1 Introduction

A quantitative evaluation of climate change such as the global warming is impossible without a high-quality numerical model which incorporates the dynamical and physical processes of the climate system including the circulations of the momentum, energy and materials. The numerical models are also useful tools to help our understanding of the dynamics of climate system. Although there already are a number of comprehensive climate models, none of them are complete and further improvements are still needed. One of the most important points is the incompleteness of parameterization schemes of physical processes. A number of parameterization schemes are proposed for each process, but there is no common agreement which parameterization scheme is the best one.

We have developed an atmospheric part of the model, *i.e.*, an atmospheric general circulation model, which is named CCSR/NIES AGCM. The model is based on a simple global atmospheric model developed at University of Tokyo (Numaguti, 1993). The basic standpoint in the development is to build the model based on simple but sound physical basis and to be less dependent on empirical parameters. Effective model code was employed to make possible a long-term integration with high resolution. In addition, considerable attention was paid to the readability and module compatibility of the model code to enable community use of the model.

2 Description of the Model

2.1 Overview of the Model

The model is based on the global three-dimensional primitive equations and uses spectral transformation method in horizontal and grid differentiation on sigma coordinate in vertical. The physical parameterization includes a radiation scheme with two-stream k-distribution method, simplified Arakawa-Schubert cumulus scheme, prognostic cloud water scheme, turbulence closure scheme with cloud effect, orographic gravity wave drag, and a simple land-surface model. The characteristics of the model are summarized as follows.

Basic Equations :

Three-dimensional hydrostatic primitive equations on sphere with normalized pressure (σ) coordinate.

Prognostic Variables :

Horizontal velocities $\boldsymbol{v} = (u, v)$, temperature T, surface pressure p_S, specific humidity q, cloud liquid water l, soil temperature T_g, soil moisture W_g, snow amount W_y, river water storage W_r.

Discretezation :

Spectral transformation method with Gaussian grid in horizontal and an grid differentiation (Arakawa and Suarez, 1983) in vertical. Leap-frog scheme and semi-implicit scheme for time integration.

Resolution :

Variable, standard resolutions are T42 (2.8° grid) and T21 (5.6° grid) in horizontal and 20 levels and 11 levels in vertical.

Physical Processes :

- Two-stream k-distribution scheme for radiative transfer (Nakajima and Tanaka, 1986).
- Simplified Arakawa-Schubert cumulus parameterization (based on Arakawa and Schubert, 1974, Moorthi and Suarez, 1992).
- Large-scale condensation with prognostic cloud water scheme (based on Le Treut and Li, 1991).
- Orographic gravity-wave drag scheme (McFarlane, 1987).
- Mellor-Yamada level 2 turbulence scheme (Mellor and Yamada, 1974), with simple cloud effect.
- Bulk scheme for surface fluxes (Louis, 1979, Uno *et al.*, 1995)
- Multi-layer treatment of land-surface energy budget.
- Bucket model for land-surface hydrology.
- River runoff routing model (Kanae *et al.*, 1995).

Optional Features :

- Plug-in-compatible alternative physical parameterization schemes
- Transport of arbitrary number of scaler variables including transport by cumulus convection.
- Thermodynamical ocean mixed layer model with prognostic sea ice.
- One-dimensional physics-only model.
- Two-dimensional longitude-height model.
- Two-dimensional latitude-height model.
- Divergent and non-divergent barotropic model.

2.2 Prognostic Variables and Equations

The prognostic variables of the atmospheric part of the model are zonal and meridional velocity (u, v), temperature T, surface pressure p_S and mixing ratios of arbitrary number of components q_i, including water vapor (specific humidity) q, and cloud liquid water l. In addition to them, there are prognostic variables in the land surface sub-model, namely soil temperature T_g, soil moisture W_g, snow amount W_y, and river water storage W_r.

Atmospheric prognostic equations are zonal and meridional momentum equations, thermodynamic equation, and continuity equations of total mass and components. The integration of these prognostic equations are divided into two steps. The first is the "dynamics step" and treats adiabatic advective processes in resolvable scale. The second is the "physics step" and treats other processes, namely diabatic heating, source and sink of materials, and advective processes in subgrid un-resolvable scale:

$$\frac{\partial \chi}{\partial t} = \left(\frac{\partial \chi}{\partial t}\right)_{dyn} + \left(\frac{\partial \chi}{\partial t}\right)_{phys} . \tag{1}$$

In descretized form,

$$\begin{aligned} \chi^{t+\Delta t} &= \chi^{t-\Delta t} + 2\Delta t \left(\frac{\partial \chi}{\partial t}\right)_{dyn} + 2\Delta t \left(\frac{\partial \chi}{\partial t}\right)_{phys} \\ &= \tilde{\chi}^{t+\Delta t} + 2\Delta t \left(\frac{\partial \chi}{\partial t}\right)_{phys} . \end{aligned} \tag{2}$$

In the dynamics step, the zonal and meridional momentum equations are converted into the vorticity and divergence equations. The continuity equation of total mass is converted into the tendency equation of surface pressure by use of σ vertical coordinate. In the physics step, the zonal and meridional momentum equations, the thermodynamic equation, the equations of total mass, and the equations of mixing ratio of the components are integrated in each vertical one-dimensional column.

2.3 Model Dynamics

In the dynamics step, tendencies by advective processes are calculated and time integration is done without the effect of physical processes. The time integration are done by use of semi-implicit method (*e.g.*, Bourke, 1988). In this method, the tendencies in the dynamics step are divided into three categories. The first is the linear advective terms and represent the contribution of linear gravity wave-type motions. The second is the nonlinear advective terms and represent the contribution of nonlinear advection including Coriolis and metric effect. The last is the horizontal diffusion terms. The linear advective terms are treated in trapezoidal implicit scheme, the nonlinear advective terms are treated in leap-frog scheme, and the horizontal diffusion terms are treated in the backward scheme.

$$\tilde{\chi}^{t+\Delta t} = \chi^{t-\Delta t} + 2\Delta t \left[\frac{1}{2} \left\{ \left(\frac{\partial \chi}{\partial t} \right)_{linear}^{t-\Delta t} + \left(\frac{\partial \chi}{\partial t} \right)_{linear}^{t+\Delta t} \right\} + \left(\frac{\partial \chi}{\partial t} \right)_{NL}^{t} + \left(\frac{\partial \chi}{\partial t} \right)_{diff}^{t+\Delta t} \right] \tag{3}$$

The spectral transformation method is used in this calculation. The nonlinear advective terms are calculated in the grid space and converted into the spectral space. The prognostic variables themselves are also converted into the spectral space. Then the liner advective terms and the horizontal diffusion terms are calculated and the time integration is applied in the spectral space. After integration, the prognostic variables are converted into the grid space.

The basic equation for the dynamics steps are (Haltiner and Williams, 1980), the continuity equation:

$$\frac{\partial \pi}{\partial t} + \boldsymbol{v}_H \cdot \nabla_\sigma \pi = -\nabla_\sigma \cdot \boldsymbol{v}_H - \frac{\partial \dot{\sigma}}{\partial \sigma}, \tag{4}$$

the hydrostatic equation:

$$\frac{\partial \Phi}{\partial \sigma} = -\frac{R T_v}{\sigma}, \tag{5}$$

the vorticity and divergence equations:

$$\frac{\partial \zeta}{\partial t} = \frac{1}{a\cos\varphi}\frac{\partial A_v}{\partial \lambda} - \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}(A_u \cos\varphi) - \mathcal{D}(\zeta), \tag{6}$$

$$\frac{\partial D}{\partial t} = \frac{1}{a\cos\varphi}\frac{\partial A_u}{\partial \lambda} + \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}(A_v \cos\varphi) - \nabla_\sigma^2(\Phi + R\bar{T}\pi + E) - \mathcal{D}(D), \tag{7}$$

the thermodynamic equation:

$$\begin{aligned} \frac{\partial T}{\partial t} &= -\frac{1}{a\cos\varphi}\frac{\partial u T'}{\partial \lambda} - \frac{1}{a}\frac{\partial}{\partial \varphi}(v T' \cos\varphi) + T'D \\ &\quad -\dot{\sigma}\frac{\partial T}{\partial \sigma} + \kappa T\left(\frac{\partial \pi}{\partial t} + \boldsymbol{v}_H \cdot \nabla_\sigma \pi + \frac{\dot{\sigma}}{\sigma}\right) - \mathcal{D}(T) \end{aligned} \tag{8}$$

the equations of mixing ratio of component i :

$$\frac{\partial q_i}{\partial t} = -\frac{1}{a\cos\varphi}\frac{\partial u q_i}{\partial \lambda} - \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}(v q_i \cos\varphi) + q_i D - \dot{\sigma}\frac{\partial q_i}{\partial \sigma} - \mathcal{D}(q_i), \tag{9}$$

In these equations,

$$\pi \equiv \ln p_S, \tag{10}$$

$$T \equiv \bar{T}(\sigma) + T', \tag{11}$$

$$\zeta \equiv \frac{1}{a\cos\varphi}\frac{\partial v}{\partial \lambda} - \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}(u\cos\varphi), \tag{12}$$

$$D \equiv \frac{1}{a\cos\varphi}\frac{\partial u}{\partial\lambda} + \frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}(v\cos\varphi) \,, \tag{13}$$

$$A_u \equiv (\zeta + f)v - \dot{\sigma}\frac{\partial u}{\partial\sigma} - \frac{RT'}{a\cos\varphi}\frac{\partial\pi}{\partial\lambda} \,, \tag{14}$$

$$A_v \equiv -(\zeta + f)u - \dot{\sigma}\frac{\partial v}{\partial\sigma} - \frac{RT'}{a}\frac{\partial\pi}{\partial\varphi} \,, \tag{15}$$

$$E \equiv \frac{u^2 + v^2}{2} \,, \tag{16}$$

where λ is longitude, φ latitude, $\sigma = p/p_S$ normalized pressure, p pressure, p_S surface pressure, $\boldsymbol{v} = (u, v)$ zonal and meridional velocity, $\dot{\sigma}$ vertical velocity, $f = 2\Omega\cos\varphi$ Coriolis factor, T temperature, q_i mixing ratio of components, R atmospheric gas constant, $\kappa = R/C_p$, where C_p atmospheric specific heat in constant pressure. T_v is virtual temperature, which is defined as,

$$T_v \equiv T(1 + \epsilon_v q - l) \,, \tag{17}$$

where q is mixing ratio of water vapor (specific humidity) and l is mixing ratio of cloud liquid water, and $\epsilon_v = 0.608$ is ratio of specific mass of water vapor and atmospheric gas.

The terms $\mathcal{D}(\zeta), \mathcal{D}(D), \mathcal{D}(T), \mathcal{D}(q_i)$ are horizontal diffusion terms and represented in hyper-viscosity type formula,

$$\mathcal{D}(\chi) = -(-1)^n K \nabla^{2n} \chi \tag{18}$$

where $n = 4$ is used for default.

Applying the boundary condition of vertical velocity at top and bottom of the atmosphere,

$$\dot{\sigma} = 0 \quad at \quad \sigma = 0,\, 1. \tag{19}$$

the continuity equation is converted to the tendency equation of surface pressure,

$$\frac{\partial\pi}{\partial t} = -\int_0^1 \boldsymbol{v}_H \cdot \nabla_\sigma \pi d\sigma - \int_0^1 D d\sigma \,, \tag{20}$$

and diagnostic equation of vertical velocity,

$$\dot{\sigma} = -\sigma\frac{\partial\pi}{\partial t} - \int_0^\sigma D d\sigma - \int_0^\sigma \boldsymbol{v}_H \cdot \nabla_\sigma \pi d\sigma \,. \tag{21}$$

These equations are vertically differentiated following Arakawa and Suarez (1983).

$$\begin{aligned}(A_u)_k = {} & (\zeta_k + f)v_k - \frac{1}{2\Delta\sigma_k}[\dot{\sigma}_{k-1/2}(u_{k-1} - u_k) + \dot{\sigma}_{k+1/2}(u_k - u_{k+1})] \\ & - \frac{C_p \hat{\kappa}_k T'_{v,k}}{a\cos\varphi}\frac{\partial\pi}{\partial\lambda} \,,\end{aligned} \tag{22}$$

$$
\begin{aligned}
(A_v)_k &= -(\zeta_k + f)u_k - \frac{1}{2\Delta\sigma_k}[\dot\sigma_{k-1/2}(v_{k-1} - v_k) + \dot\sigma_{k+1/2}(v_k - v_{k+1})] \\
&\quad - \frac{C_p \hat\kappa_k T'_{v,k}}{a}\frac{\partial\pi}{\partial\varphi} , \qquad (23)
\end{aligned}
$$

where,

$$
\begin{aligned}
\hat\kappa_k &= \frac{\sigma_{k-1/2}\alpha_k + \sigma_{k+1/2}\beta_k}{\Delta\sigma_k} && (24)\\
\alpha_k &= \left(\frac{\sigma_{k-1/2}}{\sigma_k}\right)^\kappa - 1 && (25)\\
\beta_k &= 1 - \left(\frac{\sigma_{k+1/2}}{\sigma_k}\right)^\kappa , && (26)
\end{aligned}
$$

$$
\Phi_k - \Phi_{k-1} = C_p\alpha_k T_{v,k} + C_p\beta_{k-1}T_{v,k-1} . \qquad (27)
$$

$$
\dot\sigma\frac{\partial T}{\partial\sigma} = \frac{1}{\Delta\sigma_k}[\dot\sigma_{k-1/2}(\hat T_{k-1/2} - T_k) + \dot\sigma_{k+1/2}(T_k - \hat T_{k+1/2})] , \qquad (28)
$$

where,

$$
\hat T_{k-1/2} = \alpha_k\left[1 - \left(\frac{\sigma_k}{\sigma_{k-1}}\right)^\kappa\right]^{-1} T_k + \beta_{k-1}\left[\left(\frac{\sigma_{k-1}}{\sigma_k}\right)^\kappa - 1\right]^{-1} T_{k-1} , \qquad (29)
$$

$$
\begin{aligned}
\kappa T\left(\frac{\partial\pi}{\partial t} + \boldsymbol{v}_H\cdot\nabla_\sigma\pi + \frac{\dot\sigma}{\sigma}\right) &= \hat\kappa_k\boldsymbol{v}_k\cdot\nabla\pi T_{v,k} \\
&\quad - \alpha_k\sum_{l=k}^{K}(D_l + \boldsymbol{v}_l\cdot\nabla\pi)\Delta\sigma_l\frac{T_{v,k}}{\Delta\sigma_k} \\
&\quad - \beta_k\sum_{l=k+1}^{K}(D_l + \boldsymbol{v}_l\cdot\nabla\pi)\Delta\sigma_l\frac{T_{v,k}}{\Delta\sigma_k} , \qquad (30)
\end{aligned}
$$

$$
\dot\sigma\frac{\partial q}{\partial\sigma} = \frac{1}{2\Delta\sigma_k}[\dot\sigma_{k-1/2}(q_{k-1} - q_k) + \dot\sigma_{k+1/2}(q_k - q_{k+1})] , \qquad (31)
$$

$$
\dot\sigma_{k-1/2} = -\sigma_{k-1/2}\frac{\partial\pi}{\partial t} - \sum_{l=k}^{K}(D_l + \boldsymbol{v}_l\cdot\nabla\pi)\Delta\sigma_l , \qquad (32)
$$

The spectral transformation using spherical harmonic functions are applied as,

$$
\chi_{ij} \equiv \chi(\lambda_i, \mu_j) = \mathcal{R}e\sum_{m=-N}^{N}\sum_{n=|m|}^{N}\chi_n^m {Y_n^m}_{ij} , \qquad (33)
$$

and

$$
X_n^m = \frac{1}{I}\sum_{i=1}^{I}\sum_{j=1}^{J}X_{ij}{Y_n^{m*}}_{ij}w_j \equiv \mathcal{W}\{X_{ij}\} , \qquad (34)
$$

where ${Y_n^m}_{ij} \equiv Y_n^m(\lambda_i, \varphi_j)$ is spherical harmonic functions of total wavenumber n and zonal wavenumber m. In the latter formula, the Gauss-Legendre method of numerical integration is used and thus w_j is the Gaussian weight and λ_j is the Gaussian latitude. The terms including horizontal derivative can be precisely calculated in the spectral space. For example, spectral values of the vorticity and divergence calculated from the grid values of zonal and meridional wind,

$$\zeta_n^m = \mathcal{W}_x\{v_{ij}\} - \mathcal{W}_y\{u_{ij}\} \tag{35}$$

$$D_n^m = \mathcal{W}_x(u_{ij}) + \mathcal{W}_y(v_{ij}) \tag{36}$$

where $\mathcal{W}_x\{\}$ and $\mathcal{W}_y\{\}$ represents spectral transformation with horizontal differencing and defined as,

$$\mathcal{W}_x\{\chi_{ij}\} \equiv \frac{1}{I}\sum_{i=1}^{I}\sum_{j=1}^{J} im\chi_{ij}\cos\varphi_j Y_n^{m*}{}_{ij}\frac{w_j}{a(1-\mu_j^2)}, \tag{37}$$

$$\mathcal{W}_y\{\chi_{ij}\} \equiv -\frac{1}{I}\sum_{i=1}^{I}\sum_{j=1}^{J}\chi_{ij}\cos\varphi_j(1-\mu_j^2)\frac{\partial}{\partial\mu}Y_n^{m*}{}_{ij}\frac{w_j}{a(1-\mu_j^2)}. \tag{38}$$

The equations in descretized form in spectral space are:
the tendency equation of surface pressure,

$$\frac{\partial\pi_m^m}{\partial t} = -\sum_{k=1}^{K}(D_n^m)_k\Delta\sigma_k + \mathcal{W}_x\{(Z_{ij}\}, \tag{39}$$

the vorticity and divergence equations,

$$\frac{\partial\zeta_{n,k}^m}{\partial t} = \mathcal{W}_x\{(A_v)_{ijk}\} - \mathcal{W}_y\{(A_u)_{ijk}\} - (\mathcal{D}_M)_n^m\zeta_{n,k}^m, \tag{40}$$

$$\frac{\partial D_{n,k}^m}{\partial t} = \mathcal{W}_x\{(A_u)_{ijk}\}+\mathcal{W}_y\{(A_v)_{ijk}\}-\mathcal{W}\{E_{ijk}\}+\frac{n(n+1)}{a^2}(\Phi_n^m+C_p\hat{\kappa}_k\bar{T}_k\pi_n^m)-(\mathcal{D}_M)_n^m D_{n,k}^m, \tag{41}$$

the thermodynamic equation,

$$\frac{\partial T_{n,k}^m}{\partial t} = -\mathcal{W}_x\left\{u_{ijk}T'_{ijk}\right\} - \mathcal{W}_y\left\{v_{ijk}T'_{ijk}\right\} + \mathcal{W}\{H_{ijk}\} - (\tilde{\mathcal{D}}_H)_n^m T_{n,k}^m, \tag{42}$$

and the equation of each component (omitting subscript i for brevity),

$$\frac{\partial q_{n,k}^m}{\partial t} = -\mathcal{W}_x\left\{u_{ijk}q'_{ijk}\right\} - \mathcal{W}_y\left\{v_{ijk}q'_{ijk}\right\} + \mathcal{W}\{R_{ijk}\} - (\tilde{\mathcal{D}}_E)_n^m q_{n,k}^m. \tag{43}$$

Dividing the terms of linear advective terms and nonlinear advective terms (with subscript NG) using vector form $\boldsymbol{D} = \{D_k\}$, $\boldsymbol{T} = \{T_k\}$, the equations can be written as,

$$\frac{\partial\pi}{\partial t} = \left(\frac{\partial\pi}{\partial t}\right)_{NG} - \boldsymbol{C}\cdot\boldsymbol{D}, \tag{44}$$

$$\frac{\partial D}{\partial t} = \left(\frac{\partial D}{\partial t}\right)_{NG} - \nabla_\sigma^2(\Phi_S + \underline{W}T + G\pi) - \mathcal{D}_M D \ , \tag{45}$$

$$\frac{\partial T}{\partial t} = \left(\frac{\partial T}{\partial t}\right)_{NG} - \underline{h}D - \mathcal{D}_H T, \tag{46}$$

where C, G are vectors and $\underline{h}$, $\underline{W}$ are $(k \times k)$ matrices. By using expressions,

$$\delta_t \chi \equiv \frac{1}{2\Delta t}\left(\chi^{t+\Delta t} - \chi^{t-\Delta t}\right) \ , \tag{47}$$

$$\overline{\chi}^t \equiv \frac{1}{2}\left(\chi^{t+\Delta t} + \chi^{t-\Delta t}\right) \ , \tag{48}$$

they become,

$$\delta_t \pi = \left(\frac{\partial \pi}{\partial t}\right)_{NG} - C \cdot \overline{D}^t \ , \tag{49}$$

$$\delta_t D = \left(\frac{\partial D}{\partial t}\right)_{NG} - \nabla_\sigma^2(\Phi_S + \underline{W}\overline{T}^t + G\overline{\pi}^t) - \mathcal{D}_M(D^{t-\Delta t} + 2\Delta t \delta_t D) \ , \tag{50}$$

$$\delta_t T = \left(\frac{\partial T}{\partial t}\right)_{NG} - \underline{h}\overline{D}^t - \mathcal{D}_H(T^{t-\Delta t} + 2\Delta t \delta_t T) \ . \tag{51}$$

These equations are easily solved by using linear matrix calculation.

2.4 Physical Parameterizations

The physics step treats radiative transfer, clouds and large-scale condensation, cumulus convection, vertical turbulent fluxes, gravity wave drag, and land surface processes. Prognostic estimation of temperature and sea-ice volume in the oceanic mixed layer is considered as an optional feature.

The physics step is divided in three parts. In the first step, the time change of the the prognostic variables by large-scale condensation and cumulus convection are calculated. In the second step, the time change of the prognostic variables by radiative transfer, vertical turbulent fluxes, gravity wave drag, and land surface processes are calculated as,

$$\frac{\partial \chi}{\partial t} = -\frac{\partial F_\chi}{\partial z} \tag{52}$$

where F_χ represents the vertical fluxes of χ $(\rho v, \rho C_p \theta, \rho q_i)$. This equation is vertically descretized in grid difference and solved in an implicit manner. In the last step, the dry convective adjustment and adjustment of total mass are done.

2.4.1 Radiative Transfer

Radiative transfer scheme employed in CCSR/NIES AGCM is based on the two-stream discrete ordinate method and the k-distribution method (Nakajima and Tanaka, 1986). The details of the scheme is described by Nakajima *et al.* (1995). The radiative fluxes at each level interface is calculated considering solar incidence, absorption, emission and scattering by gases clouds and aerosols. The calculation of the flux is done in 18 wavelength regions. Band absorption by H_2O, CO_2, O_3, N_2O, CH_4 and 16 species of CFCs are considered by k-distribution method with one to six sub-channels in each wavelength region. Continuum absorption by H_2O, O_2, O_3 are also included. Rayleigh scattering by gases and particle scattering and absorption by clouds and aerosol particles are considered.

The spectrum of solar and terrestrial thermal radiation are divided into 18 wavelength regions (channels). The boundaries of the channels are (in cm^{-1}) 50, 250, 400, 550, 770, 990, 1100, 1400, 2000, 2500, 4000, 14500, 31500, 33000, 34500, 36000, 43000, 46000, 50000. Each channel is divided into several (one to six for standard version) subchannels by use of k-distribution method. The total number of subchannel is 37 for standard version. The radiative flux of each channel is calculated as the weighted sum of the radiative flux of each subchannel,

$$F^{\pm}(z) = \sum_{i}^{n} w_i F_i^{\pm}(z) , \tag{53}$$

where $F_i^+(z)$ and $F_i^-(z)$ are the downward and upward radiative flux components and and w_i is k-distribution weight for sub-channel i. In the selection of k-distribution subchannel and weight, the correlated k-distribution method is used and objective minimizing of number of subchannels is considered.

The radiative flux components of each sub-channel is calculated with two-stream discrete ordinate method (DOM). Each atmospheric layer is considered as a homogeneous layer. The transmissibity $T^{\pm}$, reflectivity $R^{\pm}$ and source function $\epsilon^{\pm}$ (+: for downward incidence, −: for upward incidence) of each layer for each subchannel are calculated as functions of optical depth τ, single scattering albedo ω, asymmetry factor g (the first term of Legendre expansions of phase function) cutoff factor f (the second term of them), Plank function B, solar incidence S and solar zenith angle μ_0 (Nakajima and Tanaka, 1986). In this calculation, the delta-two stream truncation (Joseph *et al.*, 1976)

$$\tau \leftarrow (1 - \omega f) , \omega \leftarrow \frac{1-f}{1-\omega f} , g \leftarrow \frac{g-f}{1-f} , \tag{54}$$

is used. The Plank function is calculated from the temperature using polynomial functional fitting and expanded as a quadratic function of optical depth in each layer (Stamnes *et al.*, 1988).

The transmissibity, reflectivity, and source function for the layer which is the combination of adjacent two layers (for example, a lower layer 1 and an upper layer 2) are calculated by the adding method as,

$$R_{1,2}^{+} = R_2^{+} + T_2^{-}(1 - R_1^{+}R_2^{-})^{-1}R_1^{+}T_2^{+} \tag{55}$$
$$T_{1,2}^{+} = T_1^{+}(1 - R_2^{+}R_1^{-})^{-1}T_2^{+} \tag{56}$$
$$\epsilon_{1,2}^{+} = \epsilon_1^{+} + T_1^{+}(1 - R_1^{+}R_2^{-})^{-1}(R_2^{-}\epsilon_1^{-} + \epsilon_2^{+}) \tag{57}$$

R^-, T^-, ϵ^- are calculated in the similar manner by exchanging the role of the upper and the lower layer in the formula. By iterating this procedure, the bulk transmissibity, reflectivity, and source function of arbitrarily thick layer can be calculated. Then the diffusive radiative fluxes of each layer interface are calculated as,

$$f_{k-1/2}^{+} = (1 - R_{k,K}^{-}R_{1,k-1}^{+})^{-1}(\epsilon_{k,K}^{+} + R_{k,K}^{-}\epsilon_{1,k-1}^{-}) \tag{58}$$
$$f_{k-1/2}^{-} = R_{1,k-1}^{+}f_{k+1/2}^{+} + \epsilon_{1,k-1}^{-} , \tag{59}$$

with boundary conditions at the top,

$$f_{k-1/2}^{+} = 0 \tag{60}$$
$$f_{k-1/2}^{-} = \epsilon_{1,K}^{-} . \tag{61}$$

The total radiative fluxes are the sum of diffusive fluxes and the direct solar fluxes,

$$F_{k-1/2}^{+} = f_{k-1/2}^{+} + \mu_0 e^{-\tau_{k,K}/\mu_0} S \tag{62}$$

and $F_{k-1/2}^{-} = f_{k-1/2}^{-}$, where $\tau_{k,K}$ is the integrated optical thickness between layer k and K.

The optical depth of each sub-channel τ_i is calculated as,

$$\tau_i = \sum_j \tau_{ij}^{g} + \sum_j \tau_j^{CON} + \sum_m \tau_m^{s} , \tag{63}$$

where τ_{ij}^{g} is optical depth of band absorption by gases j, τ_j^{CON} is optical depth of continuum absorption by gases j, and τ_m^{s} is optical depth of extinction by particle m. These optical depth are calculated as,

$$\tau_{ij}^{g} = a_{ij}(p, T) r_j , \tag{64}$$

where r_j is partial volume of gases j and the factor $a_{ij}(p,T)$ is expressed in polynomial functional fitting deduced from HITRAN and LOWTRAN database in the form,

$$a(p,T) = \exp\left\{ \sum_n \sum_m A_{nm} (\ln p)^n (T - T_0)^m \right\} . \tag{65}$$

The optical depth of continuum absorption of H_2O is calculated in proportion to square of the partial volume of H_2O

$$\tau_{CON}^{H_2O} = (A^{H_2O} + f(T)\hat{A}^{H_2O})(r_{H_2O})^2\rho\Delta z \tag{66}$$

The single scattering albedo ω, asymmetry factor g, cutoff factor f are also calculated as optical thickness-weighted mean of the each components.

Two type of cloud is considered in the flux calculation. One is the cumulus cloud and the other is large-scale cloud. In each vertical column, the fractional cloudiness of cumulus cloud is assumed to be a constant between the cloud base and the cloud top of tallest cloud and zero elsewhere. For the large-scale cloud, random overlapping of clouds is assumed. There are two options in accounting for the effect of cumulus cloud. On option is to calculate radiative flux for two cases separately, one with only cumulus cloud and one with only large-scale cloud, and calculating the weighted average of them:

$$F = C^c F^c + (1 - C^c)F^l \ , \tag{67}$$

where C^c is cumulus cloudiness, F^c is radiative flux with cumulus cloud, F^l is radiative flux with randomly overlapping large-scale clouds. The former is calculated with the cloud liquid water mixing ratio of

$$r_k^c = \frac{l_k^c}{C^c} \ . \tag{68}$$

In the calculation of F^l, two values of the transmissibility R_k of each layer is evaluated with (cloudy) and without (clear) the cloud water,

$$r_k^l = \frac{l_k}{C_k} \ , \tag{69}$$

and the average of the transmissibility R_k (function of cloud water) with the weight of cloudiness

$$\bar{R}_k = C_k R_k(r_k^l) + (1 - C_k)R_k(0) \ , \tag{70}$$

as well as the average of the reflectivity and source function are used for the flux calculation with adding method.

The other option is to mix the cumulus cloud and large-scale cloud before calculating the radiative flux. In this option, only one flux calculation is done assuming the random overlap with the averaged transmissibility,

$$\bar{R}_k = \tilde{C}_k R_k(\tilde{r}_k^l) + (1 - \tilde{C}_k)R_k(0) \ , \tag{71}$$

where mixed cloudiness $\tilde{C}_k$ and mixed cloud water $\tilde{r}_k^l$ are,

$$\tilde{C}_k = 1 - (1 - C_k)(1 - C\ C) \tag{72}$$

$$\tilde{r}_k^l = \frac{l_k + l_k^c}{\tilde{C}_k} \ . \tag{73}$$

In the calculation of radiative properties of clouds, two type of cloud particle, liquid cloud and ice cloud, are considered. The fraction of the liquid cloud f_L is the same as (80).

The calculation of radiative transfer is usually done every three hours. The fluxes of longwave region is corrected at each time step accounting for the change of surface skin temperature. The fluxes of shortwave region is assumed proportional to cosine of solar zenith angle at each time step.

2.4.2 Large-scale Condensation

The large-scale condensation scheme describes grid-scale condensation and precipitation processes and gives condensational heating, precipitation, cloud fraction as diagnostic variables and time-change of mixing ratios of components, especially for the water vapor and cloud-water. The scheme is developed based on the scheme of Le Treut anmd Li (1991).

The total water mixing ration is defied as the sum of the specific humidity q and cloud-water mixing ratio l.

$$q^t = q + l \tag{74}$$

Subgrid probability distribution of total water mixing ration in each grid box is assumed as an uniform distribution between $(1-b)\bar{q}^t$ and $(1+b)\bar{q}^t$, where $\bar{q}^t$ is the grid-averaged value. Then the fraction of the supersaturated part in the grid box C' is,

$$C' = \begin{cases} 0 & (1+b)\bar{q}^t \le q^* \\ \dfrac{(1+b)\bar{q}^t - q^*}{2b\bar{q}^t} & (1-b)\bar{q}^t < q^* < (1+b)\bar{q}^t \\ 1 & (1-b)\bar{q}^t \le q^* \end{cases} \tag{75}$$

where q^* is the saturated specific humidity. Cloud water mixing ratio is estimated as the integral of supersaturated part of the water:

$$l = \begin{cases} 0 & (1+b)\bar{q}^t \le q^* \\ \dfrac{[(1+b)\bar{q}^t - q^*]^2}{4b\bar{q}^t} & (1-b)\bar{q}^t \le q^* \le (1+b)\bar{q}^t \\ \bar{q}^t - q^* & (1-b)\bar{q}^t \ge q^* \end{cases} \tag{76}$$

Given the specific humidity q and cloud water mixing ratio l, combining these equations with the conservation of moist internal energy, $C_pT + Lq$, the new value of q and l is obtained after iterations.

Cloud water mixing ratio in the cloudy part l' is,

$$l' = \frac{l}{C'} . \tag{77}$$

It is divided into two parts, liquid part l'_L and frozen part l'_F,

$$l'_L = f_L(T)l' \tag{78}$$
$$l'_F = (1 - f_L(T))\, l' \tag{79}$$

and the liquid fraction $f_L(T)$ is parameterized as,

$$f_L(T) = e^{[-(T-T_{L0})/T_{Le}]^2} , \tag{80}$$

where $T_{L0} = 273.15$K and $T_{Le} = 18$K.

Flux of precipitating water F_P is calculated by

$$F_P = \int_0^p (P_L - E)\frac{dp}{g} + F_F . \tag{81}$$

where P_L, E, F_F are the conversion of liquid water, the evaporation of precipitating water, and the sedimentation flux of ice particles, respectively.

The conversion of cloud liquid water into precipitation P_L is parameterized as,

$$P = \frac{l_L}{\tau_L} + C_c l_L F_P , \tag{82}$$

where the first term parameterizes the autoconversion process and the second term parameterizes the correction terms. τ_L is the time scale of precipitation and given as,

$$\tau_L = \tau_0 \left\{1 - \exp\left[-\left(\frac{l'_L}{l_C}\right)^2\right]\right\}^{-1} , \tag{83}$$

where $\tau_0 = 1 \times 10^4$s and $l_C = 3 \times 10^{-4}$, while the coefficient for correction is $C_c = 1\text{m}^2/\text{kg}$. The sedimentation flux of ice particles is parameterized as,

$$F_F = V_F \rho l_F = V_F^0 (\rho l_F)^\gamma , \tag{84}$$

where $V_F^0 = 6$m/s and $\gamma = 0.17$.

Evaporation of precipitating water is parameterized as,

$$E = k_E(q_w - q)\frac{F_P}{V_T} , \tag{85}$$

where q_w is wet-bulb saturation humidity and $k_E = 1.0$.

The precipitation is regarded as solid precipitation (snow) when the wet-bulb temperature is below the freezing point (273.15K). Melting of the solid precipitation is then considered where the wet-bulb temperature exceeds the freezing point. The precipitation at the surface is regarded as snow (rain) when the wet-bulb temperature in the lowest atmospheric layer is below (over) the freezing point.

Finally, the 2/3 power of the cloud area fraction C' is considered as the horizontal cloud coverage for the radiative flux calculation.

$$C = (C')^{2/3} \tag{86}$$

2.4.3 Cumulus Convection

The cumulus parameterization scheme is based on Arakawa and Schubert (1974) with several simplifications. The tendency of large-scale field by cumulus convection is expressed in the mass flux form as:

$$\frac{\partial \bar{h}}{\partial t} = M\frac{\partial \bar{h}}{\partial z} + D(h^d - \bar{h}) , \quad (87)$$

$$\frac{\partial \bar{q}}{\partial t} = M\frac{\partial \bar{q}}{\partial z} + D(q^d + l^d - \bar{q}) , \quad (88)$$

where, $\bar{h} = C_P\bar{T} + L\bar{q} + gz, \bar{q}$ are the moist static energy and specific humidity of the large-scale filed, M is the vertical mass flux of cumulus cloud, D is the detrainment mass flux of cloud into large-scale environment, and h^d, q^d, l^d are moist static energy, specific humidity and cloud water mixing ration of the cloud air detraining into the large-scale environment.

Considering spectrum of clouds with different cloud top height, the vertical mass flux M and detrainment flux D are written as the sum of the contribution of the each cloud:

$$M(z) = \sum_i M_B^i \eta^i(z) , \quad (89)$$

$$D(z) = \sum_i M_B^i D^i(z) , \quad (90)$$

where M_B is mass flux at cloud base $z = z_B$ and η^i is the non-dimensionalized vertical distribution of mass flux. The non-dimensionalized vertical mass flux is assumed as a linear function of height (Moorthi and Suarez, 1992),

$$\eta^i(z) = \begin{cases} 1 + \lambda^i(z - z_B) & z > z_B \\ z/z_B & z \le z_B \end{cases} \quad (91)$$

where λ is entrainment rate. The thermodynamic properties in the cloud are determined as,

$$\frac{\partial}{\partial z}(\eta h^i) = \lambda^i \bar{h} , \quad (92)$$

$$\frac{\partial}{\partial z}(\eta w_a^i) = \lambda^i \bar{q} , \quad (93)$$

where h^i is the moist static energy of i-th cloud, w_a^i is the adiabatic total water mixing ratio of i-th cloud in which the precipitation is neglected. The liquid water mixing ratio are then calculated assuming that the prescribed ratio $r(z)$ of the adiabatic liquid water $w_a^i(z) - q^i(z)$ precipitates below the height z:

$$l^i(z) = (1 - r(z)) \left[w_a^i(z) - q^i(z) \right] . \quad (94)$$

The specific humidity $q^i(z)$ is the saturated specific humidity in the cloud and calculated from $h^i(z)$. The precipitating ratio $r(z)$ is parameterized as,

$$r(z) = e - (z - z_B)/h_r \tag{95}$$

where $h_r = 4000$m.

The cloud base height is determined as lifting condensation level of the surface air. The relation between the entrainment ratio λ^i and the cloud top height z_T^i is determined by the requirement that the buoyancy of the cloud,

$$B = \frac{g}{\bar{T}}(T_v^i - \bar{T}_v) \tag{96}$$

vanishes there. The mass flux η^i is zero over the cloud top height and the detrainment is occur at the cloud top,

$$D^i(z) = \eta^i(z_T^i)\delta(z - z_T^i) \ , \tag{97}$$

where $\delta(x)$ is the delta function. Given the large-scale thermodynamic state $\bar{h}$ and $\bar{q}$, the thermodynamic state of cloud h^i and q^i are calculated as a linear function of entrainment ratio λ^i. The entrainment ratio corresponding to given cloud top height z_T^i is then easily calculated with the requirement of vanishing buoyancy $B = 0$ at $z = z_T^i$.

In order to determine the cloud base mass flux M_B, the cloud work function (Arakawa and Schubert, 1974),

$$A \equiv \int_{z_B}^{z_T} B\eta dz \ , \tag{98}$$

is used. The cloud mass flux is nonzero where $A > 0$. Usually the cloud work function is reduced by the warming and drying of large-scale field with cumulus convection. The cloud mass flux is determined so as to the cloud work function vanishes in a specified time scale of τ. To estimate this, the method of virtual displacement is used. A unit mass flux M_0 is specified and the warming and drying of large-scale field is calculated. Then the cloud work function after the change of large-scale field A' is calculated and,

$$M_B = M_0 \frac{A}{A - A'} \frac{\Delta t}{\tau} \ . \tag{99}$$

A part of the precipitating water P is evaporated as,

$$E = \rho a_e (\bar{q}_w - \bar{q}) \left(\frac{P}{V_T}\right) \ , \tag{100}$$

where $\bar{q}_w$ is wet-bulb saturated specific humidity, V_T is terminal velocity of precipitation, and a_e, b_e is a constant. The standard value is taken as $a_e = 0.3$ and $V_T = 10$m/s.

A fraction f_d of the evaporation of precipitating water creates the downdraft. The downdraft start the level of minimum moist static energy $\bar{h}$ and the air which is just saturated by evaporation entrains at each level.

$$\frac{\partial M^d}{\partial z} = -f_d \frac{E}{\bar{q}_w - \bar{q}} , \tag{101}$$

The fraction f_d is assumed constant and its standard value is 0.5. Freezing and melting of precipitation is considered in the same way for the large scale condensation.

The fractional cloudiness for the calculation of radiation is parameterized as,

$$C^c = C_0 \max \left\{ \ln \frac{M_m}{M_0}, 0 \right\} , \tag{102}$$

where M_m is the maximum value of cumulus mass flux within the vertical column. The averaged liquid water mixing ratio is parameterized as,

$$l^c = \sum_i \alpha M^i l^i . \tag{103}$$

The standard value is $\alpha = 0.3$, $M_0 = 2 \times 10^{-3}$kg/m^2/s. and $C_0 = 8 \times 10^{-2}$.

Vertical transport of components other than the water is calculated similar to that of the water vapor.

2.4.4 Gravity Wave Drag

The effects of orographically exited subgrid scale internal gravity wave is parameterized following McFarlane(1987)

The vertical momentum flux of internal gravity wave is given,

$$\tau_a = \rho_a E_f z_{SD}^2 N_a \boldsymbol{v}_a , \tag{104}$$

where z_{SD} is the standard deviation of the surface height within the grid box, N_a is Brunt-Baisala frequency, and E_f is a constant. The vertical momentum flux in the atmosphere is,

$$\tau(z) = \min \left\{ \tau(z - \Delta z),\ \frac{\rho E_f F_c^2}{N} \left(\frac{\boldsymbol{v} \cdot \boldsymbol{v}_a}{|\boldsymbol{v}_a|} \right)^3 \right\}, \tag{105}$$

where F_c is critical Frude number.

2.4.5 Turbulent Fluxes within the Atmosphere

The level 2 scheme of turbulence closure model by Mellor and Yamada(1974, 1982) is used for the subgrid vertical fluxes of prognostic variables.

$$F_\chi = -K_\chi \frac{\partial \chi}{\partial z} . \tag{106}$$

The diffusion coefficient K_x is calculated by

$$K_x = l^2 \frac{\partial |v|}{\partial z} S_x \,. \tag{107}$$

Here l is mixing length and estimated as

$$l = \frac{kz}{1 + kz/l_0} \,, \tag{108}$$

where k is Kárman constant z is the height over the surface and l_0 is asymptotic mixing length and taken to be 300m. S_x is estimated as functions of Richardson number R_i

$$R_i = \frac{\frac{g}{\theta_s}\frac{\partial \theta}{\partial z}}{\left|\frac{\partial v}{\partial z}\right|^2} \,. \tag{109}$$

S_x for potential temperature and mixing ratios S_H and for momentum S_M are,

$$S_H = \left(B_1(1 - R_f)\tilde{S}_M\right)^{1/2} \tilde{S}_H \tag{110}$$

$$S_M = \left(B_1(1 - R_f)\tilde{S}_M\right)^{1/2} \tilde{S}_M \tag{111}$$

where,

$$\tilde{S}_H = \frac{\alpha_1 - \alpha_2 R_f}{1 - R_f} \tag{112}$$

$$\tilde{S}_M = \frac{\beta_1 - \beta_2 R_f}{\beta_3 - \beta_4 R_f} \tilde{S}_H \,, \tag{113}$$

Here the flux Richardson number R_f is the solution of

$$\beta_2 R_f^2 - (\beta_1 + \beta_4 R_i) R_f + \beta_2 \beta_3 \,, \tag{114}$$

and, $\alpha_1 = 3A_2\gamma_1, \alpha_2 = 3A_2(\gamma_1+\gamma_2), \beta_1 = A_1 B_1(\gamma_1 - C_1), \beta_2 = A_1[B_1(\gamma_1 - C_1) + 6A_1 + 3A_2]$, $\beta_3 = A_2 B_1 \gamma_1$, $\beta_4 = A_2[B_1(\gamma_1 + \gamma_2) - 3A_1]$, $\gamma_1 = 1/3 - 2A_1/B_1$, $\gamma_2 = B_2/B_1 + 6A_1/B_1$. The nondimensional constants are given as, $A_1 = 0.92, A_2 = 0.74, B_1 = 16.6, B_2 = 10.1, C_1 = 0.08$.

Effect of condensation for the turbulent flux is incorporated by applying a correction for the Richardson number.

$$R_i^m = R_i + aC(R_i^* - R_i) \,, \tag{115}$$

where C is cloudiness, a is a tunable parameter (0.5 in default) and R_i^* is the Richardson number with saturation equivalent potential temperature θ_e^*,

$$R_i^* = \frac{\frac{g}{\theta_s}\frac{\partial \theta_e^*}{\partial z}}{\left|\left(\frac{\partial v}{\partial z}\right)^2\right|} \,. \tag{116}$$

A minimum value of diffusion coefficient K_{min} is specified as a function of height from the surface z,

$$K_{min} = K_0 \left\{1 + a_K \left[1 - \tanh\left(\frac{z - z_c}{z_e}\right)\right]\right\} \tag{117}$$

where $K_0 = 0.15\text{m}^2/\text{s}$, $a_K = 2$, $z_c = 3000\text{m}$, and $z_e = 1000\text{m}$.

2.4.6 Turbulent Fluxes at the Surface

The vertical fluxes at the surface is estimated by a bulk formula based on Louis(1979) with modification by Uno *et al.* (1995).

$$F_{\boldsymbol{v}} = -\rho C_M |\boldsymbol{v}_a| \boldsymbol{v}_a \tag{118}$$

$$F_\theta = \rho C_p C_H |\boldsymbol{v}_a| (T_s - T_a \sigma_a^\kappa) \tag{119}$$

$$F_q = \beta F_q^P = \beta \rho C_E |\boldsymbol{v}_a| (q^*(T_s^P) - q_a) \tag{120}$$

Here T_s is the surface skin temperature described in the next subsection. The subscript a represents the lowest level of the model. β is the evaporation efficiency and represents the effect of water stress of the land surface. F_q^P is the potential evaporation calculated without the effect of soil water stress ($\beta = 1$; see next subsection). The bulk coefficients C_M, C_H, C_E are calculated from the bulk Richardson number R_{iB} and the roughness length of the surface, $z_0 = z_{0M}, z_{0H}, z_{0E}$. The bulk Richardson number R_{iB} is calculated as,

$$R_{iB} = \frac{\frac{g}{\theta_s} \Delta\theta \Delta z}{|\Delta \boldsymbol{v}|^2} . \tag{121}$$

where $\Delta\chi$ means the difference of χ between the lowest model level and the surface.

The bulk coefficients C_M, C_H, C_E are represented by,

$$C_M = F_M C_D \tag{122}$$

$$C_M = F_H g_H C_D \tag{123}$$

$$C_E = F_H g_E C_D \tag{124}$$

where,

$$C_D = \frac{k^2}{\ln\left(\frac{z_a}{z_0}\right)} \tag{125}$$

is the bulk coefficient for the neutral atmosphere. The stability factor F_M and F_H are represented as,

$$F_M = 1 - \frac{bRi_0}{1 + c_M |Ri_0|^{1/2}} \tag{126}$$

$$F_H = 1 - \frac{bRi_0}{1 + c_H |Ri_0|^{1/2}} \tag{127}$$

for $R_{iB} < 0$ (unstable case) and

$$F_M = F_H = \frac{1}{(1 + b' Ri_0)^2} \tag{128}$$

for $R_{iB} \geq 0$ (stable case). Here,

$$\begin{aligned}
Ri_0 &= g_H Ri_{iB} && (129)\\
g_H &= \left(1 + \mu_H \frac{F_H}{\sqrt{F_M}}\right) && (130)\\
g_E &= \left(1 + \mu_E \frac{F_H}{\sqrt{F_M}}\right) && (131)\\
\mu_H &= \frac{\ln\left(\frac{z_0}{z_{0H}}\right)}{\ln\left(\frac{z_a}{z_0}\right)} && (132)\\
\mu_E &= \frac{\ln\left(\frac{z_0}{z_{0E}}\right)}{\ln\left(\frac{z_a}{z_0}\right)} && (133)\\
c_M &= d_M C_D b \left(\frac{z_a}{z_0}\right)^{1/2} && (134)\\
c_H &= d_H C_D b \left(\frac{z_a}{z_0}\right)^{1/2} && (135)
\end{aligned}$$

and $b = 9.4, b' = 4.7, d_M = 7.4, d_H = 5.3$. These equations are solved iteratively.

The effect of free convective motion is incorporated to the surface wind speed $|\boldsymbol{v}_a|$ following to Miller *et al.*(1992).

$$|\boldsymbol{v}_a| = \left(u_a^2 + v_a^2 + (w^*)^2\right)^{1/2} , \tag{136}$$

where

$$w^* = (\max\{H_B(F_T/C_p + \epsilon_v T_0 F_q^P), 0\})^{1/3} , \tag{137}$$

and $H_B = 2000$m. Further a minimum value of the surface wind speed $|\boldsymbol{v}_a|$ is introduced and taken to be 4m/s.

2.4.7 Surface Submodel

The surface submodel gives the surface skin temperature T_s, the evaporation efficiency β, roughness length z_{0M}, z_{0H}, z_{0E}, and surface albedo α_s for the calculation of surface fluxes of momentum, heat, and water vapor, and radiative fluxes. In order to estimate these values considering the internal dynamics of the surface, three prognostic variables are incorporated. They are, ground temperature T_g, ground wetness W_g, snow amount W_y. Also, the river water storage W_r is prognostically determined to calculate the runoff to the ocean.

The surface skin temperature T_s on the land is implicitly determined by the energy balance at the infinitely thin skin of the surface,

$$F_T(T_s, T_a) + L\beta F_q^P + F_R(T_s) - Fg(T_s, T_g) = 0 , \tag{138}$$

where F_T, F_R, F_g are temperature (sensible heat) flux, net radiative flux, and ground heat flux at the surface. All fluxes are positive upward. The potential evaporation F_q^P is determined by the same equation with no surface water stress $\beta = 1$:

$$F_T(T_s^P, T_a) + LF_q^P(T_s^P, q_a) + F_R(T_s^P) - Fg(T_s^P, T_g) = 0 . \tag{139}$$

The ground heat flux is estimated by,

$$F_g = -K_g \frac{\partial T_g}{\partial z} , \tag{140}$$

where with heat conductivity K_g is specified depending on the surface type. The ground temperature is prognostically determined by,

$$C_g \frac{\partial T_g}{\partial t} = -\frac{\partial F_g}{\partial z} , \tag{141}$$

where C_g is specific heat of the ground and is also specified depending on the surface type. These equations are solved in vertically descretized form in implicit time integration with three layers for the standard version. The evaporation efficiency β is estimated from the ground wetness W_g and stomata resistance r_s as,

$$\beta = \min \left\{ W_g/W_{sat}, \frac{r_a}{r_a + r_s/(1+\gamma)} \right\} , \tag{142}$$

where W_{sat} is saturation ground wetness, $r_a = (C_E|v|)^{-1}$ is atmospheric surface layer resistance and $\gamma = L/C_P \partial q^* / \partial T$. The stomata resistance is specified depending on the surface type. The ground wetness W_g is prognostically determined by the "bucket model" (Manabe *et al.*, 1965),

$$\rho_w D_w \frac{\partial W_g}{\partial t} = P_l - F_q - R . \tag{143}$$

Here P_l is the precipitation in liquid form, F_q is evaporation and R is runoff flux, and ρ_w is the density of the water and D_w is the depth of the active layer of the ground. The runoff occurs whenever the ground wetness exceeds the saturation ground wetness W_s to keep $W_g \leq W_s$. The roughness length and albedo of the snow-free land surface are prescribed depending on the surface type. In default, the roughness length for heat and moisture is one tenth of that for momentum.

In the presence of snow or continental ice, the surface heat balance is modified when the solution of (138) violates the condition $T_s \leq T_m$,

$$F_T(T_m, T_a) + (L + L_m)\beta F_q^P + F_R(T_m) - Fg(T_m, T_g) = -L_m M_y , \tag{144}$$

where $T_m = 273.15\text{K}$ is the melting temperature L_m is the latent heat of fusion, and M_y is the snow melt. The snow layer is considered as a part of the uppermost layer of the soil. Thus the heat content of the first layer and the heat conductivity between the first layer and its adjacent layers are modified by the presence of snow.

$$C_g = C_s \Delta z + C_i W_y \tag{145}$$

$$F_g = -\left[(K_g/\Delta z)^{-1} + (K_y/z_y)^{-1}\right]^{-1}(T_s - T_g) \tag{146}$$

where C_s is the specific heat of surface soil, Δz the thickness of upper-most layer, C_i the specific heat of snow, K_y the heat conductivity of snow, and z_y is the snow depth,

$$z_y = \frac{W_y}{\rho_y} \tag{147}$$

where ρ_y is the density of snow, which is assumed to be $\rho_y = 400\text{kg/m}^3$. In the presence of snow, the evaporation efficiency, albedo and roughness length (for momentum, heat, water vapor) are modified as,

$$\beta = (1 - f_y)\beta_f + f_y \tag{148}$$

$$\alpha = (1 - f_y)\alpha_f + f_y\alpha_y \tag{149}$$

$$z_0 = (1 - f_y)z_{0,f} + f_y \max\{h_y z_{0,f}, z_{0,y}\} \tag{150}$$

where the subscript f represents snow-free values and the subscript y represents the values with full snow cover.

$$f_y = \min\left\{\sqrt{W_y/W_{yc}}, 1\right\} \tag{151}$$

is the snow-covered ratio of the surface, and

$$h_y = \max\{1 - z_y/z_{0,M}, 0\} \ , \tag{152}$$

where h_y is the rough measure of the ratio of surface obstacles (such as trees) exists over the snow surface, assuming that the average hight of the surface obstacles is proportional to the momentum roughness length.

The prognostic equation of the snow amount is

$$\frac{\partial W_y}{\partial t} = P_y + \delta_m f_y P_l - F_q - M_y \ , \tag{153}$$

where P_y is snow fall (solid precipitation) P_l is rain (liquid precipitation), M_y is snow melt, and $\delta_m = 1$ for non-melting condition and $\delta_m = 0$ for melting condition. The equation of ground wetness W_g is modified as,

$$\rho_w D_w \frac{\partial W_g}{\partial t} = (1 - \delta_m f_y)P_l + M_y - R \ . \tag{154}$$

When $P_l > 0$ and $\delta_m = 1$, the first-layer ground temperature is adjusted accounting for the latent heat of freezing.

The surface skin temperature on the ice-free ocean is usually externally specified as a function of location and time. The surface skin temperature over the sea ice is calculated using (144) and

$$Fg(T_s, T_g) = F_g(T_s, T_I) = k_I(T_I - T_s)/h_I\ , \tag{155}$$

here k_I is heat conductivity of the sea ice, $T_I = 271.35\text{K}$ is the freezing point of the sea ice, and h_I is the thickness of the sea ice. The evaporation efficiency of the ocean surface is unity. The albedo and roughness length of the snow-free ice-covered ocean are given constant ($0.5\ 1\times10^{-2}$m by default). The albedo of the ice-free ocean is calculated from the solar zenith angle and optical thickness of the atmosphere (Payne, 1972). The roughness length of the ice-free ocean is calculated with the surface momentum flux by the formula of Miller *et al.*(1992).

The runoff flux R is routed by the rivers. The river water storage W_r is calculated by,

$$\frac{\partial W_r}{\partial t} = R - \mathrm{div}\boldsymbol{F}_r \tag{156}$$

$\boldsymbol{F}_r$ is the river water flux which is parameterized as,

$$\boldsymbol{F}_r = v_r W_r \boldsymbol{i}\ , \tag{157}$$

where v_r is the representative speed of bulk river flow and the standard value is $v_r = 0.3$m/s. $\boldsymbol{i}$ is the unit vector of the direction of river, which is assigned to each model grid over the land.

2.4.8 Boundary Conditions

The required boundary conditions are the surface type i, surface height z, standard deviation of the surface height z_{SD}. The sea surface temperature (SST) T_{sea}, sea ice thickness h_I are also externally prescribed by monthly dataset. The daily values are used after interpolating in time. The surface type are divided into 32 categories and sampled from the 1x1degree grid vegetation type dataset of Matthews (1983). The surface height and its standard deviation is calculated from 5-minutes resolution ETOPO5 dataset.

The surface properties snow-free value of roughness z_0, albedo α, specific heat capacity C_g, heat conductivity K_g, water holding capacity $D_w W_s$, stomata resistance r_s are specified as the constants depending on the surface type.