1. Introduction

The ubiquitous clouds of Venus can be divided into several layers, based on results from the Pioneer Venus entry probes (Esposito et al., 1983). The divisions are based on changes in cloud particle number density and particle size. In this scheme, Venus’ upper cloud reaches to an altitude of ~70 km, above which lies a layer of smaller haze particles. Before the arrival of Venus Express, Venus’ upper haze was probed through polarimetry (Kawabata et al., 1980; Esposito and Travis, 1982; Braak et al., 2002) and daytime limb scans (Lane and Opstbaum, 1983; Krasnopolskii, 1983). The divisions are based on changes in cloud particle number density and particle size. In this scheme, Venus’ upper haze was probed through polarimetry (Kawabata et al., 1980). The divisions are based on changes in cloud particle number density and particle size. In this scheme, Venus’ upper haze was probed through polarimetry.
Venus' low latitudes, and 47–50°N, which is at the edge of the ultraviolet-bright region (Titov et al., 2008). We used the same two latitude bins for all observations. We limit our analysis here to these two latitude bins, because the scattering calculations in our reevals are rather time-consuming. Future analysis will include a more complete mapping of the Venus limb. The latitude bins were chosen according to variability in the radiances from the tropospheric windows, i.e. we selected latitudes that had similar radiances in the tropospheric windows. We tried to bin observations from as wide range of latitudes as possible, whilst not being from dissimilar regions in the atmosphere. Fig. 2 shows an example of the data between 4.5 and 5 μm for the 47–50°N bin, which shows no large deviations with latitude.

After binning of the data it became apparent that there was a small, but significant, signal at altitudes that should reveal no signal at all. Fig. 3 shows the spectrum for one latitude bin, averaged between altitudes of 150 and 200 km. It shows clear structure, which was seen in all high altitude data from all the flybys analysed here. We attributed this to a small straylight effect. Although these radiances are small compared to typical nadir radiances, they are significant compared to limb radiances close to 90 km. Hence, we corrected all spectra by subtracting the spectrum averaged between 150 and 200 km in the corresponding latitude bin. Also visible is an odd–even effect that is also visible in some parts of the lower altitude spectra. It shows up as a saw-tooth pattern superimposed on the spectrum, giving the appearance of increased random noise. Work is ongoing within the VIRTIS team to correct for this systematic detector effect, but improvement of the VIRTIS calibration pipeline is beyond the scope of this paper. The odd–even effect results in slightly higher error bars on the retrievals, but it is a relatively minor effect. When repeating the retrievals with only every other wavelength included, the retrieval changed only by amounts that were well within the error bars.

Errors on the binned spectra were determined by dividing the error on one spectrum (see Piccioni et al. (2007) for the noise performance of the instrument) by the square root of the number of spectra in one bin.

3. Initial data assessment

Initial inspection of the VIRTIS limb spectra, including those not analysed here, showed a somewhat unexpected behaviour: even at altitudes as high as 85 km the infrared spectra often look very much like scaled downward-looking spectra (see Fig. 4). The limb spectra showed identical behaviour regardless of the slit orientation with respect to the limb and signals are higher than expected from stray light, so the limb spectra do not seem to be an instrument artefact. Surface and tropospheric emission bands can clearly be seen below 3 μm, even though the line-of-sight does not cross the surface, or even the upper cloud deck. Also above 3 μm the shape of the spectrum looks much like that coming from the cloud tops. At 85 km CO₂ becomes optically thin in the 3–5 μm region and hence the absorption features at 4.3 and 4.8 μm, seen as dips in the downward-looking spectra, will show up in emission in the limb if there is not scattering by particles (see also Fig. 8). The observed limb spectra do not show this behaviour and radiance levels far exceed those expected from a clear atmosphere.

### Table 1: Observations used. N is the total number of spectra between 65 and 90 km.

<table>
<thead>
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<th>Name</th>
<th>Date</th>
<th>Latitude (°N)</th>
<th>Local time (h)</th>
<th>N</th>
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<td>0</td>
<td>6742</td>
</tr>
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<td>24 December 2007</td>
<td>18–55</td>
<td>0</td>
<td>6616</td>
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<tr>
<td>VI0614_09</td>
<td>26 December 2007</td>
<td>18–55</td>
<td>0</td>
<td>1606</td>
</tr>
<tr>
<td>VI0628_09</td>
<td>9 January 2008</td>
<td>18–55</td>
<td>2</td>
<td>1493</td>
</tr>
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</table>
A physical explanation for these limb spectra is that thermal radiation from below is scattered into the line-of-sight by haze or cloud particles at high altitudes. Small sulfuric acid haze particles, which were thought to be the only particles above the cloud tops before the Venus Express mission, are very weak scatterers above 3 μm (see Fig. 5) and hence cannot reproduce the spectral behaviour seen in the VIRTIS limb spectra. Larger particles are more scattering at higher wavelengths, so the presence of larger particles above the upper cloud deck may explain the observed limb spectra.

We can test this by plotting the ratio of the limb spectra to the downward-looking spectra. The wavelength-dependent thermal radiation, $I(\lambda)$, coming from the limb can be written as the following integral over the line-of-sight:

$$ I(\lambda) = \int_{-\infty}^{\infty} J(\lambda) e^{-\tau(\lambda)} d\tau $$

Here, $\tau$ is the optical depth along the line-of-sight and $J(\lambda)$ is the source function, which contains a thermal emission part and a scattering part:

$$ J(\lambda) = (1 - \tilde{\omega}(\lambda))B(\lambda, T) + \tilde{\omega}(\lambda) \frac{1}{4\pi} \int_{-\pi}^{\pi} I(\lambda, \theta) P(\lambda, \theta) d\Omega $$

The emission part is determined by the blackbody radiation $B(\lambda, T)$, whereas the scattering part is determined by the internal radiation field $I(\lambda, \theta)$ that is scattered according to the phase function $P(\lambda, \theta)$ and integrated over solid angle $\Omega$. Both terms are weighted according to the single-scattering albedo $\tilde{\omega}(\lambda)$. As mentioned previously, VIRTIS nightside spectra of Venus’ limb at high altitudes seem to be dominated by the scattering part. If the optical depth along the line-of-sight is low, which it is around 85 km (see Fig. 7), Eq. (1) can be expanded around $\tau = 0$:

$$ I(\lambda) \approx J(\lambda) \tau(\lambda) $$

The Venus limb spectra seem to be dominated by radiation from below, scattered in the limb at an angle of 90° by an optical thin layer of particles. In this case, the received radiance is proportional to:

$$ I(\lambda) \propto I_{up}(\lambda) [\tilde{\omega}(\lambda) P(\lambda, 90°) \sigma(\lambda)] $$

where $\sigma(\lambda)$ is the extinction cross-section of the scattering particles. $I_{up}(\lambda)$ is the hemispherically averaged upward-going radiance, which will be heavily weighted towards radiances emerging with low emission angles. All terms in square brackets in this relation are only dependent on particle properties and hence the wavelength dependence of $I_{up}$ contains direct information about the particle in the case that scattering is dominating and optical thickness is small.

This ratio is plotted for observation VI0612_02, between 20 and 30°N and at 82.5 km in Fig. 6. Because the limb observations in Table 1 were taken at such close range, there were actually no spectra that have a low emission angle, which we then could use directly to estimate $I_{up}$. Instead, spectra with an emission angle of 85° were taken and a limb darkening correction was applied, based on Galileo NIMS observations and radiative transfer modeling. Above 3 μm the high emission angle spectrum was multiplied by a factor of 5 (Roos et al., 1993), and below 3 μm the spectrum was multiplied by a factor of 2.3 (R. Carlson, personal communication), which was confirmed by the model of Tsang et al. (2008). Also plotted are the products in square brackets of Eq. (4) for different sizes of sulfuric acid particles. The sulfuric acid concentration is taken to be 75% here, but increasing this to 85% has only little effect in the spectral regions where there are spectral features in the measurements. We used Mie theory, which is applicable to spherical particles, to calculate the various terms in Eq. (4) using the
optical constants of Palmer and Williams (1975). Particle sizes were distributed according to a log-normal distribution with constant log(variance) values in all cases. The lines were scaled to match the observations at 5 μm. It is clear from Fig. 6 that small particles cannot reproduce the observed spectra, since their wavelength dependence of II/Im is too steep at short wavelengths. Instead, particles with a mean radius of 1 μm are in excellent agreement with the observations. This particle size distribution happens to coincide exactly with that of the ‘mode 2’ particles of Grinspoon et al. (1993). So, VIRTIS limb spectra show that mode 2 particles are present high in the atmosphere, as also recently observed by Wilquet et al. (2009). We observe this behaviour for all observations at 82.5 km where optical thickness is low. For the low latitude observations, this behaviour is even seen as high as 85 km. The optical thickness of different tangent altitudes is illustrated in Fig. 7, which shows normalised contribution functions, i.e. they show where in the atmosphere the observation is sensitive to. For the three low altitudes the atmosphere is optically thick and one is sensitive to altitudes higher than the tangent altitude. Adding more haze there will increase the contribution of the upper (colder) atmosphere, hence this has a negative effect on the observed radiance. When the atmosphere becomes thicker the tangent altitude is probed directly. Adding more haze here will increase the amount of scattering of the upward radiation into the line-of-sight, hence increasing the observed radiance at these tangent altitudes. Having shown that the observed limb spectra can indeed be caused by scattering particles, we proceeded to retrieve vertical profiles of cloud particles using a radiative transfer model.

4. Retrieval method

We used the NEMESIS radiative transfer and retrieval code (Irwin et al., 2008b) to retrieve vertical profiles of cloud particles in Venus’ upper atmosphere from VIRTIS nightside limb spectra. NEMESIS has the option to include scattering as well as emission of thermal radiation in the limb. In this scattering model, the internal radiation field J in Eq. (2) is calculated using a doubling–adding code (e.g. Plass et al., 1973), which includes multiple scattering, albeit in a plane-parallel atmosphere. The internal radiation field is then scattered once in Eq. (1) in the proper spherical geometry.

In this paper, we used spectra between 4.3 and 5 μm for the retrievals. This region is most sensitive to large particles and hence gives useful constraints on the mode 2 vertical profile. Additional constraints on the smaller particles can possibly be obtained by including shorter wavelengths, but modelling is more complicated for this region, since the hot troposphere has to be taken into account and more variables come into play. Including shorter wavelength in the retrievals is therefore subject of future work.

We included opacity by CO₂, CO, H₂O gas and clouds. Since radiation in the 4.3–5 μm region emanates from the cloud tops these gases will be at Earth-like pressures and temperatures, so the regular HITRAN database (Rothman et al., 2005) was used to calculate opacity for gas lines. Voigt line shapes were used, except for CO₂ for which a sub-Lorenzian lineshape (Tonkov et al., 1996) was used. Our initial temperature profile is that of Seiff (1983) and our initial mode 2 vertical profile is based on that of Roos et al. (1993) and Grinspoon et al. (1993). Gases were assumed to be uniformly mixed with volume mixing ratios of 100 ppm for CO (Irwin et al., 2008a) and 2 ppm (Fedorova et al., 2008) for water. Note that in the altitude range of our retrievals the spectrum is dominated by scattered radiation from the cloud tops and hence the limb spectra are not sensitive to actual vertical variations of trace gases and temperatures at the tangent heights, but are mainly sensitive to the density of the scattering cloud particles.

When fitting, we added an additional error term that estimates systematic errors, in either model or spectra, by looking at the fit of the three lowest altitude spectra (below 80 km). With the initially estimated errors, we could not fit these spectra with high signal to within the error bars. In order to obtain more realistic error bars on our retrieved profiles, we quadratically added a constant error of 0.08 μW cm⁻² sr⁻¹ μm⁻¹ to all spectra, which makes the lower spectra fit with to:

$$\chi^2 = \frac{1}{N} \sum_{i=1}^{N} \frac{(y_{\text{meas}}(i) - y_{\text{mod}}(i))^2}{\epsilon(i)^2} \approx 1$$

Here, y_{\text{meas}} and y_{\text{mod}} are the measured and modelled spectral points and \( \epsilon \) is the error on the measurements.

Our retrieval process was then as follows. In order to obtain the correct upward spectrum from the cloud tops, we first retrieved a temperature profile using only one limb spectrum at a relatively low altitude (65 km). The retrieval using a low limb spectrum is effectively identical to temperature retrievals using nadir observations (Irwin et al., 2008a; Grassi et al., 2008), since the atmosphere is optically thick for both cases. The scattering cloud is not altered in this retrieval. Note that for the retrievals of particles above the cloud tops the concentration of clouds below the cloud top is not very important, as long as the upward radiation is modelled correctly. After retrieving temperature, the mode 2 vertical distribution was retrieved by fitting limb spectra between 65 and 90 km simultaneously. We did not include particles other than the mode 2 particles, since the previous section shows that these particles determine most of the spectra below 85 km for the low latitude observations. In the next section it is shown that at higher altitudes smaller particles can also be used to obtain good fits to the spectra. Hence the retrieved mode 2 profile above this altitude indicates an upper limit of the mode 2 particles.

5. Results

As can be seen in Fig. 8, the spectra can be generally fitted relatively well by the scattering mode 2 particles. Furthermore, we performed model calculations and calculated the ratio like in Fig. 6 and we find a very similar spectral behaviour for the high altitude mode 2 particles as estimated in Section 3. However, at the highest altitudes the structure in the spectra looks less like nadir spectra. For instance, the high altitude spectra show less pronounced local minima at 4.8 μm. Instead, the spectra show a more smooth increase with wavelength than the modelled spectrum, as indicated by the relatively bad fit between 4.45 and
4.75 μm. However, it must be noted that errors are large here. For these high altitudes, particles that are less scattering provide better fits to the spectra. Fig. 9 also shows a fit using particles that have a mean radius of 0.2 μm and a slightly adjusted temperature (dashed lines). In this case the thermal radiation from the haze at the tangent altitude is dominating, giving rise to a spectrum that is almost a scaled Planck function. Further analysis, using also shorter wavelengths, will be needed to disentangle contributions from mode 2 particles and smaller ones.

For some of the high latitude observations the transition from a scaled nadir spectrum to a smoother spectrum occurs at lower altitudes (see Fig. 9). This also causes relatively bad fits at some altitudes. Also the radiances are lower than for the low latitude cases at high altitudes. This results in low retrieved cloud
particle densities. However, it is likely that the mode 2 densities are even lower in this case and we are seeing smaller particles as well.

The retrieved profiles for the four observations are shown in Fig. 10. Fig. 11 shows the errors on the retrieved profiles. The vertical resolution of the profiles is generally 2 km, which is caused by a combination of the contribution functions (Fig. 7) and the smoothing applied in the retrieval process. The particle number density is decreasing monotonously with height for all profiles, mainly caused by the change in atmospheric density. Generally, the high latitude particle density above 80 km is lower than that at low latitudes and they are more variable than at low latitudes. For the profiles with the lowest densities, also the spectral shape is smoother at these high altitudes, indicating relatively low concentrations of scattering mode 2 particles. Below ~75 km we have less constraint on the cloud number density, since the atmosphere becomes optically thick, and it becomes difficult to disen-tangle temperature effects from cloud effects.

![Fig. 10. Retrieved mode 2 number densities for the four observations between 20 and 30°N and between 47 and 50°N. The numbers in the legend refer to the orbit numbers of the observations (see Table 1).](image1)

![Fig. 11. Errors on the retrieved profiles of Fig. 10. The line styles indicate the same observations as in that figure.](image2)
6. Conclusion and discussion

Nightside infrared spectra of Venus’ limb show thermal radiation from below the tangent altitude that is scattered into the line-of-sight by 1 µm-sized particles, up to altitudes as high as 85 km. These mode 2 particles at such high altitudes were not expected before the Venus Express mission, but recent solar occultation measurements at high northern latitudes by SPICAV/SOIR also indicate the existence of these particles at high altitudes (Wilquet et al., 2009). Our number densities of the mode 2 particles are similar in magnitude to those of Wilquet et al. (2009), although our number densities are higher at 75 km. This is expected, since Wilquet et al. (2009) probe polar latitudes, where cloud tops are observed to be lower (Titov et al., 2008; Ignatiev et al., 2009).

Our results nicely complement previous measurements, since we probe at the nightside and cover a range of latitudes simultaneously. Also, the infrared wavelengths covered by VIRTIS are more sensitive to the large particles than observations at visible or UV wavelengths. It shows that the boundary between the upper cloud and the upper haze layer is not as abrupt as previously thought, but that the mode 2 particle density is gradually decreasing with altitude.

We have analysed two latitude regions: one between 20 and 30°N and one between 47 and 50°N. The 20–30°N bin should be indicative of low latitudes, whereas the 47–50°N bin probes the edge of the ultraviolet-bright region. The higher latitude particle number density profiles show more variability between orbits than the low latitudes. Also, the number densities above 80 km are generally lower. This behaviour is consistent with behaviour seen in images of ultraviolet brightness and cloud top altitudes in the southern hemisphere (Titov et al., 2008; Ignatiev et al., 2009). Latitudes around 50° form a boundary, above which the atmosphere becomes ultraviolet-bright and clouds tops are lower.

From our retrieved particle number densities, scale heights of the clouds can be determined directly. If we evaluate each model layer individually, we obtain a mean cloud scale height of 2.8 km between 75 and 90 km, with a variance of at least 1.2 km for each profile. The high variance is mostly caused by the high altitudes, where noise is relatively high. Around 75 km the scale height is consistent with a value of 3.3 ± 0.3 km for all observations presented here. These values are consistent with the mode 2 scale height values determined previously for the cloud tops based on limb darkening of the same wavelength interval by Roos et al. (1993) (4.1 ± 0.6 km) and Roos-Serote et al. (1996) (3.9 ± 1 km). However, also considering a possible dominance by smaller particles above 85 km, a slight decrease of mode 2 scale height with altitude in the upper atmosphere would fit our profiles better.

Acknowledgments

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References


