Chapter 2
Radiation

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2.1 RADIATIVE HEATING

The radiative heating rate is computed as the divergence of net radiation fluxes $\mathcal{F}$ so that

$$\left(\frac{\partial T}{\partial t}\right)_{\text{rad}} = -\frac{g}{c_p}\frac{\partial \mathcal{F}}{\partial p} \tag{2.1}$$

where c_p is the specific heat at constant pressure of moist air

$$c_p = c_{p_{\text{dry}}}\{1 + (c_{p_{\text{vap}}} - c_{p_{\text{vap}}})q/c_{p_{\text{dry}}}\}$$

and $c_{p_{\text{dry}}}$ and $c_{p_{\text{vap}}}$ are the specific heats at constant pressure of dry air and water vapour, respectively. Sections 2.2 and 2.3 describe the computation of the longwave and shortwave radiative fluxes respectively. The solution of the radiative transfer equation to obtain the fluxes is unfortunately very expensive, and we

cannot afford to do it more than every 3 hours at every fourth grid point. The interpolation scheme used for obtaining the radiative fluxes at every grid point and every time step for the relevant instantaneous temperature profile and solar zenith angle is described in Section 2.4.

A description of the inputs, in particular the climatologically defined quantities of radiative importance is given in Section 2.5. Finally, an alphabetical list of the subroutines of the radiation scheme is given in Section 2.6.

2.2 LONGWAVE RADIATION

Since cycle Cy22r3, two longwave radiation schemes are available in the ECMWF model, the pre-cycle Cy22r3 by Morcrette (1991), and the current longwave radiation transfer scheme, the Rapid Radiation Transfer Model (RRTM).

The rate of atmospheric cooling by emission-absorption of longwave radiation is

$$\frac{\partial T}{\partial t} = \frac{g}{c_p}\frac{\partial \mathcal{F}_{\mathrm{LW}}}{\partial p} \tag{2.2}$$

where $\mathcal{F}_{\mathrm{LW}}$ is the net longwave radiation flux (the subscript 'LW' is omitted in the remainder of this section).

Assuming a non-scattering atmosphere in local thermodynamic equilibrium, $\mathcal{F}$ is given by

$$\mathcal{F} = \int_{-1}^{1} \mu \, \mathrm{d}\mu \left[\int_{0}^{\infty} \mathrm{d}v \left\{ \mathcal{L}_v(p_{\mathrm{surf}}, \mu) t_v(p_{\mathrm{surf}}, p, \mu) + \int_{p'=p_{\mathrm{surf}}}^{0} \mathcal{L}_v(p', \mu) \, \mathrm{d}t_v \right\} \right] \tag{2.3}$$

where $\mathcal{L}_v(p, \mu)$ is the monochromatic radiance at wavenumber v at level p, propagating in a direction θ (the angle that this direction makes with the vertical), where $\mu = \cos\theta$ and $t_v(p, p'; r)$ is the monochromatic transmission through a layer whose limits are at p and p' seen under the same angle θ, with $r = \sec\theta$. The subscript 'surf' refers to the earth's surface.

Subsections 2.2.1 to 2.2.4 describe the pre-cycle Cy22r3 scheme, and Subsections 2.2.5 describes the RRTM scheme in Cy22r3.

2.2.1 The pre-cycle Cy22r3 scheme

After separating the upward and downward components (indicated by superscripts + and −, respectively), and integrating by parts, we obtain the radiation transfer equation as it is actually estimated in the longwave part of the radiation code

$$\begin{aligned} \mathcal{F}_v^+(p) &= [B_v(T_{\mathrm{surf}}) - B_v(T_{0+})] t_v(p_{\mathrm{surf}}, p; r) + B_v(T(p)) + \int_{p'=p_{\mathrm{surf}}}^{p} t_v(p, p'; r) \, \mathrm{d}B_v \\ \mathcal{F}_v^-(p) &= [B_v(T_\infty) - B_v(T_{\mathrm{top}})] t_v(p, 0; r) + B_v(T(p)) + \int_{p'=p}^{0} t_v(p', p; r) \, \mathrm{d}B_v \end{aligned} \tag{2.4}$$

where, taking benefit of the isotropic nature of the longwave radiation, the radiance L_v of (2.3) has been replaced by the Planck function $B_v(T)$ in units of flux, Wm^{-2} (here, and elsewhere, B_v is assumed to always includes the π factor). T_{surf} is the surface temperature, T_{0+} that of the air just above the surface, $T(p)$ is the temperature at pressure-level p, T_{top} that at the top of the atmospheric model. The transmission t_v is evaluated as the radiance transmission in a direction θ to the vertical such that $r = \sec\theta$ is the diffusivity factor (Elsasser, 1942). Such an approximation for the integration over the angle is usual in radiative transfer calculations, and tests on the validity of this approximation have been presented by Rodgers and Walshaw (1966) and Liu and Schmetz (1988) among others. The use of the diffusivity factor gives cooling rates within 2% of those obtained with a 4-point Gaussian quadrature.

2.2.2 Vertical integration

The integrals in (2.4) are evaluated numerically, after discretization over the vertical grid, considering the atmosphere as a pile of homogeneous layers. As the cooling rate is strongly dependent on local conditions

of temperature and pressure, and energy is mainly exchanged with the layers adjacent to the level where fluxes are calculated, the contribution of the distant layers is simply computed using a trapezoidal rule integration, but the contribution of the adjacent layers is evaluated with a 2-point Gaussian quadrature, thus at the ith level

$$\int_{p'=p_{\rm surf}}^{p_i} t_v(p,p';r)\,{\rm d}B_v = \sum_{l=1}^{2} {\rm d}B_v(l) w_l t_v(p_i,p_l;r) + \frac{1}{2}\sum_{j=1}^{i-2} {\rm d}B_v(j)[t_v(p_i,p_j;r) + t_v(p_i,p_{j-1};r)] \quad (2.5)$$

where p_l is the pressure corresponding to the Gaussian root and w_l is the Gaussian weight. ${\rm d}B_v(j)$ and ${\rm d}B_v(l)$ are the Planck function gradients calculated between two interfaces, and between mid-layer and interface, respectively.

2.2.3 Spectral integration

The integration over wavenumber v is performed using a band emissivity method, as first discussed by Rodgers (1967). The longwave spectrum is divided into six spectral regions.

(i) 0–350 cm^{-1} and 1450–1880 cm^{-1}
(ii) 500–800 cm^{-1}
(iii) 800–970 cm^{-1} and 1110–1250 cm^{-1}
(iv) 970–1110 cm^{-1}
(v) 350–500 cm^{-1}
(vi) 1250–1450 cm^{-1} and 1880–2820 cm^{-1}

corresponding to the centres of the rotation and vibration-rotation bands of H_2O, the 15 µm band of CO_2, the atmospheric window, the 9.6 µm band of O_3, the 25 µm "window" region, and the wings of the vibration-rotation band of H_2O, respectively. Over these spectral regions, band fluxes are evaluated with the help of band transmissivities precalculated from the narrow-band model of Morcrette and Fouquart (1985) – See Appendix of Morcrette *et al.* (1986) for details.

Integration of (2.4) over wavenumber ν within the kth spectral region gives the upward and downward fluxes as

$$\begin{aligned}\mathcal{F}_k^+(p) &= \{B_k(T_{\rm surf}) - B_k(T_{0+})\} t_{B_k}\{r\mathcal{U}(p_{\rm surf},p), T_{\mathcal{U}}(p_{\rm surf},p)\} + B_k(T_p) \\ &\quad + \int_{p'=p_{\rm surf}}^{p} t_{{\rm d}B_k}\{r\mathcal{U}(p,p'), T_{\mathcal{U}}(p,p')\}\,{\rm d}B_k \end{aligned} \quad (2.6)$$

$$\begin{aligned}\mathcal{F}_k^-(p) &= \{B_k(T_0) - B_k(T_\infty)\} t_{B_k}\{r\mathcal{U}(p,0), T_{\mathcal{U}}(p,0)\} - B_k(T_p) \\ &\quad - \int_{p'=p}^{0} t_{{\rm d}B_k}\{r\mathcal{U}(p',p), T_{\mathcal{U}}(p',p)\}\,{\rm d}B_k \end{aligned} \quad (2.7)$$

The formulation accounts for the different temperature dependencies involved in atmospheric flux calculations, namely that on T_p, the temperature at the level where fluxes are calculated, and that on $T_{\mathcal{U}}$, the temperature that governs the transmission through the temperature dependence of the intensity and half-widths of the lines absorbing in the concerned spectral region. The band transmissivities are non-isothermal accounting for the temperature dependence that arises from the wavenumber integration of the product of the monochromatic absorption and the Planck function. Two normalized band transmissivities are used for each absorber in a given spectral region: the first one for calculating the first right-hand-side term in (2.4), involving the boundaries; it corresponds to the weighted average of the transmission function by the Planck function

$$t_B(\overline{\mathcal{U}p}, T_p, T_{\mathcal{U}}) = \frac{\int_{v_1}^{v_2} B_v(T_p) t_v(\overline{\mathcal{U}p}, T_{\mathcal{U}})\,{\rm d}v}{\int_{v_1}^{v_2} B_v(T_p)\,{\rm d}v} \quad (2.8)$$

the second one for calculating the integral term in (2.4) is the weighted average of the transmission function by the derivative of the Planck function

$$t_{{\rm d}B}(\overline{\mathcal{U}p}, T_p, T_{\mathcal{U}}) = \frac{\int_{v_1}^{v_2} \{{\rm d}B(T_p)/{\rm d}T\} t_v(\overline{\mathcal{U}p}, T_{\mathcal{U}})\,{\rm d}v}{\int_{v_1}^{v_2} \{{\rm d}B(T_p)/{\rm d}T\}\,{\rm d}v} \quad (2.9)$$

where $\overline{\mathcal{U}p}$ is the pressure weighted amount of absorber.

The effect on absorption of the Doppler broadening of the lines (important only for pressure lower than 10 hPa) is included simply using the pressure correction method of Fels (1979). A finite line width (assumed to represent the Doppler half-width of the line) is retained under low pressure conditions where the pure Lorentz line width (proportional to pressure) would normally become negligible (Giorgetta and Morcrette, 1995).

In the scheme, the actual dependence on T_p is carried out explicitly in the Planck functions integrated over the spectral regions. Although normalized relative to $B(T_p)$ or $\mathrm{d}B(T_p)/\mathrm{d}T$, the transmissivities still depend on $T_{\mathcal{U}}$, both through Wien's displacement of the maximum of the Planck function with temperature and through the temperature dependence of the absorption coefficients. For computational efficiency, the transmissivities have been developed into Pade approximants

$$t(\overline{\mathcal{U}p}, T_u) = \frac{\sum_{i=0}^{2} c_i \mathcal{U}_{\mathrm{eff}}^{i/2}}{\sum_{j=0}^{2} d_j \mathcal{U}_{\mathrm{eff}}^{i/2}} \tag{2.10}$$

where $\mathcal{U}_{\mathrm{eff}} = r(\overline{\mathcal{U}p})\Psi(T_{\mathcal{U}}, \overline{\mathcal{U}p})$ is an effective amount of absorber which incorporates the diffusivity factor r, the weighting of the absorber amount by pressure $\overline{\mathcal{U}p}$, and the temperature dependence of the absorption coefficients. The function $\Psi(T_{\mathcal{U}}, \overline{\mathcal{U}p})$ takes the form

$$\Psi(T_{\mathcal{U}}, \overline{\mathcal{U}p}) = \exp[a(\overline{\mathcal{U}p})(T_{\mathcal{U}} - 250) + b(\overline{\mathcal{U}p})(T_{\mathcal{U}} - 250)^2] \tag{2.11}$$

The temperature dependence due to Wien's law is incorporated although there is no explicit variation of the coefficients c_i and d_j with temperature. These coefficients have been computed for temperatures between 187.5 and 312.5 K with a 12.5 K step, and transmissivities corresponding to the reference temperature the closest to the pressure weighted temperature $T_{\mathcal{U}}$ are actually used in the scheme.

2.2.4 The incorporation of the effects of clouds

The incorporation of the effects of clouds on the longwave fluxes follows the treatment discussed by Washington and Williamson (1977). Whatever the state of the cloudiness of the atmosphere, the scheme starts by calculating the fluxes corresponding to a clear-sky atmosphere and stores the terms of the energy exchange between the different levels (the integrals in (2.4)) Let $\mathcal{F}_0^+(i)$ and $\mathcal{F}_0^-(i)$ be the upward and downward clear-sky fluxes. For any cloud layer actually present in the atmosphere, the scheme then evaluates the fluxes assuming a unique overcast cloud of emissivity unity. Let $\mathcal{F}_n^+(i)$ and $\mathcal{F}_n^+(i)$ the upward and downward fluxes when such a cloud is present in the nth layer of the atmosphere. Downward fluxes above the cloud, and upward fluxes below the cloud, are assumed to be given by the clear-sky values

$$\begin{aligned} \mathcal{F}_n^+(i) &= \mathcal{F}_0^+(i) \quad \text{for } i \le n \\ \mathcal{F}_n^-(i) &= \mathcal{F}_0^-(i) \quad \text{for } i > n \end{aligned} \tag{2.12}$$

Upward fluxes above the cloud ($\mathcal{F}_n^+(k)$ for $k \le n+1$) and downward fluxes below it ($\mathcal{F}_n^-(k)$ for $k > n$) can be expressed with expressions similar to (2.5) provided the boundary terms are now replaced by terms corresponding to possible temperature discontinuities between the cloud and the surrounding air

$$\begin{aligned} \mathcal{F}_n^+(k) &= \{\mathcal{F}_{\mathrm{cld}}^+ - B(n+1)\}t(p_k, p_{n+1}; r) + B(k) + \int_{p'=p_{n-1}}^{p_k} t(p_k, p'; r)\,\mathrm{d}B \\ \mathcal{F}_n^-(k) &= \{\mathcal{F}_{\mathrm{cld}}^- - B(n)\}t(p_k, p_n; r) + B(k) + \int_{p'=p_k}^{p_n} t(p_k, p'; r)\,\mathrm{d}B \end{aligned} \tag{2.13}$$

where $B(i)$ is now the total Planck function (integrated over the whole longwave spectrum) at level i, and $\mathcal{F}_{\mathrm{cld}}^+$ and $\mathcal{F}_{\mathrm{cld}}^-$ are the longwave fluxes at the upper and lower boundaries of the cloud. Terms under the integrals correspond to exchange of energy between layers in clear-sky atmosphere and have already been computed in the first step of the calculations. This step is repeated for all cloudy layers. The fluxes for the actual atmosphere (with semi-transparent, fractional and/or multi-layered clouds) are derived from

a linear combination of the fluxes calculated in previous steps with some cloud overlap assumption in the case of clouds present in several layers. Let N be the index of the layer containing the highest cloud, $C_{\rm cld}(i)$) the fractional cloud cover in layer i, with $C_{\rm cld}(0)=1$ for the upward flux at the surface, and with $C_{\rm cld}(N+1)=1$ and $\mathcal{F}^-_{N+1}=\mathcal{F}^-_0$ to have the right boundary condition for downward fluxes above the highest cloud.

Whereas the maximum and random overlap assumptions are also available in the code (Morcrette and Fouquart, 1986), the maximum-random overlap assumption is operationally used in the ECMWF model, and the cloudy upward $\mathcal{F}^+$ and downward $\mathcal{F}^-$ fluxes are obtained as

$$\begin{aligned}
&\mathcal{F}^+(i)=\mathcal{F}^+_0(i) && \text{for } i=1\\
&\mathcal{F}^-(i)=C_{\rm cld}(i-1)\mathcal{F}^+_{i-1}(i-)+\sum_{n=0}^{i-2}C_{\rm cld}(n)\mathcal{F}^+_n(i)\prod_{l=n+1}^{i-1}\{1-C_{\rm cld}(l)\} && \text{for } 2\le i\le N+1\\
&\mathcal{F}^+(i)=C_{\rm cld}(N)\mathcal{F}^+_N(i)+\sum_{n=0}^{N-1}C_{\rm cld}(n)\mathcal{F}^+_n(i)\prod_{l=n+1}^{N}\{1-C_{\rm cld}(l)\} && \text{for } i\ge N+2
\end{aligned} \tag{2.14}$$

In the case of semi-transparent clouds, the fractional cloudiness entering the calculations is an effective cloud cover equal to the product of the emissivity due to the condensed water and the gases in the layer by the horizontal coverage of the cloud layer, with the emissivity, $\varepsilon_{\rm cld}$, related to the condensed water amount by

$$\varepsilon_{\rm cld}=1-\exp(-k_{\rm abs}\mathcal{U}_{\rm LWP}) \tag{2.15}$$

where $k_{\rm abs}$ is the condensed water mass absorption coefficient (in $\mathrm{m^2kg^{-1}}$) following Smith and Shi (1992).

2.2.5 The Rapid Radiation Transfer Model (RRTM)

As stated in Mlawer *et al.* (1997), the objective in the development of RRTM has been to obtain an accuracy in the calculation of fluxes and heating rates consistent with the best line-by-line models. It utilizes the correlated-k method and shows its filiation to the Atmospheric and Environmental Research, Inc. (AER) line-by-line model (LBLRTM, Clough *et al.*, 1989, 1992, Clough and Iacono, 1995) through its use of absorption coefficients for the relevant k-distributions derived from LBLRTM. Therefore the k-coefficients in RRTM include the effect of the CKD2.2 water vapour continuum (Clough *et al.*, 1989).

The main point in the correlated-k method (Lacis and Oinas, 1991; Fu and Liou, 1992) is the mapping of the absorption coefficient $k(\nu)$ from the spectral space (where it varies irregularly with wavenumber ν) to the g-space (where $g(k)$ is the probability distribution function, i.e. the fraction of the absorption coefficients in the set smaller than k). The effect of this reordering is a rearrangement of the sequence of terms in the integral over wavenumber in the radiative transfer equation (RTE), which makes it equivalent to what would be done for monochromatic radiation.

In the ECMWF model, no provision is presently taken for scattering in the longwave. Therefore, in order to get the downward radiance, the integration over the vertical dimension is simply done starting from the top of the atmosphere, going downward layer by layer. At the surface, the boundary condition (in terms of spectral emissivity, and potential reflection of downward radiance) is computed, then, in order to get the upward radiance, the integration over the vertical dimension is repeated, this time from the surface upward.

The spectrally averaged radiance (between ν_1 and ν_2) emerging from an atmospheric layer is

$$\bar{R}=\frac{1}{(\nu_1-\nu_2)}\int_{\nu_2}^{\nu_1}\mathrm{d}\nu\left\{R_0(\nu)+\int_{t_\nu}^{1}[B(\nu,T(t'_\nu))-R_0(\nu)]\,\mathrm{d}t'\right\} \tag{2.16}$$

where R_0 is the incoming radiance to the layer, $B(\nu,T)$ is the Planck function at wavenumber ν and temperature T, t_ν is the transmittance for the layer optical path, and t'_ν is the transmittance at a point along the optical path in the layer. Under the mapping $\nu\rightarrow g$, this becomes

$$\bar{R}=\int_0^1\mathrm{d}g\left\{B_{\rm eff}(g,T_g)+[R_0(g)-B_{\rm eff}(g,T_g)]\exp\left[-k(g,P,T)\frac{\rho\Delta z}{\cos\phi}\right]\right\} \tag{2.17}$$

where $B_{\text{eff}}(g,T)$ is an effective Planck function for the layer that varies with the layer's transmittance such as to ensure continuity of flux across layer boundaries for opaque conditions. The dependence of the transmittance is now written in terms of the absorption coefficient $k(g,P,T)$ at layer pressure P and temperature T, the absorber density ρ, the vertical thickness of the layer Δz, and the angle ϕ of the optical path.

For a given spectral interval, the domain of the variable g is partitioned into subintervals (see Table 2.6, number of g-points), each corresponding to a limited range of $k(g)$ values and for which a characteristic value κ_j of the absorption coefficient is chosen. These κ_j are then used to compute the outgoing radiance

$$\bar{R} = \sum_j W_j \left[B_{\text{eff}_j} + (R_{0j} - B_{\text{eff}_j}) \exp\left(-\kappa_j \frac{\rho \Delta z}{\cos\phi} \right) \right] \tag{2.18}$$

where W_j is the size of the sub-intervals ($\sum W_j = 1$).

The accuracy of these absorption coefficients has been established by numerous and continuing high-resolution validations of LBLRTM with spectroscopic measurements, in particular those from the Atmospheric Radiation Measurement program (ARM). Compared to the original RRTM (Mlawer *et al.*, 1997), the version used at ECMWF has been slightly modified to account for cloud optical properties and surface emissivity defined for each of the 16 bands over which spectral fluxes are computed. For efficiency reason, the original number of g-points ($256 = 16 \times 16$) has been reduced to 140 (see Table 2.6). Other changes are the use of a diffusivity approximation (instead of the three-angle integration over the zenith angle used in the original scheme) to derive upward and downward fluxes from the radiances, and the modification of the original cloud random overlapping assumption to include (to the same degree of approximation as used in the operational SW scheme) a maximum-random overlapping of cloud layers. Given the monochromatic form of the RTE, the vertical integration is simply carried out one layer at a time from the top-of-the-atmosphere to the surface to get the downward fluxes. The downward fluxes at the surface are then used with the spectral surface emissivities and the surface temperature to get the upward longwave fluxes in each of the 140 subintervals. Then the upward fluxes are obtained in a similar fashion from the surface to the ToA.

For the relevant spectral intervals of the RRTM schemes, ice cloud optical properties are derived from Ebert and Curry (1992), and water cloud optical properties from Fouquart (1987). Whereas in the previous operational scheme the cloud emissivity used to compute the effective cloud cover is defined over the whole LW spectrum from spectrally averaged mass absorption coefficients and the relevant cloud water and/or ice paths (following Smith and Shi, 1992), in RRTM, the cloud optical thickness is defined as a function of spectrally varying mass absorption coefficients and relevant cloud water and ice paths, and is used within the true cloudy fraction of the layer. Alternate sets of cloud optical properties are also available for RRTM, based on Savijarvi and Raisanen (1997) for liquid water clouds, and Fu *et al.* (1998) for ice clouds.

2.3 SHORTWAVE RADIATION

The rate of atmospheric heating by absorption and scattering of shortwave radiation is

$$\frac{\partial T}{\partial t} = \frac{g}{c_p} \frac{\partial \mathcal{F}_{\text{SW}}}{\partial p} \tag{2.19}$$

where $\mathcal{F}_{\text{SW}}$ is the net total shortwave flux (the subscript SW will be omitted in the remainder of this section).

$$\mathcal{F}(\delta) = \int_0^\infty \mathrm{d}\nu \left[\int_0^{2\pi} \mathrm{d}\phi \left\{ \int_{-1}^{+1} \mu \mathcal{L}_\nu(\delta,\mu,\phi)\, \mathrm{d}\mu \right\} \right] \tag{2.20}$$

is the diffuse radiance at wavenumber ν, in a direction given by the azimuth angle, ϕ, and the zenith angle, θ, with $\mu = \cos\theta$. In (2.20), we assume a plane parallel atmosphere, and the vertical coordinate is the optical depth δ, a convenient variable when the energy source is outside the medium

$$\delta(p) = \int_p^0 \beta_v^{\text{ext}}(p')\, \mathrm{d}p' \tag{2.21}$$

$\beta_\nu^{\text{ext}}(p)$ is the extinction coefficient, equal to the sum of the scattering coefficient β_ν^{sca} of the aerosol (or cloud particle absorption coefficient β_ν^{abs}) and the purely molecular absorption coefficient k_ν. The diffuse radiance $\mathcal{L}_\nu$ is governed by the radiation transfer equation

$$\mu\frac{\mathrm{d}\mathcal{L}_\nu(\delta,\mu,\phi)}{\mathrm{d}\delta} = \mathcal{L}_\nu(\delta,\mu,\phi) - \frac{\varpi_\nu(\delta)}{4}P_\nu(\delta,\mu,\phi,\mu_0,\phi_0)\mathcal{E}_\nu^0\exp(-\delta/\mu_r) - \frac{\varpi_\nu(\delta)}{4}\int_0^{2\pi}\mathrm{d}\phi'\left\{\int_{-1}^{+1}\Phi_\nu(\delta,\mu,\phi,\mu',\phi')\mathcal{L}_\nu(\delta,\mu',\phi')\,\mathrm{d}\mu'\right\} \tag{2.22}$$

$\mathcal{E}_\nu^0$ is the incident solar irradiance in the direction $\mu_0 = \cos\theta_0$, ϖ_ν, is the single scattering albedo $(= \beta_\nu^{\text{sca}}/k_v)$ and $\Phi(\delta,\mu,\phi,\mu',\phi')$ is the scattering phase function which defines the probability that radiation coming from direction (μ',ϕ') is scattered in direction (μ,ϕ). The shortwave part of the scheme, originally developed by Fouquart and Bonnel (1980) solves the radiation transfer equation and integrates the fluxes over the whole shortwave spectrum between 0.2 and 4 µm. Upward and downward fluxes are obtained from the reflectances and transmittances of the layers, and the photon-path-distribution method allows to separate the parametrization of the scattering processes from that of the molecular absorption.

2.3.1 Spectral integration

Solar radiation is attenuated by absorbing gases, mainly water vapour, uniformly mixed gases (oxygen, carbon dioxide, methane, nitrous oxide) and ozone, and scattered by molecules (Rayleigh scattering), aerosols and cloud particles. Since scattering and molecular absorption occur simultaneously, the exact amount of absorber along the photon path length is unknown, and band models of the transmission function cannot be used directly as in longwave radiation transfer (see Section 2.2). The approach of the photon path distribution method is to calculate the probability $\Pi(\mathcal{U})\,\mathrm{d}\mathcal{U}$ that a photon contributing to the flux $\mathcal{F}_{\text{cons}}$ in the conservative case (i.e., no absorption, $\omega_\nu = 1, k_\nu = 0$) has encountered an absorber amount between $\mathcal{U}$ and $\mathcal{U} + \mathrm{d}\mathcal{U}$. With this distribution, the radiative flux at wavenumber v is related to $\mathcal{F}_{\text{cons}}$ by

$$\mathcal{F}_v = \mathcal{F}_{\text{cons}}\int_0^\infty \Pi(\mathcal{U})\exp(-k_\nu\mathcal{U})\,\mathrm{d}\mathcal{U} \tag{2.23}$$

and the flux averaged over the spectral interval $\Delta\nu$ can then be calculated with the help of any band model of the transmission function $t_{\Delta\nu}$

$$\mathcal{F} = \frac{1}{\Delta\nu}\int_{\Delta\nu}\mathcal{F}_\nu\,\mathrm{d}\nu = \mathcal{F}_{\text{cons}}\int_0^\infty \Pi(\mathcal{U})t_{\Delta\nu}(\mathcal{U})\,\mathrm{d}\nu \tag{2.24}$$

To find the distribution function $\Pi(\mathcal{U})$, the scattering problem is solved first, by any method, for a set of arbitrarily fixed absorption coefficients k_1, thus giving a set of simulated fluxes $\mathcal{F}_{k_1}$. An inverse Laplace transform is then performed on (2.23) (Fouquart, 1974). The main advantage of the method is that the actual distribution $\Pi(\mathcal{U})$ is smooth enough that (2.23) gives accurate results even if $\Pi(\mathcal{U})$ itself is not known accurately. In fact, $\Pi(\mathcal{U})$ needs not be calculated explicitly as the spectrally integrated fluxes are

$$\begin{array}{ll}\mathcal{F} = \mathcal{F}_{\text{cons}}t_{\Delta v}(\langle\mathcal{U}\rangle) & \text{in the limiting case of weak absorption}\\ \mathcal{F} = \mathcal{F}_{\text{cons}}t_{\Delta v}(\langle\mathcal{U}^{1/2}\rangle) & \text{in the limiting case of strong absorption}\end{array}$$

where $\langle\mathcal{U}\rangle = \int_0^\infty \Pi(\mathcal{U})\mathcal{U}\,\mathrm{d}\mathcal{U}$ and $\langle\mathcal{U}^{1/2}\rangle = \int_0^\infty \Pi(\mathcal{U})\mathcal{U}^{1/2}\,\mathrm{d}\mathcal{U}$.

The atmospheric absorption in the water vapour bands is generally strong, and the scheme determines an effective absorber amount $\mathcal{U}_{\text{e}}$ between $\langle\mathcal{U}\rangle$ and $\langle\mathcal{U}^{1/2}\rangle$ derived from

$$\mathcal{U}_{\text{e}} = \ln(\mathcal{F}_{k_{\text{e}}}/\mathcal{F}_{\text{cons}})/k_{\text{e}} \tag{2.25}$$

where k_{e} is an absorption coefficient chosen to approximate the spectrally averaged transmission of the clear sky atmosphere

$$k_{\text{e}} = \frac{1}{\mathcal{U}_{\text{tot}}/\mu_0}\ln(t_{\Delta v}(\mathcal{U}_{\text{tot}}/\mu_0)) \tag{2.26}$$

where U_{tot} is the total amount of absorber in a vertical column and $\mu_0 = \cos\theta_0$. Once the effective absorber amounts of H_2O and uniformly mixed gases are found, the transmission functions are computed

using Pade approximants

$$t_{\Delta\nu}(\mathcal{U}) = \frac{\sum_{i=0}^{N} a_i \mathcal{U}^{i-1}}{\sum_{j=0}^{N} b_j \mathcal{U}^{j-1}} \tag{2.27}$$

Absorption by ozone is also taken into account, but since ozone is located at low pressure levels for which molecular scattering is small and Mie scattering is negligible, interactions between scattering processes and ozone absorption are neglected. Transmission through ozone is computed using (2.24) where $\mathcal{U}_{O_3}$ the amount of ozone is

$$\mathcal{U}^{\mathrm{d}}_{O_3} = M \int_{p}^{0} \mathrm{d}\mathcal{U}_{O_3} \qquad \text{for the downward transmission of the direct solar beam}$$

$$\mathcal{U}^{\mathrm{u}}_{O_3} = r \int_{p_s}^{0} \mathrm{d}\mathcal{U}_{O_3} + \mathcal{U}^{\mathrm{d}}_{O_3}(p_{\mathrm{surf}}) \qquad \text{for the upward transmission of the diffuse radiation}$$

$r = 1.66$ is the diffusivity factor (see Section 2.2), and M is the magnification factor (Rodgers, 1967) used instead of r to account for the sphericity of the atmosphere at very small solar elevations

$$M = 35/\sqrt{\mu_0^2 + 1} \tag{2.28}$$

To perform the spectral integration, it is convenient to discretize the solar spectrum into subintervals in which the surface reflectance, molecular absorption characteristics, and cloud optical properties can be considered as constants. One of the main causes for such a spectral variation is the sharp increase in the reflectivity of the vegetation in the near-infrared. Also, water vapour does not absorb below 0.69 μm nor do liquid water clouds. Till June 2000, the ECMWF shortwave scheme considered only two spectral intervals, one for the visible (0.2–0.69 µm), one for the near-infrared (0.69–4.00 µm) parts of the solar spectrum. From June 2000 to April 2002, the near-infrared interval was sub-divided into three intervals (0.69–1.19–2.38–4.00 µm) to account better for the spectral variations of the cloud optical properties. Till April 2002, all the molecular absorption coefficients (for O_3, H_2O, uniformly mixed gases) were derived from statistical models of the transmission function using spectroscopic parameters derived from various versions of the HITRAN database (Rothman *et al.*, 1986, 1992). In April 2002, following the recomputation of all the molecular absorption coefficients from an updated version of the shortwave line-by-line model of Dubuisson *et al.* (1996) using spectroscopic data from HAWKS (2000), the ultraviolet and visible part of the spectrum are now considered in three spectral intervals (0.20–0.25–0.69 µm) making the scheme having a total of six spectral intervals over which the aerosol and cloud optical properties are also defined. The cut-off at 0.69 µm allows the scheme to be more computational efficient, in as much as the interactions between gaseous absorption (by water vapour and uniformly mixed gases) and scattering processes are accounted for only in the near-infrared interval(s).

2.3.2 Vertical integration

Considering an atmosphere where a fraction $C^{\mathrm{tot}}_{\mathrm{cld}}$ (as seen from the surface or the top of the atmosphere) is covered by clouds (the fraction $C^{\mathrm{tot}}_{\mathrm{cld}}$ depends on which cloud-overlap assumption is assumed for the calculations), the final fluxes are given as a weighted average of the fluxes in the clear sky and in the cloudy fractions of the column

$$\mathcal{F}^{-}(j) = C^{\mathrm{tot}}_{\mathrm{cld}} \mathcal{F}^{-}_{\mathrm{cld}}(j) + (1 - C^{\mathrm{tot}}_{\mathrm{cld}}) \mathcal{F}^{-}_{\mathrm{clr}}$$

where the subscripts 'clr' and 'cld' refer to the clear-sky and cloudy fractions of the layer, respectively. In contrast to the scheme of Geleyn and Hollingsworth (1979), the fluxes are not obtained through the solution of a system of linear equations in a matrix form. Rather, assuming an atmosphere divided into homogeneous layers, the upward and downward fluxes at a given layer interface j are given by

$$\begin{aligned} \mathcal{F}^{-}(j) &= \mathcal{F}_0 \prod_{k=j}^{N} \mathcal{T}_{\mathrm{bot}}(k) \\ \mathcal{F}^{+}(j) &= \mathcal{F}^{-}(j) \mathcal{R}_{\mathrm{top}}(j-1) \end{aligned} \tag{2.29}$$

where $\mathcal{R}_{\rm top}(j)$ and $\mathcal{T}_{\rm bot}(j)$ are the reflectance at the top and the transmittance at the bottom of the jth layer. Computation of the values of $\mathcal{R}_{\rm top}$ starts at the surface and works upwards, whereas determining values of $\mathcal{T}_{\rm bot}$ starts at the top of the atmosphere and works downward. $\mathcal{R}_{\rm top}$ and $\mathcal{T}_{\rm bot}$ account for the presence of cloud in the layer by using

$$\begin{aligned}\mathcal{R}_{\rm top} &= C_{\rm cld}\mathcal{R}_{\rm cld} + (1 - C_{\rm cld})\mathcal{R}_{\rm clr}\\ \mathcal{T}_{\rm bot} &= C_{\rm cld}\mathcal{T}_{\rm cld} + (1 - C_{\rm cld})\mathcal{T}_{\rm clr}\end{aligned} \tag{2.30}$$

where $C_{\rm cld}$ is the cloud fractional coverage of the layer within the cloudy fraction $C_{\rm cld}^{\rm tot}$ of the column.

(a) Cloudy fraction layer

$\mathcal{R}_{t_{\rm cdy}}$ and $\mathcal{R}_{b_{\rm cdy}}$ are the reflectance at the top and transmittance at the bottom of the cloudy fraction of the layer calculated with the Delta-Eddington approximation. Given $\delta_{\rm c}$, $\delta_{\rm a}$, and $\delta_{\rm g}$, the optical thicknesses for the cloud, the aerosol and the molecular absorption of the gases ($= k_{\rm e}U$), respectively, and $g_{\rm c}$ and $g_{\rm a}$ the cloud and aerosol asymmetry factors, $\mathcal{R}_{t_{\rm cdy}}$ and $\mathcal{R}_{b_{\rm cdy}}$ are calculated as functions of the total optical thickness of the layer

$$\delta = \delta_{\rm c} + \delta_{\rm a} + \delta_{\rm g} \tag{2.31}$$

of the total single scattering albedo

$$\varpi^* = \frac{\delta_{\rm c} + \delta_{\rm a}}{\delta_{\rm c} + \delta_{\rm a} + \delta_{\rm g}} \tag{2.32}$$

of the total asymmetry factor

$$g^* = \frac{\delta_{\rm c}}{\delta_{\rm c} + \delta_{\rm a}} g_{\rm c} + \frac{\delta_{\rm a}}{\delta_{\rm c} + \delta_{\rm a}} g_{\rm a} \tag{2.33}$$

of the reflectance $\mathcal{R}_-$ of the underlying medium (surface or layers below the jth interface), and of the cosine of an effective solar zenith angle $\mu_{\rm eff}(j)$ which accounts for the decrease of the direct solar beam and the corresponding increase of the diffuse part of the downward radiation by the upper scattering layers

$$\mu_{\rm eff}(j) = [(1 - C_{\rm cld}^{\rm eff}(j))/\mu + rC_{\rm cld}^{\rm eff}(j)]^{-1} \tag{2.34}$$

with $C_{\rm cld}^{\rm eff}(j)$ the effective total cloudiness over level j

$$C_{\rm cld}^{\rm eff}(j) = 1 - \prod_{i=j+1}^{N} (1 - C_{\rm cld}(i)E(i)) \tag{2.35}$$

and

$$E(i) = 1 - \exp\left[-\frac{(1 - \varpi_{\rm c}(i)g_{\rm c}(i)^2)\delta_{\rm c}(i)}{\mu}\right] \tag{2.36}$$

$\delta_{\rm c}(i)$, $\varpi_{\rm c}(i)$ and $g_{\rm c}(i)$ are the optical thickness, single scattering albedo and asymmetry factor of the cloud in the ith layer, and r is the diffusivity factor. The scheme follows the Eddington approximation first proposed by Shettle and Weinman (1970), then modified by Joseph *et al.* (1976) to account more accurately for the large fraction of radiation directly transmitted in the forward scattering peak in case of highly asymmetric phase functions. Eddington's approximation assumes that, in a scattering medium of optical thickness δ^*, of single scattering albedo ω, and of asymmetry factor g, the radiance $\mathcal{L}$ entering (2.17) can be written as

$$\mathcal{L}(\delta, \mu) = \mathcal{L}_0(\delta) + \mu\mathcal{L}_1(\delta) \tag{2.37}$$

In that case, when the phase function is expanded as a series of associated Legendre functions, all terms of order greater than one vanish when (2.20) is integrated over μ and ϕ. The phase function is therefore given by

$$P(\Theta) = 1 + \beta_1(\Theta)\mu$$

where Θ is the angle between incident and scattered radiances. The integral in (2.20) thus becomes

$$\int_0^{2\pi} d\phi' \left\{ \int_{-1}^{+1} p(\mu, \phi, \mu', \phi')\mathcal{L}(\mu', \phi')\, {\rm d}\mu' \right\} = 4\pi(L_0 + \pi\mathcal{L}_1) \tag{2.38}$$

where

$$g = \frac{\beta_1}{3} = \frac{1}{2}\int_{-1}^{+1} P(\Theta)\mu \, \mathrm{d}\mu$$

is the asymmetry factor.

Using (2.38) in (2.20) after integrating over μ and dividing by 2π, we get

$$\mu\frac{\mathrm{d}}{\mathrm{d}\delta}(\mathcal{L}_0 + \mu\mathcal{L}_1) = -(\mathcal{L}_0 + \mu\mathcal{L}_1) + \varpi(\mathcal{L}_0 + g\mu\mathcal{L}_1) + 1/4\varpi\mathcal{F}_0 \exp(-\delta/\mu_0)(1 + 3g\mu_0\mu) \tag{2.39}$$

We obtain a pair of equations for $\mathcal{L}_0$ and $\mathcal{L}_1$ by integrating (2.39) over μ

$$\begin{aligned}\frac{\mathrm{d}\mathcal{L}_0}{\mathrm{d}\delta} &= -3(1-\varpi)\mathcal{L}_0 + \frac{3}{4}\varpi\mathcal{F}_0 \exp(-\delta/\mu_0)\\ \frac{\mathrm{d}\mathcal{L}_1}{\mathrm{d}\delta} &= -(1-\varpi g)\mathcal{L}_1 + \frac{3}{4}\varpi g\mu_0\mathcal{F}_0 \exp(-\delta/\mu_0)\end{aligned} \tag{2.40}$$

For the cloudy layer assumed non-conservative ($\varpi < 1$), the solutions to (2.39) and (2.40), for $0 \le \delta \le \delta^*$, are

$$\begin{aligned}\mathcal{L}_0(\delta) &= C_1 \exp(-K\delta) + C_2 \exp(+K\delta) - \alpha \exp(-\delta/\mu_0)\\ \mathcal{L}_1(\delta) &= P\{C_1 \exp(-K\delta) - C_2 \exp(+K\delta) - \beta \exp(-\delta/\mu_0)\}\end{aligned} \tag{2.41}$$

where

$$\begin{aligned}K &= \{3(1-\varpi)(1-\varpi g)\}^{1/2}\\ P &= \{3(1-\varpi)/(1-\varpi g)\}^{1/2}\\ \alpha &= 3\varpi F_0\mu_0\{1 + 3g(1-\varpi)\}/\{4(1-K^2\mu_0^2)\}\\ \beta &= 3\varpi F_0\mu_0\{1 + 3g(1-\varpi)\mu_0^2\}/(4(1-K^2\mu_0^2))\end{aligned}$$

The two boundary conditions allow to solve the system for C_1 and C_2; the downward directed diffuse flux at the top of the atmosphere is zero, that is

$$\mathcal{F}^-(0) = \left[\mathcal{L}_0(0) + \frac{2}{3}\mathcal{L}_1(0)\right] = 0$$

which translates into

$$(1 + 2P/3)C_1 + (1 - 2P/3)C_2 = \alpha + 2\beta/3 \tag{2.42}$$

The upward directed flux at the bottom of the layer is equal to the product of the downward directed diffuse and direct fluxes and the corresponding diffuse and direct reflectance ($\mathcal{R}_\mathrm{d}$ and $\mathcal{R}_-$, respectively) of the underlying medium

$$\begin{aligned}\mathcal{F}^+(\delta^*) &= \left\{\mathcal{L}_0(\delta^*) - \frac{2}{3}\mathcal{L}_1(\delta^*)\right\}\\ &= \mathcal{R}_-\left\{\mathcal{L}_0(\delta^*) + \frac{2}{3}\mathcal{L}_1(\delta^*)\right\} + \mathcal{R}_\mathrm{d}\mu_0\mathcal{F}_0 \exp(-\delta^*/\mu_0)\end{aligned}$$

which translates into

$$\begin{aligned}&\{1 - \mathcal{R}_- - 2(1+\mathcal{R}_-)P/3\}C_1 \exp(-K\delta^*) + \{1 - \mathcal{R}_- + 2(1+\mathcal{R}_-)P/3\}C_2(+K\delta^*)\\ &\quad = \{(1-\mathcal{R}_-)\alpha - 2(1+\mathcal{R}_-)\beta/3 + \mathcal{R}_\mathrm{d}\mu_0\mathcal{F}_0\} \exp(-\delta^*/\mu_0)\end{aligned} \tag{2.43}$$

In the Delta-Eddington approximation, the phase function is approximated by a Dirac delta function forward-scatter peak and a two-term expansion of the phase function

$$P(\theta) = 2f(1-\mu) + (1-f)(1 + 3g'\mu)$$

where f is the fractional scattering into the forward peak and g' the asymmetry factor of the truncated phase function. As shown by Joseph *et al.* (1976), these parameters are

$$\begin{aligned} f &= g^2 \\ g' &= g/(g+1) \end{aligned} \tag{2.44}$$

The solution of the Eddington's equations remains the same provided that the total optical thickness, single scattering albedo and asymmetry factor entering (2.39) and (2.43) take their transformed values

$$\begin{aligned} \delta'^{*} &= (1+\varpi f)\delta^{*} \\ \omega' &= \frac{(1-f)\varpi}{1-\varpi f} \end{aligned} \tag{2.45}$$

Practically, the optical thickness, single scattering albedo, asymmetry factor and solar zenith angle entering (2.39)–(2.43) are δ^{*}, ϖ^{*}, g^{*} and μ_{eff} defined in (2.33) and (2.34).

(*b*) *Clear-sky fraction of the layers*

In the clear-sky part of the atmosphere, the shortwave scheme accounts for scattering and absorption by molecules and aerosols. The following calculations are practically done twice, once for the clear-sky fraction $(1-C_{\text{cld}}^{\text{tot}})$ of the atmospheric column μ with equal to μ_0, simply modified for the effect of Rayleigh and aerosol scattering, the second time for the clear-sky fraction of each individual layer within the fraction $C_{\text{cld}}^{\text{tot}}$ of the atmospheric column containing clouds, with μ equal to μ_{e}.

As the optical thickness for both Rayleigh and aerosol scattering is small, $\mathcal{R}_{\text{clr}}(j-1)$ and $\mathcal{T}_{\text{clr}}(j)$, the reflectance at the top and transmittance at the bottom of the jth layer can be calculated using respectively a first- and a second-order expansion of the analytical solutions of the two-stream equations similar to that of Coakley Jr. and Chylek (1975). For Rayleigh scattering, the optical thickness, single scattering albedo and asymmetry factor are respectively δ_{R}, $\varpi_{\text{R}}=1$ and $g_{\text{R}}=0$, so that

$$\begin{aligned} \mathcal{R}_{\text{R}} &= \frac{\delta_{\text{R}}}{2\mu+\delta_{\text{R}}} \\ \mathcal{T}_{\text{R}} &= \frac{2\mu}{(2\mu+\delta_{\text{R}})} \end{aligned} \tag{2.46}$$

The optical thickness δ_{R} of an atmospheric layer is simply

$$\delta_{\text{R}} = \delta^{*}\{p(j)-p(j-1)\}/p_{\text{surf}} \tag{2.47}$$

where δ_{R}^{*} is the Rayleigh optical thickness of the whole atmosphere parametrized as a function of the solar zenith angle (Deschamps *et al.*, 1983)

For aerosol scattering and absorption, the optical thickness, single scattering albedo and asymmetry factor are respectively δ_{a}, ϖ_{a}, with $1-\varpi_{\text{a}} \ll 1$ and g_{a}, so that

$$\begin{aligned} \text{den} = 1 &+ \{1-\varpi_{\text{a}}+\text{back}(\mu_{\text{e}})\varpi_{\text{a}}\}(\delta_{\text{a}}/\mu_{\text{e}}) \\ &+ (1-\varpi_{\text{a}})\{1-\varpi_{\text{a}}+2\,\text{back}(\mu_{\text{e}})\varpi_{\text{a}}\}(\delta_{\text{a}}^2/\mu_{\text{e}}^2) \end{aligned} \tag{2.48}$$

$$\begin{aligned} \mathcal{R}(\mu_{\text{e}}) &= \frac{(\text{back}(\mu_{\text{e}})\varpi_{\text{a}}\delta_{\text{a}})/\mu_{\text{a}}}{\text{den}} \\ \mathcal{T}(\mu_{\text{e}}) &= 1/\text{den} \end{aligned} \tag{2.49}$$

where $\text{back}(\mu_{\text{e}}) = (2-3\mu_{\text{e}}g_{\text{a}})/4$ is the backscattering factor.

Practically, $\mathcal{R}_{\text{clr}}$ and $\mathcal{T}_{\text{clr}}$ are computed using (2.49) and the combined effect of aerosol and Rayleigh scattering comes from using modified parameters corresponding to the addition of the two scatterers with provision for the highly asymmetric aerosol phase function through Delta-approximation of the

forward scattering peak (as in (2.40) and (2.41)).

$$\begin{aligned}\delta^+ &= \delta_{\mathrm{R}} + \delta_{\mathrm{a}}(1 - \varpi_a g_a^2)\\ g^+ &= \frac{g_{\mathrm{a}}}{1+g_{\mathrm{a}}}\frac{\delta_{\mathrm{a}}}{(\delta_{\mathrm{R}}+\delta_{\mathrm{a}})}\\ \varpi^+ &= \frac{\delta_{\mathrm{R}}}{\delta_{\mathrm{a}}+\delta_{\mathrm{a}}}\varpi_{\mathrm{R}} + \frac{\delta_{\mathrm{a}}}{\delta_{\mathrm{R}}+\delta_{\mathrm{a}}}\frac{\varpi_{\mathrm{a}}(1-g_{\mathrm{a}}^2)}{1-\varpi_{\mathrm{a}}g_{\mathrm{a}}^2}\end{aligned} \tag{2.50}$$

As for their cloudy counterparts, $\mathcal{R}_{\mathrm{clr}}$ and $\mathcal{T}_{\dot{\mathrm{c}}\mathrm{lr}}$ must account for the multiple reflections due to the layers underneath

$$\mathcal{R}_{\mathrm{clr}} = \mathcal{R}(\mu_{\mathrm{e}}) + \mathcal{R}_{-}\mathcal{T}(\mu_{\mathrm{e}})(1 - \mathcal{R}^{*}\mathcal{R}_{-}) \tag{2.51}$$

and $\mathcal{R}_{-}$ is the reflectance of the underlying medium $\mathcal{R}_{-} = \mathcal{R}_t(j-1)$ and r is the diffusivity factor.

Since interactions between molecular absorption and Rayleigh and aerosol scattering are negligible, the radiative fluxes in a clear-sky atmosphere are simply those calculated from (2.27) and (2.45) attenuated by the gaseous transmissions (2.25).

2.3.3 Multiple reflections between layers

To deal properly with the multiple reflections between the surface and the cloud layers, it should be necessary to separate the contribution of each individual reflecting surface to the layer reflectance and transmittances in as much as each such surface gives rise to a particular distribution of absorber amount. In the case of an atmosphere including N cloud layers, the reflected light above the highest cloud consists of photons directly reflected by the highest cloud without interaction with the underlying atmosphere, and of photons that have passed through this cloud layer and undergone at least one reflection on the underlying atmosphere. In fact, (2.22) should be written

$$\mathcal{F} = \sum_{i=0}^{N} \mathcal{F}_{\mathrm{cl}} \int_0^{\infty} \mathcal{P}_1(\mathcal{U}) t_{\Delta v}(\mathcal{U}) \mathrm{d}v \tag{2.52}$$

where $\mathcal{F}_{\mathrm{cl}}$ and $\mathcal{P}_1(\mathcal{U})$ are the conservative fluxes and the distributions of absorber amount corresponding to the different reflecting surfaces.

Fouquart and Bonnel (1980) have shown that a very good approximation to this problem is obtained by evaluating the reflectance and transmittance of each layer (using (2.39) and (2.45)) assuming successively a non-reflecting underlying medium ($\mathcal{R}_{-} = 0$), then a reflecting underlying medium ($\mathcal{R}_{-} \neq 0$). First calculations provide the contribution to reflectance and transmittance of those photons interacting only with the layer into consideration, whereas the second ones give the contribution of the photons with interactions also outside the layer itself.

From those two sets of layer reflectance and transmittances ($\mathcal{T}_{\mathrm{t}0}$, $\mathcal{T}_{\mathrm{b}0}$) and ($\mathcal{R}_{\mathrm{t}\neq}$, $\mathcal{T}_{\mathrm{b}\neq}$) respectively, effective absorber amounts to be applied to computing the transmission functions for upward and downward fluxes are then derived using (2.23) and starting from the surface and working the formulas upward

$$\begin{aligned}\mathcal{U}_{\mathrm{e}0}^{-} &= \ln(\mathcal{T}_{\mathrm{b}0}/\mathcal{T}_{\mathrm{bc}})/k_{\mathrm{e}}\\ \mathcal{U}_{\mathrm{e}\neq}^{-} &= \ln(\mathcal{T}_{\mathrm{b}\neq}/\mathcal{T}_{\mathrm{bc}})/k_{\mathrm{e}}\\ \mathcal{U}_{\mathrm{e}0}^{+} &= \ln(\mathcal{R}_{\mathrm{t}0}/\mathcal{R}_{\mathrm{tc}})/k_{\mathrm{e}}\\ \mathcal{U}_{\mathrm{e}\neq}^{+} &= \ln(\mathcal{R}_{\mathrm{t}\neq}/\mathcal{R}_{\mathrm{tc}})/k_{\mathrm{e}}\end{aligned} \tag{2.53}$$

where $\mathcal{R}_{\mathrm{tc}}$ and $\mathcal{T}_{\mathrm{bc}}$ are the layer reflectance and transmittance corresponding to a conservative scattering medium.

Finally the upward and downward fluxes are obtained as

$$\mathcal{F}^{+}(j) = \mathcal{F}_0\{\mathcal{R}_{t0}t_{\Delta\nu}(\mathcal{U}_{\mathrm{e}0}^{+}) + (\mathcal{R}_{t\neq} - \mathcal{R}_{\mathrm{t}0})t_{\Delta\nu}(\mathcal{U}_{\mathrm{e}\neq}^{+})\} \tag{2.54}$$

$$\mathcal{F}^{-}(j) = \mathcal{F}_0\{\mathcal{T}_{b0}t_{\Delta\nu}(\mathcal{U}_{\mathrm{e}0}^{+}) + (\mathcal{T}_{b\neq} - \mathcal{T}_{\mathrm{b}0})t_{\Delta\nu}(\mathcal{U}_{\mathrm{e}\neq}^{-})\} \tag{2.55}$$

2.3.4 Cloud shortwave optical properties

As seen in Subsection 2.3.2(a), the cloud radiative properties depend on three different parameters: the optical thickness δ_c, the asymmetry factor g_c, and the single scattering albedo ϖ_c.

Presently the cloud optical properties are derived from Fouquart (1987) for the water clouds, and Ebert and Curry (1992) for the ice clouds.

The optical thickness δ_c is related to the cloud liquid water amount U_{LWP} by

$$\delta_c = \frac{3\mathcal{U}_{LWP}}{2r_e}$$

where r_e is the mean effective radius of the size distribution of the cloud water droplets. Presently r_e is parametrized as a linear function of height from 10 µm at the surface to 45 µm at the top of the atmosphere, in an empirical attempt at dealing with the variation of water cloud type with height. Smaller water droplets are observed in low-level stratiform clouds whereas larger droplets are found in mid-level cumuliform water clouds.

In the two-, four-, and six-spectral interval versions of the shortwave radiation scheme, the optical properties of liquid water clouds are defined from Fouquart (1987) and those for ice clouds from Ebert and Curry (1992). Alternative optical properties are also available for liquid water clouds (Slingo, 1989) and ice clouds (Fu, 1996).

The effective radius of the liquid water cloud particles is computed from the cloud liquid water content using the diagnostic formulation of Martin *et al.* (1994) and specified concentrations of cloud concentration nuclei over land and ocean. For ice clouds, the effective dimension of the cloud particles is diagnosed from temperature using a revision of the formulation by Ou and Liou (1995).

2.4 HORIZONTAL INTERPOLATION

As stated in the introduction, the cost of the radiation scheme described in the previous sections is prohibitive if it were used to compute the radiative fluxes at every time step and every grid point of the model.

In order to cut down the computing costs, the full radiation scheme is only called every 3 hours (every 1 hour during the first 12 hours used for data assimilation) (the so-called full radiation time steps) and on a reduced grid interpolated from the full physical grid. A spatial and temporal interpolation thus provides the relevant interaction of the shortwave radiative fluxes with the solar zenith angle at every time step and every grid point.

2.4.1 Temporal interpolation

To do so, a shortwave transmissivity is defined at each model level such that

$$\mathcal{F}_s = \tau_e \mathcal{S}_0 \tag{2.56}$$

where $\mathcal{F}_s$ is the net solar (shortwave) flux and $\mathcal{S}_0$ is the solar flux at the top of the atmosphere. $\mathcal{F}_s$ is defined only for a full radiation time step. At every time step, the net solar fluxes are computed therefore from the transmissivity derived for the last full radiation time step, using (2.56) with the correct solar angle for every grid point. The net longwave fluxes at kept at the values given by the full radiation calculation.

2.4.2 Spatial interpolation

Full radiation computations are now performed using the so-called halo configuration that can be defined according to needs for the various spatial resolutions.

The previous spatial sampling (operational till Cy26r1), was done only in the longitudinal direction. It was going from one out of four points prevalent in sub-tropical and tropical latitudes and reduced gradually to every point in polar areas. On output, Lagrangian cubic interpolation was used. The scheme

Table 2.1 *Possible resolutions of the new interpolation scheme for radiation computations for the various dynamical resolutions of the ECMWF forecast system.*

Res	95	159	255	319	399	511	639	799	1023
NDLON	192	320	512	640	800	1024	1280	1600	2048
RadRes									
2	**95**	**95**	159	255	255	399	399	511	799
3	21	63	**95**	**159**	**159**	**255**	**319**	**399**	**511**
4	N/A	N/A	63	95	95	159	159	255	399

NDLON is the maximum number of longitude points for the reference configuration with radiative (and other physics) computations at all grid points, 2, 3 and 4 correspond to a larger grid for radiative computations. Default values for model configurations from Cy26r3 are in bold. Note that default T95 does not use a larger grid for radiation. The maximum number of longitude points for the radiative computations can be obtained from the equivalent value of Res. A maximum of 42 and 128 longitude points is respectively used for radiative computations for RadRes = 21 and 63.

Table 2.2 *Speed-up factor of the various radiation configurations relative to a computation at all grid points (configuration 1), for different horizontal resolutions.*

Res	95	159	255	319	511
−1	2.76	2.89	2.28	2.89	2.61
2	**1.00**	**2.47**	2.03	1.47	1.52
3	5.60	4.88	**4.23**	**3.47**	**3.48**
4	N/A	N/A	4.60	7.73	7.28

−1 is the previous operational configuration with sampling up to one point out of four in each latitude band, 2, 3 and 4 correspond to the relevant resolution in Table 2.1. Default values for model configurations from Cy26r3 are in bold.

worked efficiently on vector systems with less than 100 processors and scalar systems with about 1000 processors. The only real problem was the complexity of the message passing, a direct result of the use of a non-standard grid for radiation calculations.

The new interface for radiation computations was developed to address this complexity, and uses a standard IFS model grid, but with a coarser resolution than the current model grid. Further, interpolation between model and radiation grids are performed using the interfaces already existing within the IFS for the semi-Lagrangian interpolation, and as a result should reduce future code maintenance. By using such a standard grid for radiation computations, there is no longer a load balance issue, as each processor is given an equal number of grid points for model and radiation grids.

A new grid is computed, independent of that for the rest of the physics, over which input fields are averaged using the standard interpolation routines. Then radiation computations are done, and output fluxes are interpolated back to the reduced grid, at times of full radiation computations. This new halo-related grid can be chosen differently with the forecast application (seasonal runs, EPS, high-resolution 10-day forecasts). Table 2.1 presents the various basic model resolutions together with the resolution made available, by default, for radiation computations by the new interface, whereas Table 2.2 presents the speed-up factor introduced by the various radiation configurations corresponding to horizontal resolutions used for different applications.

2.5 INPUT TO THE RADIATION SCHEME

2.5.1 Model variables

Temperature values are needed at the boundaries of the layers, where the fluxes are computed. They are derived from the full level temperatures with a pressure weighted interpolation

$$T_{k+1/2} = T_k \frac{p_k(p_{k+1} - p_{k+1/2})}{p_{k+1/2}(p_{k+1} - p_k)} + T_{k+1} \frac{p_{k+1/2}(p_{k+1/2} - p_k)}{p_{k+1/2}(p_{k+1} - p_k)} \tag{2.57}$$

At the bottom of the atmosphere, either the surface temperature or the temperature at 2 m is used, while at the top of the atmosphere the temperature is extrapolated from the first full level and second half level temperatures.

2.5.2 Clouds

Cloud fraction, and liquid/ice water content is provided in all layers by the cloud scheme.

2.5.3 Aerosols

The aerosol climatology used in the operational model up to Cy26r1 was given as annual mean geographical distributions defined from T5 spectral coefficients, for different aerosol types, respectively, maritime, continental, urban and desert, plus a uniformly distributed stratospheric background aerosols, with fixed vertical distributions, following Tanre *et al.* (1984). In the last fifteen years, chemical and/or transport models have addressed the life cycles of various aerosol types and attempted an inventory of their spatio-temporal distributions. Out of these studies, a new climatology for the annual cycle of the aerosol distribution of various aerosol types has been compiled by Tegen *et al.* (1997), which has been implemented in the ECMWF forecast system from Cy26r3 onwards. Table 2.3 describes the characteristics of the aerosol components for each tropospheric aerosol type and Table 2.4 compares the maximum optical thicknesses in the old and new climatologies.

2.5.4 Carbon dioxide, ozone and trace gases

Carbon dioxide, methane, nitrous oxide, CFC-11 and CFC-12 have constant volume concentrations of 353 ppm, 1.72 ppm, 0.31 ppm, 280 ppt, and 484 ppt, respectively (IPCC/SACC, 1990), except in ERA-40 for the variation in concentrations is derived from (IPCC/SACC, 1995).

Two climatologies are available for the ozone distribution. In the first one (NOZOCL = 0), the ozone mixing ratio q_{O_3} depends on height, latitude, longitude and season. Its vertical distribution is assumed to be such that its integral from 0 to the pressure p is

$$\int_0^p q_{O_3} \mathrm{d}p = \frac{a}{1 + (b/p)^{3/2}} \tag{2.58}$$

The constants a and b are related to the total amount of ozone and the height of its maximum mixing ratio. They are imposed in terms of a limited series of spherical harmonics (T10) for the geographical distribution and a Fourier series for the seasonal variation. The total amount of ozone was taken from London *et al.* (1976) and the altitude of the maximum concentration was derived from Wilcox and Belmont (1977). Plots of these values can be found in the Appendix. In the second climatology (NOZOCL = 1), the ozone mixing ratio q_{O_3} depends on height, latitude and month, and is taken from Fortuin and Langematz (1994).

2.5.5 Ground albedo and emissivity

The background land albedo, α_{sb}, is interpolated to the model grid from the monthly mean values of a snow-free albedo produced for the combined 1982–1990 years. The albedo for that dataset was computed using the method of Sellers *et al.* (1996), but with new maps of soil reflectance, new values of vegetation reflectance and the biophysical parameters described in Los *et al.* (2000). More information on the original data and plots of the monthly mean albedo are shown in Chapter 10.

Spectral albedos for parallel and diffuse radiation are needed by the radiative code. In addition, the surface energy balance equation (see Chapter 3 on vertical diffusion) needs a spectrally integrated

Table 2.3 *Characteristics of the aerosol components for each tropospheric aerosol type in the new climatology for cycle CY26R3 of the ECMWF model (adapted from Hess et al. (1998)).*

Type	RH(%)	Component	Number (cm^{-1})	Volume ($\mu m^3/m^3$)	Mass ($\mu g^3/m^3$)	Density (g/cm^3)
"Continental" organic	80					
		Insoluble	4.00E-01	4.75E+06	9.49E+00	2.00
		Water soluble	7.00E+03	1.57E+07	1.99E+01	1.27
		soot	8.30E+03	4.96E+05	4.96E-01	1.00
"Maritime" sulphate	95					
		Water soluble	1.50E+03	7.45E+06	8.35E+00	1.12
		sea salt (accum.)	2.00E+01	1.64E+08	1.72E+02	1.05
		sea salt (coarse)	3.20E-03	9.85E+05	1.04E+00	1.05
"Desert" dust-like	50					
		Water soluble	2.00E+03	2.81E+06	4.00E+00	1.42
		Mineral (nuclei)	2.70E+02	2.88E+06	7.49E+00	2.60
		Mineral (accum.)	3.05E+01	6.47E+07	1.69E+02	2.60
		Mineral (coarse)	1.42E-01	1.77E+07	4.60E+01	2.60
"Urban" black carbon	80					
		Insoluble	1.50E+00	1.78E+07	3.56E+01	2.00
		Water soluble	2.80E+04	6.28E+07	7.97E+01	1.27
		Soot	1.30E+05	7.78E+06	7.78E+01	1.00

Type: First definition (e.g. continental) is the aerosol component as known within both the ECMWF model and the OPAC software; second definition (e.g. organic) is the 3D distribution to which it is linked in the climatology of Tegen *et al.* (1997). RH is the relative assumed for the computations of the relevant optical properties. The *nuclei*, *accumulation*, and *coarse* modes refer to various size ranges for the component particles.

Table 2.4 *Maximum optical thickness in the two aerosol climatologies.*

OLD	Annual	January	July	NEW
Continental	0.2	0.235	0.231	Organic
Maritime	0.05	0.099	0.232	Sulphate
Desert	1.9	0.184	1.01	Dust-like
Urban	0.1	0.039	0.039	Black carbon
Background trop.	0.03			
Background stratos	0.045	0.045	0.045	Background stratos.

Aerosol types of the new and old climatologies are paired according to the dominant components in each mix.

Table 2.5 *Diffuse and parallel albedo and window emissivity for each tile.*

Tile	1	2	3	4	5	6	7	8
Description	**Open sea**	**Sea ice**	**Interception layer**	**Low vegetation**	**Exposed snow**	**High vegetation**	**Shaded snow**	**Bare ground**
Diffuse albedo	0.06	Ebert and Curry (1993)	α_{sb}	α_{sb}	α_{sn}	α_{sb}	0.15	α_{sb}
Parallel albedo	Taylor *et al.* (1996)	Ebert and Curry (1993)	α_{sb}	α_{sb}	α_{sn}	α_{sb}	0.15	α_{sb}
Window emissivity	0.99	0.98	0.96	0.93–0.96	0.98	0.93–0.96	0.93–0.96	0.93–0.96

parallel+diffused albedo, specified for each independent surface functional unit, tile. The procedure is summarized in Table 2.5. Over open water, the surface albedo for direct parallel radiation is a fit to low-flying aircraft measurements over the ocean given by Taylor *et al.* (1996)

$$\alpha_{sp} = \frac{0.037}{1.1\mu_0^{1.4} + 0.15} \tag{2.59}$$

For sea ice, monthly values based on Ebert and Curry (1993) albedos for the Arctic Ocean are interpolated to the forecast time. The bare sea ice albedo value in Ebert and Curry is taken as a representative value for summer, and the dry snow albedo value is used for the winter months. Values for the Antarctic are shifted by six months. Separate values for visible and near-infrared spectral bands are used. The time-varying snow albedo (α_{sn}, see Chapter 7), is used for the exposed snow tile only. Finally, the average of the diffuse and parallel albedos are spectrally integrated for each tile.

The thermal emissivity of the surface outside the 800–1250 cm^{-1} spectral region is assumed to be 0.99 everywhere. In the window region, the spectral emissivity is constant for open water, sea ice, the interception layer and exposed snow tiles. For low and high vegetation and for shaded snow the emissivity depends on the water content in the top soil layer. Emissivity decreases linearly from 0.96 for soils at or above field capacity to 0.93 for soils at or below permanent wilting point. The same formulation is used for bare ground, except for desert areas ($\alpha_{sb} > 0.3$), where a value of 0.93 is used independently of the soil water content. Finally, a broadband emissivity is obtained by convolution of the spectral emissivity and the Planck function at the skin temperature.

2.5.6 Solar zenith angle

Equations to compute the annual variation of the solar constant I, the solar declination δ_s and the difference between solar time and official time can be found in Paltridge and Platt (1976). These equations are used to give the cosine of the solar angle at the ground. Because of the curvature of the earth, the zenith angle is not quite constant along the path of a sun ray. Hence the correction applied to $\mu^a 0$ to give an average μ_0 for the atmosphere is

$$\mu_0 = \frac{\frac{H}{a}}{(\mu_0^a)^2 + \frac{H}{a}\left(2 + \frac{H}{a}\right)} - (\mu_0^a)^2 \tag{2.60}$$

where a is the earth radius and H is the atmospheric equivalent height. H/a is fixed at 0.001277.

2.6 THE RADIATION CODE

Routine RADHEAT or RADHEATN (depending whether the diagnostic or prognostic cloud scheme is used) is called at every time step to compute the radiative fluxes and heating using the solar zenith angle computed in CPGLAG and emissivities and transmissivities (PEMTU, PTRSOL) computed at full radiation time steps in RADINT. or RADINTG (see Subsection 2.6.2). The other routines are called either once at the beginning of the run (SUECRAD and below) or once per full radiation step at the first row (ECRADFR and below), or at every full radiation time step for all rows. In this section, we briefly describe the function of each routine.

Table 2.6 *Spectral distribution of the absorption by atmospheric gases in RRTM.*

		Gases included	
Spectral intervals cm^{-1}	**Number of g-points**	**Troposphere**	**Stratosphere**
10–250	8	H_2O	H_2O
250–500	14	H_2O	H_2O
500–630	16	H_2O, CO_2	H_2O, CO_2
630–700	14	H_2O, CO_2	O_3, CO_2
700–820	16	H_2O, CO_2, CCl_4	O_3, CO_2, CCl_4
820–980	8	H_2O, CFC11, CFC12	CFC11, CFC12
980–1080	12	H_2O, O_3	O_3
1080–1180	8	H_2O, CFC12, CFC22	O_3, CFC12, CFC22
1180–1390	12	H_2O, CH_4	CH_4
1390–1480	6	H_2O	H_2O
1480–1800	8	H_2O	H_2O
1800–2080	8	H_2O	
2080–2250	4	H_2O, N_2O	
2250–2380	2	CO_2	CO_2
2380–2600	2	N_2O, CO_2	
2600–3000	2	H_2O, CH_4	

Note: CCl_4 and CFC22 are presented not accounted for in the ECMWF model.

2.6.1 Set-up routines

- SUECRAD provides the interface with the user, via the namelist NAERAD. It defines the constants of Table 2.6 and sets the configuration for the radiative computations (from SUPHEC).
- ECRADFR modifies the frequency of full radiative computations (from CNT4).
- SUAERLandSUAERSNset up the longwave and shortwave radiative characteristics of the aerosols (from SUECRAD).
- SUECRAD defines the geographical distribution of aerosols, in terms of spectral coefficients (from UPDTIER).
- SUAERV defines the globally averaged vertical distribution of the aerosols (from SUECRAD).
- SUCLOP sets up the longwave and shortwave radiative properties of the ice and water clouds (from SUECRAD).
- SUECOZO computes the Legendre coefficients for the ozone distribution according to the time of the year, using the Fourier coefficients defined in DATA statements (from UPDTIER).
- SULWN sets up the coefficients for the longwave radiative computations (from SUECRAD).
- SURDI sets up the concentrations of radiatively active gases and security parameters for the radiative computations (from SUECRAD).
- SUSAT sets up position and altitude of geostationary satellites in case of diagnostic simulation of radiances by the model radiation scheme (from SUECRAD).
- SUSWN sets up the coefficients for the shortwave radiative computations (from SUECRAD).
- UPDTIER updates the time for full radiative computations (from ECRADFR).
- The routines SUAERH, SUECOZO are called only once per full radiation step, at the first row.
- SURRTAB precomputes the array linking gaseous optical thickness and the transmission function (RRTM). (called from SUECRAD).
- SURRTFTR includes all coefficients related to the g-point configuration (RRTM). (called from SUECRAD).
- SURRTPK defines the limits of the spectral intervals, and the coefficients of the spectrally defined and spectrally integrated Planck functions (RRTM). (called from SUECRAD).
- SURRTRF defines the pressure and temperature reference profiles used for the tabulation of the absorption coefficients (RRTM). (called from SUECRAD).
- RRTM_CMBGBn, for each of the 16 spectral intervals, remaps the absorption coefficients from 16 to the final number of g-points (called from RRTM_INIT_140GP).

- RRTM_INIT_140GP performs the g -point reduction from 16 per band to a band-dependant number (column 2 in Table 2.1). It also computes the relative weighting for the new g-point combinations (called from SUECRAD).
- RRTM_KGBn contain the various absorption coefficients for all gases relevant to the different spectral bands.

2.6.2 Main routines

- RADINT or RADINTG is called by RADDRV to launch the full radiation computations, depending on whether the pre-Cy26r1 sampling configuration or the Cy26r1 halo configuration is used for spatial interpolation (see Subsection 2.4.2). Zonal mean diagnostic of the temperature, clouds and albedo are computed. Temperature is vertically interpolated. Depending on the value of the variable NRINT an interpolation of all input variables to a coarser grid may be carried out. It may be necessary to subdivide the latitude belt in a few parts for the actual calculation of radiative fluxes because of storage space limitations. For this reason a loop over these parts follows. Inside this loop a call to routine RADLSW provides solar and thermal fluxes for a subset of points of that latitude row. These fluxes are converted into transmissivities and emissivities and after completion of the whole latitude circle they are transferred to the full grid when the calculations are carried out with the coarse resolution (NRINT > 1).
- RADLSW is the driver routine of the solar and thermal fluxes by calling specialized routines SW for shortwave radiation and either RRTM_RRTM_140GP or LW for longwave radiation.

2.6.3 Specialized routines

- RADSRF is called from RADPAR/CALLPAR to compute surface albedo and emissivity. It computes the gridpoint diffuse and parallel spectral albedos and a spectrally integrated albedo (for postprocessing). It also computes the emissivity inside and outside the window region, and the spectrally integrated emissivity. Finally, it computes spectrally integrated tile albedos to be used by the surface energy balance routine (see Chapter 3 on vertical diffusion).
- LW organizes the longwave computation by calling in turn LWU, LWBV, LWC.
- LWU computes the effective absorber amounts including the pressure and temperature dependencies in the spectral intervals of the longwave radiation scheme.
- LWBV calls LWB and LWV.
- LWB computes the Planck function with relation to temperature for all levels and spectral intervals.
- LWV organizes the vertical integration by calling LWVN which deals with the contribution to the flux of the layers adjacent to the level of computation of flux, LWVD which deals with the contribution from the more distant layers, and LWVB which computes the contribution of the boundary terms.
- LWTT and LWTTM compute the relevant transmission functions needed in LWVN, LWVD, and LWVB.
- LWC introduces the effect of clouds on the longwave fluxes.
- SW organizes the shortwave computation by calling in turn SWU, SW1S, and SW2S.
- SWU computes the effective absorber amounts including the pressure and temperature dependencies of the absorption.
- SW1S and SW2S deal with the shortwave radiation transfer in the two spectral intervals used to describe the solar spectrum. They both call SWCLR, which deals with the conservative scattering processes (Rayleigh) and the scattering/absorption by aerosols in the totally clear sky part of the atmospheric column, then SWR which deals with the same processes for the clear sky layers in an otherwise cloudy column, and SWDE which computes the reflectivity and transmissivity of a layer including non-conservative scatterers (cloud particles) with the Delta-Eddington approximation.
- SWTT andSWTT1, computes the relevant transmission functions.
- RRTM_RRTM_140GP organizes the longwave computation by calling in turn, within a loop on the individual vertical columns, RRTM_ECRT_140GP, RRTM_SETCOEF_140GP, RRTM_GASABS1A_140GP and RRTM_RTRN1A_140GP.
- RRTM_ECRT_140GP defines the surface spectral emissivity, and the spectral aerosol thickness, and the layer absorber amounts and cloud quantities as used in RRTM.

- RRTM_SETCOEF_140GP computes the indices and frcations related to the pressure and temperature interpolations. It also calculates the values of the integrated Planck function for each spectral band at the level and layer temperatures.
- RRTM_GASABS1A_140GP launches the calculation of the spectrally defined optical thickness for gaseous absorption. It calls RRTM_TAUMOLn.
- RRTM_RTRN1A_140GP computes the downward then upward fluxes, using a diffusivity-type approximation for the angle integration. Cloud overlap is treated with a generalized maximum/random overlap method. Adjacent layers are treated with maximum overlap, non-adjacent cloud groups are treated with random overlap. For adjacent cloud layers, cloud information is carried from the previous two layers.

2.6.4 Heating rate computation

- RADHEAT or RADHEATN, depending whether the diagnostic or the prognostic cloud scheme is used, recomputes at each time step the net radiative fluxes from the layers' effective emissivity and transmissivity, using the actual temperature and solar zenith angle. It also computes the downward longwave and shortwave radiation at the surface.

APPENDIX A. LIST OF SYMBOLS

B_ν	Planck function integrated over the half sphere with the factor involving π absorbed: in units of flux (Wm^{-2})
C_{cld}	fractional cloud cover
c_p	specific heat at constant pressure of moist air
$c_{p_{\mathrm{dry}}}$	specific heat at constant pressure of dry air
$c_{p_{\mathrm{vap}}}$	specific heat at constant pressure of water vapour
$\mathcal{E}^0_\nu$	incident solar radiance in the direction θ_0
$\mathcal{F}$	radiative flux
f	fractional scattering into the forward peak
g	acceleration of gravity
g	asymmetry factor for aerosol scattering
k	absorption coefficient
$\mathcal{L}_\nu$	monchromatic radiance at wavenumber ν
M	magnification factor $(= 35/\sqrt{(\mu_0^2 + 1)})$
m_{O_3}	ozone mixing ratio
P	scattering phase function
p	pressure
$\Pi(\mathcal{U})\,\mathrm{d}\mathcal{U}$	probability of a photon encountering an absorber amount between $\mathcal{U}$ and $\mathcal{U} + \mathrm{d}\mathcal{U}$
q	specific humidity
r	diffusivity factor $(= \sec\theta)$
r_e	mean effective radius of cloud water droplets
$\mathcal{R}$	reflectance
S_0	solar flux at the top of the atmosphere
$\mathcal{T}$	transmittance
T	temperature
t_ν	monchromatic transmission at wavenumber ν
$\mathcal{U}$	absorber amount
α	surface albedo
β^{abs}_ν	cloud particle absorbtion coefficient
β^{ext}_ν	extinction coefficient
β^{sca}_ν	scattering coefficient
δ_{g}	molecular absorption of gases
δ	optical depth
$\varepsilon_{\mathrm{cld}}$	cloud emissivity
μ	$= \cos\theta$

ν	wavenumber
ϖ_ν	single scattering albedo ($= \beta_\nu^{\mathrm{sca}}/k_\nu$)
Φ	scattering phase function
φ	azimuth angle
θ	zenith angle
θ_0	direction of incident solar beam
Θ	angle between incident and scattered radiances

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