Abstract

A number of significant scientific changes have been made to the RTTOV fast radiative transfer model. A major new feature of RTTOV is the inclusion of a parameterization of multiple scattering that is performed by scaling the optical depths by a factor derived by including the backward scattering in the emission of a layer and in the transmission between levels. The introduction of multiple scattering allows RTTOV to simulate AIRS and IASI radiances in presence of eleven different types of aerosol components, five different types of water clouds and 30 different types of ice clouds. In RTTOV, ice clouds can be assumed to be composed either of randomly oriented hexagonal ice crystals or ice aggregate. RTTOV can now simulate IASI and AIRS radiances which include CO, CH$_4$, N$_2$O and CO$_2$ as profile variables and a solar term to evaluate the solar radiance reflected by a land or a wind roughened water surface. For a land surface, the reflecting surface is treated as a perfect diffuser following the Lambert law whereas for a full-gravity-capillary wave water surface the bidirectional reflectivity of the surface is computed explicitly for any geometry using a wave model to evaluate the variance of the wave slope. The treatment of the angular dependence of the optical depths has also been improved by introducing an altitude dependent value of the local zenith angle that takes into account the curvature of the Earth and its surrounding atmosphere. To improve the accuracy of the radiance computation in presence of optically thick layers, RTTOV features a new parameterization of the Planck function based on the linear in $\tau$ assumption that the source function throughout the layer is linear with the optical depth of the layer. Finally, to solve the radiative transfer for an atmosphere partially covered by clouds, RTTOV uses a scheme (stream method) that divides the field of view into a number of homogeneous columns, each column containing either cloud-free layers or totally cloudy layers. Each column is assigned a fractional coverage and the number of columns is determined by the cloud overlapping assumption (maximum-random overlap in RTTOV) and the number of layers the atmosphere is divided into. The total radiance is then obtained as the sum of the radiances for the single columns weighted by the column fractional coverage. It is planned that these upgrades are included in the next release of RTTOV, RTTOV-9.

Introduction

A prerequisite for exploiting radiances from conventional and high-resolution sounders in Numerical Weather Prediction (NWP), is the availability of a fast and accurate radiative transfer model to predict a first guess radiance from the model fields (temperature, water vapour, ozone, surface emissivity and perhaps clouds in the future).

In this paper we describe a new version of the fast radiative model currently operational at ECMWF, RTTOV. The new model is based on the work of Matricardi (2003) and Matricardi (2005). The updated
version of RTTOV allows for variable profiles of CO$_2$, CO, N$_2$O, and CH$_4$ and features the inclusion of a solar term to evaluate the contribution of the solar radiance reflected by a land or water surface. A new model has been introduced for the prediction of the water continuum absorption and a vertical pressure grid with increased number of levels is now used for the computation of radiances for the infrared high resolution sounders. Multiple scattering in the infrared has been introduced using a parameterization (scaling approximation) that scales the optical depth by a factor derived by including the backward scattering in the emission of a layer and in the transmission between levels. In the new RTTOV, optical properties are available for eleven aerosols components, five different types of water clouds and cirrus clouds. For cirrus clouds, optical properties are available either for hexagonal ice crystals or ice aggregates. Finally, the accuracy of the scaling approximation has been assessed by comparing approximate radiances with references radiances computed by using a doubling-adding algorithm.

The formulation of the fast radiative transfer model

The basic methods applied in the development of RTTOV are documented in Saunders et al. (1999) and Matricardi et al. (2001). In this section only the main components are discussed. Any major change to the model is documented in detail in Matricardi (2003) and Matricardi (2005).

In the RTTOV fast transmittance model the computation of the optical depth for the layer from pressure level $j$ to space along a path at angle $\theta$ involves a polynomial with terms that are functions of temperature, absorber amount, pressure, and viewing angle. The effective optical depth for channel $i$ from level $j$ to space can be written as:

$$\hat{\rho}_{j,i} = \hat{\rho}_{j-1,i} + \sum_{k=1}^{M} a_{j-1,i,k} X_{k,j-1}$$

(1)

where $M$ is the number of predictors and the functions $X_{k,j}$ constitute the profile-dependent predictors of the fast transmittance model. To compute the expansion coefficients $a_{j,i,k}$ (sometimes referred to as fast transmittance coefficients), a set of diverse atmospheric profiles is used to compute, for each profile and for several viewing angles, and for various absorbing constituents, accurate LBL transmittances for each level defined in the atmospheric pressure layer grid. The effective optical depths $\hat{\rho}_{j,i}$ are then used to compute the $a_{j,i,k}$ coefficients by linear regression of $\hat{\rho}_{j,i} - \hat{\rho}_{j-1,i}$ versus the predictor values calculated from the profile variables for each profile at each viewing angle.

In RTTOV H$_2$O, CO$_2$, O$_3$, N$_2$O, CO, CH$_4$ are allowed to vary, the other gases are held constant and will be referred to as fixed. Gases are considered as fixed if their spatial and temporal concentration variations do not contribute significantly to the observed radiance. In RTTOV we define as fixed gases N$_2$, O$_2$, HNO$_3$, OCS, CCl$_4$, CF$_4$, CCl$_2$F (CFC-11) and CCl$_2$F$_2$ (CFC-12).

Regression coefficients for RTTOV are generated from a database of LBL transmittances computed using the GENLN2 line-by-line (LBL) model (Edwards, 1994). In RTTOV the molecular parameters for the LBL computations are taken from the year 2000 release of the HITRAN database (Rothman et al. 2003).
The model for the prediction of water vapour continuum

For water vapour the continuum type absorption is of particular importance. In RTTOV the regression for the water continuum is handled separately from the other gases. The advantage of having a separate fast model for the continuum is that any change in the continuum coefficients can be addressed without the need of generating a new LBL database. A considerable amount of time and disk space can then be saved. It also allows the reduction of the number of predictors used in the water vapour model helping improving the accuracy of the regression since the interaction of some of the predictors can cause numerical instabilities in the results of the regression.

In RTTOV the water continuum transmittance is parameterized using the model described in Matricardi (2003). Regression coefficients are generated using a database of monochromatic transmittances computed using version 2.4 of the CKD continuum model (Clough et al, 1989). The prediction of the continuum optical depths in RTTOV is performed using a total of four predictors: two for the self-continuum absorption and two for the foreign-continuum absorption.

Inclusion of trace gases CO$_2$, N$_2$O, CO and CH$_4$ as profile variables in RTTOV

In RTTOV, CO$_2$, N$_2$O, CO and CH$_4$ profiles are allowed to vary and are profile variables in the fast model with H$_2$O and O$_3$. For each of the gases allowed to vary, the profiles used to compute the database of LBL transmittances are chosen to represent the range of variations in absorber amount found in the real atmosphere and should be representative of the gas observed behavior. The trace gases profiles used in RTTOV are a blend of profiles from in-situ measurements and profiles generated using chemistry models.

The CO$_2$ profiles were assembled assuming that the vertical distribution of this gas is constant in the troposphere and decreases by 5 to 10 ppmv between the tropopause and about 22 km altitude (see Bischof et al, 1985). No further change is assumed above this layer. CO$_2$ profiles were generated on the basis of the season/latitude classification used for the original temperature/humidity 43 profile set. For the generation of the N$_2$O profiles we assumed N$_2$O is well mixed in the troposphere. The N$_2$O profile set was generated from Cryogenic Limb Array Etalon Spectrometer (CLAES) (Reber et al. 1993) measurements in the stratosphere and from CMDL and AGAGE measurements at the surface (for further information see http://www.cmdl.noaa.gov and http://www.cdiac.ornl.gov/ndps/alegage.html). Profiles from CLAES were joined by parabolae to a constant tropospheric mixing ratio based on measurements obtained at the surface.

The CO$_2$ profile set was generated assembling profiles based on MOZART 3D model calculations (Brasseur et al., 1998; Cunnold, 2001) and measurements taken during the STRATOZ III and TROPOZ II campaign in the Austral summer, 4-26 June 1984, and in the Austral winter, 9 January to 1 February 1991 (Marenco et al., 1995). Since no stratospheric data were available from these sources, mixing ratios in the troposphere were extrapolated to the stratosphere assuming a lapse rate equal to the one from the corresponding seasonal USAFGL CO profile.

For CH$_4$, profile concentrations in the troposphere are based on the IMAGES model calculations (Müller and Brasseur, 1995; Clerbaux et al., 1998). The profile set covers the seasonal cycle of the gas. Although the latitudinal gradient was retained, absolute values at all levels were scaled to reflect recent estimates of surface values from measurements made at the stations of the CMDL network. Tropospheric mixing ratios were joined by parabola to stratospheric measurements made by CLAES.
Finally, details of the predictors used for the CO₂, N₂O, CO and CH₄ transmittance model can be found in Matricardi (2003).

The formulation of the model for the solar term

In RTTOV we have introduced a solar term to evaluate the effect of solar radiance that is transmitted through the atmosphere and then partially reflected back upward through the atmosphere to the receiver. For the case of solar radiance reflected by a land surface, a proper treatment of the solar term would then require the knowledge of the bi-directional reflectance of the considered surface. Given that the bi-directional reflectance is not currently available in RTTOV, we treat the reflecting surface as a perfect diffuser following the Lambert law. For a Lambertian surface the bidirectional reflectance is constant and is equal to the surface albedo. If we assume that the atmosphere along the downward and upward path is the same, the solar radiance \( \hat{L}_\psi \) that reaches the detector can be written as

\[
\hat{L}_\psi = \frac{1}{\pi} \rho^L \hat{I}^\oplus \mu_\odot \hat{t}_\psi (\mu_{\text{eff}})
\]

(2)

where \( \rho^L \) is the surface albedo, \( \hat{I}^\oplus \) is the solar irradiance at the top of the atmosphere and \( \hat{t}_\psi (\mu_{\text{eff}}) \) is the surface-to-space transmittance for the path angle \( \mu_{\text{eff}} \) defined as

\[
\frac{1}{\mu_{\text{eff}}} = \frac{1}{\mu} + \frac{1}{\mu_\odot}
\]

(3)

This is equivalent to say that the reflected solar radiance depends on a single transmittance whose secant is equal to the sum of the secants of the viewing (\( \mu \)) and solar zenith angles (\( \mu_\odot \)).

For the case of solar radiance reflected by a wind roughened water surface, the reflective characteristics of the wind roughened water surface are modeled following the approach by Yoshimori et al. (1995). In this model the probability density \( P \) of the wave slope obeys a Gaussian distribution whereas the spectrum of the wave slope is specified by the Joint North Sea Wave Project (JONSWAP) (Hasselmann et al.1973) wave-spectral model. The total variance of the slope is given by

\[
\sigma_{\text{tot}}^2 = \int_0^\infty \Psi(\omega)[k(\omega)]^2 d\omega
\]

(4)

where \( k(\omega) \) is the inverse function of the dispersive relation of the full-gravity capillary wave and \( \Psi(\omega) \) is the frequency spectrum of the surface wave as specified by the JONSWAP wave model.

In the model by Yoshimori et al. (1995) shadowing (the fact that the slopes on the back sides of the waves and deep in the troughs between waves are hidden from view) is treated explicitly and allows the estimate of the reflected solar radiance for large solar zenith angles.

The solar source function used in RTTOV is based on theoretical radiative transfer calculations for the solar atmosphere made by Kurucz (1992). In the infrared spectral region it is strongly dependent on measurements made by the ATMOS instrument on the Space Shuttle.
For the case of solar radiance reflected by a water surface the solar radiance that reaches the detector can be written as

\[
\hat{L}_{\nu} = \hat{I}_{\nu} \circ \tilde{w}_{\nu}(\mu, \varphi, \mu_{\oplus}, \varphi_{\oplus}) \tilde{L}_{\nu}(\mu_{\text{eff}})
\]  

(5)

Here \( \tilde{w}_{\nu}(\mu, \varphi, \mu_{\oplus}, \varphi_{\oplus}) \) is the effective reflectivity of the wind roughened water surface whereas \( \varphi \) is the azimuth angle of the receiver and \( \varphi_{\oplus} \) is the azimuth angle of the sun.

The computation of \( \tilde{L}_{\nu}(\mu_{\text{eff}}) \) in equations (2) and (5) requires the evaluation of transmittances at large zenith angles. For the fast transmittance model to be able to simulate transmittances for a wider range of zenith angles, we have extended the database of line-by-line transmittances by computing data for an additional number of eight more zenith angles, namely, the angles for which the secant assumes the following values: 2.58, 3.04, 3.72, 4.83, 6.1, 7.2, 9, 12. This extended range allows evaluating the solar term for zenith angles as large as \( \approx 85^\circ \). The additional database of line-by-line transmittances was generated only for the shortwave channels (\( \tilde{\nu} \geq 2000 \text{ cm}^{-1} \)).

Since the larger range of zenith angles increases the difficulty of fitting the line-by-line optical depths, we have developed a dedicated transmittance model for the shortwave. The new model uses a revised and larger number of predictors (Matricardi 2003).

### The radiative transfer for multiple scattering

The scheme we have introduced in RTTOV to parameterize multiple scattering is based on the approach followed by Chou et al. (1999). In this scheme (referred to hereafter as scaling approximation), the effect of scattering is parameterized by scaling the optical depth by a factor derived by including the backward scattering in the emission of a layer and in the transmission between levels. Since this parameterisation of multiple scattering rests on the hypothesis that the diffuse radiance field is isotropic and can be approximated by the Planck function, we can expect it to have an effect on the accuracy of the radiance calculations. However, the scaling approximation does not require explicit calculations of multiple scattering and since the radiative transfer equation can be written in the same form as in clear sky conditions, the computational efficiency of RTTOV can be retained. In the scaling approximation, the absorption optical depth, \( \tau_a \), is replaced by an effective extinction optical depth, \( \tau_e \), defined as:

\[
\tau_e = \tau_a + b \tau_s
\]  

(6)

where \( \tau_s \) is the scattering optical depth and \( b \) is the integrated fraction of energy scattered backward for incident radiation from above or below. If \( \bar{P}(\mu, \mu') \) is the azimuthally averaged value of the phase function and \( \mu \) is the cosine of the scattering angle, \( b \) can be written in the form:

\[
b = \frac{1}{2} \int_0^1 \int_{-1}^0 \bar{P}(\mu, \mu')d\mu' \]

(7)
The introduction of scattering in RTTOV required the introduction of an improved parameterization of the Planck function, $B$. In the previous version of RTTOV a layer average value of the Planck function was used that gives equal weight to the radiance emitted from all regions within the layer. In presence of optically thick clouds this would put too much weight on the radiance coming from the lower regions of the layer. To improve the accuracy of radiance calculations in RTTOV, we have introduced a new parameterisation of the Planck function based on the linear in $\tau$ assumption that the atmospheric emission source function throughout the layer is linear with the optical depth of the layer:

$$B(T(\tau)) = B_u + (B_1 - B_0) \frac{\tau}{\tau'}$$

(8)

Here $B_u$ is the Planck function for the top of the layer, $B_1$ is the Planck function at the bottom of the layer and $\tau'$ is the optical depth of the layer. The parameterisation is exact at the top ($\tau = 0$) and bottom ($\tau = \tau'$) of the layer. When the new formulation of the source function was applied to RTTOV, we found that the error introduced by the use of the mean layer source function could be as large as 2K for an optically thick water cloud.

The dataset of optical properties for aerosols

The optical properties for the aerosols included in RTTOV were computed using the microphysical properties assembled in the Optical Properties of Aerosols and Clouds (OPAC) software package (Hess et al. 1998). This database provides the microphysical properties (i.e. size distribution and refractive indices) for ten aerosol components. To be able to simulate the radiative properties of the atmosphere in presence of a volcanic eruption, we have supplemented the OPAC database with the microphysical properties of the volcanic ash component.

The database of optical properties used by RTTOV was generated using the Lorentz-Mie theory (Van de Hulst 1981) for spherical particles. A lognormal size distribution was used for all the aerosol components except for the volcanic ash component for which a modified Gamma distribution was used. Values of the phase function were computed for every 0.1º from 0º to 3º otherwise they are given for every 1º. For those aerosols that can take up water, we have computed the optical properties for eight different values of the relative humidity assuming the width of the distribution does not change. The optical properties for an arbitrary value of the relative humidity can then by obtained by linear interpolation. Aerosols can be classified in terms of their location (i.e. continental, maritime, polar, etc.) and type (i.e. clear, polluted, desert, urban, etc.) Aerosols are formed as a mixture of several aerosol components. RTTOV can compute optical properties for any mixture of aerosol components or, alternatively, it can compute optical properties for ten pre-defined mixtures of aerosol components, each mixture forming a typical climatological aerosol.

The dataset of optical properties for water clouds

The OPAC package gives the microphysical properties for 5 types of water clouds: two stratus clouds (Stratus Continental and Stratus maritime) and 3 cumulus clouds. Analogous to the database of optical properties for aerosols, we have generated a database of optical properties for every single water cloud type using the Lorentz-Mie theory assuming a modified Gamma size distribution.
The dataset of optical properties for cirrus clouds

Cirrus clouds are made of ice crystals. Typical shapes for ice crystals include bullet rosettes, hollow and solid columns, plates and aggregates. A shape often used in the literature is the hexagonal prism either in the form of column or plate. An exact solution for the interaction of a plane wave with a hexagonal ice crystal cannot be sought using the Lorentz-Mie theory. The problem is complicated by the fact that there is no practical solution that can be used to cover for all the crystal sizes that occur in the Earth’s atmosphere.

In RTTOV, optical parameters are available for ice clouds made of hexagonal ice crystals or ice aggregates. A composite database of optical properties for hexagonal ice crystals randomly oriented in space was generated using the geometric optics (GO) method (Macke et al. 1996) for large crystals and the T-matrix method (Kahnert et al. 2001) for small crystals. The database of optical properties for randomly oriented ice aggregates used in RTTOV is described in Baran and Francis (2004). The exact aggregate geometry is due to Yang and Liou (1998) and consists of eight hexagonal columns attached to one other. In the Yang and Liou model the aspect ratio (i.e. the ratio of the major and minor axes of a circumscribed ellipse) does not vary with size and is close to unity. For wavelengths longer than 5 μm the database described in Baran and Francis (2004) has been computed using an approximate method that represents the complex geometry of the ice aggregates by a size/shape distribution of circular ice cylinders (Baran 2003). Simulated ice aggregates are assumed to be composed of ensembles of circular cylinders. The aspect ratio (i.e. diameter-to-length ratio) of the cylinders can vary but the volume-to-area ratio of the aggregate is conserved and is equal to the value of the exact aggregate geometry due to Yang and Liou (1998). For cylinders with maximum dimensions of 3-225 μm the optical parameters of the size/shape distribution of circular ice cylinders are computed using the T-matrix method (Mishchenko 1991) whereas for cylinders with maximum dimension up to 3500 μm optical parameters are computed using the complex angular momentum (CAM) method (Nussenz and Wiscombe 1980). For wavelengths less than 5 μm the database is supplemented with optical parameters computed using the exact model due to Yang and Liou (1998). Note that the values of the phase function included in this database were not computed analytically but derived instead using a modified Henyey-Greenstein phase function (Baran et al. 2001).

To represent the microphysical properties of ice clouds, in RTTOV we used the size distributions prepared Fu (1996). The size distribution functions used in RTTOV are plotted in Figure 1. These size distributions are representative of cirrus clouds from midlatitude regions (Heymsfield and Platt, Takano and Liou, FIRE I, FIRE II, and FU) and from tropical regions (CEPEX IWC and CEPEX). All the size distributions have been obtained from in situ aircraft observations. However, for the size distributions measured in midlatitude regions, the technique used (optical array probes) could not allow the measure of particles smaller than 20-40 μm whereas the replicator sonde used in the tropical regions could measure the small ice crystals that cannot be measured by the optical array probes. To account for the radiative effect due to the presence of small ice crystals, the midlatitude size distributions have been extrapolated to small values of the particle size assuming that the logarithm of the number concentration varies linearly with the logarithm of the particle dimension (Heymsfield and Platt 1984).
To better resolve the structure of the spectra we have discretized the size distribution into 24 bins with the midpoint crystal length, $L$, varying from 4μm to 3500μm. The width, $D$, of the crystal has been derived from the length, $L$, of the crystal using the aspect ratio given in Yang et al. (2003).

The data from the 30 size distributions have been used to parameterize the absorption coefficient, the scattering coefficient, the extinction coefficient and the backscatter parameter, $b$, as a function of the ice water content, $IWC$, and the generalized effective diameter, $D_{ge}$ (Fu 1996). It can be shown that in the limit of geometric optics (i.e. the size of the crystal is much larger than the wavelength of the incident radiation), since the extinction cross section is twice the projected area $A_c$, the extinction coefficient $\beta_e$ can be related to $IWC$ and $D_{ge}$ in the form

$$\beta_e = \frac{4\sqrt{3} IWC}{3\rho_i D_{ge}}$$  \hspace{1cm} (9)$$

Although a theoretical relationship cannot be derived for the more general case, we can still expect that the extinction coefficient, the scattering coefficient, the absorption coefficient, the backscatter parameter and, possibly, the phase function can be related to the ice water content and the generalized diameter. In RTTOV we assume that...
\[
\frac{\beta_e}{IWC} = r_{e,0} + r_{e,1}D_{ge} + \frac{r_{e,2}}{D_{ge}} + \frac{r_{e,3}}{D_{ge}^2} \tag{10}
\]

\[
\frac{\beta_s}{IWC} = r_{s,0} + r_{s,1}D_{ge} + \frac{r_{s,2}}{D_{ge}} + \frac{r_{s,3}}{D_{ge}^2} \tag{11}
\]

\[
\frac{\beta_a}{IWC} = r_{a,0} + r_{a,1}D_{ge} + \frac{r_{a,2}}{D_{ge}} + \frac{r_{a,3}}{D_{ge}^2} \tag{12}
\]

\[b = r_{b,0} + r_{b,1}D_{ge} + r_{b,2}D_{ge}^2 + r_{b,3}D_{ge}^3 \tag{13}\]

where \(\beta_e\) is the extinction coefficient, \(\beta_s\) is the scattering coefficient, \(\beta_a\) is the absorption coefficient and \(b\) is the backscatter parameter.

The stream method

To solve the radiative transfer for a horizontally non-homogeneous atmosphere (i.e. an atmosphere partially covered by clouds) we follow the approach (referred to hereafter as stream method) of dividing the atmosphere into a number of homogeneous columns (Amorati and Rizzi 2002). Each column is characterized by a different number of cloudy layers, hence different radiative properties, and contributes to a fraction of the overcast radiance that depends on the cloud overlapping assumption. To describe the stream method we give here an example where seven atmospheric layers are considered. Once the cloud fractional cover in each layer is known (\(CFR\) in Figure 2), we compute the cumulative cloud coverage, \(N_{\text{tot}}(j)\), from layer 1 to layer \(j\) using the maximum-random overlap assumption. For a slab extending from layer 1 to layer \(j\), the total cloud cover is written as

\[N_{\text{tot}} = 1 - (1 - N_{1}) \prod_{i=2}^{j} \frac{1 - \max(N_{i-1},N_i)}{1 - N_{i-1}} \tag{14}\]

A cloud (blue shaded region in Figure 1) is then placed in layer \(j\) that covers the range from \(N_{\text{tot}}(j) - CFR(j)\) to \(N_{\text{tot}}(j)\). To determine the number of columns, we consider all the cloud configurations that result in a totally overcast column. In our example we obtain five, \(n_c\), columns and one clear column. Once the top of the atmosphere radiance has been computed for each homogeneous column, the cloudy radiance is written as the sum of all the single column radiances weighted by the column fractional coverage

\[L^{\text{cloudy}} = \sum_{i=1}^{n_c} (X_{s,i} - X_i)L^{\text{overcast}} + L^{\text{clear}}(1 - X_{n_c+1}) \tag{15}\]

Note that if a clear column is present, this will be given a weight equal to \((1 - X_{n_c+1})\).
The results for aerosols

The accuracy of the scaling approximation has been studied by comparing approximate radiances with reference radiances obtained using a doubling-adding algorithm. Results shown in this paper have been obtained by adapting to the infrared the code originally developed by Bauer (2002) for the microwave. Approximate and reference radiances have been computed using line-by-line optical depths for the molecular species whereas for aerosol and clouds we have used optical depths from the respective databases assuming the optical depth is constant within the width of a channel. The accuracy of the scaling approximation is expressed in terms of the difference between approximated and reference radiances in presence of a scattering medium. To assess the impact of each aerosol/cloud on the top of the atmosphere radiance we have also computed the difference between clear-sky and reference spectra.

For each climatological aerosol we have computed spectra for a tropical and arctic profile. To obtain the aerosol vertical profile we have applied the methods described in Hess et al. (1998). Since these profiles are representative of global average conditions we have computed additional spectra assuming a value of the number density four times the global average value.

The desert aerosol type has by far the largest impact on the radiance. Figure 3 shows that for the extreme condition case (bottom panel) the presence of desert dust in a tropical profile can result in a reduction of the top of the atmosphere radiance (black curve) by 4K in the thermal infrared and by 1.8K in the short wave. Smaller differences are observed for the average condition case. Errors introduced by the scaling approximation (red curve) are less than 1K in the thermal infrared and less than 0.25 K in the short wave. For the Urban aerosol, the radiance attenuation for the extreme condition case can be as large as 1 K in the thermal infrared and 0.6 K in the short wave whereas for the average concentration case the radiance attenuation is typically less than 0.2 K for the whole spectrum. For the other aerosols the radiance attenuation seldom exceeds 0.1 K for the average condition case and 0.5 K for the extreme condition case. Errors introduced by the scaling approximation never exceed 0.1 K for the Urban aerosol whereas for the other aerosol they are typically below 0.05 K.
Fig. 3: The radiative impact of the Desert aerosol type (black curve) and the error introduced by the scaling approximation (red curve) for the tropical profile for two different aerosol number densities.

The results for water clouds

Results for the Stratus Continental type are shown in Figure 4 for a tropical profile.

Fig. 4: The radiative impact of the Stratus Continental cloud type (black curve) and the error introduced by the scaling approximation (red curve) for the tropical profile for two different values of the cloud thickness.
The radiance attenuation resulting from the introduction of the cloud is larger in the shortwave and can reach 16 K for the tropical profile and 11 K for the arctic profile. No appreciable difference can be observed when the cloud thickness is doubled, thus suggesting that the cloud is rendered opaque by extinction. The error introduced by the scaling approximation for the tropical profile is less than 1 K in the thermal infrared and can be as large as 5 K in the short wave. Smaller values are observed for the arctic profile.

For the Stratus Maritime cloud, a smaller radiance attenuation is observed. The error introduced by the scaling approximation is still less than 1 K in the thermal infrared but it is now significantly smaller in the short wave where it does not exceed 3 K. This can be partly explained by the fact that in the short wave the scaling factor $b$ for the Stratus Maritime cloud is smaller than that for the Stratus Continental type.

Results for the Cumulus Continental Clean and the Cumulus Continental Polluted show that the radiance attenuation can be as large as 40 K in the short wave and 30 K in the thermal infrared. The error introduced by the scaling approximation does not exceed 2 K in the thermal infrared and is less than 7 K in the short wave. A smaller radiance attenuation is observed for the Cumulus Maritime cloud mainly in the shortwave. For this cloud type the error introduced by the scaling approximation is significantly lower. It is smaller than 0.8 K in the thermal infrared and does not exceed 1.2 K in the short wave. In fact, among the middle level cloud types, the Cumulus Maritime cloud is characterized by a smaller value of the scaling parameter $b$ across the whole spectral range.

The results for cirrus clouds

The spectra for the cirrus cloud types have been computed by placing the top height of the cloud at 12 km, 10 km and 7 km. Results plotted in Figure 5 show that for the 12 km case the radiance attenuation can reach 10 K in the thermal infrared and 5 K in the short wave. Because of the moderate optical thickness of the cirrus cloud, doubling the thickness of the cloud results in a considerably larger attenuation.

![Fig. 5: The radiative impact of the Cirrus cloud type (black curve) and the error introduced by the scaling approximation (red curve) for the tropical profile for two different values of the cloud thickness.](image)
A remarkable feature is the accuracy achieved by the scaling approximation. The error never exceeds 0.5 K. This can be justified by the very small values of the $b$ parameter. Note how for the cirrus cloud case the scaling approximation tends to underestimate the radiance in contrast to what happens for the aerosols and water clouds where the scaling approximation always overestimates the radiance. For the arctic profile case the error introduced by the scaling approximation is below 0.1 K. For the lower cloud height of 10 km we observe a reduction of the radiance departure; less than 5 K for the tropical profile and less than 0.3 K for the arctic profile. The scaling approximation now introduces an error that is typically less than 0.2 K.

## Conclusions

A new version of RTTOV has been developed. The new RTTOV features a revised vertical pressure grid that allows the integration of the radiative transfer equation to be performed with significantly increased accuracy. The water vapour transmittance model has been significantly improved by introducing a dedicated transmittance model for the continuum absorption. In the new RTTTOV profiles of CO$_2$, N$_2$O, CO, and CH$_4$ are allowed to vary and a solar term has been introduced to evaluate the solar radiance reflected by a land or water surface.

A parameterization of multiple scattering has been introduced by scaling the optical depth by a factor derived by including the backward scattering in the emission of a layer and in the transmission between levels.

RTTOV can include by default eleven aerosol components, five water cloud types and eight cirrus cloud types. The database of optical properties for aerosols and water droplets has been generated using the Lorentz-Mie theory for spherical particles using the microphysical parameters assembled in the OPAC software package. Cirrus clouds are assumed to be composed of hexagonal ice crystals randomly oriented in space. For these particles, a database of optical properties has been generated using the geometric optics (GO) method for large crystals and the T-matrix method for small crystals.

To improve the accuracy of the radiance computation in presence of optically thick layers, RTTOV features a new parameterisation of the Planck function based on the 'linear in ' assumption that the source function throughout the layer is linear with the optical depth of the layer. When compared to the parameterization used in the previous version of the code, the application of the new source function results in top of the atmosphere radiance differences up to 2 K in cloudy conditions.

To solve the radiative transfer for a partly cloudy atmosphere, a scheme (the stream method) has been implemented that divides the atmosphere into a number of homogeneous columns. Each column is characterized by different radiative properties and contributes to a fraction of the overcast radiance that depends on the cloud overlapping assumption.

To assess the accuracy of the scaling approximation we have compared approximate radiances with reference radiances computed by using a doubling-adding algorithm. Results have been obtained for each aerosol and cloud type for a tropical and arctic profile. For the desert aerosol, errors introduced by the scaling approximation are less than 1 K in the thermal infrared and less than 0.25 K in the short wave when a concentration four times the climatological value is assumed. For the Urban aerosol errors never exceed 0.1 K whereas for the other aerosols errors are typically below 0.05 K. For the low level clouds the error introduced by the scaling approximation for the tropical profile is less than 1 K in the thermal infrared and can be as large as 5 K in the short wave for the Stratus Continental cloud. For the Stratus Maritime cloud the error introduced by the scaling approximation is still less than 1 K in the thermal infrared and does not
exceed 3 K in the short wave. Results for the middle level clouds show that the error introduced by the scaling approximation does not exceed 2 K in the thermal infrared and is less than 7 K in the short wave for the Cumulus Continental Clean and Cumulus Continental Polluted clouds. For the Cumulus Maritime cloud the error introduced by the scaling approximation is significantly lower. It is smaller than 0.8 K in the thermal infrared and does not exceed 1.2 K in the short wave. Finally, for the cirrus cloud types we found a remarkable agreement between approximate and reference radiances. For the tropical profile the error introduced by the scaling approximation never exceeds 0.5 K whereas for the arctic profile is typically below 0.1 K.

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