Water vapor in Titan’s stratosphere from Cassini/CIRS Far-infrared spectra

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Abstract

Since the first detection of water vapor in Titan's stratosphere by disk-average observations from the Infrared Space Observatory (Coustenis et al. 1998) we report here the successful detection of stratospheric water vapor using the Cassini Composite Infrared Spectrometer (CIRS, Flasar et al. 2004). CIRS senses water emissions in the far infrared spectral region near 50 microns, which we have modeled using two independent radiative transfer codes (NEMESIS, Irwin et al. 2008 and ART, Coustenis et al. 2007, 2010). From the analysis of nadir spectra we have derived a mixing ratio of \(0.14 \pm 0.05\) ppb at an altitude of 97 km, which corresponds to an integrated (from 0 to 600 km) surface normalized column abundance of \((3.7 \pm 1.3) \times 10^{14}\) molecules/cm\(^2\). In the latitude range \(80^\circ S\) to \(30^\circ N\) we see no evidence for latitudinal variations in these abundances within the error bars. Using limb observations, we obtained mixing ratios of \((0.13 \pm 0.04)\) ppb at an altitude of 115 km and \((0.45 \pm 0.15)\) ppb at an altitude of 230 km, confirming that the water abundance has a positive vertical gradient as predicted by photochemical models (e.g. Lara et al. 1996, Wilson and Atreya 2004, Horst et al. 2008); retrieved scaling factors (from \(-0.1\) to \(-0.6\)) to the water profile suggested by these models show that water vapor is present in Titan's stratosphere with less abundance than predicted.

1. Introduction

Water is present in its various forms in many regions of the solar system, from the atmospheres of the inner planets and shadows of lunar craters, to the mantles of icy satellites and beyond to the Kuiper Belt Objects (KBOs) and Oort Cloud Comets. Liquid water is also an essential
ingredient for life on Earth and a potential clue in the search for life or habitability conditions in
the rocks of Mars, the internal ocean of Europa or Titan, and the volcanic vents of Enceladus. On
 Titan, Saturn's largest satellite that hosts a dense nitrogen-dominated atmosphere, water is a trace
species in the stratosphere. However, water plays a significant role since it is one of the sources
of oxygen for the observed active photochemistry on Titan (e.g. Lara et al. 1996, Wilson and

Titan's known oxygen compounds to date are carbon monoxide (CO, ~47 ppm), carbon dioxide
(CO$_2$, ~15 ppb) and water vapor (H$_2$O), where the abundances are quoted for the low-latitude
stratosphere (de Kok et al. 2007a). CO$_2$ was first detected by Voyager 1 (Samuelson et al. 1983),
while CO was first seen by ground-based observations in the near-IR (Lutz et al. 1983). Subsequent observations in the sub-millimeter led to controversy as to whether CO was well-
mixed or not (Hidayat et al. 1998, Gurwell et al. 2004). CO emission lines were later observed
by Cassini/Composite Infrared Spectrometer (CIRS), thus improving the previous abundance
estimate (de Kok et al. 2007a, Teanby et al. 2009). Water was detected at a mixing ratio of 0.4
ppb, assumed to be uniform above the condensation level, by two lines near 40-micron observed
in Short Wavelength Spectrometer (SWS) spectra acquired by the Infrared Space Observatory
(ISO) in 1997 (Coustenis et al. 1998). An early attempt to measure it with Cassini CIRS was
unsuccessful due to poorer signal-to-noise (S/N) ratios in early versions of the calibration
pipeline spectra and fewer spectra were available. Therefore, only an upper limit of 0.9 ppb
could be retrieved (de Kok et al. 2007a). Since then, water emission in CIRS data have been
definitely observed, albeit without deriving any further information on its abundance and
distribution (Bjoraker et al. 2008).
While the presence of these oxygen compounds is now well-established, the question of their origin remains controversial. Early photochemical models assumed that CO originated from episodic outgassing from Titan's interior along with nitrogen (N2) or ammonia (NH3) and methane (CH4), whereas water molecules entered the top of the atmosphere and photochemically produced oxydryl radicals (OH) (Wong et al. 2002; Wilson and Atreya 2004). The combination of OH and CO led to the production of CO2. However, Horst et al. (2008) have recently challenged this model, arguing instead that both CO and CO2 are the result of upper-atmospheric chemistry which occurs between in-falling oxygen species reacting with carbon produced by CH4 photodissociation. In this hypothesis, water enters Titan's atmosphere either in the form of H2O or OH (since the latter is quickly converted to H2O within the atmosphere) together with atomic oxygen (O and O+). These forms of oxygen are deposited at two different altitudes on Titan. The O+ ions are deposited in the upper atmosphere around 1100 km (Hartle et al. 2006a,b) where their interaction with methyl (CH3) radicals leads to the formation of CO. Water is instead deposited at 750 km due to micrometeoritic ablation (English et al. 1996) where it is photolyzed to OH. The latter finally combines with CO to form CO2 and possibly other complex species. Saturn's rings and the icy satellites that surround the giant planets, and also interplanetary dust, are probable sources of the water (oxygen) in Titan's atmosphere and recent results from the Ion and Neutral Mass Spectrometer and magnetometer on board Cassini indicate that the plumes of Enceladus are the dominant source (e.g. Dougherty et al. 2006). Based on Herschel measurements of the Enceladus torus combined with modeling of the fate of the species within the torus, Hartogh et al. (2011) showed that the flux of O / O+ into Titan is consistent with an Enceladus source for the oxygen seen in Titan CO, except for the fact that Enceladus does not seem to provide enough OH/H2O. Characterizing the occurrence of these oxygen species has
important implications for understanding the origin and evolution of Titan and the synthesis of
the complex molecules found in its atmosphere.

In this paper, we analyze the spectra acquired by CIRS in the far infrared spectral region in order
to retrieve the water vapor verticals or spatial distribution in Titan's atmosphere. CIRS has been
acquiring spectra of Titan since the beginning of the Cassini prime mission (July 2004). After
two years of the extended mission (XM), which included the 2009 equinox, in July 2010 Cassini
entered in the Solstice Mission (SM), which will last until 2017. Since the upper limit to H$_2$O
was reported by de Kok et al. (2007) there has been a considerable increasing of the number of
data collected by CIRS during the length of the mission and significant improvements to their
calibration. The increased signal to noise (S/N) ratio not only permits a definitive detection of
H$_2$O from the analysis of CIRS far infrared spectra, but it allows us to constrain its vertical and
latitudinal profile.

2. Selected dataset

CIRS (Flasar et al. 2004) is comprised of three Focal Planes observing in the spectral range 10 -
1400 cm$^{-1}$ with spectral resolutions from 0.5 to 15.5 cm$^{-1}$. Focal Plane 1 detector (FP1) is
characterized by a circular field of view of 3.9 mrad. It records data in the far infrared spectral
range (10 – 600 cm$^{-1}$) with a spectral resolution of 0.5 cm$^{-1}$, allowing us to observe the water
vapor signature, and by modeling, to retrieve its abundance. Water presents its rotational lines in
the CIRS FP1 spectral region up to 400 cm$^{-1}$, with the strongest and most visible lines in the
range positioned between 90 and 260 cm$^{-1}$. We focus here on the range from 150 to 260 cm$^{-1}$ for
the water detection, as this is the range of maximum responsivity of FP1. At lower wavenumbers
the on-board electronics of CIRS create a moving interference spike that can affect the spectrum up to 150 cm\(^{-1}\). Therefore, we exclude wavenumbers shorter of 150 cm\(^{-1}\). We use data from two different types of observations to obtain independent measurements: the far infrared on-disk integrations (FIRNADCMP) and the far infrared limb integrations (FIRLMBINT). Water is a trace species with relatively weak lines and therefore it cannot be observed in an individual spectrum. An average of a few thousand spectra of on-disk observations and a few hundred spectra of limb observations is necessary to achieve sufficient signal-to-noise (Figure 1).

Limb observations have the FP1 focal plane centered around two different altitudes – hereafter limb 1 and 2 – and are therefore used to constrain water vapor abundance in the stratosphere, around 115 and 230 km respectively, well above the tropopause. Since the contribution functions of water for on-disk observations peak around 97 km (Figure 2), the retrieved water vapor abundance derived from these measurements can be compared with the lowest altitude targeted by our limb integrations around 115 km.

For the water detection and retrieval of quantitative information together with possible latitudinal variations, multiple Titan flybys must be utilized to enhance the signal. To date, 35 limb integrations of approximately 1 hour in duration (~60 high-resolution spectra) have been obtained covering latitudes from 87°S to 80°N. The nadir integrations are more numerous (about 92 successfully executed, of typical duration 5 hrs, ~300 spectra) as they occur in a less contested observing time further from the desirable Titan closest-approach period; they have more or less complete spatial coverage of Titan’s latitudes and longitudes with an average footprint size of ~15° great circle arc.

We focus on one season of on-disk observations acquired from December 2004 to December 2008 (northern winter on Titan) in order to reach a compromise between obtaining a large
number of spectra and a sufficiently homogeneous dataset. Inside this time period for on-disk observations acquired from a maximum distance of 300,000 km and with a maximum emission angle of 60°, we selected latitudinal bins (80°S - 45°S), (45°S - 10°S), and (0°- 30°N) centered around three latitudes for which observation-derived temperature profiles were available (see model description in section 3). The numbers of spectra averaged in these latitudinal bins were respectively around 1700, 3800 and 7000 and their average emission angles were respectively 35°, 38° and 34°.

CIRS limb spectra are acquired in much smaller numbers, therefore to reach a sufficient signal-to-noise ratio we consider only one average of about 320 spectra acquired from Dec. 2004 and Sept. 2009, encompassing the entire south and mid-latitudes within the range of 90°S to 20°N; during this time period and at these latitudes data can be considered quite homogeneous as shown in Teanby et al. (2010). We exclude the higher northern latitudes where the stratospheric temperature profile changes significantly. We have also selected data acquired from a Cassini-Titan distance less than 45,000 km in order to limit the size of the projected detector footprint to less than 150 km.

3. Data analysis and model

Across the considered portion of FPI spectral range, Titan’s spectrum is formed by (i) the contribution of thermal emission of the surface and atmospheric layers, (ii) the seven pairs (Anderson and Samuelson 2011) of collision induced absorption (CIA) opacities between the main atmospheric molecules - nitrogen, methane and hydrogen - due to Titan’s dense lower atmosphere, (iii) the photochemical aerosol plus stratospheric condensates, and (iv) the ro-
vibrational emission lines corresponding to the emission of atmospheric species present at the latitudes included in our study: CH₄, CO, H₂O, C₄H₂.

These quantities were used as input to the NEMESIS retrieval code (Irwin et al. 2008) to perform a combination of correlated-k forward model computation (Lacis and Oinas 1991) and retrieval scheme based on the method of optimal estimation (Rodgers 2000) in order to determine Titan's atmospheric opacity, simulate its emerging radiation field, and retrieve Titan's water vapor abundance. This method was successfully applied to model the FP1 spectrum in Cottini et al. (2012) to retrieve surface temperature. A Hamming apodization was used, in keeping with the measured FP1 data.

We solve the radiative transfer equation for 147 spherical atmospheric layers, using as source function the thermal emission of the surface, for which a unit surface emissivity is assumed, and that of the atmospheric layers. The retrieval algorithm then iteratively computes a synthetic spectrum, compares it to the data and after applying a cost function, determines the best estimate for the physical parameters in the model (the stratospheric aerosol profile and any necessary adjustments to the temperature profile and the mole fraction of included atmospheric gas). The cost function includes two components: one that measures the quality of the fit to the spectra (similar to a $\chi^2$ test) and the other that considers the deviation of the retrieved parameters from a set of a priori quantities.

We include in the model the CIA of the atmospheric molecules N₂, CH₄ and H₂, which contribute to the opacity affecting the level of the spectrum continuum; they were calculated according to Borysow and Frommhold (1986a), Borysow and Frommhold (1986b), Borysow and Frommhold (1986c) and Borysow and Frommhold (1987), Borysow (1991) and Borysow and Tang (1993). For the N₂–CH₄ pair, we have used CIA coefficient values increased by 50% as
recommended in Tomasko et al. (2008) and then confirmed in de Kok et al. (2010) based on a comparison between model prediction and the far-infrared continuum data.

We have modeled the haze emission/absorption using the extinction cross sections of the hazes included in de Kok et al. (2007b). Since scattering is negligible at these wavelengths for particles smaller than few microns, we have omitted it from our computations.

We have adopted the atmospheric vertical temperature–pressure profiles retrieved from CIRS data for three latitudes (15°N, 15°S, 58°S) from Anderson and Samuelson (2011) from the surface to 3.3×10⁻⁷ bar, corresponding to an altitude range of 0-600 km. Spectroscopic information of the gas rotational lines in the far-infrared range was extracted from the HITRAN 2004 database (Rothman et al. 2005). For CH₄ we have adopted the revised mole fraction of 1.48% in the stratosphere (Niemann et al. 2010) acquired by the Gas Chromatograph Mass Spectrometer (GCMS) on the Huygens probe in its descent to Titan’s surface. In the stratosphere for H₂ we assumed a uniform volume mixing ratio of 0.1% (Courtin et al. 2008). The geometry of the observations was also included in the computations.

An accurate model of the emerging radiation field is required to successfully reproduce CIRS spectra and measure water abundance. The FP1 FOV is circular and has a sensitivity that decreases from the center to the edge (0-3.9 mrad) where it drops to zero. For on-disk spectra the homogeneity of the field of view usually permits simple modeling with a single ray calculated for the detector center. For limb spectra, we have to take into account the rapid decrease in atmospheric density with the height and the variations of temperature and gas volume mixing ratio profile with altitude. In such a case, the FOV is not assumed to be uniform and a multiple ray model is required to fit the data. We modeled the FOV using the minimum number of rays for which the synthetic spectrum and the water retrieval computation results became stable; this
corresponds to 9 rays with a step in altitude of 25 km. We also recomputed some of the results using 39 rays (step of 5 km) in order to show a smoother limb contribution function. Spectral radiance measured by the FPI detector is modeled by a convolution of the emerging radiance at each point in the FOV (as described in Nixon et al. 2009a and in Teanby and Irwin 2007), weighted by a Gaussian response function for CIRS FPI detector. This response function – the actual beam profile – was determined for CIRS FPI (Flasar et al. 2004) during Cassini flybys of Jupiter which, due to the large distance of observation, was considered as a point source. As 95% of the integrated response is contained in a radius of 1.95 mrad from the FOV center, the detector observes an actual maximum altitude range of about 70 km.

Line-by-line independent calculations to simulate the same on-disk FPI selections were also made using the Atmospheric Radiative Transfer (ART) code used by Coustenis and co-authors in previous papers and more recently applied to CIRS data in Coustenis et al. (2010). The code uses the most recent aerosol extinction dependence inferred from Vinatier et al. (2012) and temperature profiles derived by fitting the $v_4$ methane band at 1304 cm$^{-1}$ in FP4 averages taken at similar conditions as the FPI spectra. The spectroscopic parameters for all the observed molecules and isotopes are from GEISA 2009 (Jacquinet et al. 2011) and HITRAN 2008 (Rothmann et al. 2009). The results from these two different codes are quite similar and largely within error bars.

In Figure 1 (upper panel) the on-disk and two limb observation averages are shown together with their fit. Figure 1 (lower panel) shows only the spectral range used for the water line analysis.
4. Results

We have retrieved water vapor abundance from both on-disk and limb data assuming different water vapor mole fraction dependences on altitude: a constant water mixing ratio profile and three vertical distribution profiles with the mole fraction increasing with height as predicted by recent photochemical models: a) Horst et al. (2008) and b) Wilson and Atreya (2004) and c) Lara et al. (1996), a previous model adopted in Coustenis et al. (1998) for the first water detection by ISO. The constant profile along the atmosphere (a priori assumed to be 0.1 ppb) at the altitude where the water vapor freezes is forced to decrease to zero following the saturation law. Using the saturation vapor pressure equation of water over ice of Murphy and Koop (2005) and assuming the temperature profile retrieved at 15° N we find a condensation altitude of about 93 km. We have also computed the contribution functions - normalized inversion kernels - showing the sensitivity of each atmospheric layer to a variation of the H₂O mixing ratio. These contribution functions were computed for each profile and for all of the most intense water lines in order to provide an altitude range of validity of the retrieved values (Figure 2 and 3). Figure 2 shows the contribution functions for on-disk observations computed at four different wave numbers; at 254 cm⁻¹ (one of the two lines used for the ISO water retrieval) the upper shoulder of the contribution function is wider and sensitive to higher altitudes compared to the other wave numbers used for the water retrieval in this work. In our case the fit of the 254 cm⁻¹ line improves when using a profile increasing with altitude rather than a constant profile. In Figures 3 we show only the contribution functions computed at wavenumber 202.75 cm⁻¹, where the most intense water line in the far infrared occurs (discounting the line at 150.5 cm⁻¹ due to enhanced electronic noise).
For water retrievals obtained using a constant water profile we show the retrieved mixing ratio values at the altitude where the water functional derivative peaks for the assumed profile (Table 1). We associate to this altitude an error equal to the FWHM of the contribution function for the corresponding water profile. We also retrieve a scaling factor to the water profile associated with each of the photochemical model considered in this work (Table 1).

4.1 On – disk water retrieval

To measure the water abundance from the on-disk average (0° – 30° N) data we first use a constant water profile. We retrieve a volume mixing ratio of (0.14 ± 0.05) ppb for altitudes ranging between ~93 and 130 km, where the contribution function peaks at around 97 km. This value corresponds to a surface-normalized H₂O total column density on the order of (3.7±1.3) × 10^{14} molecules/cm².

The sensitivity of the retrieved results at the altitude where the water vapor freezes (this strongly depends on the assumed temperature profile and the water mixing ratio) is the most important source of error for the on-disk observations since condensation occurs in the altitude region where the spectrum is sensitive to water vapor. Small variations of the temperature profile in the stratosphere, random noise, and a small dependence in altitude sensitivity with wavenumber are other included sources of error.

We also fit the water lines for the three other water vertical distribution profiles (Horst et al. 2008, Wilson and Atreya 2004 and Lara et al. 1996) and obtained the necessary scale factors to fit the data, which are shown in Table 1. These values, ranging between 0.11 and 0.63 times the considered profiles, show the retrieved water mole fraction to be less than predicted from these
previous models.

We have analyzed two additional latitudinal bins from 45°S to 10°S and from 80°S to 45°S, centered on the corresponding temperature profiles retrieved from CIRS, at 15°S and 58°S respectively. The observed water mixing ratio indicates the absence of any significant latitudinal variations within the data uncertainties in the considered latitude range. It should be stressed in fact that in this work we did not analyze the water stratospheric content at high northern latitudes that were experiencing winter during this time period. To model spectra at these latitudes for a large average is particularly complex since the stratospheric temperature change quickly with latitude; besides, a temperature profile for high northern latitudes is not yet available for the stratospheric region where water saturates and to which CIRS on-disk spectra are more sensitive. The results obtained simultaneously for on-disk data only using the independent line-by-line ART code simulations of the same FPI selections confirm well within the error bars the retrieved water vapor values reported above.

### 4.2 Limb water retrieval

The measurement of water vapor obtained by modeling the limb 1 spectrum under the assumption of a constant mixing ratio profile is equal to \((0.13 \pm 0.04)\) ppb. This value is relative to an altitude around 115 km according to the position of the peak of the corresponding contribution function.

Modeling the limb 2 spectrum we retrieved a water mixing ratio of \((0.45 \pm 0.15)\) ppb using a constant water profile. The radiance mostly originates from a region centered at 230 km (190-275 km at half maximum of sensitivity). These values indicate an increase of the water mole
fraction with altitude in the stratosphere from 115 km to 230 km of about 3 times. Figure 1 depicts a comparison of the strength of the water lines in the spectrum targeted at limb 2 altitude to the ones in the limb 1 spectrum. Water lines have decreased proportionately less in radiance than the methane lines, indicating qualitatively an increase of water with altitude, as methane is uniformly mixed in this range: water lines around 115 km are only 1-2 times stronger than the corresponding lines at 230 km while methane lines are 3-5 times stronger.

The scaling factors to the model water profiles obtained from the two types of limb retrievals are shown in Table 1 and again illustrate the smaller amount of stratospheric water vapor detected by CIRS in respect to the one predicted by models.

5. Conclusions

In this work we modeled CIRS data with a constant-with-height water vapor profile and assigned the retrieved mixing ratio to the altitude where the contribution function peaks.

By combined on-disk and limb observations we are able to constrain the vertical profile of water in the region of the stratosphere — from 12 mbar to $10^{-3}$ mbar, corresponding to altitudes between 93 and 280 km (considering the widths of the contribution functions).

In Figure 4 we summarize our water vapor retrieved values and infer vertical profiles showing also the models of Horst et al. (2008), Wilson and Atreya (2004) and Lara et al. (1996) for comparison.

The measurement of the stratospheric vertical profile of water adds useful constraints to the photochemical models of Titan’s atmosphere. The increase of the water mixing ratio with altitude is in agreement with an external source of oxygen and lower altitude a sink due to
photolysis and condensation. The abundance instead seems to be less (from ~0.1 to ~0.6) than predicted from the models considered in this work (see Table 1). We also observe that since the scaling factors to the photochemical models a), b) and c) in Table 1 are more or less different for the two limb altitudes, it implies that these models might have a slope for H₂O not quite consistent with CIRS data. However, due to the quick variation of the water vapor mole fraction with altitude in the atmospheric region where water freezes and where CIRS is actually observing (on-disk and limb 1 spectra), we should be careful in the evaluation of the water profile slope and include the measurements errors to estimate it.

ISO retrieved a water vapor abundance of 0.4 ppb assuming a constant mole fraction above the condensation level (Coustenis et al. 1998); we associate to this value the same relative error bars as those derived by ISO for the scaled Lara et al. (1996) profile: the data were fit with a scaling factor to the water profile of (0.4 +0.3 −0.2). From CIRS on-disk observations we retrieved a volume mixing ratio of (0.14 ± 0.05) ppb around 97 km for latitudes (0° − 30° N) which is only marginally consistent with the ISO determination of 0.4 ppb, considering the error bars and the different atmospheric models and geometry of observation of the two instruments.

We also fit the CIRS data assuming the water profile from Lara et al. (1996) multiplied for a scaling factor of 0.48 ± 0.07 (corresponding to a water column density of 3.8 ± 1.0 ×10¹⁴ molecules cm⁻²). This result agrees with the scaling factor of (0.4 +0.3 −0.2) retrieved from ISO observations by Coustenis et al. (1998) but this agreement may be fortuitous as, in the ISO geometry (integrated flux), most of the emission originates from above 300 km with the contribution functions peaking around 400 km (Fig. 3 of Coustenis et al. 1998). This is due to a strong emission from the limb occurring at high altitudes when 0.4 times the Lara et al. (1996) profile is used. In contrast, using this H₂O profile, the contribution functions for the CIRS nadir
selection cover the range 95-145 km (at half maximum).

In previous photochemical models, as in Lara et al. (1996), in order to include external sources of CO, it was postulated that CO could be produced through a chemical reaction between OH (available from H₂O influx into the upper atmosphere) and CH₃. It was found instead by Wong et al. (2002) that this reaction produces H₂O and not CO as previously assumed.

Hence an influx of H₂O or OH does not produce any significant abundance of CO and therefore CO₂ can be produced by an H₂O influx only with CO already present (OH + CO -> CO₂ + H). For this reason these models were unable to reproduce the observed CO abundance and were substituted by other models that suggest the existence of primordial CO in the atmosphere (Wilson and Atreya 2004) or consider for CO a solely external origin but this requires an influx of O⁺ rather than H₂O or OH.

In the pre-Cassini model of Wilson and Atreya (2004) water is photolyzed to OH, which combines with CO to form CO₂ and other complex species. In this model, CO is assumed to be primordial on Titan and the water abundance profile derives from the amount necessary to form the observed CO₂. This assumption was challenged by the Horst et al. (2008) model, in which oxygen species are assumed to arrive from outside the moon and form carbon monoxide as well as carbon dioxide in the atmosphere. The values of the input fluxes of O and OH were adjusted to reproduce the observed abundances of CO and CO₂. In the Horst et al. (2008) model water profiles were produced for six different values (from K = 100 cm² s⁻¹ to K = 1000 cm² s⁻¹) of the eddy coefficient in the lower atmosphere, since the stratospheric abundances of photochemically produced species are highly dependent on this parameter. As shown in Fig. 4, the water abundance retrieved in our study is best fit by the water profile with the lowest eddy diffusion coefficient value considered in their model (K = 100 cm² s⁻¹), which is far from the value (K =
they identified as better reproducing CIRS observations of hydrocarbon species and adopted in this work for comparison with our retrievals. Our results show that even the Horst model with K=100 cm$^2$ s$^{-1}$ still has too much water.

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Figure 1. Upper panel: In black, the average of CIRS far-IR on-disk observations is plotted...
(~7,000 spectra acquired from Dec. 2004 to Dec. 2008 in the latitudinal range of (0° – 30° N),
limb observations centered around 115 and 230 km (respectively ~320 and ~280 spectra acquired
from Dec. 2004 to Sep. 2009 in the latitudinal range of 90°S– 20°S) and their fit (in green, blue,
red respectively) assuming a constant water mole fraction above the condensation altitude.
Lower panel: the retrieval spectral range is shown with the main water lines indicated by vertical
dotted lines.

Figure 2. Contribution functions of the different atmospheric layers to the water vapor line
emission computed for four wave numbers. In solid line is shown also a temperature profile.
Figure 3. Contribution functions of water vapor line emission and temperature profile. In Figures 3 a,b,c,d we show the contribution functions computed at 202.75 cm\(^{-1}\) for different water profile models: (a) a constant water vertical profile; (b) profile from Horst et al. (2008); (c) profile from Wilson and Atreya (2004) and (d) profile from Lara et al. (1996).
Figure 4. Water vapor mole fraction retrieved from CIRS on-disk and two limb observations assuming a water profile constant with altitude over the condensation level. Water profiles from photochemistry models are also shown for comparison: Horst et al. (2008) water vapor profile derived assuming two different eddy diffusion coefficients - 100 cm² s⁻¹ and 400 cm² s⁻¹ (the second being the one recommended in their model; dotted curve plus solid curve), the profile in Wilson and Atreya (2004; dashed curve) and in Lara et al. (1996; dot-dash curve).
Table 1. Retrieval of water abundance from on-disk observations (0°– 30°) N and from limb observations targeted at two altitudes assuming a constant water mixing ratio over the condensation region (second row). The altitude associated with the retrieval and its error is given by respectively the peak and the FWHM of the contribution function relative to the assumed water profile. In the subsequent rows are shown the retrieved scaling factor to water profiles given in the models of Horst et al. (2008), which assumes $K = 400 \text{ cm}^2 \text{s}^{-1}$, Wilson and Atreya (2004) and Lara et al. (1996).