Fast radiative transfer models for assimilation of radiance observation: issues for the next generation of advanced sounders

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Abstract

A fast radiative transfer model, RTIASI, has been developed at the European Centre for Medium-Range Weather Forecasts (ECMWF) for prelaunch simulation studies of Infrared Atmospheric Sounding Interferometer (IASI) data and for the exploitation of IASI radiances within the framework of a numerical weather prediction (NWP) variational analysis scheme. The fast model for IASI uses profile-dependent predictors to parameterize the atmospheric transmittances. In this paper we give a description of a number of features of RTIASI that are relevant to the fast radiative transfer models for the next generation of advanced sounders and discuss some of the most relevant issues related to the spectroscopy used to generate the parameterized transmittances.

1. Introduction

A potentially useful addition to the current satellite sounders is the Infrared Atmospheric Sounding Interferometer (IASI) (Cayla 1993). As part of the preparations being made at ECMWF to exploit the IASI high-spectral resolution datasets, RTIASI, the ECMWF fast radiative transfer model for IASI, has been developed (Matricardi and Saunders 1999, Matricardi 2003). RTIASI contains a model where the transmittances of the atmospheric gases are expressed as a function of profile dependent predictors. The accuracy of RTIASI is largely determined by the quality of the transmittance parameterization and by the spectroscopy used to generate the parameterized transmittances.

Spectroscopic errors are introduced by spectroscopic parameters such as line width, line strength and the temperature dependence of the line width. The most important source of spectroscopic errors is the spectral line shape. This is especially true for high-spectral resolution sounders and in particular for molecular species that are characterized by very high optical depths, such as H_2O and CO_2.

Since the inclusion of minor gases in the state vector can be fully exploited by high-spectral resolution sounders such as IASI, in RTIASI the profiles of water vapour (H_2O), ozone (O_3), carbon monoxide (CO), carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O) can be varied. The vertical resolution of IASI requires that not to limit the accuracy of the radiative transfer equation, the numerical integration of the former be performed on a adequate number of levels. In RTIASI, 90 pressure levels that extend from 1050 hPa to 0.005 hPa are used. The accuracy of the radiance computation in presence of optically thick layers is improved by introducing a parameterization of the Planck function based on the linear in J assumption that the Planck function throughout an atmospheric layer is linear with the optical depth of the layer. To enable the assimilation of high-resolution radiances in the short wave region of the infrared spectrum, a solar term has been introduced in RTIASI that covers the spectral region from 5 :m to 3.6 :m and solar zenith angles from 0° to 8°. Multiple scattering in presence of clouds and aerosols is parameterized 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. Finally, to solve the radiative transfer for an atmosphere partially covered by clouds, we have implemented a scheme where the instrument field of view (FOV) is divided into a number of homogeneous columns. Each column is assigned an area coverage and contains a different number of either cloud-free layers or totally cloudy layers. Once the top of the atmosphere radiance is computed for each column, the total radiance is obtained as the sum of all the radiances weighted by the column area coverage.

2. The formulation of the fast radiative transfer model

The basic methods applied to develop RTIASI are documented in Matricardi and Saunders (1999) and Matricardi (2003). In this section, only the main components are discussed.

In RTIASI the atmosphere is divided into 89 layers defined by 90 pressure levels from 0.005 hPa to 1050 hPa and is assumed to be plane-parallel in local thermodynamic equilibrium. The model uses the polychromatic form of the radiative transfer equation (i.e. convolved transmittances are used in the radiative transfer equation based on the assumption that this is equivalent to the convolution of the monochromatic radiances).

RTIASI contains a fast transmittance model where the computation of the transmittances involves a polynomial with terms that are functions of temperature, absorber amount, pressure, and viewing angle. To generate the database of parameterized transmittances, the GENLN2 (Edwards, 1994) line-by-line model (LBL) is used. The molecular parameters are taken from the year 2000 version of the HITRAN database (Rothman et al. 2003).

In RTIASI the profiles of H_2O, CO_2, O_3, N_2O, CO and CH_4 can be varied whereas the profiles of the other gases included in the LBL computations (N_2, O_2, HNO_3, OCS, CCl_4, CF_4, CCl_3F (CFC-11) and CCl_2F_2 (CFC-12)) are held constant.

Multiple scattering is parameterized by expressing the extinction optical depth, τ_e, in the form

$$\tau_e = \tau_a + b\,\tau_s \tag{1}$$

where τ_a is the absorption optical depth, τ_s is the scattering optical depth and

$$b = \frac{1}{2}\int_0^1 d\mu \int_{-1}^0 \overline{P}(\mu,\mu')\,d\mu' \tag{2}$$

is the integrated fraction of energy scattered backward for incident radiation from above or below. Here $\overline{P}(\mu,\mu')$ is the azimuthally averaged phase function and $\mu\,(\mu')$ is the cosine of the zenith angle of the incident (scattered) radiation.

3. The atmospheric layering

The layering choice made for the numerical integration of the radiative transfer equation should be in principle determined by the vertical resolution of the instrument: high-spectral resolution sounders require a finer vertical grid. In principle, the greater the number of layers the higher the accuracy of the radiance computation. On the other hand, the execution time of the LBL computations and the size of the associated database depends largely on the choice of the number layers. There is no need however to go to a disproportionate number of layers since we can consider not being relevant any gain in accuracy that is below the noise level of the instrument.

The vertical levels used for low-resolution sounders do not usually exceed the number of ~43 (e.g. RTTOV, see Saunders et al. 1999, for details). For IASI we found that radiative transfer errors introduced by the atmospheric layering can be kept below the instrument noise using ~90 pressure levels.

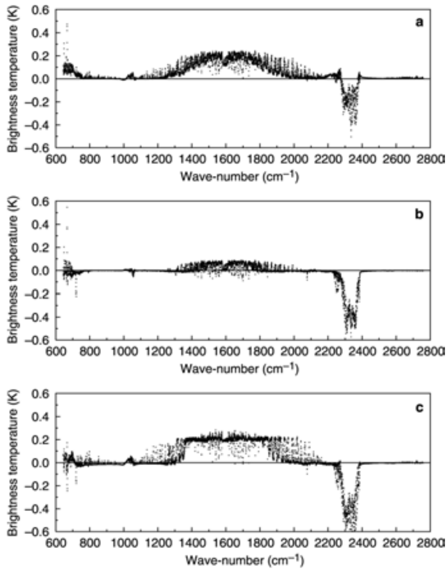

Figure 1: Difference between simulated IASI spectra obtained using 43 and 90 vertical pressure levels. Results are shown for (a) Mid-latitude spectrum; (b) Artic spectrum; (c) Tropical spectrum.

The impact of the choice of layering on LBL radiances is shown in figure 1 where the difference between simulated IASI spectra computed using the RTTOV 43 layer grid and the RTIASI 90 level grid is plotted for three different atmospheres. Larger differences are observed in the 6.3 μm (1594 cm^{-1}) water vapour band and in the 4.3 μm (2325 cm^{-1}) band with a lesser impact in the 15 μm (660 cm^{-1}) band. It is evident from Fig.1 that differences depend on the atmospheric state and in particular on the vertical gradients of water vapour and temperature in the upper troposphere and lower stratosphere.

4. The introduction of scattering in RTIASI

The radiative transfer in presence of multiple scattering can be solved using numerical solutions (i.e. discrete ordinate and doubling adding method). However, within the framework of RTIASI we can only consider analytical solutions given by approximate methods since numerical solutions are too computationally expensive. In principle the approximation known as two-stream could be used (Liou 2002). However, since RTIASI uses polychromatic transmittances, we found that, in this form, the two-stream approximation is not amenable to incorporation in RTIASI since this would result in too large errors.

On this basis we have looked into an alternative treatment of multiple scattering in RTIASI that would allows us to retain some degree of accuracy without compromising the computational efficiency of the code. The scheme we have introduced is based on the approach followed by Chou et al. (1999). In this scheme (henceforth referred to as scaling approximation), the effect of scattering by clouds and aerosols 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 (see Eq. (2)). Since this parameterization of multiple scattering rests of 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 has the advantage that the form of the radiative transfer equation does not differs too much from the one used in clear sky conditions and thus the computational efficiency of the code is not significantly degraded.

RTIASI includes eleven aerosol components and five water cloud types. The database of optical properties for aerosols and water droplets has been generated using the Lorentz-Mie theory (Van de Hulst 1981) adopting the dataset of microphysical properties assembled in the Optical Properties of Aerosols and Clouds (OPAC) software package (Hess et al. 1998). RTIASI-5 can also includes eight different types of cirrus clouds. Cirrus clouds are assumed composed of hexagonal ice crystals randomly oriented in space. For RTIASI we have generated a composite database of optical properties for ice crystals that uses the geometric optics (GO) method for large crystals (Macke et al. 1996) and the T-matrix method for small crystals (Kahnert et al. 2001).

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. The desert aerosol type has the largest impact on the radiance. The inclusion of desert dust in a tropical profile can result in a reduction of the top of the atmosphere radiance by 4K in the thermal infrared and by 1.8K in the short wave if an above the average concentration is assumed. For this case, errors introduced by the scaling approximation are less than 1K in the thermal infrared and less than 0.25 K in the short wave. For the urban aerosol type the radiance attenuation can be as large as 1 K in the thermal infrared and 0.6 K in the short wave. For the other aerosol types a smaller radiance attenuation is observed; this seldom exceeds 0.1 K for the average concentration case and 0.5 K for the above the average concentration case. Errors introduced by the scaling approximation are typically small. For the Urban type these never exceeds 0.1 K whereas for the other aerosol types they are typically below 0.05 K. For all aerosol types, errors observed in the thermal infrared are larger than the ones observed in the short wave.

For the low level clouds (Stratus Continental and Stratus Maritime) 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. 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. 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 (Cumulus Continental Clean, Cumulus Continental Polluted and Cumulus Maritime) 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 for the Cumulus Continental Clean and Cumulus Continental Polluted. For the Cumulus Maritime type a smaller radiance attenuation is observed above all in the shortwave. Te error introduced by the scaling approximation is significantly lower for this cloud type. 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 type 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.

5. The stream method

The radiative transfer equation in RTIASI is formulated for plane-parallel atmospheres (i.e. horizontally homogeneous). For an atmosphere partially covered by clouds, we have in principle to solve the radiative transfer for a horizontally non-homogeneous atmosphere. A widely used approach is the reduction of a partly cloudy layer to an equivalent homogeneous layer. This is achieved by re-distributing the optical thickness of the cloud over the entire layer in such a way that the reflectance of the cloud is the same as the one for the partly cloudy layer. The implementation of this technique would require a non-opaque cloud with fractional cover *CFR* to be treated as an opaque cloud (i.e. a black body) with an effective cloud fraction equal to *CFR* times the emissivity of the cloud. The alternative approach is to divide the atmosphere into a number of homogeneous columns, each column containing either cloud-free layers or completely cloudy layers. Despite being potentially more accurate, this technique (henceforth referred to as "stream method") has seldom been exploited (Amorati and Rizzi 2002) since depending on the number of layers and the overlapping assumption, the number of columns can become so large that it is impractical to use when lower orders approximations are used for multiple scattering (i.e two stream method). However, given the nature of the multiple scattering parameterization used in RTIASI, the computational burden of the stream method is only a fraction of the total and consequently we have implemented it in RTIASI. Note that preliminary results obtained from data collected during a number of field campaigns (Rizzi 2004) suggest that radiances computed using the stream method are in better agreement with measurement than those computed using the grey cloud approach.

In the stream method, the atmosphere is divided into a number of homogeneous columns. Each column is assigned a fractional area coverage that depends on which cloud overlap assumption has been made (the maximum-random overlap is used in RTIASI) and is characterized by different radiative properties since it can contain a different number of either cloud-free layers or completely cloudy layers. Once the top of the atmosphere radiance is computed for each column, the total radiance is obtained as the sum of all the radiances weighted by the column area coverage.

6. Source function for atmospheric emission

The accuracy of the radiative transfer equation depends on the way we parameterize the Planck function in the atmospheric emission source term. It is generally assumed that the atmospheric layer can be considered sufficiently optically thin that equal weight can be given to radiance emitted from all regions within the layer, i.e. the value of the Planck function evaluated at the average temperature of the layer is used. In presence of optically thick clouds one expects only the upper regions of the layer to give a significant contribute to the radiance. In this case, the use of the average value of the Planck function would put too much weight on the radiance coming from the lower regions of the layer. To improve the accuracy of radiance calculations in RTIASI, we have introduced a parameterization of the Planck function based on the linear in τ assumption that the source function throughout the layer is linear with the optical depth of the layer:

$$B[T(\tau)] = B_u + (B_1 - B_0)\frac{\tau}{\tau^*} \tag{3}$$

where 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 τ^* is the optical depth of the layer. The parameterization is exact at the top ($\tau = 0$) and bottom ($\tau = \tau^*$) of the layer. The new formulation of the layer source function has been applied to RTIASI and top of the atmosphere brightness temperatures, B_τ, have been compared with the ones, B_m, resulting from the application of the mean layer source function. Results are shown in Figures 2 and 3 where the difference $B_m - B_\tau$ is plotted for a tropical and arctic profile. In clear sky conditions the application of the new source function results in differences across the spectrum up to 0.15 K and, with the exception of the 4.2 μ m band, smaller values are found for the arctic profile. For both profiles, the sign of the departures changes for the stratospheric channels (the temperature of the bottom of the layer is greater than the temperature of the top of the layer) and larger errors are found for the bands with larger optical depths. When a low cloud with cloud top height of 2 km is introduced in a 320-m-thick layer, a different behavior is observed with departures significantly larger than the ones for the clear sky case. As shown in Figure 3, the application of the new source function now results in differences across the spectrum up to 0.7 K with smaller values found for the arctic profile. As expected, the presence of the cloud does not affect the spectrum for the stratospheric channels and larger differences are found in the bands with smaller optical depths. For these bands, the change of the sign of the departures for the arctic profile is the result of a temperature inversion in the layer where the cloud was placed. Results shown in Figure 3 can serve as a guide to the improved accuracy that can be achieved by introducing the linear in τ source function in the radiative transfer in presence of clouds. As shown in Amorati and Rizzi (2002), the error introduced by the mean layer source function increases with the optical depth of the cloud and tends to an asymptotic value for large optical depths. Given the vertical pressure grid used in RTIASI-5, we expect this error not to exceed 2 K.

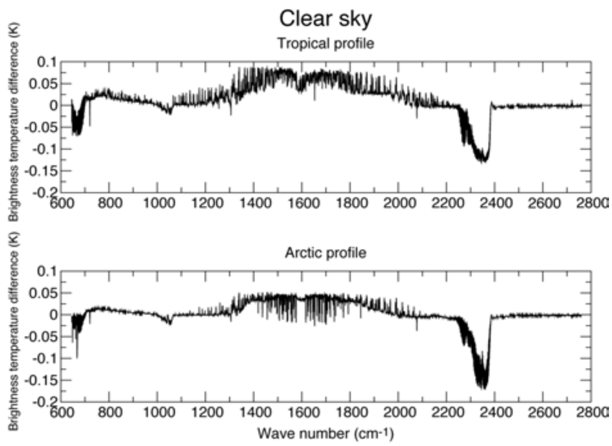

Figure 2: Differences in the top of the atmosphere brightness temperatures calculated using RTIASI with the mean layer and linear in τ source function definitions.

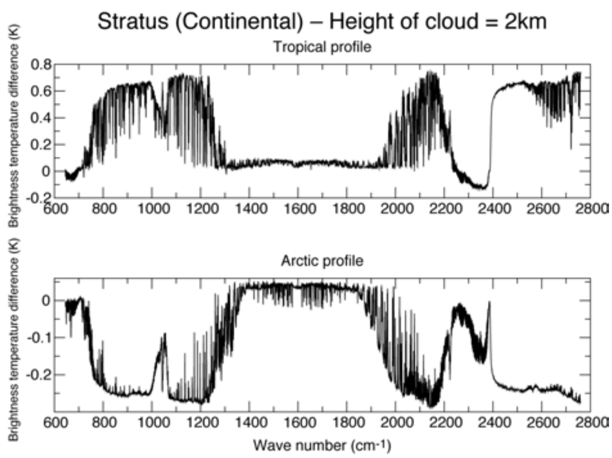

Figure 3:. Differences in the top of the atmosphere brightness temperatures calculated using RTIASI with the mean layer and linear in τ source function definitions.

7. Inclusion of CO_2, N_2O, CO and CH_4 trace gases as profile variables in RTIASI

To exploit the inclusion of trace gases in the state vector, RTIASI features variable profiles of CO_2, N_2O, CO and CH_4 . The trace gases profiles used to generate the database of LBL transmittances are selected to represent the range of variations in absorber amount found in the real atmosphere and I are a blend of profiles from in-situ measurements and profiles generated using chemistry models.

The CO_2 profiles are 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.

For the generation of the N_2O profiles we assume N_2O is well mixed in the troposphere. The N_2O profile set is 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 are joined by parabolae to a constant tropospheric mixing ratio based on measurements obtained at the surface.

The CO profile set is 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 campaigns (Marenco et al., 1995). Since no stratospheric data are available from these sources, mixing ratios in the troposphere are 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 is retained, absolute values at all levels are scaled to reflect recent estimates of surface values from measurements made at the stations of the CMDL network. Tropospheric mixing ratios are joined by parabola to stratospheric measurements made by CLAES.

8. The formulation of the model for the solar term

To enable the assimilation of high-resolution radiances in the short wave region of the infrared spectrum, a solar term has been introduced in RTIASI that covers the spectral region from 5 : m to 3.6 : m and solar zenith anglesd from 0° to 87°. 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 RTIASI, we treat the reflecting surface as a perfect diffuser following the Lambert law.

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. Since shadowing is treated explicitly, the model allows the estimate of the reflected solar radiance for large solar zenith angles. If we have the same atmosphere along the satellite and solar zenith angle, the solar radiance that reaches the detector after reflection by a water surface can be written in the form

$$\hat{I}_{\tilde{\nu}^*} \cong \hat{L}_{\tilde{\nu}^*}^{\oplus} \quad w_{\tilde{\nu}^*}(\mu, \varphi, \mu_{\oplus}, \varphi_{\oplus}) \quad \hat{\Gamma}_{\tilde{\nu}^*}(\mu_{eff}) \tag{4}$$

Here $\hat{L}^{\oplus}_{\tilde{v}^*}$ is the solar irradiance at the top of the atmosphere, $w_{\tilde{v}^*}(\mu,\varphi,\mu_{\oplus,}\varphi_{\oplus})$ is the effective reflectivity of the wind roughened water surface, $\hat{\Gamma}_{\tilde{v}^*}(\mu_{eff})$ is the surface-to-space transmittance, φ is the azimuth angle of the receiver, $\varphi_{\oplus}$ is the azimuth angle of the sun and μ_{eff} defined as

$$\frac{1}{\mu_{eff}}=\frac{1}{\mu}+\frac{1}{\mu_{\oplus}} \tag{5}$$

where μ is the cosine of the satellite viewing angle and μ_{eff} is the cosine of the solar zenith angle. The computation of $\hat{\Gamma}_{\tilde{v}}(\mu_{eff})$ in Eq. (4) requires the evaluation of transmittances at large zenith angles. Since the larger range of zenith angles increases the difficulty of fitting the LBL optical depths, we have developed a dedicated transmittance model for the shortwave that allows evaluating the solar term for solar zenith angles as large as ~87°. The new model uses a revised and larger number of predictors (Matricardi 2003).

9. Spectroscopic errors

Spectroscopic errors in LBL computations are introduced by spectroscopic parameters such as line width, line strength and the temperature dependence of the line width. Line widths and strengths are typically known within 10% accuracy. The temperature dependence of the line widths is known within 20% accuracy but is a less important source of error than line widths and strengths. In principle, LBL models should be updated on a regular basis with the latest line parameters available from molecular databases.

The most important source of error in the LBL computations is the spectral line shape. This is especially true for the high-spectral resolution sounders in that they allow the use of channels in-between the centers of absorption lines. Since high-spectral resolution sounders are particularly sensitive to the line shape of molecular species that are characterized by high atmospheric optical depths, we need a good knowledge of the line shape of species such as H_2O and CO_2. For CO_2, efforts have been recently made to improve the quality of the line shape (Strow et al. 2003). These include the formulation of Q-branch and P/R-branches line mixing in the 4.3μm and 15μm bands.

Although great improvements have been made in the knowledge of the H2O line shape in the strong 6.7 :m water vapour band, more work is still needed above all in the spectral region (7 to 8 :m) influenced by lower tropospheric H_2O. In fact, given the long path lengths involved, laboratory measurements are very difficult and because of the opacity of the atmosphere in this spectral region, often is too difficult to obtain good ground based measurements.

For water vapour, no theoretical breakthrough has yet been made to explain the nature of the continuum type absorption and LBL models still rely on semi-empirical formulations of the water continuum (Clough et al. 1989). For H_2O, we need to improve our knowledge of the continuum in the 10 :m window region and above all in the spectral region influenced by lower tropospheric water vapour. To this end, field measurements are of particular importance and laboratory measurements in the 7 to 8 :m region (if long cells are available) would be beneficial.

Finally, it should be noted that the accuracy of LBL computations also depend on the way molecular absorption due to heavy molecules (e.g. CFCs) is treated and on the implementation of models that account for the continuum like absorption of CO_2, N_2 and O_2.

10. Conclusions

A number of issues relevant to the fast radiative transfer models for the next generation of advanced sounders have been addressed in RTIASI, the ECMWF fast radiative transfer model for IASI.

To keep radiative transfer errors below the instrument noise, the numerical integration of the radiative transfer equation is performed on a grid of 90 vertical pressure levels. This is to be compared with the ~50 levels deemed necessary for low-resolution sounders.

Fast radiative transfer models usually assume a constant value of the Planck function (i.e. equal weight is given to the radiance emitted from all the regions within an atmospheric layer). In presence of an optically thick layer (e.g. clouds) this would result in too weight put on the radiance emitted in the lower parts of the layer. To address this issue, RTIASI parameterize the Planck function using the linear in J assumption that the Planck function is linear within the optical depth of the layer. The accuracy of the radiance computation is typically improved by 0.15K in clear sky conditions and by 1K in cloudy conditions.

Since the inclusion of minor gases in the state vector can be fully exploited by the next generation of high-resolution infrared sounders, in RTIASI the basic elements of the state vector have been supplemented by variable profiles of CO, N_2O, CH_4 and CO_2. This requires the availability of an adequate set of training profiles that represent the profile behaviour in the whole atmosphere.

To enable the assimilation of high-resolution radiances in the short wave region of the infrared spectrum, a solar term has been introduced in RTIASI that covers the spectral region from 5:m to 3.6:m and solar zenith anglesd from 0° to 87°. The introduction of the solar term has required a dedicated fast transmittance model in the shortwave to be able to cope with the large zenith angles involved in the solar geometry. The bidirectional reflectance for a wind roughened water surface is computed explicitly and it includes shadowing effects. The wave slope parameter is obtained from a wave model.

Since RTIASI uses the polychromatic approximation to solve the radiative transfer, a two-stream model for multiple scattering was found not amenable for incorporation into the code. Scattering by aerosols and clouds has been introduced by scaling the optical depth with a factor that is function of the backscattering of upward and downward radiation. This parameterization rests on the hypothesis that the diffuse radiance field is isotropic and can be approximated by the Planck function. For aerosols, errors introduced by the scaling approximation are less than 1K in the thermal infrared and less than 0.25 K in the shortwave. For water clouds, errors are less than 2K in the thermal infrared and can as be as large as 5K in the shortwave. For cirrus clouds, errors are remarkably low never exceeding 0.5K.

Finally, to solve the radiative transfer in presence of partially cloudy atmospheric layers we have followed an approach that divides the satellite field of view into a number of columns (or streams) that contain either cloud-free or completely cloudy layers. Each column is assigned a fractional area coverage and the total radiance is obtained by performing a weighted sum of the radiance in each column.

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