Dry Bias in Vaisala RS90 Radiosonde Humidity Profiles over Antarctica

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ABSTRACT

Middle to upper tropospheric humidity plays a large role in determining terrestrial outgoing longwave radiation. Much work has gone into improving the accuracy of humidity measurements made by radiosondes. Some radiosonde humidity sensors experience a dry bias caused by solar heating. During the austral summers of 2002/03 and 2003/04 at Dome C, Antarctica, Vaisala RS90 radiosondes were launched in clear skies at solar zenith angles (SZAs) near 83° and 62°. As part of this field experiment, the Polar Atmospheric Emitted Radiance Interferometer (PAERI) measured downwelling spectral infrared radiance. The radiosonde humidity profiles are used in the simulation of the downwelling radiances. The radiosonde dry bias is then determined by scaling the humidity profile with a height-independent factor to obtain the best agreement between the measured and simulated radiances in microwindows between strong water vapor lines from 530 to 560 cm\(^{-1}\) and near line centers from 1100 to 1300 cm\(^{-1}\). The dry biases, as relative errors in relative humidity, are 8\% (microwindows; 1\sigma) and 9\% (line centers) for SZAs near 83°; they are 20\% ± 6\% and 24\% ± 5\% for SZAs near 62°. Assuming solar heating is minimal at SZAs near 83°, the authors remove errors that are unrelated to solar heating and find the solar-radiation dry bias of 9 RS90 radiosondes at SZAs near 62° to be 12\% ± 6\% (microwindows) and 15\% ± 5\% (line centers). Systematic errors in the correction are estimated to be 3\% and 2\% for microwindows and line centers, respectively. These corrections apply to atmospheric pressures between 650 and 200 mb.

1. Introduction

Terrestrial outgoing longwave radiation is very sensitive to atmospheric humidity because water vapor is the most effective greenhouse gas. The outgoing longwave flux is particularly sensitive to middle and upper tropospheric humidity (M/UTH, 650–250 mb; see, e.g., Sinha and Harries 1995; Ferrare et al. 2004); accordingly, measurements of M/UTH globally are important for understanding the radiation budget. Furthermore, accurate measurements of upper tropospheric humidity are needed to determine the magnitude of water vapor feedback as atmospheric carbon dioxide increases and thus to predict future climate change (Held and Soden 2000).

Currently, atmospheric humidity is predominantly measured using radiosondes (see Turner et al. 2003 and references therein). Radiosonde humidity sensors have fine vertical resolution but can experience errors due to inaccuracies in calibration, time lag, contamination, and solar heating of the humidity sensor (Wang et al. 2002; Miloshevich et al. 2001, 2004, 2006; Turner et al. 2003). Much work has been done to identify and minimize sources of error in different generations of radiosondes.
between strong absorption lines from 530 to 560 cm⁻¹. Two distinct sets of infrared frequencies were used, one dominated by uncertainties in the water vapor continuum (between strong lines from 530 to 560 cm⁻¹) and water vapor lineshape parameters (near strong lines from 1100 to 1300 cm⁻¹) and by uncertainties in the temperature profiles. Uncertainties in radiative transfer and atmospheric trace gases are also considered.

We determine the total dry biases for 19 Vaisala RS90 radiosondes at SZAs near 83° and 62°. Because the solar dry bias should be small for SZAs near 83°, we use the mean dry bias at 83° as an upper limit to error from sources unrelated to solar heating (such as radiosonde calibration error) and determine the solar dry bias for 9 RS90 radiosondes at SZAs near 62°. These apply to atmospheric pressures of 650 to 200 mb, or an effective atmospheric pressure of 570 mb. We compare our results to the RS90 and RS92 results of Cady-Pereira et al. (2008) and the height-dependent RS92 correction of Vömel et al. (2007) for SZAs between 10° and 30°.

2. Instruments

a. Vaisala RS90 radiosondes

Vaisala RS90 radiosondes were used to measure atmospheric temperature, pressure, and humidity. Turner et al. (2003) found that the mean error in water vapor mixing ratio for an RS80-H radiosonde batch varied from −2% to 24% for different batches. [They define “batch” by the date of calibration (Miloshevich et al. 2004, appendix A)]. Turner et al. attribute this error in part to variations in the calibration procedure. RS90 and RS92 radiosondes are now calibrated in a new calibration facility with a standardized procedure, which
should reduce the batch-to-batch calibration variability (Paukkunen et al. 2001). Of the RS90 radiosondes for which we determine the dry bias, all were from the same calibration batch (23 October 2003) except for one (24 October 2003).

Storing and launching the radiosondes at near-ambient temperatures minimized errors due to thermal shock. [See the suggestions of Hudson et al. (2004)]. The launch procedure typically included allowing the radiosonde to equilibrate at the surface for 15–30 min before launch. Because care was not taken to keep the radiosonde in the shade, the sensor arm may have warmed while the sensor was still on the surface, although this was probably offset somewhat by ventilation by the wind at the surface. The method of Miloshevich et al. (2004) was used to correct for the time lag in the response of the humidity sensors.

Radiosonde measurements were made during the afternoon (around 1600 LT; LT = Local Time = UTC + 8), when SZAs were on average 62° (ranging from 59° to 66°), and during the late evening (around 2300 LT), when SZAs were on average 83° (ranging from 81° to 85°). Figure 1 shows the profiles of temperature and relative humidity (RH) with respect to water measured on 13 January 2004 during the afternoon and late evening. In Fig. 1a, we see that the temperature profiles are almost identical except near the surface (within the first 50 m). By contrast, radiosonde humidities differ markedly (Fig. 1b). The relative humidity at which ice saturation occurs for the afternoon temperature profile is also shown. The ice saturation for the evening temperature profile is almost identical except near the surface, where the ice saturation relative humidity is 72% instead of 80%. These radiosoundings are fairly typical of those used in this study and will serve as afternoon and evening case studies throughout the rest of this paper. Although the humidity is supersaturated with respect to ice at some heights, visual observations and inspection of concurrent radiances indicate that the sky was clear. Supersaturation with respect to ice is common in the Antarctic atmosphere (Gettelman et al. 2006). Jensen et al. (2001) have reported areas of ice supersaturation near the tropopause in the absence of ice crystals. Regions of supersaturation with respect to ice have also been observed by a water vapor Raman lidar in cloud-free regions in the middle and upper troposphere at a midlatitude site (Comstock et al. 2004).

b. Polar Atmospheric Emitted Radiance Interferometer

The PAERI is an AERI modified for use in polar regions. AERI instruments, developed by the Space Science and Engineering Center at the University of Wisconsin in Madison, are ground-based Fourier transform infrared spectrometers that remotely sense emission from the atmosphere. The PAERI measured downwelling infrared radiance in the spectral region between 500 and 3000 cm⁻¹ (3–20 μm) at 1 cm⁻¹ resolution for a zenith view. AERI instruments are described in detail by Knuteson et al. (2004a,b), who include a careful assessment of instrument accuracy that we use for estimating our sources of uncertainty. Each spectrum was generated by taking the Fourier transform of an interferogram (Beer 1992) measured by the PAERI; each interferogram is comprised of between 20 and 90 consecutive measurements (called coadditions) taken by the interferometer to reduce instrument noise. Spectra of hot and ambient temperature blackbodies of known temperature and emissivity are used to calibrate measurements of the sky, as described by Revercomb et al. (1988) and Knuteson et al. (2004b). PAERI measurements that were made concurrently with radiosoundings under clear skies were selected for this study. Between 4 and 13 spectra were averaged during the radiosonde flight to further reduce instrument noise.

Figure 2 shows a typical PAERI spectrum. The region of low radiance between 800 and 1300 cm⁻¹ (excluding ozone at 1050 cm⁻¹) is termed the “atmospheric window.” Because the Dome C atmosphere is so cold and dry, radiances in the atmospheric window are very close to zero. The center of the weak CO₂ band

![Fig. 1. Profiles of (a) temperature and (b) relative humidity with respect to water (RH) measured by Vaisala RS90 radiosondes at Dome C, Antarctica, on 13 Jan 2004 during the afternoon (1525 LT) and evening (2336 LT). The RH at which ice saturation occurs for the afternoon temperature profile is also shown.](image)
centered at 961 cm\(^{-1}\) is the clearest portion of the spectrum, where the simulated radiances is only 0.1 RU. For reference, the Planck function for the temperature near the PAERI is shown in Fig. 2. The two spectral regions used to scale the relative humidity are circled. From 530 to 630 cm\(^{-1}\) wavenumbers in microwindows between strong lines are used; from 1100 to 1300 cm\(^{-1}\) wavenumbers at strong (but unsaturated) water vapor line centers are used. The dotted line indicates the Planck function for the temperature near the PAERI.

Figure 3a shows the total systematic uncertainty (one standard deviation) in an afternoon PAERI measurement. The major sources of systematic uncertainty are uncertainties in the calibration, the nonlinear response of the detector to signal (detector nonlinearity; DN), the effective laser wavenumber (ELW), and the correction for instrument lineshape effects (Knuteson et al. 2004b). The contributions of the first three to systematic uncertainty in the PAERI are shown in Fig. 3b. Uncertainty in the calibration is determined by propagating uncertainties in the blackbody emissivities and temperatures through the calibration equation (Revercomb et al. 1988). The uncertainties used are those given in Knuteson et al. (2004b), except for the emissivity of the hot blackbody, which is taken to be 8.7 × 10\(^{-5}\) (one standard deviation). The uncertainty due to the detector nonlinearity is not well known, but is assumed to be less than 0.2% (three standard deviations; one standard deviation shown) based on Knuteson et al. The effective laser wavenumber for 20 PAERI measurements had a standard deviation of 0.3 ppm; the more conservative value of 1 ppm is used for the uncertainty. This error causes error in the wavenumber scale of the measured radiances. Propagating this to an uncertainty in the measured radiance gives rise to derivative-shaped peaks near line centers. Uncertainties in the correction for the instrument lineshape effects are negligible and not shown. The rms uncertainty due to systematic errors (one standard deviation) is 0.15 ± 0.05 RU for microwindows and 0.04 ± 0.02 RU for line centers.

Instrument noise varies randomly with both frequency and measurement. The magnitude depends on the number of coadditions made and the number of spectra averaged. The noise equivalent spectral radiance (NESR) is roughly estimated to be 0.1 RU near 550 cm\(^{-1}\) (microwindows) and 0.02 RU near 1200 cm\(^{-1}\) (line centers). Using 25 microwindow frequencies and 36 line center frequencies to scale the relative humidity
further reduces the NESRs to about 0.02 RU and 0.003 RU, respectively.

3. Scaling the water vapor profile

a. Simulating radiances using radiosonde profiles

The Line-by-Line Radiative Transfer Model (LBL-RTM; Clough et al. 2005), version 10.3, was used to simulate downwelling radiances. Lineshape parameters were taken from the HITRAN 2004 database (Rothman et al. 2005). The database is a compilation of measured and calculated parameters from a large number of researchers. Of particular interest to this study are the water vapor lineshape parameters (line positions, strength, and air and self-broadened half-widths) between 500 and 1800 cm\(^{-1}\), which come from R. A. Toth (http://mark4sun.jpl.nasa.gov/data/spec/H2O). The MT-CKD 1.3 water vapor continuum is used (Tobin et al. 1999; Turner et al. 2004; Clough et al. 2005). Atmospheric profiles having approximately 200 layers were created for LBLRTM, using the temperature, pressure, and humidity profiles from the radiosonde data. The humidity was set to 5 ppmv from about 7.7 to 24 km above the surface (i.e., from 11 to 27 km above sea level, because Dome C is at an altitude of 3.3 km), and to 4 ppmv from 24 to 60 km, in keeping with the suggestion of Walden et al. (1998). The temperature profile above the sounding was determined from the South Polar model atmosphere for January also developed by Walden et al. (1998). Profiles of N\(_2\)O, CH\(_4\), and CO\(_2\) were derived from surface measurements made at South Pole Station during our field seasons [National Oceanic and Atmospheric Administration–Earth System Research Laboratory (NOAA–ESRL); http://www.esrl.noaa.gov/gmd/; Dlugokencky et al. 2001; Conway et al. 1994]. Profiles of chlorofluorocarbons (CFCs) were generated from surface measurements and ozone profiles from ozone soundings made at South Pole Station during previous austral summers (Rowe et al. 2006 and references therein). For other trace gases, the subarctic winter standard atmosphere of McClatchey et al. (1972) was used.

The mean radiance residual (measured – simulated radiance) for afternoon cases is shown in Fig. 4a. The systematic uncertainty in the PAERI is also shown. The residual is typically within the PAERI uncertainty in the atmospheric window but not at frequencies where water vapor emission is significant. Figures 4b and 4c show expanded views of the two spectral regions used to scale water vapor; typical frequencies used for scaling are marked with dots and circles. Close to the centers of emission lines in Fig. 4c, the measured radiances are greater than the simulated radiances. In general, the simulated radiances increase when the RH at all heights is increased.

b. Temperature correction

To evaluate and improve the accuracy of the radiosonde temperature profiles, the measured and simulated radiances were compared in the CO\(_2\) band from 670 to 760 cm\(^{-1}\). The difference between measured and simulated radiance \(g\) can be written in terms of a kernel \(K\) and a temperature correction \(f\) as

\[
g_i = \int_{P_0}^{0} K(P)f(P)\,dP,\tag{1}
\]

where \(P\) is the atmospheric pressure and \(i\) is the frequency index. The kernel is given by Rowe (2004). Constrained linear inversion (CLI) is used to solve for the temperature correction (Twomey 1977; Rowe 2004; Rowe et al. 2006). Principal component analysis suggests that, to within experimental uncertainty, there are four unique kernels (the singular values that characterize the set of kernels were compared to the experimental uncertainty), and thus four basis functions are warranted. For most of the profiles, these basis functions are step functions corresponding to atmospheric layers of 0–10, 10–180, 180–1500, and 1500–4000 m in a verti-
cal distance above the PAERI. For a few profiles, better agreement between simulation and measurement was obtained when basis functions were determined using singular value decomposition of the kernels rather than the step functions. When the basis set was formed from the kernels, the correction is not constant within a layer, so the mean correction within the layer is used in determining the statistics given below. The mixing ratio of water vapor was held constant, so that changing the temperature affected the RH.

The temperature corrections were typically small and unbiased, except within strong near-surface temperature inversions. For the afternoon profiles, the temperature corrections within the first 10 m were 0.1 ± 0.4 K (one standard deviation). For the evening profiles, when a steep surface temperature inversion was typically present, the corrections were 0.5 ± 0.7 K. When the temperature changes quickly with altitude, a time lag in the temperature sensor can have a large effect (Miloshevich et al. 2004). Temperature corrections within inversions could be explained by a shift in temperature profile of roughly 0 to 12 m, which would correspond to a time lag of roughly 0 to 4 s, with different adjustments needed for different profiles. [Note that Paatukpen et al. (2001) state that RS90 temperature sensors have a time lag of only 0.2 s (at 1000 mb, for an ascent rate of 6 m s⁻¹).] For some datasets, a shift in the temperature profile provides a correction as good as that provided by the CLI. Temperature corrections from 10 to 1500 m were 0.0 ± 0.2 K (one standard deviation) with no diurnal bias.

Within the first 1500 m, uncertainty in the temperature correction can be estimated from uncertainties in the measured and simulated radiances from 670 to 700 cm⁻¹. In this spectral region, the temperature correction reduces the radiance residual to 0.0 ± 0.2 RU; the residual varies rapidly with frequency and probably results from noise in the PAERI measurement and errors in the fine temperature structure. If the measured or simulated radiances are biased because of systematic errors, the bias will result in a bias in the temperature correction. Near the band center, uncertainty due to systematic error in the PAERI is ~0.1 RU and errors in lineshape parameters have little effect on simulated radiances. A bias in radiance of 0.1 RU would correspond to a bias in temperature of approximately 0.1 K for the first 1500 m of atmosphere.

Temperature corrections between 1500 and 4000 m were typically <0.2 K for both afternoon and evening datasets. This temperature correction depends on radiances in the wing of the CO₂ band, where uncertainties in the CO₂ absorption coefficient are important. The radiance residuals from 700 to 780 cm⁻¹ are as large as 0.3 RU, have a spectral signature that cannot be removed by correcting the temperature profile, and are similar for all PAERI measurements (see Fig. 4a, but note that residuals greater than one in this region are due to water vapor error, not temperature error). It is therefore assumed that the temperature corrections are not significant to within the uncertainty.

c. Identifying clear-sky spectra

To verify the absence of hydrometeors and thereby ensure clear-sky conditions, we initially rely on visual observations made during the radiosonde launches. The datasets selected based on the visual observations are then subjected to further testing. The measured downwelling radiances are compared with clear-sky simulations in the atmospheric window from 800 to 1220 cm⁻¹. This spectral region is fairly transparent to trace gases; there is minor emission from CFCs, nitric acid, ozone, water vapor, N₂O, and carbon dioxide. Simulations indicate that the clearest region of the spectrum is 961 cm⁻¹, where the simulated downwelling radiance is 0.1–0.2 RU for the atmospheric profiles over Dome C. Of the RS90 measurements chosen based on the visual sky observations, all but one had measured radiances of 0.2 RU at 961 cm⁻¹. For these cases, the mean of the radiance residual between strong lines from 1120 to 1220 cm⁻¹ is within the experimental uncertainty (~0.06 RU for one standard deviation). One dataset was rejected as potentially having subvisible clouds. At 1518 LT 21 December 2003, the PAERI radiance was 0.8 RU at 961 cm⁻¹ and the mean radiance difference from 1120 to 1220 cm⁻¹ was 0.2 RU; visual observations indicated a cirrus cloud on the horizon but none in the view of the PAERI (zenith).

d. Scaling the water vapor profiles

A height-dependent water vapor retrieval was attempted using the measured and simulated radiance spectra at frequencies near strong lines of water vapor between 1100 and 1300 cm⁻¹ and between 1800 and 2000 cm⁻¹. The radiance residual is again expressed according to Eq. (1), where f is now a correction to the water vapor amount and the kernel is again given by Rowe (2004). Principal component analysis suggests that, to within experimental uncertainty, there is only one unique kernel for water vapor. Thus, the altitude dependence of the water vapor uncertainty cannot be retrieved using PAERI data in this dry environment.

Therefore, the relative humidity is scaled by a single factor at all heights. In a related study, Turner et al. (2003) found in dual RS80-H radiosonde launches that the ratio of relative humidities is fairly constant with
height, and they scaled the mixing ratio of water vapor to bring the PWV from the radiosonde into agreement with the PWV from a microwave radiometer. In this study, the radiosonde RH, used to simulate the downwelling radiance, is scaled until the rms difference between the measured and simulated downwelling radiance is minimized. The rms difference is

$$E(y) = \sqrt{\frac{1}{N} \sum_i [M_i - S_i(y)]^2},$$

(2)

where $y$ is the scale factor, $M$ is the measured radiance, $S$ is the simulated radiance, $i$ is the index to frequency, and $N$ is the number of frequencies. Letting “min” represent the minimization operation and defining “inv” such that $\text{inv}[E(y)] = y(E)$, we write

$$y^* = \text{inv}\left\{\min \left\{ \sqrt{\frac{1}{N} \sum_i [M_i - S_i(y)]^2} \right\} \right\}. \quad (3)$$

The rms difference is calculated for about 25 frequencies in the microwindows between strong lines from 530 to 560 cm\(^{-1}\) and for about 40 frequencies near strong line centers of water vapor between 1100 and 1300 cm\(^{-1}\). [The frequencies used on 1524 LT (0724 UT) 13 January 2004 are listed in Table 1; similar frequencies are used for all other cases.] The two spectral regions are complementary in that the water vapor continuum is negligible at the line centers used for the atmospheric conditions experienced at Dome C (Rowe et al. 2006), but uncertainties in water vapor lineshape parameters are very important; the reverse is true in the microwindows. The rms differences for all datasets are plotted as a function of the relative change in RH (%) in Fig. 5. The retrieved water vapor scale factors $y^*$ are indicated with stars. Values of $y^*$ retrieved using microwindows correspond to rms radiance differences of 0.3 to 0.9 RU, and values retrieved using line centers correspond to rms radiance differences of 0.1 to 0.2 RU.

Figure 6a shows the mean radiance residual for all afternoon cases after scaling the RH. Figures 6b and 6c are expanded views of the two spectral regions used for determining the scale factor. Scaling the water vapor decreases the rms difference from 0.8 to 0.5 RU in microwindows and from 0.4 to 0.1 RU at line centers. The residual above 1300 cm\(^{-1}\) could be caused by uncertainties in the foreign-broadened water vapor continuum. For evening cases the decrease was less dramatic but resulted in similar reductions in rms differences.

4. Analysis of systematic errors

Calculation of the water vapor scale factor depends on the comparison of measured spectral radiances to radiances simulated using the radiosonde profiles, as shown in Eq. (3). Thus, it is important to accurately

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**Table 1.** Frequencies used to scale the water vapor profile for 1524 LT 13 Jan 2004. Frequencies used for other datasets are similar.

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characterize systematic errors in both the PAERI measurements and the LBLRTM simulations. Uncertainties in the PAERI measurements have been discussed above and so are only mentioned briefly in this section. Uncertainties in LBLRTM simulations include those resulting from spectroscopic errors, uncertainties in atmospheric variables, and radiative transfer errors. Eq. (3) can be rewritten to show the dependence of the measurement and simulation on parameters relating to these sources of error; thus,

\[ y^* = \text{inv} \left\{ \min \left\{ \frac{1}{N} \sum_i \left[ M_i - S(p_{\text{WLS}}, p_C, p_T, p_{\text{CH4}}, p_{\text{N2O}}, p_{\text{RT}}, y) \right]^2 \right\} \right\}; \]

where the parameters \( p \) that the simulation depends on are, respectively, water vapor lineshape parameters (\( p_{\text{WLS}} \)), the water vapor continuum (\( p_C \)), temperature (\( p_T \)), the atmospheric concentrations of CH\(_4\) and N\(_2\)O (\( p_{\text{CH4}} \) and \( p_{\text{N2O}} \)), and the widths of layers used in performing the radiative transfer (\( p_{\text{RT}} \)). Because \( y^* \) is a function of many variables, we refer to the most appropriate value of \( y^* \) for a given dataset as \( y' \), to avoid confusion.

The equation for propagation of errors is therefore

\[
\langle \Delta y' \rangle^2 = \left( \frac{\partial y^*}{\partial M} \right)^2 \sigma_M^2 + \left( \frac{\partial y^*}{\partial p_{\text{WLS}}} \right)^2 \sigma_{\text{WLS}}^2 + \left( \frac{\partial y^*}{\partial p_C} \right)^2 \sigma_C^2 + \cdots. \tag{4}
\]

We calculate the contribution of the systematic uncertainty in the PAERI measurement by making the approximation

\[ \left( \frac{\partial y^*}{\partial M} \right)^2 \sigma_M^2 \]

where, for instance, \( y^*(M) \) indicates that \( y^* \) is a function of \( M \). The quantity \( \sigma_M \) is the total systematic error in the PAERI spectrum. For convenience, \( \Delta M \) is chosen to be equal to \( \sigma_M \). To get \( y^*(M + \Delta M) \), the measurement is perturbed by \( \Delta M \) and the scale factor \( y^* \) is recalculated.

The contribution of a source of error in the simulated radiance to uncertainty in the water vapor scale factor is calculated similarly; thus,

\[ \left( \frac{\partial y^*}{\partial p_x} \right)^2 \sigma_x^2 = \{ y^*[S(p_x + \sigma_x)] - y^*[S(p_x)] \}^2. \tag{5} \]

where \( p_x \) is a parameter the simulated radiance depends on, such as temperature, and \( \sigma_x \) is the uncertainty in the parameter.

Tables 2 and 3 summarize uncertainties for micro-windows and line centers. The first column in each table lists the sources of error. If the source of error is an input parameter to LBLRTM, the second column lists the uncertainty in the parameter. The third column lists the resulting uncertainties in measured or simulated radiances in RU; because the uncertainties are frequency-dependent, the rms value for each set of frequencies is given. The final column lists the uncertainties in the water vapor scale factor in units of relative change in RH (%). The standard deviation of the uncertainties in the scale factor is also given where possible. The last row of each table gives the combined uncertainty, calculated as the sum of squared uncertainties. Uncertainties in the simulated radiance are discussed below.

The temperature correction reduces uncertainties in the temperature profile to 0.13 K below 1.5 km. Above 1.5 km we use the uncertainty (1σ) given by Paukkunen et al. (2001) for the Vaisala RS90 radiosonde temperatures, 0.25 K.

Emission from the trace gases CH\(_4\) and N\(_2\)O only affects the 1100 to 1300 cm\(^{-1}\) region. Uncertainties in these trace gases (5%) result in uncertainties in the scale factor of 0.2%.
Radiative transfer error is mainly due to the finite size of the atmospheric layers used. Care was taken to minimize this source of error by using nearly 200 layers for each simulation. The pressure change across each layer was 3.5 mb, corresponding to layers of 40 to 100 m throughout the troposphere. Sensitivity studies were performed by comparing simulations at this vertical resolution to simulations at the vertical resolution of the radiosonde; as shown in the table, radiative transfer error is a minor contributor to the total error.

Uncertainties in line shape parameters need to be well characterized for results using frequencies near line centers. Fortunately, a lot of work has gone into calculating and measuring water vapor line shapes from 1100 to 1300 cm$^{-1}$. Uncertainty ranges for water vapor line strengths, foreign-broadened line widths, self-broadened line widths, and line positions based on the work of Toth et al. are given by the HITRAN database (Rothman et al. 2005). To test the importance of these errors, line shape parameters were perturbed by the uncertainties listed, a modified line shape parameter database was constructed, and the simulated spectra were recalculated. It was determined that uncertainties in air-broadened line widths (<1%–20%), self-broadened line widths (<1%–10%), and line positions (0.001–0.01 cm$^{-1}$) are dominated by uncertainties in line strengths (5%–10%). To propagate errors in line strengths, the uncertainty in each line strength ($\sigma_{\text{WLS}}$) was multiplied by a random number $r$ taken from a normal distribution with a standard deviation of one. Using $\sigma_{\text{WLS}}$ as the uncertainty in line strength, Eq. (5) was used to solve for the resulting uncertainty in the water vapor scale factor for several sets of random numbers; the mean values for each spectral region are reported in Tables 2 and 3.

Uncertainties in the foreign-broadened water vapor continuum are important for microwindows. The uncertainty in the foreign-broadened water vapor continuum (MT-CKD 1.3) is estimated to be 20% (Tobin et al. 1999; Clough et al. 2005).

The total uncertainties in the water vapor scale factor, resulting from sources of error that produce bias, are 9% and 4% (relative error in relative humidity) for microwindows and line centers, respectively.

### 5. Effective atmospheric pressure

Emission at the frequencies used to scale the relative humidity is influenced by water vapor throughout the troposphere. The sensitivity of the downwelling radiance to changes in water vapor in atmospheric layers was investigated to determine the effective height or

<table>
<thead>
<tr>
<th>Source of error</th>
<th>Uncertainty</th>
<th>Radiance error$^a$ (RU)</th>
<th>Uncertainty in scale factor (relative error in RH, %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAERI uncertainty$^b$</td>
<td>—</td>
<td>0.15</td>
<td>4 ± 1</td>
</tr>
<tr>
<td>WV continuum$^c$</td>
<td>20%</td>
<td>0.2</td>
<td>8</td>
</tr>
<tr>
<td>Temperature</td>
<td>0.13–0.25 K</td>
<td>0.009</td>
<td>1 ± 2</td>
</tr>
<tr>
<td>Radiative transfer</td>
<td>—</td>
<td>0.002</td>
<td>&lt;0.1</td>
</tr>
<tr>
<td>Total</td>
<td>—</td>
<td>0.25</td>
<td>8.7</td>
</tr>
</tbody>
</table>

$^a$ The value given is the mean radiance error averaged over frequencies used to scale water vapor (shown in Table 1).
$^b$ Includes only errors that may be correlated from measurement to measurement (i.e., does not include noise).
$^c$ Uncertainties in line strengths are assumed to be random with frequency, but correlated from measurement to measurement.
atmospheric pressure of the water vapor corrections. The sensitivity is a weighting function \( W \), defined to be the change in radiance per percent change in water vapor, per bar of layer pressure. To calculate \( W \) at a given layer, the water vapor amount is perturbed only at that layer; it is held constant at all other layers. The height coordinate is pressure, rather than altitude, because we are assuming that the solar dry bias itself is a function of pressure. Note that \( W \) is analogous to the kernel for the water vapor retrieval; it is the weight, or contribution, of each layer to the total change in downwelling radiance due to a perturbation. The magnitude of \( W \) varies with frequency, but the shape only varies slightly with frequency (otherwise it would be possible to retrieve water vapor in a height-dependent manner). The sensitivity at 1174.5 cm\(^{-1}\) is shown in Fig. 7 for the average atmospheric profile.

The sensitivity allows us to calculate the mean, or effective, atmospheric pressure as

\[
P = \frac{\int_{650}^{0} W(P)P \, dP}{\int_{650}^{0} W(P) \, dP}.
\]

The effective atmospheric pressure represents the effective height of emission of radiation that accounts for the difference in water vapor. The effective pressure was determined to be 565 mb (dashed line in Fig. 7) at 1174.5 cm\(^{-1}\) and was found to vary with frequency by less than 8 mb. Therefore, our water vapor corrections for Vaisala RS90 radiosondes at SZAs near 62° should best compare to other such results at atmospheric pressure near 565 mb.

6. Results and discussion

Figure 8 shows a profile of water vapor before and after the correction for the afternoon (Fig. 8a) and evening (Fig. 8b) case studies. The mean of the values determined using microwindows and line centers was used. The change in water vapor is considerably greater in the afternoon than in the evening, when the SZA was smaller.

Figure 9 shows the dry biases needed for all afternoon and evening cases. Figure 9a shows the percentage change in RH needed to attain agreement at line centers, and Fig. 9b shows the change needed for microwindows. The means and standard deviations of all evening and afternoon cases are indicated. In the evening (SZAs \( \approx 83° \); gray), the mean correction for microwindows is 8% ± 5% and for line centers is 9% ± 3%. In the afternoon (SZAs \( \approx 62° \); white), the mean correction for microwindows is 20% ± 6% and for line centers is 24% ± 5%. The variation among the results is due to random error in both measured and simulated radiances as well as variability within the batch.
instrument noise is expected to be important only for microwindows and is probably responsible for the 1%–2% greater variability for microwindow results. Noise is probably the cause of the wet bias observed using microwindows for one evening case (−2.4%; Fig. 9b). For the same case, using strong lines results in a dry bias (3.7%; Fig. 9a). Variability in solar heating may explain the 1%–2% greater variability in the afternoon than evening because solar heating is expected to be minimal in the evening.

As discussed previously, the systematic uncertainty is ±9% for results using microwindows and of ±4% for results using line centers. Because the second largest source of error differs for microwindows (the foreign-broadened water vapor continuum) and line centers (water vapor lineshape parameters), using the two different sets of frequencies provides an important check on our scaling method.

For one evening profile, at 2336 LT 12 December 2003, the water vapor correction was determined to be 24.4% using strong lines; this is more than four standard deviations from the mean, so this result is rejected. An important difference for this dataset is that the PAERI measurement was one of the first of the field season, and the PAERI was installed on the surface (0.4 m) instead of on the tower (24 m). Thus, the first 24 m of the atmosphere are included in the analysis. As part of the radiosonde launch procedure, the radiosonde was typically allowed to equilibrate at the surface for about 15 min. During this time, the RH increased from 55% to 80%. It is possible that the radiosonde humidity sensor had not yet equilibrated. If such a problem exists for other radiosondes, the RH would have the first 24 m of the sounding to further equilibrate. After rejection of this measurement, there are 19 RS90 radiosondes remaining for analysis: 10 evening measurements and 9 afternoon measurements.

Previous work (Miloshevich et al. 2006) has shown that RS90 radiosondes can have a significant calibration error. For the present work, the evening measurements were made when the sun was at an angle of about 83°, and solar heating of the sensor arm is expected to be minimal. Thus, we use the evening measurements as an upper limit on the size of errors that are not due to solar heating. Subtracting the mean evening correction from the mean afternoon correction has the additional advantage that it removes systematic errors that are the same in the evening and afternoon. This error is thus 8% relative error in RH for microwindows and 9% for line centers. Dividing the standard deviation in evening dry biases (±5% and ±3%) by the square root of the number of radiosonde observations (10) gives uncertainties of 2% and 1% for microwindows and line centers, respectively. Further uncertainty arises from the assumption that there is no solar heating at SZAs near 83°. Cady-Pereira et al. (2008) estimate the dry bias for an SZA of 83° to be about 1%. The dry bias could be closer to 1.5% at 565 mb if it increases with pressure as suggested by Vömel et al. (2007). The combined uncertainty in the correction is then 3% for microwindows and 2% for line centers. After applying the corrections, we find a solar dry bias for SZAs near 62° of 12% ± 6% (relative error in RH; one standard deviation) using microwindows and of 15% ± 5% using line centers, with systematic uncertainties of 3% and 2%, respectively.

Figure 10 shows the solar dry bias determined in this work for the RS90 radiosondes for SZAs near 62° for an effective pressure of 565 mb (the results for microwindows and line centers have each been offset slightly in Fig. 10 for viewing purposes). The thick gray horizontal lines indicate the standard deviation among measurements; the thin gray horizontal lines indicate uncertainties due to systematic errors. Also shown in Fig. 10 are the water vapor corrections derived by Vömel et al. (2007) for the RS92 radiosonde as a function of pres-
Measurements and simulations of downwelling spectral infrared radiance over Dome C, Antarctica, are used to correct 10 Vaisala RS90 radiosonde humidity profiles measured in the late evening (SZAs near 83°) and 9 measured in the afternoon (SZAs near 62°). The relative humidity profiles are scaled by a constant factor that minimizes the differences between measured and simulated downwelling infrared radiances in microwindows between strong lines of water vapor between 530 and 560 cm\(^{-1}\) and near strong line centers between 1100 and 1300 cm\(^{-1}\). Expressing the dry bias as the relative error in relative humidity (\%), evening dry biases are found to be 8% ± 5% (one standard deviation) using microwindows and 9% ± 3% using line centers. Afternoon dry biases are found to be 20% ± 6% using microwindows and 24% ± 5% using line centers. Systematic uncertainties are 9% using microwindows and 4% using line centers. The corrections correspond to a range of pressures from 650 to 200 mb with an effective pressure of 565 mb.

The correction is considerably larger in the afternoon than in the evening, as expected for a dry bias caused by solar heating of the humidity sensor. Because the sun was so low during the evening measurements, solar heating is expected to cause a dry bias of less than 2% (relative error in relative humidity) in the evening. Thus, we use the mean evening corrections as approximations of error in RH unrelated to solar heating and subtract them from the dry biases measured for SZAs near 62° to estimate the dry bias caused by solar heating. Using microwindows between strong lines, we find that 9 RS90 sensors experience a solar dry bias of 12% ± 6% (relative error in RH; one standard deviation), with an uncertainty due to systematic error of 3%. Using frequencies near strong line centers, we find a solar dry bias of 15% ± 5%, with an uncertainty due to systematic error of 2%. These results complement those of Vömel et al. (2007) for RS92 radiosondes at a variety of atmospheric pressures and at SZAs between 10° and 30° and the results of Cady-Pereira et al. (2008) for a variety of SZAs.

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**Fig. 10.** The solar radiation dry-bias correction for nine RS90 radiosondes determined using atmospheric radiance measurements in microwindows between strong water vapor lines from 530 to 560 cm\(^{-1}\) (circle) and near strong line centers from 1100 to 1300 cm\(^{-1}\) (square). The effective pressure is 565 mb (the results have been shifted slightly in pressure so each symbol can be viewed). The thick horizontal gray lines through the symbols indicate the std dev in the results; the thin gray lines indicate the uncertainty due to systematic error. For comparison, also shown are correction factors derived by Vömel et al. for the RS92 radiosonde, using a Cryogenic Frostpoint Hygrometer (CFH; solid line) and by Cady-Pereira et al. for RS90–92 radiosondes, using a microwave radiometer (MWR; dashed line; RS90–92 radiosondes). Solar zenith angles are indicated.
REFERENCES


