an HPLA radar is biased towards faster meteors.
Interpreting radar measurements requires a model to predict the evaporation rate and subsequent ionization rate of the elemental constituents during ablation. A recent example is the Chemical Ablation Model (CABMOD), which includes sputtering by inelastic collisions with air molecules before the meteoroid melts, evaporation of atoms and oxides from the molten particle, and impact ionization of the ablated fragments (19). CABMOD predicts differential ablation, i.e. the most volatile elements – Na and K – ablate first, followed by the main constituents Fe, Mg and Si, and finally the most refractory elements such as Ca. The model has been used successfully to model meteor head echo height profiles, using the Arecibo HPLA radar (20). Chemical ablation models should in future be used to correct for biases in the meteor mass/velocity distributions measured by radars. An important reason for doing this is to refine zodiacal cloud models by using radar observations to constrain the predicted IDP orbits and velocities. An initial attempt in this direction was published recently, showing that the IDP input could be reduced to only 41 t d⁻¹ by changing the initial orbital characteristics of IDPs ejected by Jupiter family comets to better match terrestrial radar measurements (10).

The population of IDPs smaller than 10⁻⁹ g can only be measured by impact detectors on satellites. An important estimate of the IDP input was provided by the Long Duration Exposure Facility (LDEF), an orbital impact detector placed on a spacecraft for several years, which yielded an estimate of 110 t d⁻¹ (21). However, the LDEF experiment measured crater size, which was treated as a proxy for particle kinetic energy. Hence, the particle velocity distribution had to be assumed in order to determine the mass distribution. If the average velocity is 30 km s⁻¹, compared with the value of only 18 km s⁻¹ that was employed in the LDEF analysis, then the corresponding mass distribution would be shifted down by more than an order of magnitude (18).

2. Metallic Neutral and Ion Layers

Ablation produces layers of neutral metal atoms, such as Fe, Mg and Na, which peak between 85 and 95 km in the terrestrial atmosphere (3). Several of these layers - Na, K, Li, Ca, Ca⁺ and Fe – can be observed using ground-based resonance lidars. One constraint is that the optical transition must be at wavelengths greater than about 300 nm; otherwise, strong absorption by the Hartley band of O₃ in the stratospheric ozone layer prevents optical transmission between the ground and the mesosphere, which rules out observations of important metallic species such as Mg, Mg⁺ and Fe⁺. There have been a number of measurements of the concentrations of positive metallic ions made by rocket-borne mass spectrometry (22). Metallic ions such as Mg⁺ and Fe⁺ have also been observed by resonant scattering of sunlight, using spectrometers on space vehicles. More recently, observations of the neutral metal atoms such as Na and Mg have been made by satellites in polar sun-synchronous orbits, providing near-global coverage (23). Figure 2 illustrates the Na column abundance (i.e. the concentration of the Na layer integrated over height), as a function of latitude and season. The satellite data-set (23) has been supplemented with ground-based lidar data during the polar winter when the Na layer is not solar-illuminated. Figure 2 illustrates that there is very little seasonal variation at low latitudes (less than a factor of 2), but the winter/summer ratio increases to nearly an order of magnitude at high latitudes.
Understanding the characteristic features of the metallic layers in the upper atmosphere has required studying the reaction kinetics of neutral and ionized metallic species with atmospheric constituents such as O$_3$, O$_2$, O and H. Over the past 30 years, the two classical techniques of flash photolysis and the fast flow tube have provided a great deal of kinetic data on the pertinent reactions of the meteoric metals Fe, Mg, Na, Ca and K. The results have been used to construct atmospheric models which successfully explain these metal atomic neutral and ion layers above 80 km (3). The metals exist mostly as ions above 100 km. Fe$^+$ and Mg$^+$ are the major constituents of sporadic E layers, thin layers of concentrated plasma which occur in the lower thermosphere (95–130 km) (22). Sporadic E layers are important for radio communications, both enabling over-the-horizon radio propagation and attenuating ground-to-space communications. The chemical lifetimes (against neutralisation) of these metallic ions are controlled by their ion-molecule chemistry. For instance, the lifetime of Fe$^+$ changes from more than a day above 100 km, to only minutes at 90 km (24).

Recently, sporadic ion layers have been observed, using radio occultation from spacecraft, to occur around 90 km on Mars (25, 26), 120 km on Venus (27) and 550 km on Titan (28). The CO$_2$ atmospheres of Mars and Venus pose a particular challenge to the existence of metallic ions, because they should form CO$_2$-clusters very rapidly and undergo dissociative recombination. In fact, it turns out that sporadic layers in the Martian atmosphere are most likely Mg$^+$ rather than Fe$^+$, because of differences in the ion-molecule chemistry (29). Titan is interesting because ablation occurs over a much greater altitude range, as a result of its small atmospheric scale height (30). An interesting next step would be to observe some of the neutral metal layers which should be present in the atmospheres of these solar system bodies. An important reason for carrying out comparative studies with these other atmospheres is that this provides a self-consistency check of the meteor input functions produced from the models of cosmic dust in the solar system.

Below 85 km in the terrestrial atmosphere, the metal atoms are oxidised in a series of reactions involving O$_3$, O$_2$, CO$_2$ and H$_2$O to form hydroxides and bicarbonates (3). Although not yet well understood, these metal reservoir species appear to be permanently removed by polymerising with each other and SiO$_2$ vapour over several days to form nanometre-sized meteoric smoke particles (MSPs), which provide a permanent sink for gas-phase metallic compounds (see below) (31).
In a recent collaboration between the National Center for Atmospheric Research and the University of Leeds, the chemistry of Na and Fe has been incorporated into the Whole Atmosphere Community Climate Model (WACCM), which is a 3D chemistry-climate model extending from the surface to ~140 km. An important outcome of this project is that an IDP input of only 6 t d\(^{-1}\) is required to model the observed Na abundance, and only 2 t d\(^{-1}\) for the Fe layer. However, this depends crucially on the modeled rate of vertical transport of the ablated metal atoms to below 80 km, where they are presumed to be permanently removed as MSPs. Four components of vertical transport have been identified: the residual mean circulation (downwards in winter, reverse in summer); turbulent (eddy) diffusion, produced by breaking gravity waves; downwards dynamical transport caused by dissipating gravity waves; and chemical transport, where wave action and irreversible chemical loss at a lower altitude (e.g. to form MSPs) produces a net flux (32). A high performance metal resonance lidar has been used to measure the Na atom density and vertical wind profiles simultaneously, from which the vertical Na atom flux can be calculated as a function of height (32). The annual-average downward Na flux corresponds to an IDP input of about 24 t d\(^{-1}\), which is ~4 times higher than that needed for the WACCM simulation of Na. This discrepancy needs to be resolved, because it may also affect the vertical transport of heat and major chemical species such as atomic oxygen.

The relative concentrations of the different meteoric metal atoms also provide valuable information. Now that many of the relevant ion-molecule and neutral reactions of Na-, Fe-, Mg- and Ca-containing species have been studied in detail in the laboratory, the atmospheric chemistries of these metals can be reasonably well quantified in a model. In the case of Na and Fe, modelling a set of lidar observations of both metals at South Pole showed that the Fe ablation rate, relative to that of Na, needed to be reduced by a factor of ~4 (33). That is, the Fe:Na ablation rates need to be ~4:1, compared to their chondritic ratios of 15:1. The Mg ablation rate needs to be decreased relative to that of Na by a factor of ~5, while Ca ablation rate needs to be decreased by a factor of ~40. These relative ablation factors therefore increase as the element becomes more refractory. Significant differential ablation will occur if most of the incoming IDPs are small and/or slow, i.e. in close to prograde orbits. If there is a population of meteoroids that lose all their volatile elements (Na and K) but do not ablate completely, this implies that the particles must have melted to allow rapid diffusion of Na and K to the gas phase. If a particle of mass greater than 10\(^{-9}\) g survives atmospheric entry, then it should sediment rapidly (within a day) to the earth’s surface. Hence, there should be a fairly homogeneous scattering of once-molten IDPs at the surface.

These particles are termed cosmic spherules, and can be identified in polar ice because they are close to perfectly spherical and glassy, having melted during atmospheric entry. The accretion rate of spherules has been measured by retrieving them from the bottom of an ice chamber at the Scott-Amundsen base at South Pole (34). The flux and size distribution of 50-700 \(\mu\)m diameter particles, corresponding to particles in the mass range from 0.2 to 500 \(\mu\)g, were used in the study. This mass range covers the bulk of IDP particles (Figure 1), so these cosmic spherules should provide a useful measure of the bulk of IDPs which underwent partial ablation. This flux corresponds to a global input of about 10 t d\(^{-1}\) (2, 34).

3. Formation and impact of MSPs in the Upper Mesosphere

A measurement of the size distribution of MSPs in the middle mesosphere should provide another constraint on the IDP input. MSPs can be measured directly above 70 km by rocket-borne particle detectors (35). However, these detectors measure only those particles that are charged. Thus, the total MSP concentration is obtained by dividing the measured number by the estimated fraction of charged particles in the plasma. Because the plasma density in the \(D\) region is in the range 100 – 1000 cm\(^{-3}\), similar to the number density of MSPs, the modelled fraction of MSPs which are charged is sensitive to a number of poorly known parameters (e.g. electron-particle attachment rates, positive ion-charged particle recombination rates etc.). In order to improve this situation, a new particle detector has recently been flown which contains a pulsed VUV lamp to photo-detach electrons from negatively charged particles (35).
An important reason for studying MSPs in the mesosphere is their relation to noctilucent ice clouds (NLCs). These clouds were first reported in 1885, and have been growing brighter and spreading to lower latitudes through much of the last century, so that they appear to be a clear signal of climate change (36). NLCs occur between 80 and 86 km, at high latitudes in the summer (4). MSPs are widely considered to be the nuclei on which the ice particles grow. This is important because changes to the dominant meridional circulation in the mesosphere (which is driven by gravity waves from the lower atmosphere) may alter the supply of nuclei, which would then affect both the occurrence frequency and brightness of the clouds. Furthermore, the increasingly consistent estimates of ice cloud particle numbers obtained from lidar, radar and satellite observations (4) can be linked back to the MSP number density and hence to the IDP input.

4. MSPs in the Lower Mesosphere, Stratosphere and Upper Troposphere

Below 80 km, ultrafine particles (diameter < 10 nm) do not sediment rapidly. Instead, 3D general circulation models predict that MSPs should be swept to the winter pole by the mean meridional circulation in the mesosphere before downward transport within the polar vortex to the lower stratosphere (37). Recent airborne measurements have revealed a 3-fold increase of the meteoritic content of stratospheric sulphate aerosol inside the winter Arctic vortex (38). An important constraint on the IDP input is the optical extinction caused by MSPs between about 40 and 75 km. It has recently become possible to measure these small extinctions (as low as $10^{-8}$ km$^{-1}$), using a visible/near-IR spectrometer on the Aeronomy of Ice in the Mesosphere (AIM) satellite (39). Extinction measurements indicate that the MSP composition is probably olivine (MgFeSiO$_4$). However, the refractive indices used for this study were for bulk crystalline minerals, whereas MSPs are nm-size (and likely amorphous) particles, and the particles are likely to be chemically weathered by H$_2$O and H$_2$SO$_4$ during the months they spend descending into the stratosphere. Measured refractive index data on realistic particles are needed to relate the observed atmospheric extinctions to the MSP volume densities, and hence through a general circulation model to the IDP input.

Metal-rich MSPs should readily remove acidic species (e.g. H$_2$SO$_4$, HCl, HNO$_3$) from the gas phase. This may explain the unexpected decrease of H$_2$SO$_4$ measured by mass spectrometry on balloons in the upper stratosphere (40). Airborne flights of an aerosol mass spectrometer in the mid-latitude lower stratosphere have shown that sulphate particles contain ~1 wt % of meteoric Fe and Mg (5). These fractions are even higher inside the winter polar vortices (38). Solid particles can act as efficient heterogeneous nuclei. Explaining nitric acid trihydrate (NAT) formation from the tertiary HNO$_3$-H$_2$SO$_4$-H$_2$O system in the winter polar stratosphere has been a long-standing problem in understanding ozone depletion. Homogeneous nucleation of NAT is too slow to account for observed NAT particles in the polar vortex (41), and MSPs have therefore been proposed as likely heterogeneous nuclei (42). Although experiments have been performed with silica particles and ground-up meteorites, neither of these surrogates is representative of the expected amorphous, fractal-like nature of MSPs, so that the nucleating ability of realistic MSP analogues should be investigated.

MSPs could therefore modify O$_3$ depletion in the lower stratosphere resulting from chlorine activation on PSCs and de-nitrification (42). A detailed understanding of interactions between MSPs and stratospheric aerosols will be important for accurate predictions as the stratosphere cools over the next century (43). Also, in the context of proposed geo-engineering plans to increase the sulphate aerosol by pumping SO$_2$ into the stratosphere (in order to increase the amount of solar radiation scattered directly back to space, thus counteract greenhouse gas-driven warming in the troposphere) (44), a quantitative assessment should be made of the possible effects caused by meteoric debris.
5. Deposition to the Surface

The IDP input has been estimated from the accumulation of several different elements - Ir, Pt and super-paramagnetic Fe - in ice cores (45, 46). The deposition flux is determined by measuring the concentration of the element in the ice sample, and using the snow accumulation rate to obtain the flux. Super-paramagnetic Fe occurs in Fe-rich particles trapped in the ice. Measurements in ice cores in central Greenland, and the Eastern Antarctic highlands, reveal a consistent picture: the deposition rate in Greenland is ~10 times higher than at Vostok, and the Greenland estimate of the IDP input is at the high end of the range in Table 1 (175 - 224 t d⁻¹). Similarly, the measured accumulation of Ir (47) and Os (48) in ocean-floor sediments indicates that the meteoric influx is around 240 t d⁻¹.

How can these very high fluxes be reconciled with the estimates from within the atmosphere, which seem to be consistent with a flux of much less than 50 t d⁻¹ (Table 1)? Interpreting the ice core flux measurements requires understanding the transport of MSPs down through the stratosphere and into the troposphere, and their subsequent deposition mechanisms. A recent attempt to model both dry and wet deposition of MSPs using the UK Met Office Unified Model produced a deposition flux in central Greenland ~7 times larger than in eastern Antarctica, in agreement with measurements (46). One intriguing result that emerged was the prediction of a large MSP deposition flux into the Southern Ocean around Antarctica, where the supply of bio-available iron to phytoplankton is limited (6). The estimated input was ~10% compared with the Aeolian dust input. However, unlike continental mineral dust which has a low solubility (estimates vary from <1 to 10%), the MSP Fe should be in the form of highly soluble Fe₂SO₄ after processing in the stratospheric sulphate layer. Thus, the input of bio-available Fe from IDPs is likely to at least as large as (and perhaps an order of magnitude greater than) the Aeolian dust input. This could have significant climate implications because phytoplankton both draw down CO₂ and produce dimethyl sulphide, which evades into the atmosphere and contributes to the formation of ultra-fine aerosol which may grow large enough to act as cloud condensation nuclei.

6. Water and Life

Two potentially important consequences of the input of cosmic dust into planetary atmospheres have been hypothesised: a source of water, and a source of life through the injection of organic precursors and microbes. Although these theories are outside the scope of this report, they provide a further rationale for understanding the origins of cosmic dust in the solar system, and so are briefly discussed here.

There is widespread agreement that most of the water in the Earth's oceans has an extra-terrestrial origin, but it has not yet been established whether from comets or asteroids. The key test has been to find a source with the same deuterium/hydrogen ratio as that observed in the ocean (D/H = 1.6 × 10⁻⁴). Spectroscopic observations of six Oort Cloud comets – including Halley and Hale-Bopp – reveal a ratio that is approximately double that of the ocean (D/H = 3.0 × 10⁻⁴) [Hartogh et al., 2011]. In contrast, the ratio in carbonaceous chondrites (D/H = 1.4 × 10⁻⁴), was thought to be good evidence that at least 90% of the Earth's water came from the outer asteroid belt close to Jupiter [Morbidelli et al., 2000]. However, a recent measurement of the D/H ratio in a Jupiter-family comet (103P/Hartley 2), found a value identical to that of ocean water [Hartogh et al., 2011]. This finding opens the possibility that the accretion of comets originating in the Kuiper belt beyond Neptune, rather than the Oort Cloud, provided a significant fraction of ocean water. Clearly, more cometary D/H observations are required to determine the relative importance of comets versus asteroidal dust.

There is evidence that many organic compounds which are components of DNA-based life on Earth were present in the early Solar System, and could thus have played a key role in the development of life. The amorphous organic solids found in meteorites appear to be similar to the mixed aliphatic-aromatic molecules which are synthesised in stellar outflows and probably give rise to the diffuse interstellar bands [Kwok and Zhang, 2011]. The sugar glycolaldehyde, which is needed to form ribonucleic acid (RNA), has recently been discovered close to a protostar 400 light years from the Earth [Jorgensen et al., 2012]. The 12C/13C isotopic ratios in organic compounds found in the Murchison meteorite indicate a non-terrestrial
origin; biologically relevant molecules identified so far include uracil (an RNA nucleobase) and xanthine [Martins et al., 2009]. It has also been proposed that deep-frozen bacteria and viruses adhering to cosmic dust particles could spread between planetary solar systems, seeding life around the galaxy. This “panspermia” theory postulates that large bolide impacts on a planet such as Earth, where life has already developed, eject boulder debris carrying microbes into space. If these boulders are then eroded by collisions with interplanetary dust particles and ground down to micron-sized dust, such small particles will be ejected rapidly (in less than $10^4$ years) from the solar system under radiation pressure before the organisms are destroyed by the harsh radiation environment. The dust may then become incorporated into protoplanetary systems elsewhere in the galaxy [Napier, 2004].

References


Incoherent scatter radar is a powerful technique for remotely measuring the bulk parameters of ionospheric plasmas. The basic theory behind the technique has proven to be extremely robust for plasmas near, but not necessarily at, thermodynamic equilibrium. The basic plasma parameters that can be estimated from ISR spectra are shown in Table 1.

### Table 1. Basic plasma parameters measured from ISR spectra as a function of range

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$N_e$</td>
<td>Electron number density</td>
</tr>
<tr>
<td>$T_e$</td>
<td>Electron temperature</td>
</tr>
<tr>
<td>$T_i$</td>
<td>Ion temperature</td>
</tr>
<tr>
<td>$V_i$</td>
<td>Ion velocity (Doppler)</td>
</tr>
<tr>
<td>$v_{in}$</td>
<td>Ion-neutral collision frequency</td>
</tr>
<tr>
<td>composition</td>
<td>Ion composition</td>
</tr>
</tbody>
</table>

Depending on the particular system and the ionospheric conditions, these can be measured continuously as functions of space and time from approximately 60 km to nearly 1000 km altitude and sometimes higher. Also depending on the specifics of the waveforms used, the spatial resolution can range from ~150 meters (E-region) to ~70 kilometers (F-region) and the temporal resolution can be as short as a fraction of a second (e.g. for density estimates in auroral arcs) to a few minutes (e.g. for temperatures in the day-side E region).

In addition to these basic measurements, a number of parameters can also be derived with the help of antenna motion, external estimates of the neutral atmosphere, and assumptions about the steady-state plasma. A few of these derived parameters are listed in Table 2.

### Table 2. Derived plasma parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$E$</td>
<td>Vector electric field</td>
</tr>
<tr>
<td>$J$</td>
<td>Current density</td>
</tr>
<tr>
<td>$V_n$</td>
<td>Neutral winds (lower E region)</td>
</tr>
<tr>
<td>$\sigma_p$</td>
<td>Pedersen conductivity</td>
</tr>
<tr>
<td>$\sigma_h$</td>
<td>Hall conductivity</td>
</tr>
</tbody>
</table>

The scientific capabilities of an incoherent-scatter radar result from a combination of its sensitivity, flexibility, spatial coverage, and the geophysical environment in which it is deployed. The existing ISR facilities demonstrate this in that they are technically very diverse and each has evolved to exploit specific characteristics of its environment. For example, Jicamarca utilizes its unique magnetic field geometry to measure one component of the plasma drift with unparalleled accuracy. Arecibo takes advantage of its large antenna aperture to probe to very low electron densities in the mid-latitude ionosphere, and uses its large signal-to-noise ratio (SNR) to measure ion composition to high altitudes. Millstone Hill puts its full steerability to good use in making measurements with large latitudinal and longitudinal coverage, including both mid-latitude and, sometimes, high-latitude measurements. Sondrestrom uses its antenna pointing flexibility to probe the highly structured and rapidly varying high-latitude auroral and polar cap ionosphere. In each case, the system evolution was driven in large part by the unique plasma environment accessible to the instrument. In contrast, the AMISR, by virtue of its
relocatability, was designed to probe a variety of plasma environments. It can, by design, adapt to each new location to best fit the needs of the new geophysical conditions. This need is addressed, in part, by the highly modular approach used in the system design.

The use of phased array radar for incoherent scatter measurements has opened a wide variety of possibilities for scientific investigation. Large reflector-based systems have traditionally been fraught with the problem of distinguishing between temporal changes in the ionospheric plasma and simple advection of ionospheric structures. For example, sporadic-E layers have previously been shown to vary significantly on timescales of seconds in a given volume. There is evidence that much of this variability is due to advection of horizontally inhomogeneous structures, but it is also clear that the structures evolve with time. A phased array system can support the short timescale imaging of such structures and the tracking of their evolution. A similar argument applies to the probing of other atmospheric and ionospheric phenomena, including auroral ionization structures. An example of this capability is shown in Figure 1, where an auroral arc is imaged in 3 dimensions and compared to optical images of the aurora (Semeter et al., 2009). Time sequences of such images show the development and evolution of auroral features.

Figure 1. Imaging of auroral ionization observed by PFISR in 3 spatial dimensions [Semeter et al., 2009].

Another case where pulse-to-pulse steering is of enormous advantage is the estimation of the latitudinal distribution of electric field vectors from F-region measurements of ion drifts. Such estimation with dish-based systems requires the physical motion of tons of antenna hardware and results in combining line-of-sight Doppler velocity measurements separated by at least several minutes. This clearly is sub-optimal for high latitude plasmas where the electric field often changes on timescales of seconds and on spatial scales of tens of kilometers. The two existing AMISR radars utilize inertia-less beam steering for vector drift estimation in nearly all of their observations. An example of this capability is shown in Figure 2, where vector drifts from an active but typical Poker Flat event are shown (Heinselman and Nicolls, 2008). The ability to determine time resolution after the fact via post-integration allows one to
“zoom-in” on features varying rapidly in time. Spatial imaging of vector electric fields has recently been extended to two spatial dimensions by Butler et al. [2010] and Nicolls et al. AGU [2010].

**Figure 2.** Vector velocity estimates on the night of 13–14 February 2007. (top) The panels show the perpendicular east, perpendicular north, and antiparallel velocities along with error estimates as a function of time and magnetic latitude. The top quiver plot shows the vector velocity estimates with arrows (blue indicating eastward drifts and red indicating westward) at 10-minute resolution. The lower quiver plot shows the zoomed-in region indicated by the dashed vertical lines at 2-minute resolution. Adapted from Heinselman and Nicolls [2008].

The lowest altitude at which ISRs can make measurements of plasma parameters is highly dependent on the electron density profile. Figure 3 shows an example of measurements as low as 60 km during an auroral event. The Doppler shifts on the spectra show clear signs of gravity waves in the neutral atmosphere from 60 to 80+ km altitude.
AMISR, using a solid-state transmitter and no moving parts on the radar itself, is very well suited to 24/7 operations. The AMISR presently located at Poker Flat (PFISR) has been operating in such a synoptic mode between other operations since shortly after its first light as a full ISR. PFISR’s long-term database of ionospheric measurements is unparalleled in its completeness. Since March of 2007, PFISR has operated in a low duty cycle mode, initially as part of the International Polar Year, for all times when high duty cycle user operations are not scheduled. These operations, which provide ionospheric density, temperatures, electric fields, and derived parameters at 10-15 minute cadence, have opened doors to...
investigations typically not accessible by ISR data. The initial goals revolved around distinguishing
between ionospheric climate and weather (variability), investigating the role of geomagnetic weather
driven by magnetospheric interactions as compared to variability driven by lower atmospheric
interactions, and proving a fiducial dataset for ionospheric modeling efforts.
Introduction

Over a million times a second, radar-detectable meteoroids ablate in the lower thermosphere, leaving behind plasma trails that the neutral winds transport. These trails create intense radar echoes and so, for over 50 years, radars have tracked these trails as an easy way to monitor winds in the lower thermosphere. These measurements have become important for understanding and modeling winds, tides and gravity waves in this region.

A number of meteor wind tracking techniques exist. Specular meteor radar measures the Doppler shift of meteors that pass through the atmosphere precisely perpendicular to the pointing direction of the radar. This relatively inexpensive technique can measure winds between 85 and 100 km altitude depending on the broadcast frequency. However, these radars must make many measurements to build up a wind profile, typically over the course of an hour. As a result, such measurements are averages over both space and time.

Modern specular meteor radars typically include some degree of interferometry, allowing one to determine the meteor's location and trajectory. This allows for better resolution of wind measurements. Nevertheless, this technique produces wind profiles with a vertical resolution generally larger than 1 km, often far larger. Over the past six decades, over 400 rockets released chemical tracers to measure winds in the LT. This represents an enormous investment in obtaining these wind profiles. For a far smaller cost, one can obtain similar measurements using radar tracking of meteor trails.

The meteor height-ceiling limits the top altitude detectable by a specular radar. When the trail cross-sectional diameter exceeds the transmit wavelength it becomes effectively invisible to the radar. This occurs roughly when the neutral-plasma mean-free-path exceeds the same wavelength. Hence, lower frequency radars can detect higher altitude winds. However, ionospheric effects such as phase-delay and refraction more strongly modify these measurements. The lack of frequent, intense meteors sets the lower altitude.

Non-Specular Meteor Radar

A newer method of remotely measuring winds also takes advantage of the billions of meteors that enter the atmosphere every day, leaving behind plasma trails which the neutral wind carries along. However, instead of using a small radar to track the entire trail, this method uses large radars to follow reflections from plasma irregularities that develop in or near many trails. This requires that the radars have interferometric capabilities and can point roughly perpendicular to the geomagnetic field. Since it first presented in 2009, only two radars have applied this method, the Jicamarca Radio Observatory and a smaller radar in China. They can infer wind velocities as a function of altitude.

By measuring winds between 93 and 110 km, this approach explores a region that overlaps the highest altitudes monitored by specular meteor radars but extends considerably higher. This technique does not require the temporal and spatial averaging typically used when obtaining winds from specular meteor radars or LIDARs, but instead measures detailed wind profiles with an altitude resolution of less
than a few hundred meters. Further, between midnight and dawn, this method can be used to monitor winds on a nearly continuous basis. The phase difference between channels gives the meteor trail position and, as seen in Figure 1, shows a clear evolution with time. One can see that the horizontal wind has a strong shear in the zonal (E-W) direction, ranging from 7m/s at 98km altitude to 102m/s at 96km while the meridional (N-S) wind shows far less shear.

**Figure 1.** Left image shows the SNR in dB from a large non-specular meteor trail and the right image shows the phase difference between the B and C quarters of the antenna array. This phase difference reflects the wind speed as the wind transports the trail. The data gap at 98.5km altitude occurs because the meteor trail passes through a null in the antenna pattern.

In a 2 minute period, JRO will often detect more than a dozen meteors lasting 10s or longer in the hours before dawn. Figure 2 shows two wind profiles obtained by combining all the meteors in a two data set. These measurements extend from 93km to 109km and indicate that the east winds have less variability than the north winds.

Non-specular trails result from radar scattering from field aligned irregularities which, in turn, derive from instabilities evolving into turbulence. This means that radars measure wave-fronts within a turbulent plasma. These wavefronts may have non-zero average phase velocities which may affect the phase of the returning signal and potentially add error to our results. Nevertheless, the measured results appear highly accurate and consistent from meteor to meteor.

While figure XX shows a clear overall slope, enabling us to make an accurate determination of the horizontal wind speed, the signal includes substantial deviations from this trend (e.g., see the phases changes at 6s). These deviations make it challenging to obtain wind data from meteor trails lasting less than 3 seconds. Only radars that can point nearly perpendicular to B can measure winds from non-specular trails. Hence, the Arecibo, EISCAT, and Sondrestrom radars cannot make these measurements but JRO, MU, and EAR can.
Low latitude radars can use this method to obtain altitude resolutions below a few hundred meters. Higher latitude radars will need to point at lower elevation angles which will reduce their altitude resolution but will increase the observing volume and likelihood of detecting a strong meteor trail. Also, they will need to use the Doppler shift to calculate the component of the wind velocity away from the radar to obtain meridional winds.

This technique works best between midnight and dawn. After dawn, meteor trails become substantially less frequent and less enduring because of the effects of the background ionosphere on trail formation and duration. We estimate it will require about 6-8 times as long an observing period to produce similar wind data. Between noon and dusk, we have the added problem that the meteor rate drops. After sunset, trail durations should increase but the count rate remains quite low until we approach midnight.

While the JRO is a good tool for performing this type of monitoring, far smaller radar can do this quite well. In fact, a group in China recently demonstrated this technique using just six simple antennas. They demonstrated reasonable agreement between the non-specular and classical radar approaches.

The horizontal wind model (HWM93) predicts considerably lower wind velocities between 90 and 115 km altitude than we see in our current measurements. These winds generally reach less than 50m/s for the zonal winds and less for the meridional one. These measurements derive from MF radar wind and specular meteor radar data. Our observations and the in-situ rocket measurements show consistently higher winds, reaching twice or more the HWM93 values.
Coherent Scatter Radars

David Hysell
Department of Earth and Atmosphere Sciences
Cornell University
Ithaca, NY

1. Introduction

The ionosphere is a remarkable medium for electromagnetic wave propagation, exhibiting a birefringent (double-valued) index of refraction that can be less than unity. Since the beginning of the radio age, it has been known that radio signals reflect from the ionosphere when their frequency is less than the ionospheric critical frequency and refract back to Earth when it is less than the maximum usable frequency (MUF). This is the principle behind radio sounding and ionosondes, the primary means of probing the ionosphere prior to the space age. An important limitation of ionospheric sounding is that it cannot be used to probe altitudes above the F peak. Around the time of the start of the space age, considerable effort and resources were consequently devoted to investigating “incoherent scatter,” weak radio scattering from thermal fluctuations in the ionospheric plasma. Incoherent scatter provides relatively unambiguous information about the state parameters (number density, temperature, composition, collisionality, magnetic field, drifts) at all altitudes of the ionosphere. However, incoherent scatter radars (ISRs) are large, expensive facilities requiring considerable resources to field, operate, and maintain. The power-aperture product of incoherent scatter radars is conveniently measured in megawatt-acres. There are just a dozen ISRs operating around the world.

In addition to incoherent scatter, “coherent scatter” can also be detected by radars and used to probe the ionosphere at many altitudes. The meaning of this term is ambiguous but generally means scattering that is stronger than incoherent scatter but weaker than total reflection. The detection of coherent radar scatter generally implies the existence of nonthermal irregularities in the plasma, indicative of free energy, waves, and instabilities. The resulting echoes can be very strong, meaning that coherent scatter radars can function at modest power levels even with relatively compact antennas. Autonomous, long-term coherent scatter radar observations are practical, economical, and commonplace. Coherent scatter can be used to monitor and study a number of different phenomena, particularly but not only during disturbed conditions.

Coherent scatter can also occur in the mesosphere, where the scattering cross section of neutral density fluctuations driven by neutral turbulence is enhanced by ambient D-region electrons. In the polar summer mesosphere, the scattering cross section is further enhanced by reduced diffusion rates related to charged ice crystals, the root cause of polar mesospheric summer echoes (PMSE). In the ionospheric F region, Farley-Buneman and gradient drift plasma instabilities produce intense field-aligned plasma irregularities (FAIs) which are easily detected by coherent scatter radars situated and configured so as to meet the conditions for field-aligned backscatter. The FAIs are particularly strong in the equatorial and auroral electrojets but also occur at middle latitudes in conjunction with irregular sporadic E layers. FAIs also occur in meteor trails throughout the MLT region. Near the dip equator, coherent scatter arises from altitudes near 150 km during the day for reasons that are not understood.

In the F region, Rayleigh-Taylor and/or E×B instabilities also generate strong FAIs. At high latitudes, the resulting coherent scatter is utilized to monitor the convection pattern. At low latitudes, the FAIs are responsible for so-called “equatorial spread F.” This is a common phenomenon that is very disruptive to radio communication and navigation systems and is consequently a focus of space weather research. At middle latitudes, FAIs can accompany medium-scale traveling ionospheric disturbances (MSTIDs) and are associated with midlatitude spread F. MSTIDs and irregular E layers appear to be related, but the connection is not well understood. Finally, FAIs can be produced through ionospheric modification (heating) in the E and F regions and detected using coherent scatter.
2. Methods

The methodology for using coherent scatter data varies widely. Sometimes, the backscattered power intensity is informative in itself, particularly when it is sorted into range and time bins (range-time-intensity format). Data such as those in Figure 1 were sufficient to demonstrate that equatorial spread F was caused by a Rayleigh-Taylor plasma instability that drove regions of depleted plasma through the F peak into the topside in its nonlinear stage of evolution. Note that calibrating the backscattered power in an absolute sense is very difficult and that the signal-to-noise ratio (SNR) is the most common data product.

The range resolution of a pulsed radar is limited mainly by the system bandwidth but is as fine as 150 m for many systems in use today. The time resolution is determined by the intensity of the backscatter and the desired level of statistical confidence. In the high SNR limit under which many coherent scatter radars operate, depending on the radar pulse rate and the manner in which spectral processing is performed a high degree of statistical confidence can often be obtained with integration times on the order of a few seconds.

![Figure 1. Coherent scatter observations of equatorial spread F event over Jicamarca plotted in range-rime-intensity format.](image)

Sometimes even the absence of ionospheric backscatter can be informative. For example, SuperDARN HF radars often receive strong ground clutter — backscatter from the ground detected along refracted ray paths. Variations in the backscatter intensity versus group delay are indicative of changes in the ionosphere through which the signals propagate. Periodic variations indicative of gravity waves and traveling ionospheric disturbances are often evident in the data. The characteristics of the waves can be deduced by comparing signals received through different radar beams.

Additional information is contained in the Doppler spectrum of the echoes. The Doppler spectrum can be measured with high precision by CW radars from which long, unbroken time series data are available. For pulsed radars, we must distinguish between underspread and overspread targets. With the former, range and frequency aliasing can be avoided with the use of simple pulse-to-pulse sampling schemes. This is untrue for the latter, necessitating the use of more complicated pulsing and/or data analysis methodologies which often result in greater ambiguity. The precision of Doppler shift
measurements depends on the length of the time series analyzed as well as on the radar wavelength. Precision on the order of a few m/s is typical in many applications.

![Radar image of coherent backscatter from Farley Buneman wave turbulence in the auroral zone.](image)

*Figure 2.* Radar image of coherent backscatter from Farley Buneman wave turbulence in the auroral zone. Pixel brightness, hue, and saturation reflect the backscatter intensity, Doppler shift, and spectral width according to the legend shown. Optical (red line) imagery from Poker Flat is superimposed in white. The green lines show the flight corridors from the Poker Flight Rocket Range. Here, ‘P’ and ‘F’ denote the locations of Poker Flat and Fairbanks respectively.

Finally, spatial information about the scattering medium in the direction normal to the radar beam can be obtained through electronic beam steering or through the combination of spaced antenna methods and radar interferometry. The former method is used by SuperDARN-class radars, and the latter by partial-reflection drift (PRD) and meteor radars. Both are used by large facilities like Jicamarca and the MU radar in Japan. With radar interferometry, the bearing to a target can be determined from differences of the phases of signals received on antennas separated in space. The angular width of the targets can be determined from the coherence, the modulus of the cross-correlation of signals from spaced antennas. In the event that multiple antenna pairs are available for reception, the angular distribution of the target can be determined continuously. Combining this information with information from range gating, volumetric images of the radar target can be made. The spatial resolution of the images in the direction normal to the beam is not diffraction limited, and in the high SNR case the resolution of the images can be considerably finer. Fully exploiting the capabilities of radar imaging is a frontier research area in radio science.

Coherent scatter clearly provides contextual information for studying a number of important aeronomic phenomena. Its main shortcoming, however, is that it is not always directly or obviously related to ionospheric state parameters. Notable counterexamples exist: The Faraday rotation of radar signals returned to Earth by coherent scatter can be used to determine electron density profiles unambiguously. Meteor trails drift with the background neutral wind which can consequently be inferred from the Doppler shift of the coherent scatter echoes. The Doppler shifts of coherent echoes from F-region irregularities at high latitudes are useful proxies for the line-of-sight \(\mathbf{E} \times \mathbf{B}\) convection speed, although refraction and other issues make the correspondence imperfect. The Doppler shift of coherent scatter from Farley-Buneman turbulence is closely related to the ion acoustic speed which is itself related to electron and ion temperatures.
A few studies have been performed to associate fundamentally the characteristics of coherent scatter with atmospheric or ionospheric turbulence parameters. For observations of mesospheric turbulence, the Doppler spectrum of the coherent scatter can be rigorously related to the turbulent energy dissipation rate. For coherent scatter from the $F$ region, attempts have been made to associate the shape of the Doppler spectrum with the underlying turbulence wavenumber spectrum in the context of so-called “collective wave scattering theory,” although the results of these analyses remain to be validated. Ultimately, the methods for analyzing and interpreting coherent scatter data vary from one ionospheric region to another and rely on the experience of the analyst.

For example, as mentioned above, the Doppler shift of coherent echoes from $F$-region irregularities at high latitudes is known to be a good proxy for the line-of-sight plasma convection speed. Furthermore, it has been determined empirically that a broadening in the Doppler spectrum can be associated with the open/closed field line boundary. The Doppler shifts of coherent echoes from the ionosphere are often presumed to be mainly indicative of $E\times B$ drifts, although this association must be applied with care. Coherent echoes from ESF plumes are usually dominated by Doppler shifts indicative of ascent even though the $F$-region plasma is mainly descending during ESF. The discrepancy arises because the backscatter comes preferentially from ascending, depleted regions. In the equatorial electrojet, the Doppler shift is indicative not of the background electric field but of the polarization electric field, which is an order of magnitude larger. To the degree that neutral winds establish $E$- and $F$-region polarization electric fields, they also affect the Doppler shifts of coherent echoes, although extracting that information from the data generally requires a substantial modeling effort. The Doppler shift of coherent echoes from Farley-Buneman waves is closely related to the ion acoustic speed. In the auroral zone, this speed is elevated by heating related to the convection electric field, and so the electric field also influences the Doppler spectrum, albeit indirectly.

Coherent scatter is measured in all latitude regimes in the $D$, $E$, and $F$ regions. (At Jicamarca, coherent scatter has been detected at over 2000 km altitude!) Measurements are fast and finely resolved in space and time compared to many other classes of upper atmospheric and ionospheric observations. However, the interpretation of coherent scatter results is not rigorously defined. The real strength of the coherent scatter technique is realized when it is combined with observations from other instruments, particularly incoherent scatter radars and lidars. The coherent scatter can provide the context surrounding the regions where a lidar provides unambiguous information about atmospheric state variables. That information can feed back into the interpretation of the coherent scatter.

For example, among the more persistent sources of coherent scatter at all latitudes are irregular sporadic $E$ layers. These layers of metallic ions are thought to be formed through the action of neutral wind shear. A leading theory states that the layers become irregular as the wind shear drives convective instability that deforms the layers. Once deformed, electrodynamic processes take over, leading to the generation of field-aligned irregularities (FAIs) and coherent scatter. Or so the theory goes. The existing radar datasets supporting research into sporadic $E$ layers barely capture the essential neutral dynamics, and so the theory remains speculative.

A wind temperature lidar could measure both the winds and wind shears necessary to form and destabilize the layers and the temperature profiles needed to assess the criterion for convective instability. A colocated ionosonde could establish the formation of the layers in the presence of wind shear, and a coherent scatter radar could identify layer structuring and subsequent FAI generation in a common volume. Thus, one of the oldest problems in radio and space physics could be resolved.

Another more fundamental issue is the ionospheric dynamo and the fact that coherent scatter radars are sensitive to dynamo-induced electric fields throughout the geospace system. Lack of detailed knowledge about the dynamo driver has made it difficult to capitalize on this information. A high-altitude Doppler lidar colocated with one or more coherent scatter radars would facilitate a quantitative assessment of dynamo theory and of the influence the MLT winds have on ionospheric stability, irregularity generation, and related aspects of space weather.
In-Situ Observations

M. F. Larsen
Department of Atmospheric and Space Physics
Clemson University
Clemson, South Carolina

1. Introduction

Ground-based measurement techniques for observing the mesosphere-lower thermosphere (MLT) region have advanced significantly in the last few decades. Techniques in lidar, passive optical, and radar have all improved the temporal and spatial resolution as sensitivity has increased, as well as the number and type of parameters that can be measured directly or inferred from the direct measurements. Other sections of this report give detailed descriptions of the advances in the various techniques. These improvements notwithstanding, there are unique measurement capabilities provided by sounding rockets. The advantages of the rocket techniques are that they provide measurements of parameters that cannot be obtained directly with ground-based instrumentation and they provide spatial resolution that often makes it possible to look at the details of the structure within the sampling volume of the ground-based instruments. The disadvantages of the in-situ measurement techniques are the expense and complexity associated with each individual launch and the limited number of sites where such measurements can be carried out.

Efforts have been made to develop extensive ground-based instrumentation at two of the permanent rocket ranges, Poker Flat Research Range located near Fairbanks, Alaska, and the Andøya Rocket Range in Norway. Both ranges have lidar and passive optical instrumentation either at the launch site or nearby locations, as well as radar instrumentation. The instrumentation is a regular part of the support for nearly all science launches carried out at those sites.

An indication of the science enhancement that can be achieved by combining ground-based and in-situ measurements is the fact that temporary rocket launch facilities have been installed specifically to take advantage of existing ground-based instrumentation on several occasions. The first notable example is the Coqui-2 rocket campaign staged in Puerto Rico in February and March 1998 to take advantage of new sodium lidar instrumentation that was available there in addition to the extensive radar instrumentation (see, e.g., Larsen, 2000; Larsen et al., 2004). Another is the Turbulent Oxygen Mixing Experiment (TOMEX) carried out in New Mexico in October 2000 to take advantage of simultaneous, coincident measurements from the sodium lidar at Starfire Optical Range (SOR) (Hecht et al., 2004). The latter launches made use of the White Sands Missile Range but required the installation of launch support equipment closer to the north end of the range to get close to the lidar facility. Other notable campaigns, including the MacWAVE (Williams et al., 2006a) and DELTA (Williams et al., 2006c) experiments at Andøya and the Turbopause experiment (Lehmacher et al., 2011) at Poker Flat, have made extensive use of the lidar instrumentation at the existing launch ranges. The combination of ground-based optical and in-situ measurements has been used in one of three ways, namely the validation of the measurements by comparison of two sets of measurements of the same parameter, enhancement of the combined measurements by in-situ probing of the sub-resolution volume, and enhancement of the combined measurements by adding measurements of additional parameters. The proposed facility further adds the possibility of extending the measurements to higher altitude ranges where critical neutral atmosphere parameters have been difficult to measure directly. Each of these topics is addressed separately below.
2. Measurement validation

The techniques available for measurements in the mesosphere and thermosphere have increased and improved significantly during the last two decades, but the validation of both new and established techniques has been a difficult problem since there are generally no direct measurements suitable for comparison. In-situ observations can, in some cases, provide independent measurements suitable for such comparisons. This type of measurement validation has been carried out extensively in the context of ionospheric measurements with comparisons of radar and in-situ electron density profiles, electric fields, and Doppler spectra. Comparisons of lidar and in-situ measurements have been limited, but a notable example is the collocated lidar and rocket wind measurements carried out as part of the TOMEX mission. The steerability and slew rate of the SOR telescope made it possible to scan the region between upleg and downleg trimethyl aluminum trails released from rockets launched from the White Sands Missile Range located south of SOR. The comparison of the wind profiles obtained from triangulation of the trail movements and the line-of-sight and vector wind profiles obtained from the Doppler lidar measurements showed excellent agreement (Larsen et al., 2003). Both the vertical and horizontal flow gradients were taken into account in the comparison.

Both passive and active optical techniques are dependent on emissions from atmospheric components that are often minor constituents, such as atomic oxygen or metallic atoms. High-resolution measurements of the vertical emission profiles are useful both for validating the ground-based measurements and for the interpretation of those measurements. Both the TOMEX and the earlier Coqui-2 in-situ photometer measurements (Hecht et al., 2004) showed complex variations with altitude in the region sampled by ground-based imagers and resonance lidars, indicating that the emission profiles are strongly influenced by dynamics as well as chemistry.

3. Probing the sub-resolution volume

The in-situ measurement techniques typically offer considerably better measurement resolution than the remote ground-based techniques (see, e.g., Hecht et al., 2004, and Lehmacher et al., 2011). The majority of these measurements are associated with electronic instrumentation carried on the vehicle, including high-resolution measurements of the atmospheric and plasma density, composition, and optical emission spectra or intensity. Another possibility is to use a flow tracer to provide a visual indicator of the small-scale tracer of the atmospheric structure. Figure 1 shows an image of a TMA trail released from the TOMEX rocket, launched as described above. The turbulent structure in the trail is evident up to an altitude of approximately 100 km. Between 100 km and 110 km the trail becomes laminar but shows evidence of enhanced horizontal diffusion rates associated with anisotropic turbulence. At still higher altitudes, the trail expansion rates are consistent with the expectations for molecular diffusion (Bishop et al., 2004). Figure 2 shows a detailed view of the same trail. Specifically, there is a high-shear region that is evident in the lower left-hand section of the image. The nearly horizontal section of the trail shows the presence of billow structure associated with a shear instability. In this case, the Starfire resonance lidar provided detailed time histories of the wind shears and vertical temperature gradients across the region, showing the development of the conditions required for shear instability. The TMA trail provided direct visual evidence of the billows associated with the instability, including the horizontal wavelength of the unstable structures. A similar result was obtained in the August 2002 SEEK-2 launch from Japan (Larsen et al., 2005). The trail from that launch, shown in Figure 3, was released across the rocket apogee, producing a nearly horizontal trail. That release occurred in a high-shear region associated with a sporadic E layer, and the Kelvin-Helmholtz billows are clearly seen in the trail structure.
Figure 1. Trail image from the October 2000 TOMEX launch showing the variation of the turbulent structure with altitude across the mesosphere-lower thermosphere region.

Figure 2. Billow structure in high-shear region is evident in the lower left-hand side of the image.
4. Combined measurements

The combination of ground-based lidar and in-situ measurements offers the possibility of measuring parameters that neither technique can provide by itself. The expense and logistical complexity associated with the rocket launches will always limit the number of launches that can be carried out. The ground-based instrumentation therefore provides critical time history and contextual information about the dynamical conditions in the background atmosphere. The studies described by Hecht et al. (2004), Williams et al. (2006b), and Collins et al. (2011) present examples in which the ground-based instrumentation provides contextual information that is critical for the interpretation of the in-situ measurements. The lidar and in-situ measurements are also complementary in the sense that lidar, for example, can provide height profiles of the Prandtl number that cannot be obtained directly from in-situ measurements. In-situ measurements can provide high-resolution measurements of turbulent fluctuations or turbulent diffusion effects that cannot be obtained directly with the ground-based measurements. A notable example is TOMEX, described in more detail above, in which lidar measurements provided detailed information about the range of altitudes where the conditions for convective and dynamical instability were met. The trail expansion rates in those regions then showed that the horizontal eddy diffusion was larger in the regions characterized by convective instability than in the regions with dynamical instability (Hecht et al., 2004; Bishop et al., 2004).

5. Extension to higher altitudes

The various studies described above have focused on neutral dynamics in the mesosphere and lower thermosphere, but such studies represent a relatively small fraction of the in-situ measurements that have been carried out. The majority of sounding rocket experiments have been related to plasma physics and the electrodynamics of the ionosphere. A significant limitation of such studies, especially those that deal with the electrodynamics of the strong current region in the lower E region, is the lack of detailed information about the neutral behavior in that altitude range. In the critical altitude range between 90 and 120 km, for example, the neutrals are the ultimate drivers for the plasma processes at mid and low latitudes. At high latitudes, the neutrals compete with magnetospheric processes, acting as drivers in some cases and as energy and momentum sinks in others. The improved capabilities
envisioned for the proposed lidar instrumentation can provide important new perspectives on the physics of the ionosphere, especially in combination with in-situ and ground-based measurements of ionospheric parameters.

6. Summary

It is clear that the combination of ground-based lidar measurements and in-situ rocket measurements can produce significantly greater insights than either measurement technique alone. The value of this combination has already offered enough motivation to justify the investment of resources required to establish temporary launch facilities in the vicinity of state-of-the-art lidar instrumentation in at least two campaigns in the past. Significant resources have also been invested in developing lidar and other optical instrumentation at existing launch facilities, notably Poker Flat and Andøya. It is important that options for carrying out in-situ measurements close to the proposed lidar instrumentation be considered in planning the new facility, either by establishing the new facility at or near an existing launch facility or at a location where temporary launch operations can be carried out.

References


Williams, B. P., D. C. Fritts, J. D. Vance, C.-Y. She, T. Abe, and E. Thrane (2006c), Sodium lidar measurements of waves and instabilities near the mesopause during the DELTA rocket campaign, *Earth Planets Space*, 58, 1131-1137.
Global Navigation Satellite Systems

Jonathan Makela
Department of Electrical and Computer Engineering
University of Illinois at Urbana-Champaign
Urbana, Illinois

Introduction

The use of signals transmitted by the various Global Navigation Satellite Systems (GNSS; e.g., the US Global Positioning System, the Russian GLONASS, and the European GALILEO system) in studying the ionosphere has played an important role over the past decade in unraveling the dynamics of this region of the upper atmosphere. A dual-frequency receiver can estimate the ionospheric integrated total electron content between the satellite and itself by making measurements of the relative arrival times of the signals transmitted by a given GNSS satellite. Due to the dispersive nature of the ionosphere, the signals transmitted at the two frequencies (i.e., for GPS L1 = 1.57542 GHz and L2 = 1.2276 GHz) are delayed with respect to one another by an amount proportional to the integrated electron density along the raypath:

$$\Delta(\delta t^{\text{ino}}) = \frac{40.3}{c} TEC \left( \frac{f_{L1}^2 - f_{L2}^2}{f_{L1}f_{L2}} \right),$$

(1)

where $\Delta(\delta t^{\text{ino}})$ is the relative delay in the reception of the two signals at the receiver, $f_{L1}$ and $f_{L2}$ are the carrier frequencies of the two signals, $c$ is the speed of light in vacuum, and $TEC$ is the total electron content. A dual-frequency receiver estimates the integrated electron density between itself and the transmitting satellite by measuring $\Delta(\delta t^{\text{ino}})$ and solving the above equation for $TEC$.

In addition to the propagation delays caused by the free electrons in the ionosphere, an additional delay source comes from the changes in the refractive index in the lower atmosphere. As with the $TEC$, measurements of the delays on GNSS signals can be analyzed to provide estimates of the water vapor in the lower atmosphere (e.g., Bevis et al., 1992). The GPS-derived technique provides a low-cost alternative to traditional (e.g., radiosondes) methods for making larger-scale observations of the lower atmosphere.

The power in utilizing GNSS-based TEC estimates comes from the fact that a given dual-frequency receiver can simultaneously make an estimate for each transmitting satellite in view. For the current GPS constellation, this means that a single dual-frequency GPS receiver at midlatitudes can typically make between 6 and 10 estimates distributed about the sky. Moreover, if an array of receivers is available, measurements can be combined and spatial maps of the TEC distribution over an area can be produced. For example, The ionospheric community has leveraged the deployments undertaken by the seismic community and routinely produced maps of the TEC at a variety of spatial and temporal scales.
Ground-based deployments

In North America, the Continuously Operating Reference Station (CORS) network (http://geodesy.noaa.gov/CORS), managed by the National Geodetic Survey within the National Oceanic and Atmospheric Administration’s National Ocean Service, provides data from over 1800 receivers at cadences ranging from 1 to 30 seconds. The current distribution of receivers in the contiguous United States is shown in Figure 1. Additional networks of receivers exist around the world and data from these networks can be combined to create global maps of the TEC distribution (e.g., Iijima et al., 1999; Rideout and Coster, 2006).

![Figure 1. Current deployment of dual-frequency GNSS receivers that are part of the CORS network operated by the National Geodetic Survey (http://geodesy.noaa.gov/CORS_Map).](image)

A large body of literature has developed around the use of the TEC maps derived from these networks. The spatio-temporal context that these TEC maps provide complements the detailed local observations of the ionospheric state provided by other measurement techniques such as incoherent scatter radars, lidars, and in-situ satellite probes. As an example, Figure 2 shows the TEC derived from the CORS network at 00:50 UT on 08 November 2004 during a major geomagnetic storm. In addition, the ion drifts measured at an altitude of ~840 km by two Defense Meteorological SATellite Program (DMSP) satellites are presented, showing enhanced westward flow over the northeast, typical of the sub-auroral polarization stream (SAPS) phenomena (e.g., Foster and Burke, 2002). From the spatial information provided by the GPS maps, the enhanced velocities are seen to be collocated with the auroral oval (indicated by the reduced electron content), which expanded over the northern United States during this storm. Importantly, by combining these two measurement techniques, Basu et al. (2008) studied how this storm adversely affected critical technologies such as the FAA’s Wide Area Augmentation System (WAAS), which was unavailable over North America for the duration of the storm. In this study, as with many similar studies, the TEC maps provided the crucial linkage between several other datasets.
Even during non-magnetically disturbed days, dense networks of GNSS receivers can be used to study the spatio-temporal aspects of the ionospheric electron density. Two examples are shown in Figure 3 and Figure 4. Figure 3 shows propagating medium-scale traveling ionospheric disturbances over North America (Tsugawa et al., 2007). These waves had long been studied using airglow imaging systems (see review by Makela and Otsuka, 2012), but the TEC maps provided a much larger context from which it was possible to understand the extent of these structures. Figure 4 shows the ionospheric response to the atmospheric waves launched by the large Tohoku earthquake in 2011 (Rolland et al., 2011). The TEC data collected during this event and other moderate-to-large earthquakes are providing valuable insight into the generation and properties of atmospheric gravity waves that reach into the thermosphere and interact with the near-space plasma environment. Data from a lidar system could provide invaluable direct measurements of the properties of the generated gravity waves which, to date, have had to be studied through simulations and models (e.g., Occhipinti et al., 2011).

Figure 3. Examples of medium-scale traveling ionospheric disturbances (MSTIDs) captured using the CORS network of GNSS receivers. The MSTIDs are seen as alternating bands of enhanced and depleted electron content aligned from the northeast to southwest. After Tsugawa et al. (2007).
Figure 4. Waves seen in the TEC measured by the Japanese GEONET network after the Tohoku earthquake in March 2011. The circular rings of alternating enhancement and depletions in TEC are seen to propagate away from the earthquake’s source region. After Rolland et al. (2011).

Space-based deployments

An alternate viewing geometry for utilizing the signals transmitted by GNSS satellites to probe the atmosphere is by flying a receiver on a low-earth orbiting satellite. Although an analog to the ground-based TEC exists for measurements from the LEO satellite to the GNSS satellites above, resulting in TEC estimates for plasma above the dense ionospheric region, perhaps the more useful measurement comes from observations whose lines-of-sight are to GNSS satellites setting behind the earth as shown in Figure 5. These occultation measurements can be inverted to provide altitudinal profiles of the water vapor and electron content of the ionosphere (e.g., Schreiner et al., 1999; Hajj et al., 2002). Recently, the joint US-Taiwanese COSMIC program has utilized the occultation geometry to provide global measurements of the ionospheric density profiles, launching 6 satellites in April 2006. The data from the COSMIC program form the basis for a space-weather assimilative model, providing valuable measurements of the state of the ionosphere in regions where ground-based observations are impossible (i.e., over the ocean). There is a planned COSMIC-2 follow-on mission which will further enhance the amount of observations available from the radio occultation technique in the future (currently, launch of the first phase of COSMIC-2 is scheduled for 2016).
Concluding Remarks

With the anticipated increase in the number of GNSS-based signals in the near future with the completion of several additional systems to the current GPS and GLONASS systems (e.g., GALILEO, COMPASS), as well as the completion of the ongoing modernization of the GPS which will increase the number and utility of the signals available to the non-military communities, it is expected that the GNSS-based TEC technique will be able to provide excellent measurements of the background electron density. This would complement and provide context for the detailed observations to be made by a large aperture lidar facility. These capabilities are leveraged from investments made by a variety of agencies and would come at no cost to the large aperture lidar facility.

References


Correlative and in-situ measurements complementing the Large Aperture Lidar

Patrick Espy
Department of Physics
Norwegian University of Science and Technology
Trondheim, Norway

Introduction

In order to quantify the state of the space-atmosphere interaction region (SAIR) from 30 to 1000 km in terms of variability, energy inputs and energy transfer, there are several correlative and in-situ observations that would represent value added to the main lidar measurements. The scientific questions to be addressed by the suite of instruments can be divided into two broad atmospheric regions.

Lower atmosphere (30-100 km)

- Dynamical energy inputs into the upper atmosphere
- Chemical and dynamical effects of energy from the upper atmosphere (solar and particle)

Upper atmosphere (100-1000 km)

- Energy inputs of solar and particle energy
- Direct effects of solar and particle energy from above on ions and neutrals
- Effects of dynamical energy inputs from below on ions and neutrals

The proposed lidar will be able to address many of these goals with unparalleled accuracy as well as time and altitude resolution. However, there are several additional instruments that can give value added information as well as observations unavailable to the lidar. This instrumentation falls broadly into two different categories: optical and radar observations. These two categories may also be characterized as those observations requiring clear skies, like lidar, and those that do not require optical seeing conditions, like radar.

Optical Instrumentation

After sunset, many atmospheric species that have been dissociated or ionized by sunlight begin to recombine in a series of reactions that emit radiation unique to those species. Since these nightglow emissions take place in relatively narrow layers at different altitudes, spectroscopic observations have been used to remotely sense upper atmospheric temperature, winds, composition, and chemistry, as well as the effects of dynamics on these parameters since the early 1950’s. As an example, Figure 1 shows the emission layers in the earth limb from the hydroxyl (OH) near infrared emissions at 87 km, the atomic sodium (Na) 589 nm doublet at 90 km, the oxygen (O) at 557 nm emission at 95 km, and the OI 630 nm emission at 250 km that results from dissociative recombination of the oxygen molecular ion. Similarly, particle precipitation from the magnetosphere into the atmosphere creates excitation, dissociation, and ionization processes, many of which result in characteristic emissions. Spectroscopy of these emissions has been used since the time of Ångström to ascertain the effect of these energy inputs into the atmosphere and the chemical and dynamical processes that result.
Although the spectroscopic instruments themselves have remained largely unchanged, detector technology has made rapid advances in the visible and near-infrared regions driven by the camera and communications industry. Array detectors have replaced film in cameras, and the industry has rapidly expanded the size, resolution, sensitivity and wavelength response of these devices to unprecedented levels. As a result, optical detectors are now available commercially that would have been unaffordable in the past. Spectrometric devices utilizing these detectors have seen improvements in signal to noise levels by factors of 15 to 30. Additionally, the size and resolution of array detectors have virtually eliminated the need for scanning, allowing additional gains from the resulting Fellgett advantage. As a result, spectroscopic night-airglow and auroral instruments are now able to observe emissions from multiple species over large spectral regions, as well as atomic line shapes and Doppler shifts using ultra-high resolution on time scales commensurate with the proposed lidar. Hence, while the lidar observes concentration, winds, and temperatures on one species, other species chemically and dynamically related to those observed by the lidar can also be quantified. Similarly, monochromatic imaging is now capable of routinely observing the two-dimensional temperature and density structure of waves propagating upward into the thermosphere, and multiple imagers may be able to observe the 3-d structure of these waves.

Airglow Imagers: One of the most important complements to the lidar observations would be the observation of the two dimensional structure at multiple altitudes using multiple wavelengths at moderate resolution. This will allow the wave scales, orientation, momentum flux into the SAIR, and the wave forcing of the mesosphere and lower thermosphere (MLT) to be quantified. By monitoring OI emissions in the thermosphere, it is possible to map regions of plasma instability associated with F-region plumes.

In order to accommodate the different integration times required for the different intensity airglow emissions and mindful of the decreasing costs of such systems, multiple imagers, each monitoring a specific emission, are suggested. Additionally, multiple sites around the lidar with key emissions duplicated would allow triangulation on the emission in order to establish its altitude as well as to determine the three dimensional structure of waves or F-region instabilities, as shown in Figure 2. Such measurements would be critical for supplementing the vertical measurements of the lidar so as to uniquely quantify the momentum flux and forcing of the atmosphere in relation to the lidar observed small-scale instabilities, heat, and momentum flux.
Figure 2. Tomography of the 3-D structure of wave structure in the hydroxyl airglow showing the altitude of the emission and the six horizontal slices, separated by 1.2 km (M. Taylor).

A table of suggested emissions to be monitored as well as their wavelengths and emission altitudes is given in the table below. Also noted in the table is the susceptibility of contamination by aurora, should the facility be located at a high-latitude site.

### Table 1: Emissions to be Monitored

<table>
<thead>
<tr>
<th>Emitting species</th>
<th>Wavelength</th>
<th>Altitude km</th>
<th>Background Contamination</th>
</tr>
</thead>
<tbody>
<tr>
<td>OH</td>
<td>650-800 nm</td>
<td>87</td>
<td>Heavily contaminated by aurora</td>
</tr>
<tr>
<td>OH</td>
<td>1500-1600 nm</td>
<td>87</td>
<td>Lightly contaminated by aurora</td>
</tr>
<tr>
<td>Na</td>
<td>589 nm</td>
<td>90</td>
<td>Lightly contaminated by aurora</td>
</tr>
<tr>
<td>OI</td>
<td>557.7 nm</td>
<td>95</td>
<td>Heavily contaminated by aurora</td>
</tr>
<tr>
<td>OI</td>
<td>630.0 nm</td>
<td>250</td>
<td>Heavily contaminated by aurora</td>
</tr>
<tr>
<td>Background</td>
<td>600 nm</td>
<td></td>
<td>Heavily contaminated by aurora</td>
</tr>
<tr>
<td>Background</td>
<td>1490 nm</td>
<td></td>
<td>Lightly contaminated by aurora</td>
</tr>
<tr>
<td>N2 1P</td>
<td>670 nm</td>
<td>91-150 km</td>
<td>Aurora, no airglow contamination</td>
</tr>
</tbody>
</table>
**Spectrometers**: Operating at moderate to high resolution (0.5 to 2 nm), spectrometers may be used to measure large free-spectral ranges. In this way, multiple spectral intensities from both the aurora and the nightglow emissions can be observed simultaneously and compared to give information on chemical processes and the auroral characteristic energy. This large wavelength coverage and high resolution allows any background light emissions to be quantified unambiguously. This allows precision measurement of rotational temperatures of nightglow and auroral molecular bands, in turn allowing the determination of the neutral temperature at the height of the airglow emission and, in the case of aurora, at the peak deposition altitude.

For a mid- to low-latitude facility, a spectrometer covering the various night-glow emissions to be observed using imagers would allow backgrounds, line ratios, and rotational temperatures to be observed simultaneously with the imaging measurements. For a high latitude facility, a range of allowed and forbidden auroral emissions should be observed to allow characteristic precipitation energies to be observed. For example, forbidden emissions from OI (557 nm and 630 nm) and NI (3F(1040 nm), which are quenched, can be compared to allowed transitions from N$_2^+$-1N system (390-1800 nm) and N$_2$1P (650-1300 nm) to determine energy deposition altitude. In addition, the rotational structure of excited molecular bands can be used to determine the local temperature, and lines such as the He-1083 nm line can be used to indicate high-altitude deposition and solar fluorescence effects. Finally, observation of hydrogen Balmer emissions can be used to deduce the proportion of proton flux in the aurora, and their shift from the rest wavelength used as an indicator of the incoming proton energy.

**Fabry-Perot Interferometers**: Operating with exceptional resolving power, albeit limited free-spectral range, these systems can measure the profile of an emission line and its offset from the rest wavelength. Thus, these systems can be used to observe temperature through the Doppler line width, the excitation mechanism (direct or dissociative recombination) through the total line width, and the line-of-sight velocity through the offset from the rest position. Both mesospheric and thermospheric emissions have been observed and, most critically, vertical winds may be deduced if an optical frequency standard is available. Atomic line profiles from, for example, OI, NaI, and He $^2$S, as well as individual rotational lines from OH Meinel molecular bands, have been used to measure neutral temperatures and winds near the mesopause (87-95 km) and the thermosphere (300-800 km).

Simultaneous observations of multiple lines, providing winds and temperatures throughout the mesosphere and thermosphere, would be optimized with separate optical front ends. If these systems could be co-located with the multiple imager observation stations then the tri-static measurements could provide meridional and zonal wind components on rapid time scales. These observations would greatly enhance the lidar measurements that require separate look directions.

**Radio wave Instrumentation**

**Active Radars measurements**: Radar observations fall broadly into two categories: coherent scatter using the plasma resonance frequency, and incoherent or Thomson scatter making use of free electrons in the ionospheric plasma. Coherent-scatter radars, such as MF or meteor radars, are typically used in the mesosphere and lower thermosphere to measure electron densities as well as neutral winds, temperatures, and the momentum flux from waves. Crossing over in the upper mesosphere, incoherent-scatter techniques measure ion and electron temperatures and winds, and can be used to infer neutral properties up to 1000 km. In addition, these systems can be used to measure the profile of energetic particle deposition directly and thus determine the characteristic energy. Thus, a combination of these techniques would characterize the wind and temperature fields throughout the radar region: approximately 65 to 1000 km with temporal resolutions of ~30 minutes, making them a powerful complement to the lidar techniques proposed.

Recently, a multiple-beam meteor radar technique has been used to measure the variance due to gravity waves and the direction of propagation of the waves, allowing the vertical flux of horizontal
momentum to be inferred. As shown in Figure 3, the radar-measured gravity wave variance indicates a momentum flux towards the NE. A simultaneous all-sky OH imager shows the wave pattern and its motion towards the NE, in agreement with the radar observations. The distinct advantage of the radar technique, however, is that such measurements are available during cloudy weather and moon-up conditions that render optical measurements impossible. Thus, similar radar measurements would provide continuous monitoring of not only the background winds and tides but also of the momentum flux into the thermosphere and ionosphere system. A companion incoherent scatter radar would be able to track the growth rate of tides and perhaps gravity waves high into the thermosphere. This combination of lidar with coherent and incoherent radars would quantify the forcing and coupling of the SAIR from 30 to 1000 km.

![Figure 3](image1.png)

**Figure 3.** Top: gravity wave momentum flux measured in the beams of the meteor radar showing a maximum towards the NE between 20 and 22 UT. Bottom: All sky OH image of wave structure at 21 UT with direction of propagation towards NE marked. (R. Hibbins)

**Passive radiometer measurements:** Recently a standard technique used in the stratosphere has been adapted for mesospheric use. Using passive thermal molecular emissions in the atmosphere at mm and sub-mm wavelengths, the collision broadened line shape of the altitude integrated spectrum can be inverted to yield a profile of the emission. Thus, the volume mixing ratio (vmr) of ozone, nitric oxide, and other molecules can be measured between 30 and 75 km with an altitude resolution of 8-15 km and temporal resolutions of ~30 min. Since the chemical lifetime of ozone determines its concentration above 30 km, variations in its vmr can be used to infer the temperature oscillations associated with wave motions. For example, temperature oscillations occurring in the OH nightglow at 87 km, indicating the passage of a 16-day planetary wave, can be correlated with the observed ozone oscillations to map the phase fronts of the wave down to 30 km, as shown in Figure 4a. Here one can see that the vertical phase fronts associated with this normal mode oscillation are interrupted in the region of large negative wind gradients above the zonal wind maximum due to absorption and transient propagation of the wave.
Given multiple spectrometers, the signals from ozone and nitric oxide may be simultaneously measured to allow the effects of energetic particle precipitation on the neutral atmosphere to be observed. Since nitric oxide (NO) reacts catalytically with ozone, NO produced by the particle precipitation associated with very moderate magnetic storms can lead to significant ozone destruction. When this occurs during the polar winter night, the effects of the NO are long lived and are transported downward into the polar stratosphere. Figure 4b shows the Dst and AE storm indexes in the top panel and the ~70% ozone loss due to catalytic chemistry in the lower panel. The bottom panel shows the nitric oxide produced by energetic particle precipitation during the storm.

Figure 4. Left panel (a): The time-lagged cross-correlation between winter mesospheric temperature and ozone at different levels. Red color represents positive and blue denotes negative correlation.

Right panel (b): Catalytic ozone destruction from 30 to 85 km by nitric oxide produced by energetic particle precipitation over Troll Research Station, Antarctica.
Exospheric Studies

E.J. Mierkiewicz
Department of Physical Sciences
Embry-Riddle Aeronautical University
Daytona Beach, Florida

J.W. Meriwether
Department of Physics and Astronomy
Clemson University
Clemson, South Carolina

In the tenuous uppermost reaches of an atmosphere, often referred to as a corona or exosphere, important interactions take place in which mass, momentum and energy are exchanged between the atmosphere and its interplanetary environment. What processes drive the development and evolution of an exosphere and how do processes taking place within the exosphere (e.g., escape) influence the evolution of the atmosphere below? The answers to these questions are almost certain to be found by comparing and assessing planetary and minor body exospheres throughout our solar system and perhaps beyond.

Like all exospheres, low ambient number densities and correspondingly large mean free paths characterize the terrestrial exosphere. The net result is a vanishingly small collision frequency that ultimately leaves the exospheric velocity distribution to be determined solely by the independent motion of gravitationally bound particle populations traveling on Keplerian trajectories (e.g., ballistic, satellite, escape) [Chamberlain and Hunten, 1987].

The lower boundary of the terrestrial exosphere, defined by the transition between the collision dominated regions below and the nearly collisionless regions above, is typically centered near 450 km. The upper boundary of the exosphere is somewhat ill defined as the neutral atmosphere gradually merges with the interplanetary medium. A theoretical outer boundary, however, can be considered as the level at which solar radiation pressure exceeds the earth’s gravitational influence on the exospheric particle populations (∼200,000 km) [Bishop, 1985, 1991].

Since the 1970s [Reynolds et al., 1973] ground-based Fabry-Perot observations of geocoronal hydrogen Balmer-alpha nightglow (6563 Å) have become one of the primary methods to study the terrestrial exosphere. Observations are made throughout the night and the base of the earth’s shadow is used as a first-order probe of the exosphere’s altitude structure. Major areas of scientific focus include: (1) high-resolution observations of the geocoronal hydrogen Balmer-alpha line profile and its relation to excitation mechanisms, effective temperature, and exospheric physics [Atreya et al., 1975; Meriwether et al., 1980; Yelle and Roesler, 1985; Kerr et al., 1986; Kerr and Hecht, 1996; Nossal et al., 1997, 1998; Mierkiewicz et al., 2012]; (2) retrieval of geocoronal hydrogen parameters such as the hydrogen column abundance [H], the hydrogen density profile H(z), and the photochemically initiated hydrogen flux Φ(H) [e.g., Bishop et al., 2004; He et al., 1993; Kerr and Tepley, 1988]; and (3) long-term observations of the geocoronal hydrogen column emission intensity for the investigation of natural variability, such as solar cycle trends, and of potential anthropogenic change due to increases in atmospheric concentrations of greenhouse gases [e.g., Nossal et al., 2008; Kerr et al., 2001a].

Building on the application of advanced Fabry-Perot spectrometers for geocoronal hydrogen observations (see e.g., Mierkiewicz et al., 2006a), refined radiative transport codes with realistic atomic hydrogen profiles (Bishop, 1999, 2001; Bishop et al., 2001), and encouraging results from parametric data-model search procedures (Bishop et al., 2004), we are uniquely placed to make new progress in these important areas of research; see e.g., Figure 1. In addition, a new device, the Balmer-alpha Field-Widened Spatial Heterodyne Spectrometer (FW-SHS) is poised to make significant contributions to
geocoronal science in the next few years. The FW-SHS is modeled after the successful application of the SHS technique for ionospheric and astrophysical O+ 3727 Å emission line studies (Mierkiewicz et al., 2006b; Fallest et al., 2005). The SHS itself is an interferometric Fourier transform spectrometer, but unlike conventional Fourier transform spectroscopy (FTS) it requires no moving parts to obtain a spectrum. Field-widening can be achieved to obtain throughput increases of about 100 over Fabry-Perot instruments of similar size. These characteristics make the FW-SHS a superior instrument in many applications (Roesler et al., 2003).

The FW-SHS is expected to have the sensitivity and resolving power of current 150mm Fabry-Perot instruments while being smaller, more robust, and more easily transported and set up at remote sites. In this regard the FW-SHS will be an ideal instrument to deploy in support of a future lidar facility. As colleagues at the University of Illinois prepare to deploy a helium (He) 10830 Å resonance lidar system to make the first lidar measurements of the Earth’s upper thermosphere, the Balmer-alpha FW-SHS will be capable of being located at the lidar site for coincident observations. This campaign will allow the first combined passive and active sounding of the upper thermosphere (near 500 km altitude) by optical techniques.

![Figure 1](image.png)

*Figure 1. Exospheric effective temperatures derived from Balmer-alpha line width observations as a function of LOS shadow altitude (km) separated into pre- and post-midnight sets. A persistent narrowing of the profile with shadow altitude is apparent in the majority of the data with a decrease of ∼470 K in terms of effective temperature from ∼840 K near 500 km to ∼370 K near 20,000 km (Mierkiewicz et al., 2012).*
Figure 2. Exospheric effective temperatures derived from line width observations (for shadow altitudes below 1500 km). Observations each month (i.e., dark moon observing window) were averaged and plotted by (new moon) day number. A semi-annual variation is observed in the column averaged exospheric effective temperature with maxima near day numbers 100 and 300 and minima near day numbers 1 and 200. Solid lines correspond to daily MSIS exobase temperatures, run with daily and average \(F_{10.7}\) flux inputs set to 181 (corresponding to the 2000-2001 average) and 75 (Mierkiewicz et al., 2012).
References


Optical/lidar Workshop Report

Introduction

Although generally invisible to the naked eye, the night sky is a rich source of UV, visible, and infrared emissions that are signatures of the chemistry, physics, and dynamics that are taking place in the upper atmosphere. We are familiar with the remarkable phenomena of the aurorae at high latitudes, and the so-called nightglow bears many similarities but on a sub-visual scale. Thus, progress in studying and utilizing the nightglow depends on advances in technology that permit measuring low-level light quickly and accurately.

The complexity of the nightglow has been revealed over the years, starting with early studies that identified only a few atomic emissions, to the current realization that there are thousands of atomic and molecular lines [Cosby et al., 2006]. There is in fact no clear viewing region through which the nightglow can be avoided, and thus astronomers are well aware of the importance of making corrections for the nightglow when viewing their target objects. This fact has led to the interesting situation that some of the best nightglow spectra are produced by astronomical instruments at large telescopes, where the intent is not to study the nightglow emissions at all, but to learn how to compensate for them. Such data are referred to as sky spectra.

Sky spectra

A sample sky spectrum is shown in Figure 1, where one sees that the intensities are heavily weighted towards the infrared region, where the dominant emission, from the OH (hydroxyl) Meinel bands, becomes progressively stronger [Cosby and Slanger, 2007]. Figure 2 shows an expanded region, and we see that the line density is quite high. All of these emission lines can be identified, and most of them are due to the excited states of either OH or molecular oxygen, the latter being substantially weaker than the former.

Figure 1. Average nightglow spectrum, from ESI/Keck II
The two sky spectra in Figures 1 and 2 were obtained with one of the 8-meter VLT telescopes in Chile with relatively long exposures, typically 40 minutes. For instruments dedicated to aeronomical studies, we are of course interested in much higher acquisition rates.

The goals of any given research project dictate the type of instrumentation that is appropriate. The typical question is whether a study focuses on a single atomic feature (cf. the oxygen red and green lines), or whether the emphasis is on comparison of features seen over a range of wavelengths. In the former case a technique such as Fabry-Perot interferometry is a possible choice [Meriwether, 2006], while in the latter instance one needs to use a dispersing instrument, one of which was used to obtain the spectrum in Figure 1.

The nightglow is an active phenomenon, exhibiting periods of relative stability interspersed with times when wave motion is very prominent. All-sky imaging is often used to show these large scale movements to best advantage [Taylor et al., 2001].

**CESAR (Compact Echelle Spectrograph for Aeronomic Research)**

The extraordinary quality of the sky spectra from large telescopes prompted the aeronomy community to support the construction of a large echelle spectrograph (CESAR) with many of the same properties, but with the ability to relocate as needed, particularly to higher latitudes. A.L. Broadfoot was instrumental in creating the initial design. CESAR will have the study of aeronomical issues as its primary function.

The spectrograph with grating post-dispersion will be dedicated to nightglow studies at high spectral resolution (R \(\sim\) 20000) between 300 and 1000 nm, and will be easily deployable at different sites. The development of CESAR was conducted by SRI International, and the optical design and integration of the spectrograph camera is based on the camera of the HIRES spectrograph at the Keck I telescope.
The detailed optical design is used to calculate the position of the spectral elements on the detector, predict their image quality, and estimate the level of stray light [Lavigne et al., 2010]. Figure 3 shows the instrumental components, and Figure 4 shows a calibration spectrum from a nitrogen discharge, which will be similar to an auroral spectrum.

**Figure 3. Schematic of CESAR light path**
Combined Lidar/Optical Studies

One of the accomplishments of recent sky spectra studies has been to identify a nightglow component originating with emission from the FeO molecule [Evans et al., 2010; Saran et al., 2011]. This presents us with an interesting opportunity to combine lidar and optical instrumentation to study atmospheric iron, since the FeO source is the reaction of atomic iron and ozone. The missing information in a ground-based study of FeO is the altitude of the emission, and simultaneous iron lidar measurements will at least provide the atomic iron altitude profile.

This line of inquiry can then be extended because reaction with ozone produces other emitting species, in particular the OH emission (Figures 1/2) and the sodium emission at 589 nm (Figure 1). Thus, three distinct emissions are produced, all dependent on ozone, and with an instrument such as CESAR, with broad spectral coverage, it is in principle possible to carry out multi-dimensional analysis related to ozone, iron, sodium, and H-atom profiles. Figure 1 shows the continuum associated with FeO at 500-650 nm, and Figure 5 shows an expanded version in which the various additional features are identified in the spectral region [Saran et al., 2011].

![Figure 5. FeO quasi-continuum, with identification of nearby nightglow features](image)

**OH Nightglow and Temperature**

The OH molecule has long been a target of atmospheric research, for a variety of reasons. The most evident is that it provides an easily accessible and intense emission, as seen in Figure 1. Because it is produced by the reaction of H-atoms with ozone, the intensity is related to the concentrations of each. The initial OH product is highly vibrationally excited, and the numerous bands appearing in Figure 1 originate with levels from $v = 3$ to $v = 9$ in the OH ground state. A further important property is that since the radiator is a molecule, the distribution of rotational transitions (Figure 2, a portion of the OH 7-3 band) is considered a measure of the atmospheric temperature at the altitude of emission.

If this last point is verified, then measurement of the intensity distribution in an OH band is a direct temperature probe. However, there are still numerous caveats, and it is hoped that the envisioned
A lidar/optical facility will contribute to clarifying the situation. These issues include questions of rotational equilibration [Dodd et al., 1994], the distribution with altitude of the vibrationally-excited molecules [McDade, 1991], the specific pathways of vibrational relaxation [Llewellyn], and the effects of collisional quenching by relevant air particles, in particular atomic and molecular oxygen.

**Twilight/Dayglow Investigations**

Astronomical sky spectra are generally limited to times when the sky is sufficiently dark for observations, between the hours of evening and morning astronomical twilight. That is not a necessary restriction for an aeronomical instrument. Therefore an optical facility at intermediate and low latitudes can be used to observe what is called the twilight glow but is actually high-altitude dayglow. The dayglow has many of the characteristics of aurorae, so extending sky spectra in this manner will prove to be a valuable tool for atmospheric studies. An example of a twilight glow spectrum and a simulation is shown in Figure 6, where early evening data acquisition showed an O₃ band (b-X 1-1) that is only produced at high altitude, verified by the fact that the associated temperature from the rotational distribution is 900 K, far higher than the 200 K normally found for this oxygen band system [Slanger et al., 2003].

![Twilight glow spectrum of the O₂(b-X) 1-1 band and a 900 K simulation, demonstrating that the band originates at an altitude well above the mesosphere](image)

**Figure 6.** Twilight glow spectrum of the O₂(b-X) 1-1 band and a 900 K simulation, demonstrating that the band originates at an altitude well above the mesosphere

**Atmospheric Helium**

Another example of a quasi-dayglow spectrum is shown in Figure 7, where the helium line at 388.9 nm appears as a feature caused by the solar pumping of metastable triplet He [Sharpee, O’Neill, and Slanger, 2008]. It should be noted that this same transition can be accessed via lidar pumping in the absence of solar illumination, also the case for the helium line at 1.083 μm [Waldrop et al., 2005]. Prominent in the figure is emission from the N₂⁺ (B-X) 0-0 band.
A further investigation of interest related to the twilight/dayglow makes it possible to study the effects of solar flares and coronal mass ejections on the atmosphere. New emissions can be produced when high energy photons slam into the upper atmosphere, and it is likely that such enhanced excitation will even be reflected in the nightglow.

**Relationships to Extra-solar Planets**

The in-depth study of the terrestrial atmosphere that will be afforded by the envisioned lidar/optical facility can be viewed, in the long term, as a contribution to the burgeoning field of extra-solar planet investigations. Studies with the Kepler telescope are making it clear that there are enormous numbers of planets in the universe, and that our solar system can be viewed as a not uncommon example of a star surrounded by a number of diverse planets. It seems probable that many of the planets will be found to have oxygen/nitrogen atmospheres, and, from our local sampling, atmospheres of CO₂. It is therefore to be expected that much of what we learn from our sun’s planets will be directly transferable to the investigation of extra-solar planet atmospheres within the next couple of decades. In a sense, it is not necessary to reinvent the wheel for each atmosphere.

**References**


Lavigne et al. (2010), Proc. SPIE 7735, doi:10.1117/12.857509
Sharpee, B.D. et al. (2008), J. Geophys. Res. 113, A12301
Taylor initials et al. (2001), Geophys. Res. Lett. 28, 1899-1902
Waldrop, L.S. et al. (2005), J. Geophys. Res. 110, A08304
Signal-to-Noise Ratio Calculations for Imaging Bi-Static Rayleigh Lidars

Chester Gardner and Gary Swenson
Department of Electrical and Computer Engineering
University of Illinois at Urbana-Champaign
Urbana, Illinois

Since the invention of the laser in 1961, lidar systems have been developed to measure a wide variety of atmospheric constituents and parameters. Although the mono-static configuration is the most prevalent, bi-static techniques have been proposed for several upper atmospheric studies [e.g. Reagan et al., 1982; Welsh and Gardner, 1989]. In fact Elterman [1951; 1952; 1953] made the very first Rayleigh lidar measurements of stratospheric temperature and aerosol profiles sixty years ago using a bi-static configuration consisting of a vertically pointed searchlight and an off-axis scanning telescope.

For mono-static systems, the laser and receiving telescope are typically co-located. Range resolution is achieved by transmitting a pulsed laser beam and range-gating the backscattered signal collected by the receiving telescope. The time of flight of the laser pulse corresponds to range. For bi-static systems, the laser and telescope are separated horizontally by up to several km. Range resolution is achieved by imaging the backscattered beam onto a detector array. The position of the imaged beam on the detector corresponds to range. Mono-static lidars require pulsed lasers while bi-static systems can employ either pulsed or cw lasers.

Because of the different viewing geometry and detection strategy, the classic lidar equations must be modified to properly calculate the backscattered signal photon count \( N_S \) and background noise count \( N_B \) for the bi-static technique. We assume the laser and telescope are separated by the distance \( d \) (see Figure 1). The laser beam is pointed vertically and the telescope is pointed off-zenith to image the beam onto the detector array. The signal count is given by

\[
N_S(z) = \frac{P_{\text{Laser}} \tau}{\hbar c / \lambda_{\text{Laser}} \rho_c(z) \sigma_{\text{eff}}^C(\lambda_{\text{Laser}}) \Delta z} \frac{A_{\text{Tele}}}{4\pi(z^2 + d^2)} \eta_{\text{Tele}} T_{\text{Atmos}}^2
\]  

Where \( z \) is the altitude of the integrated volume (m), \( P_{\text{Laser}} \) is the average power of the laser beam (W), \( \tau \) is the integration period (s), \( \hbar c / \lambda_{\text{Laser}} (J) \) is the photon energy, \( h = 6.63 \times 10^{-34} \text{ J} / \text{s} \) is Planck’s constant, \( c = 3 \times 10^8 \text{ m} / \text{s} \) is the velocity of light, \( \lambda_{\text{Laser}} (m) \) is the optical wavelength of the laser beam, \( \rho_c(z) \) is the density of the constituent being measured \( (m^3) \), \( \sigma_{\text{eff}}^C \) is the effective backscatter cross-section of the species \( (m^2) \), \( A_{\text{Tele}} \) is the aperture area of the imaging telescope \( (m^2) \), \( \eta_{\text{Tele}} \) is the optical efficiency of the imaging telescope including the quantum efficiency of the detector and \( T_{\text{Atmos}}^2 \) is the 2-way optical transmittance of the atmosphere.

Because of the finite divergence of the laser beam \( (\theta_{\text{Laser}}) \) and the imaging geometry, the actual vertical resolution of the bi-static lidar (see Figure 1 and Welsh and Gardner [1989]) is limited to

\[
\Delta z_{\text{BS Stat}} = \frac{\theta_{\text{Laser}} z^2}{d}.
\]  

Alternatively, to achieve a given vertical resolution, the laser and telescope must be separated by
\[ d = \frac{\theta_{\text{Laser}} z^2}{\Delta z_{\text{BiStat}}}. \]  

(3)

To compute the background noise count we assume the worst-case situation in which the detector integrates for the complete observation period. If a pulsed laser is employed and the detector is only activated when the pulse illuminates the scattering volume, the background noise count could be several orders of magnitude smaller than the worst-case values. For this worst-case scenario, the background noise count is given by

\[ N_B = \frac{S_{\text{Sky}}(\lambda_{\text{Laser}}) A_{\text{Te}} \Delta \lambda \tau}{hc / \lambda_{\text{Laser}}} \Omega_{\text{FOV}} \eta_{\text{Te}} \]  

(4)

where \( S_{\text{Sky}}(\lambda) \) is the sky’s spectral radiance (W/m²/nm/sr), \( \Delta \lambda \) is the optical bandwidth of the imaging telescope (nm) and \( \Omega_{\text{FOV}} \) is the solid angle field-of-view of the telescope. For the bi-static geometry illustrated in Figure 1

\[ \Omega_{\text{FOV}} = \theta_{\text{Laser}} \theta_{\text{sz}} = \frac{\theta_{\text{Laser}} d \Delta z}{z^2 + d^2} \]  

(5)

and \( \Delta z \) is the effective vertical resolution of the detector (m), which may be different from the fundamental resolution of the bi-static configuration, which is given by (2). If the detector resolution is matched to \( \Delta z_{\text{BiStat}} \), then

\[ \Omega_{\text{FOV}} = \frac{z^2}{z^2 + d^2} \theta_{\text{Laser}}^2; \quad \theta_{\text{Laser}}^2 \]  

(6)

which is the solid angle field-of-view of a conventional mono-static lidar. In this case, the signal and background counts and their ratio are given by

\[ N_S(z) ; \quad \frac{P_{\text{Laser}} \tau}{hc / \lambda_{\text{Laser}}} \rho_C(z) \sigma_{\text{off}}^C(\lambda_{\text{Laser}}) \frac{\theta_{\text{Laser}} A_{\text{Te}}}{4 \pi d} \eta_{\text{Te}} \Theta_{\text{Atmos}}^2 \]

\[ N_B ; \quad \frac{S_{\text{Sky}}(\lambda_{\text{Laser}}) A_{\text{Te}} \Delta \lambda \tau}{hc / \lambda_{\text{Laser}}} \theta_{\text{Laser}}^2 \eta_{\text{Te}} \]

\[ \frac{N_B}{N_S(z)} ; \quad \frac{S_{\text{Sky}}(\lambda_{\text{Laser}}) \Delta \lambda}{P_{\text{Laser}}} \frac{4 \pi d \theta_{\text{Laser}}}{\rho_C(z) \sigma_{\text{off}}^C(\lambda_{\text{Laser}}) \Theta_{\text{Atmos}}^2}. \]  

(7)

Notice that the signal count (\( N_S \)) and vertical resolution (\( \Delta z_{\text{BiStat}} \)) are largest (see (7) and (2)) when the separation between the telescope and laser is smallest.

The major disadvantage of the bi-static technique compared to a mono-static lidar is the larger background noise contamination that results when the detector integrates for the full observation period, instead of just when the scattering volume is illuminated by the pulsed laser for a mono-static lidar. For example, if a mono-static system is designed to observe Rayleigh scattering to 200 km altitude at 5 km vertical resolution, then a ranged-gated detector would observe a minimum factor of 40 = 200/5 reduction.
in background noise for the maximum possible laser pulse rate of 750 pps. Larger noise reductions are possible when lower pulse rates are employed.

**Figure 1.** Schematic of bistatic and monostatic lidar configurations
References

L. B. Elterman (1951), The measurement of the stratospheric density distribution with the search light technique, *Journal of Geophysical Research*, 56, 509-520

L. B. Elterman (1953), A series of stratospheric temperature profiles obtained with the search light technique, *Journal of Geophysical Research*, 58, 519-530

L. B. Elterman (1954), Seasonal trends of temperature, density and pressure to 67.5 km obtained with the search light probing technique, *Journal of Geophysical Research*, 59, 351-358


Participants of the NSF-sponsored workshop on Exploring the Interaction of Earth's Atmosphere with Space, 15-17 May 2012, University Club of Chicago, IL.


**Acknowledgements:** The Chicago Workshop and the preparation of the reports OASIS – Exploring the Interaction of Earth’s Atmosphere with Space and OASIS – Engineering and Technical Supplement were funded by the NSF Division of Atmospheric and Geospace Sciences under grants AGS 11-15725 and AGS 11-62271 to the University of Illinois at Urbana-Champaign.

We thank Joyce C. Mast for her graphic design contributions to both documents and Stephen A. Hall and Gary R. Swenson for editing the Supplement.