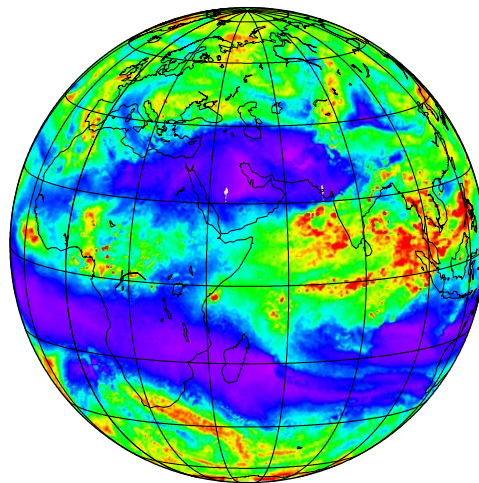


Remote Sensing of Atmospheric Composition for Climate Applications

dem Fachbereich Physik/Elektrotechnik
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1 Introduction

Global change is a threat to human society, not just since ‘The day after Tomorrow’ has hit the cinemas in 2004. Indeed, how naive does one have to be, to think that changing atmospheric composition on a global scale would have no consequences? Since pre-industrial times, the atmospheric carbon dioxide concentration has increased by approximately 30%, according to the third assessment report of the Intergovernmental Panel on Climate Change [48]. The carbon dioxide rise has concerned atmospheric scientists since the 1960s [52, 96] and has become one of the big topics of atmospheric science, if not the biggest.

A lot of progress has been made in understanding the climate system, and also in our predictive capabilities. Nevertheless there are still large uncertainties in the predictions, arising to a significant part from deficiencies in the global observing system, most prominently the ones concerning water vapor and clouds. One should think that the concentration of water vapor, a gas, is easy to measure, but it is not so. The reasons for the difficulty are largely connected to water vapor’s very high variability on small spatial scales, and to its correlation with the occurrence of clouds, which contaminate remote measurements. And yet, water vapor is the most important greenhouse gas. The figure on the cover page shows the high variability of the water vapor distribution. How it was generated is explained in Section 3.3.

Chapter 2 and paper [H 2] deal with the role of water vapor in the earth’s radiation budget. Chapter 3 and papers [H 1], [H 4], and [H 6] deal with measurements of water vapor in the upper troposphere, where it is particularly important for the radiation budget, but poorly characterized. Chapter 4 and papers [H 5] and [H 8] deal with the development of radiative transfer algorithms, including the effects of clouds, which are difficult to handle because the cloud particles scatter the radiation in all directions, prohibiting a straightforward integration of the radiative transfer equation. This work is important, because without accurate radiative transfer algorithms remotely sensed data are useless.

There is a limit to what can be learned about the atmospheric state from present-day data. We need better remote sensing techniques and new instrument designs that can provide measurements with high vertical resolution and high absolute accuracy. Chapter 5 and papers [H 3], [H 7], [H 9], and [H 10] deal with this issue, focusing on the limb sounding technique. Chapter 6 contains a summary and outlook. The selected publications [H 1] to [H 10] can be found after page 54, or on the Internet at <http://www.sat.uni-bremen.de/publications/>, where also many of the other cited publications can be found.

2 Outgoing Longwave Radiation

The planet Earth is in radiative equilibrium with its surroundings. It receives energy in the form of shortwave radiation from the sun and loses energy in the form of longwave radiation to space (Figure 2.1). These two radiation streams can be represented approximately by blackbody radiation of 6000 K for the solar shortwave radiation and 290 K for the terrestrial longwave radiation. The balance between the incoming shortwave radiation and the outgoing longwave radiation (OLR) determines the temperature in the atmosphere and on the earth's surface [75, 43, 44].

The OLR originates partly from the surface but to a significant part from higher levels of the atmosphere. Because of the lower temperature at these levels, the OLR is reduced compared to a hypothetical earth without atmosphere. Figure 2.2 shows a high resolution radiative transfer model simulation of clear-sky monochromatic radiance at the top of the atmosphere, which illustrates this. Besides the calculated spectrum, it shows Planck curves for different temperatures. An integration over all frequencies and directions yields the OLR. The reduction of OLR compared to a hypothetical earth without atmosphere is of course nothing else than the atmospheric “greenhouse” effect. From the known incoming solar shortwave radiation we can easily infer the global average OLR to be close to 240 Wm^{-2} because the incoming and outgoing radiation fluxes must balance [43]. However, there is considerable variability for different latitudes and weather conditions, so that local OLR values vary between about 160 Wm^{-2} and 320 Wm^{-2} .

Figure 2.3 displays a monthly mean map of measured OLR data from the Earth Radiation Budget Experiment (ERBE) on the NOAA 9 satellite of the US National Oceanic and Atmospheric Administration. The figure shows the large regional variability in OLR. It has its lowest values near the poles and its highest values in the subtropics, where the atmosphere is very dry and warm due to large scale subsidence in the descending branch of the Hadley circulation. What the figure does not show is that OLR also has a large day-to-day variability.

Paper [H 2] focuses on the clear-sky OLR, i.e., the OLR in the absence of clouds. This case is simpler to deal with than the one with clouds. A high frequency resolution radiative transfer model was used to simulate the clear-sky OLR flux, Jacobians for the clear-sky monochromatic zenith radiance, and monochromatic clear-sky cooling rates. Details of the model are discussed in Section 4.1 and in paper [H 5]. Compared to the simulations by Clough and Iacomo [15] (CLI) the model used updated spectroscopic data from the current version of HITRAN [74] and updated continuum parameterizations provided by Mlawer et al. [65].

2 Outgoing Longwave Radiation

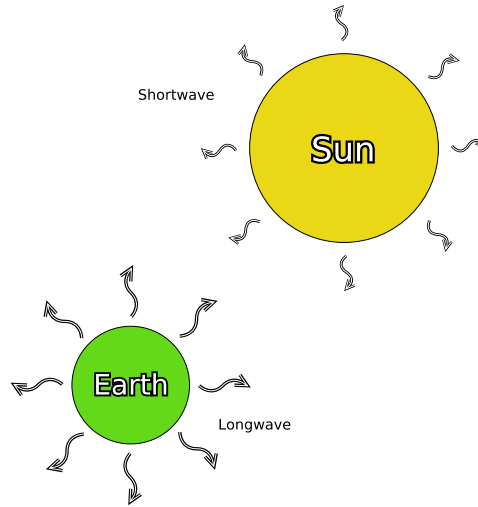


Figure 2.1: The sun-earth system. The earth receives shortwave radiation from the sun and emits longwave radiation to space.

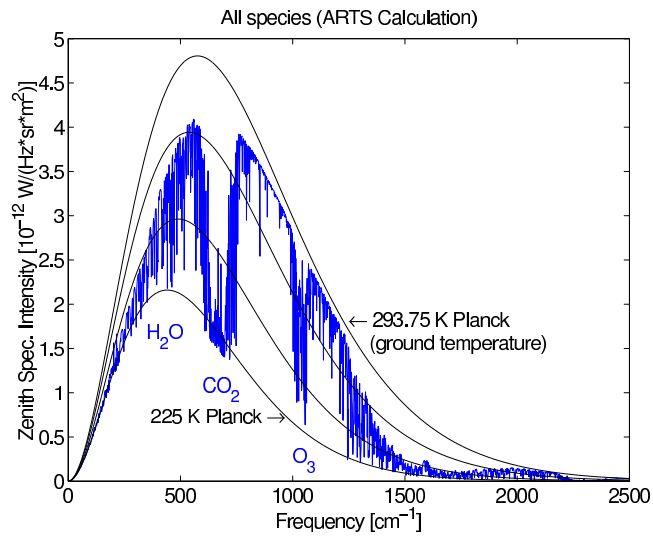


Figure 2.2: A radiative transfer model simulation of the top of the atmosphere zenith monochromatic radiance for a midlatitude summer atmosphere. Smooth solid lines indicate Planck curves for different temperatures: 225 K, 250 K, 275 K, and 293.75 K. The latter was the assumed surface temperature. The calculated quantity has to be integrated over frequency and direction to obtain total OLR.

2 Outgoing Longwave Radiation

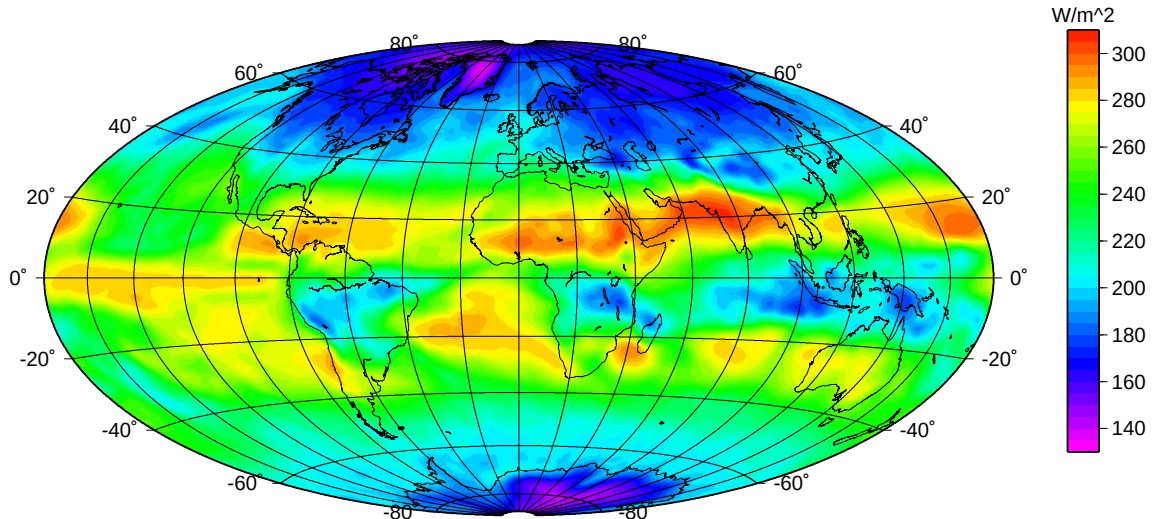


Figure 2.3: Outgoing Longwave Radiation (OLR) measured by the ERBE instrument on NOAA 9. This is a monthly mean for February 1985. The data were obtained from the NASA Langley Research Center, the unit is W m^{-2} .

Compared to the calculations by CLI, our OLR at the top of the atmosphere is approximately 4.1% smaller for all investigated scenarios. This is partly due to the fact that CLI assumed the top of the atmosphere to be at 65 km altitude and make a plane parallel approximation, whereas we assume the top of the atmosphere to be at 95 km altitude and do not make a plane parallel approximation. The combined effect of these two differences in setup explains a difference of approximately 2.5%. The remaining 1.7% difference is probably due to the differences in spectroscopy and continuum models, since all other likely explanations were excluded. Both the CLI calculations and our calculations are in reasonable agreement with CERES/TRMM data. For our calculations this is shown by Figure 2.4. The CLI results lie on the upper end of the CERES clear-sky OLR variability range, whereas our results lie on the lower end.

Figure 2.5 summarizes the OLR sensitivity to some large scale perturbations for a tropical atmosphere (TRO, top) and a subarctic winter atmosphere (SAW, bottom). In the TRO case a 20% humidity increase has a larger impact on OLR than a CO_2 doubling, in the SAW case the CO_2 doubling has the slightly larger impact. This is consistent with the findings of Brindley and Harries [9], who stated that humidity increases of 12% and 25%, for the TRO and SAW cases, respectively, have an OLR impact equivalent to doubling CO_2 . However, OLR is also strongly sensitive to changes in temperature because of the positive temperature dependence of the Planck function. The ‘T+1 K’ bar in Figure 2.5 shows that for the TRO case a 1 K temperature increment throughout the atmosphere produces roughly the same effect on OLR as a 20% humidity decrease. For the SAW case the +1 K temperature effect is even roughly twice the -20% humidity effect. Because of the partial cancellation of OLR changes due to a simultaneous increase of temperature and absolute humidity, the OLR sensitivity to a 1 K temperature increase under fixed relative humidity is much smaller than to a pure temperature increase, which is the so-called water

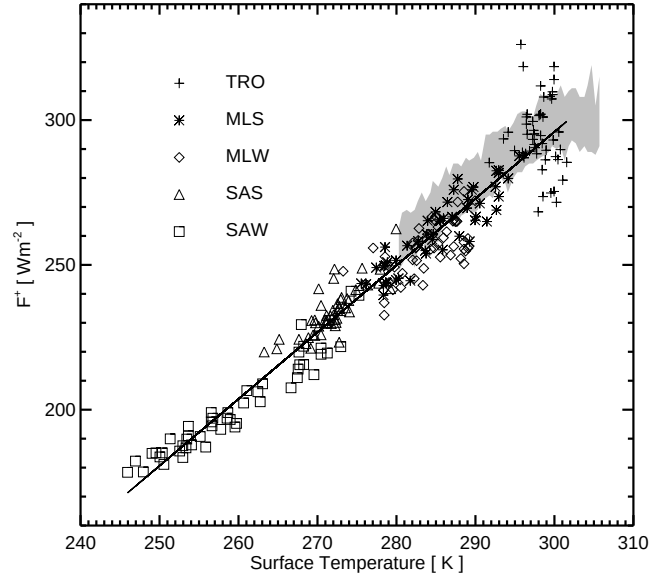


Figure 2.4: Calculated OLR as a function of surface temperature for five different radiosonde classes: tropical (TRO), midlatitude summer (MLS), midlatitude winter (MLW), subarctic summer (SAS), and subarctic winter (SAW). The solid line is a linear fit to the data from all five classes. The grey shaded area shows the one standard deviation variability of CERES/TRMM data taken from Inamdar et al. [49]. Unfortunately, these data are only available for surface temperatures above 280 K.

2 Outgoing Longwave Radiation

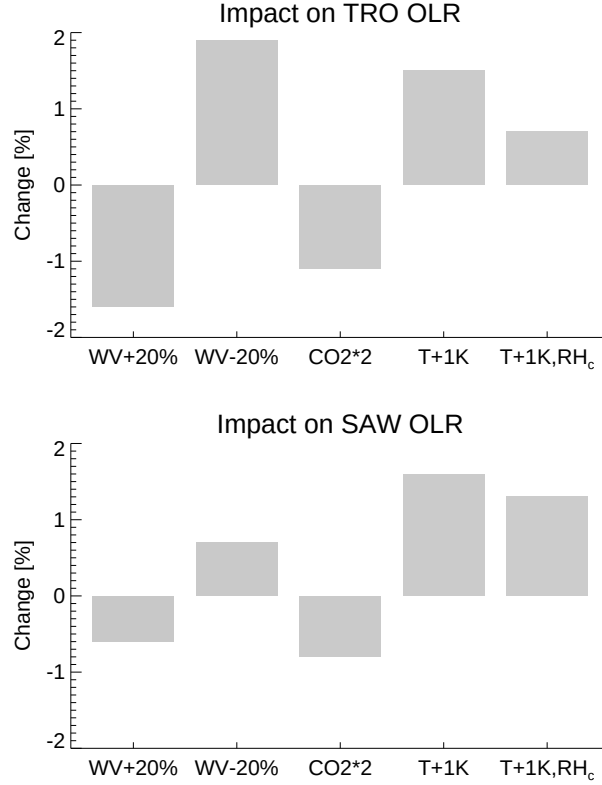


Figure 2.5: OLR impact of a 20% water vapor increase or decrease, CO₂ doubling, a 1 K increase in temperature with fixed absolute humidity, and a 1 K increase in temperature with fixed relative humidity. Results are shown for the TRO scenario (top) and the SAW scenario (bottom). Shown is the relative deviation in OLR from the reference case.

vapor feedback (see Held and Soden [46] for a more detailed discussion).

The global variability in clear-sky OLR is approximately 33 W m^{-2} , estimated by the standard deviation of all OLR values calculated from a global set of radiosondes. This large variability can be explained to a large extent by variations in the effective tropopause temperature, or in the surface temperature as a proxy. That component of the variability can be removed by making a linear fit of OLR versus surface temperature. The remaining variability is approximately 8.5 W m^{-2} . A significant part of this remaining variability can be explained by variations in the total tropospheric humidity (TTH). Making a linear fit of the temperature-independent OLR variations versus the logarithm of TTH reduces the remaining variability to only approximately 3 W m^{-2} .

This remaining variability must be due to vertical structure. It was shown in paper [H 2] that humidity structures on a vertical scale smaller than 4 km contribute a variability of approximately 1 W m^{-2} , but no significant bias if the smoothing is done in the right way. The right way to smooth is in relative humidity. If the smoothing is done for example in volume mixing ratio (VMR) it leads to a substantial bias. This result means that measure-

ments from sensors with coarse vertical resolution may be used to predict OLR with the correct mean values, but will not be able to fully reproduce the variability due to vertical structure, as almost half of that can come from structures on a scale smaller than 4 km. This calls for sensors that can sound the troposphere, including the upper troposphere, with good vertical resolution. Such sensors could use the radio occultation technique [31], as the proposed ACE+ instrument [7], or the passive microwave limb sounding technique, as the MLS instrument that is planned to be launched with the Aura satellite in summer 2004 [95]. Instruments of this type are discussed in Chapter 5 and in papers [H 3], [H 7], [H 9], and [H 10].

Another big issue is the absolute accuracy of the sensor for the humidity measurement, because there are large discrepancies between the upper tropospheric humidity (UTH) measured by the different sensors currently in use. For example, according to Soden et al. [80] the relative difference in UTH between Vaisala radiosondes and the High Resolution Infrared Sounder (HIRS) satellite sensor is approximately 40%. The 8.5 W m^{-2} variability due to humidity changes may seem small compared to the large temperature variability, however, it still significantly exceeds the effect of double CO_2 , which is only 1.6 to 3.0 W m^{-2} , depending on the scenario. If we take the $\approx 40\%$ discrepancy in the humidity datasets found by Soden and coworkers as the uncertainty in UTH, it may introduce a bias in clear-sky OLR of approximately 3.8 W m^{-2} for the MLS scenario, a number derived by sensitivity calculations similar to those presented in Figure 2.5. This shows that, although the radiative effects of clouds and aerosols currently receive more attention, we should not forget the uncertainty in OLR associated with our limited knowledge of UTH.

3 Upper Tropospheric Humidity

The last chapter showed that upper tropospheric humidity (UTH) has an important influence on OLR. Other publications also confirm that UTH is a crucial parameter for meteorology and climate research [83, 6, 78, 44]. The purpose of this chapter is to show what can be learned about UTH from existing datasets. Papers [H 1], [H 4], and [H 6] deal with this issue.

3.1 Upper Tropospheric Humidity Measured by Radiosonde and Satellite

There are two global and continuous datasets for UTH, one from polar orbiting meteorological sensors, the other from synoptic meteorological radiosondes.

The satellite data come from two series of satellites, one from the National Oceanic and Atmospheric Administration (NOAA), the other from the Defense Meteorological Satellite Program (DMSP). UTH is measured by infrared instruments at $6.7\text{ }\mu\text{m}$ [79, 80], and by microwave instruments at 183 GHz [25]. The activities in Bremen focus on the microwave measurements. They have the advantage of being much less affected by clouds than the infrared measurements. These data are used operationally by the numerical weather prediction (NWP) community, which have developed NWP model based tools to monitor satellite radiometer performance [2], valuable for identifying inter-satellite differences and changes over time, but problematic if one is interested in absolute UTH. Operational products are generated by NOAA/NESDIS who also routinely perform radiosonde inter-comparison activities [69].

Microwave humidity data exist from SSM-T2 (Special Sensor Microwave Water Vapor Sounder) on the DMSP satellites, from AMSU-B (Advanced Microwave Sounding Unit) on the NOAA satellites, and from HSB (Humidity Sounder for Brazil) on the Aqua satellite. Information on the available data is summarized in Table 3.1. The study described in paper [H 4] focuses on NOAA-15 and NOAA-16 for the years 2001–2002.

Details on the AMSU-B instrument can be found in an article by Saunders et al. [76]. It is a cross-track scanning microwave sensor with channels at 89.0, 150.0, 183.31 ± 1.00 , 183.31 ± 3.00 , and 183.31 ± 7.00 GHz. These channels are called Channel 16 to 20 of the overall AMSU instruments, Channels 1 to 15 belong to AMSU-A. The instrument has a swath width of approximately 2300 km, which is sampled at 90 scan positions. The satellite viewing angle for the innermost scan positions is 0.55° from nadir, for the

Table 3.1: A summary of currently operating microwave satellite humidity sensors.

Platform Name	Instrument Name	Launch
DMSP F-13	SSM-T2	March 1995
DMSP F-14	SSM-T2	April 1997
NOAA-15 (NOAA-K)	AMSU-B	May 1998
NOAA-16 (NOAA-L)	AMSU-B	September 2000
NOAA-17 (NOAA-M)	AMSU-B	June 2002
Aqua	HSB	May 2002

AMSU-20 ($183.31 \pm 7\text{GHz}$); May 15, 2001 16:14

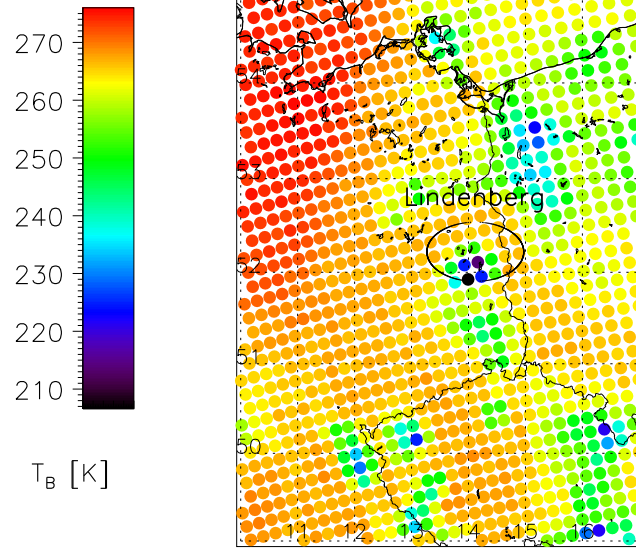


Figure 3.1: An AMSU overpass over station Lindenberg. The circle drawn around the station has a radius of 50 km.

outermost scan positions it is 48.95° from nadir. This corresponds to surface incidence angles of 0.62° and 58.5° from nadir, respectively. The footprint size is $20 \times 16 \text{ km}^2$ for the innermost scan positions, but increases to $64 \times 52 \text{ km}^2$ for the outermost positions. Figure 3.1 shows an example of AMSU data from Channel 20. This is an overpass over the radiosonde station Lindenberg.

Figure 3.2 shows the frequency positions of the AMSU-B humidity channels 18 to 20 relative to the atmospheric zenith opacity spectrum. Channels 16 and 17 (not shown) are surface channels. Channels 18 to 20 sample the free troposphere. The sampling altitude for each channel can best be seen from the Jacobian [72], i.e., the derivative of the channel radiance with respect to a change in local atmospheric humidity. Examples are displayed in Figure 3.3. Note that the Jacobians move significantly with changing atmospheric state. In particular, Channel 20, which normally cannot see the surface, can see the surface under very dry conditions.

The synoptic radiosonde data go back to the nineteen-forties [23]. Records are kept at

3 Upper Tropospheric Humidity

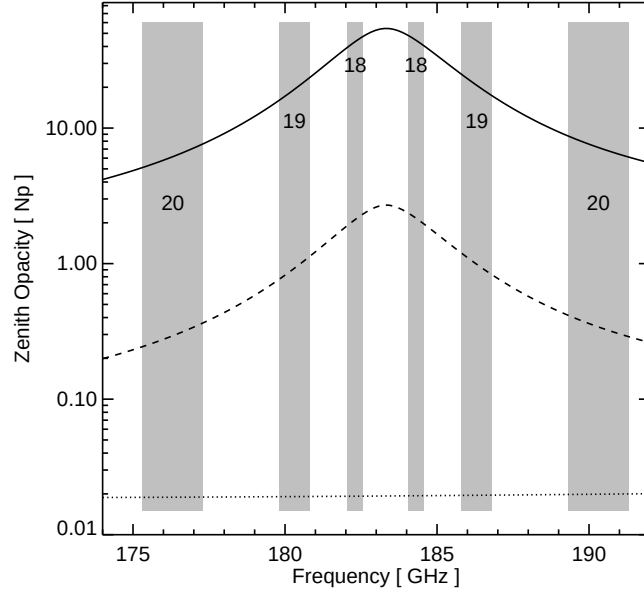


Figure 3.2: Atmospheric zenith opacity as a function of frequency. AMSU-B channel positions for the three humidity channels are indicated. The dashed and solid opacity curve corresponds to the driest and wettest Lindenberg radiosonde profile, respectively. The dotted line is the dry air opacity. The grey shaded areas indicate the bandwidth of the channels.

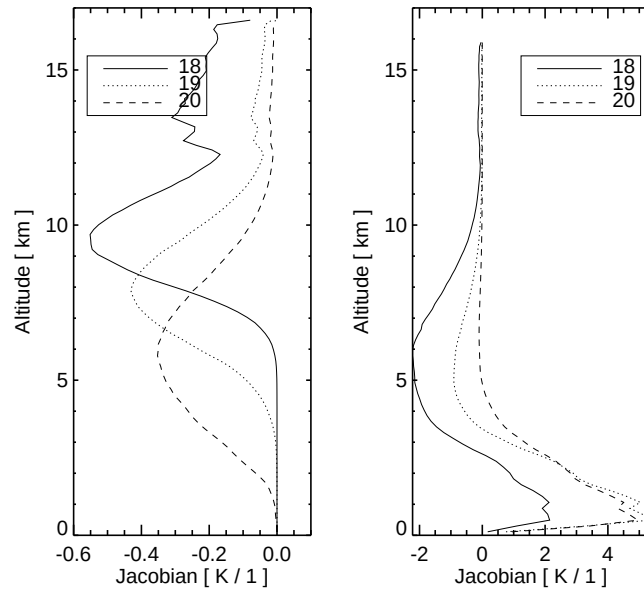


Figure 3.3: AMSU-B humidity Jacobians in fractional units. These units are such that the values correspond to the change in brightness temperature for a doubling of the mixing ratio at one grid point. Left: for wettest Lindenberg radiosonde profile, right: for driest profile.

3 Upper Tropospheric Humidity

meteorological agencies. Data can be obtained for example from the British Atmospheric Data Centre (BADC). There are about 850 stations in the data record at the BADC, but only about 250 stations have at least 10 launches per month reaching 100 hPa. Ideally, each station should launch a radiosonde four times a day at the synoptic observation times 0, 6, 12, and 18 UTC. However, many stations only launch sondes irregularly. The quality of the data from the different stations is also believed to vary considerably.

For the case study described in paper [H 4] it was decided to focus on the radiosonde data from one station. This has the advantages that the properties and quality of the data are well understood, and that high vertical resolution data, which are not in the BADC archive, can be used. The Meteorological Observatory Lindenberg (MOL), located at 52° 22' N, 14° 12' E, is one of the reference stations of the DWD. The radiosonde record there goes back to 1905. Recently, great effort has been made to improve the calibration of Humicap humidity sensors (Leiterer et al. [59]), together with the manufacturer Vaisala.

The basic idea of the study [H 4] was to compare satellite and radiosonde data. However, the satellite measures radiances, not humidity directly. While obtaining radiances from given temperature and humidity profiles is straightforward, obtaining humidity concentrations from radiances is complicated and requires additional assumptions (a classical inverse problem [72]). To avoid having to deal with the inverse problem, the comparison can be done in radiance space rather than state space: A radiative transfer (RT) model can be used to simulate satellite measurements from the radiosonde data, which can then be compared to the real satellite measurements. This approach has already been used for infrared data, for example by Soden and Lanzante [79], but not for microwave data. The RT model employed was the same as for study [H 2], described in Chapter 2. See Section 4.1 and paper [H 5] for further details.

A robust method to compare radiosonde humidity data to AMSU data was developed, which is planned to be used for future global studies. The new method has some unique features: Firstly, the comparison is done for a target area, allowing an estimation of the atmospheric variability. (The target area is a 50 km radius circle, as indicated in Figure 3.1.) Secondly, displacement and cloud filters are applied. Thirdly, a complete and consistent error model is used.

The method was validated by a detailed case study, using the high quality Lindenberg radiosonde data and the NOAA-15 and 16 satellite data for the time period from 2001 to 2002. The overall agreement is very good, with radiance biases below 1.5 K and standard deviations below 2 K. The main source of ‘noise’ in the comparison is atmospheric inhomogeneity on the 10 km scale. The study confirmed that low vertical resolution data, as found in operational archives, are sufficient to accurately predict AMSU radiances. However, it also demonstrated that corrections applied in Lindenberg to the standard Vaisala data processing make a significant difference, particularly in the upper troposphere.

Overall, the AMSU data are in very good agreement with the radiosonde data, with the notable exception of a slope in AMSU Channel 18, as shown by Figure 3.4. By re-processing with perturbed RT model parameters, RT model error was ruled out as a possible explanation for the slope, leaving only AMSU data and radiosonde data. Of these two, the

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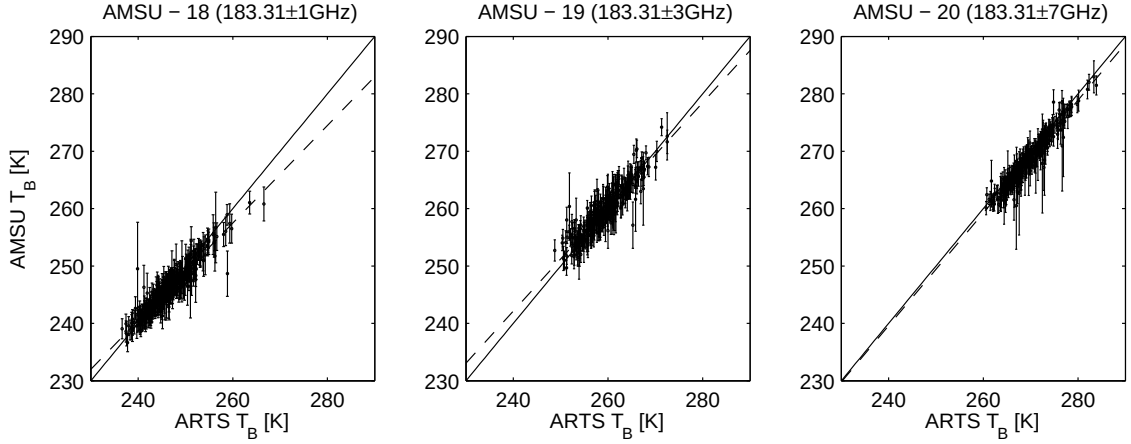


Figure 3.4: Average AMSU radiance for a 50 km radius target area versus ARTS modeled radiance based on radiosonde data. Time period: 2002; satellite: NOAA-15; number of points in the plot: 290. Vertical bars indicate $1 - \sigma(i)$ errors. The dashed line is a linear fit. Figures from left to right are for channels 18 to 20.

latter seem the more likely explanation, which would mean that the corrected Lindenberg radiosonde data have a small residual dry bias at low humidities, giving 0 %RH when the true humidity is still approximately 2–4 %RH.

3.2 Influence of Temperature Errors on Perceived Humidity Supersaturation

The equilibrium vapor pressure of water molecules over a plane surface of liquid water (e_w) or ice (e_i) depends only on temperature (T). (Strictly, this is only true for water vapor in the pure phase. If water vapor is mixed in air, e_w and e_i are slightly enhanced. However, the enhancement is below 0.5 % according to Sonntag [82], and hence can be safely neglected here.) There are a number of empirical formulas in use to calculate $e_w(T)$ and $e_i(T)$. The calculations presented here are based on the formulas by Sonntag [82], but differences between the different parameterizations are small and have no impact on the results. Figure 3.5 shows $e_w(T)$ and $e_i(T)$. The curves separate at $T = 0^\circ\text{C}$, at higher temperature, $e_i(T)$ is not defined. Note the strong and non-linear temperature dependence.

The equilibrium water vapor pressure is used to define relative humidity with respect to liquid water (RH_w) and ice (RH_i):

$$\text{RH}_w = \frac{e}{e_w(T)} \quad \text{RH}_i = \frac{e}{e_i(T)}, \quad (3.1)$$

where e is the actual water vapor pressure. Because e_i is lower than e_w , RH_i will always be higher than RH_w . Thus, it is possible that RH_i exceeds 100 %, while RH_w is still below 100 %. (This is the reason why in mixed-phase clouds the ice particles grow on the

3 Upper Tropospheric Humidity

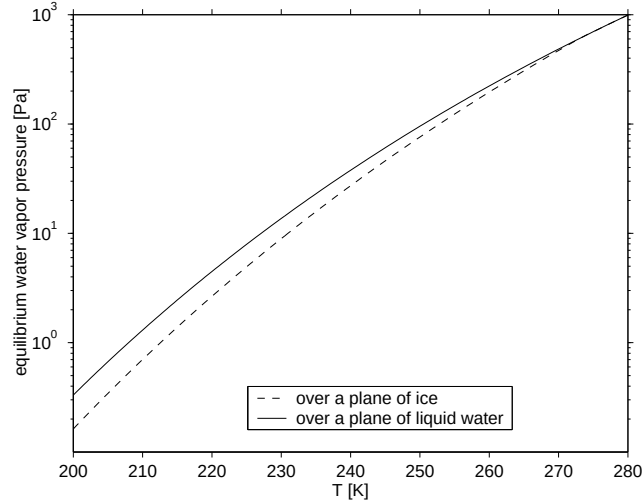


Figure 3.5: Equilibrium water vapor pressure over liquid water (solid line) and over ice (dashed line) as a function of temperature. Note the logarithmic scale.

expense of the liquid particles. This so-called Bergeron-Findeisen process is responsible for the formation of precipitation at midlatitudes.)

While $\text{RH}_w > 100\%$ does not occur in the earth's atmosphere, $\text{RH}_i > 100\%$ does occur quite frequently, as is well documented for example by Wallace and Hobbs [92]. The phenomenon can be explained by the absence of ice nuclei. Such supersaturation with respect to ice recently has received a lot of attention (Spichtinger et al. [84], Gierens et al. [40], Jensen et al. [51]). As an example, Figure 3.6 shows a distribution of RH_i at 215 hPa derived from the UARS MLS UTH dataset, which has been described by Read et al. [68]. It should be noted that Read et al. [68] themselves recommend to set data above 120% RH_i to 100%, for completely different reasons related to the radiative effect of cirrus clouds.

A closer inspection of Figure 3.6 reveals that some of the data points show even supersaturation with respect to liquid water. Consider the ratio of $e_w(T)/e_i(T)$, which is displayed in Figure 3.7. At a temperature of 220 K, consistent with the chosen altitude, the ratio is approximately 1.7, which means that RH_i values above 170% are above liquid water saturation and hence rather unlikely.

Ignoring these problems and using the data anyway for argument's sake, one can say that the problem with the distribution shown in Figure 3.6 is that the remote sensing method used relies on the fundamental law stating that the amount of radiation absorbed, emitted, or scattered is proportional to the amount of the interacting substance. (For extinction, this is stated by the Lambert-Beer law.) Hence, it measures absolute humidity, not relative humidity. An absolute humidity parameter is for example the specific humidity (q), defined as

$$q = \frac{m_w}{m_w + m_a} \left[\frac{\text{kg}}{\text{kg}} \right], \quad (3.2)$$

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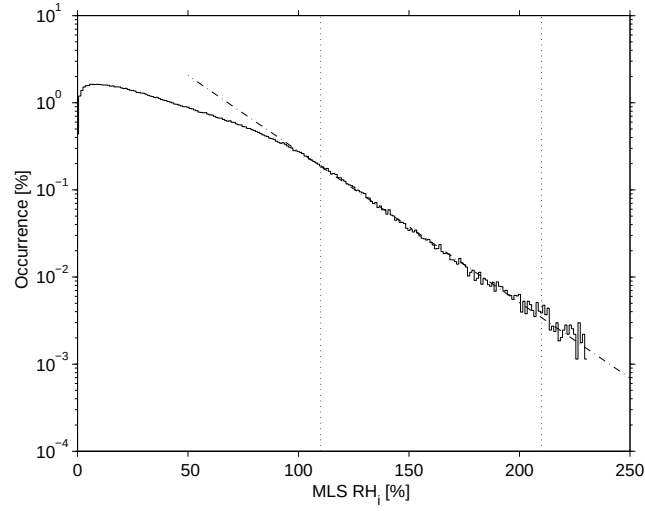


Figure 3.6: Distribution (fractional occurrence per 1 % RH_i bin) of relative humidity with respect to ice (RH_i) at 215 hPa, derived from the UARS MLS UTH dataset (solid line). A fitted exponential function (dashed-dotted line) shows that the number of occurrences decreases exponentially with increasing RH_i between 100 % and 230 %. Vertical dotted lines indicate fitting boundaries.

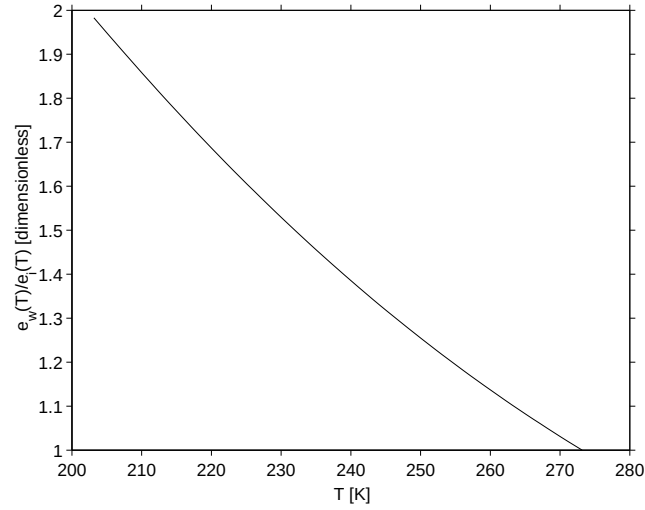


Figure 3.7: Ratio of equilibrium water vapor pressure over a plane of liquid water ($e_w(T)$) and equilibrium water vapor pressure over a plane of ice ($e_i(T)$) plotted against temperature (T).

where m_w is the mass of water molecules in a unit volume, and m_a is the mass of other air molecules.

To convert from q to RH_i , one must know the equilibrium water vapor pressure $e_i(T)$, hence the temperature T . However, T will generally be known only with a limited accuracy. The purpose of paper [H 6] is to demonstrate how uncertainties in T would lead to apparent supersaturation, even if there were no true supersaturation.

A Monte Carlo approach was chosen for the study. This had the advantage that non-Gaussian statistics and non-linearities could be correctly taken into account. The conclusion was that temperature uncertainties have a strong impact on perceived supersaturation if the relative humidity is calculated from measurements of absolute humidity. Even for moderate temperature uncertainties, very high perceived supersaturation can occur because the strongly non-linear temperature dependence of the equilibrium water vapor pressure enhances the ‘tail’ of the distribution towards high RH values. The resulting distribution for a flat q distribution and a Gaussian T error distribution is non-Gaussian, featuring an exponential drop-off behavior towards high RH values.

With an assumed T uncertainty of 2.7 K, the RH_i distribution measured by MLS can be reproduced without assuming any ‘real’ supersaturation. However, the point of the study was not to deny the reality of supersaturation, but to emphasize that the use of remotely sensed data for studies of supersaturation is problematic, and in particular requires a careful analysis of the influence of temperature uncertainties. A T uncertainty of 2 K, as assumed in Read et al. [68], would still account for a large part of the observed supersaturation.

3.3 A Simple Method to Relate Microwave Radiances to Upper Tropospheric Humidity

Radiosonde humidity measurements tend to have problems under the dry and cold conditions in the upper troposphere [23]. Furthermore, the radiosonde network is sparse, particularly over the oceans and in the equatorial regions. Thus, the only global upper tropospheric humidity measurements come from satellites. Infrared data at $6.7 \mu\text{m}$ from geostationary and polar orbiting satellites have been used extensively for this purpose. Soden and Bretherton [78], henceforth referred to as SOB, derived a simple relation between infrared radiances and upper tropospheric humidity:

$$\ln(\text{UTH}) = a + b T_b, \quad (3.3)$$

where UTH is a weighted mean of the fractional relative humidity in the upper troposphere, $\ln()$ is the natural logarithm, T_b is the radiance expressed in brightness temperature, and a and b are constants. The original relation by SOB contains also a $\cos(\theta)$ term, where θ is the zenith angle, which was omitted here for simplicity. SOB used the radiance Jacobian with respect to relative humidity for the weights in the calculation of UTH.

In the derivation of Relation 3.3, SOB made use of a reference pressure and a dimensionless lapse rate parameter. Various later studies made explicit use of these parameters to improve upon the simple relation. An overview on the different variants of the relation used over the years is given by Jackson and Bates [50]. We will henceforth refer to the method of using Relation 3.3 to transform radiances (expressed as brightness temperatures) to UTH as the BT transformation method.

The great advantage of this method is that radiances and radiance differences can be easily transformed to a more intuitive quantity, without using any a priori information. It is thus very well suited for climatological studies. A disadvantage at first sight is that the UTH defined as the weighted mean relative humidity of the upper troposphere can not be directly compared to other humidity measurements. In particular, the weights in the definition of UTH depend on the atmospheric state, as for a drier state lower altitudes are sampled. However, this difficulty can be easily overcome by doing the comparison in the proper way, which is to use a radiative transfer model to simulate radiances for all humidity datasets to be compared, and then use the transformation of Relation 3.3 to map the radiance differences back to UTH differences.

Quite a number of studies have used the BT transformation method to transform infrared radiances into UTH [33, 87, 5, 80], including a recent study on humidity supersaturation with respect to ice as seen by the High Resolution Infrared Sounder (HIRS) [39]. For microwave sensors, on the other hand, the method has not been much used. While there are many publications about microwave humidity profile retrieval, for example [98, 26, 73, 81], there appear to be only three publications using the BT transformation method.

The first to have used it for microwave data appear to be Spencer and Braswell [83], who applied the method to data from the Special Sensor Microwave humidity sounder (SSM/T-2) in order to study the UTH in the subtropical subsidence zones. They used simulated radiances for radiosonde data from one tropical station to determine the parameters a and b in Relation 3.3, but neither give the values of a and b , nor a detailed error analysis for the derived UTH, since the focus of the article is on the application rather than on the methodology. Shortly afterwards Engelen and Stephens [25] published a study comparing HIRS and SSM/T-2 UTH, derived by the BT transformation method. They used a regression on radiances generated for the TOVS Initial Guess Retrieval (TIGR-3) dataset [14] to determine a and b . Compared to Spencer and Braswell [83] there is a more detailed error analysis, but also no explicit values for a and b . Finally, Greenwald and Christopher [42] used the BT transformation method in their analysis of the effect of cold clouds on UTH derived from the Advanced Microwave Sounding Unit (AMSU) B. Since their main focus is on clouds, there is not much discussion on the BT transformation method, but at least the values $a = 20.95$ and $b = -0.089 \text{ K}^{-1}$ are given for the transformation coefficients.

The goal of the study presented in paper [H 1] was to demonstrate how the BT transformation method can be applied to AMSU data, to explicitly document the transformation coefficients to use, to discuss the method's performance, and to point out limitations. Although the analysis was carried out for microwave data, some of the new findings can also be applied to the more traditionally used infrared data. To keep things simple the study

focused only on the clear sky case, although the impact of clouds is an important issue for climatological applications, as shown by Greenwald and Christopher [42], even if the impact of clouds is much less dramatic than in the infrared.

The result is that the method can be used to retrieve Jacobian-weighted upper tropospheric humidity (UTH) in a broad layer centered roughly between 6 and 8 km altitude. The bias in UTH is always below 4 %RH, where the largest values are found for high humidity cases. The relative bias in UTH is always below 20%, where the largest values are found for low humidity cases. The UTH standard deviation is between 2 and 6.5 %RH in absolute numbers, or between 10 and 27% in relative numbers. The standard deviation is dominated by the regression noise, resulting from vertical structure not accounted for by the simple transformation relation. The part of the UTH error resulting only from radiometric noise scales with the UTH value and has a relative standard deviation of approximately 7% for a radiometric noise level of 1 K. The UTH retrieval performance was shown to be of almost constant quality for all looking angles and latitudes, except for problems at high latitudes due to surface effects.

A comparison of AMSU UTH and radiosonde UTH for the radiosonde station Lindenberg was used to validate the retrieval method. The agreement is good if known systematic differences between AMSU and radiosonde are taken into account.

Additionally, a method similar to the one discussed in Section 3.2 and paper [H 6] was used to investigate whether the BT transformation method is suitable to study humidity supersaturation in the upper troposphere. In principle it is, but it has to be kept in mind that regression noise and radiometric noise would lead to apparent supersaturation even if there were no real supersaturation. For a radiometer noise level of 1 K the drop-off slope of the apparent supersaturation is $0.17 (\%RH)^{-1}$, for a noise level of 2 K the slope is $0.12 (\%RH)^{-1}$.

The mathematics leading to the exponential drop-off behavior of the supersaturation curve are the same as discussed in Section 3.2: an enhancement of the tail of a Gauss distribution due to the mapping by a nonlinear function. However, the underlying physics are quite different. In the case discussed in Section 3.2, Gaussian errors in the temperature measurement are mapped to an exponential supersaturation curve by the nonlinear saturation water vapor pressure curve, displayed in Figure 3.5. In the case discussed here and in paper [H 1], Gaussian radiometric noise is mapped to an exponential supersaturation curve by the transformation according to Equation 3.3.

The main conclusion from the study was that the BT transformation method is very well suited for microwave data. Its particular strength is in climatological applications where the simplicity and the independence of a priori information are key advantages. Further studies applying the method to global and regional data are planned. To give a flavor of the method's capability, one can apply the method to an arbitrary AMSU overpass. Figure 3.8 shows AMSU radiances and derived UTH for a pass over Europe. The top plot shows the original radiances, displayed as brightness temperatures in Kelvin, the bottom plot shows the derived UTH in relative humidity with respect to ice. The image on the cover page shows global UTH data for the same day in the same color scale as the bottom plot

3 Upper Tropospheric Humidity

of Figure 3.8. Compare for example the dry air extrusion reaching from the subtropics across Europe.

3 Upper Tropospheric Humidity

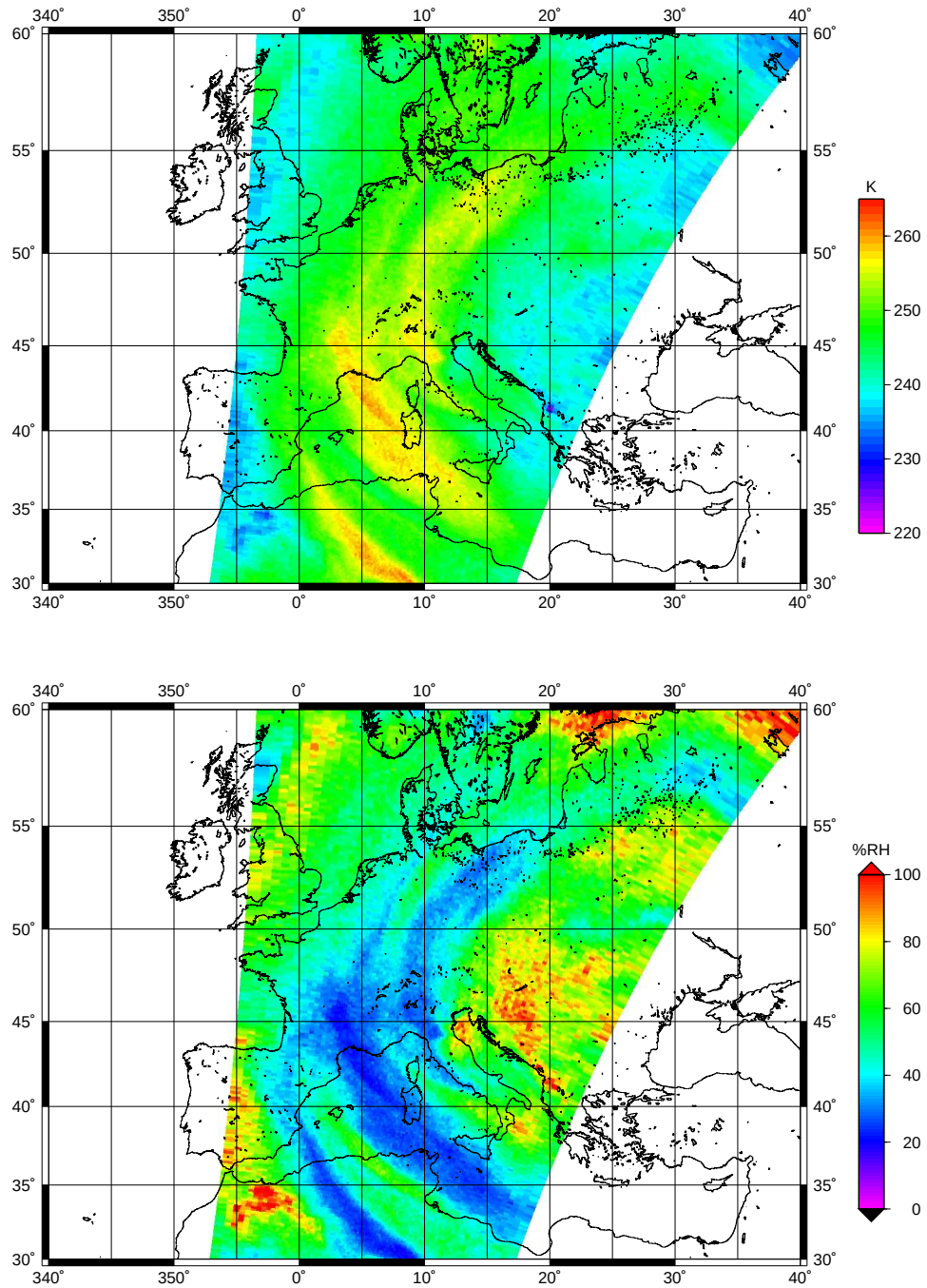


Figure 3.8: An AMSU pass over Europe. These data were taken by the NOAA 16 satellite on June 6, 2004. Displayed are Channel 18 brightness temperatures in K (top) and UTH in %RH_i (bottom).

4 Radiative Transfer Model Development

Radiative transfer (RT) models are the most crucial tools for passive remote sensing of the atmosphere. They are needed to relate the radiance received by a remote sensing instrument to the physical state of the system observed. For atmospheric applications the state consists of temperature, pressure, and the concentration of various trace gases. Additionally there can be hydrometeors, such as ice cloud particles, liquid cloud particles, or precipitation. The atmosphere contains also aerosols, but these are invisible to microwave and infrared instruments.

In 1999 no adequate model for the millimeter to submillimeter wave spectral range was available. To solve this problem, our group at the University of Bremen and Patrick Eriksson from Chalmers University of Technology (Gothenburg, Sweden) started a joint initiative for the development of a general, flexible, and freely available radiative transfer model. The new model was called ARTS, the Atmospheric Radiative Transfer Simulator. The first version of ARTS, described in paper [H 5], simulated only clear-sky radiative transfer, without clouds or precipitation. It is then fairly straightforward to integrate the radiative transfer equation, as explained in Section 4.1.

The ARTS model has in the meantime become a standard tool for a growing community. To help the development, a yearly radiative transfer workshop takes place in Bredbeck near Bremen since 1999 (the sixth workshop will take place in June 2004). These workshops are an important forum for the ARTS user community, and also for the developers. The results of the first two workshops are summarized in two books, Eriksson and Buehler [28] and Buehler and Eriksson [11]. The two books together summarize the state of the art in clear sky microwave radiative transfer modeling. At a later workshop a systematic intercomparison of all participating models was undertaken, which is summarized in Melsheimer et al. [62].

A much bigger problem is the impact of cirrus clouds on the microwave radiances measured by the sensor, since the ice particles in the clouds scatter the microwave radiation. There is no analytical solution to the radiative transfer equation in the presence of scattering, only a large number of approximate numerical methods. Unfortunately, none of the available methods was adequate for our purpose, since we wanted to be able to model all four components of the Stokes vector (to account for polarization effects introduced by the scattering) in spherical geometry (to account for the spherical symmetry of the atmosphere and allow the simulation of limb sounding measurements). Paper [H 8], which is introduced in Section 4.2, describes an iterative discrete order scheme to solve this

problem for a plane parallel atmosphere. The scheme, called Discrete Order Iterative solution method (DOIT) was later extended to handle also 1D spherical and 3D spherical atmospheres [24].

The DOIT scattering algorithm was implemented in a completely new version of the ARTS radiative transfer model [29], again in close collaboration with Patrick Eriksson from Chalmers University. In parallel, Cory Davis from the University of Edinburgh developed a Monte Carlo radiative transfer scheme for ARTS. The possibility to compare Monte Carlo and DOIT method within the same framework has since proven to be very useful. Further validation was done against satellite data using fields of a mesoscale numerical weather prediction model as input [27]. The DOIT scheme has also been compared against a simpler single scattering scheme in the infrared [47].

All programs described in this chapter, along with extensive documentation, are freely available on the Internet under <http://www.sat.uni-bremen.de/arts/>.

4.1 A Public Domain Clear-Sky Radiative Transfer Model

The ARTS-1-0-x version discussed in paper [H 5] is limited to cases where scattering can be neglected and local thermodynamic equilibrium applies. At millimeter and sub-millimeter wavelengths these assumptions are valid from the troposphere up to the mesosphere, but only in the clear-sky case.

The model carries out scalar radiative transfer calculations, that means it treats only the first component of the Stokes vector, corresponding to the total intensity. This is a good approximation in the absence of polarization effects. The only sources of polarization effects in the atmosphere are scattering, which has already been excluded, and Zeeman splitting of some spectral lines due to the earth's magnetic field. Hence, the scalar treatment implies that the Zeeman effect can not be modeled explicitly.

The model assumes a one-dimensional spherical atmosphere, in other words, the atmosphere is assumed to be spherically symmetric, with all parameters varying as a function of the vertical coordinate only. The primary vertical coordinate is pressure. All other quantities, such as temperature, geometric altitude, and trace gas concentrations, are given on pressure grids.

ARTS can be used to simulate measurements for any observation geometry: up-looking, down-looking, or limb-looking, and for any sensor position: on the ground, inside the atmosphere, or on a satellite above the atmosphere. The model has been developed having passive emission measurements in mind, but pure transmission measurements are also handled. In the case of pure transmission measurements, atmospheric emission can be neglected. This is the case for occultation measurements towards the sun or an active source.

The model works with arbitrary frequency grids, hence it can be used both for the simula-

tion of high spectral resolution sensors, and for the simulation of broad frequency ranges. The applicable spectral range is from the microwave up to the thermal infrared. In that frequency range, particular care has been taken to make the absorption calculation consistent with state of the art continuum models for water vapor and nitrogen, and with continuum and line mixing models for oxygen.

Besides providing sets of spectra, ARTS can calculate Jacobians for a number of variables. Analytical expressions are used to calculate Jacobians for trace gas concentrations, continuum absorption, and ground emissivity. A perturbation method is used to calculate Jacobians for pointing offsets, frequency offsets, and spectroscopic parameters. For temperature Jacobians, the user can choose between an analytical method, which does not assume hydrostatic equilibrium, and a perturbation method, which does assume hydrostatic equilibrium.

Under the conditions defined above, the radiative transfer through the atmosphere can be described by a simple differential equation for the specific intensity I , which is defined as the power traveling in a given direction, per unit area, per unit solid angle, and per unit frequency interval. The applicable simplified form of the radiative transfer equation is:

$$\frac{dI(\nu, s)}{ds} = -\alpha(\nu, s)I(\nu, s) + \alpha(\nu, s)B(\nu, T(s)) , \quad (4.1)$$

where α is the absorption coefficient, B is the Planck function, ν is the frequency, and T is the physical temperature. The equation describes the change in I as the radiation travels along a path, where the distance along the path is given by s . It should be noted that the equation assumes that the path is known, so the problem to determine the path has to be solved separately, as will be described below. Equation 4.1 is a monochromatic equation, i.e., it is valid independently for each frequency, but not valid for frequency averages. This equation is significantly simpler than the general form of the radiative transfer equation, which is described in Section 4.2 and in paper [H 8].

The absorption coefficient α , as defined by Equation 4.1, can generally be calculated as a sum of different spectral lines of the different gaseous species, plus some additional terms related to absorption continua:

$$\begin{aligned} \alpha(\nu, p, T, x_1, \dots, x_N) = & \\ & \sum_{i=1}^N \frac{p x_i}{k_B T} \sum_{j=1}^{M_i} S_{ij}(T) F(\tilde{\nu}_{ij}, \nu, p, T, x_1, \dots, x_N) \\ & + C_1(\nu, p, T, x_1, \dots, x_N) + \dots + C_L(\nu, p, T, x_1, \dots, x_N) , \end{aligned} \quad (4.2)$$

where p is the pressure and $x_1 \dots x_N$ are the volume mixing ratios of the various gas species. The index i goes over all N gas species and the index j over all M_i spectral lines of each gas species. The k_B in the $p x_i / (k_B T)$ term is Boltzmann's constant, which means that this term is nothing else than the partial density n_i of gas species i .

The contribution of each spectral line is given by the product of the line intensity $S_{ij}(T)$ and the line shape function $F(\tilde{\nu}_{ij}, \dots)$. ARTS allows the user to select between different

line shapes and line shape combinations, for details see paper [H 5]. The first argument of F , $\tilde{\nu}_{ij}$, is the line center frequency, which follows directly from the energy difference of the two states involved in the transition, plus a possible pressure shift. (On pressure shift see paper [H 10], paper [H 3], and Section 5.2.) In addition to the line spectrum one has to take into account several continua, C_1 to C_L , which in the general case are functions of frequency, pressure, temperature, and gas volume mixing ratios.

According to Pickett et al. [67] the line strength $S_{ij}(T)$ of Expression 4.2 can be calculated as

$$S_{ij}(T) = S_{ij}(T_0) \frac{Q_i(T_0)}{Q_i(T)} \frac{e^{-L_{ij}/(k_B T)} - e^{-U_{ij}/(k_B T)}}{e^{-L_{ij}/(k_B T_0)} - e^{-U_{ij}/(k_B T_0)}}. \quad (4.3)$$

Here, $S_{ij}(T_0)$ is the line strength at a reference temperature T_0 , which is obtained from a spectroscopic database. The function $Q_i(T)$ is the partition function, more correctly the total internal partition sum as defined for example by Gordy and Cook [41]. More information on partition functions can be found in Verdes et al. [88]. The parameters L_{ij} and U_{ij} are the energies of the lower and upper state, respectively. The lower state energy is obtained from the database, the upper state energy calculated by $U_{ij} = L_{ij} + h\tilde{\nu}_{ij}$. The $e^{-(...)/(k_B T_x)}$ terms reflect the Boltzmann distribution of the energy level population.

Overall, all the equation does is scale the line strength from the reference temperature to a different temperature, using that

$$S_{ij}(T) = \text{const} \frac{e^{-L_{ij}/(k_B T)} - e^{-U_{ij}/(k_B T)}}{Q_i(T)}. \quad (4.4)$$

The next task in solving the radiative transfer equation 4.1 is to determine the propagation path, i.e., the path through the atmosphere traveled by the radiation reaching the sensor. Refraction affects the radiative transfer in several ways. The most notable effect is that for limb sounding the tangent point is displaced vertically. The tangent point is displaced downwards compared to the pure geometrical case (for a fixed observation direction), therefore inclusion of refraction in general gives higher intensities. There is also a horizontal displacement of the tangent point, but that is not important for a spherically symmetric atmosphere, except for the fact that the distance traversed through the atmospheric layers is changed.

As the refractive index is frequency dependent [60], the atmosphere is in principle dispersive. Each frequency component has its own propagation path. For measurements using frequencies below 1000 GHz, notable dispersion will only take place in the troposphere and in the ionosphere. The strongest effect occurs for limb sounding into the troposphere. However, even in that case the dispersion is noticeable only around a few strong water vapor transitions, as shown by Eriksson et al. [32]. The atmosphere is totally opaque at lower altitudes around these transitions and the propagation path through the troposphere does not influence the radiance seen by a sensor placed in space. Thus, dispersion can be neglected for the discussed frequency range, and is in fact neglected in ARTS. The

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(non-dispersive) refractive index n is calculated following Elgered [22]:

$$n(p, T, e) = 1 + 77.593 \cdot 10^{-8} \frac{p - e}{T} + e \left(\frac{72 \cdot 10^{-8}}{T} + \frac{3.754 \cdot 10^{-3}}{T^2} \right), \quad (4.5)$$

where p is the total air pressure in Pascal, T the temperature in Kelvin, and e the water vapor pressure in Pascal.

Propagation paths can be calculated neglecting refraction effects. This results in pure geometrical calculations, handled by the Pythagorean rule and standard trigonometric expressions (see Buehler et al. [12] for details). The calculations including refraction effects are based on Snell's law, applied to a spherically symmetric atmosphere. One can easily show that the product of the radius, refractive index n , and zenith angle θ is constant along the propagation path (see for example Kyle [58], Balluch and Lary [3], Rodgers [72]):

$$\gamma = (R_e + z)n(z) \sin(\theta), \quad (4.6)$$

where γ is called the path constant, R_e is the earth geoid radius, and z the altitude above the geoid. The path constant is determined by the zenith angle of the sensor's line-of-sight, and the radius and refractive index at the sensor position. The calculations start by determining the lowest altitude of the path. For downward observations, the tangent altitude z_t is found by the implicit expression

$$(R_e + z_t)n(z_t) = \gamma. \quad (4.7)$$

The tangent altitude will normally not happen to be on a model grid point, but between two grid points. It is determined practically by linear interpolation in $\gamma'(z) = (R_e + z)n(z)$. If the tangent altitude would be below the surface the sensor sees the ground. In that case the zenith angle of the propagation path at the ground, θ_g , is given by

$$\theta_g = \sin^{-1} \left(\frac{\gamma}{(R_e + z_g)n(z_g)} \right). \quad (4.8)$$

The radiative transfer is evaluated along the propagation path, while Equation 4.6 is expressed for vertical altitudes. The relationship between a change in vertical altitude and the corresponding distance along the path is here denoted as the geometrical term and it is

$$g(z) = \frac{ds}{dz} = \frac{1}{\cos(\theta)}, \quad (4.9)$$

which can be rewritten, using trigonometric identities and Equation 4.6, to

$$g(z) = \frac{(R_e + z)n(z)}{\sqrt{(R_e + z)^2 n^2(z) - \gamma^2}}. \quad (4.10)$$

If the refractive index is assumed to be constant between altitude z_1 and z_2 , Equation 4.10 gives

$$\Delta s = \sqrt{(R_e + z_2)^2 - \left(\frac{\gamma}{\bar{n}}\right)^2} - \sqrt{(R_e + z_1)^2 - \left(\frac{\gamma}{\bar{n}}\right)^2}, \quad (4.11)$$

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where Δs is the distance along the propagation path between z_1 and z_2 , $z_2 > z_1$ and $\bar{n} = (n(z_1) + n(z_2))/2$. Further details of these calculations are given in Buehler et al. [12].

The simplified form of the radiative transfer equation 4.1 can be solved analytically:

$$I(\nu) = I_0(\nu)e^{-\tau(\nu,0,l)} + \int_0^l B(\nu, T(s))\alpha(\nu, s)e^{-\tau(\nu,0,s)}ds, \quad (4.12)$$

where I_0 is the intensity at the practical starting point of the propagation path, at the distance l from the observation position, and

$$\tau(\nu, l_1, l_2) = \int_{l_1}^{l_2} \alpha(\nu, s)ds \quad (4.13)$$

is the optical depth along (a part of) the propagation path.

Different numerical schemes to calculate Expression 4.12 exist (see for example Rodgers [72]). In the approach selected here, the radiation is followed along the path from one point to the next, using the expression

$$I_{q+1} = I_q e^{-\tau_q} + \bar{B}_q(1 - e^{-\tau_q}), \quad (4.14)$$

where

$$\tau_q = \Delta s \frac{\alpha_q + \alpha_{q+1}}{2}, \quad (4.15)$$

$$\bar{B}_q = \frac{B_q + B_{q+1}}{2}. \quad (4.16)$$

Here I_q is the intensity reaching path point q , α_{q+1} is the absorption at path point $q + 1$, and so on. Equation 4.14 is valid under the condition that the Planck function is constant for each step, and can be approximated by the mean of the values at the end points. Absorption variations are assumed to be piecewise linear.

Using precalculated absorption coefficients according to Equation 4.2, path points according to Equation 4.6, and the numerical integration scheme of Equation 4.14, the model can calculate $I(\nu)$ for any viewing direction. To simulate measurements by real instruments this quantity has to be integrated with the instrument's frequency response and viewing angle response. This integration, which can be formulated mathematically as a simple matrix-vector multiplication, is not done by ARTS, but by the external program Qpack, which is described in Eriksson et al. [30].

Due to its flexible implementation, ARTS can not only be used for its original purpose of sensor simulation, as documented in paper [H 4], but also to simulate radiative fluxes over wide frequency ranges, as documented in paper [H 2].

4.2 A Simple Iterative Method for Scattering Radiative Transfer Calculations in a 1D Plane Parallel Atmosphere

Satellite remote sensing data are extensively used by the numerical weather prediction community for global measurements of geophysical parameters. The millimeter wave channels on AMSU-B [19] and SSM/T-2 [56] provide global data on the distribution of humidity in the upper troposphere. However, in cloudy areas these data are influenced by the presence of cirrus clouds, which have distinct radiative transfer properties compared to other cloud types. The scattering radiative transfer model described in paper [H 8] was the first step towards a better quantification of the cirrus cloud impact, and ultimately towards the utilization of these data for climate research.

Another important motivation for this work was the recently proposed ESA opportunity mission Cloud Ice Water Submillimeter Wave Imaging Radiometer (CIWSIR). CIWSIR is mainly based on the studies of Miao et al. [63] and Evans et al. [34], in which they have discussed the rationale for adopting submillimeter wave channels for retrieving cirrus cloud ice water path (IWP) and characteristic particle size. As the water vapor absorption is relatively strong in the submillimeter frequency range, the lower atmosphere is opaque so that the effects of surface and lower clouds on the upwelling radiation are negligible. Moreover, at these frequencies ice particles are weak absorbers, which makes the physical temperature of the cirrus clouds unimportant. As a first application of the new model the paper describes the effect of cirrus IWP on the measured radiance at the CIWSIR frequencies.

There was already a lot of work done in the area of RT models with scattering. For example, Evans and Stephens [35] developed a numerical model that solves the polarized radiative transfer equation for a plane parallel vertically inhomogeneous scattering atmosphere. In this model the solution method for the multiple scattering aspect of the problem is doubling and adding. Another example is the model described by Czekala and Simmer [17] that uses the successive order of scattering method for nonspherical particles in a plane parallel atmosphere. However, developing a new model from scratch was necessary to allow modifications and extensions to suit the requirements. The RT model discussed in paper [H 8] calculated only the first component of the Stokes vector for a plane parallel atmosphere. However, the model was formulated in such a way that the later extension to all Stokes components and a spherical atmosphere, as described in Emde et al. [24], was feasible. That extended model was used for the simulations discussed in Section 5.3 and paper [H 7].

According to Mishchenko et al. [64], the general radiative transfer equation for an atmosphere taking into account extinction, emission and scattering is:

$$\frac{d\mathbf{I}(\mathbf{n})}{ds} = -\mathbf{K}(\mathbf{n}) \mathbf{I}(\mathbf{n}) + \mathbf{a}(\mathbf{n}) B(T) + \int_{4\pi} d\mathbf{n}' \mathbf{Y}(\mathbf{n}, \mathbf{n}') \mathbf{I}(\mathbf{n}') , \quad (4.17)$$

where $\mathbf{I}(\mathbf{n})$ is the four-component specific intensity vector of multiply scattered radia-

tion propagating in the direction $\mathbf{n}$. The first component of this vector, I , gives the total intensity, the second component Q and the third component U give the degree of linear polarization, and the fourth component V gives the degree of circular polarization. The pathlength element measured in the direction of $\mathbf{n}$ is ds . The extinction of radiation is described by the 4x4 extinction coefficient matrix $\mathbf{K}(\mathbf{n})$ which includes the extinction due to gas and particles.

The second term in Equation 4.17 is the thermal source term where $\mathbf{a}(\mathbf{n})$ is the absorption coefficient vector and $B(T)$ is the Planck function at temperature T . The absorption coefficient vector $\mathbf{a}(\mathbf{n})$ is the sum of particle absorption and gaseous absorption.

The last term in Equation 4.17 is the scattering source term. It describes the contribution of radiation illuminating a small volume element from all directions $\mathbf{n}'$ and scattered into the direction $\mathbf{n}$. The matrix $\mathbf{Y}(\mathbf{n}, \mathbf{n}')$ is called the scattering efficiency matrix. It can be expressed in terms of the well known phase matrix $\mathbf{Z}(\mathbf{n}, \mathbf{n}')$ as

$$\mathbf{Y}(\mathbf{n}, \mathbf{n}') = n^p \mathbf{Z}(\mathbf{n}, \mathbf{n}') , \quad (4.18)$$

where n^p is the particle number density. The Phase matrix gives the efficiency of scattering between directions $\mathbf{n}$ and $\mathbf{n}'$. The name is historic and has nothing to do with the phase of an electromagnetic wave.

In the algorithm presented in paper [H 8], Equation 4.17 is solved iteratively in a plane-parallel atmosphere for the scalar case, i.e., only for the first component of the specific intensity vector. The plane parallel approximation holds when the radius of curvature is much larger than the thickness of the layer. Also keeping in mind that solving this equation for $\mathbf{I}(\mathbf{n})$ may cause computational difficulties as it involves the exponential of a matrix, a linear approximation is considered assuming optically thin layers. The advantage of this approach is that it is conceptually simple but fully general and can be extended easily for the full vector equation.

Assuming that all cloud ice particles are spherical in shape and that the incident radiation is unpolarized, the radiative transfer equation for total intensity is written as

$$\frac{dI(\mathbf{n})}{ds} = -K_{11}(\mathbf{n})I(\mathbf{n}) + a_1(\mathbf{n})B(T) + \int_{4\pi} d\mathbf{n}' Y_{11}(\mathbf{n}, \mathbf{n}') I(\mathbf{n}') , \quad (4.19)$$

where K_{11} is the (1,1) element of the extinction coefficient matrix $\mathbf{K}$, I is the first Stokes component, a_1 is the first component of the absorption vector $\mathbf{a}$, and Y_{11} is the (1,1) element of the scattering efficiency matrix $\mathbf{Y}$.

Equation 4.17 can be solved iteratively with a first guess radiation field. This field can either be the clear sky radiation field or the Extinction and Thermal Source (ETS) field. The ETS field is the radiation field generated by the first two terms of Equation 4.17.

Assuming local thermodynamic equilibrium, an analytical solution to Equation 4.19 for homogeneous layers in the absence of the scattering integral is

$$I_{i,\theta} = I_{i-1,\theta} \exp(-\tau_{\text{ext}}) + B(T) [1 - \exp(-\tau_{\text{abs}})] , \quad (4.20)$$

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where i is the level index. The opacities τ_{ext} and τ_{abs} are defined by

$$\tau_{\text{ext}} = \frac{1}{\cos(\theta)} \int K_{11}(\mathbf{n}) dz \quad (4.21)$$

and

$$\tau_{\text{abs}} = \frac{1}{\cos(\theta)} \int a_1(\mathbf{n}) dz , \quad (4.22)$$

where θ is the viewing angle. The factor $1/\cos(\theta)$ follows from the plane parallel geometry. Each layer is assumed to have a representative extinction coefficient, absorption coefficient, and temperature. The representative value of each parameter is calculated by taking the mean of the values at adjacent levels. When considering the clear sky case, $\tau_{\text{ext}} = \tau_{\text{abs}} = \tau$.

Assuming optically thin layers, a linear approximation to Equation 4.20 is written as

$$I_{i,\theta} = I_{i-1,\theta}(1 - \tau_{\text{ext}}) + B(T)\tau_{\text{abs}} , \quad (4.23)$$

which can be solved by traversing the atmosphere layer by layer in the direction of propagation starting with appropriate boundary conditions.

In order for the linear approximation to be valid each layer should be optically thin which is not always the case. Opacity can take large values if the pathlength is large as it is the case when θ is close to 90° . Here an opacity limit is defined for which the linear approximation is valid within the required accuracy. If the opacity of a layer is higher than this limiting value, the layer is divided further into sub-layers until the opacity is below the limiting value. For $\theta < 90^\circ$, the iteration starts with the intensity array initialized with the cosmic background radiation, corresponding to a brightness temperature of 2.735 K. The iteration is performed from one layer to the next storing the intensity value for each θ . Upon reaching a layer where clouds are included, extinction as well as emission due to particles are also taken into account. At the ground, the effect of ground reflection is modeled as

$$I_{0,\theta'} = I_{0,\theta}(1 - \sigma) + \sigma B(T_G) , \quad (4.24)$$

where σ is the ground emissivity, $I_{0,\theta'}$ is the intensity after reflection, $I_{0,\theta}$ is the intensity before reflection and T_G is the ground temperature. After reflection, the radiation is propagated along the angle $\theta' = 180^\circ - \theta$ and the intensity values are stored for θ' . Once the iteration is completed the intensity reaching a point from all directions is known. This is essential for the evaluation of the scattering integrals. The scattering source terms which are computed using this radiation field as a first guess are used to integrate the full radiative transfer equation. The resulting radiation field is used to generate new scattering source terms and the iteration is repeated until the solution converges.

Paper [H 8] also documents some test calculations with the new model. For example, Figure 4.1 shows the angular variation of brightness temperature for the clear sky case and a cloudy case at different heights in the atmosphere. Figure 4.1a is for an altitude below the cloud, Figure 4.1b inside the cloud, and Figures 4.1c and 4.1d above the cloud.

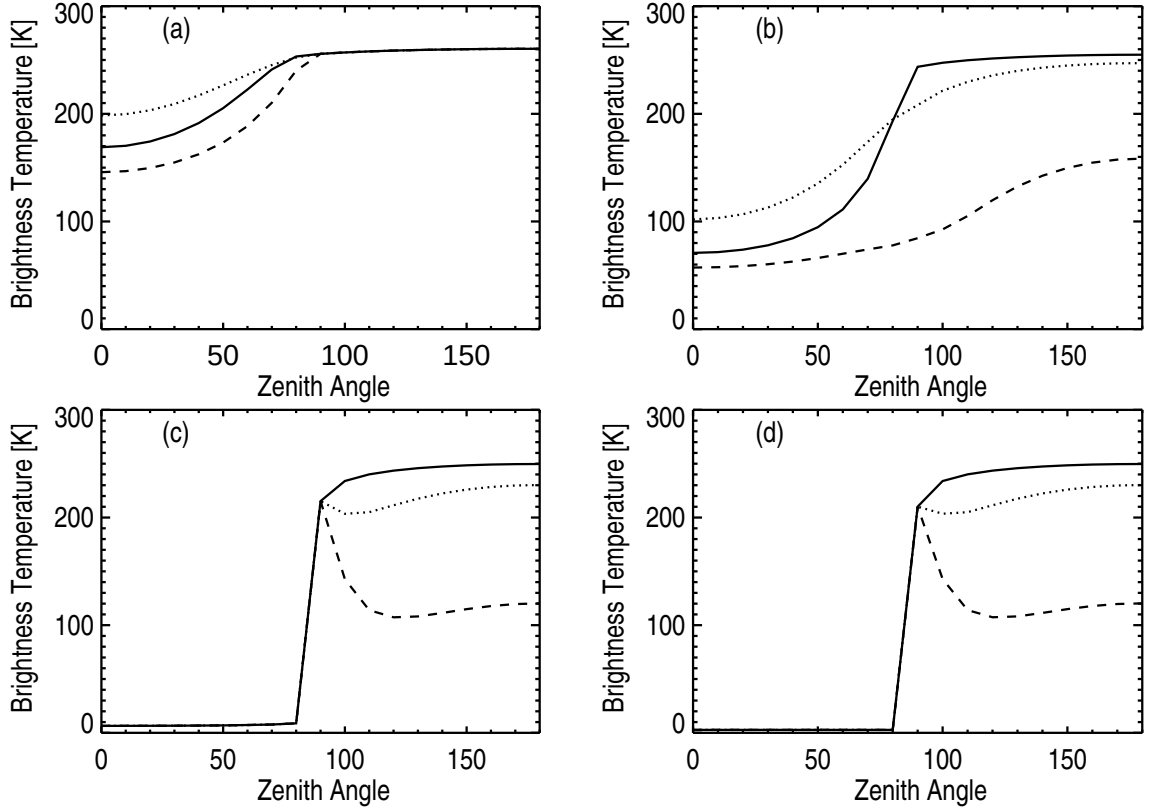


Figure 4.1: The radiation field inside the atmosphere for a strong cirrus case. The radius of the spherical ice particles was $100 \mu\text{m}$, the temperature was 250 K, the ice mass content (IMC) was 0.02 g m^{-3} , the cloud thickness was 2.5 km, and the frequency was 874 GHz. The four sub-plots are for altitudes of: (a) 4 km, (b) 6 km, (c) 25 km, and (d) 80 km. The solid line is the clear sky field, the dashed line is the ETS field, and the dotted line is the multiply scattered field after convergence.

The clear sky field shown by the solid line remains almost constant for the downlooking angles at all altitudes. In the up-looking case at lower altitudes emission from the cloud above is seen whereas at higher altitudes only cosmic background is seen.

The dashed line shows the ETS field. The dotted line shows the scattered field computed by iterating equation 2 with the ETS field as the first guess field. The scattered field converges after about 15 iterations. For downlooking angles viewing from above the cloud (25 km and 80 km) scattering decreases brightness temperature compared to the clear sky case. For uplooking cases looking from below the cloud (4 km) scattering increases the brightness temperature. Of particular interest is the field inside the cloud at 6 km (Figure 4.1b) where we see how scattering actually does a redistribution of radiation by comparing the scattered field to the clear sky field.

5 Limb Sounders

A limb sounder is a receiver that looks tangentially through the atmosphere, as indicated schematically in Figure 5.1. It is equipped with a movable antenna that makes it possible to adjust the viewing direction so that different tangent altitudes can be selected. The received radiation is analyzed by a spectrometer. In the case of a microwave instrument the spectral resolution can in principle be designed to be arbitrarily fine, but for practical reasons there is a tradeoff between spectral resolution and total bandwidth. The planned MASTER instrument will have a bandwidth of approximately 8 GHz per observation band (there will be 5 observation bands altogether), and a spectral resolution of the order of a few MHz.

The technique was originally developed in order to satisfy the need of stratospheric trace gas measurements with a good altitude resolution. A good introduction to this classical application of limb sounding can be found in Waters [94]. Recently, the investigation of the upper troposphere and tropopause region has received some attention as a new application that has its own advantages and limitations.

The limb sounding technique can be used with or without an external source. In the case of a microwave instrument no external source is needed because the thermal emission from the atmosphere itself is sufficiently strong. Figure 5.2 shows a typical limb spectrum. The intensities of the trace gas emission lines contain information about the trace gas concentrations. The widths of the lines contain information about the pressure, which can be used to determine the instrument's viewing angle. The overall intensity of the signal, as well as the relative strengths of the different lines, contain information about the temperature. Paper [H 9] describes a study on the quality of the temperature information obtained in this way.

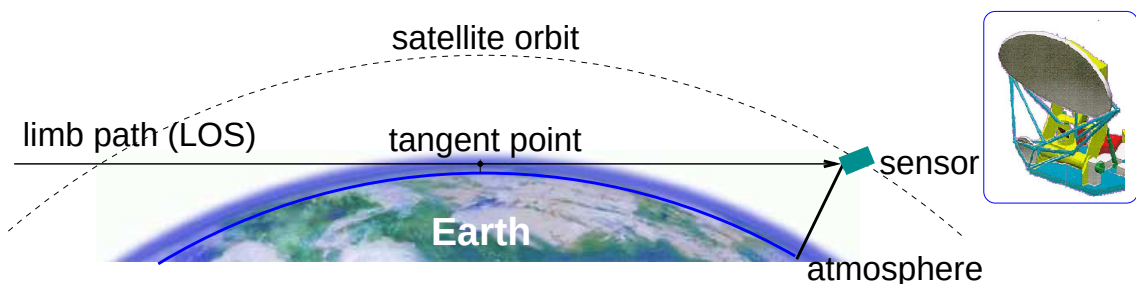


Figure 5.1: A schematic drawing of the limb sounding geometry. The receiver is in orbit and looks tangentially through the atmosphere.

5 Limb Sounders

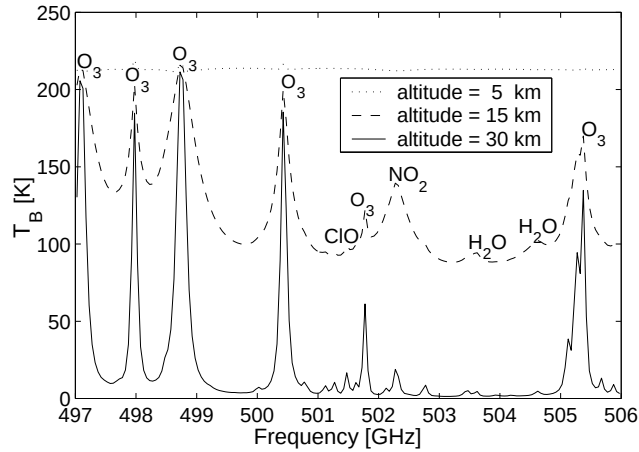


Figure 5.2: Simulated limb spectra for tangent altitudes of 5 km (dotted), 15 km (dashed), and 30 km (solid). The species to which the main transitions belong are indicated.

Limb sounding in the millimeter and submillimeter wave spectral range is a very suitable remote sensing technique for atmospheric chemistry and climate studies. Its advantage compared to downlooking techniques is the better altitude resolution, which is crucial for radiation studies as argued in paper [H 2]. The advantage of microwave and infrared techniques compared to visible and ultraviolet techniques is that the former use thermal radiation emitted by the earth and the atmosphere itself, whereas the latter use sunlight and thus work only in daytime. Compared to infrared techniques the millimeter and submillimeter techniques have the additional advantage that they are less affected by clouds (see for example Hoepfner and Emde [47]). Some clouds do have an influence, however, which is the subject of Section 5.3 and paper [H 7].

The typical spectral resolution of limb sounding instruments is so fine that the measurements are made at a number of frequencies across the width of a single spectral line. It is therefore important to model the line shape and the instrument response accurately as inaccurate modeling will lead to a poor match between the measured radiances and those re-calculated from the retrieved atmosphere. In paper [H 10], the 183.3 GHz rotational transition of water vapor and the 184.4 GHz transition of ozone were considered. These transitions have been used to measure water vapor and ozone in the middle atmosphere by the two classical millimeter wave limb sounding instruments, the Microwave Limb Sounder (MLS) [4] and the Millimeter-wave Atmospheric Sounder (MAS) [45]. Both instruments have had similar problems matching measured and re-calculated 183 GHz radiances. The study showed that by allowing the retrieval algorithm to fit certain spectral and instrumental parameters in addition to the mixing ratio profiles, one can improve both the quality of the retrieved profiles and our knowledge of spectroscopic parameters, in particular the pressure shift of the line.

In recent years the receiver technology has advanced considerably, which makes it possible to envisage instruments of a far more ambitious character. The new generation in-

cludes MASTER (Millimeter Wave Acquisitions for Stratosphere/Troposphere Exchange Research), SOPRANO (Sub-Millimeter Observation of Processes in the Atmosphere noteworthy for Ozone), and JEM/SMILES (Superconducting Sub-Millimeter Wave Limb Emission Sounder on the Japanese Experimental Module of the International Space Station). It also includes the EOS/MLS (Microwave Limb Sounder as part of the Earth Observing System) and Odin, an instrument shared between astronomers and upper atmosphere researchers. (The name Odin is not an acronym, but the name of a character from the nordish mythology.) Odin has been operating in orbit since February 2001, EOS/MLS is planned to be launched in June 2004 with the Aura satellite.

This chapter discusses some aspects of limb sounding instruments. Section 5.1 discusses the retrieval of atmospheric temperature profiles, Section 5.2 presents a study with aircraft data, and Section 5.3 presents first simulations of the effect of cirrus clouds on the limb radiances.

5.1 Pointing and Temperature Retrieval

The retrieval of trace gas species profiles from limb sounder data requires good knowledge of the atmospheric temperature as well as the instrumental pointing direction. Temperature information can be obtained from external sources, for example climatological models. The pointing can be determined from star tracker data. However, the knowledge of pointing and the atmospheric temperature profile is not perfectly accurate, a fact which will have an impact on the quality of the trace gas retrieval. Alternatively, both temperature profile and information on the pointing can be retrieved directly from suitable features in the recorded atmospheric spectra.

The standard approach for microwave limb sounding instruments was the derivation of temperature and pointing information from oxygen emission lines [97, 90, 13]. The oxygen volume mixing ratio is constant and therefore the intensity of the line is only a function of atmospheric temperature and total pressure. The question raised in paper [H 9] is to what extent pointing and temperature information can be separated from unknown mixing ratios. With the technical development of microwave instruments, total bandwidths of 10 GHz are feasible in the future. Such wide bands observe several emission lines simultaneously and make the additional retrieval of temperature and pointing information possible.

The temperature information in that case comes primarily from the fact that the limb path becomes opaque for certain tangent altitudes and frequencies. In general, the limb path for spectral line centers becomes opaque at higher tangent altitudes, whereas the limb path for spectral line wings becomes opaque at much lower tangent altitudes only. In the opaque case there is a direct relation between the brightness temperature received by the instrument and the physical temperature of the atmosphere. Information on pointing, on the other hand, can be derived from line-widths, which are proportional to the total pressure. The retrieval can be performed with the Optimal Estimation Method (OEM) as

described, for example, by Rodgers [70, 71]. The solution provided by the OEM can be understood as a weighted mean of the measurement and a priori information, assuming Gaussian statistics. The OEM allows a rigorous error analysis. This includes an assessment of the resolution, of the statistical error, and of possible correlations with other retrieved parameters. The objective of the paper was to assess the capability of pointing and temperature retrieval from millimeter/submillimeter observations, independent of oxygen emission lines.

By a detailed error analysis, including both statistical and systematic errors, it was found that the simultaneous retrieval of one pointing parameter (bias) and atmospheric temperature is possible. For the selected simulation conditions and instrumental specifications, the pointing bias can be retrieved with a very good precision, corresponding to an altitude bias of less than 24 m at the tangent point. The temperature profile can be retrieved with a precision better than 2 K at altitudes ranging from 10 to 40 km.

The retrieval is sensitive to some model parameter errors (uncertainties in instrumental and spectroscopic parameters). The total instrumental error could be greatly reduced by a reduction of the calibration uncertainties. Regarding the errors caused by uncertainties in spectroscopic parameters, it was found that only a few lines are responsible for these errors. In all cases the error is so strongly dominated by a single line, that it would be reduced significantly if the parameters of that single line were perfectly known. This observation has in the meantime been exploited by a dedicated laboratory spectroscopy study, described in Perrin et al. [66] and Verdes et al. [89].

A comparison between the information derived from a band containing an oxygen line and one without an oxygen line showed that there is no qualitative difference between the two. This is easily explained by the fact that O_2 is not the main source of pointing and temperature information, even in bands containing O_2 lines, if the bands are broad.

If pointing bias and temperature are retrieved using only the information coming from an oxygen line, the results are poorer in retrieval precision and in instrumental errors. The magnitude of the total spectroscopic error is comparable (slightly lower, but this is probably a side effect of the poorer precision). It is indeed caused exclusively by the O_2 line parameters, which might be an advantage. However, the overall performance is so much poorer that this retrieval scheme does not seem to be a good alternative.

In summary, one can say that the pointing bias and temperature retrieval capabilities are good enough to allow the investigated instruments to fulfill their scientific objectives without external data. Dedicated oxygen bands appear to be unnecessary.

5.2 The Performance of a Planned Satellite Instrument Compared to an Existing Aircraft Instrument

The Superconducting Submillimeter-Wave Limb Emission Sounder (SMILES) [61] is an instrument concept designed to measure with exceptionally low noise stratospheric trace gases that have only weak spectroscopic signatures. Its unique feature is a superconductor-insulator-superconductor (SIS) mixer, a device that has so far not been operated in space since it works only if cooled down to 4.5 K. The first SMILES type instrument will be JEM / SMILES. It is planned to operate on the exposed facility of the Japanese Experiment Module (JEM) of the International Space Station (ISS) from the year 2008. Its bands have been selected to allow the measurement of various trace gases that are important for the understanding of stratospheric ozone chemistry, notably ozone itself including its isotopes, and several chlorine compounds.

The Airborne Submillimeter Radiometer (ASUR) [54] is an aircraft-borne submillimeter sensor used for studying stratospheric trace gases. Since its development in 1991 by the University of Bremen, ASUR has contributed to all major European Arctic ozone campaigns (EASOE, SESAME I+III, THESEO, THESEO-2000/ SOLVE, EUPLEX) as well as to several satellite validation experiments (MLS, GOME, ILAS, SAGE III, SCIAMACHY). In 2002 ASUR participated in the LEONID-MAC campaign on board the NASA DC-8 to study changes of mesospheric composition during the passage of the earth through comet dust trails. For detailed information see Crewell et al. [16], de Valk et al. [18], Sugita et al. [86], Fahey et al. [38], Bremer et al. [8], von Koenig et al. [91], and Kleinboehl et al. [55].

The ASUR instrument looks upward with a constant zenith angle of 78° and can be tuned over a wide frequency range from 604 GHz to 662 GHz. For the study described in paper [H 3] it was tuned to observe the planned JEM / SMILES bands. Like the planned SMILES, ASUR also features a state of the art SIS mixer. The cooling for the ASUR mixer is achieved by a cryostat with liquid helium, whereas for the space instrument SMILES it is achieved by a three-stage mechanical cooler, using helium in a closed cycle. Figure 5.3 shows simulated spectra for both instruments.

It was investigated whether the performance of an SIS submillimeter-wave radiometer near 650 GHz is consistent with model estimates, and whether the spectroscopic properties of the atmosphere in the frequency regions selected for SMILES are consistent with state of the art absorption and radiative transfer models. The results show that the measurement errors of ASUR meet the expectations raised by the SIS technique. This is consistent with previous findings [57].

As to the spectroscopy, we observed a much stronger pressure shift for the HCl lines near 626 GHz than reported in the databases. The incorrect value of the shift in the databases leads to a noticeable fit residual, as shown by Figure 5.4. Retrieving the value of the pressure shift along with the trace gas concentrations improves the residual considerably, as

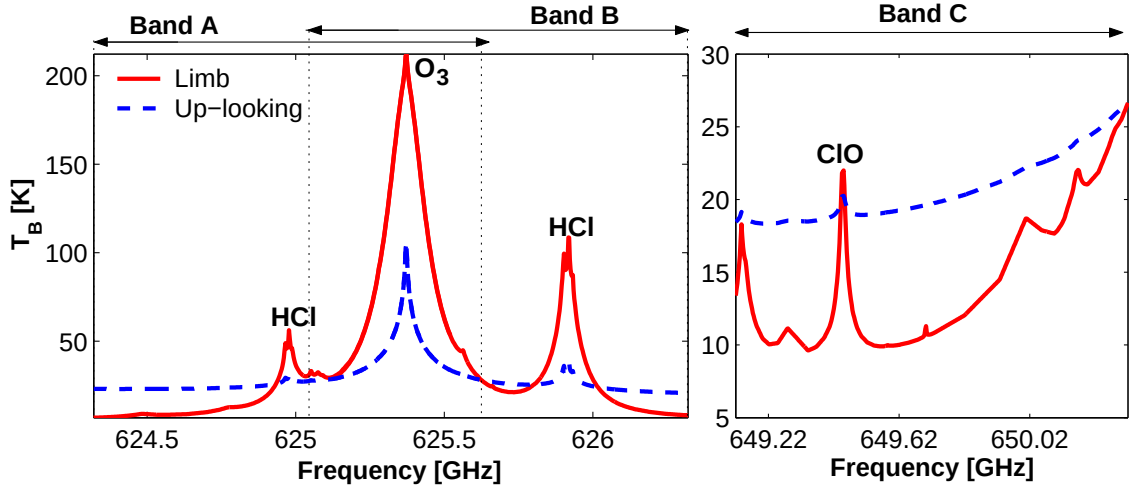


Figure 5.3: Simulated submillimeter brightness temperatures as a function of frequency. Solid: limb sensor for a tangent altitude of 25 km, antenna beam width 0.096° (HPBW, elevation). Dashed: uplooking sensor, flight altitude 10 km, zenith looking angle 78° . The main target species are indicated.

shown by the bottom plot of Figure 5.4. Depending on the assumed temperature dependence of the shift, the HCl pressure shift value consistent with the ASUR data is 0.090–0.117 MHz/hPa, instead of the 0.030 MHz/hPa reported in HITRAN [74]. This result is in good agreement with very recent independent laboratory work by Drouin [21], which suggests a value of 0.110 MHz/hPa for the shift.

Apart from the HCl pressure shift the observed spectroscopy in the designated SMILES bands seems to be consistent with current spectroscopic data.

5.3 The Effect of Cirrus Clouds on Microwave Limb Radiances

To improve existing climate models it is very important to extend the knowledge about cirrus cloud parameters, as such clouds cover more than 20% of the globe [93] and play an important role in the earth's radiation budget [1]. Depending on cloud altitude and microphysical properties, clouds can either cause warming or cooling at the earth's surface. So far clouds are not well treated in Global Climate Models (GCM) because of uncertainties concerning the properties of cirrus clouds and the complex interaction between radiation, microphysics, and dynamics in these clouds. Moreover, it is essential to consider clouds for the evaluation of limb measurements of trace gases in the upper troposphere. Measurements contaminated by clouds have to be detected and corrected appropriately (see for example Read et al. [68]).

The effective radius R_{eff} of the cloud particles is important for the cloud's radiative proper-

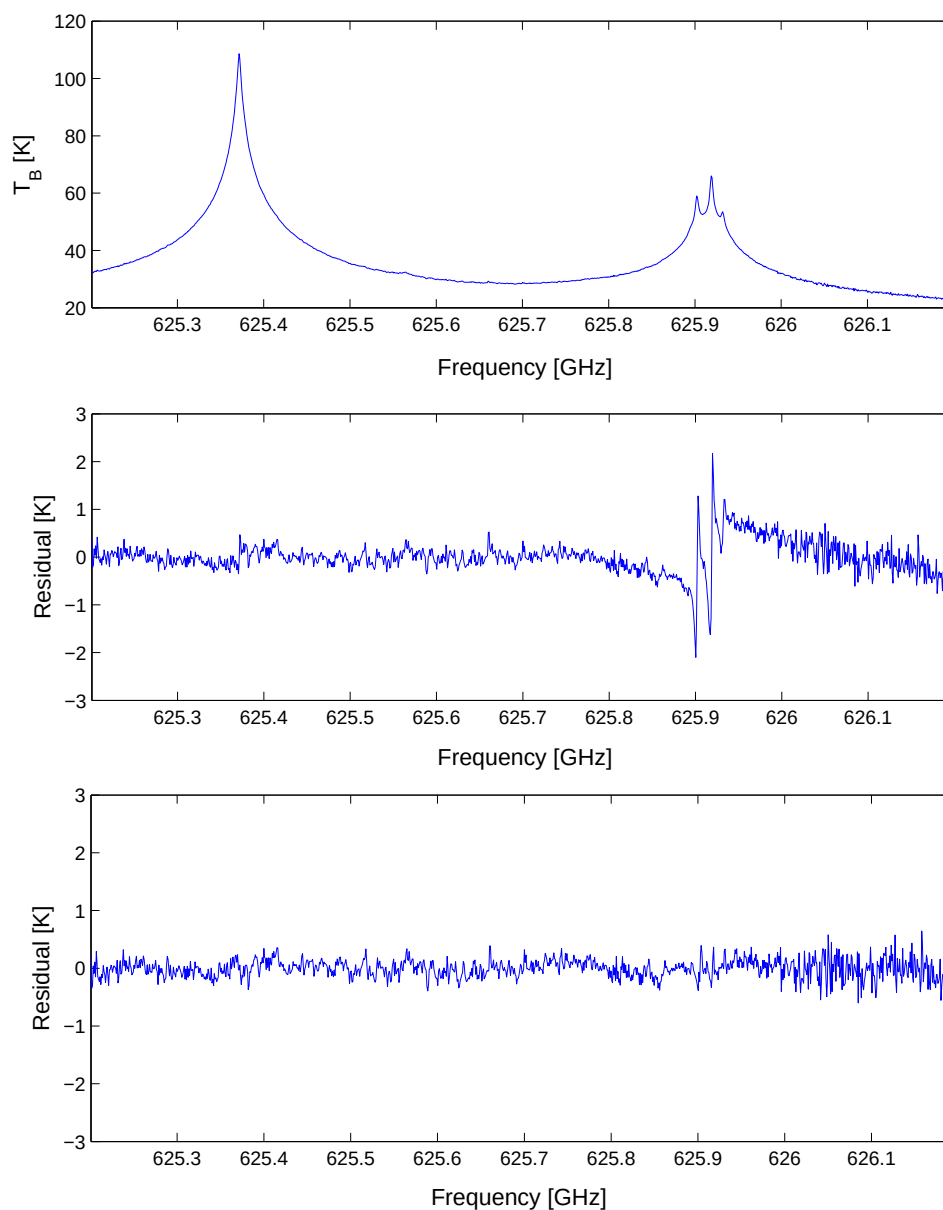


Figure 5.4: Top: measured ASUR spectrum in Band B for 19.11° E, 71.90° N, September 4, 2002, 16:14 UTC. Middle: residual (data—model) from a trace gas retrieval. The residual feature near 625.9 GHz is due to the incorrect pressure shift value in the database. Bottom: residual with pressure shift retrieval for the HCl lines.

ties. For a given frequency, R_{eff} largely determines the relation between ice mass content (IMC) and cloud optical thickness [36]. Parameterizations of R_{eff} have been retrieved from combined lidar and radar reflectivity [20] or from observations made in situ using aircraft-mounted instruments [53]. The Submillimeter-Wave Cloud Ice Radiometer (SWCIR) to fly on an aircraft has been developed to retrieve upper tropospheric IMC and R_{eff} [37].

Satellite remote sensing techniques in the thermal infrared can only be applied for thin cirrus clouds consisting of small ice particles as saturation is reached for moderate optical depths [85]. Only ice particle properties of the uppermost cloud layers can be measured. Disadvantages of visible and near-infrared solar reflection methods include that they cannot distinguish between ice and liquid water clouds and they cannot measure low optical depth clouds over brighter land surfaces.

There are several studies about the impact of cirrus clouds on microwave nadir radiances, for example Evans et al. [36] and Skofronik-Jackson et al. [77]. They show that the brightness temperature depression depends strongly on R_{eff} and IMC. Paper [H 7] is the first study of that kind for microwave limb radiances. Passive nadir-viewing techniques cannot sufficiently resolve the vertical distribution of IMC. Microwave limb sounding techniques can provide higher resolutions than nadir techniques for the same frequencies. In paper [H 7] the impact of clouds on microwave limb radiance measurements of the planned MASTER (Millimeter Wave Acquisitions for Stratosphere/Troposphere Exchange Research) [10] instrument is investigated.

A new version of the radiative transfer model ARTS (Atmospheric Radiative Transfer Simulator) is used for the simulations. The first version of ARTS is a one-dimensional (1D) unpolarized model, which is sufficiently fast for operational use but does not include scattering (see paper [H 5]). The new version was developed to simulate the influence of multiple scattering on measured radiances, polarization effects emerging from ice particle scattering, and in addition three-dimensional (3D) effects resulting from horizontal inhomogeneities in the atmosphere. It uses the scattering algorithm described in paper [H 8], but extended to spherical geometry so that the simulation of limb measurements is possible. The extended algorithm is described in detail in Emde et al. [24].

The model can be applied in the microwave and in the infrared. ARTS has been compared to a forward model developed at RAL (Rutherford Appleton Laboratory) and to the single scattering forward model KOPRA developed for MIPAS (Michelson Interferometer for Passive Atmospheric Sounding) [47]. ARTS and the RAL forward model show an excellent agreement (less than 1K difference in simulated brightness temperatures for most cloud cases) and ARTS and KOPRA agree well in the single scattering regime. For the study described in paper [H 7] the 1D unpolarized version of ARTS was used.

Figure 5.5 shows simulated limb spectra in the presence of a cirrus cloud from 10 to 12 km altitude with varying IMC and R_{eff} . Following Kinne et al. [53], it was assumed that IMC and R_{eff} are correlated. The IMC values used were $4.0 \cdot 10^{-5}$, $1.6 \cdot 10^{-3}$, $8.0 \cdot 10^{-3}$, $1.6 \cdot 10^{-2}$, and $4.0 \cdot 10^{-2} \text{ g m}^{-3}$, the associated R_{eff} values were 21.5, 34.0, 68.5, 85.5, and $128.5 \mu\text{m}$. The particles are assumed to be spherical, simulations for nonspherical particles are shown

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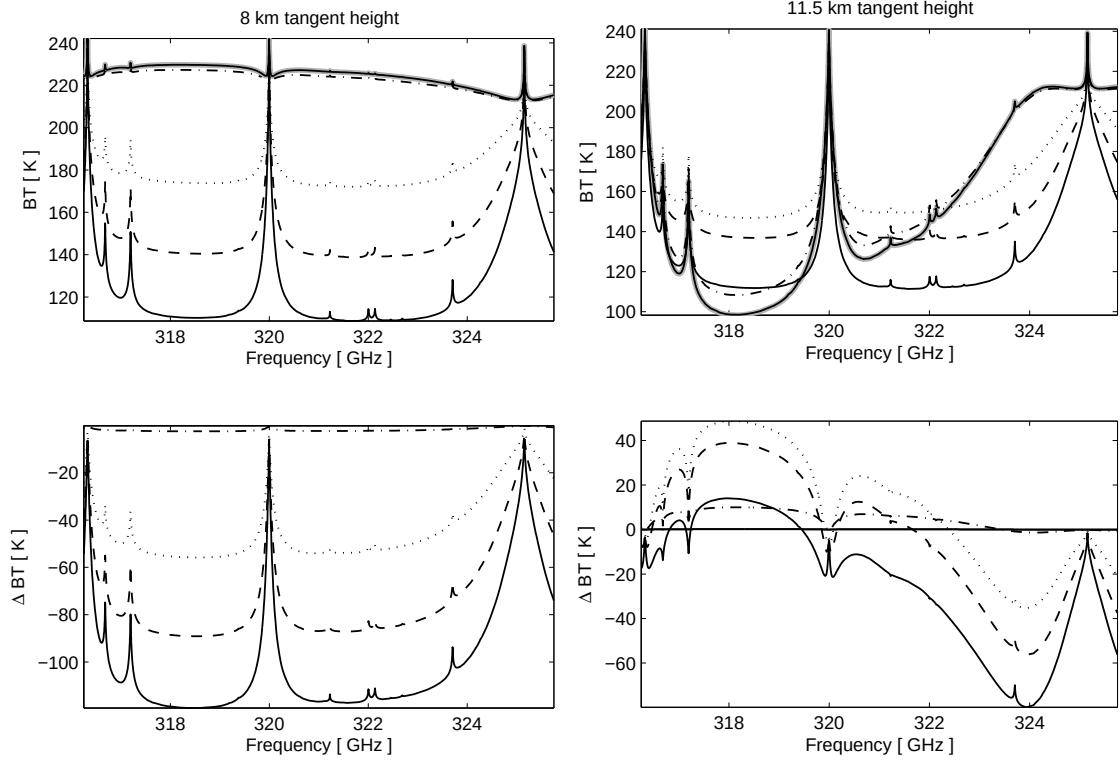


Figure 5.5: Left column: limb spectrum at 8 km tangent altitude. Right column: limb spectrum at 11.5 km tangent altitude. Clear sky spectrum (grey line) and cloudy spectra for IMCs of $4.0 \cdot 10^{-5} \text{ g m}^{-3}$ (solid), $1.6 \cdot 10^{-3} \text{ g m}^{-3}$ (dashed-dotted), $8.0 \cdot 10^{-3} \text{ g m}^{-3}$ (dotted), $1.6 \cdot 10^{-2} \text{ g m}^{-3}$ (dashed), $4.0 \cdot 10^{-2} \text{ g m}^{-3}$ (solid), and the corresponding particle sizes. Top plots show absolute brightness temperatures, bottom plots differences from the clear-sky case.

in Emde et al. [24].

The strongest cloud scattering effect can be observed in the microwave window regions. The total extinction coefficient consists of gaseous extinction and particle extinction (compare Equation 4.17). When the gaseous extinction dominates, the scattering effect becomes smaller. In the line centers there is no difference between cloudy and clear-sky spectrum. At these frequencies the water vapor opacity is so high that the transmission from the cloud to the sensor is practically zero, thus the existence of the cloud does not affect the measured radiance. As gas absorption is very high at low altitudes, the cloud altitude also has a big impact on the scattering signal. At low altitudes, gas absorption is dominant, therefore the scattering effect for low clouds is very small. Cirrus clouds can exist at altitudes above 10 km. Here gas absorption is low, so the scattering signal can be very large. The scattering coefficients increase with frequency, but the scattering effect does not necessarily increase with frequency, it also depends on the gas absorption characteristics of the considered frequency region.

The impact of the optically thinnest cloud with an IMC of $4.0 \cdot 10^{-5} \text{ g m}^{-3}$ and an R_{eff} of

$21.5\mu\text{m}$ is very small, the spectrum can not be distinguished from the clear-sky spectrum. The impact of the second thinnest cloud is also very small. For the optically thickest cloud with an IMC of 0.04 g m^{-3} and an R_{eff} of $128.5\mu\text{m}$ the brightness temperature (BT) depression reaches 120 K. The highest BT enhancement at 11.5 km is observed for the medium thick cloud with an IMC of $8.0\cdot 10^{-3}\text{ g m}^{-3}$ and an R_{eff} of $68.5\mu\text{m}$. The reason for this is that for very thick clouds there is a high probability that radiation scattered into the line of sight (LOS) in the lower or middle part of the cloud is again absorbed or scattered away from the LOS by a cloud particle. For thinner clouds most radiation that is scattered once into the LOS will continue to propagate into this direction without being disturbed by other cloud particles.

Although the microphysical cloud properties, i.e., particle size and IMC, are realistic, a limb instrument would observe smaller scattering signals. The reason is that the 1D mode of the model was taken for the study, which corresponds to a homogeneous cloud cover. This assumption is unrealistic, particularly for the intense cloud cases, which correspond to clouds of limited horizontal extent.

Overall, the study showed that both the ice mass content and the size of the cloud particles are important. When the effective particle size is doubled the BT depression can be enhanced by a factor of 9. To estimate in detail the effect of particle size and IMC a sensitivity study for more different sizes and IMC values is necessary. The strong impact of both, particle size and IMC, shows that in cloud retrievals it will be a challenge to obtain IMC and particle size simultaneously, as they both lead to an enhancement of the scattering effect. Particle shape effects will also play a role. For retrievals with clouds further investigations are required.

6 Summary and Outlook

This text and the selected publications have argued that accurate knowledge of the global distribution of water vapor, particularly in the upper troposphere, is crucial for our understanding of the climate system (Chapter 2). Studies of existing datasets (Chapter 3), radiative transfer modeling advancements (Chapter 4), and new sensor concepts (Chapter 5) have been presented.

The work described gives us the tools for global humidity climatology studies with operational microwave data. This undertaking has already been started in the framework of the AFO 2000 project UTH-MOS. Still much work lies ahead in the areas of intercalibrating the sensors on different satellites and in quantifying the cirrus cloud impact on humidity climatologies. The cirrus clouds themselves are a research area of growing interest. The Bremen group will participate or take the lead in proposals for new submillimeter wave cirrus sensors. An ESA funded radiative transfer study aimed at such a sensor is ongoing, in collaboration with the universities of Bonn, Edinburgh, and Chalmers.

Remote sensing from satellite platforms is the only way to obtain global data that are necessary to understand the climate system and predict its evolution. A lot of effort will be necessary to make the best use of the already available data record, and to deal with the wealth of new data that will be available from the upcoming sensor generation. It will be fun if we do it in the right way.

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