JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 112, D24S37, doi:10.1029/2007JD008698, 2007

Validation of Aura Microwave Limb Sounder water vapor by balloon-borne Cryogenic Frost point Hygrometer measurements

H. Vömel,[1] J. E. Barnes,[2] R. N. Forno,[3] M. Fujiwara,[4] F. Hasebe,[4] S. Iwasaki,[5] R. Kivi,[6] N. Komala,[7] E. Kyrö,[6] T. Leblanc,[8] B. Morel,[9] S.-Y. Ogino,[10,11] W. G. Read,[8] S. C. Ryan,[2] S. Saraspriya,[7] H. Selkirk,[12] M. Shiotani,[13] J. Valverde Canossa,[14] and D. N. Whiteman[15]

Received 24 March 2007; revised 15 August 2007; accepted 29 October 2007; published 29 December 2007.

[1] Here we present extensive observations of stratospheric and upper tropospheric water vapor using the balloon-borne Cryogenic Frost point Hygrometer (CFH) in support of the Aura Microwave Limb Sounder (MLS) satellite instrument. Coincident measurements were used for the validation of MLS version 1.5 and for a limited validation of MLS version 2.2 water vapor. The sensitivity of MLS is on average 30% lower than that of CFH, which is fully compensated by a constant offset at stratospheric levels but only partially compensated at tropospheric levels, leading to an upper tropospheric dry bias. The sensitivity of MLS observations may be adjusted using the correlation parameters provided here. For version 1.5 stratospheric observations at pressures of 68 hPa and smaller MLS retrievals and CFH in situ observations agree on average to within 2.3% ± 11.8%. At 100 hPa the agreement is to within 6.4% ± 22% and at upper tropospheric pressures to within 23% ± 37%. In the tropical stratosphere during the boreal winter the agreement is not as good. The "tape recorder" amplitude in MLS observations depends on the vertical profile of water vapor mixing ratio and shows a significant interannual variation. The agreement between stratospheric observations by MLS version 2.2 and CFH is comparable to the agreement using MLS version 1.5. The variability in the difference between observations by MLS version 2.2 and CFH at tropospheric levels is significantly reduced, but a tropospheric dry bias and a reduced sensitivity remain in this version. In the validation data set a dry bias at 177.8 hPa of −24.1% ± 16.0% is statistically significant.

Citation: Vömel, H., et al. (2007), Validation of Aura Microwave Limb Sounder water vapor by balloon-borne Cryogenic Frost point Hygrometer measurements, *J. Geophys. Res.*, *112*, D24S37, doi:10.1029/2007JD008698.

1. Introduction

[2] Water vapor is one of the most important trace gases in the atmosphere and contributes to many processes at different altitude regions. In the upper troposphere and lower stratosphere it contributes strongly to the radiative balance of the atmosphere and plays a major role in global climate [e.g., *Forster and Shine*, 2002]. It is a source for the OH radical, which is a key compound in atmospheric chemistry and may participate in heterogeneous chemistry involving liquid or ice clouds. Decadal trends in stratospheric water vapor have been detected but are poorly understood [*Oltmans et al.*, 2000; *Rosenlof et al.*, 2001]. Trends in upper tropospheric water vapor are difficult to detect [*Bates and Jackson*, 2001] due to its large variability and the difficulty to accurately determine its concentration in this altitude region.

[3] Accurate monitoring of water vapor is crucial for our understanding of climate processes and our abilities to detect and predict future changes in climate. Observations from space are the only method providing a global distri-

[1]Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado, USA.
[2]NOAA Earth Systems Research Laboratory, Hilo, Hawaii, USA.
[3]Atmospheric Physics Laboratory, University of San Andrés, La Paz, Bolivia.
[4]Graduate School of Environmental Earth Science, Hokkaido University, Sapporo, Japan.
[5]National Defense Academy, Yokosuka, Japan.
[6]Finnish Meteorological Institute, Sodankylä, Finland.
[7]Lembaga Penerbangan dan Antariksa Nasional, Bandung, Indonesia.
[8]Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California, USA.
[9]Laboratoire de Physique de l'Atmosphere, Reunion University, St. Denis, Reunion, France.
[10]Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan.
[11]Kobe University, Kobe, Japan.
[12]Bay Area Environmental Research Institute, Sonoma, California, USA.
[13]Research Institute for Sustainable Humanosphere, Kyoto University, Kyoto, Japan.
[14]Universidad Nacional, Heredia, Costa Rica.
[15]NASA Goddard Research Center, Greenbelt, Maryland, USA.

0148-0227/07/2007JD008698$09.00

bution of water vapor in the different layers of the atmosphere. Different remote sensing techniques have been used, with different characteristics and limitations. The Microwave Limb Sounder (MLS) on board the Aura satellite, which is part of the so called A-Train constellation of Earth observing satellites, is currently a leading satellite instrument providing vertical profiles of water vapor and other important trace gases in the upper troposphere, stratosphere, and mesosphere [*Waters et al.*, 2006]. Details on the water vapor retrievals by MLS are described in a paper by *Read et al.* [2007]. As with many other remote sensing techniques, confidence in these observations is based on independent validation observations, which characterize the remote sensing data. Limited validations for water vapor and several other trace gases were presented by *Froidevaux et al.* [2006]. These comparisons of version 1.5 with other satellite and remote sensing instruments indicated that MLS stratospheric water vapor is accurate to within roughly 10% level of accuracy. For tropospheric comparisons only data from the Atmospheric Infrared Sounder (AIRS) onboard the Aqua satellite were available and a number of open questions remained.

[4] Here we present in situ observations of water vapor obtained by the balloon-borne Cryogenic Frost point Hygrometer (CFH). Extensive observations have been timed with Aura/MLS overpasses at a number of locations covering every geographical region except the Antarctic winter stratosphere. These measurements provide independent reference observations, which we use to validate the upper tropospheric and lower to middle stratospheric water vapor product of MLS.

2. Instrumentation and Observations

2.1. Cryogenic Frost Point Hygrometer

[5] The CFH is a chilled mirror instrument capable of measuring the large range of water vapor concentrations found in the troposphere and stratosphere. It is carried up by small meteorological balloons and measures a water vapor profile between the surface and the middle stratosphere with high vertical resolution. The instrument has been described in detail by *Vömel et al.* [2007].

[6] Like many chilled mirror instruments, CFH is not calibrated for water vapor and can be considered an absolute reference for water vapor measurements. It measures the temperature at which an ice layer is in equilibrium with the gas phase of water passing over this ice layer. The largest source of uncertainty in CFH water vapor measurements is the stability of the feedback controller, which maintains the constant frost layer on the mirror. The total uncertainty in frost point is better than 0.5 K throughout the entire profile [*Vömel et al.*, 2007], which translates to a mixing ratio uncertainty of about 4% in the lower tropical troposphere to about 10% in the middle stratosphere and tropical tropopause.

[7] The only limitations are measurements inside liquid clouds, which may disable the instrument due to wetting of the detector lens, and contamination near the balloon ceiling due to outgassing from any surface of the flight train. These artifacts are screened out in the data processing.

2.2. MLS Water Vapor Retrieval

[8] The MLS version 1.5 water vapor product is the first major release for which all observations have been processed. For version 2.2, which has become the production version since April 2007, not all days have been reprocessed; however, as will be seen later, the differences between both versions appear to be small and the conclusions drawn for version 1.5 appear to be largely valid for version 2.2 as well.

[9] The most significant difference between the two versions is the different vertical gridding, on which these data are reported. Version 1.5 data are reported on 6 pressure levels per decade (lpd), whereas version 2.2 data are reported on 12 lpd up to 21 hPa, after which the resolution reverts back to 6 levels per decade. This increase in vertical gridding was achieved by a number of retrieval configuration changes, which are discussed in detail by *Read et al.* [2007].

[10] Following the recommendations for these data sets, profiles have been screened using the appropriate status flags, precision values, and quality flags. Only profiles with even values of the status field were used, which indicate that the retrieved profile passed a number of rejection criteria. Data were required to have associated quality values larger than 0.9, indicating that for these data a good fit between the observed radiances and those values computed by the forward model using the retrieved values was achieved. Furthermore, data with negative precision values were rejected, since these data points did not have a sufficient information yield from MLS.

[11] Despite these recommended filters, a few outlier values were found at tropospheric levels, which heavily skewed the fit between CFH and MLS observations described in section 4.4. Most of these outliers were filtered out using MLS status bit one, which indicates that the profile might be questionable. In all cases the flags indicated that the profiles may have been affected by low-altitude clouds. Using this filter had only a minimal impact on the comparison average and slightly decreased the standard deviation, however, it had a significant impact on the correlations described in section 4.4.

[12] Since MLS and CFH have vastly different vertical resolutions, the resolution of the CFH data has been degraded to match that of MLS. The basics of this resolution degrading is discussed by *Read et al.* [2007, equation (1)]. In short, the observed in situ profile is multiplied by the forward model smoothing function and the retrieval averaging kernel. In version 1.5, the averaging kernel for pressures larger than 68 hPa is nearly a unity matrix, which means that the forward model smoothing function describes the smoothing completely. In version 2.2, both the averaging kernel and the forward model smoothing function contribute to the smoothed downsampling of CFH data to the MLS grid points. All CFH data used in this study have their vertical resolution degraded using the forward model smoothing function and for comparisons with version 2.2 additionally using the appropriate retrieval averaging kernel.

[13] The smoothing operation is only valid for situations where the measurement system responds linearly to the profile fluctuations being smoothed. For MLS water vapor (both versions) this is unlikely to be the case for pressures

Table 1. Cryogenic Frost Point Hygrometer (CFH) Launch Locations and Number of Profiles Coincident With Microwave Limb Sounder (MLS) Observations Within 3 h and 300 km for Version 1.5 and Within 6 h and 300 km for Version 2.2

Location (Campaign)	Latitude	Longitude	Period/Campaign	Matches v1.5	Matches v2.2
Beltsville, Maryland (WAVES)	39.05	−76.88	Oct 2006	9	3
Biak, Indonesia (SOWER)	−1.17	136.06	Jan 2006 and Jan 2007	2	1
Boulder, Colorado	39.95	−105.20	Feb 2005 to Jun 2006	6	2
Hanoi, Vietnam (SOWER)	21.01	105.80	Jan 2007 (SOWER)	1	-
Heredia, Costa Rica (Ticosonde)	10.00	−84.11	Jul 2005 to Jun 2007	11	6
Hilo, Hawaii	19.43	−155.04	Dec 2005 and Apr 2006	2	-
JPL TMF, Wrightwood, California(MOHAVE)	34.38	−117.68	Oct 2006	3	4
Sodankylä, Finland[a]	67.37	26.63	Feb 2005 to Aug 2006	5	5
St. Denis, La Reunion, France	−21.06	55.48	Sep 2005 and Jun 2006	3	2
Tarawa, Kiribati (SOWER)	1.35	172.92	Jan 2007	1	-

[a]At Sodankylä the time difference was relaxed to within 6 h for version 1.5.

greater than 147 hPa and unquantifiable errors may be introduced by the smoothing.

2.3. CFH Campaigns

[14] CFH soundings have been made at a number of sites, which are listed in Table 1. The number of soundings at each station gives the number of soundings that were launched within 300 km and 3 h of an MLS overpass for version 1.5 and within 300 km and 6 h for version 2.2. These overpass criteria are somewhat arbitrary. Tighter overpass criteria provide better coincidences with less opportunity of atmospheric change between observations at the expense of a smaller number of coincidences. More relaxed overpass criteria provide a larger sample of coincident measurements, but with a possibility of larger atmospheric change between the observations. The analyses discussed below were done with different overpass criteria to test the influence of these criteria. We tested overpass criteria up to 900 km and 12 h and found that the results are largely insensitive to the overpass criteria in this range. For this reason the overpass criteria at Sodankylä were relaxed to observations within 300 km and 6 h to provide a sample large enough for a statistically relevant analysis.

[15] At Costa Rica three Ticosonde campaigns took place with a larger number of soundings. CFH/ozone sondes were launched during the Tropical Cloud Systems and Processes (TCSP) campaign in July 2005, the CR-AVE (Costa Rica Aura Validation Experiment) in January and February 2006, and the Ticosonde/Veranillo campaign in July 2006. In January 2006 CFH/ozone soundings were also launched at Biak, Indonesia as part of the Soundings of Ozone and Water in Equatorial Regions (SOWER) campaign, providing observations over the maritime continent during the same period as the CR-AVE campaign in Central America. The observations at Hanoi, Vietnam, and Tarawa in January 2007 were also part of the SOWER campaigns.

[16] The Water Vapor Validation Experiment–Satellite/Sondes (WAVES 2006) campaign, which took place in July/August 2006 at the Howard University Campus in Beltsville, Maryland, and the Measurements Of Humidity in the Atmosphere and Validation Experiment (MOHAVE), which took place at the Table Mountain Facility of Jet Propulsion Laboratory (JPL) in Wrightwood, California, were dedicated validation campaigns, which provided a substantial amount of data from different remote sensors as well as several in situ instruments. Here we only use CFH observations obtained during these campaigns, since only CFH observations covered the upper troposphere and lower stratosphere.

[17] Observations at Boulder, Colorado, and Hilo, Hawaii, are dedicated Aura validation soundings, whereas the coincident measurements at Sodankylä, Finland, and at Reunion Island are opportunity comparisons as part of the water vapor programs at these stations.

3. Validation of Version 1.5

3.1. Global Mean Difference

[18] An example for a comparison in mid latitudes is shown in Figure 1. This sounding was launched at the Table Mountain Facility of JPL in Wrightwood, California, as part of the MOHAVE campaign in October 2006. At the time of the overpass the balloon had reached 200 hPa with a horizontal separation to the MLS observation of 137 km. The comparison for this example is shown in Figure 1b and typical for the average of all soundings. Throughout this study, we use the relative difference in units of percent defined as

$$\delta q = 100 \cdot \frac{q_{MLS} - q_{CFH}}{q_{CFH}}, \quad (1)$$

where q_{MLS} is the water vapor mixing ratio observed by MLS at any pressure level and q_{CFH} the CFH observation at that level derived from the properly downsampled CFH profile measurement (see section 3.2).

[19] Between 68 hPa and 14 hPa both profiles show agreement within 8%, which is less than the instrumental uncertainty of either instrument. At 100 hPa MLS is 26% wetter than CFH and at 146 hPa MLS is 45% drier.

[20] The comparison for all soundings listed in Table 1 is shown in Figure 2. MLS v1.5 agrees on average with CFH measurements at pressures of 68 hPa and smaller to within 2.3% ± 11.8%. At 100 hPa the mean difference is 6.4% ± 22% with a median difference of 5%. At higher pressures the variability increases with increasing pressure. The mean difference indicates a 23% ± 37% dry bias at 215 hPa and an 11.3% ± 50% dry bias at 316 hPa. However, because of the large variability in the tropospheric comparisons, these mean biases are statistically not significant.

[21] Water vapor concentrations at the upper tropospheric levels (146 hPa, 215 hPa, and 316 hPa) can vary by more than two orders of magnitude. In addition to horizontal

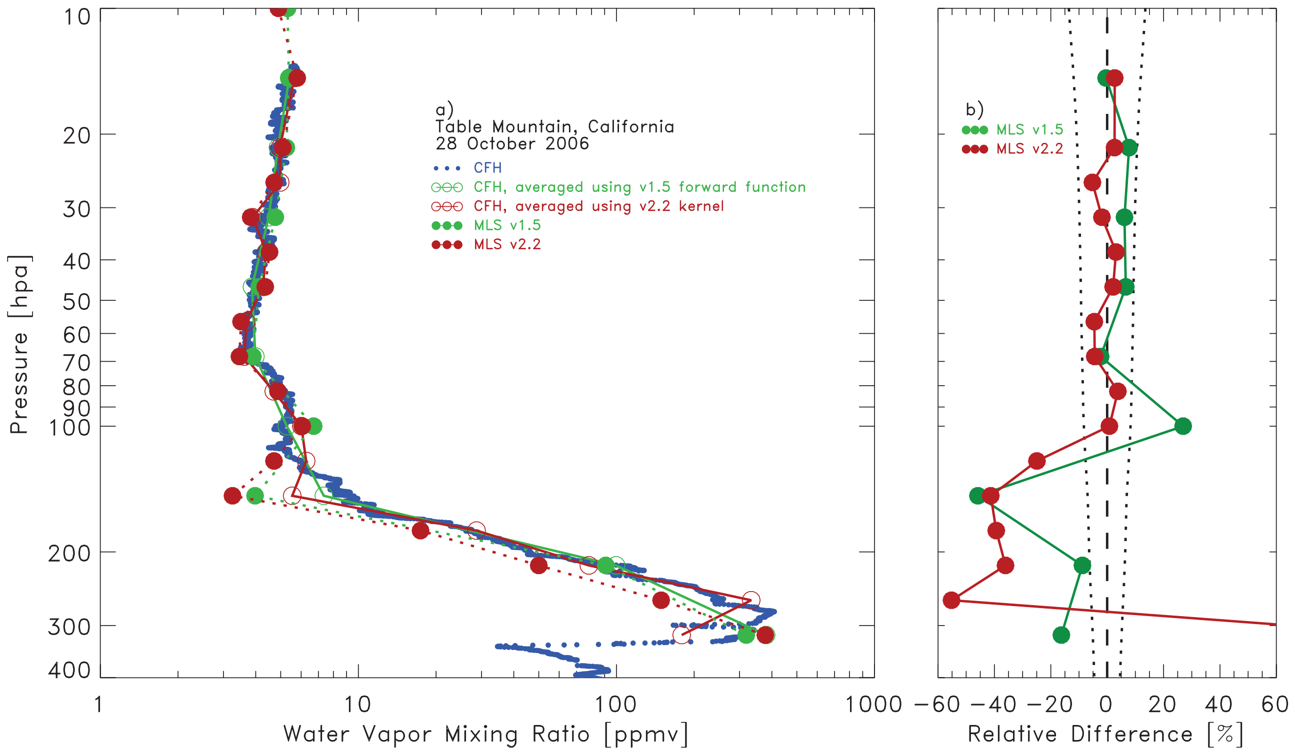

Figure 1. (a) Cryogenic Frost point Hygrometer (CFH) water vapor profile and Microwave Limb Sounder (MLS) observations near Table Mountain, Wrightwood, California. The balloon was launched 30 min prior to the overpass and reached 200 hPa at the time of the overpass. The horizontal separation was roughly 130 km. MLS v1.5 and v2.2 are shown as well as the appropriately degraded CFH data. Also shown is (b) relative difference δq. Dotted lines indicate CFH uncertainty.

gradients that are not detected by CFH and smoothed out by MLS, this large range of water vapor concentrations contributes to the large variability in the tropospheric levels. However, the difference profile remains largely unchanged if we limit the comparisons to mid latitude soundings only, which dominate the data set in Figure 2. The results for only tropical and polar soundings are discussed separately below. The difference profile is largely insensitive to the overpass criteria used to select coincident observations. This implies that for the average comparison the horizontal and temporal separation is mostly irrelevant.

3.2. Tropics

[22] The tropics are the most important region for stratospheric water vapor observations since the tropical tropopause regulates the amount of water vapor that is directly transported from the troposphere into the stratosphere. The seasonal cycle of tropical tropopause temperatures leads to a seasonal cycle in stratospheric water vapor, which is propagated upward and leads to a distinct sequence of water vapor maxima and minima, which has been dubbed "the atmospheric tape recorder" [*Mote et al.*, 1996]. A precise detection of this tape-recorder signal is essential to properly quantify the input of water vapor through the tropical tropopause.

[23] The intensive CFH observations as part of CR-AVE in January and February 2006 at Heredia, Costa Rica, and the SOWER campaigns at Biak, Indonesia, in January 2006 and January 2007 provide observations in two different boreal winters, with very different phases of the Quasi Biannual Oscillation (QBO). For comparison we only look at the period between 7 January and 17 January of 2006 and of 2007. Since each period has only few coincident observations, we use a slightly modified approach. Instead of coincident profile comparisons, here we consider all CFH observations in this 2 week time frame as well as all MLS observations within 3° of latitude and 30° of longitude. Differences of the mean profiles are a good approximation of the difference between climatological profiles measured by each instrument. Since the stratospheric variability is small, the results reflect with better statistics what coincident profile comparisons would have shown with limited statistics. Even tropospheric average comparisons are useful.

[24] The water vapor minimum in the Tropical Tropopause Layer (TTL) during the boreal winter typically lies between 70 hPa and 90 hPa (Figures 3a–3c). The MLS level nearest the water vapor minimum is the 100 hPa level, which in all in situ observations lies well within the water vapor gradient of the upper troposphere. In addition the coarse MLS sampling places the minimum at the 100 hPa level. Therefore MLS does not capture the actual minimum and averages over parts of the upper tropospheric profile. Nonetheless, with the proper smoothing applied to the in situ data CFH and MLS agree within their error bars at 100 hPa at Costa Rica as well as at Biak where the water vapor minimum is nearly half that observed at Costa Rica.

[25] The CR-AVE and SOWER data of January 2006 (Figures 3a and 3b) show that in that year MLS overestimates

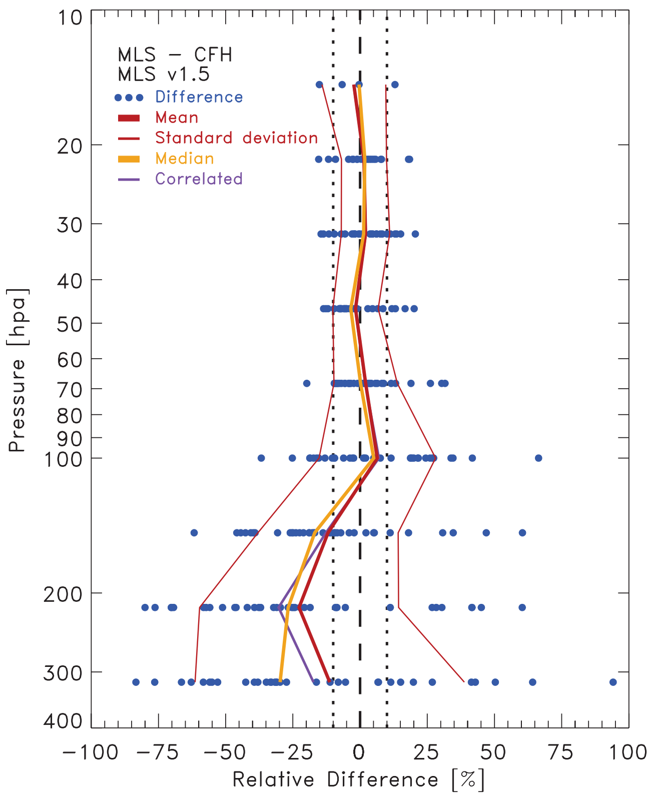

Figure 2. Mean and median of the relative difference δq between MLS v1.5 and CFH water vapor profiles for all coincident observations. The dotted vertical lines approximate the accuracy of CFH.

the water vapor concentration at 68 hPa and 32 hPa which are the pressure level below and above the stratospheric water vapor maximum at that time and underestimates the water vapor concentration at 46 hPa which is close the pressure level where CFH observed the "tape recorder" maximum. This means that the MLS profile shows a smaller vertical tape recorder amplitude than CFH. As described in section 3.2, CFH data are degraded in resolution using the MLS forward model smoothing function, which means that the in situ data represent what MLS should be detecting. The smoothing of the tape recorder shown here is in excess of the smoothing inherent to this remote sensing instrument.

[26] The SOWER data of January 2007 (Figure 3c), on the other hand, provide a very different picture. During the January 2007 campaign at Biak the tropopause was roughly 3 K warmer than in the previous year and the water vapor minimum was comparable to that in CR-AVE data in January 2006. Related to the warmer tropopause temperature is the higher pressure of the seasonal maximum which is located at 60 hPa compared to around 46 hPa in the previous year. In contrast to January 2006, the MLS water vapor profile in January 2007 agrees well with the in situ data over the entire lower tropical stratosphere.

[27] Therefore the good agreement observed in the global mean comparison does not hold to the same degree over the boreal winter tropical stratosphere, where the vertical profile of water vapor shows larger variations. The retrieval depends to some degree on the structural details of the vertical water vapor profile. Biases vary between different years, implying that biases might vary during the course of the year as the signal of the tropopause temperature propagates upward.

[28] Vertical gradients comparable to or even larger than those of the tape recorder may be found in the dehydrated stratosphere during the Antarctic winter. Although we do not have in situ comparisons for this season and region, we can speculate that detailed structures of the Antarctic dehydration may be smoothed in excess of the intrinsic MLS smoothing.

[29] During the boreal summer months the vertical profile of water vapor in the lower stratosphere shows a very broad minimum at about 62 hPa. Two campaigns took place in Costa Rica during July 2005 and July 2006. Both years show nearly the same vertical structure in the comparison between CFH and MLS (Figure 3d for July 2005). The comparison at stratospheric pressures less than100 hPa is similar to the global mean comparison.

[30] The coincident comparisons in the upper troposphere show a dry bias in MLS data, which is larger than the global mean comparison (Figures 3b–3d). It is interesting to note that the observations during the rainy season (January for Indonesia and July for Costa Rica) show the largest amounts of water vapor in the upper troposphere and the largest dry bias for MLS. The observations during the dry season at Costa Rica (January 2006) show much smaller amounts of water vapor in the upper troposphere and a smaller dry bias. Thus these comparisons provide an indication for a mixing ratio dependent dry bias in the upper troposphere.

3.3. Polar Winter

[31] Soundings were made at Sodankylä, Finland, during the months of December through March in 2004/2005, 2005/2006, and 2006/2007. Here the comparison (Figure 4) shows good agreement at pressures of 146 hPa and smaller as well as at the 316 hPa level and a small dry bias at 215 hPa. The standard deviation of this comparison is at the instrumental uncertainty up to 146 hPa and then increases to ±60% at 316 hPa. Since 146 hPa is always well within the stratosphere during the polar winter, this result is in agreement with the general stratospheric results.

[32] As mentioned above, in the Antarctic winter stratosphere, with its recurrent dehydration and strong vertical gradients, the comparisons may differ from this result; however, we do not have data to evaluate this in detail.

3.4. Linear Correlation

[33] The comparisons presented above show individual and averaged relative differences between MLS and CFH observations, without considering the systematic relationship between these observations. This simple comparison is expanded in Figure 5, which shows the relationship between MLS and CFH observations at each pressure level. A linear regression model is fit to the observations at each level according to

$$q_{i,MLS} = \alpha + \beta \cdot q_{i,CFH} + \eta_i, \quad (2)$$

where β is the linear slope, α the constant offset, and η_i the residual not captured by this regression. The regression model is weighted by the precision of each MLS

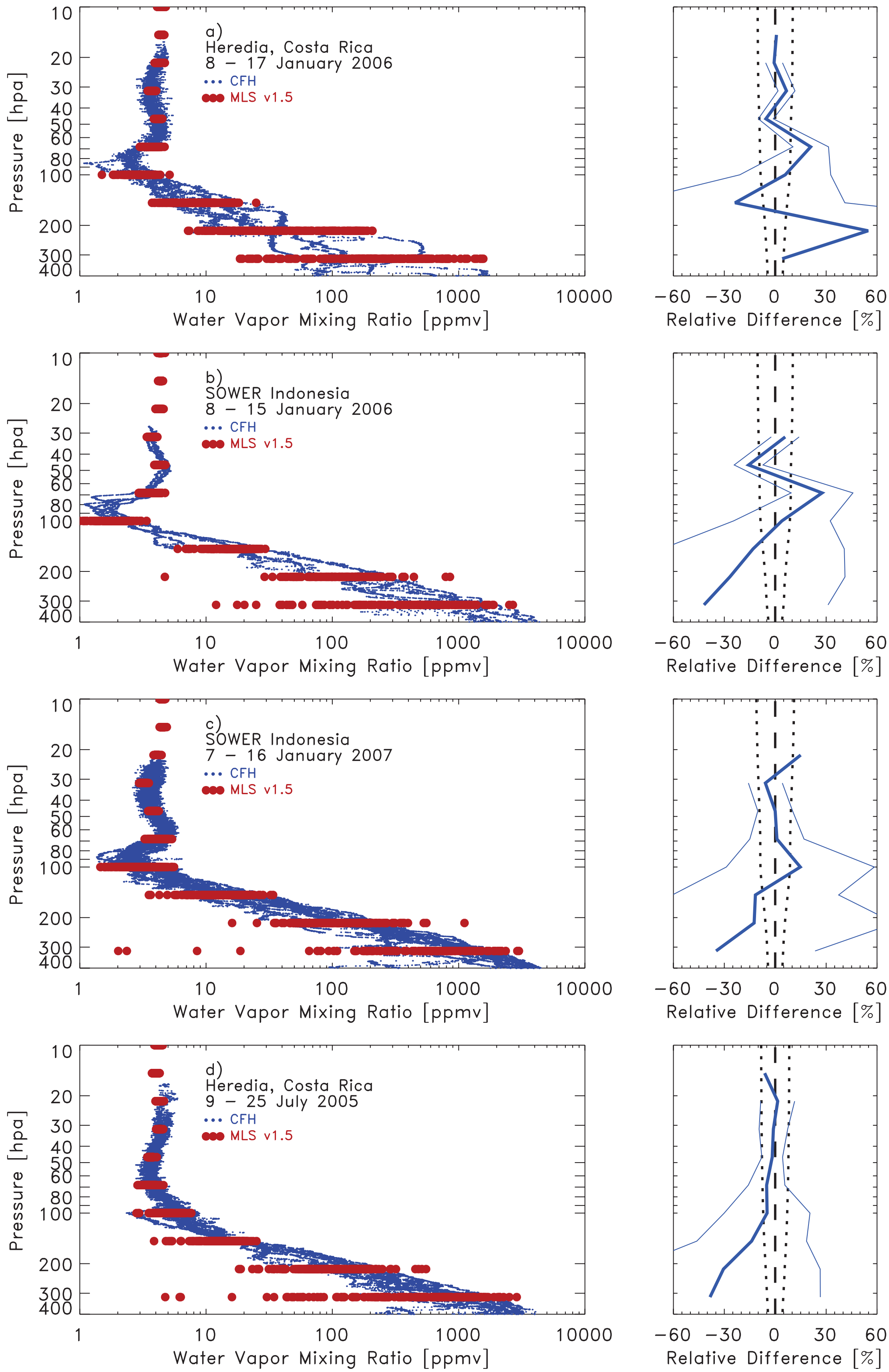

Figure 3. Tropical comparisons between MLS version 1.5 and CFH at Costa Rica and Indonesia: (a and b) January 2006, (c) January 2007, (d) July 2005. The standard deviation shown in the difference plots is the combined standard deviation of both data sets.

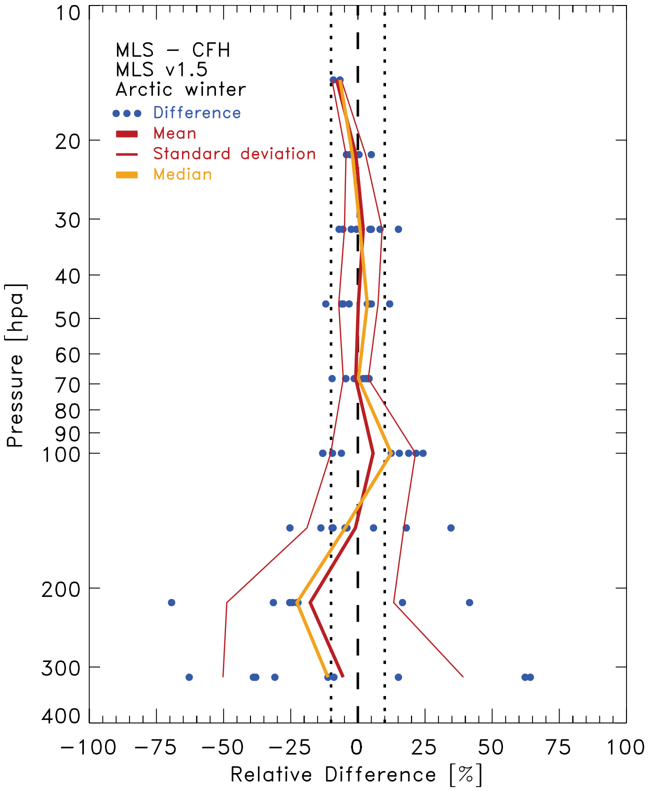

Figure 4. Same as Figure 2 for Arctic winter coincident observations.

observation, which reduces the influence of outlier observations but does not influence the correlation significantly. The regression parameters indicated in Figure 5 are listed in Table 2. For perfect agreement between MLS and CFH the slope β would take the value 1 and the offset α would take the value 0. However, the slope ranges between $\beta = 0.56$ and $\beta = 0.85$ with an average value of $\overline{\beta} = 0.71$. This implies that the sensitivity of MLS is about 30% lower than that of CFH. The offset ranges between $\alpha = -2.4$ ppmv and $\alpha = 1.8$ ppmv. The data at 316 hPa and 215 hPa show the largest variability and the poorest correlation, which, however, is still significant.

[34] The correlation provides a consistent result over all pressure levels, with stronger deviations only at the tropospheric levels. This may be due to two reasons. First, the smoothing operation makes the assumption of a linear response, which may not be a good assumption at the higher mixing ratios of the upper troposphere, introducing additional errors. Second, water vapor concentrations at tropospheric levels span a much larger range and atmospheric variability and instrumental variability contribute strongly to the larger scatter at these levels. The linear correlations at the tropospheric levels are strongly influenced by the higher mixing ratio values. In the panels for 316 hPa and 215 hPa, which are shown as log-log plots instead of linear plots, the linear fits lines exhibit a slight curvature at the lower mixing ratios as a result of this offset.

[35] Figure 5 indicates that the correlation is independent of the latitude band in which the observations were taken and appears to be a function of the observed mixing ratio only. The observed variability of the correlation coefficients at different pressure levels may be interpreted as random variability, although some dependency of the correlation coefficients between the different pressure levels cannot be excluded.

[36] A slope β less than one implies that the sensitivity of MLS is less than that of CFH and hence a dry bias of MLS. However, since the offset α is other than zero, this slope is not equal to the dry bias derived above. We can express the average relative difference shown in Figure 2 using the correlation parameters listed in Table 2:

$$\begin{aligned}\overline{\delta q} &= \overline{\left(\frac{q_{i,MLS}}{q_{i,CFH}} - 1\right)} \\ &= (\beta - 1) + \alpha \cdot \overline{\left(\frac{1}{q_{i,CFH}}\right)} + \overline{\left(\frac{\eta_i}{q_{i,CFH}}\right)}, \\ &\cong (\beta - 1) + \alpha \cdot \overline{\left(\frac{1}{q_{i,CFH}}\right)} \end{aligned} \quad (3)$$

which is shown in Figure 2 as "correlated" relative difference. Equation (3) shows that the slope parameter directly relates to the dry bias only for large mixing ratios, when the offset multiplied by the average inverse mixing ratio becomes small. This also explains why the average dry bias $|\overline{\delta q}|$ is smaller than sensitivity deficit $|\beta - 1|$. For the low mixing ratios found in the stratosphere, the reduced sensitivity is compensated by the offset resulting in a good average agreement. At the higher tropospheric mixing ratios a relative dry bias remains and is mixing ratio dependent. The average relative difference $\overline{\delta q}$, therefore, is influenced by the distribution of mixing ratios in the data set.

[37] Using the reverse of equation (2), the MLS observations at each level can be adjusted to match the sensitivity of CFH:

$$q'_{MLS} = \frac{1}{\beta} q_{MLS} - \frac{\alpha}{\beta} \quad (4)$$

where α and β are the regression parameters derived at each pressure level and listed in Table 2.

[38] This adjustment significantly improves the agreement between MLS and CFH at the tropospheric levels and by definition improves the MLS sensitivity at all levels. However, it increases the standard deviation of the relative difference by around 1/10 at the stratospheric levels and up to 1/5 at the tropospheric levels. This fact indicates that a simple linear adjustment only partially corrects the average dry bias and that other nonlinear processes are not revealed by this approach.

[39] A limitation of this simple linear approach is that it is based only on the empirical correlation between MLS and CFH water vapor. The MLS water vapor measurements are based on logarithm of the retrieved water vapor, not on the retrieved water vapor itself [see *Read et al.*, 2007, equation (1)]. This means that a better correlation could be done using $ln(q_{MLS})$ and $ln(q_{CFH})$. However, an adjustment based on this correlation does not improve the average relative difference over the more simple linear correction and leads to nearly the same increase in standard deviation. In addition it no longer allows relating the correlation with

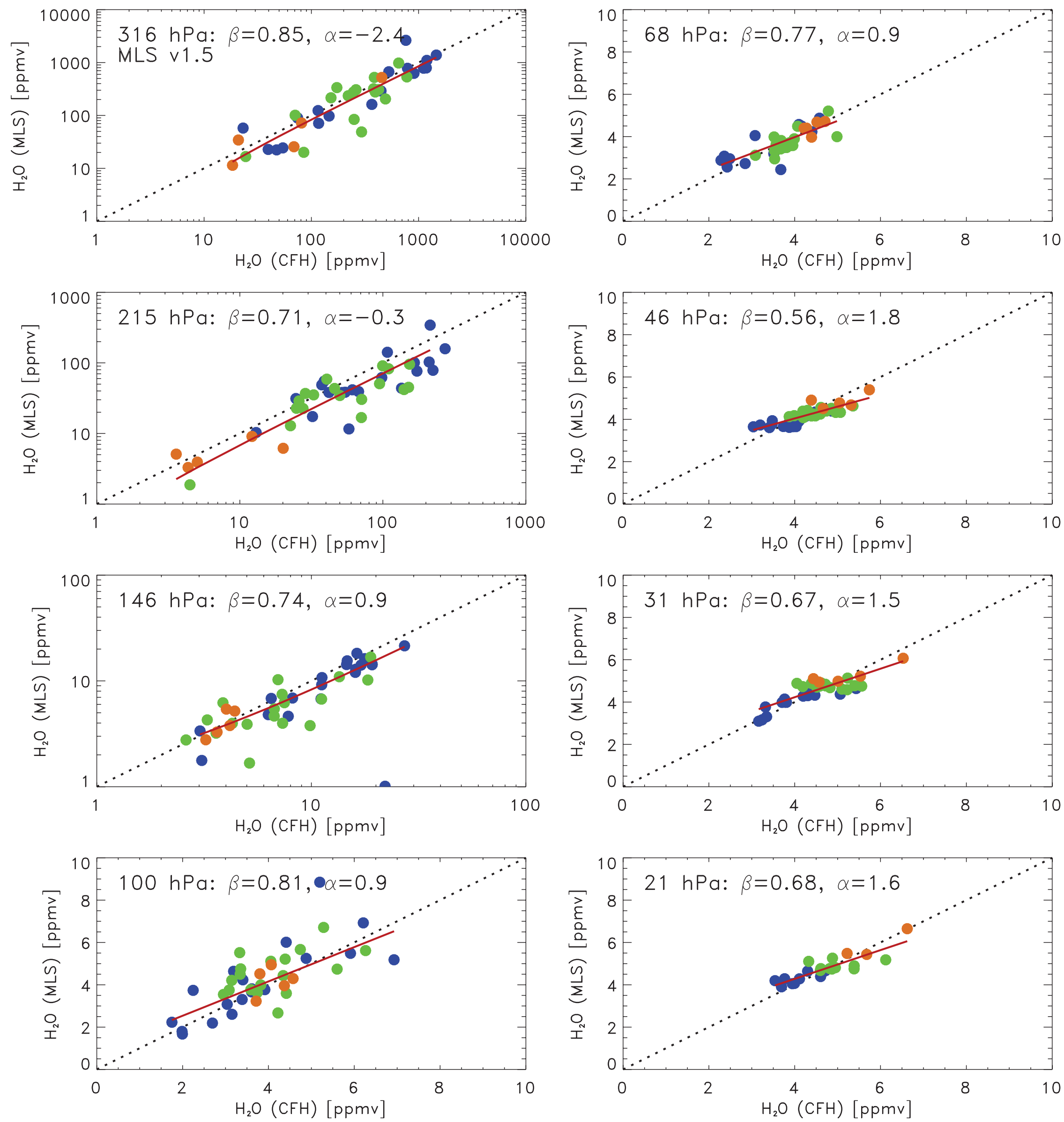

Figure 5. Correlation between MLS v1.5 and CFH at each pressure level. The slope (β) and the offset (α) is given for a linear fit to the data. The color coding indicates the latitude of the soundings, where orange represents high latitude, green represents midlatitude, and blue represents the tropics. Note the logarithmic axes for 316 hPa, 215 hPa, and 146 hPa.

the mean relative dry bias as is shown for the linear correlation in equation (3).

4. Validation of Version 2.2

[40] The most important difference between MLS version 1.5 and version 2.2 is the different vertical sampling, which also requires a slightly different smoothing (see section 3.2). An example of the impact of the smoothing for version 2.2 is given in Figure 1, which shows the difference in vertical sampling of version 1.5 and 2.2 as well as the differently degraded in situ data.

[41] Owing to the limited amount of MLS version 2.2 data the comparison with CFH observations is not as extensive as for version 1.5. In particular there are not as many low-latitude observations (see Table 1), and there are fewer tropospheric comparison with large water vapor amounts. There are also not yet enough coincident tropical observations during the boreal winter 2005/2006 and 2006/2007.

[42] Nevertheless, these observations allow a statistically significant profile comparison, which allows the evaluation of the general features of version 2.2. Figure 6 shows the relative difference profile for version 2.2 as well as the difference profile for version 1.5 using the exact same soundings and MLS overpass retrievals.

[43] In the stratosphere at pressures between 68 hPa and 21.5 hPa MLS v2.2 and CFH agree on average to within 2.7% ± 8.7%, which is comparable to version 1.5. In the vicinity of the tropical tropopause the average difference is

Table 2. Regression Coefficients for Linear Regression Between CFH and MLS v1.5 Observations

Pressure	Slope β	Offset α	$\left(\frac{1}{q_{i,CFH}}\right)$
316.2	0.85	−2.4	0.011
215.4	0.71	−0.3	0.046
146.8	0.74	0.9	0.152
100.0	0.81	0.9	0.283
68.1	0.77	0.9	0.278
46.4	0.56	1.8	0.235
31.6	0.67	1.5	0.227
21.5	0.68	1.6	0.214
14.7	0.61	1.8	0.200

−1.0% ± 9.7% and 3.6% ± 12.7% at 100 hPa and 82.5 hPa, respectively. We need to point out that despite the increase in vertical gridding, the variability did not increase; in fact nearly the opposite is the case: between pressures of 68 hPa and 215 hPa, the variability in version 2.2 is smaller compared to version 1.5. Most strikingly the variability remains low up to 177 hPa, making the average dry bias of −24.1% ± 16% at this level statistically significant. At 261 hPa and 316 hPa the dry bias apparently decreases and the variability increases, driven by a few tropical observations, which show a large wet bias of MLS.

[44] Despite the fact that the temporal coincidence criteria for version 2.2 was increased by a factor of 2, the variability of the comparison at tropospheric levels is lower in version 2.2 compared to the same data set in version 1.5. Analyses with different coincident criteria strongly affected the sample size but did not lead to different results. This indicates that the spatial and temporal coincidence criteria do not play a large role in this comparison as indicated above.

[45] At the moment we only have a limited amount of tropical comparisons with version 2.2 to test whether the measurement of the tropical tape recorder shows a similar interannual variation as was found in the comparison with version 1.5. Initial comparisons based on a small data set appear to show a slight improvement, but due to the small number of observations this result would have to be called inconclusive. For polar comparisons the results for version 2.2 are comparable with version 1.5.

[46] The linear correlation between MLS and CFH discussed in section 4.4 for version 1.5 shows very similar features in version 2.2 (Figure 7). Table 3 lists the correlation coefficients at each pressure level. At pressures of less than 177 hPa the offset varies between α = 0.19 ppmv and α = 2.4 ppmv, with an average of α = 0.93 ppmv. The slope parameter varies between β = 0.45 and β = 1.08 with an average value of β = 0.79. This means that in the TTL and stratosphere the sensitivity in version 2.2 is improved and closer to that of the CFH compared to version 1.5.

[47] At pressures of 215 hPa and higher the linear fit shows the largest offset values and deviates strongly from the data at low mixing ratios. This is a result of the larger scatter of data at the higher pressure levels and from the strong influence of the high mixing ratio values on the linear fit. For example, the linear fit at 316 hPa (Figure 7, top left) is mostly controlled by three tropical observations, contributing mixing ratios above 1000 ppmv, where MLS shows a strong wet bias. The offset has a value of α = −233 ppmv, the slope a value of β = 3.0, and the correlation parameter a value of r^2 = 0.65. Excluding the three tropical data points

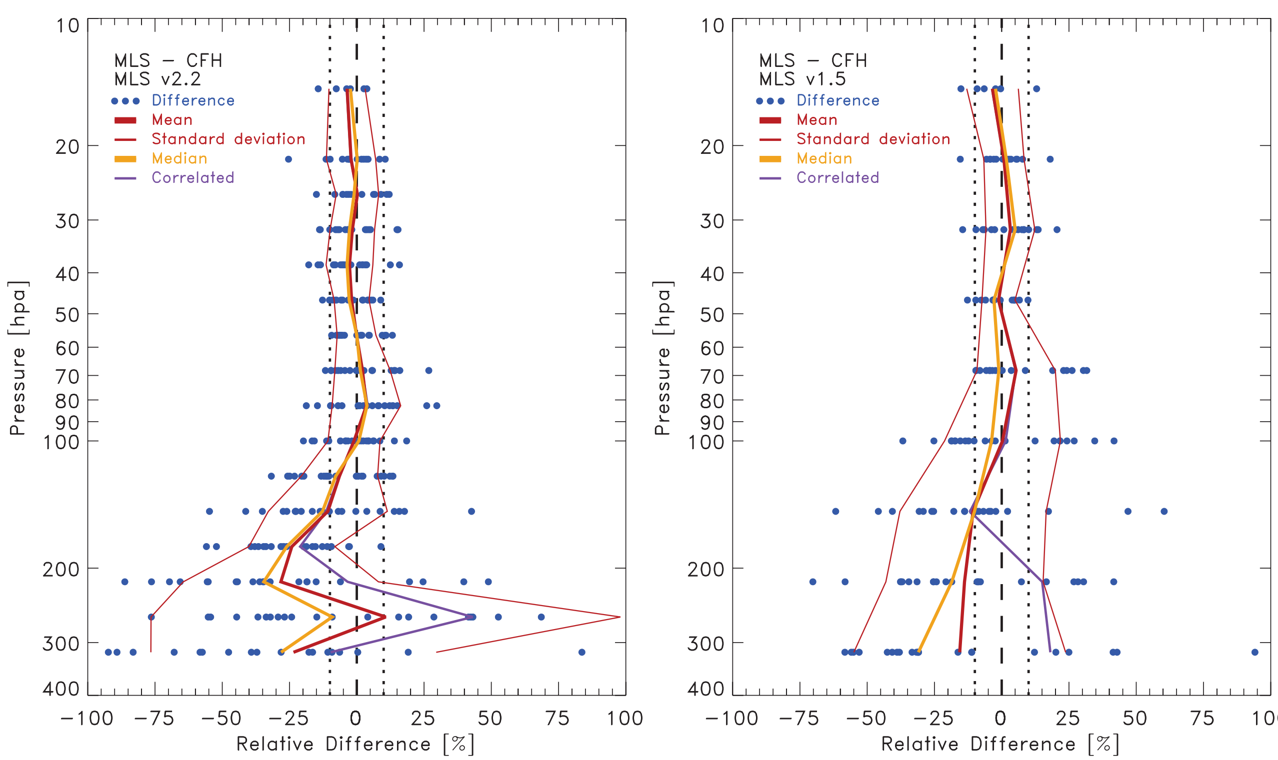

Figure 6. (a) Relative difference between MLS v2.2 and CFH and (b) relative difference for the same data set except using version 1.5. The dotted vertical lines indicate the approximate accuracy of CFH.

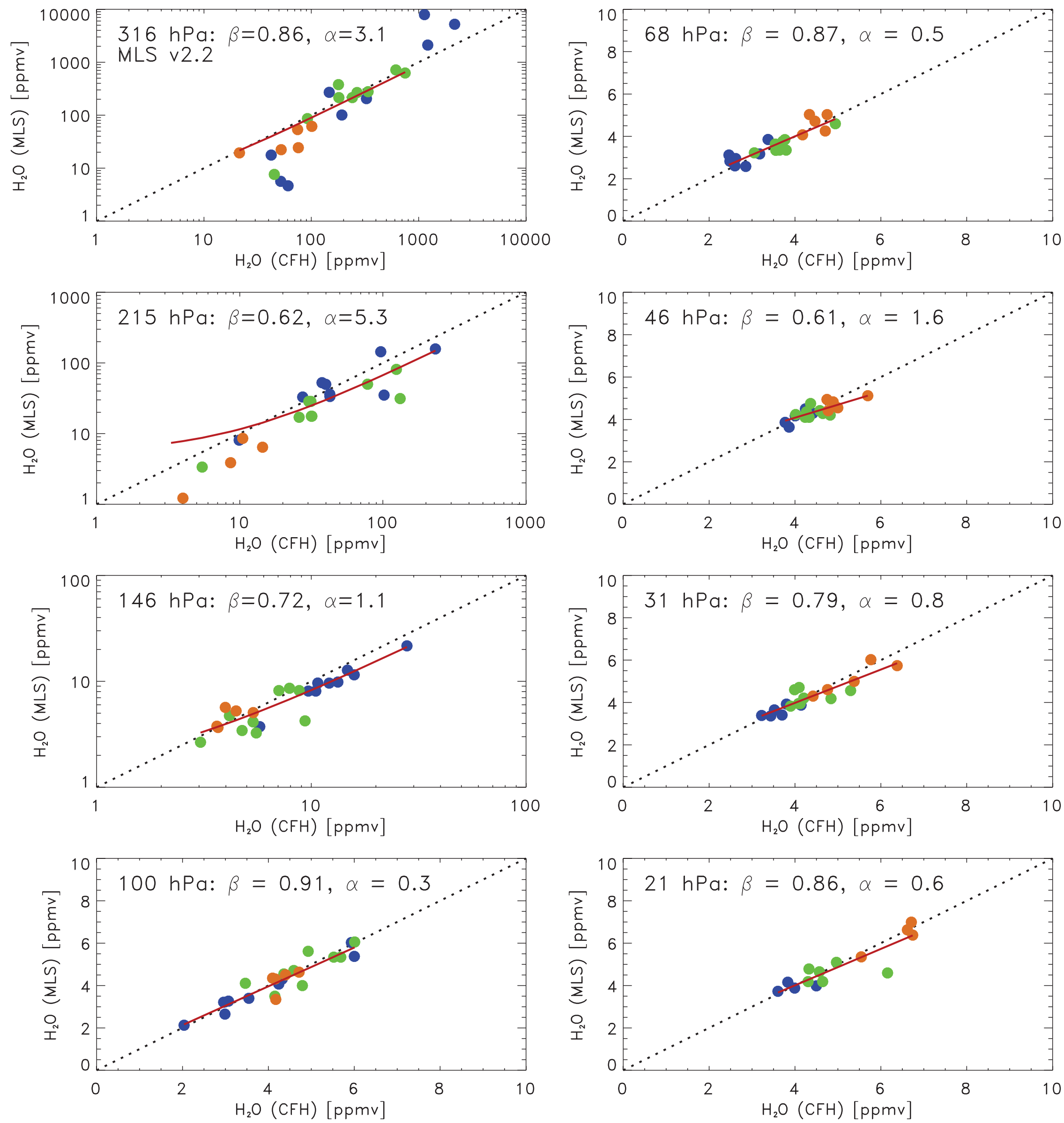

Figure 7. Same as Figure 5 for MLS version 2.2.

with mixing ratios above 1000 ppmv provides a much better fit ($r^2 = 0.81$) with $\alpha = 3.1$ ppmv and $\beta = 0.86$. Incidentally, these values are similar to the stratospheric average. Our comparison with version 2.2 is limited and is strongly influenced by large deviations at high mixing ratios. However, this example indicates, that the average of the stratospheric correlation parameters could be extended to adjust the tropospheric values.

5. Summary and Discussion

[48] MLS v1.5 and v2.2 show agreement with CFH for stratospheric pressures below and including 68 hPa to within 2.3% ± 11.8% and 2.7% ± 8.7%, respectively. At 100 hPa, which is close to the tropical tropopause, version 2.2 shows a mean difference of −1.0% ± 9.7% compared to 6.4% ± 22% for the same version 1.5 data set.

[49] The level of variability observed in the stratospheric comparison with version 1.5 and version 2.2 is comparable to the variability within CFH observations during intensive field campaigns such as TCSP [*Vömel et al.*, 2007], CR-AVE, WAVES, MOHAVE, or SOWER. Since atmospheric variability is ignored in this comparison, the precision of MLS observations is therefore significantly smaller and does not contribute much to the variability of the comparisons.

[50] The coincidence criteria to match CFH and MLS observations proved to have little influence on the average results and only changed the sample size. This is a clear indication that atmospheric variability between each pair of observations was not sufficient to explain the differences. Horizontal smoothing of the water vapor profile in MLS observations may be of concern in particular near convection. This could potentially lead to drier observations by MLS compared to CFH, since drier airmasses in the field of view of MLS away from convection are averaged into the MLS observations. However, CFH soundings were not preferentially made near convection; in fact, due to the

Table 3. Regression Coefficients for Linear Regression Between CFH and MLS v2.2 Observations

Pressure	Slope β	Offset α	$\overline{\left(\frac{1}{q_{i,CFH}}\right)}$
316.2	0.86	3.13	0.012
261.0	1.08	9.14	0.039
215.4	0.62	5.33	0.065
177.8	0.65	1.56	0.087
146.8	0.72	1.09	0.156
121.2	0.70	1.17	0.201
100.0	0.91	0.30	0.247
82.5	0.86	0.61	0.294
68.1	0.87	0.51	0.288
56.2	0.93	0.26	0.259
46.4	0.61	1.62	0.226
38.3	0.45	2.40	0.216
31.6	0.79	0.83	0.236
26.1	0.81	0.95	0.206
21.5	0.86	0.57	0.207
14.7	0.92	0.19	0.196

cloud contamination risk, there may be a tendency for sonde launches in drier regions, and horizontal smoothing by MLS is not believed to be a factor in this comparison.

[51] In version 1.5 the amplitude of the vertical structure in the tropical "tape recorder" is not correctly measured during some phases of the tape recorder and depends on the location of the seasonal maximum. We speculate that this might also affect the observations of Antarctic stratospheric dehydration. The deviation in tropical tape recorder observations may be caused by some additional smoothing in the stratosphere. The analysis presented here used only the forward model function for version 1.5 to smooth the in situ data to MLS levels under the assumption that the averaging kernel does not contribute to the smoothing significantly. While this is a good assumption in the upper troposphere and lower most stratosphere, it may not be appropriate at pressures less than 68 hPa in the stratosphere. It is possible that the version 1.5 averaging kernel needs to be considered in the smoothing of in situ data to MLS levels.

[52] In version 2.2 the averaging kernel has been used in addition to the forward model smoothing function and although we do not have sufficient data to show this, we may speculate that the problem with the tropical tape recorder does not persist in version 2.2. As the number of reprocessed MLS data for version 2.2 increases, we will, at some time, be able to test whether this is the case, and planned observations over Antarctica will test the accuracy of MLS observing highly detailed dehydration profiles.

[53] The correlation of MLS and CFH observations revealed a decreased sensitivity of MLS in both versions, which is partially compensated by a nonzero offset, leading to the average agreement described above. The sensitivity of MLS version 1.5 is less than that of version 2.2, indicating a significant improvement in version 2.2. The dry bias in MLS observations may be adjusted using the correlation parameters derived here. The correlation is able to explain the average agreement in the stratosphere, the mixing ratio dependent dry bias in the upper troposphere and the reduced lower amplitude of the tropical tape recorder.

[54] Owing to the additional pressure level in version 2.2 at 82 hPa, this version is better suited to capture the tropical tropopause during the boreal winter. The limited data indicate that no average bias is expected at this pressure level. However, owing to smoothing done by the averaging kernel, MLS observations at 82 hPa are not equivalent to in situ observations by CFH at that level (or at the cold point for that matter) and may not resolve the extremely low values observed in some balloon soundings over a shallow vertical range in the Western Pacific region. It is therefore of no surprise that high vertical resolution observations by in situ instruments are more suited for detailed dehydration studies within the TTL.

[55] Studies such as the RH distribution described by *Sherwood et al.* [2006], which use a regional, or global, or annual distribution of water vapor in the upper troposphere, need to consider the decreased sensitivity of MLS observations and a mixing ratio dependent bias. Globally averaged upper tropospheric data will lead to undefined (dry) biases of up to a few tens of percent depending on the distribution of mixing ratio values in these data sets. Stratospheric data appear to be unbiased well within the accuracy level of 10%.

[56] **Acknowledgments.** The authors would like to thank Mike Kurylo and NASA for their partial support under grant NNG05GK53G. SOWER campaigns were supported by the JSPS Grant-in-Aid for Scientific Research (A) 15204043 and 18204041 and the Global Environment Research Program (A-1) of the Ministry of the Environment of Japan. We would like to thank the Hanoi Aero-Meteorological Observatory of the National Hydro-Meteorological Service for their dedicated support to obtain the first stratospheric water vapor observations at Vietnam. For their valuable assistance we would also like to thank the staff members of the Kiribati Meteorological Service at Tarawa; the staff members of the LAPAN offices Biak, Watukosek, and Kototabang; the students of the Universidad Nacional, Costa Rica; the INAMHI offices in San Cristóbal, Galapagos and Quito, Ecuador; students and staff of Howard University for the WAVES 2006 launches. Aura MLS data were prepared at the Jet Propulsion Laboratory, California Institute of Technology, under contract with NASA.

References

Bates, J. J., and D. L. Jackson (2001), Trends in upper-tropospheric humidity, *Geophys. Res. Lett.*, *28*, 1695–1698.

Forster, P. M. de F., and K. P. Shine (2002), Assessing the climate impact of trends in stratospheric water vapor, *Geophys. Res. Lett.*, *29*(6), 1086, doi:10.1029/2001GL013909.

Froidevaux, L., et al. (2006), Early validation analyses of atmospheric profiles from EOS MLS on the Aura satellite, *IEEE Trans. Geosci. Remote Sens.*, *44*, 1106–1121.

Mote, P. W., et al. (1996), An atmospheric tape recorder: the imprint of tropical tropopause temperatures on stratospheric water vapor, *J. Geophys. Res.*, *101*, 3989–4006.

Oltmans, S. J., H. Vömel, D. J. Hofmann, K. H. Rosenlof, and D. Kley (2000), The increase in stratospheric water vapor from balloonborne frost-point hygrometer measurements at Washington D.C., and Boulder, Colorado, *Geophys. Res. Lett.*, *27*, 3453–3456.

Read, W. G., et al. (2007), Aura Microwave Limb Sounder upper tropospheric and lower stratospheric H_2O and RHi validation, *J. Geophys. Res.*, doi:10.1029/2007JD008752, in press.

Rosenlof, K. H., et al. (2001), Stratospheric water vapor increase over the past half-century, *Geophys. Res. Lett.*, *28*, 1195–1198.

Sherwood, S. C., E. R. Kursinski, and W. G. Read (2006), A distribution law for free-tropospheric relative humidity, *J. Clim.*, *19*, 6267–6277.

Vömel, H., D. David, and K. Smith (2007), Accuracy of tropospheric and stratospheric water vapor measurements by the Cryogenic Frost point Hygrometer (CFH): Instrumental details and observations, *J. Geophys. Res.*, *112*, D08305, doi:10.1029/2006JD007224.

Waters, J. W., et al. (2006), The Earth Observing System Microwave Limb Sounder (EOS MLS) on the Aura satellite, *IEEE Trans. Geosci. Remote Sens.*, *44*, 1075–1092.

J. E. Barnes and S. C. Ryan, NOAA Earth Systems Research Laboratory, Hilo, HI 96720, USA.

H. Vömel, Cooperative Institute for Research in Environmental Sciences, University of Colorado, Campus Box 216, Boulder, CO 80309, USA. (holger.voemel@colorado.edu)

R. N. Forno, Atmospheric Physics Laboratory, University of San Andrés, La Paz, Bolivia.

M. Fujiwara and F. Hasebe, Graduate School of Environmental Earth Science, Hokkaido University, Sapporo, Hokkaido 060-0810, Japan.

S. Iwasaki, National Defense Academy, 1-10-20 Hashirimizu, Natsushimacho, Yokosuka, Kanagawa 239-8686, Japan.

R. Kivi and E. Kyrö, Finnish Meteorological Institute, Tahtelantie 62, F-99600 Sodankylä, Finland.

N. Komala and S. Saraspriya, Lembaga Penerbangan dan Antariksa Nasional, Bandung, Indonesia.

T. Leblanc and W. G. Read, Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA 91109, USA.

B. Morel, Laboratoire de Physique de l'Atmosphere, Reunion University, 15, avenue Rene Cassin BP 7151, St. Denis, Reunion F-97715, France.

S.-Y. Ogino, Japan Agency for Marine-Earth Science and Technology, Yokosuka, Kanagawa 237-0061, Japan.

H. Selkirk, Bay Area Environmental Research Institute, 560 Third Street West, Sonoma, CA 95476, USA.

M. Shiotani, Research Institute for Sustainable Humanosphere, Kyoto University, Uji, Kyoto 611-0011, Japan.

J. Valverde Canossa, Universidad Nacional, Heredia, Costa Rica.

D. N. Whiteman, NASA Goddard Research Center, Greenbelt, MD 20771, USA.