A Simple Instrument and Technique for Measuring Columnar Water Vapor via Near-IR Differential Solar Transmission Measurements

J. A. Reagan, Fellow, IEEE, K. J. Thome, and B. M. Herman

Abstract—This paper describes a simple two-channel solar radiometer and data retrieval technique for sensing the columnar content of atmospheric water vapor via differential solar transmission measurements in and adjacent to the 940-nm water vapor absorption band. The instrument features two parallel channels for simultaneous measurements in and out of the absorption band to eliminate temporal variability effects in the differential comparison of the data from the two channels. The water vapor transmittance is determined by a modified Langley plot analysis of the ratio of the two channel signals. A statistical band model which closely follows the square-root law is then used to extract the columnar water vapor amount from the water vapor band transmittance. Error analyses and experimental results indicate that the instrument/technique can be reasonably employed to retrieve water vapor amounts with an error of 10% or less.

Keywords—Differential Transmission, solar radiometry, water vapor absorption.

I. INTRODUCTION

STARTING with the work of Fowle [1], many investigators have used differential solar transmission measurements in the near-IR to retrieve estimates of atmospheric columnar water vapor. Though simple in concept, practical implementation of this approach has often been limited by laborious calibration procedures, instrumental instability, and inadequate modeling of the water vapor transmittance. In an earlier paper [2], the authors presented a self-contained approach for retrieving columnar water vapor which did not require calibration by transfer/comparison to radiosonde or microwave radiometer measurements. This paper presents a further simplification of this self-contained calibration/retrieval procedure and describes a simple two-channel solar radiometer for retrieving columnar water vapor via measurements in and adjacent to the 940-nm water vapor absorption band.

II. DIFFERENTIAL TRANSMISSION TECHNIQUE

The directly transmitted solar spectral irradiance \( F_\lambda \) (in \( W \cdot m^{-2} \cdot nm^{-1} \)) at wavelength \( \lambda \) received at the Earth’s surface may be related to the exoatmospheric or zero-airmass solar spectral irradiance \( F_{0\lambda} \) by the Bouguer–Lambert law

\[
F_\lambda = \left( \frac{R_m}{R} \right)^2 F_{0\lambda} T_\lambda
\]

where \( m \) is the atmospheric relative airmass for the solar position (zenith angle) at the time of observation of \( F_\lambda \), \( \tau_\lambda \) is the atmospheric spectral optical depth at \( \lambda \) excluding any gaseous spectral absorption, \( \tau_{g\lambda} \) is the atmospheric gaseous spectral absorption optical depth, \( T_\lambda \) is the atmospheric spectral transmittance at \( \lambda \) for airmass \( m \) and total optical depth \( \tau_\lambda + \tau_{g\lambda} \), and \( F_{0\lambda} \) is the exoatmospheric solar spectral irradiance at \( \lambda \) for mean the Earth–Sun separation distance \( R_m \) and \( R \) is the separation at the time of observation. The relative airmass is the ratio of slant path to vertical distances through the atmosphere; it is equal to the secant of the solar zenith angle for a simple plane parallel Earth–atmosphere model. For the remainder of the paper, all expressions and data are normalized to the mean Earth–Sun separation (i.e., \( R = R_m \)).

For a ground-based solar radiometer with collection area \( A_R \) and bandpass \( \Delta \lambda \) centered on \( \lambda \), the quasimonochromatic output \( V_\lambda \) produced by the radiometer when pointed at the Sun (assuming a receiver field-of-view (fov) greater than that subtended by the solar disk) is related to \( F_\lambda \) by

\[
V_\lambda = A_R \int_{\Delta \lambda} R_\lambda F_\lambda d\lambda
\]

where \( R_\lambda (V \cdot w^{-1}) \) is the system spectral responsivity including the spectral dependence of both the photodetector and the wavelength/bandpass selection system of the spectroradiometer. Here, it is assumed that \( F_\lambda \) is constant over the relatively small receiver collection area \( A_R \) and diffuse light within the receiver fov is assumed to be negligible. This latter assumption should hold quite well even for rather turbid conditions for radiometers with small fov’s (receiver full-angles \( < \sim 3^\circ \)) and the longer wavelengths (\( \lambda > \sim 800 \text{ nm} \)) used for water vapor sensing.

If \( \Delta \lambda \) is fairly small, say less than about 10 nm, and \( \lambda \) is in a region excluding any gaseous spectral absorption (i.e., where \( \tau_{g\lambda} = 0 \)), then (3) is well approximated by taking \( \exp(-m\tau_\lambda) \) outside the integral (e.g., Thomason et al. [3]);

\[
V_\lambda = V_{0\lambda} e^{-m\tau_\lambda}
\]
and

\[ V_{0\lambda} = A_R \int_{\Delta \lambda} R_\lambda F_{0\lambda} d\lambda \]  \hspace{1cm} (5)

where \( V_{0\lambda} \) is the radiometer's exoatmospheric or zero-airspace signal level. In this case, the radiometer output signal is of the same form and, thus, obeys the Bouguer–Lambert law. Given that \( \tau_{0\lambda} \) is spatially/temporally constant over the range of airmasses that measurements are made, a plot of \( \ln V_\lambda \) versus \( m \), a Langley plot, will yield a set of data points distributed along a straight line with slope \(-\tau_\lambda\) and intercept \( \ln V_{0\lambda} \).

If \( \Delta \lambda \) is still fairly small but \( \lambda \) is within a gaseous spectral absorption band such as the oxygen and water-vapor bands found in the visible and near-infrared regions, (3) may be expressed in the form

\[ V_\lambda = V_{0\lambda} e^{-m \tau_{0\lambda}} T_{\lambda g} \]  \hspace{1cm} (6)

and

\[ T_{\lambda g} = \frac{A_R \int_{\Delta \lambda} R_\lambda F_{0\lambda} e^{-m \tau_{0\lambda}} d\lambda}{V_{0\lambda}} \]  \hspace{1cm} (7)

where \( T_{\lambda g} \) is the band-weighted transmittance determined by the particular spectral absorption behavior of the gas in question and the spectral features of \( F_{0\lambda} \) and \( R_\lambda \). Even if \( F_{0\lambda} \) and \( F_\lambda \) were effectively constant over \( \Delta \lambda \), the strong spectral variation of \( \tau_{0\lambda} \) typically exhibited by gaseous absorption lines is sufficient to require the band-weighted transmittance \( T_{\lambda g} \). Also, (6) does not follow the Bouguer–Lambert law as \( T_{\lambda g} \) generally cannot be accurately modeled by an exponential with a negative argument of airmass times a constant, band-weighted optical depth.

For water vapor bands in the near-IR, \( T_{\lambda g} \) is well approximated by an exponential with a negative argument proportional to the square-root of the total path absorber amount (e.g., [4], [5]). Following the form used in our earlier paper [2], the water vapor band transmittance is modeled by

\[ T_{\lambda g} = T_{\lambda w} = e^{-b \sqrt{\frac{m}{\mu}}} \]  \hspace{1cm} (8)

where \( m \) is airmass, as given earlier, \( a \) is the equivalent vertical column water vapor amount, \( b \) is a constant depending on the absorption spectra of the band in question and the form of \( R_\lambda \) over the band, and \( a \) is a constant to account for offset as the exponent of \( T_{\lambda w} \) reverts from the \( \sqrt{\mu} \) to a linear \( m \) dependence at very small values of \( \mu \).

To determine the constants \( a \) and \( b \) in (8), we use a statistical band model developed by Thomason [6]. This model transforms the wavelength (actually wavenumber) integration of \( T_{\lambda g} \), via an integral transform, to an integration over absorption coefficient space weighted by an absorption coefficient probability distribution function. The required distribution function can be estimated fairly accurately from representative line parameter information on the band in question. Temperature and pressure inhomogeneities along the path can also be handled fairly easily. As applied to the 940-nm water vapor band, subband transmittances for unity \( R_\lambda \) weights are computed for 1.8-nm intervals in 10 subbands extending from about 930 to 950 nm. The subband transmittances are then sum-weighted by the transmission profile of the interference filter used for the water vapor channel (i.e., \( R_\lambda \) over channel bandwidth is essentially determined by the interference filter transmission profile) to obtain the band transmittance. A plot of the band transmittance computed in this fashion for a nominal 10-nm bandpass interference filter profile centered on 940 nm and for surface conditions representative of Boulder, CO, in September, plotted as \( -\ln T_{\lambda w} \) versus \( \sqrt{\mu} \), is shown in Fig. 1. The straight-line behavior indicated by (8) applies quite well for \( \sqrt{\mu} \) up to about 1.5, and the constants \( a \) and \( b \) may be determined by least-squares straight-line fitting the plot in this region. Variations in surface temperature are also rather insignificant for \( \sqrt{\mu} \) less than about 1.5. To handle values of \( \sqrt{\mu} \) larger than this, the Thomason model can be run for specific surface conditions. Values of \( a \) and \( b \) computed by the Thomason model agree quite well with those obtained by empirically fitting \( -\ln T_{\lambda w} \) derived from radiometer measurements, against \( \sqrt{\mu} \) using \( m \) determined from radiosonde observations [7].

The procedure used to extract \( T_{\lambda w} \) from the radiometer measurements \( V_\lambda \) is to analyze the ratio of the two channel signals. Using the subscript 1 to denote the channel outside the water vapor band and subscript 2 for the water vapor band, the in-band to out-of-band signal ratio is

\[ \frac{V_2}{V_1} = \frac{V_{02} e^{-m \tau_{02} T_{\lambda w}}}{V_{01} e^{-m \tau_{01}}} \]  \hspace{1cm} (9)

Taking the natural log of this ratio yields

\[ \ln(V_2/V_1) = \ln(V_{02}/V_{01}) + m(\tau_1 - \tau_2) + \ln T_{\lambda w} \]  \hspace{1cm} (10)

and using (7) to model \( T_{\lambda w} \) gives

\[ \ln(V_2/V_1) = \ln(V_{02}/V_{01}) + m(\tau_1 - \tau_2) + a - b \sqrt{\mu} \]  \hspace{1cm} (11)

For channels 1 and 2, respectively, centered at \( \approx 870 \text{ nm} \) and \( \approx 940 \text{ nm} \), \( \tau_1 - \tau_2 \) is quite small (less than \( \approx 0.03 \) for even moderately hazy conditions), and this term may often be neglected to a good approximation. A plot of \( \ln(V_2/V_1) \) versus \( \sqrt{\mu} \) (referred to as a modified Langley plot [2]) as modeled in (11), with the \( m(\tau_1 - \tau_2) \) term neglected or subtracted, yields a straight line with intercept \( = \ln(V_{02}/V_{01}) + a \) and
slope = \sqrt{\mu u}. An example plot is shown in Fig. 2 for a day that yielded a very straight-line fit. The ratio \( V_{02}/V_{01} \) may be estimated quite accurately from such plots even when \( \sqrt{\mu u} > 1.5 \) so long as the value assumed for \( a \) is representative of a straight-line fit to \( \ln T_{\lambda w} \) versus \( \sqrt{\mu u} \) for the same range of \( \sqrt{\mu u} \) as the observations.

Given the intercept as determined by straight-line fits on clear, stable days when \( a \) is nearly constant, \( a \) may then be retrieved for other days by

\[
u = \frac{1}{m} \left( \ln \left( \frac{V_{01}}{V_{02}/V_{01}} \right) + a + m(\tau_1 - \tau_2) \right)^2 \tag{12} \]

using \( a \) and \( b \) determined from the band model. Alternately, \( u \) may be retrieved from the band model derived relation between \( \ln T_{\lambda w} \) and \( \sqrt{\mu u} \), as in Fig. 1, with \( \ln T_{\lambda w} \) extracted from the measurements using the relation

\[-\ln T_{\lambda w} = \ln \left( \frac{V_1}{V_2} \right) + m(\tau_1 - \tau_2). \tag{13} \]

Depending on atmospheric conditions and the desired retrieval accuracy, \( m(\tau_1 - \tau_2) \) in (12) and (13) may be corrected for or neglected. The effect of neglecting \( m(\tau_1 - \tau_2) \) will be discussed further later in the paper.

III. RADIOMETER INSTRUMENTATION

The radiometer has two parallel channels for simultaneous measurements in and out of the water vapor absorption band to eliminate short-term temporal variability effects in the differential comparison of data from the two channels. A schematic cross section of the instrument for the plane cutting the center of the two channels is shown in Fig. 3. The mechanical structure/housing of the radiometer is of sturdy aluminum tubing or plate (~4.5 mm thick) and machined aluminum block. Clear glass flats are mounted over the telescope entrance apertures, and the exterior surfaces are painted white or clear anodized.

The receiver f/4 for each channel is defined by simple two-aperture, nonlens telescopes about 20 cm long with unobscured full-angle f/4's of 2.5°. An interference filter is positioned immediately behind the rear aperture of each channel telescope followed by a silicon photodiode/op-amp combination detector-amplifier (EG&G HUV 1140 BG). The photodiodes have a 3.5-mm active area, which is sufficiently large to provide ample tolerance for centering the detector on the 2.5-mm spot of unmagified, filtered solar irradiance defined at the detector by the telescope-filter combination. Each interference filter (filters are 12.5 mm in diameter) has a bandpass of about 10 nm, and the center wavelength of the out-of-band filter is at 870 nm while the in-band filter is centered at 940 nm.

The detector op-amps are connected in a transimpedance amplifier configuration via an external feedback resistor, selected to be 100 kΩ, connected between pins 9 and 6 with shorts between pin pairs 9 and 10 and 1 and 3 as per manufacturer recommendations. An offset adjustment resistor pot (25 kΩ) is also connected to pins 4, 7, and 8, and DC voltages of +15V and -15V are supplied to pins 2 and 7, respectively, with pin 1 as ground, to bias the op-amps. The offset adjustment and feedback resistors are mounted on a small circuit board behind the detectors, and input bias and output signal wires run from the board to electrical connectors on the back of the radiometer.

The 100-kΩ feedback resistors provide sufficient gain to yield output signals between one and two volts for high sun and clean, dry atmospheric conditions. Decreases in signal by factors of 10 to 15, as can occur for large airmasses and moist conditions, still yield signals large enough to be directly recorded with high precision by a data logger. No attempt was made to stabilize the detector temperatures because the temperature dependence of the detector responsivity is quite small and nearly equal for both channel wavelengths (<~0.05%/°C). By using the ratio of the channel signals, this small temperature dependence is largely canceled out.

The radiometer is weather-sealed so that it can be mounted on an automated solar tracker and left to operate unattended with the channel outputs being sampled/recorded and time stamped, at a specified sampling rate, by a multichannel data logger. The first prototype version of the radiometer was provided under contract to NOAA/ERL in Boulder, CO, in July 1990. Trial measurements were made during September 1990 and March 1991. Example retrievals from these trial measurements are presented in the Results section of this paper.

IV. ERROR EFFECTS

As can be seen in (12), determination of \( u \) requires inputs of \( m, V_1/V_2, V_{02}/V_{01}, a, b \) and \( (\tau_1 - \tau_2) \). The airmass \( m \) can be determined quite accurately (within 0.1 to 1%) by using standard airmass expressions and monitoring the time of observation to within a few seconds [3, 8]. Similarly, \( V_1 \) and \( V_2 \) can be readily measured with very good accuracy (within ~0.3%) using a conventional digital data logger with sufficient bit resolution (e.g., such as the Campbell Scientific CR10 unit used to record the trial measurements in Boulder, CO). Thus, \( m, V_1/V_2 \) are assumed to contribute negligible error in the retrieval of \( u \). Applying standard error analysis to the remaining required inputs in (12), the fractional error in
\[ \frac{\Delta u}{u} \bigg|_{\Delta V_0} = \frac{2}{b \sqrt{mu}} \cdot \frac{\Delta (V_{02}/V_{01})}{(V_{02}/V_{01})} \]  

\text{(14)}

\[ \frac{\Delta u}{u} \bigg|_{\Delta \tau} = \frac{2}{b} \sqrt{\frac{m}{u}} \cdot \Delta (\tau_1 - \tau_2) \]  

\text{(15)}

\[ \frac{\Delta u}{u} \bigg|_{\Delta a} = \frac{-2}{b} \sqrt{\frac{m}{u}} \cdot \Delta a \]  

\text{(16)}

\[ \frac{\Delta u}{u} \bigg|_{\Delta b} = \frac{-2}{b} \sqrt{\frac{m}{u}} \cdot \Delta b. \]  

\text{(17)}

Determination of \( V_{02}/V_{01} \) via the modified Langley plot method discussed in connection with (11) can obviously be affected by a number of uncertainties, but temporal variabilities in \( u \) and \((\tau_1 - \tau_2)\) generally prove to be the limiting factors. Estimates of out-of-band channel intercepts, such as \( V_{01} \), obtained by averaging intercepts from Langley plot fits for several clear, stable days typically yield percent standard deviations in the range of 1 to 2\% \cite{9, 10}. Similarly, voltage ratio intercepts, \( V_{02}/V_{01} \), obtained from fits of modified Langley plots for clear, stable days (like Fig. 2) have yielded ratio averages with percent standard deviations in the range of 2 to 3\% (examples to be given in Results). Thus, a reasonable estimate of the fractional uncertainty within which \( V_{02}/V_{01} \) can be determined is \(<3\%\).

If no correction is made for the term \((\tau_1 - \tau_2)\), the modified Langley plot intercept determination is negatively biased (i.e., inferred \( V_{02}/V_{01} \) is smaller) as the slope of the plot becomes increasingly less negative, by a factor \( \sqrt{m} \) \((\tau_1 - \tau_2)\), with increasing \( \sqrt{m} \). This, in turn, reduces the estimate of \( u \). An example of this is given in Fig. 4, which shows plots of the percent error in retrieved \( u \) versus airmass for various \((\tau_1 - \tau_2)\) and \( u = 2 \) cm.

![Fig. 3. Schematic cross section of the two-channel radiometer.](image)

![Fig. 4. Percent error in retrieved columnar water vapor amount \( u \) versus airmass due to neglecting Delta Tau = \((\tau_1 - \tau_2)\) for various \((\tau_1 - \tau_2)\) and \( u = 2 \) cm.](image)
scaled to the surface pressure for the observation site. Hence, the error in determining \((r_1 - r_2)\) is essentially the same as that of \((\tau_{3a} - \tau_{3b})\). Also, the aerosol optical depth has been found to vary rather closely with wavelength as \(\tau_\lambda \propto \lambda^{3/2}\) (e.g., [9], [11]) where the Angstrom coefficient \(\alpha\) typically ranges between 0.5 and 1.5. For \(\alpha = 1.5\), which produces the greatest difference between \(\tau_{3a}\) and \(\tau_{3b}\), \(\tau_{3a} - \tau_{3b} \approx 0.11\tau_{3a}\) for \(\tau_{3a}\) at 870 nm and \(\tau_{3a}\) at 940 nm. Given that \(\tau_{3a}\) can be readily estimated to within \(\sim 0.03\) (only requires that \(V_{01}\) be known to within \(\sim 3\%\)), the uncertainty within which \((r_1 - r_2)\) can be reasonably estimated is

\[
\Delta(r_1 - r_2) = \Delta(\tau_{3a} - \tau_{3b}) \approx 0.11 \Delta \tau_{3a} < \sim 0.003.
\]

Uncertainty in implementing the model for \(T_{3a}\) to recover \(u\) stems primarily from temperature and lapse rate differences between the modeled and actual atmospheric properties. As seen in Fig. 1, rather large changes of \(\pm 10\%\) in surface temperatures produced only small variations in the \(T_{3a}\) curve for \(\sqrt{\omega u} \sim 1.5\). Similar results are obtained if the temperature lapse rate is varied between 3°C/km and 9°C/km. Straight-line fits to \(-\ln T_{3a}\) versus \(\sqrt{\omega u}\) for these rather extreme ranges in surface temperature and lapse rate yield a spread in the model coefficients \(a\) and \(b\) of \(\Delta a = 0.007\) and \(\Delta b = 0.01\) for \(\sqrt{\omega u} \sim 1.5\). The differences in \(a\) and \(b\) for different conditions are completely correlated, which is to be expected as they are analytically related \((-\ln T_{3a} = -a + b\sqrt{\omega u})\). As can be seen in Fig. 1, a higher surface temperature yields a lesser slope (smaller \(b\)), but the intercept is correspondingly less negative (smaller \(a\)). This indicates that the uncertainty due to \(a\) or \(b\) considered separately, as given in (16) and (17), should be added algebraically to obtain the net effect on \(u\). The errors in determining \(u\) due to uncertainties in \(V_{02}/V_{01}\) and \((r_1 - r_2)\), as given in (14) and (15), are reasonably treated as independent errors, including being independent of the correlated \(a\) and \(b\) uncertainty given by summing (16) and (17). Hence, the total retrieval error for \(u\) is estimated by taking the square root of the sum of the squares of these three contributions:

\[
\frac{\Delta u}{u} = \left[ \left( \frac{\Delta u}{u} \right)_{V_{02}/V_{01}}^2 + \left( \frac{\Delta u}{u} \right)_{\Delta r}^2 + \left( \frac{\Delta u}{u} \right)_{\Delta a}^2 + \left( \frac{\Delta u}{u} \right)_{\Delta b}^2 \right]^{1/2}.
\]

The resulting total errors in \(u\) computed from (20) for the uncertainties in \(V_{01}\), \((r_1 - r_2)\), \(a\) and \(b\) estimated in the foregoing discussion are given in Table I for cases where \(u = 0.2\) cm and \(2\) cm and \(m = 1.5\) and 5. Retrieval error is dominated by the uncertainty in the intercept \(V_{02}/V_{01}\) even for the larger values of \(m\) and \(u\). For the case \(m = 5\) and \(u = 2\) cm, \(\sqrt{\omega u} \approx 3.16 \text{ cm}^{1/2}\) is sufficiently large that the straight-line fit to \(-\ln T_{3a}\) using coefficients \(a\) and \(b\) determined for \(\sqrt{\omega u} \sim 1.5\) is no longer a good approximation. The actual model curve, as in Fig. 1, should be used to infer \(u\) from \(-\ln T_{3a}\) in such cases. However, for the purpose of the error analysis presented here, the straight-line approximation yields a reasonable estimate of the spread in \(\ln T_{3a}\) (i.e., estimates for \(\Delta a\) and \(\Delta b\)) for different surface temperatures and lapse rates.

<table>
<thead>
<tr>
<th>(m = 1.5)</th>
<th>(m = 5)</th>
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<tr>
<td>(u = 0.2) cm</td>
<td>(u = 2) cm</td>
</tr>
<tr>
<td>(\Delta u/u \times 100%)</td>
<td>(\Delta u/u \times 100%)</td>
</tr>
<tr>
<td>13.5%</td>
<td>4.5%</td>
</tr>
<tr>
<td>(\Delta a_u)</td>
<td>0.027 cm</td>
</tr>
</tbody>
</table>

\(\Delta a_u\) is computed from (20) for \(a = 0.076\), \(b = 0.82\), \(\Delta a_{V_{02}/V_{01}} = 0.03\), \(\Delta (r_1 - r_2) = 0.003\), \(\Delta a = 0.076\), and \(\Delta b = 0.01\).

V. RESULTS

Trial measurements were made with the two-channel radiometer in Boulder, CO, during September 1990 and at a nearby Erie, CO, site during March 1991. Ten days were sufficiently clear near sunrise or sunset during September, 1990 to permit Langley/modified Langley plot fits. Inspection of Langley plots for Channel 1, the aerosol channel at 870 nm, revealed significant temporal fluctuations and biases for four days, leaving 6 days from which to determine the intercept \(V_{01}\). Similar screening of the March 1991 data yielded five apparently suitable days for Langley plot calibrations.

Table II lists the \(V_{01}\) values for the 6 September days along with the resulting mean and standard deviation (impressively small at 0.9%). Also listed are the modified Langley plot derived intercept voltage ratios, \(V_{02}/V_{01}\), both with and without correction for \((r_1 - r_2)\) which makes a difference of about 2%. Due apparently to temporal water vapor variations, the intercept voltage ratio for 1 day (denoted by *) was sufficiently removed from the mean (~2 Std. Dev.) to be reasonably rejected. This yields means for the five remaining days nearly the same as for the 6 days, but the percent standard deviation is reduced by about a factor of two. Table III lists the intercepts for the 5 days of March 1991 data and the results are very similar (with even slightly lower standard deviations) to the September 1990 data. The fact that the mean intercepts for the two observation periods, being separated in time by some 6 months, agree with one another within their small standard deviations indicates that the calibration stability of the instrument is quite good.

Columnar water vapor retrievals obtained via (13), with \(T_{3a}\) as modeled in Fig. 1 for a 289.7 K surface temperature, and the mean intercept voltage and voltage ratios from Table II are shown in Figs. 5 and 6 for measurements made on September 7 and 12, 1990. As is to be expected from the fairly straight modified Langley plot for September 12 (Fig. 2), \(u\) in Fig. 6 is rather constant with time. In contrast, \(u\) in Fig. 5 for September 7 (a day rejected for Langley plots due to temporal variability) is quite temporally variable; significant variations in \(u\) can be seen for various time periods ranging from a few minutes to hours. Examination of fine scale variations indicates that changes in \(u\) can be resolved down to about 0.1 mm. A re-
### Table II

<table>
<thead>
<tr>
<th>DATE</th>
<th>Langley Plot Derived Intercept $V_{01}$ (mV)</th>
<th>Modified Langley Plot Derived Intercept Voltage Ratio $V_{02}/V_{01}$</th>
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</thead>
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<td>without $\tau_1 - \tau_2$ correction</td>
<td>with $\tau_1 - \tau_2$ correction</td>
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<tr>
<td>9/04/90</td>
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#### 6-day

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<th>$\sigma$ (% Std. Dev.)</th>
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<td>1784.3±15.4</td>
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<tr>
<td></td>
<td>1.3377±0.096</td>
<td>±(7.2%)</td>
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<tr>
<td></td>
<td>1.3616±0.099</td>
<td>±(7.3%)</td>
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#### 5-Day

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<td>1.2985±0.042</td>
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<tr>
<td></td>
<td>1.3211±0.045</td>
<td>±(3.4%)</td>
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*Rejected day for 5-day average.

### Table III

<table>
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<tr>
<th>DATE</th>
<th>Langley Plot Derived Intercept $V_{01}$ (mV)</th>
<th>Modified Langley Plot Derived Intercept Voltage Ratio $V_{02}/V_{01}$</th>
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<td>with $\tau_1 - \tau_2$ correction</td>
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<tr>
<th></th>
<th>Avg $\pm$ Std. Dev.</th>
<th>$\sigma$ (% Std. Dev.)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1775.4±11.6</td>
<td>±(0.7%)</td>
</tr>
<tr>
<td></td>
<td>1.2550±0.022</td>
<td>±(1.8%)</td>
</tr>
<tr>
<td></td>
<td>1.2839±0.026</td>
<td>±(2.0%)</td>
</tr>
</tbody>
</table>

The effect of correcting for $(\tau_1 - \tau_2)$ removes negative biases of $\sim$4 to 10% in the retrievals shown in Figs. 5–7. This demonstrates the benefit of making such a correction. Based on (20), $\Delta u$ and $\Delta b$ as given in Table I and the inferred intercept errors in Tables II and III, the uncertainty in the corrected curves in Figs. 5–7 is estimated to be $\sim$6% or less with the $V_{02}/V_{01}$ intercept error being the dominant error. To demonstrate that the corrected retrievals agree well with other means for measuring $u$, retrievals are included for a few times from available NOAA/WPL microwave radiometer measurements at Denver, CO, for Fig. 6 and at Erie, CO, in Fig. 7 (courtesy of J. Snider). The morning and afternoon Denver radiosonde observations are also included at the edges of the time axis in Fig. 7. Excluding the am denver radiosonde observation, which could be expected to differ significantly from the later Boulder measurements as the mixing layer had yet to develop, all the other comparison points fall within $\sim$5% of the aerosol corrected two-channel solar radiometer retrievals. A more definitive intercomparison of water vapor retrievals obtained with microwave radiometers and solar radiometers is the subject of a future paper now in preparation. The few comparisons presented here suffice to show...
that the solar radiometer technique can provide retrievals in good agreement (within 10%) with other techniques. This is consistent with retrieval comparisons that have been made to radiosonde data for many days of observations taken in Tucson, AZ, with a similar solar radiometer [7].

VI. CONCLUSION

The simple two-channel instrument and retrieval technique offers a self-contained method for sensing columnar water vapor, \( u \). Error analyses and experimental results presented here indicate that the instrument/technique can be reasonably employed to retrieve \( u \) within an error of \( \pm 5 \) to 10%. The dominant error is due to uncertainty in the intercept voltage ratio, \( V_{02}/V_{01} \). By applying modified Langley plots to selected clear, stable days, including correcting for the optical depth difference (\( \tau_1 - \tau_2 \)), it should be possible to estimate \( V_{02}/V_{01} \) with an uncertainty within 2% as was the case for the March 1991 data presented here. This permits retrievals in \( u \) with an error of \( \pm 5\% \) or less for \( \sqrt{\text{N}_u} > 1 \). However, uncertainty in the water vapor transmission model used to relate \( T_{X_u} \) to \( u \) probably limits retrieving \( u \) with an error less than a few percent.

Currently, the principal means for obtaining information on the temporal/spatial variability of columnar water vapor over land is from radiosondes which, with the standard two launches per day, give only 12-h temporal resolution and are spatially limited to sites \( \pm 200 \) km apart. Our results indicate that the simple instrument and technique described here can be used to retrieve columnar water vapor with an accuracy comparable to radiosondes, on a continuous time basis (so long as the sun is up and unobscured by clouds). Furthermore, the relative low cost of the instrument makes it feasible for use in large geographical networks which, when combined with radiosonde and surface measurements and satellite imagery, could greatly enhance information about how water vapor fields fluctuate in space and time. Such information can be used to improve the initial data input to numerical weather prediction and climate models.

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REFERENCES


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